Review: Sensitivity of the surface energy budget to drifting snow as simulated by MAR in coastal Adelie Land, Antarctica

General notes

Please see our answer to R1.Q3 ("Reviewer 1, question 3") and R2.Q3. New model simulations have been computed, consequently all statistics presented in the original manuscript have undergone modifications. Please find at the end of the review the updated figures and tables. The text in the manuscript has been updated accordingly.

Reviewer 2

This is an interesting study on the impact of drifting and blowing snow on boundary layer meteorology, surface radiation and energy balance in Terre Adélie, a region on the slopes of the East Antarctic ice sheet. A 9-yr time series (2010-18) of observations at a site near the coast is used to validate the regional climate model MAR. Methodology, presentation and discussion of results as well as the conclusions are sound. One of the main findings is that sublimation of drifting-snow particles leads at the surface to a reduction in sensible and latent heat exchange, which is compensated by an increase in net radiative forcing. While the net impact on total surface energy budget, and therefore surface temperature, is minimal, structure of the lower atmosphere is modified, which needs to be resolved in climate models to understand impact of warming on air-surface interactions and boundary layer meteorology. I have only minor comments, and recommend publication after they have been addressed.

We thank reviewer 2 for making a positive review and instructive comments. Please find below our response to each of the points raised in the review.

R2.Q1

1. It would be useful to get more detail on how sublimation rates are computed in MAR and how reliable they are. What are the model assumptions (snow particle size distribution and shape).

The sublimation in MAR is distinguished between surface sublimation and sublimation of airborne particles (including both cloud-originating particles and drifting-snow particles raised from the surface), we suggest to modify and complete paragraph line 143 as follow:

“[...]MAR is coupled to the surface scheme SISVAT (Soil Ice Snow Vegetation Atmosphere Transfer; De Ridder and Gallée (1998), Gallée and Duynkerke (1997), Gallée et al. (2001)), which handles energy and mass transfer between the atmosphere and the surface, and includes a multi layer snow/ice model representing snow properties (dendricity, sphericity and size) taken from an early version of the CROCUS snow model (Brun et al. 1992). Surface sublimation (which is distinguished in the model from atmospheric sublimation) and latent heat exchanges at the surface are computed following a bulk flux formulation in SISVAT.

*MAR includes a drifting-snow scheme originally described in Gallée et al. (2001). A detailed description of MARv3.11 latest version (including updates, changes relative to the original version and interactions with the surface and the atmosphere) can be found in Amory et al. (2020). In brief, the drifting-snow scheme simulates erosion at every grid cell in which the modelled friction velocity exceeds a threshold value, u*t, depending on local surface snow density. While former parameterisations of u*t in the model did involve other snow microstructural properties such as snow grain shape and size (Gallée et al., 2001) for which observations are virtually non-existent in Antarctica, here the formulation for u*t has been simplified and sensitivity parameters have been reduced to surface snow density only, a variable better observationally constrained (Amory et al. 2020). Once removed from the snowpack, eroded snow is mixed with the pre-existing windborne snow mass and advected to higher atmospheric levels and/or downwind grid cells by the turbulence and microphysical schemes. Interactions with the atmosphere are computed by the microphysical and the radiative transfer schemes. More particularly, atmospheric sublimation (including both cloud-originating particles and drifting-snow particles) is computed by the model microphysics (Gallée 1995). It incorporates a formulation for snow sublimation in the atmosphere (Lin et al. 1983). This formulation is based on the assumption of an exponential distribution for particle size and is a function of the air temperature, snow particles ratio and relative humidity (so that sublimation only occurs in a subsaturated environment, with respect to ice). It also considers snow particles as graupel-like snow of hexagonal type (Gallée et al. 1995, Locatelli and Hobbs 1974). Consequently, drifting-snow sublimation modifies the local humidity budget, the lower atmosphere stratification and moist air advection. Representing the contribution of drifting-snow layers to the atmospheric radiative forcing is accounted for in MAR by including suspended snow particles in the computation of cloud radiative properties (Gallée et Gorodetskaya, 2010). “*

This paragraph will be complemented by the additional information on how drifting snow affects the momentum budget in the boundary layer described in R1.Q2. Finally, we refer to our answer to R2.Q3 for a more detailed discussion on the reliability of sublimation rates.

R2.Q2

1.2 Another model parameterisation of bulk sublimation rates from blowing snow by Déry and Yau (1999) uses the mean snow particle diameter. Is this the case in MAR and is particle diameter a sensitive parameter? If yes, future studies would gain by deploying next to an electro-acoustic sensor also an optical particle counter, to measure particle diameter as well as snow mass flux more accurately.

We refer to the previous question concerning the computation of atmospheric sublimation in MAR, which is not directly based on an assumption of a particle diameter but rather on an assumption of the distribution of particle size. However, we would like to inform the reviewer and our readers that an optical snow particle counter (SPC) has been deployed at D17 in January 2014 for a single drifting snow event to initiate first comparisons with FlowCapt sensors in Antarctic conditions, and assess the ability of the FlowCapt to measure drifting snow fluxes (S1, supplement of Amory 2020). Energy supply issues have so far limited the permanent use of an SPC at D17 . However, we do believe too that deploying an SPC at D17 in complement of acoustic measurement, such as already done at the Col du Lac Blanc in the french Alps (Naaim-Bouvet et al. 2013), would complete the already existing set of measurement devices and help future studies to assess modeling hypothesis and help the scientific community better understand drifting-snow physics. We refer to our answer to R2.Q3 including a discussion in the manuscript about the potential benefit of optical measurements at D17.

R2.Q3

What are the uncertainties of calculated sublimation rates and how do calculations compared to existing observations in Antarctica (e.g. hourly blowing snow sublimation rates at Halley range 0.1-1 mm we/day (King et al., 1996))?

We thank the reviewer for that question that helped us spot a miscalculation in the initial model results, for which we apologize. The sublimation rates, as initially presented on Fig. 3 (b) and Fig. 6 (j), were not correctly computed. We corrected this issue, relaunched the simulations for the 9 year period (in addition to the correction related to R1.Q3), and we now report updated sublimation rates, computed following the method described in R2.Q1.

Sublimation rates are now expressed in [kg of sublimated snow / kg of moist air] instead of [mm w.e. year] on Fig. 3 and Fig. 6. Such a choice is justified to preserve the consistency between the model meteorological variables presented in Fig. 3. and Fig. 6 (wind speed, relative humidity, temperature…), which are representative of values averaged over each vertical model level. Such averaged values can not be directly compared to sublimation rates expressed in mm we (per unit of time), as such rates are representative of total sublimation, integrated over the thickness of each model vertical level (a better comparison could be done by comparing e.g. sublimation rates in [mm w.e.] to latent heat release integrated over the thickness of the model vertical level). The unit now presented is appropriate for estimating a sublimation rate intensity, comparable between different models levels of distinct thicknesses to other variables as presented in Fig. 3 and Fig. 6.

Additionally, we estimated drifting-snow sublimation as the difference in atmospheric sublimation over the first 1000 m above ground between MAR-DR and MAR-nDR. The averaged drifting-snow sublimation over the integration domain is 606 mm we/year and is locally higher at D17 with 719 mm we/year. Such estimates are affected, among other, by a source of uncertainty in the model microphysics: as the atmosphere contains more humidity in MAR-DR due to enhanced atmospheric sublimation in comparison with MAR-nDR (and as eroded snow particles can also act as nuclei particles), MAR-DR might simulate more clouds and associated precipitations than MAR-nDR. Newly formed cloud and/or precipitations could potentially sublimate, which could induce an overestimation of drifting-snow sublimation rates. A distinction between sublimation of snowfalls and eroded snow particles could be theoretically done in the model, but is not currently implemented in MAR. Drifting-snow sublimation rates presented above thus can not be directly interpreted as a surface mass balance component as they do include the sublimation of cloud-originating particles that have not reach the surface yet. .

King et al. (1996) reported lower estimates of drifting-snow sublimation using a sublimation model fitted on observed data at Halley station (mean values being typically from 1 to 2 order of magnitude lower, but peak values of the same order of magnitude). Similarly, King et al (2001) and Bintanja et Reijmer. (2001) report sublimations rates up to 50 mm we/year and 70 mm we/year at Halley and near Svea station in Dronning Maud Land (sometimes even including surface sublimation). Finally, Bintanja (1998) report higher values in Adelie Land (typically around 150 mm we/year), characterized by a strong spatial variability.

MAR-DR simulates stronger drifting-snow sublimation rates. However, Palm et al. 2017 estimated, using remotely sensed data, that drifting-snow sublimation rates could be up to 250 mm we/year in Adelie Land. Finally, Lenaerts and van den Broeke 2012 estimated drifting-snow sublimation in Adelie Land to be up from 150 to more than 300 mm we/year. Both Palm et al. 2017 and Lenaerts and van den Broeke 2012 reported a very high spatial variability in drifting-snow sublimation.

Important disparities between models/instruments used to estimate drifting-snow sublimation (e.g. maximum/minimum height above ground until/from which sublimation in computed) and climatic differences between sites where measures have been held exist: e.g. D17 is both windier and warmer than Halley (King et al. 2001), favoring more drifting-snow sublimation. Consequently, quantifying the potential (and probable) extent to which MAR-DR overestimate drifting-snow sublimation could hardly be done by just comparing with pre-existing estimates with different methods and from other locations, and we believe would lie beyond the scope of our study (an independent study using the same model is currently being held on the entire continent to discuss and quantify drifting-snow sublimation).

Accordingly to our answer to Reviewer 1 R1.Q5, we suggest to delete paragraph 4.3 and modify paragraph 4.4 into*"4.3 Current limitations"*. Consequently, we suggest to summarize this discussion as follows (this discussion is inserted in the main text as presented in R1.Q5):

"Drifting-snow sublimation, defined here by the difference in atmospheric sublimation between MAR-DR and MAR-nDR over the first 1000m above ground, equals on average 606 mm we/year on the all Adelie Land domain, with higher values reported at D17 (716 mm we/year). These rates are larger than previous in-situ estimates of drifting-snow sublimation held in distinct parts of the continent (King et al. 1996, King et al. 2001) where the climate differs from the windy and (relatively warm) conditions of coastal Adelie Land. However, Palm et al 2017, through remotely sensed data, and Lenaerts and van den Broeke 2012, by using a regional climate model, report sublimation rates in Adelie Land which are more in agreement with our model estimates, though still twice to three times lower. Finally, the inclusion of newly formed clouds in MAR-DR as discussed in Sect. 2.4 can contribute to the probable overestimation of drifting-snow sublimation rates in MAR-DR. The sublimation rates, as simulated by MAR, have not been yet directly compared to in-situ measurements, although indirect comparisons have been made through the evaluation of near-surface air relative humidity and temperature. Accounting for drifting-snow sublimation in the present study has proven useful to modify the relative humidity of the lower atmosphere and help the model matching with observed relative humidity from a timescale of a single event (Fig. 2) to a seasonal scale (Fig. 3). The deployment of eddy-covariance systems including highly sensitive hygrometers could provide complementary atmospheric sublimation estimates to evaluate model simulations during calm to moderate conditions. However, using eddy-covariance devices during strong drifting-snow episodes remains a challenge as drifting-snow particles alter the observed signal and limit their use in Adelie Land (e.g. Bintanja, 2001). Moreover, including drifting snow in MAR shows large impacts on turbulent fluxes which compensate (and sometimes slightly override, e.g. at D17) modifications in radiative fluxes. Such a compensation also needs to be evaluated through comparison with direct in situ measurements of latent and sensible heat fluxes during drifting-snow occurrences to determine if MAR-DR simulates (more) realistic turbulent heat exchanges at the surface. Modeling hypothesis regarding drifting-snow particle distribution and subsequent sublimation rates could be better constrained using information derived from in-situ optical measurements (e.g. Naaim-Bouvet 2013)"

R2.Q4

2. The limitations of the current evaluation method needs to be expanded (Section 4.4), in order to guide future observations, which parameters should be measured to better constrain the model. Comment also on model uncertainties in vertical profiles (e.g. T, RH, wind speed, sublimation rate) and drift layer height, and how they would impact on the main conclusions.

We fully agree that the reader should be aware of the scope of validity of the results, its assumptions and the uncertainties associated. This discussion will be added to section 4.3 (see R1.Q5).

"The vertical profiles presented in Fig. 6 have only been evaluated at 2 m. a.g.l. therefore, the behavior of the model, and the eventual benefit of accounting for drifting snow in order to capture more realistic atmospheric dynamics in the lower atmosphere still needs to be assessed. Daily radio soundings are operated at the closeby permanent station of Dumont D'Urville. However, sufficient climatic disparity exists between D17 location, situated on the marginal slope of the Antarctic continent and Dumont d'Urville station, situated beyond the continent boundaries on an island approximately 15 kilometers northeast of D17. Nevertheless a good agreement with observed values for several meteorological variables (wind speed, relative humidity, temperature, drifting-snow fluxes, incoming and outgoing radiative fluxes) and the fact that the model is well constrained at its boundaries by global reanalysis is an argument in favour of firstly studying model outputs at the first vertical level and then exploring its behavior at higher altitudes. Dropsondes observations near D17 location or operation of radiosoundings from the ground at D17 would help assess model performance and uncertainties at higher elevation in complement of near-surface observations."

We refer to line 401 for the discussion about the necessity to evaluate turbulent fluxes using methods complementary to the bulk/profile methods. Such methods, such as e.g. the deployment of eddy covariance measuring devices, are still limited by the presence of hydrometeors in the atmosphere during drifting-snow episodes (e.g. Bintanja, 2001) and highlight the difficulty to measure drifting-snow sublimation.

We propose to better discuss the current limitation of our estimation of drifting-snow layer heights by adding the following sentences (inserted in Sect 4.3, see R1.Q5):

"Additionally, we introduced a method, based on CALIPSO observations to estimate the height of a drifting-snow layer using model outputs. This method allows us to derive an objective criterion concerning snow concentration in the atmosphere to determine the presence (or not) of a drifting-snow layer and its height, during specific meteorological conditions. This method is limited by the fact that it has only been developed for 8 years

of CALIPSO observations collected near D17. Future work could focus on other locations in Antarctica to improve the determination of the snow concentration threshold by gathering more remotely sensed observations to be compared to model simulations. This could ultimately lead to an evaluation of modeled drifting-snow layer heights using CALIPSO observations on a specific test dataset. Ultimately, the use of a grounded lidar at D17 could provide complementary information concerning the vertical structure of drifting-snow layers."

Finally, we refer to our answer to R2.Q3 to underline current limitations in assessing the ability of MAR to simulate realistic sublimation rates.

SPECIFIC COMMENTS

R2.Q5

L142 Please provide detail on the model parameters for snow particles (size, shape) used in MAR.L145 '...the drifting-snow scheme simulates erosion at every grid cell in which the modelled wind shear exceeds a threshold value depending on the local surface snow density.' What is this threshold value? How is it parameterised? Does snow particle size play a role? Please expand.

The parameterisation of the threshold friction velocity for initiation of drifting snow u_{*t} (m/s) is fully described in Amory et al. (2020) and is based only on surface snow density.

$$u_{*t} = u_{*t0}\, exp(\frac{\rho_i}{\rho_0} - \frac{\rho_i}{\rho_s}) \qquad \text{Eq. R2}$$

$$u_{*t0} = \frac{log(2.868)-log(1+0.625)}{0.085} C_D^{\ 0.5} \qquad \text{Eq. R3}$$

$$C_D = \frac{u_*^{\ 2}}{U^2} \qquad \text{Eq. R4}$$

where ρ_s the surface snow density, ρ_i the density of ice, ρ_0 the density of fresh snow, C_D the drag coefficient for momentum, U the wind speed in the lowest model vertical level and u_* the friction velocity. Please find more information in Amory et al. (2020).

Please find below further elements we suggest to add to the model description line 145:

*"In brief, the drifting-snow scheme simulates erosion at every grid cell in which the modelled friction velocity exceeds a threshold value, u*t, depending on local surface snow density. While former parameterisations of u*t in the model did involve other snow microstructural properties such as snow grain shape and size (Gallée et al., 2001) for which observations are virtually non-existent in Antarctica, here the formulation for u*t*

has been simplified and sensitivity parameters have been reduced to surface snow density only, a variable better observationally constrained (Amory et al. 2020).”

R2.Q6

L231 ‘snow particle ratio’; do you mean here snow particle mixing ratio (mass of suspended snow particles to that of dry air)? Please clarify.

Here the snow particle ratio refers to the snow particle specific ratio: the mass of suspended snow particles to the total mass of air including dry air, humidity and the mass of all other hydrometeors.

We included this specification L172, to the first appearance in the text of the term snow particle ratio:
“[...] the snow particle ratio (the specific ratio, which equals the mass of snow particles per kg of air, including dry air, humidity and the mass of all other hydrometeors) [...]

R2.Q7

L243/Fig.3 Expand explanation - I assume the vertical maxima in SWnet (Fig.3d) reflect the diurnal cycle, and a small reduction is seen <100m on 3 Oct compared to 2 Oct,but the impact of drift snow on LWnet is only noticeable below 50m (Fig.3c). Why is that? And is it consistent with estimated drift snow layer heights during that time?

Your assumption is right, the diurnal cycle is indeed retrieved in the SWnet representation in Fig. 3d (intense yellow areas, corresponding to high SWD periods). We intended to point out simultaneous modifications in both modeled SWnet and LWnet during a drifting snow event occurring on the 3rd of October 2017 (Fig. 3 (e)). As pointed out by the reviewer, a first visual analysis on Fig. 3d indicates that modifications in SWnet with elevation start at approximately 100m while LWnet modifications are only visible below 50m. Small variations of LW are not retrieved in the current visualization but a finer data analysis point out increases smaller than 5 W.m^2 up to 100m.

Furthermore, we would like to underline the fact that drifting-snow layer heights were not initially retrieved using observations based on radiative modifications of longwave and shortwave fluxes. Indeed, CALIPSO observations consist of lidar measurements and the algorithm used in MAR deals with snow concentrations. We would also like to point out that drifting-snow concentration in the atmosphere decreases exponentially with height (Fig. 6 (i)). As a consequence, as snow residence in the atmosphere drives radiative modifications, more important radiative effects can be expected when approaching the surface. Consequently we believe it is consistent to simulate drifting-snow layer heights (representative of snow concentrations in the atmosphere) at a specific elevation (Fig. 3 (a)) and retrieve the more important radiative effects at a close but different elevation (e.g.

Fig 3 (c) and (d)). Finally, Fig. 3 suggests that the radiative modification with altitude, as observed on Fig. 3, depends on the type of radiation (shortwave and longwave) in MAR.

R2.Q8

L278 'Drifting snow modifies the seasonal values of incoming radiative fluxes by enhancing LWD and decreasing SWD (Fig. 4 (e) and (g)). ' - the latter does not seems to be supported by Fig.4g, both model scenarios plot on top of each other, please clarify.

SWD modifications with drifting-snow are visible on single specific events, such as presented on Fig. 3d, but such modifications are not sufficient to be retrieved in seasonal means at D17, as presented on Fig. 4g. We suggest to modify sentence L278 to:

"Drifting snow enhances the seasonal values of LWD (Fig. 4 (e)), but even if significant modifications in SWD can occur during specific events such as presented in Fig. 3d, the impact on seasonal averages is low (Fig. 4 g)."

The seasonality of SWD can partially explain such a difference between larger increases in LWD (see Sect 3.3), which are retrieved in seasonal means, and decreases in SWD, which are not visible in Fig. 4. SWD are weak or null a large part of the year in the high latitudes of Adelie Land meanwhile longwave emission of drifting-snow particles remains positive all year long. Furthermore wind speeds are stronger in winter at D17 (Fig. 4 (b)) favoring stronger and more frequent drifting snow during that part of the year, and thus greater radiative effects on LWD (see Fig. 4 (e)). However, Fig. 8 (Fig. 7 in the original manuscript) indicates that, on a yearly basis, drifting snow induces significant decreases in SWD in locations experiencing more intense drifting snow than D17.

R2.Q9

Fig.3b What is the averaging period simulated sublimation rate refers to? Per 30min or per hour? Are these values consistent with observations existing elsewhere in Antarctica?

Fig. 3 (b) (corrected, see our answer to R2.Q3) reports the quantity of snow sublimated during 30 min (the time step of model outputs), and is expressed in kg of sublimated snow per kg of air (see R2.Q3). This is now specified in Fig. 3. In the modified Fig. 6 (j), sublimation rates are expressed in kg of sublimated snow per kg of air per year. Please see our answer to R2.Q3 for further detail on the unit used for sublimation rates and comparison to observed value in Antarctica.

TECHNICAL COMMENTS

Fig.3l Place legend outside the figure panel.

The legend location has been modified.

Fig.5a,c The grey shaded area to illustrate RMSE is missing.

For better readability, we propose to keep the figure as it is and thus to delete the mention concerning missing RMSE.

Updated figures and tables

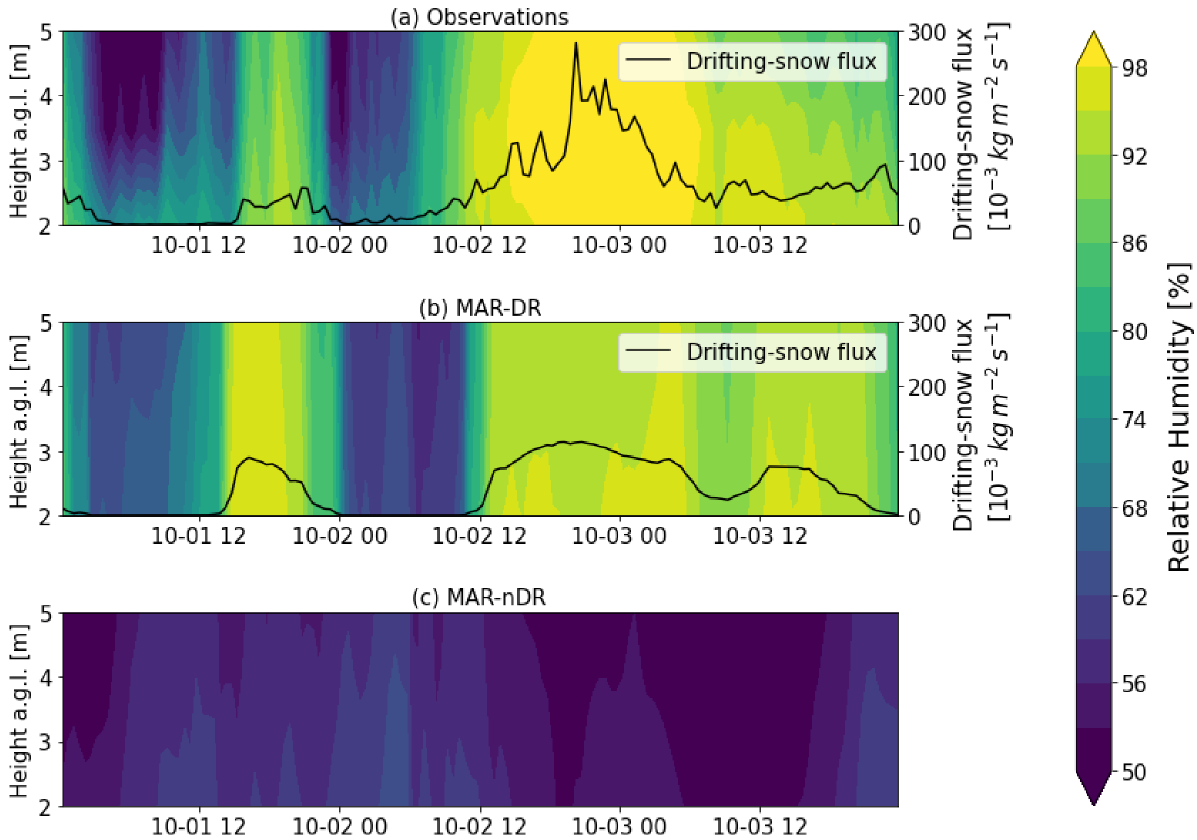


Figure 2. *(a) Observed, (b) MAR-DR and (c) MAR-nDR vertical relative humidity profile (with respect to ice, color) and drifting-snow fluxes (from the surface to 2 m a.g.l., black line) between the 1st and the 3rd of October 2017.*

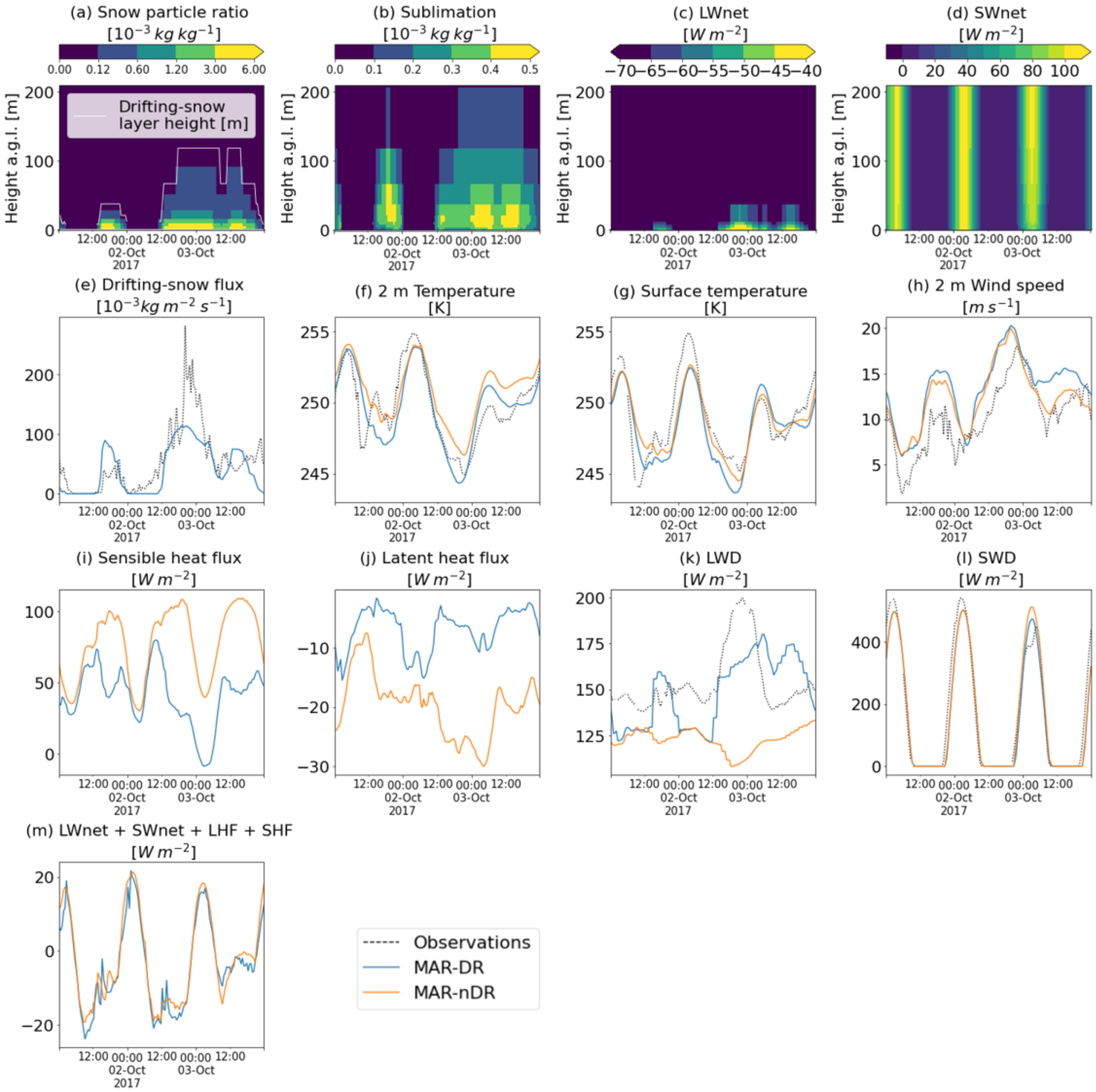


Figure 3. *(a) Snow particle ratio, (b) sublimation (expressed in g of sublimated snow per kg of moist air per 30 minutes), (c) SWnet and (d) LWnet vertical profiles as simulated by MAR-DR during a drifting-snow episode occurring between the 1st and the 3rd of October 2017. (e) to (m): 2 m and surface variables as observed and simulated by MAR-DR and MAR-nDR between the 1st and the 3rd of October 2017.*

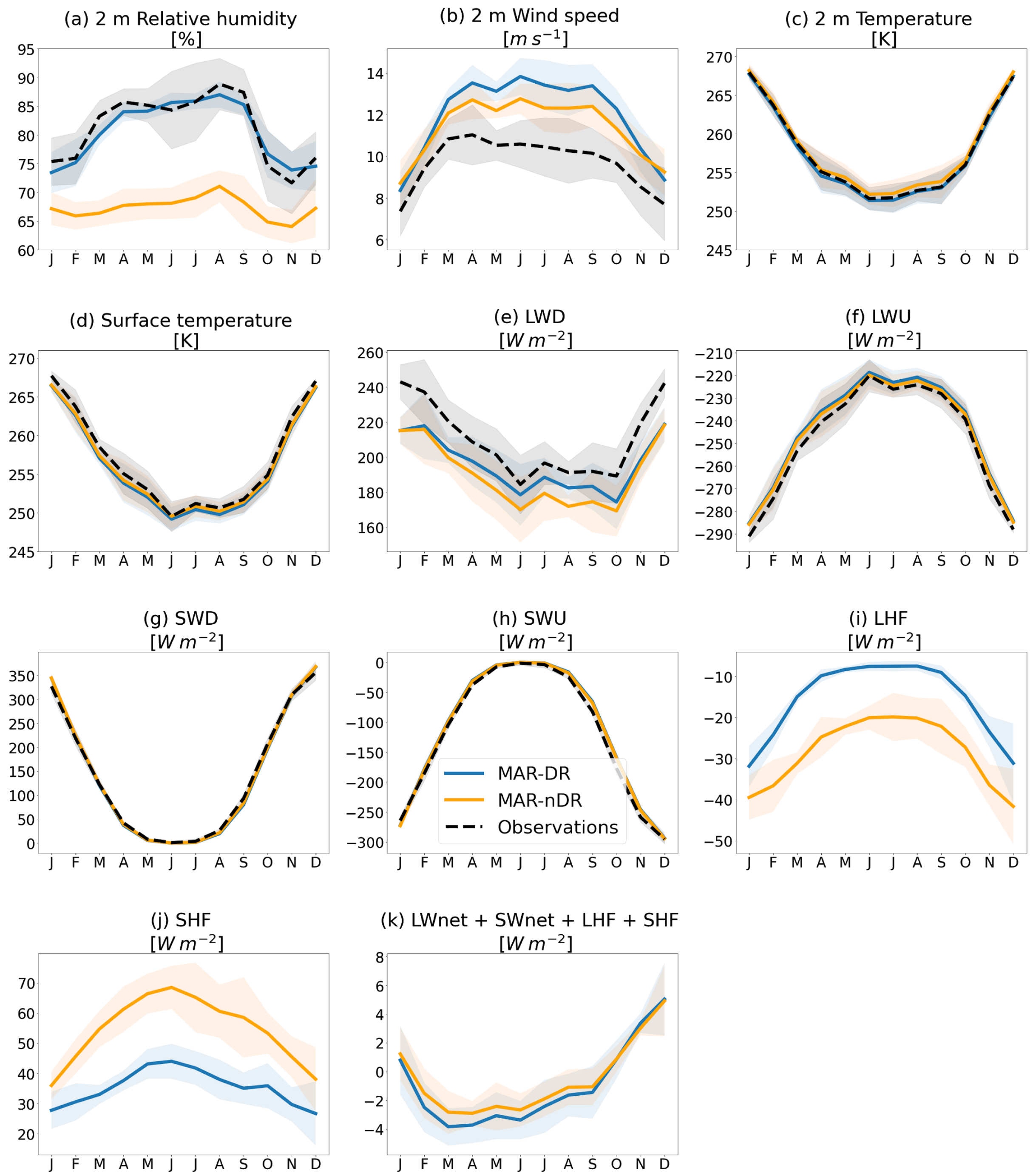


Figure 4. *2 m and near-surface variable monthly means as simulated by MAR-DR and MAR-nDR. First, data are aggregated by both months and years. Then means and standard deviations are evaluated within each group aggregated by month. Statistics are performed on 2014–2018 period for radiative fluxes and surface temperature and on 2010–2018 period for near-surface variables and turbulent fluxes.*

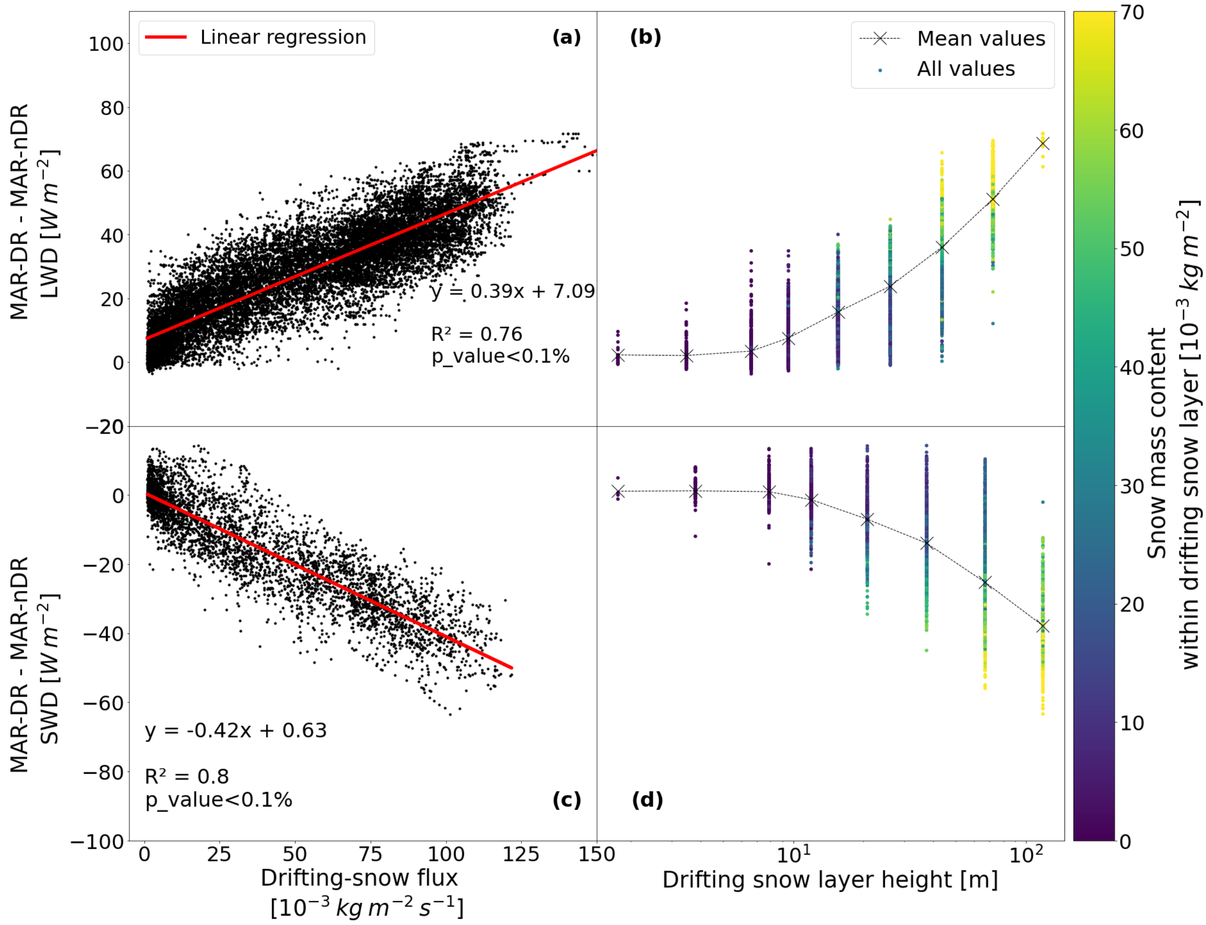
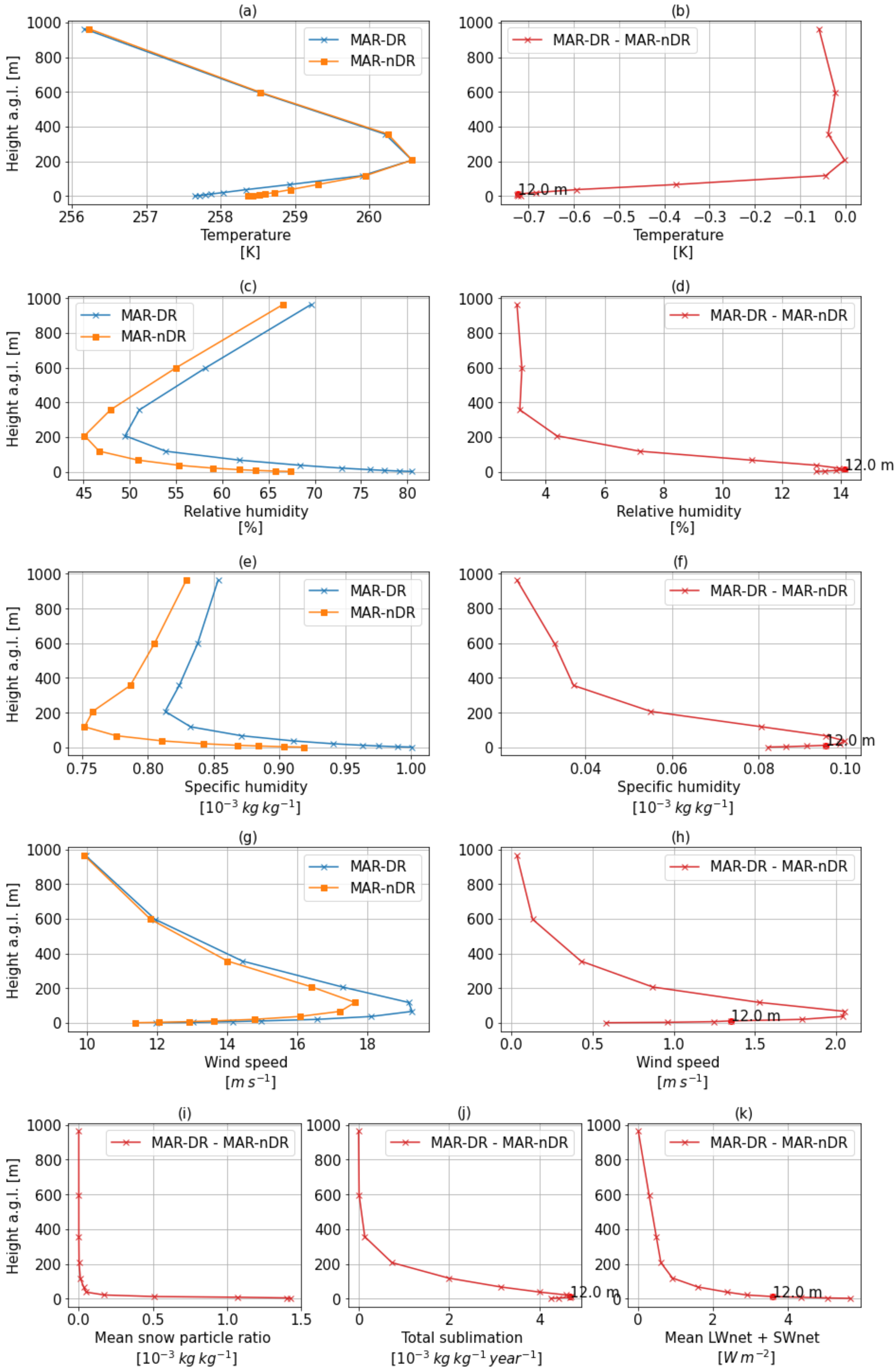


Figure 5. *Modifications in (a) LWD and (c) SWD between MAR-DR and MAR-nDR during drifting snow (drifting-snow flux >* $10^{3} kg\,m^{-2} s^{-1}$*), as a function of drifting-snow flux, for a mean flux calculated between 0 and 2m. The red line indicates the best linear regression between radiative modifications and drifting-snow fluxes. Regression functions and statistics are displayed on the corresponding panels. SWD modifications are computed when MAR-nDR simulates SWD > 50* $W m^{-2}$*. Data are filtered according to Sect. 2.4.*

Modifications in (b) LWD and (d) SWD between MAR-DR and MAR-nDR during drifting snow (drifting-snow flux > $10^{3} kg\,m^{-2} s^{-1}$*), as a function of drifting-snow layer height. The colorbar indicates the mass of snow contained between the drifting-snow layer height and the surface. Mean values are calculated for the lowest 9 model vertical levels and are represented by a grey mark. SWD modifications are computed when MAR-nDR simulates SWD > 50* $W m^{-2}$*. Data are filtered according to Sect. 2.4.*



(a)
Height a.g.l. [m]
MAR-DR
MAR-nDR
Temperature
[K]
(b)
MAR-DR - MAR-nDR
12.0 m
Temperature
[K]
(c)
MAR-DR
MAR-nDR
Relative humidity
[%]
(d)
MAR-DR - MAR-nDR
12.0 m
Relative humidity
[%]
(e)
MAR-DR
MAR-nDR
Specific humidity
[10^{-3} $kg\,kg^{-1}$]
(f)
MAR-DR - MAR-nDR
12.0 m
Specific humidity
[10^{-3} $kg\,kg^{-1}$]
(g)
MAR-DR
MAR-nDR
Wind speed
[$m\,s^{-1}$]
(h)
MAR-DR - MAR-nDR
12.0 m
Wind speed
[$m\,s^{-1}$]
(i)
MAR-DR - MAR-nDR
Mean snow particle ratio
[10^{-3} $kg\,kg^{-1}$]
(j)
MAR-DR - MAR-nDR
12.0 m
Total sublimation
[10^{-3} $kg\,kg^{-1}\,year^{-1}$]
(k)
MAR-DR - MAR-nDR
12.0 m
Mean LWnet + SWnet
[$W\,m^{-2}$]

Figure 6. *Annual mean (2010-2018) vertical profiles for near-surface and surface variables calculated at D17 on the lowest 12 vertical levels as simulated by MAR-DR, MAR-nDR or corresponding differences between both runs. In (i), sublimation rates are expressed in g of sublimated snow per kg of moist air per year.*

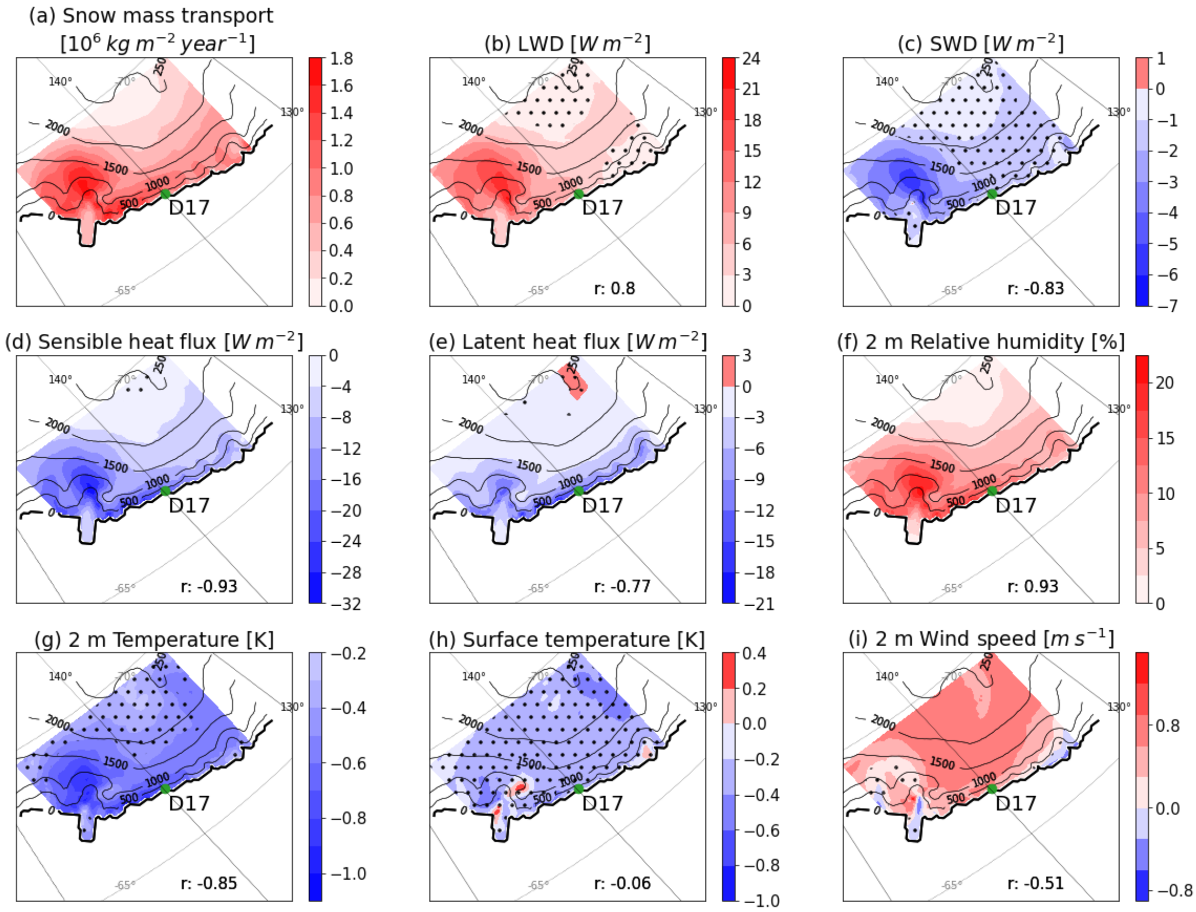


Figure 8 *(Figure 7 in the original manuscript) Annual mean (2010-2018) near-surface and surface variables modifications between MAR-DR and MAR-nDR over the integration domain. Within each panel, r indicates the Pearson correlation coefficient between the snow mass transport anomaly (a) and the considered variable (b to i). Dotted area designate areas where modifications are lower than interannual variability (taken as the standard deviation computed from annual means).*

Table 2. Root mean square error (RMSE), Pearson correlation coefficient and mean bias computed at D17 for MAR-DR and MAR-nDR half-hourly simulations in comparison with in situ observations.

	r		RMSE		Mean bias	
	MAR-DR	MAR-nDR	MAR-DR	MAR-nDR	MAR-DR	MAR-nDR
LWD [Wm^{-2}]	0.87	0.89	19.9	22.8	-14.9	-20.4
LWU [Wm^{-2}]	0.97	0.98	6.5	5.6	-4.0	-2.9
SWD [Wm^{-2}]	0.98	0.98	24.6	24.2	-1.3	0.3
SWU [Wm^{-2}]	0.98	0.98	22.4	22.0	-7.0	-5.9
Surface temperature [K]	0.97	0.98	1.7	1.4	-1.0	-0.7
2 m temperature [K]	0.97	0.98	1.3	1.2	-0.2	0.5
2 m wind speed []	0.78	0.82	3.0	2.5	2.3	1.7
2 m relative humidity [%]	0.62	0.51	9.5	15.8	-0.7	-14.0

Table 3. Half-hourly mean value and standard deviation (STD) for several near-surface and surface meteorological variables computed on Adelie Land with MAR-DR and MAR-nDR. Differences between both model runs are attributed to drifting-snow processes.

	MAR-DR		MAR-nDR		|MAR-DR| - |MAR-nDR|	
	Mean value	STD	Mean value	STD	Mean value	STD
LWD [Wm^{-2}]	162.4	22.1	156.3	20.8	6.1	1.3
LWU [Wm^{-2}]	-205.3	22.2	-206.2	22.2	-0.9	0
SWD [Wm^{-2}]	142.6	3.4	144.4	2.8	-1.8	0.6
SWU [Wm^{-2}]	-115.4	2.8	-116.0	2.4	-0.6	0.4
LHF [Wm^{-2}]	-2.7	3.0	-5.8	6.2	-3.1	-3.2
SHF [Wm^{-2}]	18.1	4.3	27.0	8.9	-8.9	-4.6
LWnet + SWnet + LHF + SHF [Wm^{-2}]	-0.3	0.3	-0.2	0.3	0.1	0
Surface temperature [K]	244.6	6.7	244.9	6.7	-0.3	0
2 m temperature [K]	245.5	6.7	246.0	6.8	-0.5	-0.1
2 m wind speed [$m\,s^{-1}$]	10.6	1.8	10.1	1.8	0.5	0
2 m relative humidity [%]	92.1	3.2	84.9	6.2	7.2	3.0

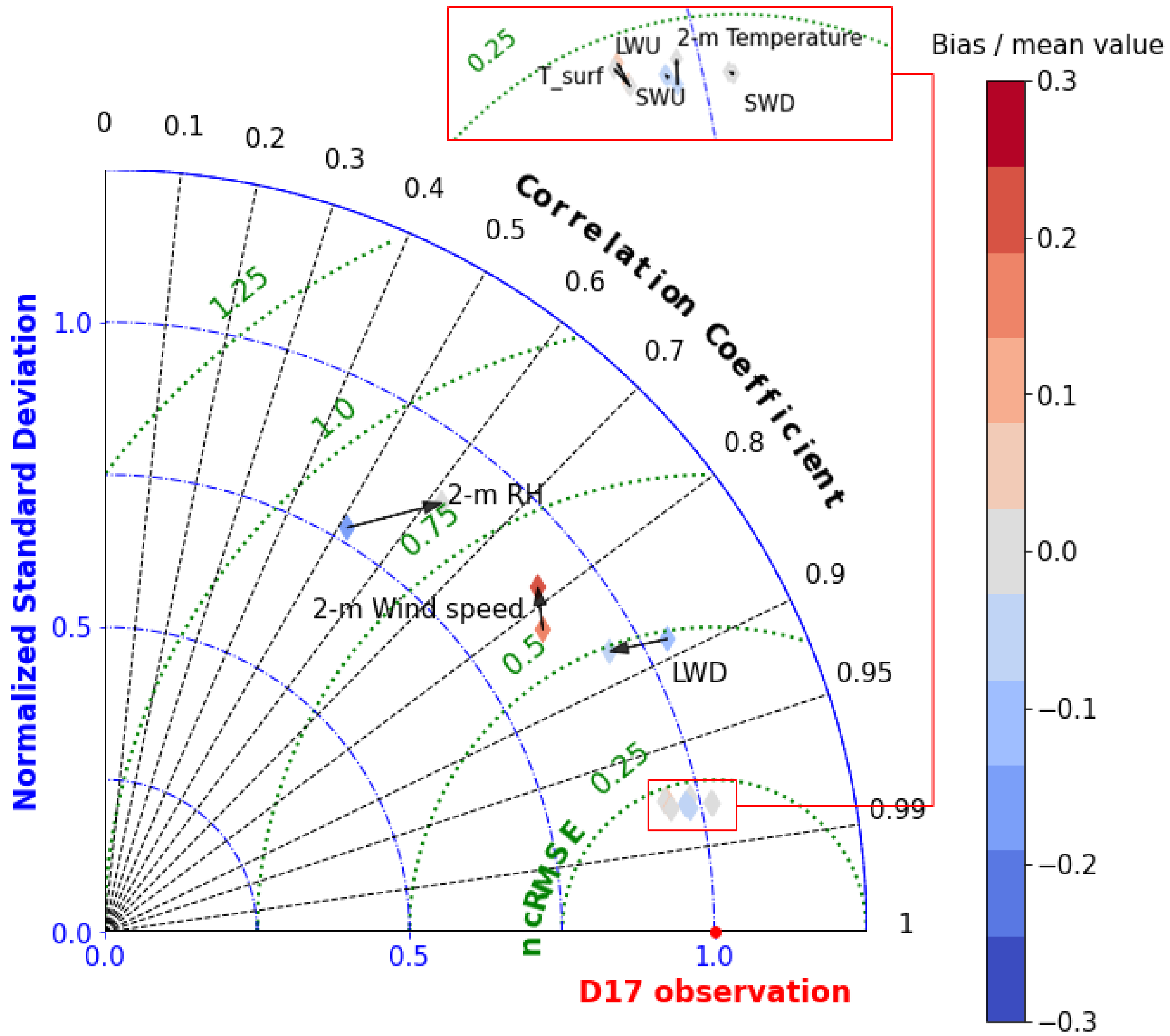


Figure S1. *Taylor diagram at D17 enables visualization of modifications between simulations, using observations as a reference. The radial distance from the origin accounts for the normalized standard deviation (standard deviation of the simulated variable divided by the observed standard deviation). Correlation coefficient is represented by the angular distance from the horizontal. Normalized and centered root mean squared error (ncRMSE) is represented by a green circle centered on the red point. A simulation matching perfectly observations would stand on the red point. The colorbar indicates the mean bias divided by the mean value of observations. The arrows point from MAR-nDR simulations to MAR-DR simulations. 2 m RH designates 2 m relative humidity and T_surf designates surface temperature. Surface temperature is computed using LWD and LWU.*

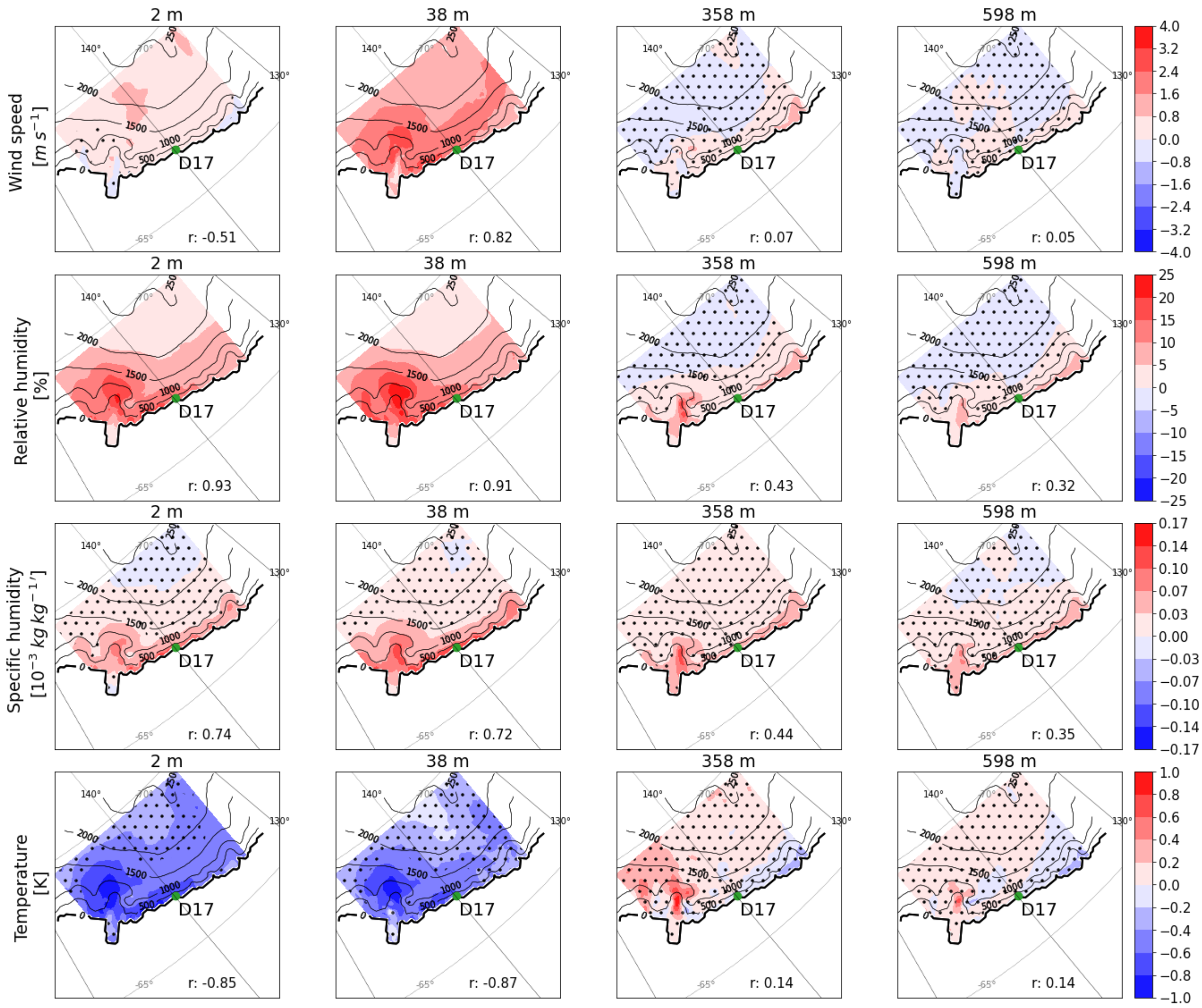


Figure S2. *Annual mean (2010-2018) near-surface and surface variables modifications between MAR-DR and MAR-nDR over the integration domain at different vertical levels in the low atmosphere.. Within each panel, r indicates the Pearson correlation coefficient between the snow mass transport anomaly (a) and the considered variable (b to i). Dotted area designate areas where modifications are lower than interannual variability (taken as the standard deviation computed from annual means).*

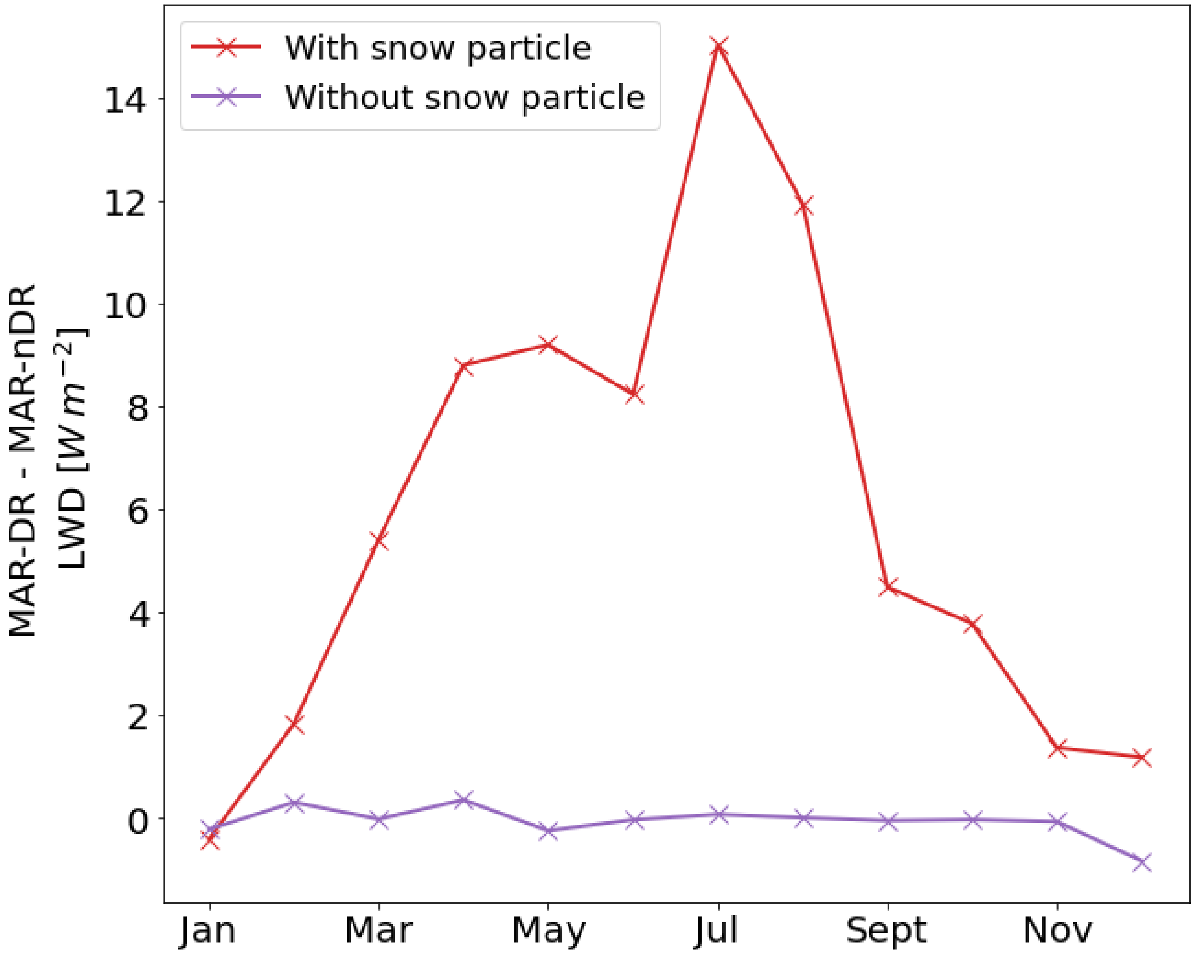


Figure S3. *Time series of LWD modifications between MAR-DR and MAR-nDR at D17 for the year 2017. Two model configurations are compared here: the reference simulation, referred as "With snow particle" includes the snow particle ratio in the LWD computation, oppositely to the "Without snow particle" configuration. LWD modifications between MAR-DR and MAR-nDR are highly influenced by the presence of snow particles in the atmosphere.*

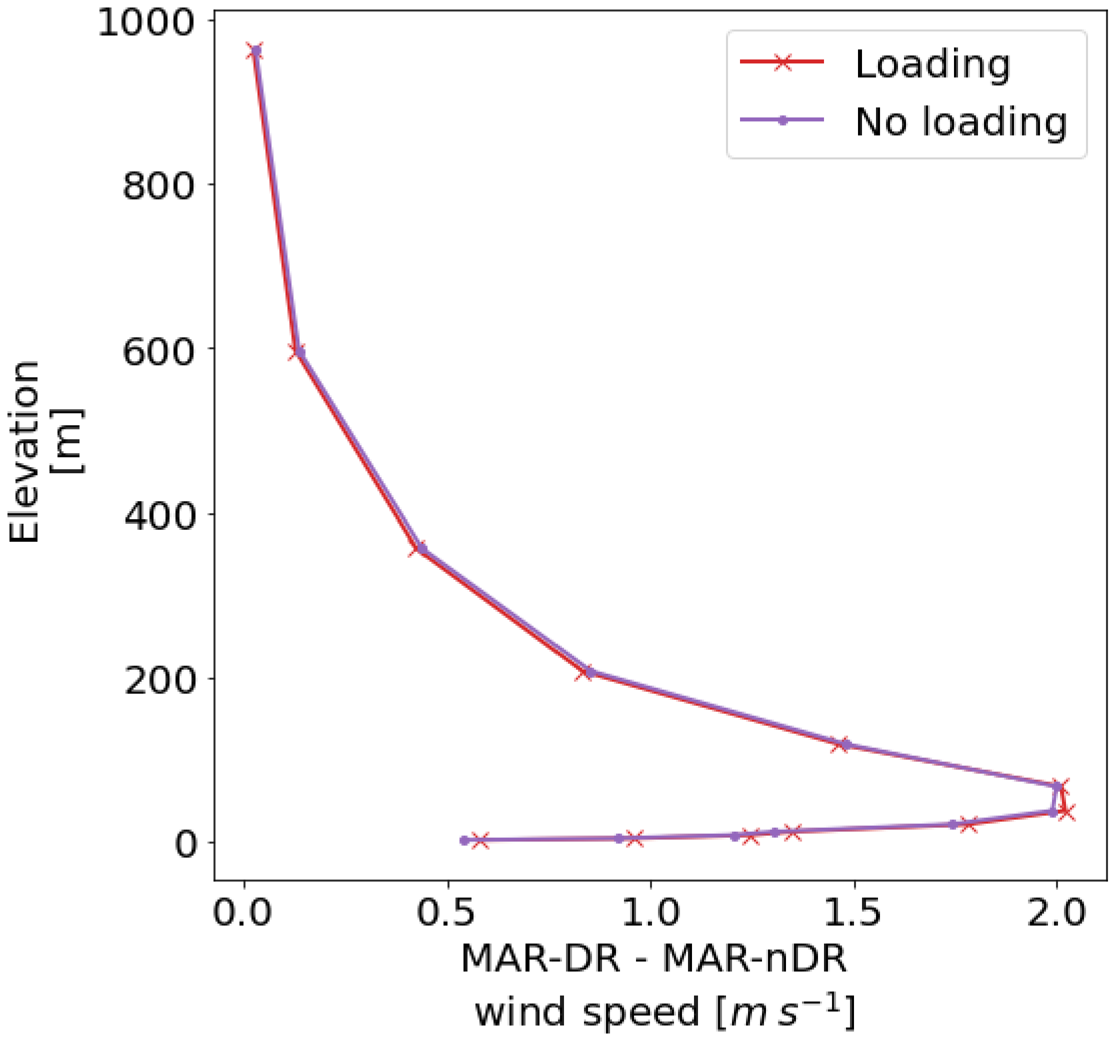


Figure S4. *Vertical profile of wind speed modifications between MAR-DR and MAR-nDR at D17 for the year 2017. Two model configurations are compared here: the reference simulation, referred as "Loading" includes the contribution of snow particles to air density, oppositely to the "No loading" configuration. The mass of snow particles is only responsible for limited wind speed modifications when the drifting-snow scheme is activated in MAR.*

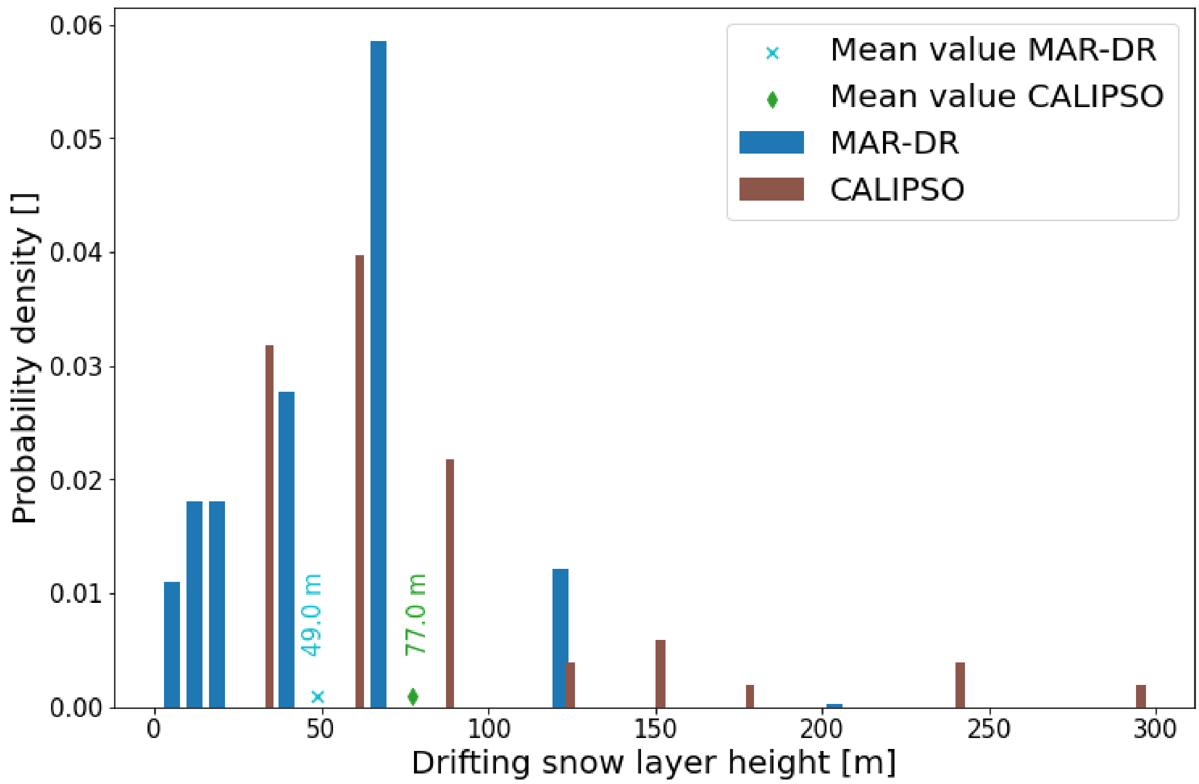


Figure S5. *Distribution of drifting-snow layer heights as simulated by MAR-DR (after the filtering process described in Sect. 2.4) and as observed by CALIPSO (Palm et al., 2011). The MAR-DR algorithm for detecting drifting-snow layer height is calibrated on CALIPSO observations. On the 2010–2018 period and after the filtering process, MAR-DR simulates a mean drifting-snow layer height of 49 m while CALIPSO detects for specific occurrences a mean value of 77 m*

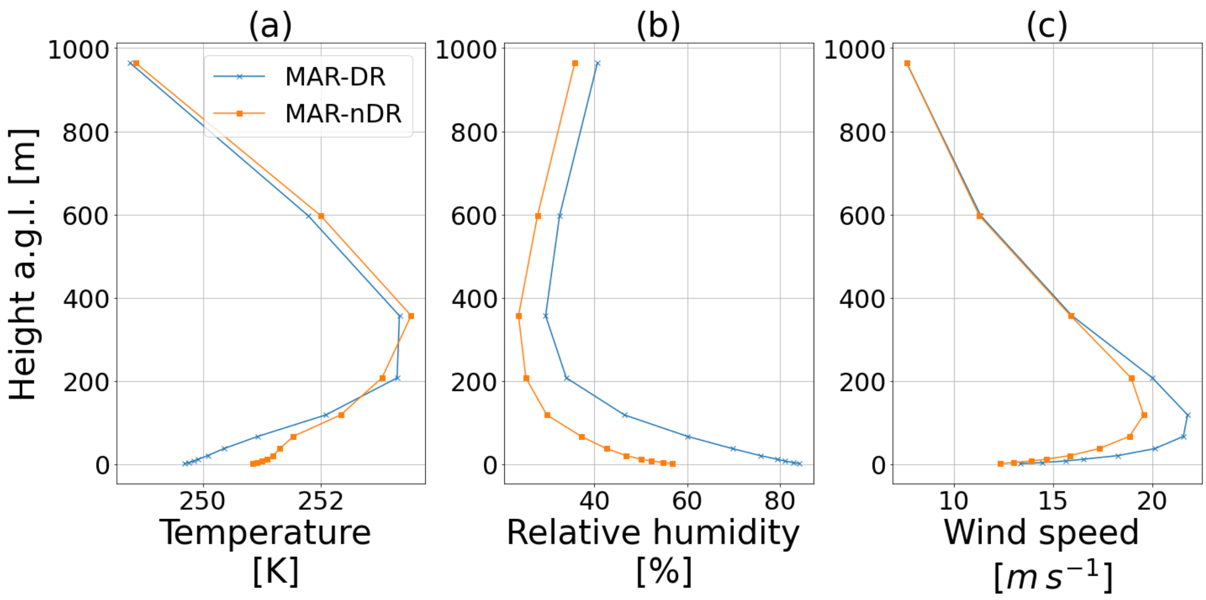


Figure S6. *(a) Temperature, (b) relative humidity and (c) wind speed mean profiles calculated during a drifting-snow event between the 1st and the 3rd of October 2017 at D17. Temperature variations are small in the katabatic layer (0.017 K m^{-1} on the first 100 m a.g.l.), relative humidity peaks near the surface, and wind speed increases with elevation in the katabatic layer*

Bibliography

Amory, C., Kittel, C., Le Toumelin, L., Agosta, C., Delhasse, A., Favier, V., & Fettweis, X. (2020). Performance of MAR (v3. 11) in simulating the drifting-snow climate and surface mass balance of Adelie Land, East Antarctica. Geoscientific Model Development Discussions, 1-35.

Bintanja, R. (1998). The contribution of snowdrift sublimation to the surface mass balance of Antarctica. Annals of Glaciology, 27, 251-259.

Bintanja, R. (2000). Snowdrift suspension and atmospheric turbulence. Part I: Theoretical background and model description. Boundary-layer meteorology, 95(3), 343-368.

Bintanja, R. (2001). Snowdrift sublimation in a katabatic wind region of the Antarctic ice sheet. Journal of Applied Meteorology, 40(11), 1952-1966.

Bintanja, R., & Reijmer, C. H. (2001). A simple parameterization for snowdrift sublimation over Antarctic snow surfaces. Journal of Geophysical Research: Atmospheres, 106(D23), 31739-31748.

Brun, E., David, P., Sudul, M., & Brunot, G. (1992). A numerical model to simulate snow-cover stratigraphy for operational avalanche forecasting. Journal of Glaciology, 38(128), 13-22.

Delhasse, A., Kittel, C., Amory, C., Hofer, S., As, D. V., S Fausto, R., & Fettweis, X. (2020). Brief communication: Evaluation of the near-surface climate in ERA5 over the Greenland Ice Sheet. The Cryosphere, 14(3), 957-965.

De Ridder, K., & Gallée, H. (1998). Land surface–induced regional climate change in southern Israel. Journal of applied meteorology, 37(11), 1470-1485.

Gallée, H. (1995). Simulation of the mesocyclonic activity in the Ross Sea, Antarctica. Monthly Weather Review, 123(7), 2051-2069.

Gallée, H., & Duynkerke, P. G. (1997). Air‐snow interactions and the surface energy and mass balance over the melting zone of west Greenland during the Greenland Ice Margin Experiment. Journal of Geophysical Research: Atmospheres, 102(D12), 13813-13824.

Gallée, H., & Pettré, P. (1998). Dynamical constraints on katabatic wind cessation in Adélie Land, Antarctica. Journal of the atmospheric sciences, 55(10), 1755-1770.

Gallée, H., Guyomarc'h, G., & Brun, E. (2001). Impact of snow drift on the Antarctic ice sheet surface mass balance: possible sensitivity to snow-surface properties. Boundary-Layer Meteorology, 99(1), 1-19.

Gallée, H., & Gorodetskaya, I. V. (2010). Validation of a limited area model over Dome C, Antarctic Plateau, during winter. Climate dynamics, 34(1), 61.

King, J. C., Anderson, P. S., Smith, M. C., & Mobbs, S. D. (1996). The surface energy and mass balance at Halley, Antarctica during winter. Journal of Geophysical Research: Atmospheres, 101(D14), 19119-19128.

King, J. C., Anderson, P. S., & Mann, G. W. (2001). The seasonal cycle of sublimation at Halley, Antarctica. Journal of Glaciology, 47(156), 1-8.

Kodama, Y., Wendler, G., & Ishikawa, N. (1989). The diurnal variation of the boundary layer in summer in Adélie Land, eastern Antarctica. Journal of Applied Meteorology and Climatology, 28(1), 16-24.

Lenaerts, J. T. M., & Van den Broeke, M. R. (2012). Modeling drifting snow in Antarctica with a regional climate model: 2. Results. Journal of Geophysical Research: Atmospheres, 117(D5).

Lesins, G., Bourdages, L., Duck, T. J., Drummond, J. R., Eloranta, E. W., & Walden, V. P. (2009). Large surface radiative forcing from topographic blowing snow residuals measured in the High Arctic at Eureka. Atmospheric Chemistry and Physics, 9(6), 1847-1862.

Lin, Y. L., Farley, R. D., & Orville, H. D. (1983). Bulk parameterization of the snow field in a cloud model. Journal of Applied Meteorology and climatology, 22(6), 1065-1092.

Locatelli, J. D., & Hobbs, P. V. (1974). Fall speeds and masses of solid precipitation particles. Journal of Geophysical Research, 79(15), 2185-2197.

Luo, L., Zhang, J., Hock, R., & Yao, Y. (2021). Case Study of Blowing Snow Impacts on the Antarctic Peninsula Lower Atmosphere and Surface Simulated With a Snow/Ice Enhanced WRF Model. Journal of Geophysical Research: Atmospheres, 126(2), e2020JD033936.

Mahesh, A., Eager, R., Campbell, J. R., & Spinhirne, J. D. (2003). Observations of blowing snow at the South Pole. Journal of Geophysical Research: Atmospheres, 108(D22).

Mahrt, L. (1982). Momentum balance of gravity flows. Journal of Atmospheric Sciences, 39(12), 2701-2711.

Naaim-Bouvet, F., Guyomarc'H, G., Bellot, H., Durand, Y., Naaim, M., Vionnet, V., ... & Prokop, A. (2013, October). Lac Blanc Pass: a natural wind-tunnel for studying drifting snow at 2700ma. sl. In International Snow Science Workshop (ISSW) (pp. p-1332). Irstea, ANENA, Meteo France.

Palm, S. P., Kayetha, V., Yang, Y., & Pauly, R. (2017). Blowing snow sublimation and transport over Antarctica from 11 years of CALIPSO observations. The Cryosphere, 11(6), 2555-2569.

Uppala, S. M., Kållberg, P. W., Simmons, A. J., Andrae, U., Bechtold, V. D. C., Fiorino, M., ... & Woollen, J. (2005). The ERA‐40 re‐analysis. Quarterly Journal of the Royal Meteorological Society: A journal of the atmospheric sciences, applied meteorology and physical oceanography, 131(612), 2961-3012.

Vignon, E., Genthon, C., Barral, H., Amory, C., Picard, G., Gallée, H., ... & Argentini, S. (2017). Momentum-and heat-flux parametrization at Dome C, Antarctica: A sensitivity study. Boundary-Layer Meteorology, 162(2), 341-367.

van Angelen, J. H., Van den Broeke, M. R., & Van de Berg, W. J. (2011). Momentum budget of the atmospheric boundary layer over the Greenland ice sheet and its surrounding seas. Journal of Geophysical Research: Atmospheres, 116(D10).

van den Broeke, M. R., Van Lipzig, N. P. M., & Van Meijgaard, E. (2002). Momentum budget of the East Antarctic atmospheric boundary layer: Results of a regional climate model. Journal of the Atmospheric Sciences, 59(21), 3117-3129.

van den Broeke, M. R., & Van Lipzig, N. P. M. (2003). Factors controlling the near-surface wind field in Antarctica. Monthly Weather Review, 131(4), 733-743.

Yamanouchi, T., & Kawaguchi, S. (1984). Longwave radiation balance under a strong surface inversion in the katabatic wind zone, Antarctica. Journal of Geophysical Research: Atmospheres, 89(D7), 11771-11778.

Yang, Y., Palm, S. P., Marshak, A., Wu, D. L., Yu, H., & Fu, Q. (2014). First satellite‐detected perturbations of outgoing longwave radiation associated with blowing snow events over Antarctica. Geophysical Research Letters, 41(2), 730-735.