

Impact of climate forcing time step ~~on the modelled ice-sheet in an ice sheet firn layer model~~

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Abstract. The firn layer regulates how an ice-sheet responds to climate change by modifying how changes in surface temperature, snow accumulation and ablation affect the ice-sheet mass balance. Firn properties are often simulated with a firn densification model. ~~Firn models are forced with surface mass balance components, surface energy balance components and/or meteorological variables. In literature, Prior studies have used~~ a variety of time steps in ~~those climate forcings have been used~~ to force the climate forcings of such firn models, ranging from 3-hours to 1-day, 1-month, or even annual. To investigate the impact of these different time steps, we force the firn densification model IMAU-FDM with different climate forcing time steps at the surface for the Antarctic Peninsula and southern Greenland Ice Sheet. IMAU-FDM is forced at its upper boundary with surface mass balance components and meteorological variables. We show that the modelled firn layer contains more pore space for larger forcing time steps, and that ~~locations~~ this effect depends on the climate regime. Locations with limited firn pore space due to seasonal melt, and regions with emerging firn aquifers, are most sensitive. The climate forcing time step impacts the creation of pore space by ~~snowfall~~ snow accumulation and the depletion of pore space by snowmelt and firn densification. The key in causing the differences in firn pore space is the presence or absence of a diurnal cycle in the input data. A climate forcing time step greater than a day allows for a non-physical coexistence of snowmelt and sub-zero surface temperatures, leading to immediate shallow refreezing of meltwater. Subsequent melting removes refrozen ice rather than porous firn, reducing the amount of firn air that is lost through melting. Therefore, for locations experiencing surface melt, the decoupled temperature and snowmelt in the upper layers ~~give results in~~ more firn air with a climate forcing time step larger than a day. We also found that firn density parameterizations can become unsuitable when applied outside the physical conditions or climate forcing time step on which they are based. These parameterizations lead to unrealistic firn densification and accumulation behavior in the model. We argue that (1) firn models forced with mass balance terms and meteorological variables require a timestep small enough to capture at least the diurnal cycle, (2) use of parameterizations should be critically assessed and used consistently with the way they were originally developed, and (3) the forcing time step should be considered when interpreting firn model output.

1 Introduction

Firn, the transitional state between snow and glacial ice, covers $\sim 90\%$ of the Greenland Ice Sheet (GrIS) (Noël et al., 2022) and $\sim 99\%$ of the Antarctic Ice Sheet (AIS) (Winther et al., 2001). Firn contains air that prevents a fraction of the meltwater to run off into the ocean, either by refreezing or retention. Approximately $\sim 39\%$ of surface meltwater on the GrIS and $\sim 94\%$ on the AIS is retained in the firn (Medley et al., 2022). Consequently, the firn layer keeps meltwater away from crevasses ~~on Antarctic ice shelves~~, where it could otherwise enter the GrIS subglacial drainage system, or engage in hydrofracturing on Antarctic ice shelves. Also, outside the percolation zone the firn layer evolution is of relevance, as variations in firn mass and thickness complicate the interpretation of altimetry observations for ice-sheet mass balance.

Firn models are used to improve our understanding of physical processes, to interpret observations ~~, to interpret~~ including remote sensing data, and to ~~assess~~ make projections of ice-sheet surface mass balance (SMB) in the future (The Firn Symposium team, 2024). Firn models differ in their parameterizations and formulations for densification, thermodynamics, and water percolation processes (Vandecrux et al., 2020). Although a variety of firn models exist, all models have in common that they use numerical methods to integrate a set of coupled differential equations in time. In this paper, we focus on the fact that the upper boundary conditions for solving the equations are provided at a discrete time step. Those boundary conditions can be components of the SMB and/or surface energy balance (SEB), or meteorological variables, which varies depending on the firn model (The Firn Symposium team, 2024). The climate forcing time step differs greatly across studies (~~Table ??~~), varying from 1 hour, to days, months, and even years ~~–~~.

~~Overview of the variety in climate forcing time steps used in different firn models: Climate forcing time step Literature~~
~~1-hour Community Firn Model (CFM) (Medley et al., 2022; Thompson-Munson et al., 2023), SNOWPACK (Thompson-Munson et al., 2023), 3-hours CFM (Verjans et al., 2019; Vandecrux et al., 2020), DTU model, HIRHAM, MeyerHewitt, UppsalaUniBucket, GEUS model (Vandecrux et al., 2020), (Sørensen et al., 2011; Simonsen et al., 2013; Langen et al., 2017; Meyer and Hewitt, 2017; Verjans et al., 2020). Here, we focus on the climate forcing time step for the IMAU firn densification model (IMAU-FDM (Vandecrux et al., 2020; Brils et al., 2020), 6-hours HIRHAM (Langen et al., 2017; Hansen et al., 2024) 1-day Firn densification model based on CFM (Zhang et al., 2024) 1-month DTU model (Sørensen et al., 2011; Simonsen et al., 2013), CFM (Stevens et al., 2020; Gkinis et al., 2021) 1-season MeyerHewitt (Meyer and Hewitt, 2017) Annual CFM (Gkinis et al., 2021)), which is forced with SMB and meteorological variables at the upper boundary.~~

Reasons for choosing a certain climate forcing time step are most often practical: the data are not available at smaller time steps, the amount of forcing data becomes too large, or the computational resources are insufficient. Evaluation of the forcing time step is often absent or based on differing lines of reasoning. For example, Medley et al. (2022) argue that the absence of a diurnal cycle in the forcing time step is a limitation of their model, whereas Meyer and Hewitt (2017) expect that the daily cycle influences only the uppermost meter of the firn column, and therefore, ignore it. We ~~want to~~ emphasize the difference between the time step used to solve the differential equations in a model, and the time step at which the forcing data is provided. The focus of this paper is on the latter. Most often, the model time step is smaller than the forcing data time step. To clarify with an example, Veldhuijsen et al. (2023) uses a 3-hourly climate forcing linearly interpolated to a 15 minute model time step.

In this paper, we ~~aim~~ demonstrate that the output of a firm model is sensitive to the forcing time step when forced with SMB and meteorological variables at the surface, while geographically focusing on the Antarctic Peninsula Ice Sheet (APIS) and the southern GrIS.

60 2 Methods: Firm modelling setup

We use ~~the IMAU Firm Densification Model (IMAU-FDM) (Brils et al., 2021; Veldhuijsen et al., 2023)~~ (Brils et al., 2021; Veldhuijsen et al. to simulate the firn layer and investigate the sensitivity of the modelled firn layer to the climate forcing time step. This section describes IMAU-FDM and specifies similarities and differences in the model for the two domains. We also describe the forcing dataset, ~~model initialization, the~~ setup for testing different forcing time steps, ~~and the~~ model initialization, and study areas.

65 2.1 IMAU-FDM

IMAU-FDM is a 1D semi-empirical model that simulates ~~densification rate, refreezing, temperature evolution~~ density, temperature, and meltwater transport content in the firn layer. The model is originally developed by Helsen et al. (2008) and has undergone multiple updates (Ligtenberg et al., 2014; Brils et al., 2021; Veldhuijsen et al., 2023). We use IMAU-FDM v1.2A (Veldhuijsen et al., 2023) for the Antarctic domain and IMAU-FDM v1.2G (Brils et al., 2021) for the Greenland domain, which only differ
70 from each other in their ice-sheet dependent calibration. Both versions have been evaluated extensively against firn density and temperature observations (Ligtenberg et al., 2014; Kuipers Munneke et al., 2015; Brils et al., 2021; Veldhuijsen et al., 2023).

The firn and ice column are divided into 200 to 3,000 layers, traced in a Lagrangian framework. Snowfall is added to the topmost layer of the column, burying other layers and thereby representing downward advection. No layers are removed during a simulation. The layers have a maximum vertical resolution of 0.15 m. Layer routines are ~~build~~ built in that split the uppermost
75 layer if it exceeds the 0.15 m threshold and merges the two uppermost layers if the thickness of the uppermost layer is smaller than 0.05 m. When the firn column has less than 200 layers, 100 very thin layers of pure ice are added to the bottom of the column.

The fresh snow density is an upper boundary condition with differing parameterizations for the AIS and GrIS domain. In IMAU-FDM v1.2A (AIS), the fresh snow density ρ_0 (kg m⁻³) depends on instantaneous surface temperature T_s (K) and 10 m
80 wind speed ~~$f f_{10m}$~~ $f f_{10m}$ (m/s) (Eq. 1) (Lenaerts et al., 2012; Veldhuijsen et al., 2023):

$$\rho_0 = 82.97 + 0.769T_s + 11.67 \underline{f f}_{10m} \quad (1)$$

In IMAU-FDM v1.2G (GrIS), ρ_0 is based on the surface temperature averaged over the year prior to the snowfall $\overline{T_s}$ (K) (Eq. 2) (Fausto et al., 2018; Brils et al., 2021):

$$\rho_0 = 362.1 + 2.78(\overline{T_s} - 273.15) \quad (2)$$

85 The ~~snow and underlying firn densification~~ densification of snow and the underlying firn in each layer is parameterized by a semi-empirical equation (Eq. 3) based on Arthern et al. (2010):

$$\frac{d\rho}{dt} = D \dot{b} g M O (\rho_i - \rho) e^{-\frac{E_c}{RT} + \frac{E_g}{RT_b}} \quad (3)$$

with $\dot{b}$ the average accumulation rate ($\text{kg m}^{-2} \text{yr}^{-1}$) over the spin-up period, g the gravitational acceleration (9.81 m s^{-2}), ρ_i the density of bubble-free ice (917 kg m^{-3}), ρ the instantaneous layer density (kg m^{-3}), E_c the activation energy for creep ($60,000 \text{ J mol}^{-1}$), E_g the activation energy for grain growth ($42,400 \text{ J mol}^{-1}$), R the universal gas constant ($8.3145 \text{ J mol}^{-1} \text{K}^{-1}$), T the instantaneous layer temperature (K), and T_b the instantaneous temperature of the bottom firn layer. D is a constant and MO is a correction term that improves model alignment with observations ~~at depths~~ of the 550 and 830 kg m^{-3} density ~~levels~~ level depths (Ligtenberg et al., 2011). The correction term differs for both domains (Brils et al., 2021; Veldhuijsen et al., 2023). In this equation, ~~$e^{-\frac{E_c}{RT}}$~~ $e^{-\frac{E_c}{RT}}$ represents the temperature-dependency of ice deformation; ~~$e^{\frac{E_g}{RT_b}}$~~ $e^{\frac{E_g}{RT_b}}$ is a correction term arising from the estimated grain size ($E_c > E_g$). Ice grains grow faster in warmer conditions, while large grains deform slower than smaller grains, hence reducing the densification. As the current grain size is the result of the grain growth since deposition, its impact on densification must be approximated by the typical temperature conditions. Therefore, the mean firn temperature, estimated by the temperature of the lowermost layer (T_b), is used in Eq. ~~(3)~~ 3.

The physical description for water percolation and ~~thermodynamics~~ heat transport is identical in both model versions. Liquid water is added to the top of the column. Its transport through the layers is calculated using the 'bucket scheme'. In this scheme, each layer has an irreducible water content which is the maximum amount of water that can be retained in the layer by capillary forces. The irreducible water content decreases with increasing density (Coléou and Lesaffre, 1998). Water refreezes when enough cold content is available. ~~This Ice layers form through refreezing of liquid water and are assigned a density of 917 kg m^{-3} . The bucket scheme does not allow standing water over ice layers. Instead, water percolates to the next layer when~~ no pore space is available. The water percolation formulation implies that, in one model time step, water can percolate through the whole firn layer until it reaches the firn-ice interface. Any leftover water is assumed to run off.

Heat conduction through the firn layer is described by the 1D heat transfer equation, with conductivity being a function of density (Calonne et al., 2019; Brils et al., 2021). Refreezing acts as a heat source within the firn. Heat transfer at the upper boundary is governed by the prescribed surface temperature. A constant heat flux through the bottom layer is applied as the lower boundary condition.

2.2 Forcing dataset

IMAU-FDM is forced at the surface with surface temperature, 10 m windspeed, snowmelt, snowfall, sublimation, snowdrift erosion, and rainfall. These climate forcings originate from the regional atmospheric climate model RACMO2.3p2 (Noël et al., 2018; Van Wessem et al., 2018), dynamically downscaling ERA5 reanalysis data (Hersbach et al., 2020). The AIS ~~/APIS~~ data covers the period 1979 - 2023 with a horizontal resolution of $27 \times 27 \text{ km}^2$ (Van Wessem et al., 2018). The GrIS data spans 1 September 1939 - 1 December 2023 with a horizontal resolution of $5.5 \times 5.5 \text{ km}^2$ and is a renewed and extended version of the simulation presented by Noël et al. (2018).

2.3 Initialization

The firn layer is initialized by spinning up the model over a repeated reference period until it reaches a steady state where the firn air content (FAC) and elevation no longer change. The full firn layer, up to the pore close-off density of 830 kg m^{-3} , is refreshed during the initialization. As a prerequisite, this reference period should not exhibit large climate trends. The AIS/APIS simulation spin-up period covers 1979–2020 (Veldhuijsen et al., 2023). After 2020, Antarctica has gained mass due to unprecedented precipitation (Wang et al., 2023), and therefore 2020–2023 is excluded from the spin-up period. The GrIS SMB has been decreasing since 1990 (Shepherd et al., 2012). Therefore, we chose the first 30 years of the run (1 September 1939–31 December 1970) as the reference period.

2.3 Climate forcing time step

We force the model with four climate forcing time steps (i.e. $\frac{dt_{force}}{dt}$) at the surface: 3-hours, 6-hours, 1-day and 1-month, referred to as $\frac{dt_{force}}{3h}$, $\frac{dt_{force}}{6h}$, $\frac{dt_{force}}{1d}$, and $\frac{dt_{force}}{1m}$ respectively. We chose these values because (1) they are typically used to force firn models (Table ??), and (2) they provide examples with and without a diurnal cycle in the input data (Fig. 1). In addition, $\frac{dt_{force}}{3h}$ is the smallest available climate forcing time step from RACMO2.3p2 and is commonly used to force IMAU-FDM. We averaged the 3-hourly surface mass fluxes, surface temperature, and 10 m wind speed data to get the input data for $\frac{dt_{force}}{6h}$, $\frac{dt_{force}}{1d}$, and $\frac{dt_{force}}{1m}$. IMAU-FDM interpolates the forcing data onto a 15-minute model time step in which we assume a constant value for the whole forcing period for the surface mass fluxes, surface temperature and 10 m wind speed (Fig. 1). To summarize, numerical time integration is done using a model time step of 15 minutes, but the climate forcing time step can be 3-hours, 6-hours, 1-day, or 1-month.

2.4 Initialization

In IMAU-FDM, the firn layer is initialized by spinning up the model over a repeated reference period until it reaches a steady state where the firn air content (FAC) and surface height no longer change. FAC is the vertically integrated pore space in a column expressed in meters, representing the change in depth that would result from compressing the firn layer to the density of glacier ice (910 kg m^{-2}). The full firn layer, up to the pore close-off density of 830 kg m^{-3} , is refreshed during the initialization. To avoid shocks after the initialization, this reference period should not exhibit large climate trends. The AIS spin-up period covers 1979 - 2020 (Veldhuijsen et al., 2023). After 2020, the AIS has gained mass due to higher than average precipitation (Wang et al., 2023), and therefore 2020 - 2023 is excluded from the spin-up period. The GrIS SMB has been decreasing since 1990 (Shepherd et al., 2012). Therefore, we chose 1 September 1939 - 31 December 1970 as the reference period. The climate forcing time step was identical for spin-up and the actual simulation.

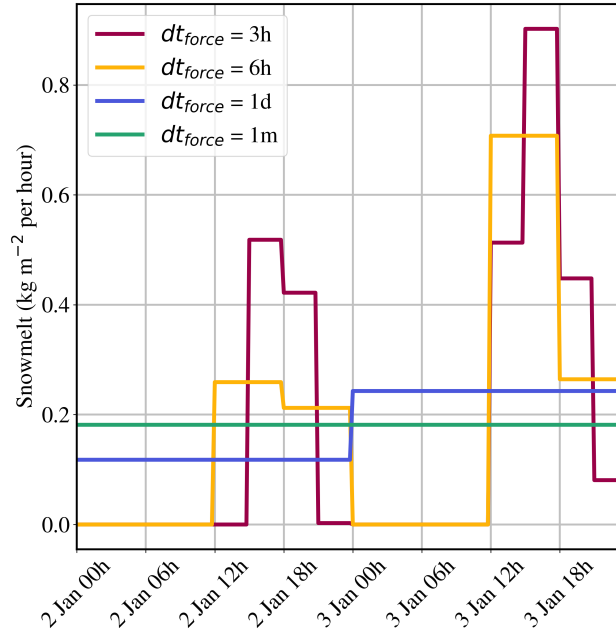


Figure 1. Example of 48 hours of snowmelt forcing at $dt_{force} = 3h$, $dt_{force} = 6h$, $dt_{force} = 1d$, and $dt_{force} = 1m$. The graph also visualizes the model's interpolation of the forcing onto a 15-minute timestep. Note the presence of a diurnal cycle in the snowmelt input for $dt_{force} = 3h$ and $dt_{force} = 6h$ and its absence for $dt_{force} = 1d$ and $dt_{force} = 1m$.

2.5 Study areas: APIS and southern GrIS

We select the APIS and the southern GrIS, two areas with a broad range of surface climate conditions, from low to high snowfall accumulation and with and without melt (Fig. 2). Accumulation represents the net deposition snowfall, sublimation and drifting snow processes (deposition/erosion).

150 In the APIS, the main climate gradient is from the wet and warm west to the dry and cold east. The ice shelves are the lowest-lying areas (Fig. 2a) and experience most snowmelt (up to $500 \text{ kg m}^{-2} \text{ yr}^{-1}$) (Fig. 2b), whereas the snowmelt is limited on the higher-elevated grounded ice. In contrast, the grounded ice has more accumulation, peaking in the north-west of the APIS. Accumulation is less on the ice shelves (Fig. 2c).

155 In the southern GrIS, the lowest-lying marginal areas (Fig. 2d) experience most snowmelt ($>750 \text{ kg m}^{-2} \text{ yr}^{-1}$) (Fig. 2e). Accumulation is highest in the east ($> 2000 \text{ kg m}^{-2} \text{ yr}^{-1}$) (Fig. 2f). The combination of Snowmelt rates exceeding $500 \text{ kg m}^{-2} \text{ yr}^{-1}$ snowmelt and combined with high accumulation, like as in the southeastern GrIS, can result in the presence of typically results in firm aquifers, or, in drier climate zones like the southwest, in ice slab formation slabs (Brils et al., 2024).

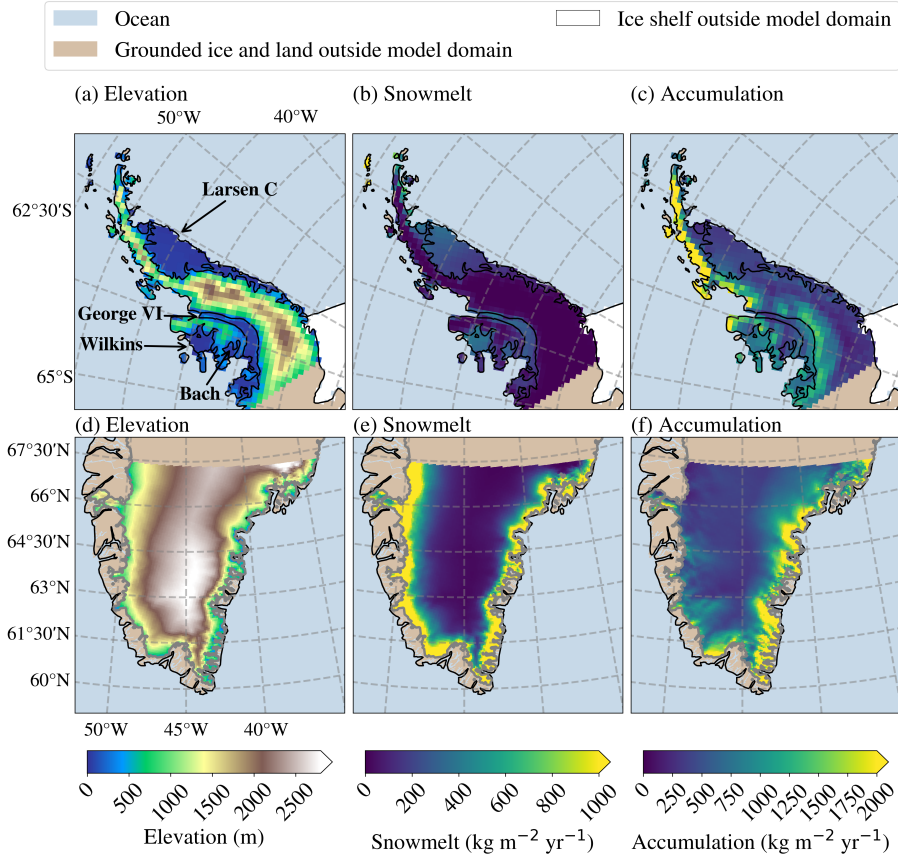


Figure 2. Maps showing elevation (a, d), snowmelt averages (b, e), and accumulation averages (c, f) for the modelled domain in the APIS (a–c) and southern GrIS (d–f) during the spin-up period, i.e. 1979 - 2020 for the APIS and 1 September 1939 – 31 December 1970 for the southern GrIS. The black contours show the coastal and ice shelf boundaries in the APIS (a–c) and the coastal boundaries in the southern GrIS (d–f). Ice shelf names within the model domain are indicated for the APIS (a). Gray contours show the ice mask boundaries in southern GrIS (d–f).

3 Effect of climate forcing timestep on firn

We look into the FAC ~~patterns for $\frac{dt^{force}}{dt_{3h}}$ for $dt_{force} = 3h$~~ for $dt_{force} = 3h$ and compare this to FAC with larger $\frac{dt^{force}}{dt_{force}}$. We show that increasing $\frac{dt^{force}}{dt_{force}}$ leads to more firn pore space. We ~~then identify the present maps of FAC for different dt_{force} to understand how climate regimes and dt_{force} influence the FAC, showing the relevance of identifying the~~ processes responsible for these differences. We discuss how these processes depend on the $\frac{dt^{force}}{dt_{force}}$ and how this affects the depletion or creation of pore space. In this section we focus on the effect on total FAC, an important quantity for retention studies.

In the discussion of these results, we use the following abbreviations: FAC_{3h} , FAC_{6h} , FAC_{1d} , and FAC_{1m} . FAC_{3h} , FAC_{6h} , FAC_{1d} , and FAC_{1m} correspond to the FAC at $\frac{dt^{force}}{dt_{3h}}$, $\frac{dt^{force}}{dt_{6h}}$, $\frac{dt^{force}}{dt_{1d}}$, and $\frac{dt^{force}}{dt_{1m}}$ $dt_{force} = 3h, 6h, 1d, \text{ and } 1m$ respectively.

The FAC_{3h} - FAC_{3h} compared to FAC simulated at a larger $\frac{dt_{force}}{dt_{force}}$, is defined as $\Delta FAC_{6h} - \Delta FAC_{6h} = FAC_{6h} - FAC_{6h} - FAC_{3h}$, $\Delta FAC_{1d} - \Delta FAC_{1d} = FAC_{1d} - FAC_{3h}$, and $\Delta FAC_{1m} - \Delta FAC_{1m} = FAC_{1m} - FAC_{3h}$. ΔFAC (without subscript) denotes the difference between FAC_{3h} - FAC_{3h} and FAC modelled with larger $\frac{dt_{force}}{dt_{force}}$. A positive ΔFAC thus indicates more pore space for the larger climate forcing time step larger than 3-hours.

3.1 FAC_{3h} - FAC_{3h} and ΔFAC

Figure 3 shows the FAC_{3h} - FAC_{3h} (Figs. 3a and e), $\Delta FAC_{6h} - \Delta FAC_{6h}$ (Figs. 3b and f), $\Delta FAC_{1d} - \Delta FAC_{1d}$ (Figs. 3c and g), and $\Delta FAC_{1m} - \Delta FAC_{1m}$ (Figs. 3d and h) for the APIS and southern GrIS. Panels 3b, c, d, f, g, and h show that FAC_{3h} - FAC_{3h} is lower than FAC with larger $\frac{dt_{force}}{dt_{force}}$ in both the APIS and southern GrIS. The larger the $\frac{dt_{force}}{dt_{force}}$ becomes, the higher the ΔFAC .

On the APIS, FAC is smallest (0-5 m for FAC_{3h} - FAC_{3h}) in the lower-lying areas, corresponding to the ice shelves (Fig. 3a). ΔFAC is largest on the ice shelves (15 % for $\Delta FAC_{1d} - \Delta FAC_{1d}$, increasing up to 44% for $\Delta FAC_{1m} - \Delta FAC_{1m}$). $\Delta FAC_{6h} - \Delta FAC_{6h}$ is relatively small (3%). Interestingly, an area with a negative ΔFAC is situated along the high-elevation southern spine of the APIS (red hues in Figs. 3b, c, and d). In a relative sense, these differences are small, of the order of ~ 0.5 m on typical FAC_{3h} - FAC_{3h} values of 15-20 m. In both an absolute and a relative sense, ΔFAC is largest for the locations with low FAC_{3h} - FAC_{3h} .

In the southern GrIS, FAC increases from the ice margin where it equals zero in the ablation zone, to higher elevations, with the steepest spatial gradients on the east coast (Fig. 3e). In the southern GrIS, in the ablation zone, most clearly visible on the western margin, ΔFAC is small (Figs. 3f, g, and h). Here, firn pore space is depleted across all simulations, regardless of $\frac{dt_{force}}{dt_{force}}$. At high elevations (>2000 m), ΔFAC is also relatively small, with average differences $< 1\%$ in $\Delta FAC_{6h} - \Delta FAC_{6h}$ up to 3% in $\Delta FAC_{1m} - \Delta FAC_{1m}$. At intermediate elevations, a band with larger ΔFAC is apparent, increasing with greater $\frac{dt_{force}}{dt_{force}}$ (Fig. 3f, g, and h). In the east and west, ΔFAC in this band is quite similar in an absolute sense: on average, $\Delta FAC_{6h} - \Delta FAC_{6h} = 0.3$ m, $\Delta FAC_{1d} - \Delta FAC_{1d} = 2.1$ m, and $\Delta FAC_{1m} - \Delta FAC_{1m} = 4.0$ m. However, relative differences are larger in the west, where absolute FAC_{3h} - FAC_{3h} values are much smaller (typically < 20 m) than in the east (typically > 20 m).

3.2 Dominant processes responsible for ΔFAC in IMAU-FDM

We discuss the processes that contribute to ΔFAC . Figure 4 gives an overview of the dominant processes in a firn layer that introduce the forcing time step dependency of the FAC and that are described in depth later in this section. The processes described below are computed every 15 minute model time step during both the spin-up and the simulation periods.

FAC is governed by a balance between the processes that create pore space (snowfall) and the processes that deplete it (densification, snowmelt, snowdrift erosion, and sublimation). Figure 5 shows the relative importance of several processes determining FAC_{3h} - FAC_{3h} and FAC_{1d} - FAC_{1d} for two example locations depicted in Fig. 3. Figure 5a represents a location

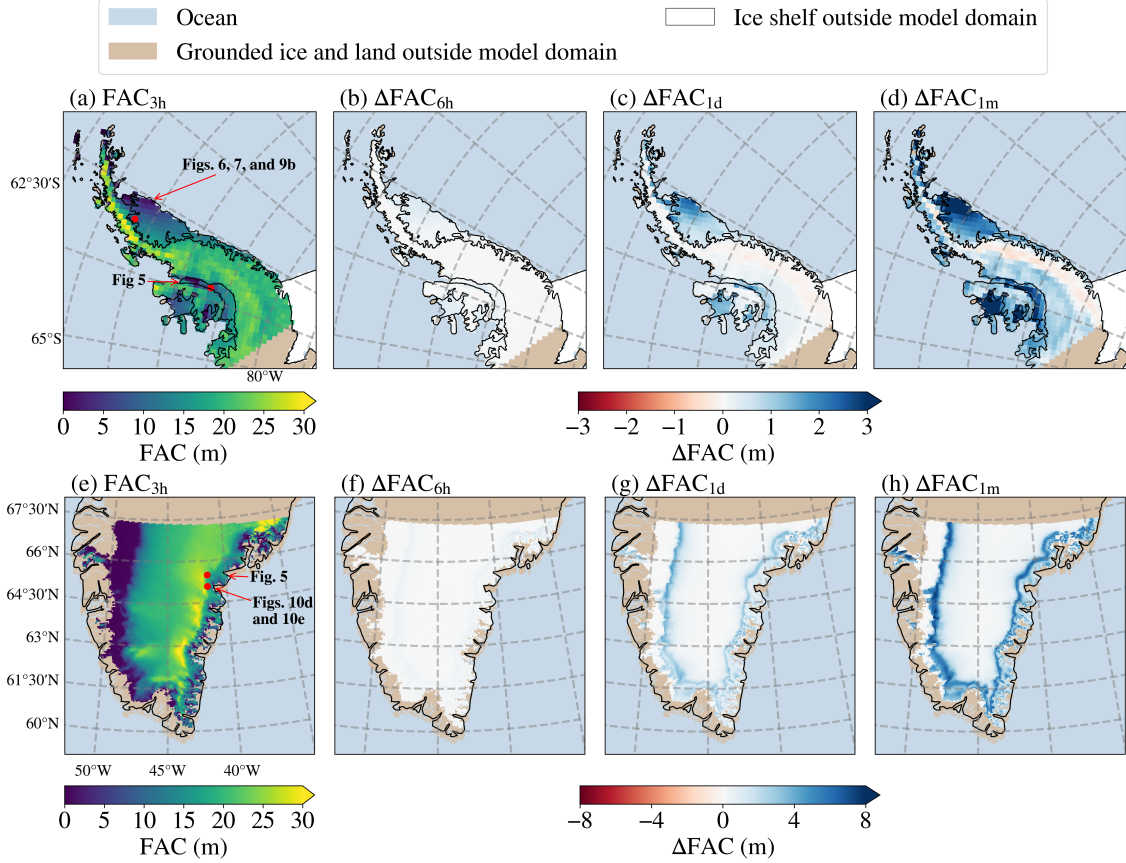


Figure 3. Maps showing average FAC_{3h} (a, e), along with average ΔFAC_{6h} (b, f), ΔFAC_{1d} (c, g), and ΔFAC_{1m} (d, h) in the APIS (a-d) and southern GrIS (e-h). The averages are over the whole simulation period, i.e. 1979 - 2023 for the APIS and 1 September 1939 – 31 December 2023 for the southern GrIS. The points highlighted in panel (a) and (e) are used as an example later in this paper.

in the APIS and Fig. 5b a location in the southern GrIS. Both locations experience similar melt and ΔFAC_{1d} , but the location in-on the GrIS has more snowfall. For clarity, a positive excursion of the elevation change difference in panels 5c and d means that the elevation increases more for $\frac{dt_{force}}{dt_{1d}} = 1d$.

Snowfall is the major source of firm air in both locations (Figs. 5a and b). The contribution of snowfall to FAC for the APIS is slightly higher for $\frac{dt_{force}}{dt_{3h}}$ than $\frac{dt_{force}}{dt_{1d}}$ than for $\frac{dt_{force}}{dt_{1d}} = 3h$ than for $\frac{dt_{force}}{dt_{1d}} = 1d$ (Fig. 5c), whereas it is almost the same for the GrIS (Fig. 5d). Larger differences arise for processes that reduce FAC. In both locations, more FAC_{3h} is removed by snowmelt than FAC_{1d} (Figs. 5c and d). For the dry location (Fig. 5a), densification is relatively slow, and FAC depletion due to melt is largest. For the wet location (Fig. 5b), the densification dominates reduction of the FAC. At both locations, melt removes more FAC_{3h} and has slower firm densification at $\frac{dt_{force}}{dt_{3h}} = 3h$ (Figs. 5c and d).

1. Melt removes more FAC with smaller forcing time steps

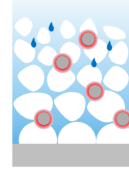
Melt event:

Melt is prescribed. Firn mass in the first layer is converted to the prescribed amount of melt.



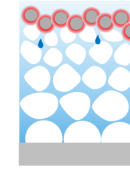
$dt_{force} = 3h$

- $T_s = 273.15\text{ K}$
- Refreezing at depth
- Deeper percolation
- ↓ Surface density



$dt_{force} = 1d$ (non-physical)

- $T_s < 273.15\text{ K}$
- Immediate refreezing at surface
- Shallower percolation
- ↑ Surface density



↓ = less / lower
↑ = more / higher

Refrozen grain
and heat release

Melt event next time step: firn → liquid water

↑ FAC removal

↓ FAC removal

2. Greater forcing timesteps both increase and decrease densification

1. Related to the amount of firn

$dt_{force} = 3h$

- ↓ FAC
- ↓ Densification



$dt_{force} = 1d$

- ↑ FAC
- ↑ Densification



2. Related to the densification equation (Eq. 3)

$dt_{force} = 3h$

- Refreezing at depth
- Heat remains in firn
- ↑ $T \rightarrow \uparrow$ densification
- ↑ $T_b \rightarrow \downarrow$ densification

$dt_{force} = 1d$

- Refreezing at surface
- Heat releases to atmosphere
- ↓ $T \rightarrow \downarrow$ densification
- ↓ $T_b \rightarrow \uparrow$ densification

T = instantaneous layer temperature

T_b = instantaneous bottom layer temperature

3. Low 10 m wind speeds during snowfall increase FAC for smaller forcing time steps

Only applies to the Antarctic fresh snow density parameterization (Eq.1)

$dt_{force} = 3h$

- ↓ 10 m wind during snowfall
- ↓ Crystal breaking
- ↓ Efficient packing
- ↓ Fresh snow density



$dt_{force} = 1d$ (non-physical)

- ↑ 10 m wind during snowfall
- ↑ Crystal breaking
- ↑ Efficient packing
- ↑ Fresh snow density

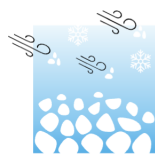


Figure 4. Processes that control Δ FAC. The uppermost row shows the effect of a diurnal cycle in dt_{force} on melt and refreezing, and how those influence FAC. The middle row shows that dt_{force} can both increase and decrease densification. In the lowermost row, the effect of fresh snow density parameterization in the AIS is shown.

Summarizing, Δ FAC- Δ FAC is mainly controlled by melt and densification rate, with a minor contribution from snowfall. Now, we further zoom in on those processes to understand the Δ FAC- Δ FAC they cause.

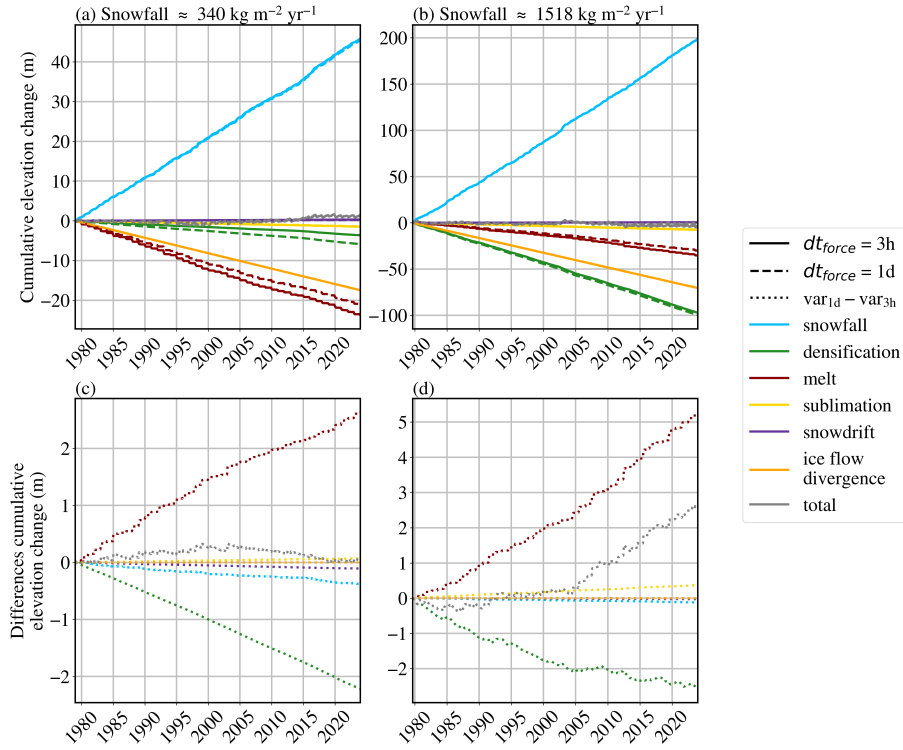


Figure 5. (a) and (b) Cumulative elevation change (m) due to the processes that influence the firn layer. (c) and (d) differences in cumulative elevation change for all variables between $\frac{dt_{force}}{dt} = 1d$ and $\frac{dt_{force}}{dt} = 3h$. Firn densification is integrated over the whole firn column. All other variables represent surface processes. Results for $\frac{dt_{force}}{dt} = 3h$ are in solid lines, for $\frac{dt_{force}}{dt} = 1d$ are in dashed lines, and the difference between the two is in dash-dot lines. Panel (a) and (c) are a location in the APIS and panel (b) and (d) are a location in southern GrIS. Both locations experience similar melt ($\sim 340 \text{ kg m}^{-2} \text{ yr}^{-1}$) and $\Delta FAC_{1d} - \Delta FAC_{3h}$ (2.1 m), but different snowfall. See Fig. 3a and e for a map with the grid cell locations.

210 3.2.1 Process 1: melt

Melt provides the largest contribution to ΔFAC , simulations with $\frac{dt_{force}}{dt} = 3h$ have more FAC reduction than for larger $\frac{dt_{force}}{dt}$ (Figs. 5c and d). Climate zones with more melt, have a higher ΔFAC . We identify two types of melt-related mechanisms responsible for ΔFAC , namely a surface-based process and a process occurring at depth.

At the surface, the model is forced with a melt flux. As melt is not generated within the model, and firn in the top layer is converted to liquid water (Fig. 4, process 1, step 1). The subsequent fate of the meltwater depends on $\frac{dt_{force}}{dt}$. With a diurnal cycle in $\frac{dt_{force}}{dt}$, i.e. 3- and 6-hours, surface temperatures reach 0°C during melt events. Therefore, the water cannot immediately refreeze and the timing of refreezing and snowmelt in the top layer alternate between day and night (Fig. 6a). In contrast, without a diurnal cycle in $\frac{dt_{force}}{dt}$, i.e. 1-day and 1-month, surface melt and temperature become decoupled. As a result, melt can occur at subfreezing surface temperatures and meltwater refreezes immediately near the

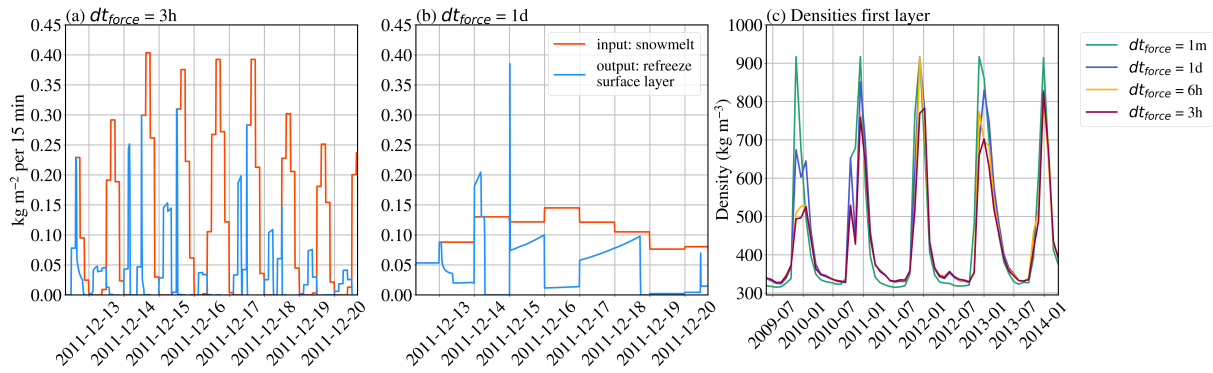


Figure 6. Modelled snowmelt-Snowmelt input and modelled refreezing in the surface layer for $\frac{dt_{force}}{dt_{3h}} = 3h$ (a) and for $\frac{dt_{force}}{dt_{1d}} = 1d$ (b). Panel (c) shows the modelled densities in the surface layer. See Fig. 3a for a map with the grid cell location.

220 surface (Fig. 6b and Fig. 4, process 1, step 2). This leads to a higher near-surface density (more ice) under larger $\frac{dt_{force}}{dt_{force}}$ (Fig. 6c). Subsequent melt events convert firn to liquid water again and remove less pore space when a high-density surface layer is present, as in $\frac{dt_{force}}{dt_{1d}}$ and $\frac{dt_{force}}{dt_{1m}}$ (Fig. 4, process 1, step 3). $dt_{force} = 1d$ and $1m$. This melt and surface temperature interaction in absence of a daily cycle is the main mechanism that leads to higher ΔFAC with increasing dt_{force} . ΔFAC with increasing dt_{force} (Fig. 4, process 1).

225 The second process is that meltwater percolates deeper at $\frac{dt_{force}}{dt_{3h}}$ (Figs. 7a and b and Fig. 4, process 1, step 2) $dt_{force} = 3h$; less water is immediately refrozen near the surface. Consequently, refreezing occurs deeper if there is enough cold content τ (Figs. 7a and b and Fig. 4 process 1). The latent heat released by refreezing at depth is largely retained-trapped due to the low thermal conductivity of firn. Thus, the deeper firn for $\frac{dt_{force}}{dt_{3h}} = 3h$ is warmer (Figs. 7c and d and Fig. 4, process 1, step 2). Warmer firn enhances firn densification, as discussed in the following subsection.

230 3.2.2 Process 2: firn densification

Several processes contribute to the firn densification rate. First, more snowfall leads to more firn that can densify τ (contribution of snowfall to $\dot{b}$ in Eq. 3). Simulations with larger $\frac{dt_{force}}{dt_{force}}$ have more firn dt_{force} have more FAC (Fig. 3), and therefore undergo more densification (Fig. 5 and Fig. 4, process 2).

235 Second, the densification rate is governed by the instantaneous layer temperature (T in Eq. 3) and the temperature of the bottom layer (T_b in Eq. 3), with higher T enhancing densification and higher T_b reducing it (Fig. 4, process 2). When both A similar increase in T and T_b increase, densification is enhanced would lead to enhanced densification (see constants in Eq. 3). As-Because the simulation with $\frac{dt_{force}}{dt_{3h}} = 3h$ has deeper refreezing (Fig. 7a compared to Fig. 7b) and warmer firn (Fig. 7c compared to Fig. 7d), the densification is indeed faster. An exception occurs when T is similar for different dt_{force} dt_{force} for the uppermost meters of the firn column, but T_b is higher for $\frac{dt_{force}}{dt_{3h}} = 3h$ due to the deeper refreezing. Then the larger approximated grain size for $\frac{dt_{force}}{dt_{3h}} = 3h$ slows down the densification rate, leaving more FAC FAC.

240

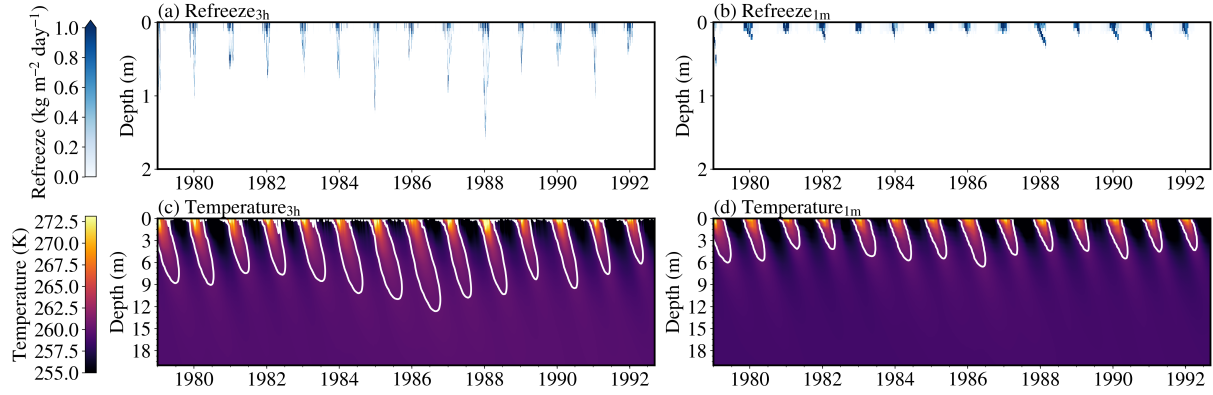


Figure 7. Refreezing amount and depth (a, b) and temperature profiles (c, d) for an example point on the APIS for $\frac{dt_{force}}{dt_{3h}} = 3h$ (a, c) and for $\frac{dt_{force}}{dt_{1m}} = 1m$ (b, d). The white contours in panels c and d represent a temperature of 260 K. See Fig. 3a for a map with the grid cell location.

To summarize, higher firm temperatures at $\frac{dt_{force}}{dt_{3h}} = 3h$ increase densification rate, ~~but overall densification remains lower as there is less firm.~~ However, the reduction in firm depth resulting from densification (Fig. 5) remains smaller as less pore space is available to densify.

3.2.3 Process 3: ~~snowfall~~fresh snow density on the APIS

- 245 An area with lower rather than higher ΔFAC ΔFAC is modelled in the APIS (red hues in Figs. 3b, c, and d); this relates to differences in the estimated fresh snow density for various $\frac{dt_{force}}{dt_{force}}$. The simulation for $\frac{dt_{force}}{dt_{3h}} = 3h$ has more 10 m wind speed ff_{10m} (ff_{10m}) variability than for larger $\frac{dt_{force}}{dt_{force}}$, which influences fresh snow density through Eq. 1: fresh snow density increases with surface temperature T_s (Judson and Doesken, 2000) and ff_{10m} (Sato et al., 2008) (Fig. 4, process 3) ff_{10m} (Sato et al., 2008). The area with negative ΔFAC ΔFAC typically experiences snowfall at low ff_{10m} ff_{10m} .
- 250 Lower wind speeds during snowfall lead to less crystal breaking, and therefore, less efficient packing at the surface, decreasing the fresh snow density (Fig. 4, process 3). Then, the snowfall-weighted average 10 m windspeed ($\overline{ff_{10m}^{sf}}$ $\overline{ff_{10m}^{sf}}$) (Eq. 4) for $\frac{dt_{force}}{dt_{3h}} = 3h$ is lower than for higher $\frac{dt_{force}}{dt_{force}}$ larger $\frac{dt_{force}}{dt_{force}}$ (Fig. 8).

$$\overline{ff_{10m}^{sf}} = \frac{\sum snowfall(t) * 10m - windspeed(t)}{\sum snowfall(t)} \frac{\sum snowfall(t) * ff_{10m}(t)}{\sum snowfall(t)} \quad (4)$$

- Lower $\overline{ff_{10m}^{sf}}$ Thus, lower $\overline{ff_{10m}^{sf}}$ is related to lower fresh snow densities, and therefore higher FAC_{3h} FAC_{3h} . In climate zones with limited melt ($< 30 \text{ kg m}^{-2} \text{ yr}^{-1}$), this effect of the fresh snow density causes a relatively small ΔFAC and possible negative ΔFAC of up to $\pm 0.5 \text{ m}$ on typical FAC_{3h} FAC_{3h} values of 15-20 m (Figs. 3a, b, c, and d).

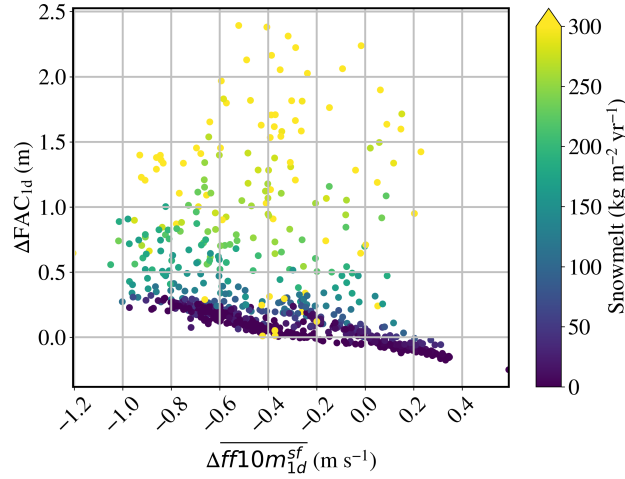


Figure 8. The ΔFAC_{1d} against $\Delta \overline{ff10m_{1d}^{sf}}$ in the APIS. $\Delta \overline{ff10m_{1d}^{sf}} - \Delta \overline{ff10m_{1d}^{sf}} = \overline{ff10m_{1d}^{sf}} - \overline{ff10m_{3h}^{sf}} - \overline{ff10m_{1d}^{sf}} - \overline{ff10m_{3h}^{sf}}$. In absence of melt, there is a clear relation between $\Delta \overline{ff10m_{1d}^{sf}}$ and ΔFAC_{1d} .

4 Implications ~~ice sheets~~for IMAU-FDM

Climatic differences between the APIS and southern GrIS lead to a different response of ΔFAC to dt^{force} . ΔFAC to dt^{force} in IMAU-FDM. In turn, ΔFAC modifies the firn correction to altimetry. These implications are discussed in the following sections for both ice sheets.

4.1 APIS

Figure 9a shows that $\Delta FAC - \Delta FAC$ increases with snowmelt. Accordingly, ice shelves have largest positive $\Delta FAC - \Delta FAC$ values (Figs. 3b, c, and d), reflecting their relatively high snowmelt rates (Fig. 2b). Consequently, the chosen $dt^{force} - dt^{force}$ is crucial in assessing the vulnerability of an ice shelf in terms of firn air; whether the firn air is depleted or could still buffer meltwater. Ice shelves are more vulnerable to hydrofracturing when the firn air is depleted. Therefore, neglecting the daily temperature and melt cycle in firn simulations could lead to underestimated ice shelf vulnerability.

Fig. 9b presents a FAC timeseries of a single grid cell on the Larsen C ice shelf (see Fig. 3 for location). The overall variations in FAC are similar across all simulations, with $\Delta FAC - \Delta FAC$ primarily established during the spin-up. This is in line with the expected behavior of firn layers. Although the main initial forcing to ΔFAC , namely reduced FAC. The equilibrium state of the firn column has higher FAC for larger dt^{force} , where reduced FAC removal during melt events for larger dt^{force} , is short-lived, the extra remaining FAC for dt^{force} is removed is balanced by enhanced densification on decadal timescales. Consequently, the equilibrium state of the firn column is with higher FAC for larger dt^{force} .

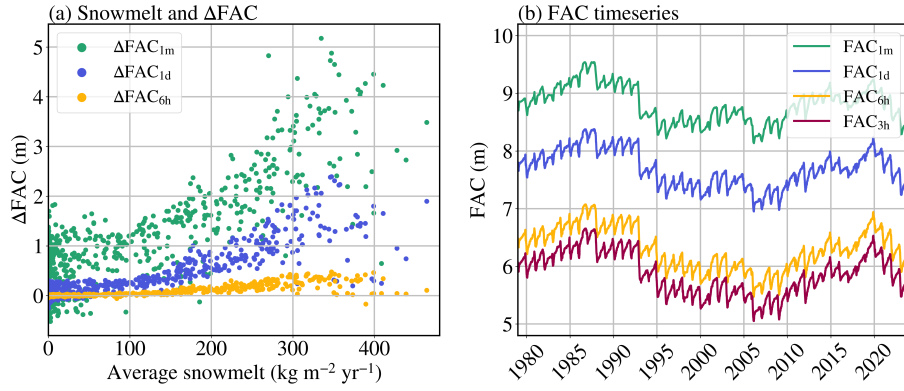


Figure 9. (a) Average ΔFAC_{6h} , ΔFAC_{1d} , and ΔFAC_{1m} over the simulation period against average snowmelt over 1979 - 2020. (b) FAC timeseries of one grid point on the APIS for the four $\frac{dt_{force}}{dt}$. See Fig. 3a for a map with the grid cell location.

4.2 Southern GrIS

The wider range of climatic conditions in terms of snowmelt and accumulation in the southern GrIS (Fig. 10a) compared to the APIS (Fig. 9a), complicates the isolation of processes governing ΔFAC . To facilitate the discussion, we divide the accumulation-snowmelt space into four domains (Fig. 10a).

The largest increases in ΔFAC_{1d} occur at snowmelt rates of $\sim 500 \text{ kg m}^{-2} \text{ yr}^{-1}$ and near a Melt-Over-Accumulation (MOA) ratio of 1 (Fig. 10a). Climate zones with snowmelt below $500 \text{ kg m}^{-2} \text{ yr}^{-1}$ (Fig. 10a, climate zone I), corresponding to the interior of the southern GrIS (Fig. 2e), behave similarly to the APIS: higher snowmelt rates increase ΔFAC_{1d} , with more ΔFAC gain for larger $\frac{dt_{force}}{dt}$.

Firn aquifers can form when snowmelt exceeds $500 \text{ kg m}^{-2} \text{ yr}^{-1}$ (Kuipers Munneke et al., 2014; Brils et al., 2024). In climate zone II (Figs. 10a and b), which are the points with annual snowmelt between 500 and $800 \text{ kg m}^{-2} \text{ yr}^{-1}$, firn aquifers formed under $\frac{dt_{force}}{dt} = 3h$ during spin-up, but did not for $\frac{dt_{force}}{dt} = 1d$. As meltwater refreezing occurs deeper for $\frac{dt_{force}}{dt} = 3h$, the firn layer $\frac{dt_{force}}{dt} = 1d$. The firn has a higher heat content throughout the entire column for $\frac{dt_{force}}{dt} = 3h$, due to the deeper refreezing of meltwater. The higher heat content limits the refreezing capacity. We hypothesize that more liquid water remains in the firn layer and the aquifer can refill during the next melt season. In contrast, the cold content for $\frac{dt_{force}}{dt} = 1d$ still facilitates refreezing of all liquid water, and therefore, aquifers did not form during spin-up.

In climate zone III, snowmelt exceeds $800 \text{ kg m}^{-2} \text{ yr}^{-1}$ and the MOA is below 1. Firn aquifers form under both $\frac{dt_{force}}{dt} = 3h$ and $\frac{dt_{force}}{dt} = 1d$ (Fig. 10b). Here, ΔFAC_{1d} increases with snowmelt as more pore space is removed for larger $\frac{dt_{force}}{dt}$ (Fig. 4, process 1), but this process is increasingly counterbalanced by enhanced firn densification at sites with higher accumulation (Fig. 10a). FAC_{3h} is lower in this climate zone, so less pore space is available for water storage, resulting in less liquid water content (LWC) for $\frac{dt_{force}}{dt} = 3h$ whereas the liquid water is still retained for $\frac{dt_{force}}{dt} = 1d$ (Fig. 10b, climate zone III). More excess water for $\frac{dt_{force}}{dt} = 3h$ leads to more runoff. This example

illustrates that the runoff limit also depends on $\frac{dt_{force}}{dt_{force}}$ (Fig. 10c). In the southeastern GrIS, the runoff limit shifts more
 295 than 10 km inland for $\frac{dt_{force}}{dt_{force}} = 3h$ compared to $\frac{dt_{force}}{dt_{force}} = 1d$, and up to 35 km compared to $\frac{dt_{force}}{dt_{force}} = 1m$. This result shows that $\frac{dt_{force}}{dt_{force}}$ plays a role in estimates of future GrIS mass balance projections.

When the MOA exceeds 1 (Fig. 10a climate zone IV), ΔFAC equals 0, as the firm is depleted in both simulations.

Unlike the example of the APIS (Fig. 9b), long-term FAC variability in Fig. 10d depends on $\frac{dt_{force}}{dt_{force}}$ (Fig. 10d). For
 300 $\frac{dt_{force}}{dt_{force}} = 3h$, firm aquifers are present at the start of the simulation, whereas for $\frac{dt_{force}}{dt_{force}} = 1d$ they develop around
 2005 (climate zone II) (Fig. 10e). In this year, FAC_{1d} diverges from FAC_{3h} , FAC_{1d} diverges from FAC_{3h} , and eventually,
 FAC_{1d} even drops below FAC_{3h} . This results from the densification equation (Eq. 3): liquid water in the
 aquifers sets the instantaneous temperature (T in Eq. 3) to 273.15 K. The bottom temperature (T_b in Eq. 3) is higher for $\frac{dt_{force}}{dt_{force}} = 3h$
 (section 3.2.1), because of the deeper refreezing and releasing of latent heat, slowing down densification.
 The faster densification under $\frac{dt_{force}}{dt_{force}} = 1d$ reduces FAC_{1d} . Hence, for locations where aquifers form
 305 during the simulation, the time series of FAC (and surface elevation) become different and their trends possibly of opposite
 sign. This has important implications for the correct interpretation of satellite altimetry.

4.3 Altimetry corrections

As simulated FAC depends on $\frac{dt_{force}}{dt_{force}}$, then so does firm depth and hence surface height change, which is in turn used to
 correct altimetry data. For the same APIS sample location as shown in Fig. 9b, surface height change under $\frac{dt_{force}}{dt_{force}} = 1d$ and
 310 $1m$ differs from $\frac{dt_{force}}{dt_{force}} = 3h$ by 0.2 and 0.4 m, respectively (Fig. 11a). ΔFAC already emerges during spin-up. At a GrIS
 location in transition to an aquifer, as shown in Fig. 10d and e, ΔFAC is primarily established after the spin-up and shows a
 stronger divergence of surface height changes, at 8 to 30% of the FAC (Fig. 11b). Altimetry corrections based on IMAU-FDM
 simulations with $\frac{dt_{force}}{dt_{force}} = 1d$ and $1m$ would give higher surface height increases for the two examples. Consequently, during
 periods of ice thinning, subtracting the modeled firm depth change from the altimetry signal leads to an underestimation of ice
 315 loss, while during periods of ice thickening it leads to an underestimation of ice gain.

5 Lessons learned for firm modelling

In this section we present the lessons learned of using different $\frac{dt_{force}}{dt_{force}}$ in firm modelling. We distinguish between
 general findings and findings specific to IMAU-FDM specifics and general findings.

First, we argue that firm models that prescribe snowmelt fluxes and surface temperature, like IMAU-FDM, need at least a
 320 diurnal cycle in the $\frac{dt_{force}}{dt_{force}}$. Snowmelt turned out to be $\frac{dt_{force}}{dt_{force}}$. Snowmelt is the major predictor of ΔFAC , increasing ΔFAC
 for larger $\frac{dt_{force}}{dt_{force}}$. Without resolving the daily cycle, snowmelt can co-exist with
 subfreezing surface temperatures, leading to immediate refreezing of meltwater. This decoupling of surface temperature and
 snowmelt leads to non-physical results. Firm models with surface energy balance schemes that compute rather than prescribe
 melt fluxes might be less sensitive to this finding, but might suffer from less accurate melt fluxes.

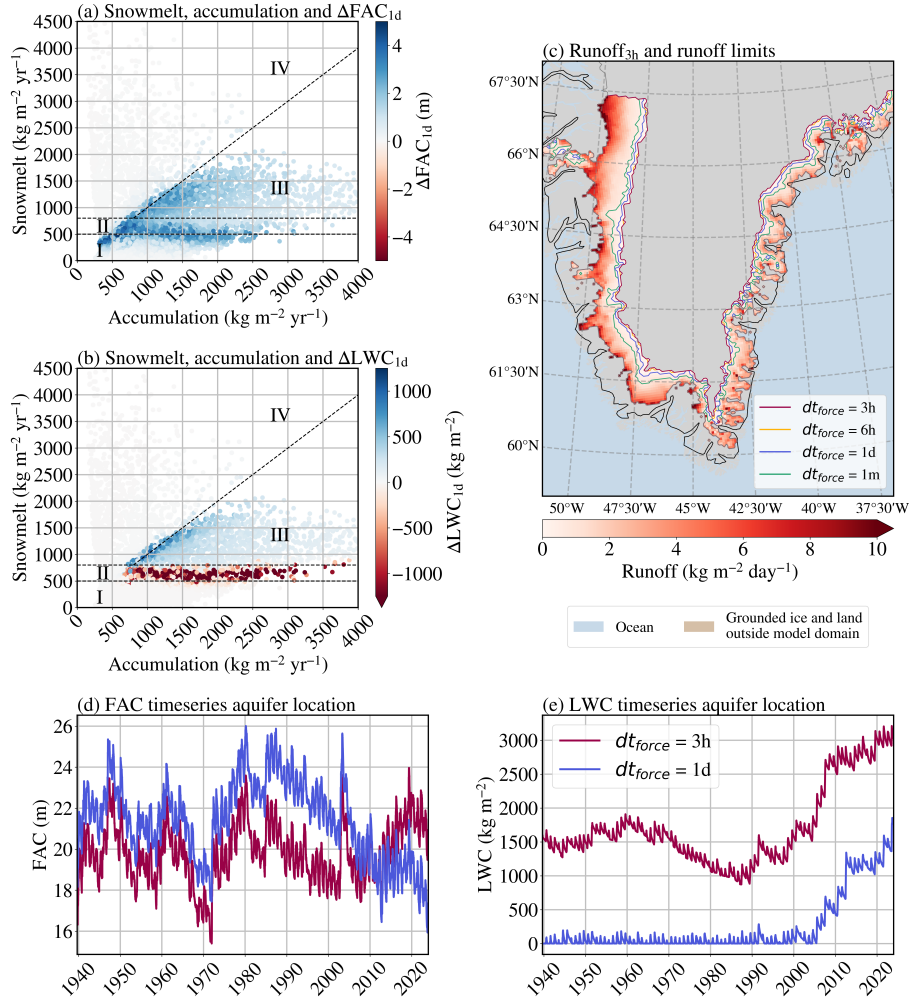


Figure 10. Average snowmelt against accumulation over 1939-1970 colour coded with average $\Delta F_{AC_{1d}}$ (a) and ΔLWC_{1d} (b) over the simulation period. The black dashed lines represent snowmelt of 500 kg m⁻² yr⁻¹ (climate zone I), 500-800 kg m⁻² yr⁻¹ (climate zone II) and a MOA of 1 (separating climate zones III and IV). Panel (c) shows the average runoff_{3h} over the simulation period and the different runoff limits for the four dt_{force} . Panel (d) and (e) show timeseries of respectively $F_{AC_{3h}}$ and $F_{AC_{1d}}$, and LWC_{3h} and LWC_{1d} . See Fig. 3e for a map with the grid cell location.

Second, a general lesson learned for modelling is that the original purpose and forcing time step of a parameterization should be respected when applying it. We showed how the fresh-snow density and densification parameterizations in IMAU-FDM are sensitive to dt_{force} . The fresh-snow density parameterization (Eq. 1) in the APIS is an example where the parameterization is tuned specifically using dt_{3h}^{force} (Veldhuijsen et al., 2023). ΔF_{AC} with $dt_{force} = 3h$ (Veldhuijsen et al., 2023). ΔF_{AC} arising from other dt_{force} are related to the parameterization (Fig. 8) and are not based on physical grounds. We also gave an example of the densification equation (Eq. 3) which is based on steady state and dry assumptions.

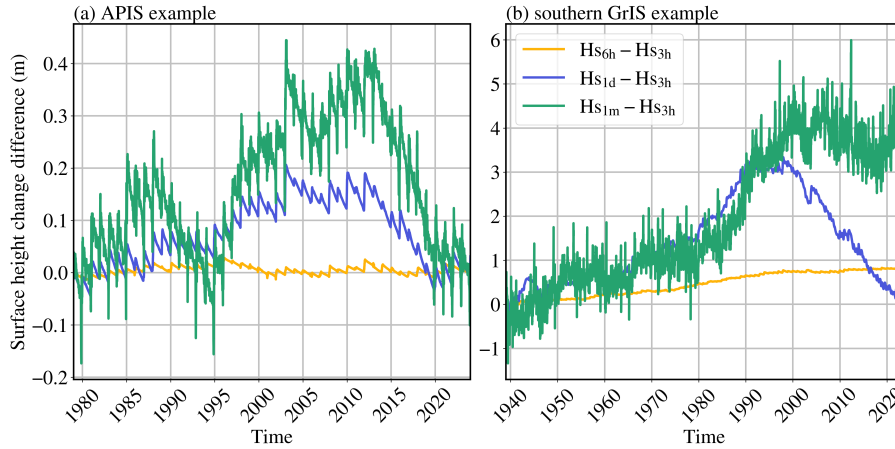


Figure 11. Differences in surface height changes over time between $dt_{force} = 3h$ and larger climate forcing time steps. Panel (a) represents the point in Fig. 9b and (b) the point in Fig. 10d and e.

snow. However, we applied the parameterization at larger dt_{force} and also for transient and wet conditions. The latter results in non-physical processes where the bottom temperature (T_b in Eq. 3) determines the densification rate in an aquifer and leads to different FAC trends when the model is forced at dt_{3h}^{force} or dt_{1d}^{force} variations over time for different dt_{force} (Figs. 10d and e). A general lesson learned for modelling is that the original purpose and forcing time step of a parameterization should be respected.

6 Conclusions

Ice-sheet firn models in literature use different surface climate forcing time steps. We In this study we investigated the effect on firn air content of using a 3-hour, 6-hour, 1-day, and 1-month climate forcing time step on firn air content modelled by IMAU-FDM in the APIS and southern GrIS, Antarctic Peninsula and southern Greenland Ice Sheet.

We find that increasing the climate forcing time step usually typically results in more firn air (ΔFAC positive ΔFAC). The strongest sensitivities are found in climate zones with low FAC_{3h} or where firn aquifers are absent under larger forcing time steps, but present under shorter ones. ΔFAC resulting from the forcing time step is governed by snowmelt, firn densification, and snowfall magnitude of this change depends on the climate regime and is most sensitive to snowmelt. More specifically, (1) snowmelt is the main contributor to the positive ΔFAC . Input data requires a the climate forcing should resolve the diurnal cycle when surface temperature and snowmelt are prescribed to a firn model at the top boundary. Otherwise, subfreezing temperatures and snowmelt can coexist, leading to immediate refreezing of the meltwater and, consequently, overestimating the amount of firn air; (2) parameterizations should be applied respecting their original purpose and temporal resolution. The fresh snow density parameterization determined the ΔFAC under limited snowmelt conditions and is related to dt_{force} on

non-physical grounds. The densification parameterization leads to non-physical ΔFAC for all dt^{force} when applied in wet
350 conditions. (3) the

We found that in IMAU-FDM, ΔFAC increases with snowmelt for melt rates up to $500 \text{ kg m}^{-2} \text{ yr}^{-1}$, across both the APIS
and southern GrIS. In the latter, snowmelt rates exceeding $500 \text{ kg m}^{-2} \text{ yr}^{-1}$ can lead to the formation of firm aquifers. A
smaller climate forcing time step result in earlier aquifer formation and therefore higher ΔFAC . Once aquifers have formed
 ΔFAC increases with snowmelt up to a melt-over-accumulation rate of 1, above which the firm becomes depleted for all climate
355 forcing time step is critical in assessing ice shelf vulnerability in APIS, the runoff limit in the GrIS, and the steps.

In IMAU-FDM, ΔFAC was also caused by the densification and fresh snow density parameterization for non-physical
reasons, i.e. when the original purpose and temporeal resolution of the parameterizations is not respected.

Ice sheet firm air content impacts the interpretation of altimetry observations, co-determines the ice sheet meltwater
retention potential and therewith runoff limits and ice shelf vulnerability to hydrofracturing. In firm models, the climate forcing
360 time step potentially influences the simulated firm air content and should therefore be carefully selected and tested.

Code and data availability. The IMAU-FDM v1.2A and v1.2G code is available on GitHub at <https://github.com/mbrils/IMAU-FDM-v1.2>
(Brils et al., 2021). For this paper we updated the subroutine interpol in ini_model.f90 to interpolate the forcing data onto a 15-minute model
time step with a constant value, instead of the linear interpolation presented in the GitHub model version. The datasets and code to create the
figures will be published on Zenodo.

365 *Author contributions.* Tvda, PKM, WJvdB and MvdB designed the study. Tvda performed and analyzed the simulations. All co-authors
contributed to discussions on the research and manuscript.

Competing interests. At least one of the (co-)authors is a member of the editorial board of *The Cryosphere*. The authors have no other
competing interests to declare.

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