Modeling drifting snow in Antarctica with a regional climate model: 2. Results

J. T. M. Lenaerts[1] and M. R. van den Broeke[1]

Received 1 December 2010; revised 12 January 2012; accepted 13 January 2012; published 6 March 2012.

[1] This paper presents a model study of the impact of drifting snow on the lower atmosphere, surface snow characteristics, and surface mass balance of Antarctica. We use the regional atmospheric climate model RACMO2.1/ANT with a high horizontal resolution (27 km), equipped with a drifting snow routine and forced by ERA-Interim (1989–2009) at its lateral and ocean boundaries. Drifting snow sublimation (SU_{ds}) is significant in Antarctica, especially in the coastal regions (>150 mm water equivalent yr^{-1}). Integrated over the ice sheet, SU_{ds} removes ∼6% of the annually precipitated snow. Drifting snow interacts with the atmosphere, as it increases the lower atmospheric moisture content and reduces surface sublimation (SU_s), and leads to increased snowfall in regions where the atmosphere usually is close to saturation. Drifting snow sublimation (SU_{ds}) is smallest in summer, when katabatic wind speeds are lower and melting and surface sublimation consolidate the snow surface. Compared to a simulation without drifting snow, total sublimation (SU_{ds} + SU_s) doubles on the grounded ice sheet if drifting snow is considered. Drifting snow erosion is locally significant, but can be neglected on a continent-wide scale.

Citation: Lenaerts, J. T. M., and M. R. van den Broeke (2012), Modeling drifting snow in Antarctica with a regional climate model: 2. Results, *J. Geophys. Res.*, *117*, D05109, doi:10.1029/2010JD015419.

1. Introduction

[2] The Antarctic ice sheet (AIS) is the largest ice volume on Earth, representing an equivalent eustatic sea level rise of 61 m [*Church et al.*, 2001]. Recent studies found that the AIS is losing mass at a rate of 30–200 Gt yr^{-1}, with uncertainties in the order of 40% [*Rignot et al.*, 2011; *Zwally and Giovinetto*, 2011]. A major reason for this large uncertainty is insufficient knowledge of the surface mass balance (SMB), including the impact of drifting snow [*Bintanja*, 1998; *Bintanja and Reijmer*, 2001; *Van de Berg et al.*, 2005; *Box et al.*, 2006; *Van den Broeke et al.*, 2010]. We define drifting snow here as the combined processes of drifting snow particles (limited to below 2 m above the surface) and blowing snow (above 2 m). The drifting snow layer is usually shallow, but on some occasions can extend hundreds of meters into the atmosphere [*Budd et al.*, 1966; *Mahesh et al.*, 2003]. For the East Antarctic ice sheet, which represents more than 90% of the AIS and which exhibits low accumulation, strong katabatic winds and negligible snow and ice melt, the impact of drifting snow on the SMB may be of particular importance.

[3] *Lenaerts et al.* [2010] found that drifting snow sublimation (SU_{ds}) removes 16 ± 8 % of the annual snowfall at Neumayer station (Dronning Maud Land, East Antarctica), representing a mass flux of around 80 mm yr^{-1}. It has been suggested that SU_{ds} can remove up to 75% of annual snowfall in dry and windy areas, such as the East Antarctic ice sheet [*Frezzotti et al.*, 2005]. Drifting snow erosion, although believed to be two or three orders of magnitude smaller than SU_{ds} on a continental scale, can become important locally, especially in dry and windy areas, such as Victoria Land [*Frezzotti et al.*, 2007; *Scarchilli et al.*, 2010].

[4] Not only does drifting snow impact the SMB, it also interacts with the physical properties of the snow surface and the lower atmosphere, with several negative feedbacks being active. Drifting snow sublimation cools and moistens the near-surface air, limiting further sublimation [*Mann et al.*, 2000]. In katabatic wind regions along the ice sheet margin, vertical entrainment of dry air in the boundary layer may enhance and maintain drifting snow sublimation [*Bintanja*, 2001b, 2001a], which would otherwise be strongly self-limiting. The characteristics of the surface snow to a large extent determine the onset of drifting snow [*Gallée et al.*, 2001], whereas drifting snow erosion and sublimation in turn lead to changes in snow surface conditions: if older, denser snow layers are exposed, this will decrease the drifting snow potential.

[5] In this study we use the regional climate model RACMO2.1/ANT at a horizontal resolution of 27 km, to model the impact of drifting snow on the SMB, surface characteristics and lower atmosphere of Antarctica. Part 1 of this paper [*Lenaerts et al.*, 2012a] describes the methodology and evaluates the near-surface (drifting snow) climate in

[1]Institute for Marine and Atmospheric Research Utrecht, Utrecht University, Utrecht, Netherlands.

0148-0227/12/2010JD015419

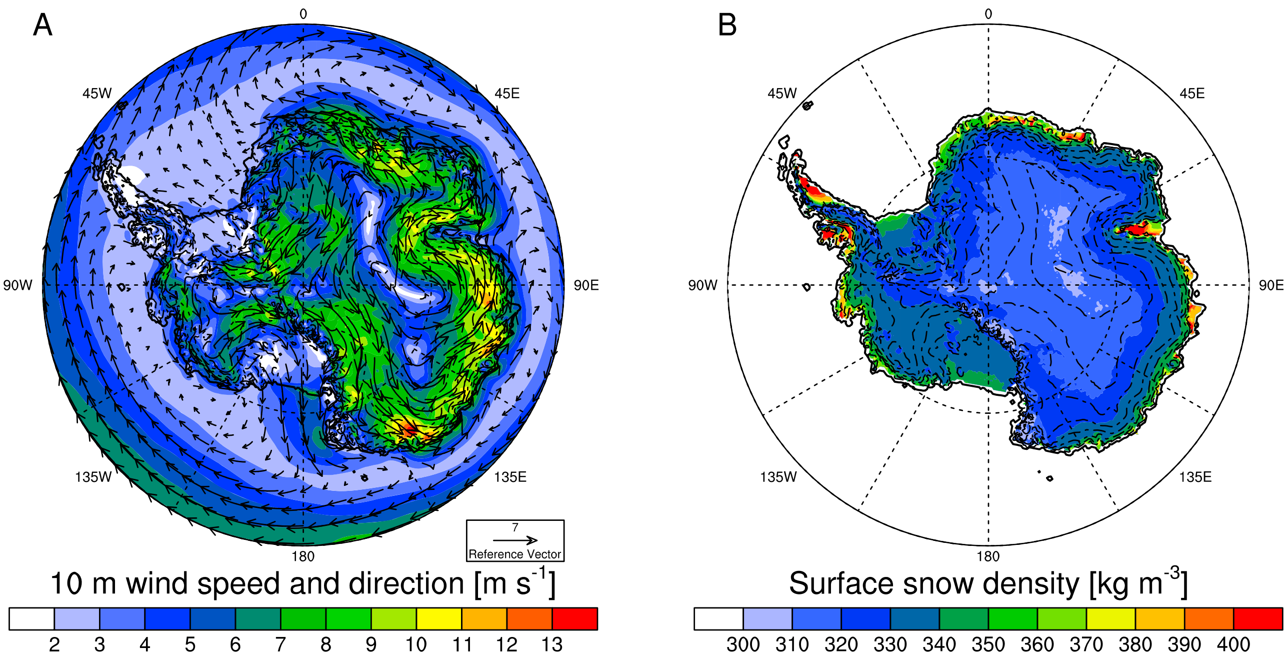

Figure 1. (a) Simulated mean (1989–2009) 10 m wind speed (m s^{-1}) and direction. (b) Surface snow density (upper 5 cm) (kg m^{-3}).

RACMO2.1/ANT. Part 2 (the present paper) discusses the drifting snow climate (section 2) and analyzes the interactions with the atmosphere and surface (sections 3 and 4) and the impact of drifting snow on the SMB (section 5). Conclusions and discussion are given in section 6.

2. Drifting Snow Climate

[6] To analyze the drifting snow climate of Antarctica, we use a regional atmospheric climate model (RACMO2.1/ANT) with a horizontal resolution of ~27 km, which is driven by ERA-Interim fields (1989–2009) at its lateral boundaries [*Lenaerts et al.*, 2012a]. RACMO2.1/ANT is coupled to a drifting snow routine based on that of *Déry and Yau* [1999], such that the interaction between the atmosphere and drifting snow (drifting snow sublimation is assumed to be the only source of latent heat during drifting snow [*Lenaerts et al.*, 2010]) and the surface and drifting snow (following *Gallée et al.* [2001]) are considered. *Lenaerts et al.* [2012a] show that the model is well able to simulate realistic near-surface conditions and spatial distribution of ablation. In addition, they develop an empirical relationship for fresh snow density in order to realistically simulate observed drifting snow frequencies in Antarctica [*Lenaerts et al.*, 2012a]. To analyze the impact of drifting snow on the SMB, surface characteristics and atmosphere over Antarctica, we performed two simulations, one with drifting snow activated (DRIFT) and one with drifting snow switched off (NODRIFT). Here, we show results of those model simulations for the period 1989–2009.

2.1. Wind and Surface Snow Density

[7] The near-surface wind speed and state of the snow surface (mainly density) determine the likelihood of snow particles to be lifted by the wind. The mean 10 m wind speed and direction as modeled by RACMO2.1/ANT is presented in Figure 1a. Largest wind speeds are found in coastal regions of East Antarctica, such as Adélie Land, with annual mean wind speeds >10 m s^{-1} [*Turner et al.*, 2009]. In the interior ice sheet, owing to its katabatic nature, the wind is persistently directed from the interior toward the coast, deflected to the left by the Coriolis effect [*Parish and Bromwich*, 2007]. In the escarpment region, synoptic and katabatic forcing are equally large, and over the flat ice shelves synoptic forcing is the dominant cause of drifting snow events. The fresh snow density (Figure 1b) is empirically derived to match modeled drifting snow frequencies with observations [*Lenaerts et al.*, 2012a]. The annual mean value of resulting surface snow density (Figure 1b) varies from ~300 kg m^{-3} on the cold East Antarctic Plateau to ~400 kg m^{-3} on some ice shelves where summer melting occurs.

2.2. Drifting Snow Variables

[8] Figure 2 depicts the modeled annual mean horizontal transport of snow (Figure 2a), drifting snow frequency (Figure 2b), drifting snow sublimation (Figure 2c) and drifting snow erosion (Figure 2d) for the period 1989–2009. On the high Antarctic Plateau, modeled snow transport equals 0.1 Mt m^{-1} yr^{-1} or less. In the escarpment and coastal areas, snow transport generally exceeds 1 Mt m^{-1} yr^{-1} and reaches 10 Mt m^{-1} yr^{-1} in Adélie Land. In these areas, modeled drifting snow frequency exceeds 80% (Figure 2b). Generally, modeled drifting snow frequency in the coastal areas varies between 30 and 80% and peaks in areas with strong katabatic winds. Figure 3 illustrates that TR_{ds} and SU_{ds} are correlated in a non-linear fashion to 10 m wind speed, and peak in regions with moderate near-surface temperatures (240–250 K) and relatively low 2 m relative humidity (~0.6–0.7). These regions represent the escarpment

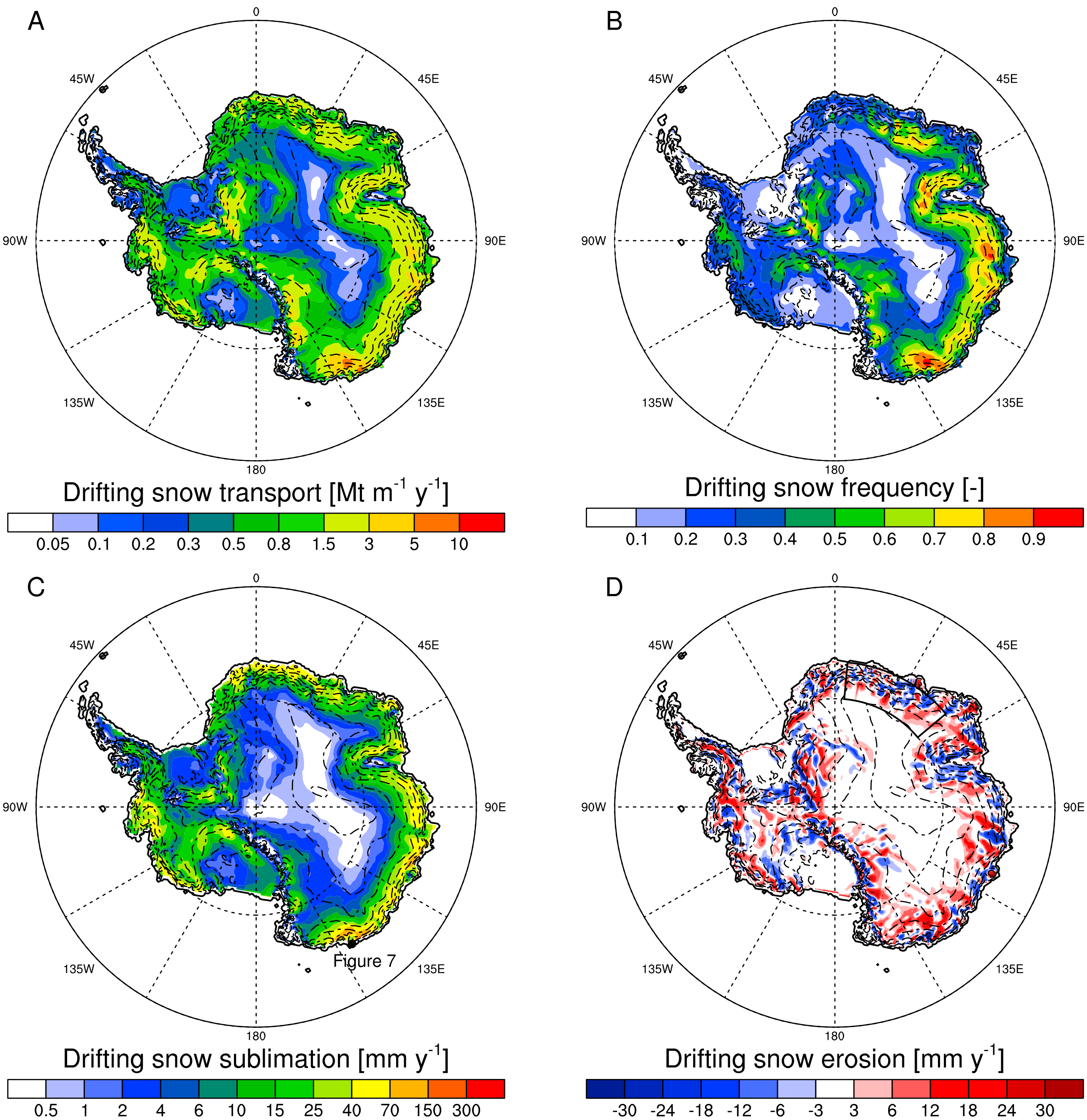

Figure 2. Simulated mean annual (1989–2009) (a) drifting snow horizontal transport (M tm^{-1} yr^{-1}), (b) drifting snow frequency, defined as the fraction of days with drifting snow in the model (TR_{ds} > 300 kg m^{-1} d^{-1}), (c) drifting snow sublimation (mm water equivalent^{-1}), and (d) drifting snow erosion (mm water equivalent^{-1}). In Figure 2d, negative numbers indicate net accumulation. The region shown in Figure 4 is indicated by the black box in Figure 2d. Dashed contours represent topography (500 m resolution).

region of East Antarctica, where katabatic winds are strong (>10 m s^{-1}), generating drifting snow and entraining relatively warm and dry air into the atmospheric boundary layer (ABL). At the same time, accumulation in these regions is significant and there is no melting, keeping surface snow density ρ_s and the corresponding drifting snow threshold friction velocity ($u_{*,t}$) low. As the point clouds in Figure 3 demonstrate, no simple relations between any of the parameters exist.

[9] Drifting snow sublimation (Figure 2c) is smaller than 1 mm yr^{-1} in large parts of the interior and increases to 50 mm yr^{-1} or more toward the coast. Largest SU_{ds} (>150 mm yr^{-1}) values are found in windy coastal regions, such as Victoria Land, Law Dome and Dronning Maud Land. On the Ross and Filchner-Ronne ice shelves, low wind speeds generally lead to small values of SU_{ds}. An exception is the southwestern part of the Ross ice shelf, where the Ross

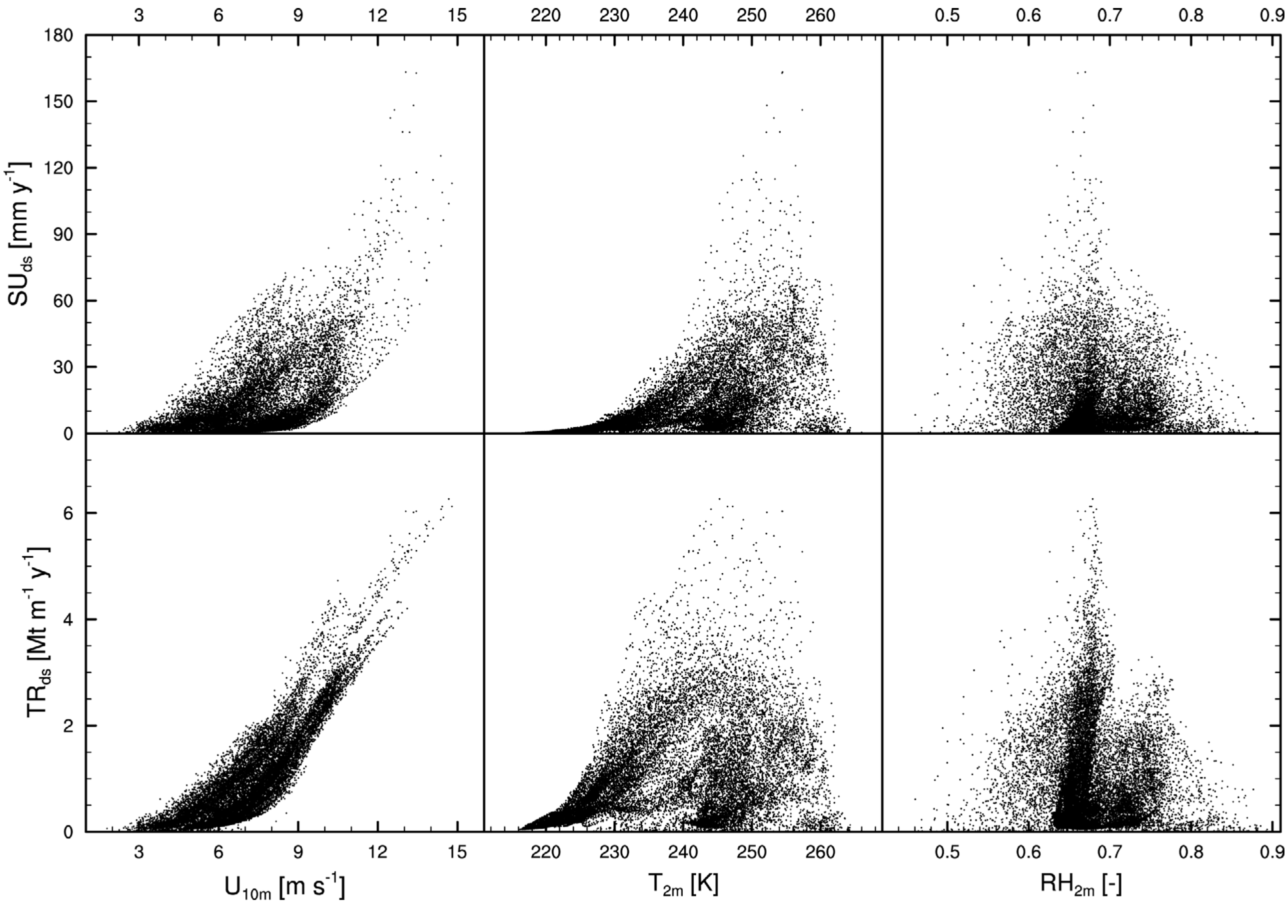

Figure 3. (top) Mean drifting snow sublimation and (bottom) horizontal transport as a function of (left) mean 10 m wind speed, (middle) 2 m temperature, and (right) 2 m relative humidity (with respect to ice) for the period 1989–2009 at all ice sheet grid points (N ~ 16000).

ice shelf airstream [*Steinhoff et al.*, 2009] blows parallel to the Transantarctic Mountains. There we find SU_{ds} values up to 40 mm yr^{-1}.

[10] Modeled drifting snow erosion (Figure 2d) is only locally important. It is negligible on the Antarctic Plateau, but its magnitude increases to 30 mm yr^{-1} or more at locations where winds are strong and the wind field is strongly divergent/convergent. In these regions, areas with strong drifting snow erosion ($ER_{ds} > 0$) are found next to locations with strong drifting snow deposition ($ER_{ds} < 0$), which indicates that ER_{ds} tends to reduce or enhance local SMB gradients. Although our results show more spatial detail, the spatial patterns of drifting snow erosion and their magnitude are very similar to those presented by *Bromwich et al.* [2004]. Owing to the relation between ER_{ds} and the divergence of the near-surface wind, ER_{ds} is strongly coupled to topographic features that follows from the terrain-following nature of the katabatic winds [*Van den Broeke et al.*, 2002]. Figure 4 zooms in on Dronning Maud Land between 10 and 45°E. Around 25°E, a topographic ridge induces divergence

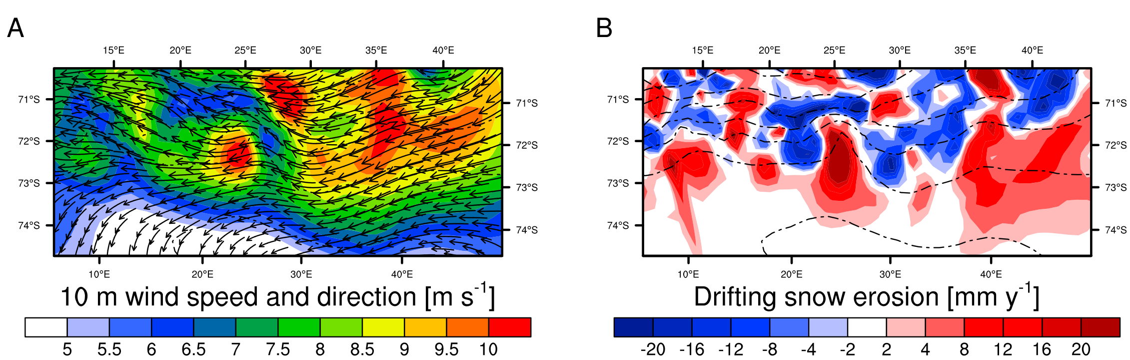

Figure 4. (a) Detail of mean 10 m wind speed (colors) and vector and (b) mean drifting snow erosion in Eastern Dronning Maud Land (70°-75°S, 5°E-50°E; see Figure 2d).

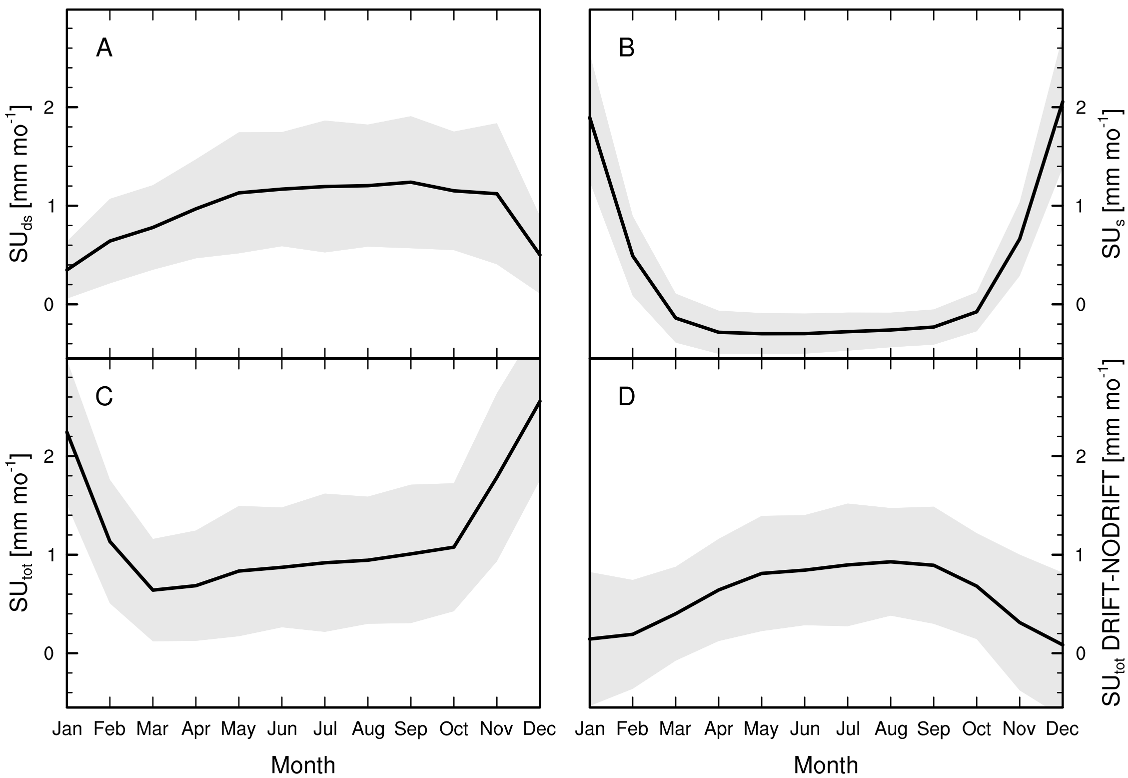

Figure 5. Monthly mean (1989–2009) ice sheet-average (a) drifting snow sublimation, (b) surface sublimation, and (c) total sublimation ($SU_{tot} = SU_{ds} + SU_s$, C) in DRIFT . (d) The difference between SU_{tot} in DRIFT and NODRIFT (SU_s only) (all in mm water equivalent^{-1}). The shaded area represents twice the standard deviation from the mean.

of the katabatic flow direction and velocity (Figure 4a). Behind the ridge, the wind speed decreases. This leads to net drifting snow erosion (~20 mm yr^{-1}) at the ridge and drifting snow deposition (~−10 mm yr^{-1}) behind the ridge (Figure 4b). Similar phenomena are seen elsewhere on the ice sheet. We expect these features be become even much more pronounced at higher model resolutions.

2.3. Seasonal Cycle

[11] Figure 5a shows the mean seasonal cycle of SU_{ds}, averaged over the AIS. Drifting snow sublimation peaks in late winter, with values ranging from 0.35 mm mo^{-1} in January to 1.2 mm mo^{-1} in September. During late summer and fall, SU_{ds} increases (January to May) and remains relatively constant throughout the period May to November. The interannual variability of monthly SU_{ds} is higher in winter than in summer, due to enhanced variability in wind speed and temperature in winter. For individual months, the year-to-year variability (in terms of standard deviation) is around 30%. However, the variability of annual mean SU_{ds} is much smaller (6%, not shown). Apparently, the variability of SU_{ds} on monthly timescales is substantial, but is greatly reduced when longer timescales are considered.

3. Interactions With the Lower Atmosphere

[12] To analyze the impact of drifting snow on the SMB, surface characteristics and atmosphere over Antarctica, we performed two simulations, one with drifting snow activated (DRIFT) and one with drifting snow switched off (NODRIFT). Figures 5b and 5c show surface (SU_s) and total sublimation ($SU_{tot} = SU_{ds} + SU_s$) in DRIFT. In contrast to SU_{ds}, SU_s peaks in summer and becomes slightly negative in winter, signifying net deposition (riming). SU_s exceeds SU_{ds} only in December and January (~2 mm) and has comparable magnitude in February and November (~1 mm). Figure 5d indicates that total sublimation (SU_{tot}) increases in DRIFT compared to NODRIFT, throughout the year but especially in winter, when SU_{ds} is large; however, the difference is smaller than SU_{ds}, because SU_s is reduced in DRIFT. Previous studies confirmed this effect [*Mann et al.*, 2000; *Bintanja*, 2001b; *Lenaerts et al.*, 2010], which is caused by the reduction of the vertical moisture gradient just above the surface when drifting snow occurs. This limits SU_s, but within that saturated layer also SU_{ds}, which thereby limits its own strength [*Mann et al.*, 2000; *Lenaerts et al.*, 2010]. Net water vapor deposition (negative SU_s) occurs in winter in the DRIFT simulation, while mean SU_s in NODRIFT is positive. An explanation for this is that in winter in the NODRIFT simulation, SU_s is mainly significant during windy conditions, making mean SU_s positive. In DRIFT, SU_{ds} takes over under windy conditions and SU_s is only non-zero during calm conditions, when deposition is promoted by the strong surface-based temperature inversion, making SU_s negative.

[13] To assess the impact of drifting snow on the atmospheric surface layer, Figure 6 shows the mean seasonal cycle of major surface layer variables for the NODRIFT

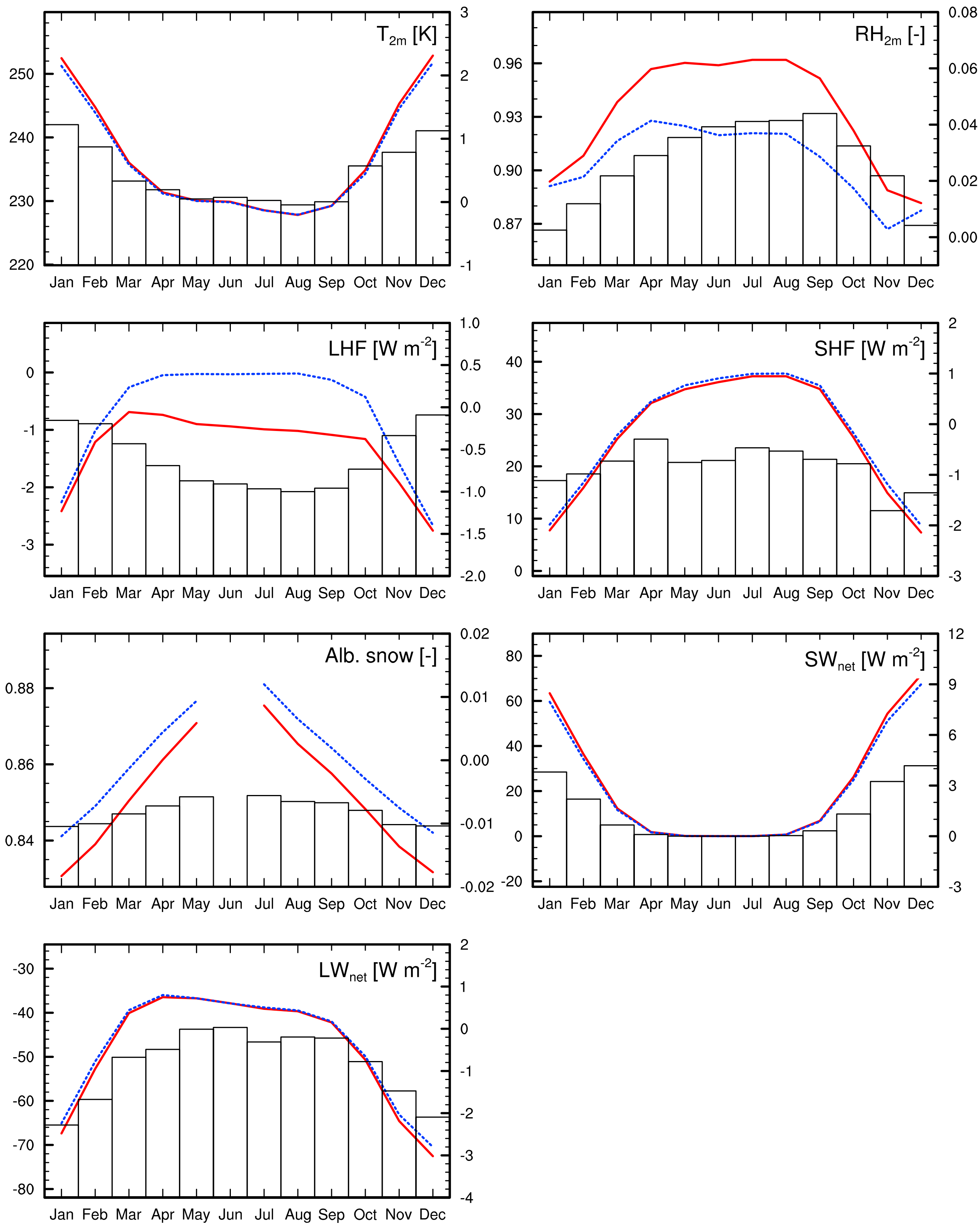

Figure 6. Mean annual cycle (1989–2009) of 2 m temperature (T_{2m}), 2 m relative humidity (RH_{2m}), latent heat flux (LHF), sensible heat flux (SHF), snow albedo, net shortwave radiation (SW_{net}), and net longwave radiation (LW_{net}), the latter five evaluated at the surface. Fluxes are defined as positive when directed toward the surface. DRIFT is shown in red (solid curve) and NODRIFT in blue (dotted curve). The differences between DRIFT and NODRIFT are shown in bars with scale on the right axis. The snow albedo is not plotted in June, because the largest part of Antarctica receives no direct sunlight during that month.

(blue) and DRIFT (red) simulations, averaged over the ice sheet. In both simulations, a pronounced annual cycle is seen in all variables. The 2 m temperature varies from 250 K in summer to 230 K in winter. Relative humidity (with respect to ice) is higher in winter, which reflects near-surface diabatic cooling. The latent heat flux is smaller than the sensible heat flux, and negative (directed away from the surface) throughout the year, while the sensible heat flux is directed toward the surface, mainly to compensate the negative LW_{net} [*Van den Broeke*, 2005]. Snow albedo is high in spring and autumn owing to a large zenith angle, but does not strongly influence the energy budget during these seasons due to the small flux of solar radiation. Typical for Antarctica is the strongly negative longwave radiation balance, due to the cold and dry atmosphere and low cloud cover. The minimum value of LW_{net} occurs in summer, when the surface is heated by SW_{net}, enhancing longwave emission.

[14] The difference between DRIFT and NODRIFT (bars) is also characterized by a seasonal cycle. Sublimation is enhanced, especially in winter, when drifting snow sublimation peaks. This leads to an increase in RH_{2m} of 1–5%. Another interesting result is the marked decrease in snow albedo in DRIFT, signifying a significant interaction of drifting snow with the surface (see section 4). This is not relevant in winter due to the polar night, but in summer this leads to increased net shortwave radiation and 2 m temperature (up to 1 K in December and January). Secondary effects include the decrease of sensible heat flux and the enhanced energy loss by longwave radiation.

[15] To analyze the interactions between drifting snow and the atmosphere in more detail, we focus on a location where drifting snow is strong (Adélie Land, 67°S, 145°E, 297 m a.s.l., see Figure 2). The annual mean snowfall is $\sim$600 mm yr^{-1} at this location, whereas drifting snow sublimation is $\sim$125 mm yr^{-1}. Surface sublimation is $\sim$50 mm yr^{-1}, which indicates that both SU_{ds} and SU_s are significant for the SMB at this location and that total sublimation removes almost one third of the snowfall. Figure 7 presents the impact of drifting snow on the near-surface atmospheric variables. Most distinctly, the latent heat flux in DRIFT shows large negative peaks in winter, which are related to drifting snow events. The largest events lead to a daily mean LHF of ~ -60 W m^{-2} ($\sim$2 mm water equivalent). Relative humidity increases in winter due to the enhanced sublimation in DRIFT. Some peaks in relative humidity co-exist in DRIFT and NODRIFT; these are related to low-pressure system passages, advecting warm and moist air inland, raising temperature and humidity throughout the troposphere. The 2 m temperature difference is mainly driven by the feedback between the surface and drifting snow sublimation. The latter exposes older snow at the surface, thereby increasing surface snow grain size and lowering the corresponding snow albedo (see section 4). As a result, SW_{net} is larger and T_s increases, which in turn leads to increased 2 m temperatures.

[16] Through turbulent mixing, the water vapor that is released by SU_{ds} is transported vertically into the atmospheric boundary layer and the lower troposphere. Figure 8 presents vertical profiles of temperature and specific/relative humidity at this location. Unlike at the surface, drifting snow incurs cooling throughout the lower troposphere (below 750 hPa), accompanied by an increase in specific humidity below 880 hPa, and a decrease above. The weakened temperature inversion enhances the vertical transport of heat and moisture. The cooling above the drifting snow layer is associated with the upward mixing of the air that was cooled closer to the surface. In combination with vertical diffusion of moisture, this elevates relative humidity occasionally above 80%, triggering the model precipitation scheme, removing moisture from the free atmosphere and resulting in a drying above 880 hPa (see section 5). Wind speed is not appreciably affected by drifting snow, in contrast to earlier studies that assumed that cooling of the near-surface air could enhance the katabatic winds [*Kodama et al.*, 1985].

4. Interactions With the Surface

[17] A critical parameter for drifting snow occurrence is surface snow density (ρ_s). The average model snow density of the upper 5 cm was shown in Figure 1b. Values higher than 400 kg m^{-3} are found on ice shelves, which experience melting in summer. This increases surface snow density and prevents drifting snow to occur, which in part explains the strong annual cycle of drifting snow sublimation and its minimum in summer (Figure 5). An example is shown in Figure 9 for the year 1993 at the location of the German station Neumayer, situated on the Ekström ice shelf. At Neumayer, weak melting occurs in summer, which increases ρ_s to more than 500 kg m^{-3}, in spite of regular accumulation events. Throughout the melting period (usually around 2–3 months), no drifting snow occurs in the model (SU_{ds} = 0). At the end of March, a strong accumulation event causes a lowering of ρ_s to around 350 kg m^{-3}, leading to the start of the drifting snow season. Drifting snow then occurs throughout the winter period from April to November, when ρ_s fluctuates around the fresh snow value, until the melting period starts again in December. The drifting snow observations at Neumayer (3-hourly [*König-Langlo*, 2005]) suggest that this model behaviour is realistic: the occurrence of drifting snow is much reduced in summer compared to winter. In the model, melt increases the surface snow density, increasing the threshold for drifting snow to values >1 m s^{-1}. In reality, this feedback may not be that strong, explaining the infrequent occurrence of drifting snow in the observations.

5. Effect on Surface Mass Balance

[18] Because drifting snow removes mass from the surface by sublimation and redistributes it horizontally by transport, it impacts the surface mass balance (SMB). RACMO2.1/ANT reliably simulates Antarctic SMB. No spatial pattern is detected in the SMB bias when a direct comparison is made with $\sim$750 observations [*Lenaerts et al.*, 2012b], so a post-processing fitting procedure, as used by *Van de Berg et al.* [2006], is not needed [*Lenaerts et al.*, 2012b]. Averaged over the ice sheet (grounded part plus ice shelves), drifting snow sublimation equals 11.7 $\pm$ 0.7 mm yr^{-1}, equivalent to a total for the ice sheet of 164 $\pm$ 10 Gt yr^{-1} (Table 1). Given a mean snowfall of 2696 $\pm$ 133 Gt yr^{-1}, SU_{ds} removes around 6% of the precipitated snow. The same fraction is

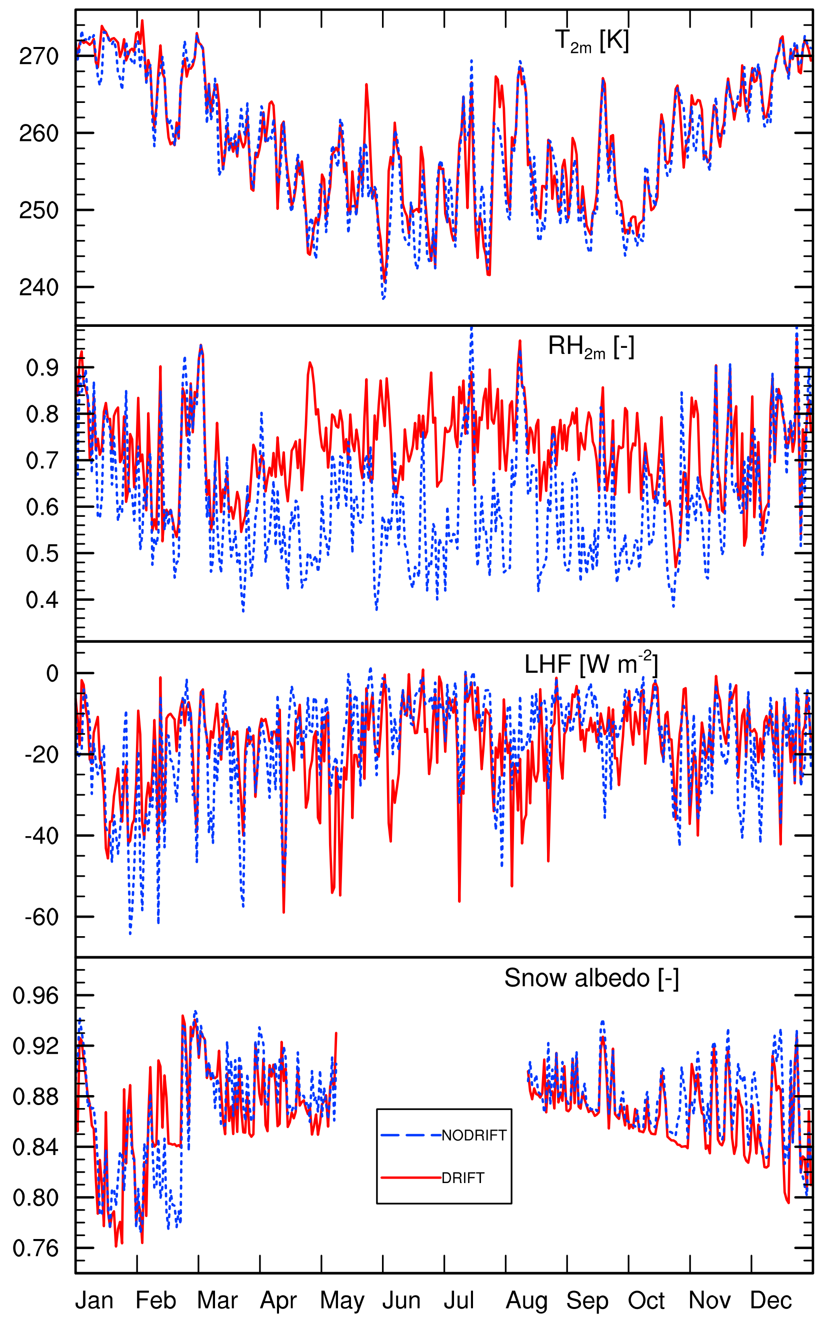

Figure 7. Daily mean (top to bottom) 2 m temperature, 2 m relative humidity with respect to ice, latent heat flux at the surface and snow albedo at 67°S, 145°E for the year 1993 (see Figure 2 for location). Snow albedo is not plotted when downward shortwave radiation is zero (polar night).

found for the grounded ice sheet, where SU_{ds} equals 143 ± 9 Gt yr^{-1} and snowfall 2244 ± 122 Gt yr^{-1}.

[19] Because the formation of large-scale precipitation in RACMO2.1/ANT depends on relative humidity [*Tiedtke*, 1993; *White*, 2001], an interesting result is that the increase of relative humidity in the lower atmosphere owing to drifting snow (Figures 6 and 8) leads to an increase in snowfall of 5–50 mm yr^{-1} in the coastal areas where drifting snow is most active (Figure 10). The largest differences occur in regions where snowfall is abundant and the air is frequently near saturation, such as coastal West Antarctica and high-accumulation areas in Adélie Land, Wilkes Land and Enderby Land. Further over the ocean no clear patterns emerge.

[20] Table 1 lists ice sheet integrated values of SMB components for DRIFT and NODRIFT. Total runoff is small in Antarctica (<5 Gt yr^{-1}), because most meltwater and rain refreezes [*Lenaerts et al.*, 2012b]. That is why with 164 ± 10 Gt yr^{-1}, SU_{ds} clearly is the largest ablation term in Antarctica, and almost three times larger than SU_s. Due to moistening of the atmospheric surface layer, drifting snow reduces surface sublimation compared to NODRIFT, but

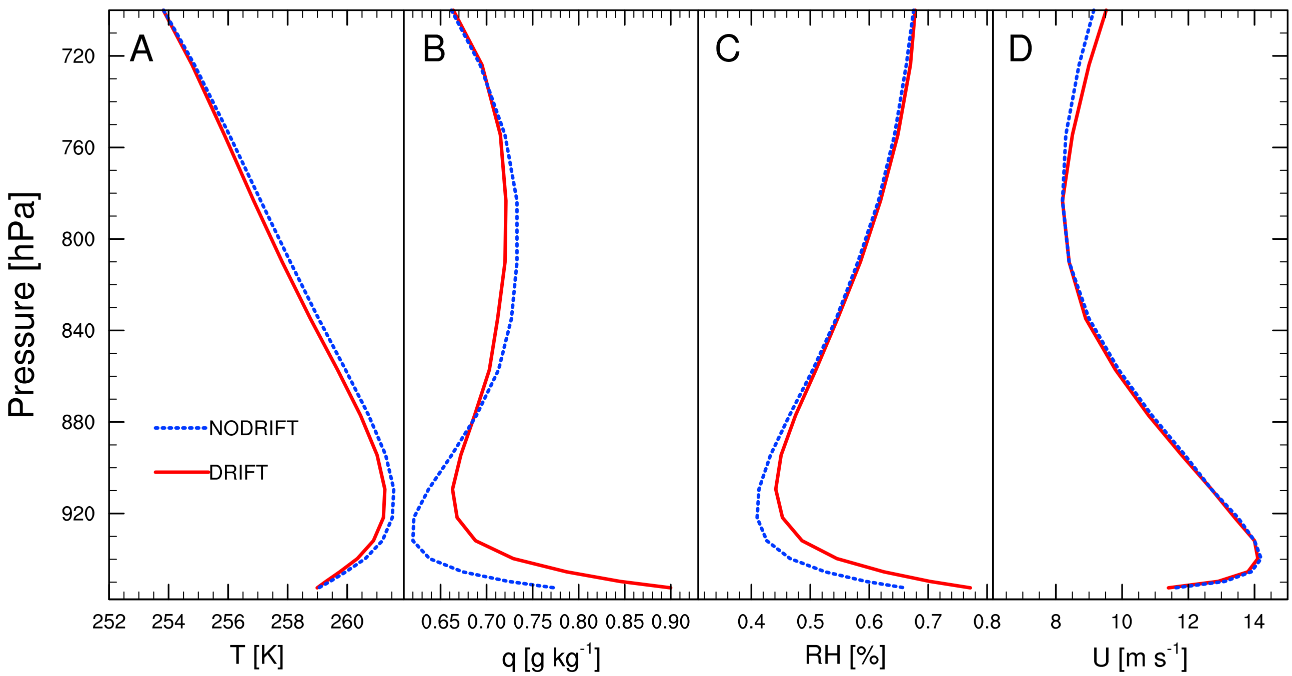

Figure 8. Vertical profiles of (a) temperature, (b) specific humidity, (c) relative humidity with respect to ice, and (d) wind speed averaged for the year 2009 at 67°S, 145°E.

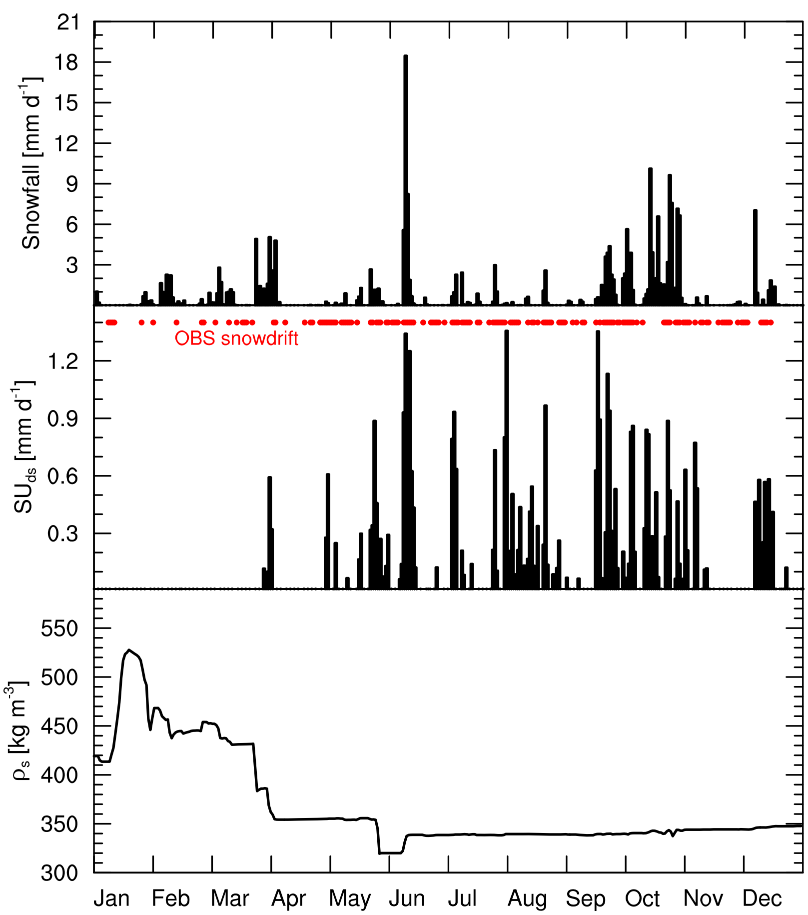

Figure 9. (top) Simulated snowfall, (middle) drifting snow sublimation, both in mm d^{-1}, and (bottom) surface snow density (upper 5 cm) in kg m^{-3} for the year 1993 at Neumayer (70°40′S, 8°16′E). Drifting snow observations are indicated by red dots in the middle plot.

Table 1. Ice Sheet–Integrated Values of Snowfall, Drifting Snow Sublimation (SU_{ds}), Surface Sublimation (SU_s), and Total Sublimation ($SU_{tot} = SU_{ds} + SU_s$) for the DRIFT and NODRIFT Simulations, for Both the Total Ice Sheet (Including Ice Shelves) and the Grounded Ice Sheet (Grounded IS)[a]

	DRIFT		NODRIFT	
	Including Shelves	Grounded IS	Including Shelves	Grounded IS
Snowfall	2699 ± 135	2246 ± 123	2643 ± 126	2196 ± 117
SU_{ds}	164 ± 10	143 ± 9	…	…
SU_s	57 ± 6	34 ± 5	128 ± 10	89 ± 8
SU_{tot}	221 ± 14	177 ± 12	128 ± 10	89 ± 8
ER_{ds}	5 ± 0.1	6 ± 0.2	…	…

[a]Values are in Gt yr^{-1}. Grounding line definition from *Le Brocq et al.* [2010].

increases the total (surface and drifting snow) sublimation [*Van den Broeke et al.*, 2010]. This is confirmed in Table 1: surface sublimation SU_s approximately doubles, in NODRIFT compared to DRIFT, but total sublimation SU_{tot} increases by 73% for the ice sheet including ice shelves, and doubles for the grounded ice sheet in DRIFT. Including drifting snow increases ablation by 93 Gt yr^{-1} for the total ice sheet and by 88 Gt yr^{-1} for the grounded ice sheet. The increase in snowfall due to increased moisture content in the atmospheric boundary layer is around 50 Gt yr^{-1} per year. Combining both effects leads to an additional ablation of 43 Gt yr^{-1} for the total ice sheet and 38 Gt yr^{-1} for the grounded ice sheet. Although important on the regional scale, the contribution of ER_{ds} to the ice sheet-integrated surface mass balance is small (~5 Gt yr^{-1}).

6. Conclusions

[21] This paper presents the drifting snow climate of Antarctica, as obtained using a high-resolution regional climate model coupled to a drifting snow routine based on the bulk drifting snow model of [*Déry and Yau*, 1999]. We found important feedbacks between the atmosphere, the surface and drifting snow. Drifting snow sublimation saturates and cools the near-surface air, limiting its own strength. Moreover, drifting snow alters the local surface mass balance. Finally, drifting snow sublimation and erosion increase the surface snow density and therewith the threshold wind velocity for drifting snow to occur. Surface sublimation and snowmelt peak in summer (November–February), consolidating the upper snow layers and limiting drifting snow sublimation, which peaks in winter (March–October).

[22] Our model results suggest that, through interactions with the atmosphere and the surface, drifting snow has a significant impact on the Antarctic climate. First, drifting snow sublimation, by exposing older snow layers, lowers snow albedo. This increases summer (near-)surface temperatures. Secondly, the humidity content of the near-surface air increases during drifting snow sublimation. This surplus of moisture is transported upwards into the lower troposphere and leads to increased snowfall in parts of coastal areas where the lower atmosphere is frequently near saturation. Finally, the energy required for SU_{ds} also lowers temperatures in the lower atmosphere.

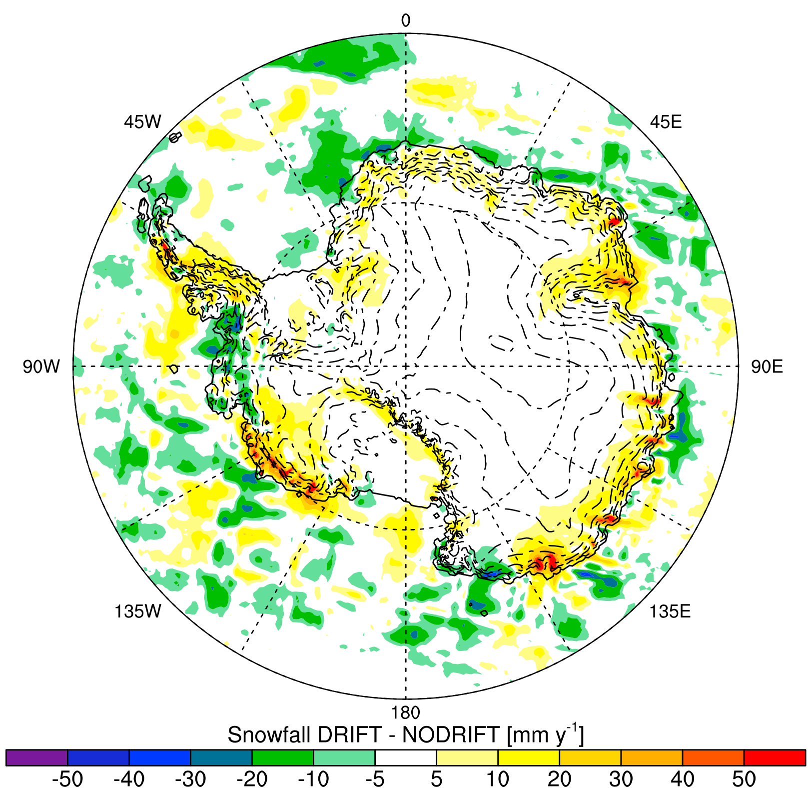

Figure 10. Difference of mean (1989–2009) snowfall (mm yr^{-1}) between DRIFT and NODRIFT.

[23] The contribution of drifting snow sublimation to the Antarctic surface mass balance is significant: 164 ± 10 Gt yr^{-1} or 6% of the annual snowfall is removed by SU_{ds}, clearly the largest surface ablation term in the Antarctic SMB. Surface sublimation is partly replaced by drifting snow sublimation (50% reduction) but total sublimation almost doubles as a result of including drifting snow. The modeled combined effect of drifting snow sublimation and enhanced precipitation is a reduction in the Antarctic SMB of ~50 Gt yr^{-1}, equivalent to ~2% of the annual snowfall. The impact of drifting snow erosion is small when integrated over the ice sheet, but locally it is significant and may lead to ablation.

[24] The current resolution of RACMO2.1/ANT (27 km) is not sufficient to resolve in detail the wind field in the rugged coastal areas, where confluence and terrain strongly influence the wind field. The results for these regions are probably not realistic. These results should therefore be perceived as a first attempt toward an Antarctic drifting snow climatology. A pilot study in which RACMO2.1/ANT is run at higher (5.5 km) resolution over Adélie Land is currently in preparation.

[25] **Acknowledgments.** We thank the reviewers for constructive comments that improved this manuscript. This work was supported by funding from the ice2sea programme from the European Union 7th Framework Programme, grant number 226375. This is Ice2sea contribution 060.

References

Bintanja, R. (1998), The interaction between drifting snow and atmospheric turbulence, *Ann. Glaciol.*, *26*, 167–173.
Bintanja, R. (2001a), Snowdrift sublimation in a katabatic wind region of the Antarctic ice sheet, *J. Appl. Meteorol.*, *40*(11), 1952–1966.
Bintanja, R. (2001b), Modelling snowdrift sublimation and its effect on the moisture budget of the atmospheric boundary layer, *Tellus, Ser. A*, *53*(2), 215–232.
Bintanja, R., and C. H. Reijmer (2001), A simple parameterization for snowdrift sublimation over Antarctic snow surfaces, *J. Geophys. Res.*, *106*(D23), 31,739–31,748.
Box, J. E., D. H. Bromwich, B. A. Veenhuis, L. S. Bai, J. C. Stroeve, J. C. Rogers, K. Steffen, T. Haran, and S. H. Wang (2006), Greenland ice sheet surface mass balance variability (1988–2004) from calibrated polar MM5 output, *J. Clim.*, *19*(12), 2783–2800.
Bromwich, D. H., Z. Guo, L. Bai, and Q. Shen (2004), Modeled Antarctic precipitation: part i. Spatial and temporal variability, *J. Clim.*, *17*(3), 427–447, doi:10.1175/1520-0442(2004)017.
Budd, W. F., W. R. J. Dingle, and U. Radok (1966), The Byrd Snow Drift project: Outline and basic results, in *Studies in Antarctic Meteorology*, *Antarct. Res. Ser.*, vol. 9, edited by M. J. Rubin, pp. 71–134, AGU, Washington, D. C.
Church, J. A., J. M. Gregory, P. Huybrechts, M. Kuhn, K. Lambeck, M. T. Nhuan, D. Qin, and P. L. Woodworth (2001), Changes in sea level, in *Climate Change 2001: The Scientific Basis—Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, edited by J. T. Houghton et al., chap. 3, pp. 659–664, Cambridge Univ. Press, New York.
Déry, S. J., and M. K. Yau (1999), A bulk blowing snow model, *Boundary Layer Meteorol.*, *93*(2), 237–251.
Frezzotti, M., et al. (2005), Spatial and temporal variability of snow accumulation in East Antarctica from traverse data, *J. Glaciol.*, *51*, 113–124.
Frezzotti, M., S. Urbini, M. Proposito, C. Scarchilli, and S. Gandolfi (2007), Spatial and temporal variability of surface mass balance near Talos Dome, East Antarctica, *J. Geophys. Res.*, *112*, F02032, doi:10.1029/2006JF000638.
Gallée, H., G. Guyomarch, and E. Brun (2001), Impact of snow drift on the Antarctic ice sheet surface mass balance: Possible sensitivity to snow-surface properties, *Boundary Layer Meteorol.*, *99*, 1–19.
Kodama, Y., G. Wendler, and J. Gosink (1985), The effect of blowing snow on katabatic winds in Antarctica, *Ann. Glaciol.*, *6*, 59–62.
König-Langlo, G. C. (2005), Meteorological synoptical observations from Neumayer Station (1993–01), report, Alfred Wegener Inst. for Polar and Mar. Res., Bremerhaven, Germany, doi:10.1594/PANGAEA.271730.
Le Brocq, A. M., A. J. Payne, and A. Vieli (2010), An improved Antarctic dataset for high resolution numerical ice sheet models (ALBMAP v1), *Earth Sys. Sci. Data*, *2*, 247–260, doi:10.5194/essd-2-247-2010.
Lenaerts, J. T. M., M. R. van den Broeke, S. J. Déry, G. König-Langlo, J. Ettema, and P. Kuipers Munneke (2010), Modelling snowdrift sublimation on an Antarctic ice shelf, *The Cryosphere*, *4*(2), 179–190, doi:10.5194/tc-4-179-2010.
Lenaerts, J. T. M., M. R. van den Broeke, S. J. Déry, E. van Meijgaard, W. J. van de Berg, S. P. Palm, and J. Sanz Rodrigo (2012a), Modeling drifting snow in Antarctica with a regional climate model: 1. Methods and model evaluation, *J. Geophys. Res.*, *117*, D05108, doi:10.1029/2011JD016145.
Lenaerts, J. T. M., M. R. van den Broeke, W. J. van de Berg, E. van Meijgaard, and P. Kuipers Munneke (2012b), Present-day (1989–2009) high-resolution surface mass balance of Antarctica from a regional atmospheric climate model, *Geophys. Res. Lett.*, doi:10.1029/2011GL050713, in press.
Mahesh, A., R. Eager, J. R. Campbell, and J. D. Sphinhirne (2003), Observations of blowing snow at the South Pole, *J. Geophys. Res.*, *108*(D22), 4707, doi:10.1029/2002JD003327.
Mann, G. W., P. S. Anderson, and S. D. Mobbs (2000), Profile measurements of blowing snow at Halley, Antarctica, *J. Geophys. Res.*, *105*(D19), 24,491–24,508.
Parish, T. R., and D. H. Bromwich (2007), Reexamination of the near-surface airflow over the Antarctic continent and implications on atmospheric circulations at high southern latitudes, *Mon. Weather Rev.*, *135*(5), 1961–1973, doi:10.1175/MWR3374.1.
Rignot, E., I. Velicogna, M. R. van den Broeke, A. Monaghan, and J. T. M. Lenaerts (2011), Acceleration of the contribution of Greenland and Antarctic ice sheets to sea level rise, *Geophys. Res. Lett.*, *38*, L05503, doi:10.1029/2011GL046583.
Scarchilli, C., M. Frezzotti, P. Grigioni, L. De Silvestri, L. Agnoletto, and S. Dolci (2010), Extraordinary blowing snow transport events in east Antarctica, *Clim. Dyn.*, *34*, 1195–1206, doi:10.1007/s00382-009-0601-0.
Steinhoff, D. F., S. Chaudhuri, and D. H. Bromwich (2009), A case study of a Ross ice shelf airstream event: A new perspective, *Mon. Weather Rev.*, *137*(11), 4030–4046.
Tiedtke, M. (1993), Representation of clouds in large-scale models, *Mon. Weather Rev.*, *121*, 3040–3061.
Turner, J., S. N. Chenoli, A. Abu Samah, G. Marshall, T. Phillips, and A. Orr (2009), Strong wind events in the Antarctic, *J. Geophys. Res.*, *114*, D18103, doi:10.1029/2008JD011642.
Van de Berg, W. J., M. R. van den Broeke, C. H. Reijmer, and E. van Meijgaard (2005), Characteristics of the Antarctic surface mass balance (1958–2002) using a regional atmospheric climate model, *Ann. Glaciol.*, *41*, 97–104.
Van de Berg, W. J., M. R. van den Broeke, C. H. Reijmer, and E. van Meijgaard (2006), Reassessment of the Antarctic surface mass balance using calibrated output of a regional atmospheric climate model, *J. Geophys. Res.*, *111*, D11104, doi:10.1029/2005JD006495.
Van den Broeke, M. R. (2005), Strong surface melting preceded collapse of Antarctic Peninsula ice shelf, *Geophys. Res. Lett.*, *32*, L12815, doi:10.1029/2005GL023247.
Van den Broeke, M. R., N. P. M. van Lipzig, and E. van Meijgaard (2002), Momentum budget of the East-Antarctic atmospheric boundary layer: Results of a regional climate model, *J. Atmos. Sci.*, *59*(21), 3117–3129.
Van den Broeke, M. R., G. C. König-Langlo, G. Picard, P. Kuipers Munneke, and J. T. M. Lenaerts (2010), Surface energy balance, melt and sublimation at Neumayer Station, East Antarctica, *Antarct. Sci.*, *22*(1), 87–96.
White, P. W. (2001), Physical processes (CY23R4), technical report, Eur. Cent. for Med. Range Weather Forecasts, Reading, U. K.
Zwally, H. J., and M. B. Giovinetto (2011), Overview and assessment of Antarctic ice-sheet mass balance estimates: 1992–2009, *Surv. Geophys.*, *32*(4–5), 351–376, doi:10.1007/s10712-011-9123-5.

J. T. M. Lenaerts and M. R. van den Broeke, Institute for Marine and Atmospheric Research Utrecht, Utrecht University, Princetonplein 5, Utrecht NL-3584 CC, Netherlands. (jtmlenaerts@gmail.com)