Drifting snow climate of the Greenland ice sheet: a study with a regional climate model

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Received: 20 March 2012 – Published in The Cryosphere Discuss.: 3 May 2012
Revised: 18 July 2012 – Accepted: 3 August 2012 – Published: 15 August 2012

Abstract. This paper presents the drifting snow climate of the Greenland ice sheet, using output from a high-resolution (~11 km) regional climate model. Because reliable direct observations of drifting snow do not exist, we evaluate the modeled near-surface climate instead, using automatic weather station (AWS) observations from the K-transect and find that RACMO2 realistically simulates near-surface wind speed and relative humidity, two variables that are important for drifting snow. Integrated over the ice sheet, drifting snow sublimation (SUds) equals 24 ± 3 Gt yr⁻¹, and is significantly larger than surface sublimation (SUₐ, 16 ± 2 Gt yr⁻¹). SUds strongly varies between seasons, and is only important in winter, when surface sublimation and runoff are small. A rapid transition exists between the winter season, when snowfall and SUds are important, and the summer season, when snowmelt is significant, which increases surface snow density and thereby limits drifting snow processes. Drifting snow erosion (ERds) is only important on a regional scale. In recent decades, following decreasing wind speed and rising near-surface temperatures, SUds exhibits a negative trend (0.1 ± 0.1 Gt yr⁻¹), which is compensated by an increase in SUₐ of similar magnitude.

1 Introduction

The Greenland ice sheet (GrIS) is the largest body of ice in the Northern Hemisphere, containing approximately 7 m sea level equivalent (IPCC AR4). In the last two decades, numerous Greenland outlet glaciers have accelerated and surface mass balance (SMB) has declined (Rignot et al., 2011), both contributing about equally to recent Greenland mass loss (Van den Broeke et al., 2009). The volume loss of outlet glaciers may be primarily related to oceanic warming (Holland et al., 2008), but the interaction between ocean and outlet glaciers is complex (Nick et al., 2009; Straneo et al., 2011). At the same time, Greenland has experienced significant atmospheric warming in the recent two decades (Box and Cohen, 2006), increasing surface meltwater production and subsequent runoff (Ettema et al., 2009), extending the length of the melt season (Fettweis et al., 2011) and triggering the melt-albedo feedback (Tedesco et al., 2010).

To assess the surface mass balance (SMB) of the GrIS, regional climate models are useful tools. In the SMB of the GrIS, precipitation (P) is the main source of mass, whereas mass is lost by surface (SUₐ) and drifting snow sublimation (SUds), drifting snow erosion (ERds), and meltwater runoff (RU). ERds is defined as the horizontal divergence of the snow transport (TRds). Until now, drifting snow processes have usually been neglected in GrIS SMB studies (Fettweis, 2007; Hanna et al., 2005; Ettema et al., 2009). On the Antarctic ice sheet, drifting snow sublimation (SUds) is an important ablation term in dry and windy areas and ERds redistributes snow on a local to regional scale (Lenaerts and van den Broeke, 2012). Several studies estimated SUds for Greenland (Déry and Yau, 2002; Box et al., 2006) and ERds (Déry and Yau, 2002), using parameterizations based on wind speed (Déry and Yau, 1999), neglecting feedbacks to the overlying atmosphere and the snow surface, which are known to be important (Bintanja, 2001; Lenaerts et al., 2010, 2012a; Gallée et al., 2001).

Published by Copernicus Publications on behalf of the European Geosciences Union.
Here we present the drifting snow climate (1960–2011) of the Greenland ice sheet using a regional atmospheric climate model (RACMO2) at relatively high horizontal resolution (11 km). RACMO2 includes an interactive drifting snow routine (Lenaerts et al., 2012a). We discuss the spatial and temporal variability of SU\textsubscript{ds} and ER\textsubscript{ds} and their impact on the SMB of the GrIS.

2 Methods

2.1 Numerical setup

The Regional Atmospheric Climate MOdel version 2 (RACMO2 hereafter; Van Meijgaard et al., 2008) combines the dynamical parameterizations from the HIgh Resolution Limited Area Model (HIRLAM) (Undén et al., 2002) with the physical schemes from the European Centre for Medium-Range Weather Forecasts model (ECMWF cycle 23r4, White, 2001). In recent years, RACMO2 has been used to estimate the SMB of Antarctica (Van de Berg et al., 2005; Lenaerts et al., 2012b) and Greenland (Ettema et al., 2009). Modeled precipitation and surface mass balance have been extensively evaluated using available in-situ observations (Ettema et al., 2009). Moreover, Ettema et al. (2010b) showed that RACMO2 provides a realistic simulation of the near-surface climate of the GrIS, although several deficiencies were detected, especially in the snow albedo scheme. To resolve this, we included a snow albedo parameterization based on the growth of snow during dry and wet metamorphosis, so that snow albedo can be physically coupled to snow grain size (Flanner and Zender, 2006). This significantly improved net shortwave radiation in RACMO2 over the Antarctic ice sheet (Kuipers Munneke et al., 2011) and the length of the melt season in Greenland (Van Angelen et al., 2012). In addition, a remote sensing-derived background albedo (from the Moderate Resolution Imaging Spectroradiometer, MODIS) is prescribed for ice (Van Angelen et al., 2012) to capture the spatial variability of albedo in the ablation area when the winter snow has melted (Van den Broeke et al., 2008). Finally, the drifting snow scheme derived from Déry and Yau (1999) has been included in RACMO2. It calculates drifting snow transport and sublimation, and accounts for interactions between the drifting snow layer with both the overlying atmosphere and the underlying snow surface (Lenaerts et al., 2010). Furthermore, we use an empirically derived parameterization for surface snow density. This was derived by Lenaerts et al. (2012a) for Antarctica, such that modeled drifting snow frequency and horizontal transport agree well with available in-situ and remote sensing observations (Lenaerts et al., 2012a).

For the GrIS, RACMO2 has 40 levels in the vertical and the model grid has a horizontal spacing of \(\sim 11\) km.

Fig. 1. Overview of model domain. Model topography is shown in colors; blue represents ocean. The white dotted boundary represents the model relaxation zone (24 grid points).

a 6-hourly resolution. The atmospheric fields from the reanalyses are relaxed towards true model resolution (11 km) in the relaxation zone (see Fig. 1). Sea surface temperature (SST) and sea ice extent are prescribed as a surface boundary condition by the reanalyses. Within the interior model domain, RACMO2 evolves freely.

2.2 Model evaluation

To date, no reliable direct observations of drifting snow processes on Greenland exist. In-situ instruments are prone to extreme weather conditions, and measurement errors increase with wind speed (Cierco et al., 2007). Remote-sensed observations of drifting snow, shown to be promising on Antarctica (Palm et al., 2011), are limited on Greenland due to enhanced cloudiness and generally shallower drifting snow layers (S. P. Palm, personal communication, 2011). Instead, we focus the evaluation of RACMO2 on the near-surface climate. Drifting snow predominantly occurs in the surface layer, in the first few meters above the surface. Near-surface parameters determine the occurrence and strength of drifting snow processes. Therefore, a key component of the model evaluation is to determine the model’s ability to simulate a realistic near-surface climate on the GrIS. Primarily near-surface wind speed drives drifting snow processes. Other important atmospheric parameters that control the amount of drifting snow sublimation are near-surface temperature \((T_2\) m) and near-surface specific humidity \((q_2\) m);
3 Results

3.1 Near-surface climate

Ettema et al. (2010a) thoroughly evaluated the near-surface climate of RACMO2, and found good agreement between model and observations. Their results are summarized in Table 1. A high correlation between modeled and observed near-surface temperature ($T_2$ m) is found. The negative bias, which is mainly confined to the winter season, is attributed to an underestimation of downward longwave radiation (Ettema et al., 2010a). Considering that, especially for coastal stations, 10 m wind speed is greatly influenced by local topography that may not always be properly represented by RACMO2, the agreement between model and observations is good. High wind speeds that drive drifting snow are also well represented by RACMO2 and no clear underestimation is seen, so the model resolution appears sufficient to represent local wind climates on the ice sheet (Ettema et al., 2010a).

To further evaluate the modeled wind climate, Fig. 2 shows the frequency distributions of daily mean wind speeds from RACMO2 and AWS observations along the K-transect in southwestern Greenland (Van de Wal et al., 2005). RACMO2 realistically simulates the observed wind speed distributions at S5, S6 and S9. For drifting snow cases, especially the tail of the frequency distributions is important, which includes the very high wind speed events. The tails of the wind distributions from RACMO2 agree well with the observations at S6 and S9. RACMO2 somewhat underestimates the frequency of high wind speeds at S5, which is located on steep
slopes in rough topography that are not fully resolved by the model.

Daily mean near-surface specific humidity at S6 (Table 1) is well represented in RACMO2. At S9, however, where drifting snow occurs frequently, relative humidity is found to be underestimated in RACMO2 simulations without the drifting snow routine (Fig. 3). Including drifting snow physics strongly reduces the negative bias (Fig. 3), especially during drifting snow sublimation events. The mean bias during the period shown in Fig. 3 decreases from 13 % to 7 %, and the RMSE reduces from 15 % to 9 %. The reason for this improvement is that water vapor is added to the surface layer in response to SU_{ds}, increasing relative humidity in the surface layer. We conclude that it is currently hard to evaluate drifting snow and its effect on the near-surface atmosphere accurately.

However, indirect evidence suggests that the occurrence of drifting snow and its effect on the near-surface atmosphere are captured well. Figure 4a–f present the mean fields of the (near-surface) climate parameters that are important for drifting snow (sublimation). The surface mass balance (SMB, Fig. 4a) is negative (< −500 mm w.e.) along major portions of the margins, indicating that, on an annual basis, surface runoff from snow- and icemelt exceeds the input of precipitation. This results in the exposure of bare glacier ice during a part of the year. On the other hand, the southeastern marginal zones are characterized by strong precipitation, and have a positive annual SMB of > 1 m w.e. The inland regions of the ice sheet are drier, but also exhibit less or no snowmelt; hence SMB > 0. The occurrence of surface melt is also reflected in the surface snow density (ρ_s, Fig. 4b). Melt strongly increases the densification rate of the snowpack (Wakahama, 1975). This, in turn, limits the onset of drifting snow. Lenaerts et al. (2012a) show that, for a surface snow density increasing from 300 to 500 kg m^{-3}, the associated threshold friction velocity (u^*_t) increases from 0.4 to 1.2 m s^{-1}, the latter which, for a typical surface roughness z_0 of 1 mm, represents a 10 m wind speed of > 30 m s^{-1}.

Measurements of surface snow density on the GrIS only exist for the summer season, when ρ_s is high, limiting drifting snow. An exception is the dry snow zone on the higher portions of the ice sheet, where melt does not occur, not even in summer. For these regions, we used ρ_s observations on the Antarctic ice sheet to introduce temporal and spatial variability of the surface snow density (ranging from 300 to 400 kg m^{-3}), depending on the local wind and accumulation climate (Lenaerts et al., 2012a). Based on the sparse observations available, this parameterization appears to be valid on Greenland as well. For example, the observations of Bolzan and Strobel (1994) suggest a mean summer surface snow...
density of 332 kg m$^{-3}$ (Zwally and Li, 2002). RACMO2 agrees very well with these observations, and gives a mean $\rho_s$ in July of 330 kg m$^{-3}$ in this region (Fig. 4b). Another way to assess model performance is to investigate how well the model captures the transition from winter (no melt) and summer (melt) season, as melt-induced densification limits the occurrence of drifting snow. Fettweis et al. (2011) and Van Angelen et al. (2012) show that RACMO2 is able to well simulate the onset, duration and extent of melt on the Greenland ice sheet. Although indirect evidence, this indicates that the timing and phase of the drifting snow season is well captured by the model.

Figure 4c shows the mean friction velocity, the parameter that combines the near-surface wind speed and surface roughness (Schmidt, 1981), (Li and Pomeroy, 1997). When the friction velocity exceeds the threshold friction velocity ($u_{\theta_i}$), drifting snow is initiated. The highest average friction velocities are found in the northeastern part of the ice sheet, where katabatic forcing is largest (Van Angelen et al., 2011), along the western coast, where katabatic and synoptic forcing lead to a persistent southerly jet (Van Angelen et al., 2011), and in southern Greenland, which is under the frequent influence of midlatitude cyclones. The near-surface temperature and relative humidity are shown in Fig. 4d and e, respectively. Evidently, near-surface temperature is largely determined by elevation. The highest temperatures are found along the coastline, where above-zero temperatures occur regularly in summer. Due to the low temperatures, RH$_t$ is very high (> 95 %) on higher portions of the ice sheet (> 2000 m), limiting the potential for SU$_{ds}$, whereas the near-surface air is less saturated at lower altitudes, where katabatic (downslope) winds give rise to adiabatic warming and subsequent drying of the near-surface air.

### 3.2 Drifting snow climate

#### 3.2.1 Spatial variability

Figure 5a–d present the mean annual drifting snow frequency, transport, SU$_{ds}$, and ER$_{ds}$. Drifting snow occurs on the entire ice sheet (Fig. 5a), with frequencies ranging from < 20 % in highly elevated regions, where calm conditions prevail, up to 50 % and more in coastal areas. Highest drifting snow frequencies are found along the northeastern, southeastern and western margins, where the surface snow density is relatively low (Fig. 4b) and mean wind speeds are highest (Fig. 4c). Horizontal transport of drifting snow (shown as a scalar in Fig. 5a) varies from less than $0.5 \times 10^6$ kg m$^{-3}$ yr$^{-1}$ on the ice sheet plateau to $> 2 \times 10^6$ kg m$^{-3}$ yr$^{-1}$ in the regions where drifting snow is most active. SU$_{ds}$ varies from < 5 mm yr$^{-1}$ in regions above 2500 m to 20–60 mm yr$^{-1}$ in some low-lying regions. In contrast, SU$_{ds}$ is small in the surrounding tundra areas, where the presence of snow on the ground is limited to the winter season, and, more importantly, near-surface wind speed is lower than on the ice sheet (Fig. 2a, Ettema et al., 2010b). ER$_{ds}$ is related to divergence/convergence of the near-surface (katabatic) winds and the resulting TR$_{ds}$. From the interior ice sheet to the margins, katabatic winds accelerate

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**Fig. 5.** Mean drifting snow frequency (A), transport (B), sublimation (C) and erosion (D). Drifting snow frequency is defined as the fraction of days with accumulated TR$_{ds}$ of 3 kg m$^{-3}$. Dashed-dotted lines represent the topography with 500 m intervals.

**Fig. 6.** Mean (1960–2011) SU$_{ds}$ and TR$_{ds}$ on all ice sheet grid points (~ 14 000 points) as a function of mean 10 m wind speed (red), 2 m relative humidity with respect to ice (blue) and surface snow density (green).
leading to drifting snow erosion (10–40 mm yr\(^{-1}\); see Fig. 5b), whereas net snow deposition occurs closer to the coast, where near-surface winds generally slow down. Patterns of alternating erosion/deposition occur where regional topography is strongly concave/convex, for example in the northwest (82° N, 50–60° W). The nonlinear nature of drifting snow is evident from Fig. 6. Both SU\(_{ds}\) and TR\(_{ds}\) strongly depend on wind speed, but SU\(_{ds}\) is limited when the surface layer becomes saturated. Drifting snow sublimation is therefore most active in relatively warm, windy and dry (in terms of relative humidity) regions.

Table 2 shows the ice-sheet integrated values of the SMB and its components. The 1960–2011 mean SMB of the Greenland ice sheet is estimated at 376 ± 118 Gt yr\(^{-1}\). This estimate is comparable to other recent GrIS SMB estimates that are assembled by Rae et al. (2012), which are mostly in the range 320–450 Gt yr\(^{-1}\). Surface mass balance is largely determined by the mass input from precipitation (732 ± 74 Gt yr\(^{-1}\)) and the mass loss via runoff (306 ± 86 Gt yr\(^{-1}\)). The mass loss through sublimation is of secondary importance on Greenland. The total sublimation (SU\(_{ds}\) + SU\(_s\)) equals 40 ± 4 Gt yr\(^{-1}\). SU\(_{ds}\) is responsible for 60 % of the total sublimation (24 ± 3 Gt yr\(^{-1}\)), and 16 ± 2 Gt yr\(^{-1}\) is lost via surface sublimation. Compared to a simulation without drifting snow (Ettema et al., 2009), total sublimation has increased by more than 40 %, although SU\(_s\) is 40 % lower; hence, including SU\(_{ds}\) reduces SU\(_s\), but increases total sublimation. Integrated over the ice sheet, SU\(_{ds}\) is clearly smaller than runoff, and contributes 7 % to the total ablation on Greenland. In comparison, SU\(_{ds}\) is estimated to be responsible for ~75 % of the ablation in Antarctica (Lenaerts et al., 2012b), which highlights the large difference in near-surface climate between the two ice sheets. The influence of ER\(_{ds}\) on the Greenland surface mass balance is small (1 ± 1 Gt yr\(^{-1}\)). Despite their relatively small effect on ice sheet integrated SMB, drifting snow processes (SU\(_{ds}\) and ER\(_{ds}\)) attribute largely to the local surface mass balance in relatively dry, windy regions of the Greenland ice sheet. Locally, they remove up to 50 % of the precipitated snow (not shown).

### 3.2.2 Seasonal variability

All SMB components, apart from snowfall, show a clear seasonal cycle (Fig. 7). Snowfall is on average largest in early winter (September–November). Runoff and SU\(_s\) peak in summer, whereas SU\(_{ds}\) is most significant in winter. It is larger than runoff during 7 months and larger than SU\(_s\) during 8 months of the year. In the winter months (December–February), SU\(_{ds}\) removes ~6 % of the snowfall.

Figure 8 shows the mean number of drifting snow days per year. Drifting snow occurs on 120–180 days in the western katabatic wind zone, and locally on >200 days in the north-east. The latter region not only experiences the highest katabatic wind speeds (Van Angelen et al., 2011), but also has a relatively short melt season (see Fig. 8), keeping \(\rho_s\) and \(u_{s,t}\) relatively low.

An example of the transition from drifting snow season to melt season is shown in Fig. 9. It shows the annual cycle (year 2005) of drifting snow-related variables at S6, located in the ablation zone of the western ice sheet (Van den Broeke et al., 2008). In winter (January–May and October–December), the friction velocity (\(u_t\)) often exceeds the threshold (\(u_{s,t}\), which depends on the density of the top snow layer, Gallée et al., 2001). The top snow layer is frequently refreshed by lower-density fresh snow, which keeps \(u_{s,t}\) low (0.3–0.4 m s\(^{-1}\)). In summer, but also during some
...and it includes a snow albedo scheme that tracks the...white dashed areas) coincide with regions...Van den Broeke et al. (2012), only...V an den Broeke Lenaerts and van den...Fig. 9. Modeled annual cycle (2005) at S6 (67°5′N, 49°23′W)...the drifting snow season (see text)....of daily mean friction velocity (blue) and its threshold (black, see...This strongly enhances ρs, preventing drifting snow to occur. This illustrates the strong feedback between the length of the melt season, snow surface conditions and length of the drifting snow season in Greenland.

3.2.3 Interannual variability and trends

Interannual variability in SUds is clearly related to interannual variability in near-surface wind speed (correlation coefficient r = 0.91). Years with high SUds (for example 1996) are characterized by anomalously high near-surface wind speed and, to a lesser extent, temperature. Interestingly, the areas with the largest inter-annual variability of drifting snow frequency (Fig. 8, white dashed areas) coincide with regions where year-to-year variations in the extent of the ablation season are also largest.

RACMO2 suggests that, in the period 1960–1990, near-surface temperature (−0.03 ± 0.02 K yr⁻¹) and wind speed (0 ± 0.003 m s⁻¹ yr⁻¹) trends were insignificant. Since 1990, however, following a rise in atmospheric temperatures, meltwater runoff has been increasing (Ettema et al., 2009), only partly compensated by increased snowfall (Van den Broeke et al., 2009), which led to a strong decline in SMB. Figure 9 shows the recent trend of SUds and SUa. Since 1990, SUds has decreased at a rate of −0.1 ± 0.1 Gt yr⁻², following a 10 m wind speed decrease of −0.009 ± 0.005 m s⁻¹ yr⁻¹ (Fig. 10) and the increased length of the summer season over Greenland (0.6 ± 0.9 days yr⁻²). The large uncertainty in the trend of SUds is due to strong interannual variability, and a longer time series is required to reach significance at a 95% level. The decline in SUds is compensated by a significant increase in SUa (+0.09±0.09 Gt yr⁻²), which is related to the less frequent occurrence of drifting snow in combination with the higher near-surface temperature (+0.09 ± 0.03 K yr⁻¹).

4 Conclusions

This paper describes the first attempt to simulate the drifting snow climate of the Greenland ice sheet, using output of a high-resolution (11 km) regional climate model (RACMO2) with interactive drifting snow routine for a period of 52 yr (1960–2011). RACMO2 is able to realistically model the near-surface climate of Greenland (Ettema et al., 2010a), and it includes a snow albedo scheme that tracks the evolution of snow grain size and a drifting snow routine that allows for interaction between drifting snow, the snow surface and the overlying atmosphere (Lenaerts and van den Broeke, 2012).
We evaluated the results in terms of the near-surface climate in further detail, because a direct evaluation of the drifting snow climate is not feasible. RACMO2 shows realistic wind speed spectra along the K-transect. Furthermore, we showed that, during drifting snow, SU_{ds} enhances near-surface relative humidity, which improves the agreement with observations. Unfortunately, because no reliable direct observations of drifting snow transport or frequency are available for Greenland, we are presently unable to explicitly/directly evaluate and/or further improve the drifting snow model.

The results show that drifting snow in Greenland is not significant during summer, when near-surface wind speeds are lower and surface snow density and associated threshold friction velocity increase due to melt, which suppresses drifting snow to occur. In winter, during 7 months of the year, SU_{ds} is the only important ablation mechanism, removing ∼6% of the precipitation during this period. In summer, SU_{ds} is negligible, and meltwater runoff becomes the primary ablation component. Integrated over the Greenland ice sheet, SU_{ds} equals 24 Gt yr^{-1}, 50% larger than SU_s (16 Gt yr^{-1}). Although the contribution of drifting snow erosion (ER_{ds}) to the surface mass balance is negligible on the ice sheet, it is locally of importance, especially in areas where topographic features induce strong wind field divergence/convergence. An abrupt transition occurs between the melting season in summer, when drifting snow is strongly suppressed, and the drifting snow season, when light, fresh snow is prone to drifting.

Acknowledgements. We thank the editor, R. Dadic and an anonymous reviewer for their helpful comments. This work was supported by funding from the ice2sea programme from the European Union 7th Framework Programme, grant number 226375. Ice2sea contribution number 092.

Edited by: V. Radic

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