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Refreezing on the Greenland ice sheet: a comparison of parameterizations

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Abstract

Retention and refreezing of meltwater are acknowledged to be important processes for the mass budget of polar glaciers and ice sheets. Several parameterizations of these processes exist for use in energy and mass balance models. Due to a lack of direct observations, validation of these parameterizations is difficult. In this study we compare a set of 6 refreezing parameterizations against output of the Regional Atmospheric Climate Model (RACMO2), applied to the Greenland ice sheet. In RACMO2, refreezing is explicitly calculated in a snow model that calculates vertical profiles of temperature, density and liquid water content. For consistency, the parameterizations are forced with output (surface temperature, precipitation and melt) of RACMO2. For the ice sheet-integrated amount of refreezing and its inter-annual variations, all parameterizations give similar results, especially after some tuning. However, the spatial distributions differ significantly. Results are especially sensitive to the choice of the depth of the thermally active layer, which determines the cold content of the snow in most parameterizations.

1 Introduction

The surface mass balance (SMB) of a glacier is defined as the sum of all processes adding mass to the surface (accumulation) minus all processes removing mass (ablation):

$$\mathrm{SMB} = \int_{1\,\mathrm{yr}} dt\,(C + \mathrm{RF} - \mathrm{SU_s} - \mathrm{SU_{ds}} - \mathrm{ER_{ds}} - \mathrm{RU})\,. \quad (1)$$

The most important contribution to accumulation is snowfall (C), with additional contributions of condensation and freezing of rainfall (RF). Removal of mass occurs by means of surface sublimation ($\mathrm{SU_s}$), sublimation of drifting snow ($\mathrm{SU_{ds}}$), erosion by drifting snow ($\mathrm{ER_{ds}}$), and melt and subsequent runoff (RU). Especially in the (sub)polar regions, where glaciers are usually polythermal, part of the meltwater percolates into the snow/firn and refreezes. Refreezing has been addressed by several authors, especially in relation to the estimated contribution of glaciers to sea level rise (e.g. Trabant and Mayo, 1985; Pfeffer et al., 1990, 1991; Braithwaite et al., 1994; Schneider and Jansson, 2004; Reijmer and Hock, 2008; Fausto et al., 2009). Although its importance for the Greenland Ice Sheet (GrIS) is acknowledged, refreezing estimates are scarce and cover a wide range of values (Box et al., 2006; Fettweis, 2007; Hanna et al., 2008; Ettema et al., 2009).

The process of refreezing can be split in two main components: refreezing of meltwater percolating in the cold snow/firn in spring, and refreezing of liquid water held by capillary forces when the winter cold wave penetrates the firn. Refreezing is an important process: it increases the temperature and density of the snow/firn and delays and reduces runoff, it reduces melt in the ablation zone since it delays bare ice exposure, and impacts mass balance profiles since it enhances mass accumulation around the equilibrium line and in the percolation zone above.

Most published work on refreezing refers to estimates for the GrIS, e.g. Pfeffer et al. (1991); Braithwaite et al. (1994); Fausto et al. (2009), although some estimates for individual glaciers in the Arctic are available (Trabant and Mayo, 1985; Schneider and Jansson, 2004; Reijmer and Hock, 2008; Wright et al., 2007). Bøggild (2007) and Wright et al. (2007) focussed on estimating superimposed ice formation, while Schneider and Jansson (2004) and Reijmer and Hock (2008) discussed the impact of refreezing on the glacier mass balance. Given the wide range of applications and parameterizations, several authors attempted to compare the available parameterizations, most notably Janssens and Huybrechts (2000) and Wright et al. (2007). These comparisons are hampered by the scarcity of refreezing observations, although Wright et al. (2007) did compare their results with observed superimposed ice layers in ice cores.

Janssens and Huybrechts (2000) studied the spatial variability of refreezing in Greenland using different parameterizations. They report a strong dependency on the chosen depth of the thermally active layer, which in these expressions largely determines the cold content of the snow before the melting season starts. To account for the effects of refreezing below this depth requires a more comprehensive calculation of the temperature profile in the upper ice and snow layers. Several authors have explicitly incorporated the refreezing process in their energy, mass balance or (regional) climate models (Bougamont et al., 2005; Reijmer and Hock, 2008; Ettema et al., 2010b). For many climate studies involving ice sheet evolution over centuries to millennia it is, however, still too computationally expensive to explicitly include this process, and parameterizations will remain necessary. Furthermore, a comparison of different models for the GrIS shows large differences in ice sheet and annual estimated amount of refreezing, illustrating the sensitivity of estimated refreezing from explicit schemes to their exact formulation (Ettema et al., 2009).

This study aims at improving our insight in the performance of various refreezing parameterizations. We use data of a regional climate model (Ettema et al., 2009) in which refreezing is explicitly calculated. Atmospheric data (temperature, precipitation, melt) from this model are used to force the selected refreezing parameterizations. In the absence of observations we compare the results to the amount of refreezing calculated by the model.

2 Parameterizations

The amount of refreezing is limited by (1) the available energy, (2) the available pore space in the snow/firn, and (3) the available amount of water from melt, condensation, and rain. Following Janssens and Huybrechts (2000), we define P_r as the potential retention mass, which is the maximum amount of water that can be refrozen, and is determined by (1) and (2). We define W_r as the available water mass (3) and E_r as the effective retention mass, which is the actual mass refrozen in the snow. P_r, W_r and E_r are related by:

$$E_r = \min[P_r, W_r] \tag{2}$$

By defining the retention mass as outlined above it equals the amount of refrozen mass. On an annual time scale, this estimate includes the meltwater that refreezes in the cold snow in spring, the meltwater that refreezes at depth to form superimposed ice and the capillary retained water that remains in the snow pack until the end of the melt season, and subsequently refreezes in winter. Note that the meltwater that refreezes in spring and superimposed ice may melt again and run off later in the melt season.

Below we describe several published parameterizations to calculate P_r and E_r, and the modifications we made where necessary. These methods do not necessarily include all important processes or, for instance, include rain in the estimation of W_r and thus E_r. Table 1 presents the selected parameterizations and their required input fields. The fields used to force the parameterizations are described in Sect. 3. Note that all parameters referring to mass are in mm water equivalent (w.e.) unless stated otherwise.

2.1 P_max formulations

P_max formulations are the simplest way to calculate refreezing. They assume runoff to occur when the amount of refreezing exceeds a maximum fraction (P_max) of the annual snowfall (C):

$$P_\mathrm{r} = P_\mathrm{max} \cdot C \, , \tag{3}$$

where P_r is the potential retention mass. Reeh (1991) used $P_\mathrm{max} = 0.6$, so that his modelled amount of melt from the GrIS agreed with other published estimates. Later research supports this value (Braithwaite et al., 1994). P_max may be varied from 0, which is the lower bound with no refreezing possible, to 1, which represents a case in which all water may be refrozen. The latter is only a meaningful solution at the higher parts of the ice sheet. In the remainder Re1991 refers to the P_max method.

2.2 Physical based formulations

A more physically based approach was proposed by Pfeffer et al. (1991) (henceforth Pf1991). Pf1991 defines a runoff elevation h_r above which all melt water refreezes, while below this elevation all melt water runs off. This run-off elevation is determined by a combination of two requirements. The first is that for part of the melt water to run off, the amount must be large enough to remove the cold content of the snow, thus enough water must first refreeze in order to raise the snow temperature to 0 °C. The second requirement is that the melt water has to saturate the snow pore space up to the maximum value. This leads to the following condition for which runoff occurs:

$$M \geq \frac{c_i}{L_f} C |T_f| + (C - M)\left(\frac{\rho_{pc} - \rho_f}{\rho_f}\right). \tag{4}$$

Here, c_i is the heat capacity of ice that is usually assumed constant ($2050\,\mathrm{J\,kg^{-1}\,K^{-1}}$), but sometimes as a function of air temperature T_a (in K): $c_i = 152.2 + 7.122 \cdot T_a$ (Paterson, 1994). L_f is the latent heat of fusion for ice ($0.334 \times 10^6\,\mathrm{J\,kg^{-1}}$), T_f is the initial firn temperature in °C at the runoff elevation, ρ_{pc} and ρ_f are the density at pore close-off and the initial firn density, respectively (in $\mathrm{kg\,m^{-3}}$). C and M are the mean annual amount of snowfall and melt, respectively (in m w.e.). The first term on the right hand side (r.h.s.) represents the removal of cold content where C, the annual mean snowfall, represents a variable thickness of the thermally active layer. The second term describes the saturation of the pore space in the remaining annual snowfall ($C - M$), i.e. the refreezing of capillary water at the end of the melt season.

Pf1991 applied this method to the GrIS where they estimated C and M from synthesized melt and accumulation profiles. These profiles provided $h_r = 1680$ m, the elevation where the transition from refreezing to runoff occurs and T_f is then the characteristic temperature at h_r. When applied to gridded data, with ρ_{pc}, ρ_f and T_f taken constant in space and time (Table 1, values taken from Pf1991), the above condition provides us with a mask defining the area where refreezing occurs and the area where runoff occurs.

Janssens and Huybrechts (2000) (henceforth JH2000) modified the condition in Eq. (4) such that it provided P_r instead of a mask:

$$P_r = \frac{c_i}{L_f} C|T_s| + (C - M)\left(\frac{\rho_{pc} - \rho_f}{\rho_f}\right) . \tag{5}$$

Here, T_s is the annual mean surface temperature (in °C). To calculate the actual amount of refreezing on the GrIS, JH2000 additionally limited E_r to the total annual precipitation (P_{tot}): $E_r = \min[P_r, W_r] \leq P_{tot}$. The forcing in JH2000 came from a degree day model providing M, an annual temperature climatology depending on latitude, surface elevation and time of year providing T_s, and a total precipitation (P_{tot}) climatology based on a.o. ice core measurements. The input fields we use are described in Sect. 3. With small variations, Eq. (5) has been applied to e.g. the GrIS by Fausto et al. (2009) and a small glacier on Svalbard by Wright et al. (2007).

Huybrechts and de Wolde (1999) (henceforth HdW1999) and Wright et al. (2007) (henceforth Wr2007) presented parameterizations based on the same principles as Pf1991 and JH2000 but neglected the refreezing due to capillary water (2nd term r.h.s. Eq. 5). The HdW1999 condition for refreezing is given by:

$$P_r = \frac{c_i}{L_f} d_{ice}|T_s| , \tag{6}$$

where d_{ice} is the thickness of the thermally active layer. HdW1999 used a value of $d_{ice} = 2$ m w.e. based on observations at the equilibrium line in central west Greenland (Oerlemans, 1991).

In Wr2007 the energy available for refreezing, represented by T_s in the above equations, is expressed in T_s and T_w, the period averaged annual and winter surface temperature. The expression was based on the integration between standard profiles of winter and summer snow temperature (based on observations at the end of winter and summer) assuming the area between these curves to be representative for the available energy:

$$P_\mathrm{r} = \frac{c_\mathrm{i}}{L_\mathrm{f}} d_\mathrm{ice}\, 0.5\left(\left(1-\frac{\pi}{2}\right)T_\mathrm{s} - T_\mathrm{w}\right). \tag{7}$$

Here, d_ice is the maximum depth to which the annual temperature cycle penetrates, similar to the thermally active layer in Eq. (6). Wr2007 used Eq. (7) to estimate the amount of superimposed ice on a glacier on Svalbard. They obtained the best agreement with observations for $d_\mathrm{ice} = 5$ m w.e.

2.3 Energy balance formulation

In the energy balance approach the amount of refreezing is linked to the sum of available energy at the surface. Oerlemans (1991) (henceforth Oe1991) applied the method in an energy balance model for the GrIS. In this method the available energy at the surface is the sum of all energy fluxes (Q):

$$Q = \mathrm{Sw}_\mathrm{net} + \mathrm{Lw}_\mathrm{net} + \mathrm{SHF} + \mathrm{LHF}, \tag{8}$$

where Sw_net is the net short wave radiation, Lw_net is the net long wave radiation, and SHF and LHF are the turbulent fluxes of sensible and latent heat, respectively. All fluxes are in $\mathrm{W\,m^{-2}}$. The partitioning of the energy per time step that can be used for refreezing (Q_ice) is determined by the average snow temperature T_sn (in °C) of the upper 2 m of snow/firn:

$$Q_\mathrm{ice} = \max[Q, 0.](1-\exp(T_\mathrm{sn})). \tag{9}$$

Thus, when temperature decreases, a larger fraction of the energy used for melt can be re-used for heating the snow through refreezing. Oe1991 initialized the model with the annual mean surface temperature. The energy released when refreezing occurs is used to increase the snow temperature. Oe1991 calculated this process each model time step of 15 min. Using this relation we define P_r as:

$$P_\mathrm{r} = \sum_{i=1}^{12} n_i \left(\frac{Q_\mathrm{ice}(i)}{L_\mathrm{f}} \right) , \tag{10}$$

where the sum is taken over 12 months since our input consists of monthly mean values, and n_i is the number of 15 min time steps in each month. Oe1991 only applied this formulation over snow surfaces since refreezing can only occur in snow or firn. We therefore limit P_r to the total annual precipitation P_tot similar to JH2000: $E_\mathrm{r} = \min[P_\mathrm{r}, W_\mathrm{r}] \leq P_\mathrm{tot}$. We furthermore use T_s to represent T_sn. Note that we do not take the heating effect of refreezing on T_sn into account.

3 RACMO2

RACMO2 (Regional Atmospheric Climate MOdel, Van Meijgaard et al., 2008) has been successful in simulating the mass budgets of the Antarctic ice sheet and the GrIS (see e.g. Van de Berg et al., 2006; Ettema et al., 2009). For the application over the GrIS the model uses a domain that includes part of Eastern Canada, Greenland, Iceland and Svalbard, on a horizontal resolution of 11 km. The model is forced at the lateral boundaries and at the sea surface by output of ERA-40 (European Centre for Medium-Range Weather Forecasts (ECMWF) 40-yr re-analysis project), supplemented by ECMWF operational analyses, and covers the period 1958–2008. RACMO2 has been two-way coupled to a physical snow model. We refer to Ettema et al. (2010b) for a more detailed description of RACMO2, the snow model is described below. Results of the application of RACMO2 to the GrIS are published in e.g. Ettema et al. (2009, 2010a,b); Van den Broeke et al. (2009).

3.1 The coupled snow model

The snow model incorporated in RACMO2 follows Greuell and Konzelman (1994); Bougamont et al. (2005); Reijmer and Hock (2008) to calculate the process of meltwater percolation, retention, refreezing and slush formation. In this model, refreezing is limited by three factors: (1) the firn/snow temperature cannot be raised above melting point, (2) the available amount of water (melt plus rain), and (3) the available pore space. To calculate these processes, the snow model uses a vertical grid that is 30 m deep and consists of grid layers with variable thickness, ranging from 6.5 cm near the surface to 4 m at 30 m depth. The thickness of the layers is allowed to change due to melt, accumulation, evaporation, and densification. Each layer is characterized by a temperature, density, liquid water content, depth and thickness. At the snow surface the snow model is forced by the surface energy balance, which determines surface temperature T_s and, when $T_s = 273.16$ K, the energy available for melting. The temperature evolution in the snow pack is calculated based on the thermodynamic equation (Paterson, 1994). When the temperature of a snow layer increases above the melting point, it is reset to the melting point and all excess energy is used for melting. Melt and rain water are allowed to percolate into lower layers where it may refreeze, raising the temperature and density. Liquid water is retained as capillary water in the pores of the snow. The maximum amount of water retained against gravity is 2 % of the pore volume. No slush layer forms, the remaining liquid water runs off without delay. Note that the model provides E_r and W_r, not P_r.

3.2 Input data

We force the various parameterizations with the following input fields from RACMO2: monthly snowfall, melt, rain, and, depending on the parameterization, surface temperature and net surface energy budget. Annual values of P_r are then calculated and Eq. (2) applied to annual values of W_r to provide annual values of E_r. RACMO2 also provides E_r, against which the parameterizations will be evaluated. Note that the annual values are based on January to December monthly means or sums.

Figure 1 presents 1958–2008 average annual sums of snowfall (a), melt (b), rain (c) and refreezing (d). The most pronounced feature in Fig. 1a is the high snowfall over the southeast. The snowfall pattern is determined by the large scale circulation around Greenland and the ice sheet topography: the Icelandic Low advects moist oceanic air westward to the GrIS, where it rises steeply from sea level to 2.5 km height. Note that only a small part of the ice sheet receives on average more than 2 m w.e. per year and no point receives on average more than 5 m w.e. per year.

Melt and runoff mainly take place along the ice margin (Fig. 1b), with the widest melt zone in the west. Rainfall is concentrated on the southern margins of the ice sheet (Fig. 1c). The percentage of the total precipitation falling as rain can be considerable (up to 50 %), and occasionally rain occurs at elevations over 2000 m, especially on the southern part of the ice sheet. Therefore, taking rain into account may (locally) have a significant impact on the estimated refreezing (see Sect. 4.2).

Most refreezing occurs along the margins (Fig. 1d), where most melt occurs. In the larger ablation areas (western ice margin) the amount of refrozen mass is limited by the rapid removal of the snow pack in spring making pore space the limiting factor for refreezing. Most refreezing occurs on the wet south and south-eastern margins, where pore space is much larger. Due to the lack of refreezing observations, the modelled refreezing cannot be robustly validated.

The interannual variability in the ice sheet integrated mass balance components is considerable (Fig. 2). No significant trend in snowfall occurs over the period 1958–2008. In contrast, melt (and melt + rain) has increased significantly over the last 20 yr (about 3 % yr^{-1}), as has refreezing (about 1.5 % yr^{-1}). With the shift of melt to ever higher elevations, melt water will not run off, but refreeze in the cold snow pack until the refreezing capacity has degraded to the point that runoff starts.

The surface temperature shows the well-known decrease of temperature with height and elevation (Fig. 3). The –15 °C isotherm corresponds to an altitude of about 1500 m (equilibrium line) on the western margin, and elsewhere ranges from sea level up to 2000 m. The annual average temperature on the GrIS is –24.3 °C, and varies between

−26.2 °C and −22.3 °C (Fig. 4). Over the past 20 yr, surface temperature on the ice sheet has increased by about 2.5 °C; the first decade of the model period exhibits similarly high temperatures, although inter-annual variability was much larger than during the last decade.

4 Results

First a comparison will be made of the parameterizations as formulated in their original papers. Then the sensitivity of the different parameterizations to their different input parameters will be discussed. All results will be compared to refreezing as calculated in RACMO2, in the absence of observations.

4.1 Comparison

4.1.1 Time series

Figure 5 shows that the inter-annual variability in parameterized values of E_r is very similar for most methods. The absolute values, on the other hand, exhibit a large range around RACMO2, with mean differences (Diff) ranging from −42.6 mm w.e. (−37.7 %) to 35.5 mm w.e. (+31.4 %) (Table 2). Over the largest part of the ice sheet, i.e. the higher parts, E_r is limited by W_r. In these higher areas the correspondence between RACMO2 and the parameterizations is good (see next Section). In RACMO2, because of this, a strong correlation exists between ice sheet annual averaged M (or M + rain) and E_r (Fig. 6). The differences in temporal variability and absolute amount in Fig. 5 are therefore mainly determined by the lower areas of the ice sheet, where E_r is at least partly determined by P_r.

In Fig. 5, Pf1991 and HdW1999 show the lowest, and Wr2007 the highest refreezing values. The low values for Pf1991 are mainly the result of the mask formulation, which only takes into account refreezing at elevations above the runoff line, while below this line all water is assumed to run off. Although JH2000 is the most physically based parameterization, and in that sense best comparable to RACMO2, the refreezing differs significantly from RACMO2. The absolute amount is lower by 17 %, as is the temporal variability, by 31 %. Of all parameterizations the average difference with RACMO2 is smallest in Re1991 (Diff = −4.2 mm w.e., Table 2). In Re1991, refreezing is determined by either the annual average snowfall C, or melt M (Eqs. 2 and 3). The value of $P_{\mathrm{max}} = 0.6$ is obviously well chosen to represent the fraction of C that is refrozen in the area where $P_{\mathrm{max}}C$ limits the amount of refreezing.

Oe1991 also corresponds well with RACMO2 (Diff = 11.3 mm w.e., Table 2). This is surprising since the formulation of E_{r} in Oe1991 is fairly different from RACMO2. Furthermore, we do not apply Oe1991 as it was original intended. Oe1991 was designed to be used in an energy balance model, were refreezing changes the snow temperature, and thereby affecting refreezing in the next time step. This interaction is not allowed in the present application. Furthermore, we use monthly data as input fields, instead of 15 min time steps in Oe1991 (Eq. 10), not taking into account variability on shorter time scales. The reason that Oe1991 results are similar to RACMO2 is twofold: firstly, by limiting E_{r} to the annual amount of total precipitation P_{tot}, the possible overestimation of refreezing in the ablation area, where ice surfaces in the course of the melt season, is prevented. Secondly, as is the case for the other parameterizations, W_{r} is the limiting factor over the remainder of the ice sheet, not P_{r}.

4.1.2 Areal distribution

Although the time series correspond reasonably well, the spatial distribution of refreezing differs significantly between the different parameterizations and RACMO2 (Fig. 7, Table 2). In general, all parameterizations show small differences with RACMO2 in the higher parts of the ice sheet (<10 mm w.e.), where W_{r} is the limiting factor for refreezing. For the parameterizations that take rain into account, the difference goes to zero (Fig. 7c, d, e), whereas the others show small negative differences in the order of the annual amount of rain. The largest differences are once more found at the margins of the ice sheet, where most of the refreezing occurs. In these areas E_r is mainly determined by P_r, not W_r. The largest underestimation of refreezing compared to RACMO2 is found in Pf1991, which does not allow refreezing to occur below the runoff line, i.e. along the ice margins.

For the parameterizations that depend on the annual amount of snowfall (Re1991, Pf1991, JH2000), E_r is larger than RACMO2 along the south and southeastern margins of the ice sheet (Fig. 7a, b, c) where the amount of snowfall is high (Fig. 1a). In these areas, the use of snowfall results in a large P_r, and consequently E_r is limited by W_r and not by P_r. In RACMO2 P_r is not set by a given snow depth and determined more by the modelled snow temperature, resulting in lower values of E_r.

HdW1999 make use of a fixed depth of 2 m w.e. for the thermally active layer. On the south and southeast margin, this value is smaller than snowfall C resulting in lower values of P_r, limiting E_r. Although Wr2007 also uses a constant value of the thermally active layer (5 m w.e.), this value is on average larger than C and therefore does not explain the smaller values of E_r on the southeast margin. The smaller difference between RACMO2 and Wr2007 in these areas is likely caused by a different representation of the cold content by using the integrated area between standard profiles of the winter and summer snow temperature instead of using the annual average temperature.

Based on Table 2, the best correspondence with RACMO2 (lowest Std2) is found for Oe1991, although Re1991 and HdW1999 also show reasonably low Std2 values. Although the principles on which Pf1991 and especially JH2000 are based are the most similar to RACMO2, the spatial correspondence is smallest (highest Std2 values, Table 2).

4.2 Sensitivity experiments

We investigate the sensitivity of the calculated amount of refreezing by varying the different parameters in the parameterizations: the period of averaging, the in- or exclusion of rain, as well as more model specific parameters such as the depth of the thermally active layer, temperature, yes or no capillary water, and density. The tests are described below and statistics are presented in Table 3.

4.2.1 Annual or period averages

The parameterizations presented in Sect. 2 are based on either annual average values of C, M and/or T_s, or period mean (1958–2008) annual values (except Oe1991). The result is an annual P_r that is either constant throughout the calculated period, or annually variable. The latter is the physically most correct approach and applied by Re1991, JH2000 and HdW1999. Pf1991 and Wr2007 make use of period average P_r (Table 1). Pf1991 motivated his choice by limited available information, while Wr2007 based their parameterization on typical profiles of T_{sn} at the end of winter and summer that were best represented by multi-year averages of T_s and T_w determined from snow/ice temperature profile measurements.

Especially the inter-annual variability in Pf1991 depends heavily on whether a period average or annual average mask is used (Table 3): inter-annual variability is much larger when using a period-averaged mask. This is due to the fact that in Pf1991 the variability is determined by the variability in the melt, which is compensated by changes in the mask if annual values are used: more/less melt results in a smaller/larger area with refreezing. Results of Wr2007 and Re1991 on the other hand are not very sensitive to this choice. For Wr2007 this is explained by the fact that with $d_{ice} = 5$ m w.e., the refreezing over most part of the ice sheet is limited by W_r and not by P_r. Thus, as long as changes in P_r do not result in a significant larger area where P_r exceeds W_r, E_r will not be sensitive to changes in P_r. In Re1991 P_r is determined by C. Using a period average C results in a larger dependency on M. Using period averages, HdW1999 and JH2000 are also more determined by variations in M. In these parameterizations the correspondence with RACMO2 increases due to the correlation between M and E_r in RACMO2. Note that HdW1999, which is a parameterization very similar to Wr2007, is more sensitive to changes in P_r. This is caused by their choice of $d_{ice} = 2$ m w.e. leading to an on average lower P_r. The sensitivity to d_{ice} will be discussed in more detail below.

Using period averages, HdW1999 and JH2000 are more determined by variations in M. As a result, both show an increase in E_r, corresponding to the increase in M, which is larger than found in RACMO2.

4.2.2 Refreezing of rain

The amount of refreezing (E_r) is (partly) determined by the available amount of water including rain (W_r). However, not all parameterizations take rain falling on cold snow into account in their estimate of W_r. Pf1991, Re1991 and Oe1991 assume the contribution of rain to be negligible, because rain constitutes only a small fraction of the total amount of precipitation. In RACMO2, about 6 % of the annual amount of precipitation over the ice sheet falls as rain, with the largest percentages (up to 50 %) on the southern ice margins. Therefore, refreezing of rain may locally constitute a significant contribution to the total.

Including rain increases the amount of refreezing in all cases (Table 3), by up to 12 %. Locally the differences can be much larger. Figure 8 illustrates this for two cases, JH2000 and HdW1999, where JH2000 shows large differences and HdW1999 only small differences. In Oe1991 the difference is smallest, which is due to the fact that in the regions with most rainfall, refreezing is limited by the annual amount of precipitation (P_{tot}), not by W_r. Note that JH2000 also limits E_r to P_{tot}, but in JH2000 calculated E_r seldom exceeds P_{tot} and rain is included in W_r. The largest differences are found for the parameterizations that use the annual snowfall as depth of the thermally active layer. In those cases, W_r limits E_r in the lower areas where C is large (such as in the south east), thus increasing W_r, which results in more refreezing as can be seen for JH2000 in Fig. 8a. In the case of JH2000 the inclusion of capillary water increases the difference even further, since it provides additional capacity to store water in areas where C is larger than M. In the case of HdW1999 the use of a constant d_{ice} results in the largest differences in the areas where available liquid water is the limiting factor, which is just above the equilibrium line.

4.2.3 Depth of the thermally active layer

All the parameterizations tested in this study, except Oe1991, make use of an estimation of the depth of the thermally active layer (d_{ice}). Parameterizations Pf1991, JH2000 and Re1991 assume that d_{ice} equals annual snowfall C, whereas HdW1999 and Wr2007 assume a constant value for d_{ice} of 2 m w.e. and 5 m w.e., respectively. In the tests we vary d_{ice}, or use $d_{\text{ice}} = C$.

The amount of refreezing is very sensitive to the choice of d_{ice} as can be seen in Table 3 and Fig. 9. Figure 9 shows the difference in E_{r} when using different values of d_{ice}, and illustrates that when d_{ice} increases, refreezing increases, the latter becoming more and more limited by the available amount of liquid water W_{r}. Using a constant d_{ice}, JH2000, HdW1999 and Wr2007, can be tuned to best represent the ice sheet and period averaged RACMO2 refreezing (Table 4). Wr2007 and HdW1999 show the smallest mean difference and the best correspondence in temporal and spatial variability to RACMO2 when d_{ice} is about 3 m w.e. (3.07 and 3.45 m w.e., respectively), while JH2000 shows the best correspondence when $d_{\text{ice}} = 1.45$ m w.e.

In all experiments, using $d_{\text{ice}} = C$ drastically reduces the amount of refreezing (Fig. 9b). The reason is that period averaged C is only 0.40 m w.e. Using $d_{\text{ice}} = C$ results in smaller P_{r} over those parts of the ice sheet where annual average C is smaller than 3 m w.e. (Fig. 1a), which is virtually everywhere. In addition, the interannual variability almost vanishes, and the spatial correspondence with RACMO2 decreases. JH2000 is the least affected by this choice because they include refreezing of capillary water, which does not depend on the depth of the thermally active layer.

4.2.4 Capillary water

Pf1991 and JH2000 are the only parameterizations that specifically take into account the refreezing of capillary water at the end of the melt season (second term r.h.s. Eqs. 4 and 5). We tested the impact by removing this term in Eqs. (4) and (5). Note that removing the capillary water in JH2000 equals using HdW1999 with the same d_{ice} as

JH2000. Including capillary water increases the amount of refreezing (Table 3). It also results in a larger temporal variability. However, although RACMO2 also includes the contribution of capillary water, including it in the parameterizations does not result in a better spatial agreement with RACMO2.

Figure 10 shows that capillary water is a significant contributor to refreezing in areas were melt M does not exceed the amount of snowfall C. This is especially the case in Pf1991 where, due to the mask formulation and the use of $T_\mathrm{f} = -15\,^\circ\mathrm{C}$, the remaining cold content is very small, resulting in only a small area where refreezing occurs. In JH2000, the areas where the difference is zero are those where M exceeds C and those where P_r minus the possible capillary contribution is larger than M. In case d_ice is constant, the latter area is larger because P_r remains larger compared to the case where $d_\mathrm{ice} = C$, since over large areas of the ice sheet C is on average smaller than 2 m w.e. (see Fig. 1a).

4.2.5 Density

When including the capillary water content, the additional amount of refreezing that may occur depends on the chosen densities. JH2000 (Eq. 5) use a pore close-off density $\rho_\mathrm{pc} = 960\,\mathrm{kg\,m^{-3}}$, which they define as the density of water saturated snow, while Pf1991 (Eq. 4) use a value of $900\,\mathrm{kg\,m^{-3}}$. They both use a firn density $\rho_\mathrm{f} = 300\,\mathrm{kg\,m^{-3}}$. Changing ρ_pc or ρ_f changes the factor determining how much water can be retained. Increasing this factor, either by increasing ρ_pc or decreasing ρ_f, results in a larger amount of refreezing. Tests with the density factor are presented in Table 3 and Fig. 11. They show an increase in the amount of refreezing, which is largest in areas around the equilibrium line. Increasing the density factor further results in a larger inter-annual variability and less spatial correspondence with RACMO2 in case of JH2000 and more in case of Pf1991 (Table 3). Changing ρ_f has the largest impact, but the change has to be considerable to have a significant effect. This is because changing density only has effect in areas were less than the annual amount of snowfall C melts away, and where P_r is the limiting factor, not W_r.

4.2.6 Temperature

In several parameterizations temperature is used as a measure for the cold content of the snow. Except for Pf1991, all parameterizations were forced by RACMO2 surface temperatures T_s. Pf1991 uses a representative value of the firn temperature at the firn limit (–15 °C). The sensitivity of the parameterizations to the temperature chosen indicates in fact how well this temperature represents the cold content of the snow.

Table 3 shows that Pf1991 and JH2000 are not very sensitive to reasonable changes in T_s while HdW1991, Wr2007 and Oe1991 are very sensitive. In Oe1991 the sensitivity to changes in T_s is strongly non-linear due to the exponential relation between T_s and P_r. In case of equal d_{ice}, the temperature description is responsible for the main difference between HdW1999 and Wr2007 (Fig. 12). Figure 4 illustrates the temperature used by both parameterizations. Due to the combined use of annual averaged and winter temperature, the temperature factor used by Wr2007 is more variable in time than the annual mean temperature used in HdW1999. This does not result in a larger sensitivity to changes in temperature in Wr2007.

4.2.7 P_{max}

Compared to RACMO2, $P_r = 0.6C$ represents the amount and temporal variability in refreezing well in areas where P_r is the limiting factor (Table 2). Tuning results in an even better correspondence in average amount, although the resulting value of P_{max} does not deviate much from 0.6 (0.65, Table 4). Increasing P_{max} increases the amount of refreezing below the elevation where M is the limiting factor and increases the area where M is the limiting factor. It also increases the temporal variability and decreases the spatial correspondence with RACMO2. Decreasing P_{max} results in the opposite: it decreases the temporal variability and increases the correspondence with RACMO2.

From RACMO2 fields of C and E_r, the fraction of C that is refrozen can be calculated (Fig. 13). Interesting feature in this figure is the northern marginal areas where the fraction is larger than 1 and thus more than the annual amount of snowfall refreezes.

This is the result of multiple cycles of melt and refreezing of the same snow/ice. This happens over the whole GrIS, but in areas where little or no runoff takes place, and C is small, this can result in $E_r/C > 1$. The cold snow pack warms up due the energy provided by refreezing. Due to the increase in melt over the period 1958–2008 (Fig. 2), the refreezing capacity in these areas will degrade to the point that runoff starts. In contrast, in the southeastern marginal zone E_r is small compared to C. Only on the western margin of the ice sheet are there significant areas where the fraction is about 0.6, similar to the value of P_{max} measured by Braithwaite et al. (1994) in this area. The ice sheet average value of E_r/C is 0.28.

5 Summary and conclusions

In this study we applied several parameterizations that calculate the annual amount of refreezing to the Greenland ice sheet. In the absence of refreezing observations we compare the results to output of the RACMO2 regional climate model, that includes an explicit scheme to calculate retention and refreezing as a function of snow depth and temperature. The parameterizations are forced with output from the same model for consistency. Almost all refreezing parameterizations discussed here use temperature and an estimate of the depth of the thermally active layer to determine the cold content of the snow. In RACMO2, water may percolate to any depth depending on the vertical temperature and density distribution in the snow/firn.

The annual, period average (1958–2008) and ice sheet averaged amount of refreezing calculated with the different parameterizations differs up to a factor 2 with RACMO (Table 2). The spatial fields show large differences as well, especially in the lower areas of the ice sheet (up to a factor 5). Janssens and Huybrechts (2000) also noted large differences in parameterized refreezing in these areas, which they related to the chosen depth of the thermally active layer. Our results confirm this large sensitivity as well as the large impact this has on refreezing in the marginal areas. All parameterizations can be tuned within realistic limits, to produce ice sheet and annual average amount of refreezing similar to RACMO2, but this does not necessarily result in better spatial correspondence (Table 4). After tuning, the spatial and temporal variability of Wr2007 is most similar to RACMO2.

Care must be taken when choosing a parameterization, because they were developed for different applications. For example, Pf1991 was not intended to be applied to the full ice sheet, but was developed to describe the effect of refreezing on the average GrIS mass balance profile. The lack of refreezing below the runoff line in this method is therefore, of limited importance, since in this area the refrozen mass melts again later in the season to run off. Note that the elevation of the chosen runoff line should be close to the equilibrium line. The P_{max} formulation works well on annual ice sheet averages, because P_{max} corresponds to the fraction of C that is refrozen in the area where most refreezing occurs.

Oe1991 is the only parameterization that does not include the depth of the thermally active layer. The amount of refreezing in Oe1991 depends on available energy and temperature. Oe1991 is very sensitive to changes in the latter. However, Oe1991 was designed for application in an energy balance model that includes a simple snow model, in which the snow temperature changes when refreezing occurs. To obtain reasonable results in our test, the refreezing is limited to the total annual precipitation P_{tot}. It is questionable whether Oe1991 will work similarly well in other settings and without those constraints.

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Table 1. The tested parameterizations. W_r refers to whether the available water mass in Eq. (2) equals melt (M) or melt plus rain. Input P_r lists the input parameters to P_r, period refers to the period over which the input to P_r is averaged.

	Abbr.	Input W_r	Input P_r	Period	Comments
Ettema et al. (2010b)	RACMO2	M + Rain	–	–	Reference
Reeh (1991)	Re1991	M	C	Annual	$P_{max} = 0.6$
Pfeffer et al. (1991)	Pf1991	M	C, M	Period	$\rho_{pc} = 900\,\mathrm{kg\,m^{-3}}$ $\rho_f = 300\,\mathrm{kg\,m^{-3}}$ $T_f = -15\,^\circ\mathrm{C}$
Janssens and Huybrechts (2000)	JH2000	M + Rain	C, M, T_s	Annual	$\rho_{pc} = 960\,\mathrm{kg\,m^{-3}}$ $\rho_f = 300\,\mathrm{kg\,m^{-3}}$
Huybrechts and de Wolde (1999)	HdW1999	M + Rain	T_s	Annual	$d_{ice} = 2\,\mathrm{m}$
Wright et al. (2007)	Wr2007	M + Rain	T_s, T_w	Period	$d_{ice} = 5\,\mathrm{m}$
Oerlemans (1991)	Oe1991	M	Q, T_s	Time step	

Table 2. Statistics of the different parameterizations compared to RACMO2. Mean = ice sheet and period (1958–2008) averaged annual E_r; Diff. = difference of mean with RACMO2 (Param.-RACMO2); Std1 = standard deviation in the ice sheet and annual averages; Std2 = standard deviation in the period averaged differences over all grid points. All values are expressed in mm w.e.

Param.	Mean	Diff.	Std1	Std2
RACMO2	112.9		19.8	
Re1991	108.7	−4.2	17.1	85.9
Pf1991	70.3	−42.6	20.6	142.8
JH2000	93.4	−19.4	13.6	129.1
HdW1999	80.1	−32.8	13.5	89.5
Wr2007	148.4	35.5	26.5	113.0
Oe1991	101.6	−11.3	18.5	74.3

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Table 3. The sensitivity of the parameterizations to their input variables. Reference statistics are given in Table 2. Headings are as in Table 2, test refers to change compared to reference, numbers between brackets denote sections where experiments are discussed.

Param.	Test	Mean	Diff.	Std1	Std2
Averaging period (Sect. 4.2.1)					
Re1991	Period	111.0	−1.9	18.0	90.2
Pf1991	Annual	66.6	−46.4	10.3	114.5
JH2000	Period	115.0	2.1	42.6	135.3
HdW1999	Period	103.8	−9.1	39.0	80.1
Wr2007	Annual	148.1	35.2	26.7	112.6
Including rain or not (Sect. 4.2.2)					
Re1991	M + Rain	116.9	4.1	17.2	106.0
Pf1991	M + Rain	76.1	−36.8	21.7	166.7
JH2000	M	82.4	−30.5	12.5	103.1
HdW1999	M	77.0	−35.9	13.6	89.1
Wr2007	M	141.4	28.5	26.5	113.0
Oe1991	M + Rain	109.2	−3.7	20.1	90.2
Thickness d_{ice} (Sect. 4.2.3) and capillary water (Sect. 4.2.4)					
Pf1991	no cap. water[1]	6.1	−106.8	4.6	147.1
JH2000	$d_{\text{ice}} = 2\,\text{m}$	124.5	11.6	17.7	112.1
JH2000	$d_{\text{ice}} = 3\,\text{m}$	139.6	28.6	20.6	115.9
JH2000	$d_{\text{ice}} = 5\,\text{m}$	164.6	51.7	26.1	141.9
HdW1999	C	29.9	−83.0	3.7	115.6
HdW1999	$d_{\text{ice}} = 1\,\text{m}$	48.6	−64.3	7.7	114.0
HdW1999	$d_{\text{ice}} = 3\,\text{m}$	103.8	−9.1	17.7	79.1
Wr2007	C	30.6	−82.3	3.7	115.3
Wr2007	$d_{\text{ice}} = 3\,\text{m}$	111.3	−1.5	19.5	79.2
Wr2007	$d_{\text{ice}} = 6\,\text{m}$	162.4	49.5	29.5	136.2

Table 3. Continued.

Density (Sect. 4.2.5)					
Pf1991	$\rho_{pc} = 960\,\mathrm{kg\,m^{-3}}$	72.4	−40.5	21.0	142.3
Pf1991	$\rho_{pc} = 830\,\mathrm{kg\,m^{-3}}$	67.6	−45.5	20.1	143.2
Pf1991	$\rho_f = 450\,\mathrm{kg\,m^{-3}}$	54.1	−58.7	17.5	144.7
Pf1991	$\rho_f = 150\,\mathrm{kg\,m^{-3}}$	84.4	−28.5	23.1	139.0
JH2000	$\rho_{pc} = 900\,\mathrm{kg\,m^{-3}}$	92.2	−20.7	13.4	127.7
JH2000	$\rho_{pc} = 830\,\mathrm{kg\,m^{-3}}$	90.5	−22.4	13.1	125.8
JH2000	$\rho_f = 450\,\mathrm{kg\,m^{-3}}$	83.2	−29.6	11.9	117.7
JH2000	$\rho_f = 150\,\mathrm{kg\,m^{-3}}$	101.6	−11.3	15.1	138.5
Temperature (Sect. 4.2.6)					
Pf1991	$T_s = \mathrm{RACMO2}^2$	70.1	−42.7	20.6	142.2
Pf1991	$T_f = -10$	69.1	−43.7	20.4	143.1
Pf1991	$T_f = -20$	71.1	−41.8	20.7	142.7
JH2000	$T_s + 5$	91.2	−21.6	13.4	130.1
JH2000	$T_s - 5$	95.5	−17.3	13.9	128.4
HdW1999	$T_s + 5$	64.2	−48.7	11.0	105.5
HdW1999	$T_s - 5$	93.8	−19.1	15.7	77.6
Wr2007	$T_s + 5$, $T_w + 5$	131.2	18.3	23.6	99.1
Wr2007	$T_s - 5$, $T_w + 5$	162.8	49.9	29.1	132.3
Oe1991	$T_s + 5$	24.1	−88.8	5.4	135.1
Oe1991	$T_s - 5$	110.4	−2.5	20.8	91.2
P_{max} (Sect. 4.2.7)					
Re1991	$P_{max} = 0.5$	98.8	−14.1	14.9	80.2
Re1991	$P_{max} = 0.7$	116.8	3.9	19.0	92.2

[1] Second term r.h.s. Eq. (4) is 0.

[2] Period averaged per grid point.

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Table 4. Statistics of the different parameterizations compared to RACMO2 after tuning. Headings are as in Table 2, comments refers to changes in parameter setting compared to the reference (Table 2). In all experiments mean = 112.9 mm w.e. and Diff = 0.0 mm w.e.

Param.	Std1	Std2	Comments
RACMO2	19.8		
Re1991	18.1	89.0	$P_{\mathrm{max}} = 0.65$
Pf1991	15.5	116.7	$d_{\mathrm{ice}} = 4.46$ m w.e., $\rho_{\mathrm{pc}} = 960\,\mathrm{kg\,m^{-3}}$, $T_{\mathrm{f}} = -24.3\,^{\circ}\mathrm{C}$, annual averages, including rain
JH2000	15.9	113.9	$d_{\mathrm{ice}} = 1.45$ m w.e.
HdW1999	19.3	79.9	$d_{\mathrm{ice}} = 3.45$ m w.e.
Wr2007	19.7	79.6	$d_{\mathrm{ice}} = 3.07$ m w.e.
Oe1991	21.1	97.2	$T_{\mathrm{s}} - 0.37$, including rain

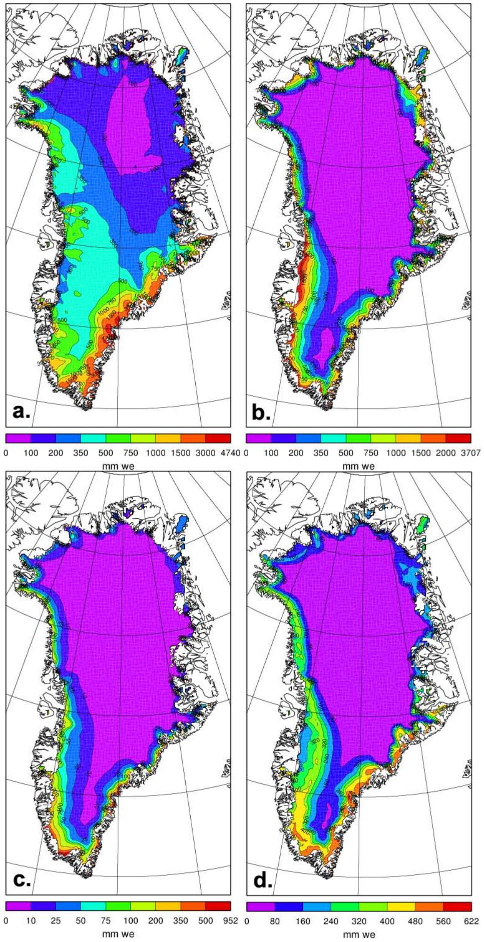

Fig. 1. Period (1958–2008) averaged anual sums of **(a)** snowfall (*C*) **(b)** melt (*M*), **(c)** rain, and **(d)** refrozen mass (E_r) (mm w.e.) as modelled in RACMO2. Note the different scales.

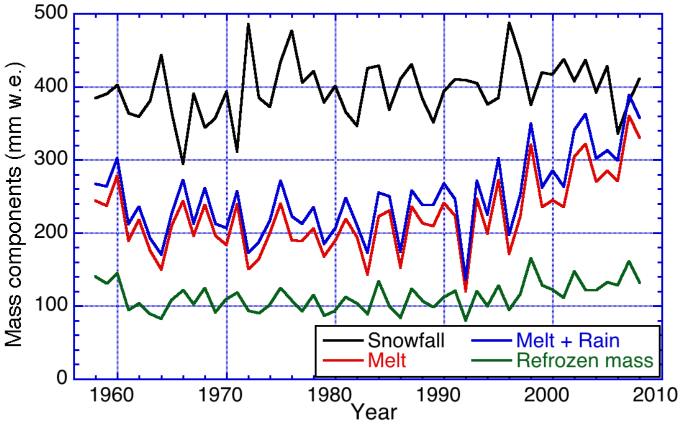

Fig. 2. Time series of ice sheet averaged annual sums of snowfall (C), melt (M), melt plus rain (W_r), and refrozen mass (E_r) as modelled in RACMO2.

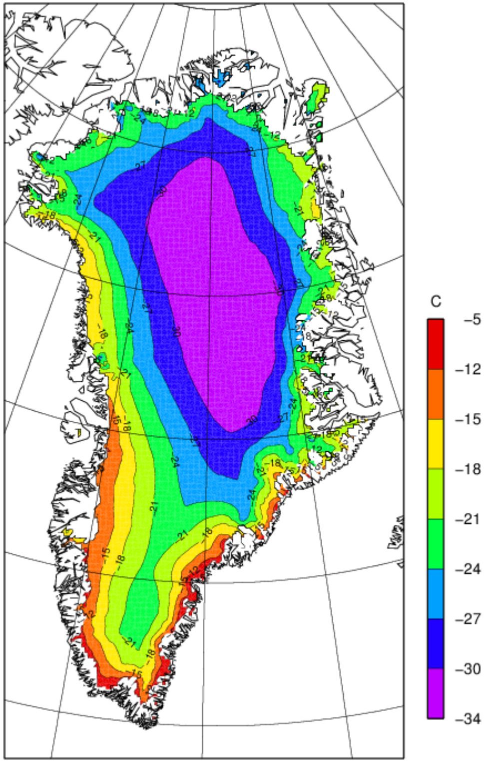

Fig. 3. The period (1958–2008) average annual mean surface temperature (T_{s} in °C) as modelled in RACMO2. The −15 °C isotherm corresponds to an altitude of about 1500 m (equilibrium line) in the ablation area on the western margin, and elsewhere ranges from sea level up to 2000 m.

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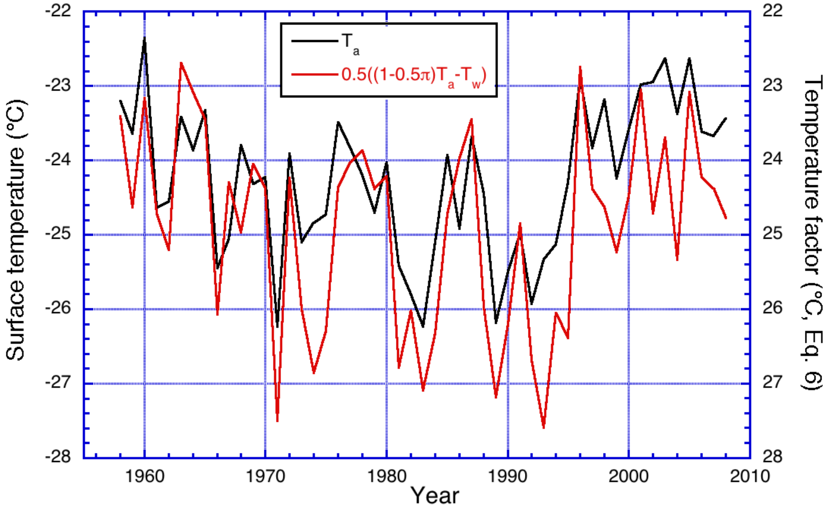

Fig. 4. Time series of ice sheet averaged annual averaged surface temperature (T_s, black) as modelled in RACMO2 and temperature factor used in Eq. (7) (red). Note the reversed axis on the right hand side.

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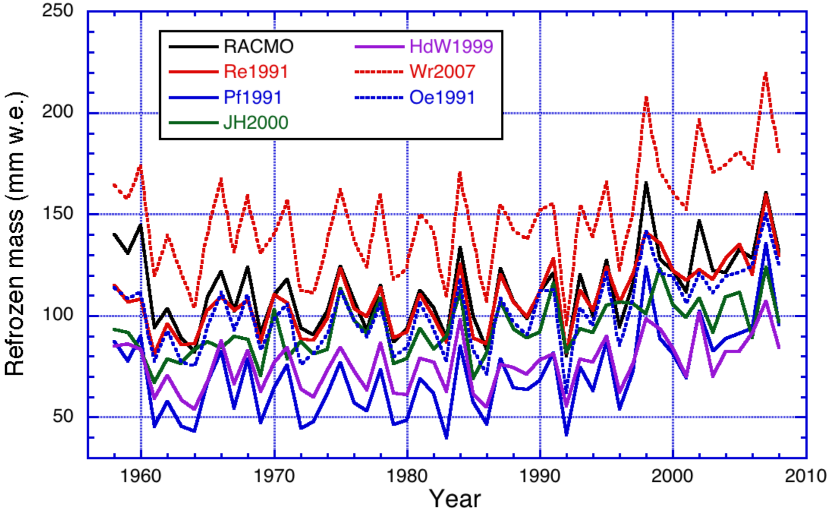

Fig. 5. Time series of ice sheet averaged annual sums of refrozen mass (E_r) as modelled using the presented parameterizations.

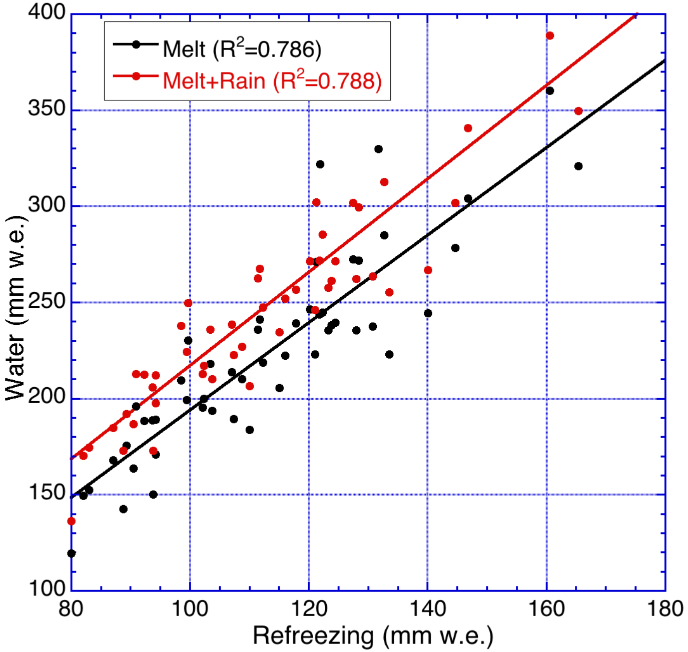

Fig. 6. Scatter plot of annual and ice sheet averaged melt (M) and melt plus rain as a function of refrozen mass (E_r).

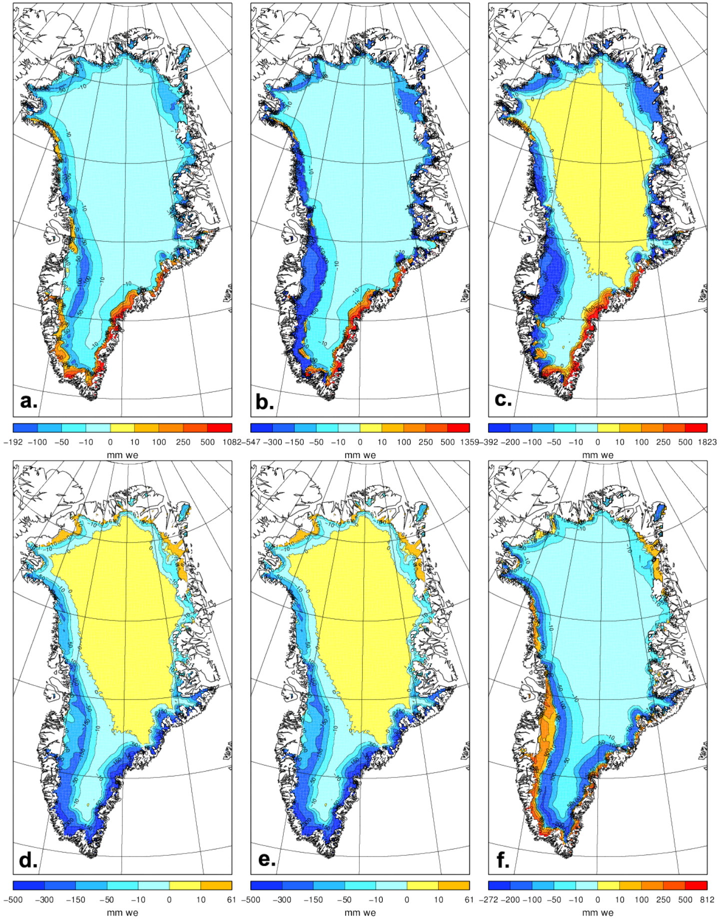

Fig. 7. Difference between RACMO2 modelled refrozen mass (E_r) and parameterized amount, Param.-RACMO2. **(a)** Re1991, **(b)** Pf1991, **(c)** JH2000, **(d)** HdW1999, **(e)** Wr2007, **(f)** Oe1991.

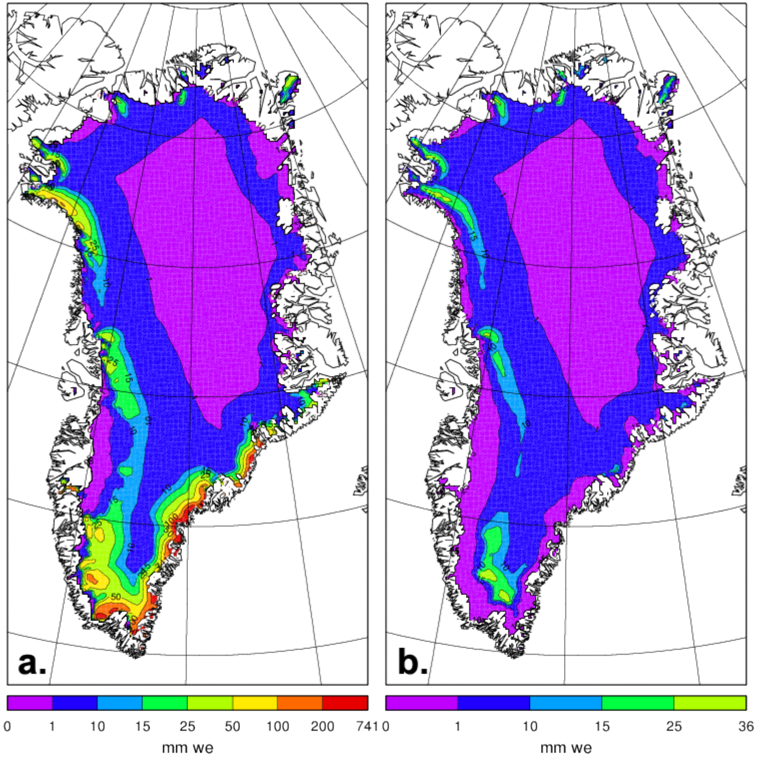

Fig. 8. Difference in refreezing when including rain in W_r (Yes – No) for JH2000 **(a)** and HdW1999 **(b)**.

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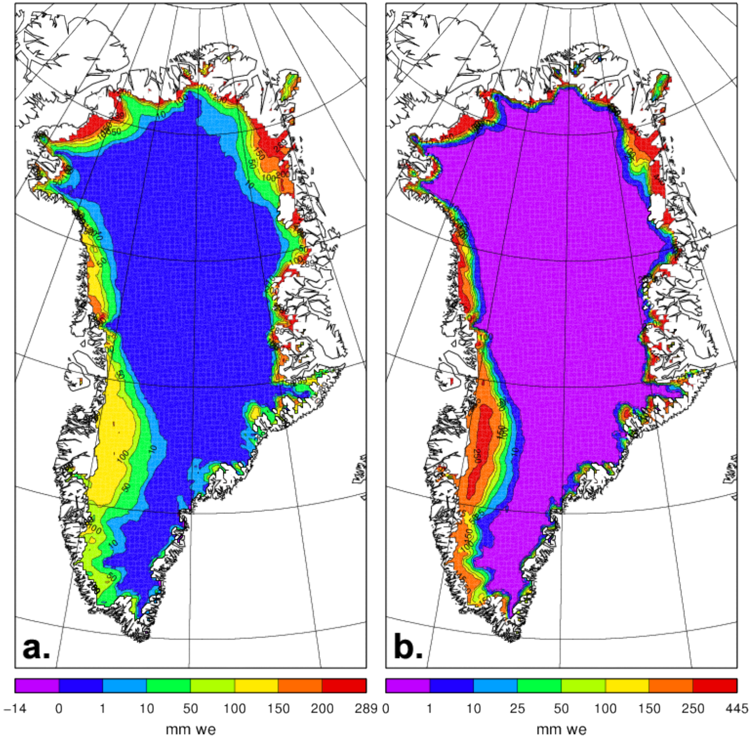

Fig. 9. Difference in refreezing when varying the depth of the thermally active layer (d_{ice}) in JH2000 (Eq. 5). **(a)** $d_{ice} = 2$ m minus C, **(b)** $d_{ice} = 5$ m minus 2 m.

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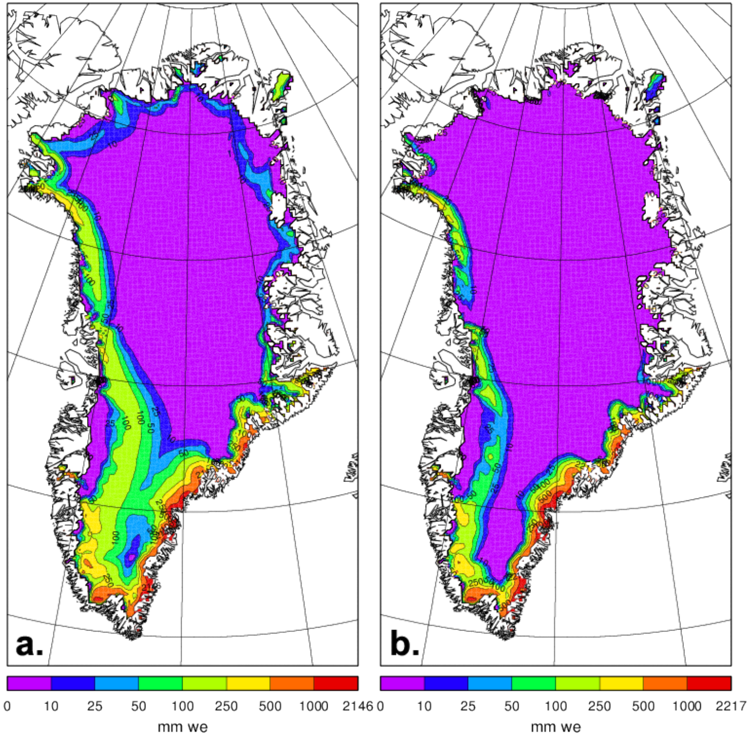

Fig. 10. Difference in refreezing when including capillary water in P_r. Both figures show JH2000 minus HdW1999 (Yes – No). **(a)** $d_{ice} = C$, **(b)** $d_{ice} = 2$ m.

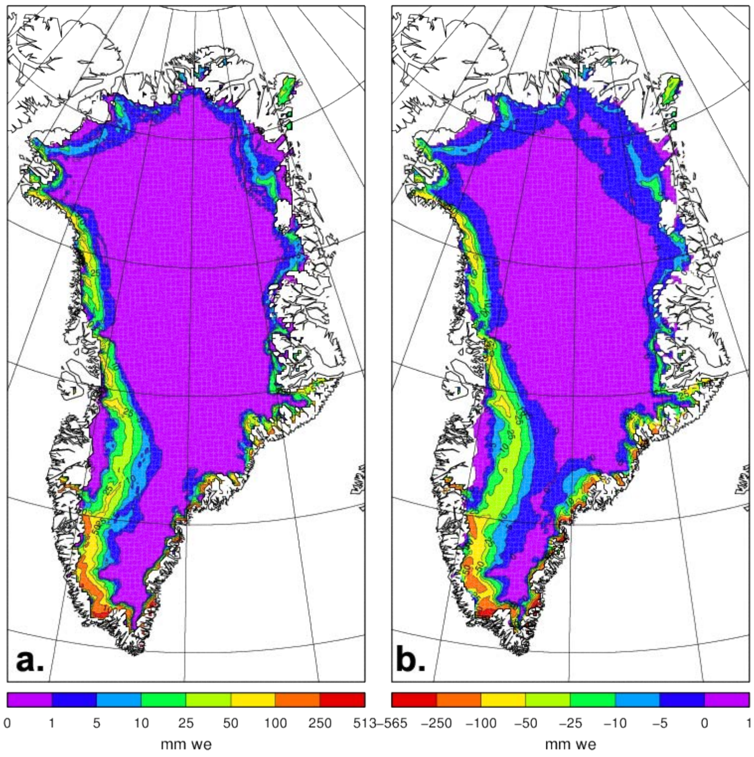

Fig. 11. Difference in refreezing when changing the density factor in JH2000 by changing $\rho_f = 300\,\text{kg}\,\text{m}^{-3}$ (Eq. 5) to **(a)** $\rho_f = 150\,\text{kg}\,\text{m}^{-3}$, and **(b)** $\rho_f = 450\,\text{kg}\,\text{m}^{-3}$ (test minus reference). Note the reversed color scale in panel **(b)**.

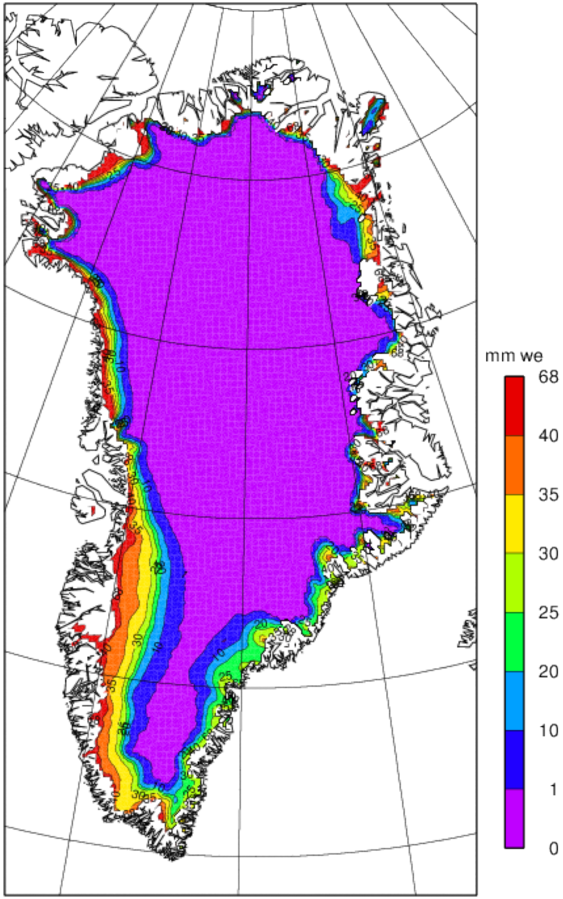

Fig. 12. Difference in refreezing when using different temperature descriptions (Wr2007 minus HdW1999 with $d_{\mathrm{ice}} = 3\,\mathrm{m}$).

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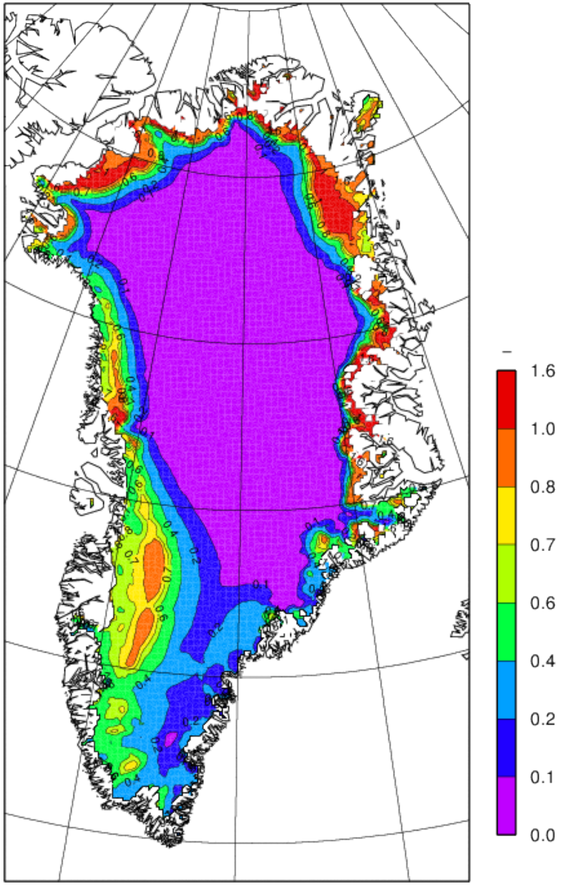

Fig. 13. $P_{\max}$ fraction calculated from RACMO2 fields of E_r and C (E_r/C).