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Event-driven deposition: a new paradigm for snow-cover modelling in Antarctica based on surface measurements

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Abstract

Antarctic surface snow is studied by means of continuous measurements and observations over a period of 3 yr at Dome C. Snow observations include precipitation, daily records of deposition and erosion, snow temperatures at several depths, and snow profiles. Together with meteorological data from automatic weather stations, this forms a unique and complete dataset of snow conditions on the Antarctic Plateau. Large differences in snow amounts and density exist between precipitation measured 1 m above the surface and deposition on the surface. We then used the snow-cover model SNOWPACK to simulate the snow-cover evolution for different deposition parameterizations. The main adaptation of the model described here is a new event-driven accumulation scheme. The scheme assumes that snow is added to the snow cover permanently only for periods of strong winds. This assumption followed from the comparison between precipitation observations and daily records of changes in snow height, which showed that over a period of 235 days there was precipitation on 40 % and deposition on 25 % of the days, but precipitation accompanied by deposition on 14 % of the days only. This confirms that precipitation is not necessarily the driving force behind snow height changes. A comparison of simulated snow height to stake farm measurements over 3 yr showed that we underestimate the total accumulation by about 64 %, when the total snow deposition is constrained by the precipitation measurements. This is because the precipitation measured above the surface and used to drive the model, even though comparable to ECMWF forecasts in its total magnitude, should be seen as a lower boundary of accumulation. As a result of the new deposition mechanism, we found a good agreement between model results and measurements of snow temperatures and recorded snow profiles. In spite of the underestimated accumulation, the results strongly suggest that we can obtain quite realistic simulations of the Antarctic snow cover by the introduction of event-driven snow accumulation.

1 Introduction

The upper meter of the snow cover on the Antarctic Plateau is exposed to extraordinary conditions. Extremely low air and snow surface temperatures (−81 to −12 °C and −80 to −19 °C, respectively, according to our observations between January 2005 and December 2009), low accumulation (39 $\text{kg}\,\text{m}^{-2}\,\text{a}^{-1}$, Frezzotti et al., 2005), and windy conditions (5-yr mean 3.1 $\text{m}\,\text{s}^{-1}$, maximum 16.2 $\text{m}\,\text{s}^{-1}$, our observations) influence the development of the snow cover. Snow is transported by the wind before it is permanently added to the underlying snow cover. This does not only affect the local surface mass balance but also the snow microstructure, as the properties of the snow particles are modified during drifting or blowing snow (e.g. Doorschot and Lehning, 2002). The effects of blowing snow on snow properties are most easily visible in the surface features they form, such as dunes and zastrugi. Watanabe (1978) describes the characteristics of surface features formed by wind and drifting snow, for example dunes, or the deposition-erosion processes at Mizuho Plateau. Doumani (1967) reports in detail on the formation of zastrugi and other erosional and depositional features. Yet another description of snow accumulation and the occurrence of dunes and zastrugi in a katabatic wind zone is given by Goodwin (1990). Although these observations are very valuable, they only qualitatively discuss the state of the surface snow after drifting snow has occurred. We miss a quantitative description and especially observations of the wind and the properties of the snow while transported. More recently, Walden et al. (2003) collected ice crystals from precipitation and deposited snow on an elevated platform on the roof of a building at South Pole Station, however, they do not describe the snow surface. Furthermore, Birnbaum et al. (2010) observed the formation of dunes at Kohnen station. They describe under what circumstances the dunes were formed and report on some snow properties such as the density and the snow particles. To our knowledge however, there are no studies that continuously and quantitatively describe the surface snow properties and drifting snow at one site over a longer period.

More quantitative descriptions of Antarctic surface snow were found in the following studies. Palais et al. (1982) made several 3 m snow profiles at Dome C and found a strong stratigraphic layering. Other than visible stratigraphy, gross β-radioactivity and microparticles were used to date the layers. However, snow properties such as density, grain and bond sizes as well as snow-cover temperatures were not measured. Gay et al. (2002) describe grain sizes at several locations in the Antarctic. They conclude that grain size is spatially homogeneous near the surface (0–0.5 m depth) and may be classified as very fine to fine (< 0.5 mm; see Fierz et al., 2009). Furthermore, they report that within one meter from the surface, the grain size at Dome C remains fine. These observations however, describe the snow cover at a given time and do not give us information on the evolution of the Antarctic surface snow. A study by Radok and Lile (1977) provides surface snow density measurements over 1 yr, but continuous studies of both surface snow and erosion are hardly available. These are, however, necessary to understand the evolution of the snow cover in a wind-prone environment.

Erosion processes are not only important for the surface snow characteristics; they also have an influence on the surface mass balance. Several studies indicate that local erosion can be on the same order of magnitude as annual precipitation (e.g. Petit et al., 1982). This, combined with a lack of seasonal melting layers, makes it difficult to date snow layers in snow profiles and to understand total accumulation. Despite this, many efforts have been made to estimate the surface mass balance (e.g. Frezzotti et al., 2007). Their research aimed to estimate snow accumulation from ice core drillings and stake measurements. They found that accuracy is low for low-accumulation sites. While accumulation is already hard to measure, it is even more difficult to distinguish between drifting snow and actual snow precipitation. Visual observations of the South Pole Weather Office distinguish diamond dust, blowing snow and snow grains. Often, however, blowing snow and diamond dust occur together (Walden et al., 2003). Furthermore, gauge measurements are problematic since drifting snow is blown into the gauges (Bromwich, 1988) and the snowfall amounts in this region do not exceed the minimum gauge resolution (Cullather et al., 1998). Combination of these facts decreases the accuracy of precipitation measurements and the estimations of the surface mass balance. This problem could be overcome, however, by continuously recording precipitation, height of snow and surface snow characteristics such as density and snow type at several locations as discussed in our contribution.

Several attempts to model Antarctic snow cover have preceded the current study. Morris et al. (1997) used DAISY, a physics-based snow model, to simulate the snow cover at Halley Bay. Initial new snow density was set to 300 or 400 $kg\,m^{-3}$ in different simulations. Dang et al. (1997) used a similar approach with the snow-cover model CROCUS, setting the initial new snow density to either 300 or 350 $kg\,m^{-3}$. In a further study with CROCUS, Brun et al. (1997) proposed yet another approach to account for new snow densification in cold and windy conditions. Their approach combines surface snow characteristics and threshold wind speeds, that is, drifting and blowing snow conditions to attain an effective compaction of surface snow down to about 10 centimetres below the surface. Compaction, however, does not necessarily happen at the time of deposition but can occur days or weeks later. Recently, Brun et al. (2011) modelled the snow cover at Dome C for 10 days in January 2010. The aim was to show that in this region a snow-cover model can be coupled successfully to an atmospheric model as long as the surface energy balance is correctly reproduced. Over such a short period of time the authors did not have to consider either settlement or accumulation though and the model reproduces quite well snow temperatures from the surface down to about 80 cm depth.

Our observations at Dome C (Fig. 1) show that the bulk density of snowfall can be very low as the average density of solid deposits on a table 1 m above the surface was approximately $83 \pm 43\,kg\,m^{-3}$ where the uncertainty here and henceforth refers to one standard deviation of the mean. On the other hand, we took 16 series of 8 measurements each of the top 10 cm snow density throughout one year; all measurements taken together averaged $357 \pm 50\,kg\,m^{-3}$, covering the range from 234 to 460 $kg\,m^{-3}$ (Fig. 1). Regarding the means of each series, they averaged $357 \pm 14\,kg\,m^{-3}$. These results suggest a considerable spatial variability but a minimal change in mean surface snow density in time. These observations agree well with observations at other high elevation locations in Antarctica. For example, during a traverse between Syowa Station and the South Pole, surface density was measured at several stakes and was never found to be less than 210 $kg\,m^{-3}$ (Fujiwara and Endo, 1971). A one year record of surface snow densities at Plateau Station showed a minimum (monthly mean) surface density of 279 $kg\,m^{-3}$ (Radok and Lile, 1977). Finally, measurements of snow density within the top meter of the snow cover performed during a recent Japanese-Swedish traverse further confirm these findings (Sugiyama et al., 2012). These authors report nearly constant values from Dome F (3800 m) to Kohnen (2890 m), ranging from 333 to 375 $kg\,m^{-3}$ and averaged 351 $kg\,m^{-3}$.

Here we present a 3 yr study of the snow cover at Dome C (Concordia) that includes meteorological observations, snow profiles, erosion studies, and numerical simulations. We use the snow-cover model SNOWPACK that has been extensively tested for Alpine regions (for example, Lehning and Fierz, 2008; Fierz and Lehning, 2001), but has not been previously applied in the Antarctic. We attempt to model the rapid densification of surface snow by a suitable parameterization depending on mean wind speed over a given period. Furthermore, care is taken to obtain realistic settling rates. We then compare model results with measurements like snow heights recorded at stake farms, emitted long wave radiation, continuous records of snow temperatures within the snowpack, and observed snow profiles.

We begin with a description of the data in Sect. 2. In Sect. 3 we briefly describe the adaptation of the model to polar deposition. In particular, a new parameterization for event-driven accumulation is introduced. Model results are compared with observations and discussed in Sect. 4, followed by conclusions.

2 Data

Concordia research station (75°06′ S, 123°24′ E, 3233 m a.s.l.) is located at Dome C on the East Antarctic Plateau. At Dome C the mean annual accumulation is about

39 kg m^{-2} a^{-1} inferred from stake measurements between 1969 and 1999, annual mean temperature is −53 °C and annual mean wind speed is 2.9 m s^{-1} (Petit et al., 1982; King and Turner, 1997; Frezzotti et al., 2005). The station is mainly used for research studies in geodesy, glaciology, human biology, seismology, astronomy, and for meteorological and geomagnetic observations as well as for environmental monitoring. Near the base there are two Automatic Weather Stations (AWS), that is, Dome C II and Concordia AWS, and a station of the Base Surface Radiation Network (BSRN) (see Fig. 2). Other sites are dedicated to snow temperature measurements and to observations of both deposition (tables above surface) and drifting snow (snow boards lying on the surface). In addition, accumulation is measured at two stake farms. Below we will only describe measurements of interest to our study that cover the period from 28 January 2005 through 1 March 2009.

2.1 Meteorological Records

Continuously recorded meteorological data at Dome C include air temperature and relative humidity, incoming shortwave and longwave radiation, and wind speed and direction. Air temperature and relative humidity were measured at 2 m height with a platinum resistance thermometer at Dome C II and a HUMICAP® at Concordia AWS, respectively. Wind speed and direction were taken from Dome C II, where a mechanic Young sensor is located at a height of about 3 m above the surface. Incoming long and shortwave radiation measurements were taken from the Dome C BSRN station, which is equipped with two normal incidence Kipp & Zonen CM21 pyranometers and a Kipp & Zonen CG4 Pyrgeometer, all operated according to BSRN guidelines (Lanconelli et al., 2011). In addition, upwelling long wave and reflected shortwave radiation was also made available by the Institute of Atmospheric Sciences and Climate of the Italian National Research Council (ISAC/CNR, V. Vitale). The above measurements were recorded at different time steps and intervals but we use hourly averages to drive the snow-cover model. Whenever necessary, data gaps were filled by interpolation.

2.2 Deposition and accumulation

Four methods of observing precipitation as well as deposition and erosion at Dome C provided data discussed in the current study. Figure 2 shows the location of the different measurement sites.

Simultaneous measurements of the depth and density of deposits, that is snowfall or hoar, on wooden tables located 1 m above the surface, have been set-up to determine the water equivalent of the solid precipitation by weighing the deposit. This was not always possible though and if necessary density was estimated based on the snow crystal forms, for which typical densities were inferred from our data set, from 44 $\mathrm{kg\,m^{-3}}$ for needles (PPnd) to 107 $\mathrm{kg\,m^{-3}}$ for small rounded particles (RGsr; for the abbreviations, see Fierz et al., 2009). These daily observations cover the entire study period and the mean density of over 200 measured deposits is $83 \pm 43\ \mathrm{kg\,m^{-3}}$.

Furthermore, two stake farms allow assessing accumulation on a larger scale around Dome C. The first farm located about 500 m away from the base but close to Concordia AWS, consists of 13 stakes arranged in two 60 m lines forming a cross. At this location observations were made weekly from January 2006 to March 2009. The other field, located about 3 km away from the base, next to Dome C II, includes 50 stakes placed 25 m apart from each other, also arranged in a cross. Due to the distance of the latter field from the base, those snow height measurements are available on a monthly basis only during the summer seasons of 2005 to 2008.

The two last methods involve measuring snow height on snow boards placed flush with the snow surface. To assess deposition at the surface, snow board $\mathrm{SB_{clear}}$ is cleared and repositioned flush with the surface snow daily. A second variant, not removing the accumulated snow but measuring daily changes of snow depth over snow board $\mathrm{SB_{acc}}$ reveal the importance of both erosion and deposition events. Observations of erosion and deposition events at the surface are available over the period running from November 2008 through June 2009.

2.3 Snow temperatures

Snow temperatures were monitored continuously by means of two strings of resistance temperature detectors (RTD, PT 100 DIN-A) placed into the snow cover at initial depths of 5, 10, 50, and 100 cm as well as 150, 200, 250, 300, 400, 500, 600, 800, and 1000 cm. The first set of four near surface measurements was acquired at a location 3 to 4 m away from a shelter, the second set 5 to 7 m away from the same shelter. Surface disturbances due to drifting and blowing snow around the shelter are known to occur but were not recorded. The upper four sensors were reinstalled in February 2006, February 2008, and December 2008 to minimize the effect of settlement and accumulation on their position relative to the surface.

2.4 Snow profiles

Several snow profiles were taken in the summers between December 2004 and December 2008. Snow profiles are records of the stratigraphy of the snow cover, that is, density, grain forms and sizes, temperature and hand hardness are usually observed layer by layer on the wall of an open pit. Most snow pits were about 1 m deep, but occasionally deeper pits were dug. We combined several profiles taken at different locations around Dome C from December 2004 to January 2005 as well as measurements from a 10 m deep shallow core to build an approximately 10 m deep initial profile for our simulation.

3 The snow-cover model SNOWPACK: adaptation to the Antarctic environment

SNOWPACK is a one-dimensional physical snow-cover model. Driven by standard meteorological observations, the model describes the stratigraphy, snow microstructure (metamorphism), temperature distribution, and settlement as well as surface energy exchange and mass balance of a seasonal snow cover. It has been extensively described in Lehning et al. (2002a,b) and Lehning and Fierz (2008). In this section we focus on the adaptations made to make the basic model (release 3.0.0) suitable for the Antarctic environment.

3.1 Event-driven snow accumulation

The snow cover at Dome C is highly influenced by single events (for example Palais et al., 1982). One cause is the relatively constant and light clear-sky precipitation at Dome C (about 20 % of precipitation according to our observations, not shown), which is easily transported by the wind. Single strong wind events have a large impact on snow properties and can erode more snow than the yearly precipitation. On the other hand, Birnbaum et al. (2010) observed near Kohnen that during drifting snow events dune formation would only occur after days when loose heavy particles were generated at the surface. This leads us to assume that longer periods of drifting snow alter the snow surface significantly and snow can only become immobile during such events. In other words, only snow that has already been repeatedly transported by the wind can be added to the snow cover during long enough drifting snow events, that is, usually not at the time of precipitation, and this gives the snow cover a strongly wind influenced stratigraphy. The amount of precipitation, however, can be retrieved from the measurements taken on the tables at 1 m above the surface (see Sect. 2.2) and cumulated between two events. That way the time at which the snow is added to the snow cover is different from the observation time but the total accumulation over the study period is left unchanged. This original mechanism allows for wind driven deposition under polar conditions and we describe it in more details below.

To estimate the wind speeds needed for such an event to occur, we compared the snow height measurements on the boards placed on the snow surface (see Sect. 2.2) with the wind speed records. We assumed that strong winds and a longer period of drifting snow would be necessary to alter the snow properties so much that the snow could be permanently added to the snow cover. Though the threshold wind speed for initiation of drifting snow depends on snow surface properties that vary in time, the data available for the present study focussed on accumulation, not on snow properties.

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

The observations on the snow boards revealed more important erosion and deposition events when the 1 h average wind speed we use as input was above 4 $\mathrm{m\,s^{-1}}$ for a day at least. This wind speed is well below an often assumed drifting snow threshold of 7 $\mathrm{m\,s^{-1}}$ at 10 m above the surface, (e.g. Pomeroy et al., 1993; Clifton et al., 2006), which is comparable to a wind speed of about 6 $\mathrm{m\,s^{-1}}$ at 3 m above the surface. However, while the initiation of drift may require high speeds, the maintenance of drifting snow does not (Mellor, 1965), and – maybe more importantly – this daily average implies high wind speeds during shorter periods. From this we assume an event to occur whenever the 100-h moving average of the wind speed measured at 3 m height, U_{event}, lies in the range of 4 to 7 $\mathrm{m\,s^{-1}}$. Such an event is assumed to include the gusts that could easily initiate the process. During our total observation period, U_{event} was within this range about 23 % of the time. Furthermore, note that the measured and reported wind speeds may be seen as lower bounds, as deposition of drifting snow on the wind sensor as well as icing may influence the measurements.

Besides the timing of the deposition, we also need to parameterize the new snow density, as the difference in density measured on the table and at the surface (Fig. 1) is too large to be overcome by settlement following deposition. It has been shown repeatedly that the wind and transport of snow by the wind have an effect on the density. For example, Birnbaum et al. (2010) observed strong densification of snowdrifts. The effect however, has not been quantified for this region. We thus need to make an assumption of this process, even though we lack measurements to confirm it. A few theories of the mechanisms behind this assumption have however been suggested, (e.g. Seligman, 1936).

First of all, during drifting snow events, snow particles will become smaller due to collisions and enhanced sublimation. These effects have been frequently observed, but not often quantified (Sato et al., 2008). The particles getting smaller and smaller may then pack closer between unaffected, larger ones that are already immobilized, increasing the density. Another theory considers the humidity. In drifting snow, the air is close to saturated. Some authors (e.g. Kotlyakov, 1966), indicate that this enhances

TCD
6, 3575–3612, 2012
Event-driven deposition
C. D. Groot Zwaaftink et al.

the density because fast sintering may be facilitated. However, in a snow cover, the air is always close to saturation and it is questionable whether supersaturation will have such a strong effect. We did try to quantify whether water vapour transport from the drifting snow layer to the snow surface may already increase the density. The added mass, however, was so small, that this effect is negligible for the density, especially when trying to explain a change from 83 to more than 300 kg m^{-3}. Thus we only consider the wind as the main driving force of the densification between the precipitation and surface snow (see Fig. 1), but do not try to distinguish physical processes. Furthermore, we do not have the data to allow us formulate a precise dependency of snow density on wind speed. Therefore we looked for a simple relation to obtain a realistic initial snow density and use a logarithmic dependency on the wind speed as we expect that snow transport will densify the snow only up to a certain level. The density ρ of the "new" surface snow added to the snow cover is thus estimated according to:

$$\rho = 361 \cdot \log(U_{\text{event}}) + 33 \tag{1}$$

Over the range of 250 to 340 kg m^{-3}, density thereby increases by about 30 kg m^{-3} for an increase in wind speed of 1 m s^{-1}. The coefficients in Eq. (1) are such that (1) the minimum density corresponds to minimum values observed in our and other studies mentioned in the introduction and (2) after a simulation period of about 10 yr, the snow reaches a mean density in the upper 10 cm of about 320 kg m^{-3}. The impact of the above adaptation will be discussed further in Sect. 4.2.

Not only density, but also the microstructure parameters of newly added snow are adapted to polar conditions: dendricity and sphericity are initially set to 0.5 and 0.75, respectively, or to 0.15 and 1 if the wind speed exceeds 5 m s^{-1}. In addition, whenever the 100-h moving average of relative humidity of the air with respect to ice is greater than 75 %, this will increase bond growth, leading to a larger bond size. The introduction of events to add snow gives the modelled permanent snow cover a pronounced stratigraphy with fewer single layers since snowfall is collected over several deposition events.

TCD
6, 3575–3612, 2012

C. D. Groot Zwaaftink et al.

3.2 Surface compaction by wind

Newly added surface snow is hardly subjected to overload and the low temperatures will not lead to a rapid settlement due to metamorphic processes either. Brun et al. (1997) pointed out that wind may further compact surface snow and they used that effect to render simulation of polar snow more realistic. The model SNOWPACK also features such a compaction mechanism that enhances the basic strain rate, $\dot{\varepsilon}$, down to a depth of 0.07 m below the snow surface according to:

$$\dot{\varepsilon}_{\mathrm{enh}} = (1 + f(u))\,\dot{\varepsilon},$$

$$f(u) = \begin{cases} A_{\mathrm{enh}}(d)(u - u_0)^n & \text{, if } u > u_0 \\ 0 & \text{, else} \end{cases} \tag{2}$$

where u is the instantaneous wind speed ($\mathrm{m\,s^{-1}}$), u_0 is a threshold velocity ($5\,\mathrm{m\,s^{-1}}$) and $A_{\mathrm{enh}}(d)$ is a function of depth, d (m), below the snow surface. In the basic version of SNOWPACK, $n = 1$ and $A_{\mathrm{enh}}(d)$ is a constant set to $A_{\mathrm{enh,0}} = 5\,\mathrm{s\,m^{-1}}$. This proves to be quite inefficient if the surface snow already reached a density of roughly $300\,\mathrm{kg\,m^{-3}}$. Thus, in this contribution, we also run simulations with $n = 3$ and the following depth dependence for $A_{\mathrm{enh}}(d)$:

$$A_{\mathrm{enh}}(d) = 2.7 A_{\mathrm{enh,0}} \left(1 - \frac{d}{1.25 d_{\mathrm{max}}}\right) \tag{3}$$

that is, the effect decreases linearly with depth to reach 20 % of its surface strength at the maximal affected depth d_{max} (0.07 m). Unfortunately, there are no data available to test these model implementations; they simply represent a conceivable additional compaction process for surface snow based on current knowledge of drifting and blowing snow.

TCD
6, 3575–3612, 2012

3.3 Snow settlement

Snow temperatures at Dome C hardly ever rise above −20 °C. In that case the temperature dependence of the snow viscosity dominates the snow settlement process. Here we describe a new temperature dependence of snow settlement covering the full temperature range that has recently been introduced to SNOWPACK. Parameterizations of the Arrhenius-type are often used to describe the temperature dependence of mechanical properties of snow. However, to avoid a hardly compressible snow cover at temperatures below roughly −50 °C, the activation energy Q must be compatible with the material snow. Schweizer et al. (2004) measured snow toughness from −20 °C near to the melting point and their results suggest a value of 16 080 J mol^{-1} for Q compared to 67 000 J mol^{-1} for ice. The same authors also showed that toughness drastically decreases for temperatures above roughly −8 °C, eventually reaching zero at the melting point. We take account of this fact by multiplying the Arrhenius term by a power law as is often done to describe critical phenomena near a phase transition. This results in the following equation:

$$f(T_s) = \exp\left(-Q_s/R\left(\frac{1}{T_{ref}} - \frac{1}{T_s}\right)\right) \left(0.3\,(T_m - T_s)^{\beta} + 0.4\right) \quad (4)$$

where T_s (K) is the snow temperature, T_{ref} a reference temperature (265.15 K), T_m the melting point of ice (273.15 K), R the gas constant (8.31 J mol^{-1} K^{-1}), β the critical exponent (0.7), and Q_s the activation energy of snow. From our calibrations with Alpine snow it turns out that a value of 26130 J mol^{-1}, that is, higher than suggested above, works best throughout the full temperature range of interest. In Fig. 3 we compare Eq. (4) to both the formerly implemented temperature term:

$$f_{old}(T_s) = 9.0 - 8.7\exp(0.015(T_s - 273.15)) \quad (5)$$

and a pure Arrhenius term taking for Q_s the ice value and T_{ref} as 263 K. It appears clearly that at temperatures below roughly 245 K, $f(T_s)$ increases much less than the pure Arrhenius law for ice while being about 50 times larger than $f_{\text{old}}(T_s)$ at 200 K. Nearing the melting point, $f(T_s)$ shows a more pronounced temperature dependence than both the other parameterizations (see insert in Fig. 3). In other words, snow will still settle a little at low temperatures while maintaining the properties of seasonal snow as temperature is nearing the melting point.

3.4 Snow albedo

The current multi-linear regression for snow albedo used in SNOWPACK is similar to the one proposed in Lehning et al. (2002b):

$$\alpha = \alpha_0 + \ln\left(1.442 + \sum_{i=1}^{12} a_i Q_i\right) \tag{6}$$

Table 1 summarizes the terms and coefficients of Eq. (6). For the Antarctic application, however, we applied the following changes to the standard parameterization: we dropped the age term $a_1 Q_1$ that describes accumulation of mineral dust and other "dark matter" at the surface and we reduced the mean value α_0 from 0.8042 to 0.7542. The latter change is based on a comparison with the albedo measured at the BSRN station.

3.5 Initial and boundary conditions

We use Neumann boundary conditions at the surface of our model snow cover. Turbulent fluxes are computed using a Monin-Obukhov scheme considering stability corrections, as described by Stearns and Weidner (1993). The deepest available snow temperature record provides the lower Dirichlet boundary condition. The snow temperature of $-54 \pm 1\,^{\circ}\mathrm{C}$ at this depth of 10 m roughly represents the mean annual air temperature at Dome C. We further assume that settlement is negligible over a few years at this depth.

Our initial snow profile was constructed from several snow profiles taken at several locations around Concordia base. We then used the first simulation year, that is, from

28 January 2005 to 15 February 2006, as a "spin-up time" for the different parts of this profile to adjust to each other regarding deformation rate and temperature. At the end of this spin-up run, we adapted the temperatures of the simulated snow profile to the values measured by the installed temperature sensors, in particular to the upper four that were relocated about 10 days earlier. This allows starting our main simulation with temperatures comparable to the measurements in the top meter of the snow cover while not affecting much the temperature profile below this level. This profile then initializes our main simulation that runs through 1 March 2009.

In order to compare modelled and measured snow temperatures, we specially mark elements at those depths of our model snow cover that correspond to the location of temperature sensors on 15 February 2006. The marked depths of the upper four sensors were reassigned whenever these sensors were relocated. We further mark the first newly deposited element at the surface as a horizon to separate newly deposited snow from the old snow cover. Finally, we mark the element located at 5 m depth on 15 February 2006 to monitor settlement rate. All these marked elements then follow the settlement of the snow cover.

4 Results and discussion

We first discuss the results of the experiment where deposition on the surface is compared to precipitation measurements above the surface, as this is the basis for our event-wise model approach. Then we validate the model by comparing modelled snow height changes to stake measurements, modelled to measured snow temperatures, and the modelled stratigraphy to snow profiles.

4.1 Deposition and erosion

We start by comparing the estimated amount of daily precipitation with the 24-h ECMWF operational forecast for a $1.125^{\circ} \times 1.125^{\circ}$ grid. After subtracting the amount of hoar deposition from the observed daily total deposition measured on a table 1 m above the surface, one obtains the estimated precipitation. Forecasted and estimated amounts compare rather well from March to November 2006, especially regarding the cumulated sums over this period, 6.58 and 5.71 $\mathrm{kg\,m^{-2}}$, respectively. These results are encouraging with regard to validating Numerical Weather Prediction (NWP) in Antarctica with the method used here to measure total deposition. A larger data set including in particular more observations from the winter season is needed to confirm the above finding, though. We consider the measurement method to give a lower bound to the true precipitation as it is much more likely that precipitation particles do not land at the table due to flow distortion or deposits are blown away from the table than that snow from the surface would additionally deposit on the table.

Next we compare the observed daily total deposition (black solid bars in Fig. 4) with the observed daily "new" snow height on the surface (open bars in Fig. 4). As can be clearly seen in the upper panel of Fig. 4, the height of deposit on the table is usually less than measured on SB_{clear}, the daily cleared board on the snow surface. This can in part be explained by the difference in height at which the measurements were performed. Drifting snow often occurs in a shallow layer above the surface while there is no deposition but most probably part of the deposit taken away by the wind on the table 1 m above the surface. Thus snow height changes at the surface and measurements of depositions performed 1 m above the surface are not necessarily linked.

Interesting is the difference between the measurements on the two boards lying on the surface. The observations on both 20 and 21 November 2008 show that in total over one day, there was deposition on SB_{clear} while snow was eroded from SB_{acc}, the accumulating board. This may indicate a large spatial heterogeneity or possibly transport between the adjacent boards. In the lower panel of Fig. 4 we therefore removed all cases where there was deposition on SB_{clear} but erosion on SB_{acc} to analyse the results over a longer period starting 18 November 2008 to end 3 July 2009. Here we focus on the change of snow height ΔHS on SB_{acc} compared to the height of deposits measured on the table. Again, the heights of deposit measured on the table clearly do not match the snow height changes at the surface.

Cumulated over this entire period of 7.5 months, 164.5 mm of snow was deposited on the table. Taking the mean deposit density of 83 $kg\,m^{-3}$ mentioned in Sect. 2.2, this amounts to 13.7 $kg\,m^{-2}$ or 35 % of the mean yearly accumulation. On 3 July 2009, deposit height on SB_{acc} amounted to 120 mm. On the other hand, summing up ΔHS over the full period yields 348 mm, where we may have missed some real erosion cases due to our filtering of spurious cases as described above. Moreover, we frequently observed changes at the surface on the order of 10 mm, corresponding to about 3.6 $kg\,m^{-2}$ or 9.2 % of the mean yearly accumulation. From May 2008 to March 2009 we also measured at least monthly the density of the top 10 cm at Dome C. Using the resulting mean surface snow density from these measurements, that is, 357 ± 14 $kg\,m^{-3}$, and considering the two aforementioned accumulation heights as lower and upper bound for the effectively deposited snow on the surface, lower and upper bounds for accumulation during this period amount to 42.8 $kg\,m^{-2}$ and 124.2 $kg\,m^{-2}$, respectively, or 110 % and 319 % of the yearly mean, respectively.

None of the measurements above thus seems to reliably represent the contribution from precipitation needed to drive snow-cover models. The solid depositions observed on the table 1 m above the surface, however, are available to a very large extent daily throughout our entire simulation period of three years. This, and the fact that after subtracting the hoar deposition there is a good correspondence with cumulated forecast precipitation, led us taking them as is to drive SNOWPACK.

4.2 Stakes

To verify how the model performs during the complete period of simulation, we show in Fig. 5 the computed height of snow relative to the snow layer which was at the surface on 15 February 2006 and we compare it with snow-height changes recorded at the 13-stakes farm since the same date. The simulation named "$\rho_{\text{n, measured}}$" uses the standard SNOWPACK set-up to add snow during snowfall periods. This means that the model takes both height and density of the deposits as observed on the table and adds this amount of snow whenever the observation was made. In the "Event" simulation, the model adds snow only at times of events as described in Sect. 3.1. "Event + SfcDens" applies in addition the surface compaction by wind described in Sect. 3.2. Both the event-driven simulations use the surface snow density given by Eq. (1). Note that the total amount of deposition and precipitation is the same in all three simulations, that is, $42.4\,\mathrm{kg\,m^{-2}}$ over 36.5 months or about $14\,\mathrm{kg\,m^{-2}\,a^{-1}}$.

In simulation "$\rho_{\mathrm{n,\ measured}}$", we clearly overestimate the changes in snow height. This indicates that the measured or estimated density of the deposits is not representative of the snow added to the snow cover but is underestimated. The very low initial overload and the prevailing low air and snow temperatures can not overcome this shortcoming by either settlement or a metamorphic process.

In the event-driven simulations "Event" and "Event + SfcDens", the snow height is increasing more stepwise. These sudden changes in snow height correspond qualitatively to the measured snow height and the model roughly catches the mean snow height increase. But several drifting snow events lead to larger mismatches. As mentioned previously, we do not have continuous observations of deposition and erosion over the complete simulation period. The stake observations, however, reveal some of the drifting snow events with a large effect on the snow height. For example, in June 2006 and in October 2007, rapid increases of snow height were followed by rapid decreases. The latter might indicate strong settlement, surface sublimation, or – most likely – erosion of the surface snow. As stated before, strong settlement of high-density snow is unlikely at these low temperatures, as is sublimation. Indeed, only little sublimation already saturates the air and the process stops. Thus we have to assume erosion to be the main mechanism at work here. Anyway, a much better characterization of drifting snow events is required to improve our knowledge about the processes leading to the final inclusion of new amounts of snow in the snow cover. It looks like there is a subtle balance between deposition, erosion, and immobilization and that these processes may occur under similar conditions but at different time scales. These ideas

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

are behind the proposed new paradigm of event-driven deposition and the subsequent surface compaction by wind but may hint at future developments.

4.3 Snow settlement

Whereas it was not the primary focus of this work, the settlement of polar snow has to be as realistic as possible (see Sect. 3.3). However, as Arthern et al. (2010) point out, there are surprisingly few in situ observations of Antarctic snow compaction. Their measurements of compaction for the 0 to 5 m depth range at Berkner Island yield a mean compaction rate of about $6.6 \times 10^{-10}\,\mathrm{s}^{-1}$. Because at initialization time we specially marked an element at 5 m depth as well as the subsequently buried surface element (see Sect. 3.5), we retrieve after three years of "Event + SfcDens" simulation a compaction rate of $0.97 \times 10^{-10}\,\mathrm{s}^{-1}$ over the 0 to 5 m depth range. This compares better with Giovinetto and Schwerdtfeger (1966) who give a compaction rate at South Pole of $5\,\mathrm{mm\,m^{-1}\,a^{-1}}$, that is, roughly $1.6 \times 10^{-10}\,\mathrm{s}^{-1}$. In addition, the computed local deformation rate at 5 m depth shows an annual cycle of about $0.1 \times 10^{-10}\,\mathrm{s}^{-1}$ in amplitude around a mean of $1.3 \times 10^{-10}\,\mathrm{s}^{-1}$, the maxima being around December of each year. Although the rates above are of the same order of magnitude, the variability found between Berkner Island and South Pole may possibly be due to variations in both annual accumulation rates and mean annual air temperatures. From this we may expect the compaction rate to be even lower at Dome C indeed.

4.4 Snow temperatures

In view of the problems regarding disturbances due to drifting and blowing snow at the site of measurements (see Sect. 2.3), we need to identify a suitable period to evaluate the performance of the model in reproducing the snow temperatures measured within the top meter of the snow cover. The time span from 15 February 2008 to 2 April 2008 shortly after a repositioning of the upper 4 temperature sensors (see Sect. 2.3) is quite appropriate in the sense that during this period there were no important deposition

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

and erosion events near the temperature measurement site as revealed by the 13-stakes farm measurements (see Fig. 5). The upper and lower panels of Fig. 6 show comparisons of computed to measured temperatures for simulation "$\rho_{\text{n, measured}}$" and the event-driven run "Event + SfcDens", respectively, each panel showing results for the snow surface temperature (T_{sfc}, upper frame) and the snow temperature at 10 cm depth ($T_{10\,\text{cm}}$, lower frame).

We now first look at the simulated snow surface temperature that represents the closure of the modelled energy balance and compare it with the measured surface temperature obtained from the measured up and downwelling longwave radiation using a snow emissivity of 0.98. It already appears from the short period shown in Fig. 6 that both runs do a very good job at reproducing the snow surface temperature even though the simulation "Event + SfcDens" performs slightly better in terms of temperature swings and bias. The regression of modelled to measured snow surface temperature over the full time span defined above corroborates this impression as the slope is closer to one and the coefficient of determination is higher for the simulation "Event + SfcDens" (see Table 2). We attribute the difference to a marked change in simulated albedo between the two runs. Indeed, over this period, the mean parameterized albedo (see Eq. (6) and Table 1) decreases from 0.88 ± 0.02 in the "$\rho_{\text{n, measured}}$" simulation to 0.85 ± 0.02 in the "Event + SfcDens" run due to differences in modelled properties at the snow surface, primarily in dendricity and sphericity. In other words, both simulations quite nicely reproduce the energy balance but the difference is a direct consequence of the model set-up.

At 10 cm depth, the "Event + SfcDens" simulation almost perfectly matches the measurements not only with respect to the amplitude of the diurnal cycle but also regarding the timing. Note that the maxima of temperature lag about 4 h behind the maximum of incoming radiation that occurs around 1300 h. Indeed, the model absorbs the net short wave radiation penetrating the snowpack within the top 5 cm. This energy must then be transported by conduction deeper in the snowpack resulting in the observed lag. On the other hand, the lower density of the top layers in simulation "$\rho_{\text{n, measured}}$" allow

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |
TCD
6, 3575–3612, 2012
Event-driven deposition
C. D. Groot Zwaaftink et al.

a deeper penetration of short wave radiation and thus the modelled temperatures at 10 cm depth show larger amplitudes compared to measurements and the maxima occur about two hours earlier than in the "Event + SfcDens" simulation. This comparison shows how valuable good measurements of snow temperature are as they can reveal intrinsic deficiencies of a model.

4.5 Snow profiles

A rather qualitative validation of SNOWPACK is possible by comparing observed and modelled snow stratigraphy. Important for such a comparison is that we keep in mind how they are retrieved. In the manual profile only layers that could somehow be detected by the observer, that is, show a sufficiently large difference in snow properties to adjacent layers, are included (Pielmeier and Schneebeli, 2003). SNOWPACK layers are comparatively thin, however, and represent a quasi-continuous snow profile with mostly much less pronounced vertical differences between them.

In Fig. 7 we compare a manual profile recorded near Concordia Base on 10 December 2008 to the three different SNOWPACK runs described in Sect. 4.2. This is approximately 34 months after the starting date of our main simulation period. We only show the top 15 to 35 cm of these measured and modelled profiles as below a model height of 10.05 m, the modelled snow originates from the common initialization profile of 15 February 2008 and therefore no large differences are expected there between different model runs.

The top panels show the density observed in the manual profile and the simulations. First of all, simulation "$\rho_{\mathrm{n,\ measured}}$", which uses the measured density has a mean density much lower than the manual profile. This confirms that a mechanism to compact the precipitation (other than overload) is needed. In "Event" the density is already larger and there are fewer layers. Stronger densification leading to a result more comparable to the manual profile is obviously seen when surface densification is added ("Event + SfcDens"). We anticipate that future work on improving the parameterizations discussed in Sects. 3.1 and 3.2 could lead to an even more realistic modelling of observed wind crusts and maybe even glazed surfaces.

The middle and lower panels show grain size and shape, respectively. Here we only compare the mean average grain size given by the observer as well as the majority grain type (Fierz et al., 2009). There are only small differences in grain size between the three simulations (middle panels). "Event + SfcDens" shows the right trend and differences between layers, but generally underestimates the grain size, even in the old snow. Note however, that we do not model erosion and deposition at the surface (drifting and blowing snow). Thus we cannot expect to find a close match between observation and simulation. The observer did find a layer of larger grains just below the top layer of rounded grains (RG) but classified the grains as FC (DH), that is, faceted crystals with a few depth hoar crystals. It is questionable, however, whether this layer can be matched to the layer of large depth hoar crystals found in the simulations. This model layer was originally surface hoar deposited and buried by a subsequent snowfall in March 2007. Later on the model turned this surface hoar to the depth hoar seen in the simulations presented in Fig. 7. Except for the rather high density of the modelled layer, this process is quite similar to the one Alley (1988) describes to account for low density depositional – or surface hoar – layers in polar firn.

5 Conclusions

Over three years, the surface snow was intensively studied at Dome C. We put a large effort in assembling a high quality and complete data set containing both information about the snow cover as well as meteorological data from automatic weather stations.

One focus of our work was the daily observation on a table 1 m above the snow surface of depth, density, and water equivalent of solid deposits. The density of these deposits could be measured over 200 times during this period of 3 yr with a mean value of $83 \pm 43\,\mathrm{kg\,m^{-3}}$. This is much lower than the density of the top 10 cm of surface snow that, from our measurements during this period, averaged to $357 \pm 14\,\mathrm{kg\,m^{-3}}$.

Comparison of these deposits to daily measurements of snow heights on two snow boards placed side by side on the surface further confirmed that the measurements on the table 1 m above the surface are not representative for the accumulation on the surface of the snow cover. We attribute the main cause of this mismatch to drifting snow that does not affect the snow surface and the deposits on the table alike. The surface is strongly influenced by the wind as erosion and deposition occur frequently while the wind may often blow away part of the shallow deposition on the table. This first of all affects the timing of deposition, which is continuous on the table, but very irregular on the surface. However, also the total amount of accumulation did not correspond. Over a period of 7.5 months between 42.8 and 124.2 $\mathrm{kg\,m^{-2}}$ accumulated on the surface while only 37.2 $\mathrm{kg\,m^{-2}}$ deposited on the table. In comparison, the mean annual accumulation at Dome C from 1996 to 1999 amounted to about 39 $\mathrm{kg\,m^{-2}}$ (Frezzotti et al., 2005).

Another aim of the present study was to model the snow-cover evolution at Dome C with SNOWPACK. That first required making SNOWPACK suitable for the Antarctic region. In particular a new temperature dependence of snow viscosity has been introduced. We further defined a mechanism by which new snow is not added at the time of precipitation but only during periods of strong winds lasting for 100 h. With this event-driven deposition, the model is suitable for long-term studies of either the snow cover or the surface mass balance in Antarctic regions. For example, snow height measured at stake farms is better represented by event-driven model runs than by a simulation where new snow is added at the time of precipitation. However, when looking at shorter time scales, erosion and deposition events are of major importance. These processes are currently not implemented in the model and therefore model results may not always be accurate on short term runs.

Comparing modelled and measured snow temperatures also allow testing the performance of the model. We showed that SNOWPACK accurately reproduces the measured snow surface temperature, that is, the energy balance is correctly computed. Furthermore, modelled snow temperatures at 10 cm depth agree very well with measurements regarding both the amplitude and the phase of temperature swings. The periods over which such comparisons could be done are limited though, again because of drifting and blowing snow events.

Finally, even though only a very qualitative comparison can be done, the top 30 to 40 cm of the simulated snow profiles show the trends observed in manually recorded profiles in terms of density, grain size, and grain shape. Here too the event-driven simulations show promising results.

Summing up, the comprehensive study presented in this paper not only produced a unique and complete data set but also allowed for developing a new paradigm in snow-cover modelling, event-driven deposition. This new deposition mechanism is based on the assumption that snow is only permanently added to the snowpack during long lasting drifting and blowing snow events, revealing the subtle balance existing between erosion, deposition, and permanent incorporation into the snowpack. The scheme is flexible enough to be developed further as our knowledge of the processes involved increases. Future research is required to investigate the influence of the new deposition mechanism on longer time scales. It remains to be shown whether total accumulation on ice sheets can be better understood when considering event-driven deposition.

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TCD

6, 3575–3612, 2012

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TCD
6, 3575–3612, 2012

Table 1. Description of the terms in Eq. (6). The coefficient values correspond to the standard SNOWPACK implementation (release 3.0.0).

index	Coefficient	Description	Units
0	0.8042	Average albedo	1
1	−0.000575	Age of surface snow, limited to 30 days at most	d
2	0.00459	Snow surface temperature	K
3	−0.006	Air temperature	K
4	0.0333	Relative air humidity	1
5	0.00762	Wind speed	$m\,s^{-1}$
6	−0.000101	Reflected short-wave radiation	$W\,m^{-2}$
7	−0.000056	Snow density	$kg\,m^{-3}$
8	−0.2735	Volumetric liquid water content	1
9	0.175	Grain size	mm
10	−0.301	Bond size	mm
11	0.064	Dendricity	1
12	−0.0736	Sphericity	1

TCD
6, 3575–3612, 2012

Table 2. Regression of modelled vs. measured snow temperatures for the period 15 February 2008 to 2 April 2008. T_{sfc} is snow surface temperature and $T_{10\,cm}$ the snow temperature measured at 10 cm depth, both in degrees Celsius. The simulation names are explained in the text.

	Simulation	slope (1)	Intercept (°C)	Coefficient of determination r^2
T_{sfc}	"$\rho_{n,\ measured}$"	0.88	−8.75	0.79
	"Event + SfcDens"	0.92	−6.01	0.86
$T_{10\,cm}$	"$\rho_{n,\ measured}$"	1.19	11.4	0.86
	"Event + SfcDens"	1.08	3.49	0.97

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

TCD

6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

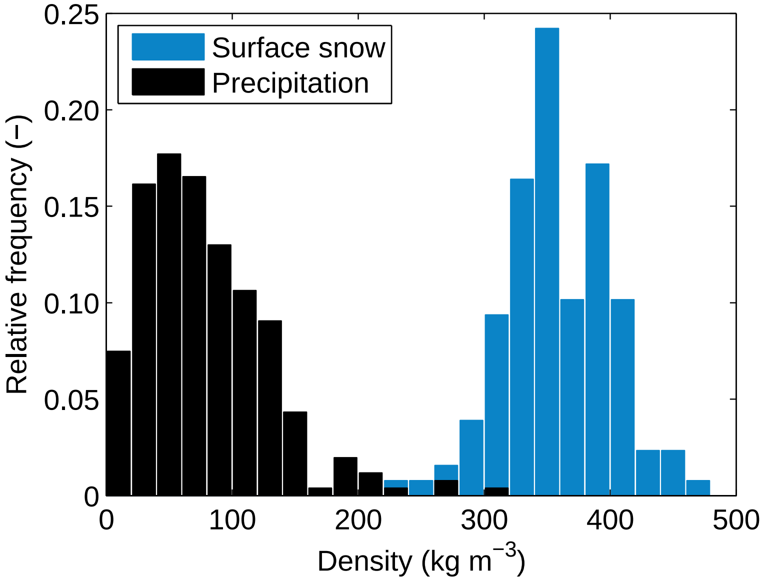

Fig. 1. Relative frequency distribution of the measured density of precipitation (254 measurements, March 2005–March 2009) and surface snow down to a depth of 10 cm (128 measurements, May 2008–March 2009).

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

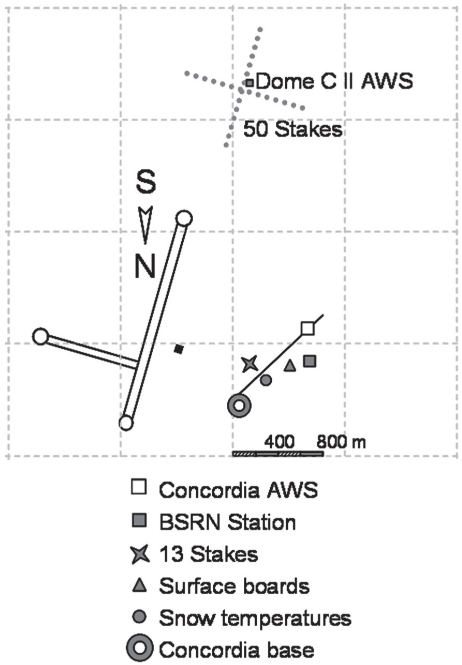

Fig. 2. Plan of the immediate surroundings of Concordia Research Station.

TCD
6, 3575–3612, 2012

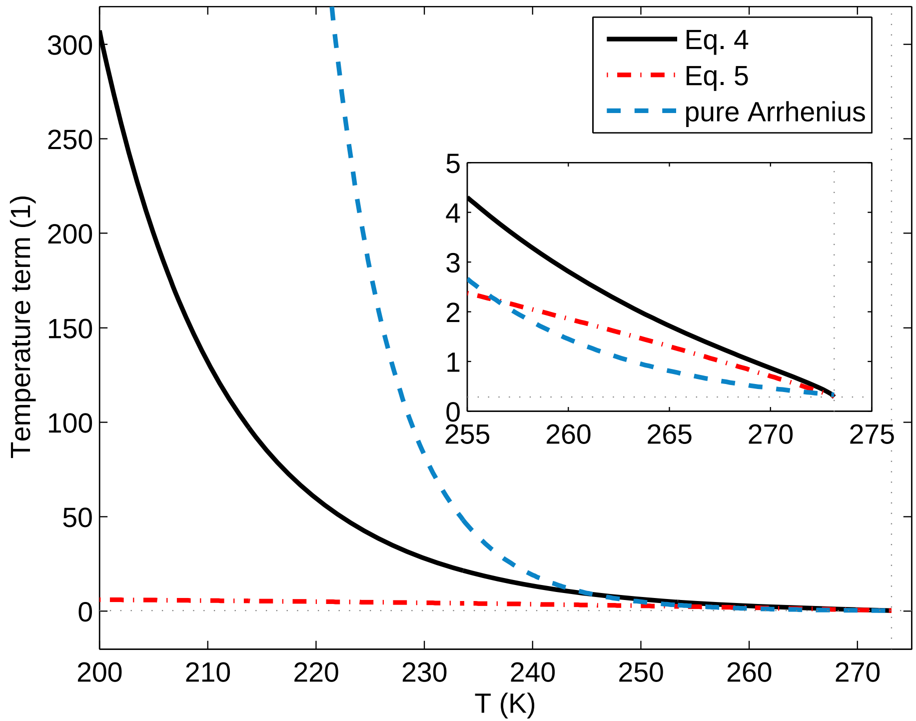

Fig. 3. Temperature term of the viscosity according to Eqs. (4) and (5).

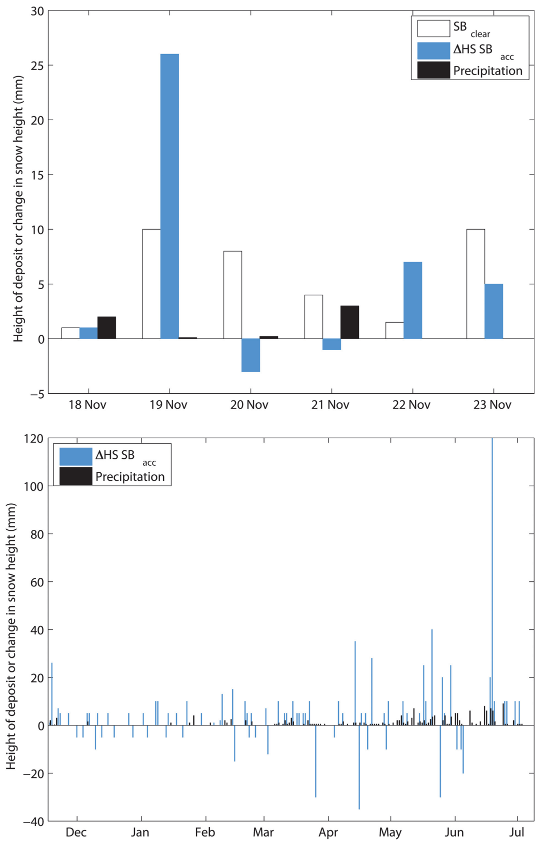

Fig. 4. Height of deposits (mm) measured daily either on a table 1 m above the surface (Precipitation) or at the surface on a daily cleared snow board (SB_{clear}), and change of snow height ΔHS over an untouched snow board (SB_{acc}). Top panel: 18–23 November 2008; Bottom panel: 18 November 2008 through 3 July 2009, board SB_{clear} not shown.

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

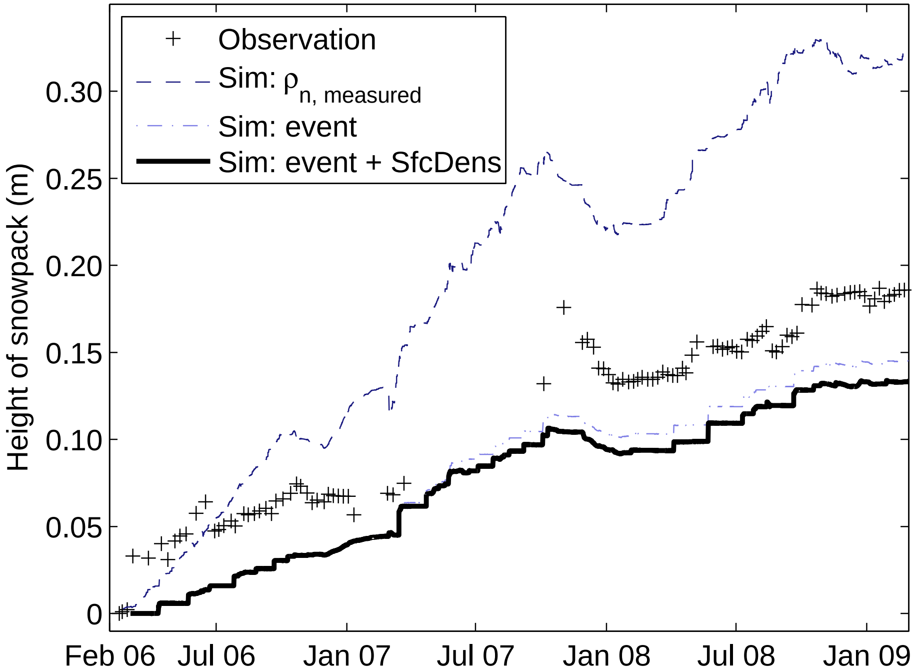

Fig. 5. Height of snow (cm) relative to the snow layer which was at the surface on 15 February 2006. Stake measurements are indicated by crosses, and lines represent results from SNOWPACK simulations using either the measured density of new snow as input ("$\rho_{\text{n, measured}}$"), or the event-driven mechanism without ("Event") or with additional densification due to wind at the surface ("Event + SfcDens").

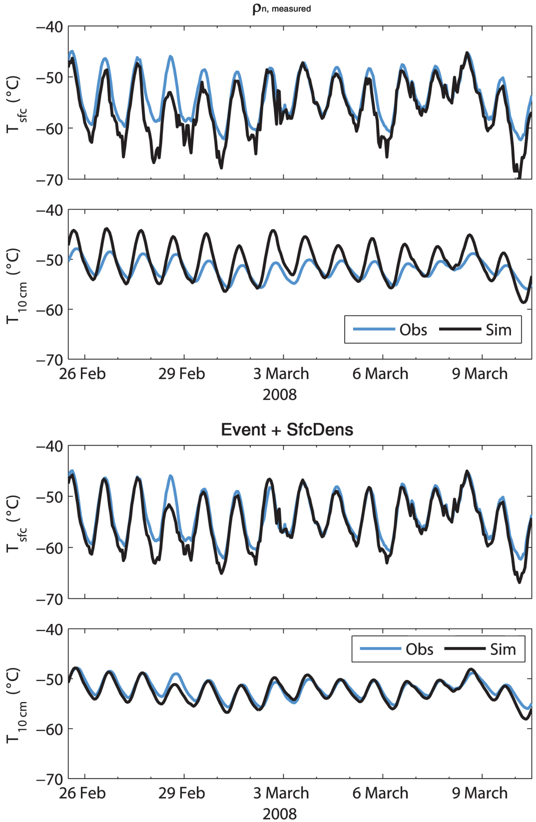

Fig. 6. Observed (Obs) and simulated (Sim) snow surface temperature T_{sfc} and snow temperature $T_{10\,cm}$ at 10 cm depth. The two upper panels relate to the simulation "$\rho_{n,\ measured}$", the two lower ones to the event-driven simulation "Event + SfcDens", that is, with additional wind densification at the surface.

TCD
6, 3575–3612, 2012

Event-driven deposition

C. D. Groot Zwaaftink et al.

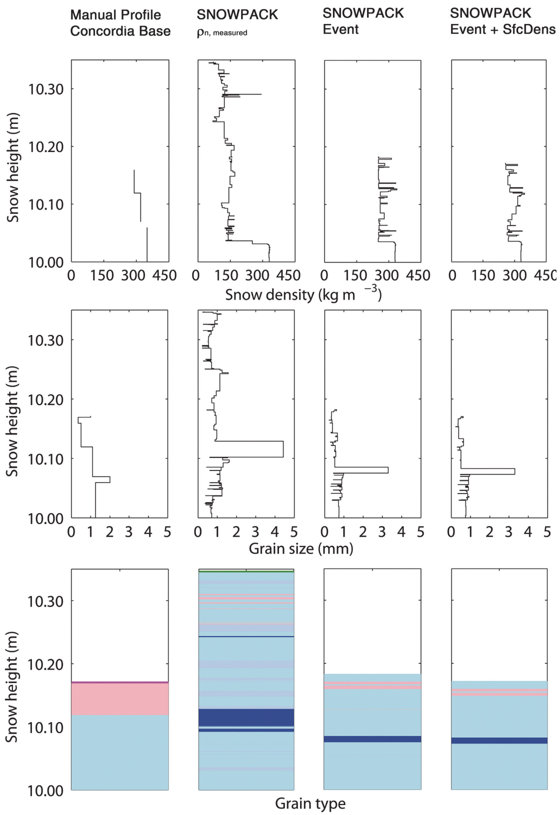

Fig. 7. Snow profile on 10 December 2008 as observed (leftmost column) and simulated with SNOWPACK using either measured new snow density as input ("$\rho_{n, measured}$") or the event-driven mechanism without or with additional densification due to wind at the surface ("Event" and "Event + SfcDens", respectively). The snow height of the observed profile is adjusted to the top of the event-driven simulation with wind densification ("Event + SfcDens"). Top panels show snow density (kg m^{-3}), middle panels grain size (mm), and bottom panels grain type colour coded according to Fierz et al. (2009): precipitation particles PP = lime, rounded grains RG = light pink, faceted crystals FC = light blue, depth hoar = blue, surface hoar SH = fuchsia. Additionally, rounding faceted particles FCxr are described by a mixture of light blue and light pink.

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6, 3575–3612, 2012