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ARTICLE

Transport of Snow by the Wind: A Comparison Between Observations in Adélie Land, Antarctica, and Simulations Made with the Regional Climate Model MAR

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Abstract For the first time a simulation of blowing snow events was validated in detail using one-month long observations (January 2010) made in Adélie Land, Antarctica. A regional climate model featuring a coupled atmosphere/blowing snow/snowpack model is forced laterally by meteorological re-analyses. The vertical grid spacing was 2 m from 2 to 20 m above the surface and the horizontal grid spacing was 5 km. The simulation was validated by comparing the occurrence of blowing snow events and other meteorological parameters at two automatic weather stations. The Nash test allowed us to compute efficiencies of the simulation. The regional climate model simulated the observed wind speed with a positive efficiency (0.69). Wind speeds higher than 12 m s^{-1} were underestimated. Positive efficiency of the simulated wind speed was a prerequisite for validating the blowing snow model. Temperatures were simulated with a slightly negative efficiency (−0.16) due to overestimation of the amplitude of the diurnal cycle during one week, probably because the cloud cover was underestimated at that location during the period concerned. Snowfall events were correctly simulated by our model, as confirmed by field reports. Because observations suggested that our instrument (an acoustic sounder) tends to overestimate the blowing snow flux, data were not sufficiently accurate to allow the complete validation of snow drift values. However, the simulation of blowing snow occurrence was in good agreement with the observations made during the first 20 days of January 2010, despite the fact that the blowing snow flux may be underestimated by the regional climate model during pure blowing snow events. We found that blowing snow occurs in Adélie Land only when the 30-min wind speed value at 2 m a.g.l. is >10 m s^{-1}. The validation for the last 10 days of January 2010 was less satisfactory because of complications introduced by surface melting and refreezing.

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Keywords Antarctica · Blowing snow · Regional climate model · Surface mass balance

1 Introduction

The surface mass balance (SMB) of the Antarctic ice sheet is probably the only important negative contribution to sea-level rise (Solomon et al. 2007). Its main components are snow precipitation, snow net erosion/deposition by the wind, and surface sublimation of snow or ice. Runoff is caused by snow/ice melting but, except for the Antarctic peninsula, its influence on the Antarctic ice sheet can be considered as negligible. Net erosion of snow by the wind may contribute significantly to the SMB of the Antarctic coastal zone. Up to 35 % of snow precipitation may be removed by the wind and deposited elsewhere or sublimated in coastal areas (Bromwich 1988), and to an even greater extent over blue ice. This may be due to the fact that the surface winds along the Antarctic coast are among the strongest observed at sea level worldwide (Parish 1988).

Here we define the margin of the Antarctic ice sheet as that part of the ice sheet whose surface elevations are lower than 1,000 m above sea level (a.s.l.). This area can contribute roughly 50 % of the SMB and as the winds there are strong, accurate assessment of snow erosion by the wind is needed. This can be achieved either by detailed observations or by using specially designed climate models. Observations of blowing snow in Antarctica are rare for logistical reasons (Budd et al. 1965; Takahashi 1985; Mellor and Fellers 1986; Mann et al. 2000; Bintanja 2001; Scarchilli et al. 2010; König-Langlo and Loose 2007). A carefully validated model plus available observations is thus a good candidate for such an exercise. The present work is motivated by the fact that previous modelling studies pointed to the need for observations to validate parametrization of the blowing snow process in the extreme Antarctic environment (e.g., Gallée et al. 2001).

The combination of atmospheric dynamics, blowing snow physics and the surface of the sloping coastal area of the Antarctic ice sheet may be influenced by several physical feedbacks. Kodama et al. (1985) discussed three types of feedback (among others): (i) surface roughness, which reduces the wind speed and the resulting erosion of the snow by the wind, (ii) an increase in air density due to the presence of airborne snow particles, and (iii) an increase in air density by cooling due to the sublimation of airborne snow particles.

First, it should be noted that the erosion of snow by the wind is responsible for changes in surface roughness through the formation of obstacles such as sastrugis, barcans, and megadunes. In turn, surface roughness is responsible for a decrease in the kinetic energy available for snow erosion, and is a negative feedback. On the other hand, the ejection of snow particles from the surface may be facilitated by the saltation process (Male 1980). It should also be noted that megadunes can generate gravity waves that modulate the surface wind speed (Frezzotti et al. 2002).

Second, the increase in air density is responsible for an increase in the along-slope pressure gradient force, and is a positive feedback in katabatic flows. A preliminary discussion of the impact of this feedback in a limited area atmospheric model is given in Gallée (1998). Such an impact is only significant with very strong winds (wind speeds >28 m s^{-1}, Gosink 1989).

Wamser and Lykossov (1995) and Bintanja (1998) identified the influence of airborne snow particles on the vertical stability of the lower atmosphere. Indeed an increase in the concentration of airborne snow particles near the surface may decrease turbulence in the surface boundary layer (SBL). This process contributes negatively to snow erosion.

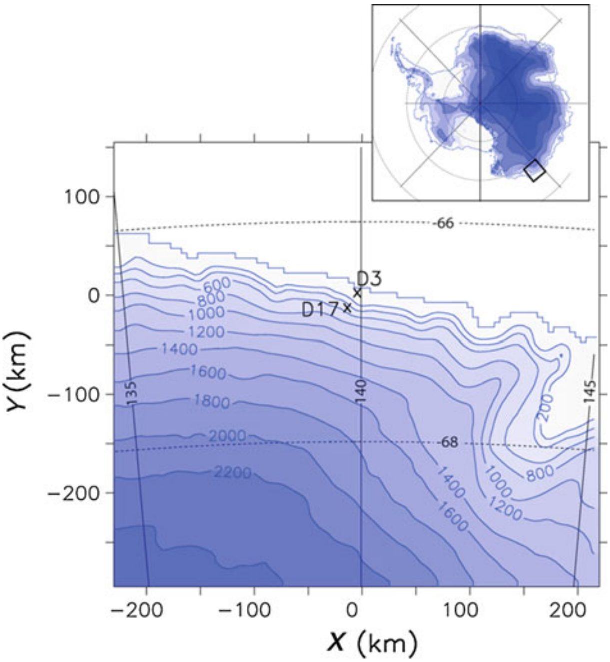

Fig. 1 MAR integration domain. Isocontours in the bottom panel are at 200-m intervals

Blowing snow may be very difficult to model because of its highly non-linear behaviour, and this is all the more true when case events are considered, in particular their timing and intensity. In turn, detailed modelling may provide physical insight into the relevant processes and possible changes in their respective importance. Such knowledge is crucial in the context of climate changes.

The objective of the work reported herein is thus to increase our understanding of the coupling of snow accumulation processes and atmospheric dynamics over the surface of a huge ice sheet. This type of coupling is strongly influenced by the wind erosion of snow, at least over the margin of the ice sheet. Our work will enable the development of simpler coupling schemes in the future, and ultimately their use in SMB studies with regional climate models (RCMs) or general circulation models (GCMs).

The *Modèle Atmosphérique Régional* (MAR) is a coupled atmosphere—blowing snow—snowpack model. It has already been used to study blowing snow in the Antarctic by analyzing the sensitivity of snow net erosion to the representation of snow properties (dendricity, sphericity, size and density, Gallée et al. 2001). A long-term simulation of the SMB of Wilkes Land has also been performed, and showed that results were improved when the net erosion of snow by the wind was taken into account (Gallée et al. 2005).

Adélie Land (Fig. 1) is probably an ideal location for observations of blowing snow, as the terrain is relatively homogeneous, allowing the development of the katabatic wind system and associated turbulence at relatively large horizontal scales. Adélie Land is also strongly influenced by synoptic weather systems. Together, these represent an almost complete sample of meteorological situations typical of the Antarctic ice sheet margin. In addition, atmospheric soundings are recorded daily at the French Dumont d'Urville station in Adélie Land, and are included in the meteorological analyses. As a result, good quality meteorological analyses can be generated, providing a good forcing for the RCMs used to validate parametrizations of blowing snow from observations made in the area. Finally our RCM MAR has already been validated over Adélie Land (Gallée et al. 1996; Gallée and Pettré 1998).

The observations of blowing snow events in this study were made in Adélie Land in January 2010. Our aim here is to validate the simulation of the atmospheric flow that drives blowing snow events and then their timing. Blowing snow erosion by the wind is a highly non-linear process, so it is important to simulate the timing of its occurrence for the right reasons. This is why we use our RCM here with a high horizontal and vertical resolution. Nevertheless the issues raised concerning the accuracy of the Ingénierie Acoustique & Vibratoire (IAV) FlowCapt acoustic sensors (Chritin et al. 1999) used in the present study have, up to now, prevented a reliable quantitative comparison of the blowing snow flux (Cierco et al. 2007; Naaim-Bouvet et al. 2010).

The paper is divided into five sections: Sect. 2 is a short description of the observations made, and a description of the coupled atmosphere—blowing snow—snowpack model is given in Sect. 3. The model is validated by comparing observations of wind speed, temperature, and the occurrence of airborne snow particles in Sect. 4. Conclusions are given in Sect. 5.

2 Observations

The observational dataset was obtained from a monitoring system deployed in Adélie Land, one of the windiest areas in Antarctica. Blowing snow was evaluated in the field in early 2009 using IAV FlowCapt acoustic sensors (Chritin et al. 1999). Although a longer record exists, we chose to concentrate on the data obtained in January 2010 when operators were present in the field who could check that all instruments were functioning correctly and could visually confirm the occurrence of meteorological events.

Issues have been raised about the accuracy and reliability of FlowCapt data in terms of snow fluxes. The calibration of FlowCapt results in overestimation of the blowing snow flux that is all the larger due to the high wind speed (Cierco et al. 2007; Naaim-Bouvet et al. 2010). As the winds responsible for the transport of blowing snow are very strong in Antarctica, considerable overestimation of the blowing snow flux can be expected. For this reason, we decided to check the model in terms of general meteorology and the occurrence of blowing snow events and to leave aside the evaluation of blowing snow fluxes. However, observed fluxes could provide qualitative information about the intensity of the events and new instruments are currently being tested to enable the evaluation of simulated blowing snow fluxes in the future.

2.1 Site Description

The experimental set-up is described in Table 1 and Fig. 2. The permanent French coastal Antarctic station named DDU (Dumont d'Urville) is located on an island (Ile des

Table 1 Characteristics of the observation stations

Station	Operating team	Latitude (S)	Longitude (E)	Distance from coast (km)	Altitude (m a.s.l.)
CP	LGGE AWS	66°41′	139°55′	0	30
D3	LGGE AWS	66°41′	139°54′	1	110
D10	AMRC AWS	66°42′	139°48′	5	246
D17	LGGE AWS	66°43′	139°43′	10	426

AWS automatic weather station, *ASS* automatic snow station

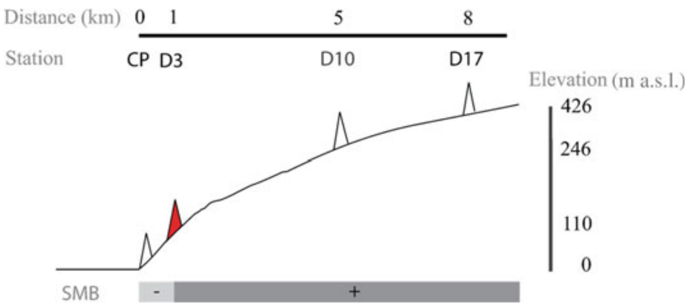

Fig. 2 Set-up of the observation stations. *White triangles* represent AWSs and the *red triangle* is the AWS + ASS

Pétrels) approximately 5 km from the ice sheet. A summer station, Cap Prudhomme (CP, 66°41′ S, 139°55′ E, 30 m a.s.l.), is operated seasonally on the coast on the ice sheet itself.

A long record of meteorological conditions is available at station DDU (e.g., König-Langlo et al. 1998). An automatic weather station (AWS) (wind, temperature, moisture, radiation) was set up on a rock outcrop at station CP in early 2005. Reports from these stations are described and used in Genthon et al. (2007) and Favier et al. (2011). An AWS was set up by the Antarctic Meteorological Research Center (AMRC) in 1984 at station D10 (see Table 1), about 5 km inland (Stearns and Wendler 1988; Colwell et al. 2010), and provides an additional long term record of local meteorological events. All these records show that strong katabatic winds occur in the area and can occasionally reach more than 50 m s^{-1} or more. As a result, drifting and blowing snow events are frequent. Stakes were set up in this area in 2004 and have been surveyed ever since (Agosta et al. 2011). They show that within an area of approximately 1 km from the sea (blue ice) the SMB is negative but becomes consistently positive further inland. Genthon et al. (2007) showed that the negative SMB on the coastal blue ice field is mostly due to erosion and export of snow by the wind. Altogether, coastal Adélie Land thus offers very appropriate conditions for detailed observations of blowing snow.

FlowCapt instruments were deployed in early 2009 at site D3, roughly 1 km from the coast (110 m a.s.l., see Table 1), along with an AWS (wind, temperature, moisture, snow surface height) thus becoming an automatic snow station (ASS). A full AWS was also deployed at site D17, 8 km inland (426 m a.s.l.), with the same characteristics as station D3 but without FlowCapt instrumentation during the period concerned. Although only 7 km apart, stations D3 and D17 are very distinct in terms of SMB (Agosta et al. 2011), with station D3 SMB being close to zero. The stations D3 and D17 provide consistent meteorological observations in terms of the types of instrument, the height of the instruments above the surface, sampling, and averaging. The Cap Prudhomme AWS is not located on the ice, and the station D10 uses different instruments and sampling. Here we only use data from stations D3 and D17 to scale our RCM with respect to the meteorology for this period, and only from station D3 for blowing snow. Measurements are made by all the AWS instruments at 10-s intervals and their 30-min averages are stored. The wind speed, direction and temperature are measured 2 m above ground level (a.g.l.). The FlowCapt consists of three vertical 1-m long tubes, each of which detects and tentatively measures the flux from the surface up to a height of 3 m a.g.l. Like the AWS, data are averaged on a 30-min basis.

3 Description of the Model

The coupled atmosphere/blowing snow/snow pack model is an improved version of the RCM MAR. An overview of MAR is given here, focused on the description of the blowing snow submodel and its coupling with the snowpack and atmospheric sub-models. A full description of atmospheric dynamics is given in Gallée and Schayes (1994). The original version of the snow and blowing snow sub-models is described in Gallée et al. (2001) and a preliminary long-term validation based on a comparison with snow stake measurements is described in Gallée et al. (2005).

MAR atmospheric dynamics are based on the hydrostatic approximation of the primitive equations. This is acceptable when the vertical extent of the circulation (here the katabatic flow) is much smaller than the size of the grid (here 5 km). Nevertheless, it should be noted that non-hydrostatic processes may be responsible for a weak deceleration of the katabatic flow (Cassano and Parish 2000). The vertical coordinate is the normalized pressure, with the model top situated at the 1-Pa pressure level. Parametrization of turbulence in the SBL takes into account the stabilization effect by the blowing snow flux, as in Gallée et al. (2001). Turbulence above the SBL is parametrized using the local $E - \varepsilon$ model of Bintanja (2000). In particular, it contains a parametrization of the turbulent transport of snow particles that is consistent with classical parametrizations of their sedimentation velocity. The influence of changes in the water phase on turbulence is included following Duynkerke and Driedonks (1987).

Prognostic equations are used to describe five water species, as in Gallée (1995): specific humidity, cloud droplets and ice crystals, raindrops and snow particles. A sixth equation has been added describing the number of ice crystals. Cloud microphysical parametrizations are based on the studies of Kessler (1969), Lin et al. (1983), Meyers et al. (1992) and Levkov et al. (1992), and the influence of hydrometeors on air specific mass is included in the model as in Gallée et al. (2001). This allows the influence of the weight of eroded particles on katabatic flow dynamics to be taken into account. Furthermore, sublimation of airborne snow particles is a direct contribution to the heat and moisture budget of the atmospheric layer in which these particles are simulated. This is important because blowing snow particles may reach a significant height above the surface (Mahesh et al. 2003; Scarchilli et al. 2010). Latent heat losses due to the sublimation of blowing snow particles are taken into account in the energy budget of the atmospheric layer in which the particles are found. Contrary to Lenaerts et al. (2010, 2012a,b), they are not included in the surface budget, because they could be responsible for the underestimation of surface temperature and subsequent underestimation of surface sublimation and spurious stabilization of the SBL.

The radiative transfer through the atmosphere is parametrized as in Morcrette (2002) and is the same as that used in ERA-40 re-analyses. As blowing snow particles are small (Walden et al. 2003), they may have an impact on the radiative transfer. In MAR, the influence of snow particles on atmospheric optical depth is included (Gallée and Gorodetskaya 2010).

In MAR, surface processes are modelled using the "soil-ice-snow-vegetation-atmosphere transfer" scheme (SISVAT, De Ridder and Gallée 1998; Gallée et al. 2001). In particular, the snow surface albedo depends on the snow properties (dendricity, sphericity and size of the snow particles). The influence of snow erosion / deposition on surface roughness (z_0) is taken into account by allowing the aerodynamic roughness length to increase linearly as a function of the wind speed 10 m a.g.l. (V_{10}), when $V_{10} > 6\,\mathrm{m\,s^{-1}}$. The time scale for sastrugi formation is assumed to be half a day, as suggested by Andreas (1995), and the asymptotic value of z_0 may increase linearly as a function of the wind speed $V(z_{0,lim} = 10\,\mathrm{mm}$ for $V = 20\,\mathrm{m\,s^{-1}}$; note that the friction velocity corresponding to $V = 20\,\mathrm{m\,s^{-1}}$ is generally slightly >1 m s^{-1}). z_0 is allowed to decrease when precipitation occurs with no erosion of the snow by the wind. Indeed the newly deposited snow progressively buries the sastrugi. Andreas et al. (2005, their Fig. 1) found values of z_0 ranging between approximately 10^{-4} and 100 mm, for friction velocities no greater than 0.6 m s^{-1}. The scatter is very high and is explained by the high dependency of z_0 on sastrugi history. Our parametrization includes that effect in a simple way, and is calibrated to obtain the best simulation of wind speed. Sastrugis contribute to surface roughness (sastrugi form drag) and hence to the loss of kinetic energy available for erosion. The contribution is represented by a decrease in the snow erosion flux and is parametrized as in Marticorena and Bergametti (1995). The increase in roughness length through the building of sastrugis is a negative feedback, and is not included e.g. in the study of Lenaerts et al. (2010, 2012a,b).

Densification of snow by the wind is included in SISVAT as in Gallée et al. (2001) with a slight modification. Snow-pack densification is allowed for erodible snow layers even deep within the snow pack to account for the redistribution of snow inside a grid cell. This is because our RCM simulates snow-pack behaviour averaged over a whole grid cell. In turn, an increase in the density of the surface snow pack is responsible for an increase in the friction velocity threshold before erosion. This is a negative feedback.

Unlike in previous versions of our model, the density of deposited blown snow particles is parametrized as a function of the wind speed, as in Kotlyakov (1961):

$$\rho = 104(V_{10} - 6)^{0.5} \tag{1}$$

where ρ is the snow density in kg m^{-3} and $V_{10} > 6\,\mathrm{m\,s^{-1}}$.

4 Results

Our RCM MAR was set up for Adélie Land with a horizontal resolution of 5 km (Fig. 1). This domain was chosen in order to include the katabatic wind system over the slopes of Adélie Land, which extend from the coast roughly 250 km inland. MAR was tested for several domain configurations including larger ones but the sensitivity of the simulated wind speed over our area of interest (station DDU—station D17 transect) was very small. Consequently we chose a small domain in order to perform simulations at an affordable numerical cost. Lateral forcing of MAR and sea surface conditions (sea-surface temperature and sea-ice fraction) are taken from ERA-Interim (Dee et al. 2011). The ratio of the ERA-Interim grid size to the MAR grid size is >10, which could have been a source of distortions over our area of interest. However this was not the case, as the simulated wind speed there is not sensitive to the size of the domain. It will be recalled that one objective of our study is to accurately simulate the forcing of blowing snow, namely wind speed, over the area of interest, so as to simulate the occurrence of blowing snow for the right reasons and consequently provide a realistic validation for the representation of such a process in our model.

There are 60 vertical levels in the atmospheric model with a high vertical resolution in the low troposphere. The spacing between the first 12 levels is 2 m and the first level is 2 m a.g.l. We realized that this vertical resolution would not allow us to represent the much higher concentration of blowing snow particles just above the saltation layer. In order to infer the consequence of our choice for the detection of blowing snow by the model, a preliminary 2-D simulation was performed with a much finer vertical resolution near the surface and with the lowest levels located 0.1, 0.2, 0.4 and 0.8 m a.g.l. We found that the concentration of blown snow particles 2 m a.g.l. was comparable to that in a simulation using the coarser vertical resolution of the present study. This made us confident in the validation of the occurrence of blowing snow in the model. The influence of blowing snow just above the surface will thus be included in the model in forthcoming studies.

The simulation was initialized once on 1 December 2009, with integration to two months. As snow surface temperatures depend on initialization at a longer time scale than is the scale for air temperatures, the simulation was started one month before the period of interest (January 2010). This allowed at least the snow layers near the surface to adjust to meteorological conditions. Note that, in our model, the snow pack near the surface is very rapidly renewed because snow precipitation and erosion are responsible for the formation of new layers and the disappearance of older ones. Furthermore as shown below (Fig. 4, in the sub-section on temperature analysis), the model does not have a temperature drift, because the adjustment period is already sufficiently long and also probably because of (i) the small size of the domain, and (ii) the good quality of the large-scale meteorological fields. The results are shown for the second month of the simulation run.

Model performances were assessed using the efficiency statistical test (E) proposed by Nash and Sutcliffe (1970):

$$E = 1 - RMSE^2/s^2 \tag{2}$$

where s and $RMSE$ are respectively the standard deviation of the observations and the root-mean-square error of the simulated variable. Note that $RMSE = 0$ implies $E = 1$. An efficiency index close to 1 means that comparing the simulated variable with the corresponding observation provides a lower $RMSE$ than that obtained when comparing it with its time average. A negative efficiency index means that the $RMSE$ is higher than the standard deviation of the observations. This then suggests that a detailed model would not improve the results compared to a simpler model providing an estimation of the variable averaged over the time period concerned. It is important to perform this statistical test for wind speed since a good simulation of it is a prerequisite for a good simulation of blowing snow.

In the remaining part of this section we analyze the wind speed, the temperature and the occurrence or absence of blowing snow. The analysis of temperature is needed because changes in snow erosion may occur when the melting point is reached. Wind direction is not analyzed since the validation of the simulated divergence of blowing snow fluxes is not the purpose of our study.

4.1 Wind Speed

The wind speed simulated at station D3 by the model is in good agreement with the observations reported in Fig. 3 and Table 2, as suggested by the high positive value of the efficiency. The timing of maximum winds is generally well simulated but their intensity during strong wind events is underestimated. The agreement is also good at station D17 (see Table 2). As expected, the good results obtained with the model can be at least partly explained by the fact that the area of interest (Adélie Land) is strongly influenced by synoptic weather systems and relatively large-scale topographic winds (both katabatic and anabatic—see, e.g., Gallée and Pettré 1998).

Note that the underestimation of the wind speed is not related to the choice of the size of the domain since no sensitivity was found to this choice. The underestimation is thus probably due to the turbulence scheme, which does not do a very good job in the case of high wind speed. In the real world, turbulent eddies have a large vertical extent during strong wind events and provide efficient downwards transfer of momentum. This is not the case with the $E - \varepsilon$ turbulence model. Possibly the use of a non-local turbulence scheme would improve this aspect of the simulation.

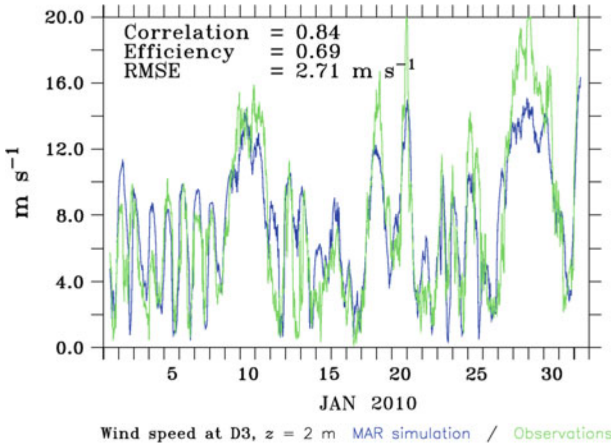

Fig. 3 Comparison between simulated and observed wind speeds (m s^{-1}) at station D3, 2 m a.g.l. Data are provided at 30-min intervals

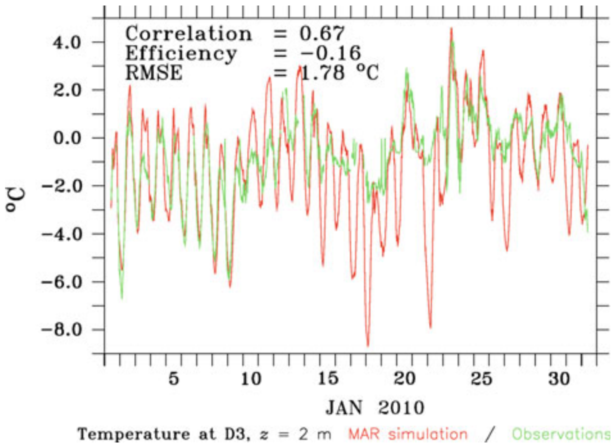

Fig. 4 Comparison between simulated and observed temperatures (°C) at station D3, 2 m a.g.l. Data are provided at 30-min intervals

Table 2 Correlation, efficiency and bias of the simulated wind speed at stations D3 and D17

	Station D3	Station D17
Correlation	0.84	0.86
Efficiency	0.69	0.72
Bias (m s^{-1})	−0.05	0.37

4.2 Temperature

The temperature at station D3 simulated by the RCM was compared with the observations in Fig. 4. Statistical properties of the simulated temperatures shown in Table 3 were computed

Table 3 Correlation and efficiency of simulated temperatures at stations D3 and D17

	Station D3	Station D3	Station D17	Station D17
		$\lvert DLW(\mathrm{MAR}) - DLW(\mathrm{OBS})\rvert < 90\,\mathrm{W\,m^{-2}}$		$\lvert DLW(\mathrm{MAR}) - DLW(\mathrm{OBS})\rvert < 90\,\mathrm{W\,m^{-2}}$
Correlation	0.67	0.78	0.68	0.82
Efficiency	−0.16	0.46	0.09	0.54
Bias (°C)	−0.52	−0.10	−0.30	0.32

after discarding values for which the observed wind speed was <2 m s^{-1} because spurious heating of the instruments has been observed in such cases (Genthon et al. 2011). Consequently only 1,373 and 1,374 measurements were included in the computations for stations D3 and D17 respectively (see columns 2 and 4 of Table 3).

The agreement is less good than for the wind speed at station D3. The poor efficiency of the simulated temperatures could be due to overestimation of the amplitude of the diurnal cycle by the RCM, especially between the 14 and 21 January. In fact there were marked differences between the simulated and observed downward longwave (DLW) fluxes during that period. A significant increase in the DLW fluxes was observed for the period concerned whereas this increase was not simulated. This suggests the presence of clouds during the period, which were not simulated by the RCM.

The statistical properties of the simulated temperatures were computed again but only using data for which $|DLW(\mathrm{MAR}) - DLW(\mathrm{OBS})| < 90\,\mathrm{W\,m^{-2}}$. In this case, only 1,017 and 773 measurements were included in the computations for stations D3 and D17 respectively. The agreement was then much better (see columns 3 and 5 of Table 3).

Note that the personnel on site confirmed that no snow was deposited on the radiative shield during the period concerned, probably because of the very high wind speed. It should also be noted that MAR also underestimated the amount of cloud cover at Dome C in summer, in a simulation domain covering the entire Antarctic Continent and a grid size (80 km) roughly equal to that used in ERA-Interim (see e.g., Gallée and Gorodetskaya 2010). The problem at Dome C persisted even when a finer grid size (20 km) was used.

4.3 Detection of Airborne Snow Particles

Here we make a preliminary validation of the blowing snow model by comparing the occurrence of blown hydrometeors in the observations and the simulation. The FlowCapts consist of a vertical assembly of three 1-m long vertical tubes. The lower 0.3 m of the bottom tube were buried in the snow pack in January 2010. So the 2 m a.g.l. line simulated by the RCM is located near the lower end of the third tube (1.70 m) and an event of blowing snow detected by this FlowCapt will correspond to the detection of an event in the simulation. A blowing snow event is considered to occur if the lowest FlowCapt gives a flux above a threshold value (0.001 g m^{-2} s^{-1}, which in those wind conditions, roughly corresponds to a concentration of hydrometeors of 0.0001 g kg^{-1}). This threshold value was obtained from a comparison between a Biral VPF730 disdrometer (Bellot et al. 2011) and the FlowCapt in January 2010. In Fig. 5, the simulated vertical profile of the concentration of hydrometeors is compared to the observed detection of hydrometeors at station D3.

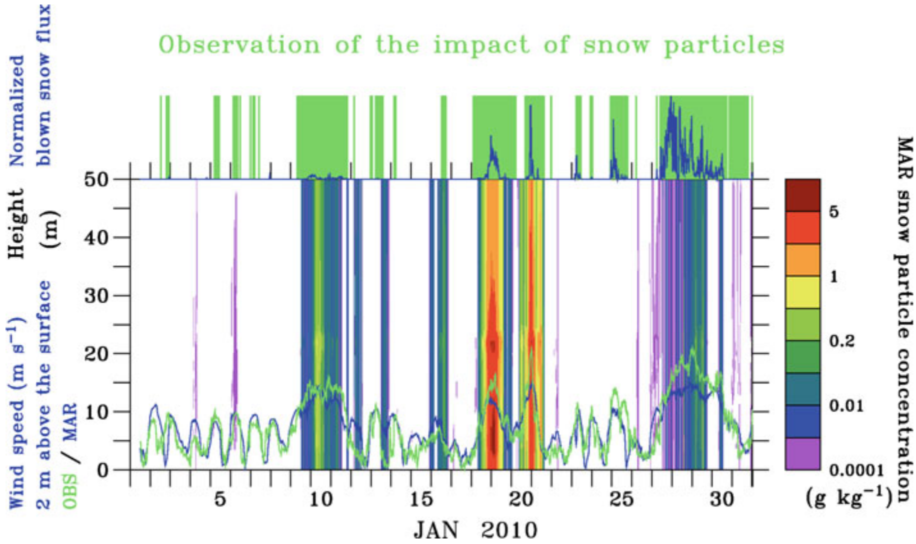

Fig. 5 Comparison of observed and simulated wind speed 2 m a.g.l. (*green and blue line*), the simulated concentration of atmospheric snow particles (*colour code*), and the observed detection of blown snow particles at station D3, with *green bars* indicating time periods when blown snow particles were detected by the two bottom tubes of the FlowCapt (detection threshold assumed to be 0.0001 $g\,m^{-2}s^{-1}$). The normalized value of the blowing snow flux measured by the bottom tube of the FlowCapt (*blue line in the top panel*) is also shown in order to provide a qualitative overview of the intensity of blowing snow events. Data are provided at 30-min intervals

Significant events (concentration of snow particles above 0.2 g kg^{-1}) are simulated for January 10, 18–21 and 28 respectively, events that correspond to strong winds and to a large observed blowing snow flux. Snow was also observed to fall during the first three events. The January 10 event was barely detected, and the field report states that a light snowfall occurred. The January 18, 19 event started with snowfall under light winds but the wind strengthened and snow drift started at 1,600 LT on 18 January. The model shows similar behaviour, as suggested by the larger simulated concentrations in the lowest part of the profile. The event on the 20, 21 of January, 2010 was a strong blowing snow event.

Two other observed events characterized by a large blowing snow flux were observed on 23 and 25 January but were not simulated by the model. The observed event which occurred on 27–30 January appears to be underestimated by the model compared to those which occurred on 18–21 January.

It was stated in the field report that no snowfall occurred after 21 January. Both observed and simulated temperatures rose above melting point from 21 to 25 January, after which they significantly decreased. Observations of the snow surface during the last 10 days of January reported the formation of a freezing crust on 22 January. The snow surface was then reported to be smooth before being progressively covered with sastrugis as blown snow particles were transported by the wind from the Antarctic interior towards the observation site.

This meteorological situation may explain why our RCM (MAR) did not simulate snow being transported by the wind on the 23 and 25 of January. Indeed MAR does not allow erosion of melt snow, even after it has frozen again, and this is a highly non-linear process that is difficult to simulate accurately. Furthermore the simulated concentration of blown snow increased with time during the event on 27 to 30 January whereas the observed concentration decreased. This suggests that the underestimation of the maximum wind speeds by MAR after 25 January may be responsible for the underestimation of snow transport by the wind and its deposition on the site, greatly reducing the availability of snow for new erosion at the site at the onset of the event.

A slightly less intense event was detected both in the simulation and in the field report on 16 January. It occurred under light winds and was considered by the observers to be a pure snowfall event. Finally it was found that blowing snow events in Adélie Land occur only when the wind speed 2 m a.g.l. is >10 m s^{-1}, and the simulation was in agreement with field reports when no surface melting occurred. This high wind threshold value is somewhat surprising but may be explained partly by the fact that we consider 30-min averages, so that a smaller instantaneous threshold is to be expected. Snowfall events were also simulated by MAR and confirmed in field reports.

5 Discussion and Conclusion

Our aim was to describe a strategy for studying the influence of blowing snow on the surface mass balance of a huge ice sheet. Adélie Land was chosen as the site for measurements of the physical processes responsible for variations in blowing snow and their consequences for the surface mass balance. A model was then developed to provide a detailed representation of the processes measured during field campaigns.

An improved version of the coupled atmosphere—blowing snow—snow-pack model was implemented. Apart from the positive feedbacks discussed by Kodama et al. (1985), it contains the representation of several significant physical processes: (i) the densification of the snow pack by the wind as a function of the wind speed and in turn the dependence of the friction velocity threshold before erosion on snow-pack properties, (ii) the influence of the erosion of snow by the wind on terrain roughness and in turn the influence of terrain roughness on erosion, (iii) the influence of snow erosion on the vertical stability of the surface layer and in turn the influence of this stability on the snow erosion flux. These physical processes are negative feedbacks whose representation is probably needed to control the strong non-linear behaviour of wind erosion of snow, especially when validation is performed by analyzing simulated case studies. The first step was to validate the consistency between the site and the model used to study that process. The set-up was obtained by choosing a small integration domain to capture the large-scale forcing as well as possible. A fine horizontal resolution was also chosen to represent the terrain topography as accurately as possible, and subsequently the katabatic flow forcing.

Results show a good behaviour of the model for wind speed, which is the most relevant forcing variable for the erosion of snow by the wind. Air temperatures at station D3 were less well simulated by the model, possibly due to a problem with the representation of clouds. On the other hand, air temperatures were better simulated at station D17.

As the use of FlowCapt acoustic sounders previously did not allow reliable quantification of the blowing snow flux, the main objective was to simulate realistic detection of blowing snow events. Results showed that the model performed well in January 2010. The difficulty with the sensors has recently been improved (V. Chritin, personal communication, 2012). Consequently, as new observations in Adélie Land become available, future work will include validation of the blowing snow flux simulated by our RCM. In this case, the simulated concentration of blowing snow particles well below 1 m a.g.l. will be considered. This will enable us to validate the simulated divergence of the blowing snow flux, in order to assess the impact of blowing snow on the surface mass balance. However, each individual component of flux divergence may be much larger than the divergence itself, so that the information on flux divergence may contain a high level of noise. Snow stake measurements will help confirm the validity of that information, but probably only for long time scales.

In short, our model is the first RCM to include so many coupling processes between a snow-pack model, a blowing snow model and an atmospheric model. To our knowledge, this is the first time the simulated detection of blowing snow events in a RCM has been adequately compared with long-term continuous observations, as up to now, it has only been possible up to detect the occurrence or absence of blowing snow events in field observations. Our model thus provides a tool for validating much simpler coupling strategies in GCMs and/or other RCMs.

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