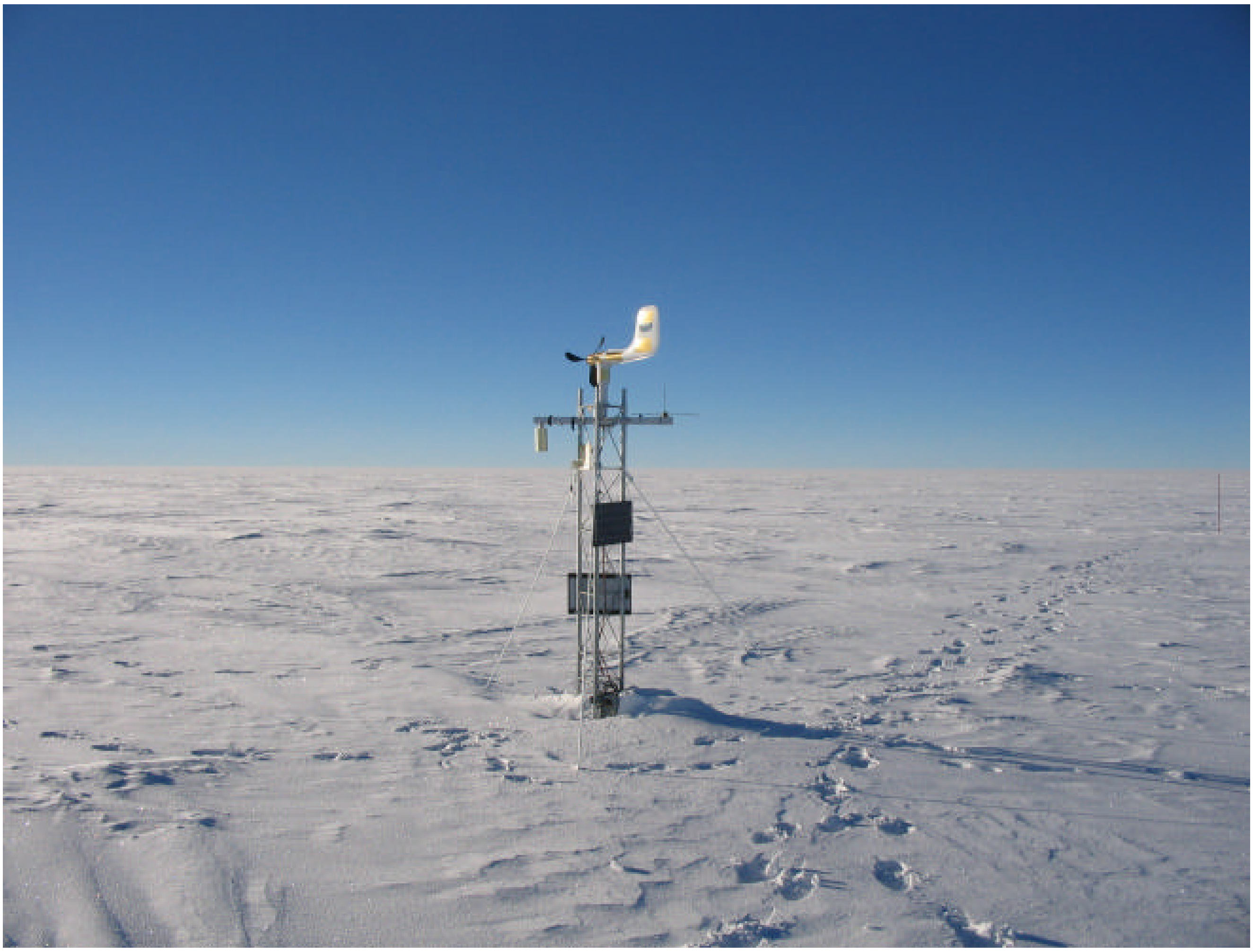
Agenzia Regionale per la Prevenzione e Protezione Ambientale del Veneto

Modelling the snow cover of Dome C (Antarctica) with SNOWPACK



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1 Introduction

ARPAV, the "*Agenzia Regionale per la Prevenzione e Protezione Ambientale del Veneto*" is involved in research projects in Antarctica. The goal is to better understand the exchange processes between the snow surface and the atmosphere but in particular the surface mass balance (SURFMASS (Cagnati, 2004, Valt, 2005)). Initial snow profiles, continuous records of in-snow temperatures as well as meteorological data are available for two Antarctic sites: Mid Point Giulia Station (Giulia, 75° 32.2' S, 145° 51.5' E, 2509 m a.s.l.) and Concordia Research Station (Dome C, 75° 06' S, 123° 24' E, 3233 m a.s.l.).

Several models have been developed to study and model the exchange processes in the surface layers of the snowpack. One such model is SNOWPACK (Bartelt and Lehning, 2002, Lehning and others, 2002a, Lehning and others, 2002b). This model, however, was designed for the temperate climate of the Alps. ARPAV has supplied data from the two above stations of the eastern Antarctic plateau to test SNOWPACK's ability to reproduce the snow-cover evolution in Antarctica and to provide plausibility tests on data quality.

In this report, a summary of the current knowledge on modelling the polar snow cover is first given. Chapter 3 and 4 focus on the preparation of input data for the model runs while chapter 5 summarizes the necessary changes to the original SNOWPACK code. Results are discussed in chapter 6 and we conclude with recommendations for further research in chapter 7.

2 Polar snow cover: review digest

The full review is to be found in Appendix B, p. 25.

2.1 Models used for the modelling of polar snow

DAISY

The snow-cover model DAISY has been described in detail by (Bader and Weilenmann, 1992). It should be noted that DAISY does not account for either metamorphism nor water transport within the snowpack.

CROCUS

The snow-cover model CROCUS has been described in detail by Brun and others (1992, 1989) and Durand and others (1999). CROCUS accounts for metamorphism based on empirical parameters – dendricity and sphericity – as well as for grain growth.

SNTHERM

The snow-cover model SNTHERM has been described in detail by Jordan (1991). It has been widely used in arctic and Antarctic studies. However, it does not include snow metamorphism and, to our knowledge, was not used in detailed snow stratigraphy studies in Antarctica. One exception may be Glendinning and Morris (1999) study on radiative transfer.

2.2 Antarctica

2.2.1 Modelling temperature variations in polar snow using DAISY. (Morris and others, 1997)

Data collected from 1956 to 1958 at Halley Bay during the International Geophysical Year IGY are used to validate the snow-cover model DAISY in polar conditions. For polar snow, the effective thermal conductivity proposed by Morris (1983) but adjusted by a multiplicative factor of 1.8 as well as the parameter values proposed by Kojima (1964) for the compactive viscosity are used. New snow density is set to 400 kg m^{-3}, albedo to 0.9, extinction depth to 20 m^{-1} and aerodynamic roughness length to 10^{-4} m.

Sensitivity analysis are also conducted with respect to the energy input at the surface. The authors conclude that "*The DAISY model has been calibrated for polar conditions using data from Halley Bay primarily because the quality of the meteorological data collected during the IGY is so good. Snow-pit measurements are available to initialise the model and test its capacity to simulate densification, … The densification of the snow was well represented by an equation based on Kojima's (1964) studies of Antarctic snow. Heat flow was best modelled assuming effective thermal conductivities in the upper part of the range reported by Mellor*

(1977). This probably reflects the relative importance of vapour diffusion and convective air and vapour movement in polar conditions. …"

2.2.2 Modelling mass and energy exchange over polar snow using the daisy model. (Morris and others, 1994)

DAISY is used to simulate the snow temperature during the STABLE II experiment using the optimized parameter values found in *2.2.1*. The STABLE II experiment, which was performed near Halley Bay station, is described in the paper.
The authors conclude that "*… the overall simulation is good and the modelling supports the suggestion made by King and Anderson (1994) on the basis of the STABLE II profile data that the roughness lengths for heat and water vapour transfer are considerably larger than the aerodynamic roughness length. We therefore suggest that atmospheric and glaciological models which use a common roughness length of around 0.01 cm may have been underestimating the energy and mass transfers at the air-snow boundary in polar regions.* "

2.2.3 Numerical modeling of snow cover over polar ice sheets. (Dang and others, 1997)

CROCUS having been validated extensively for temperate Alpine snow, an extreme test site was chosen to check its capability in polar conditions. Model forcing is produced with hourly interpolations of ECMWF 6 hour analyses and compared to measured data at the South Pole. A one year data set is used cyclically over 20 years, perturbing the one year record to mimic inter-annual variations in accumulation. Sensitivity to surface snow density (either prescribed – 350 or 300 kg m^{-3} – or parameterized as a function of temperature and wind speed) as well as to inter-annual variability in accumulation is analysed.
Mean monthly temperature profiles are compared to observations reported by Dalrymple (1966) for 1957-58 (IGY). Density and grain-size profiles are compared with measurements taken by J.R. Petit (personal communication to authors) and Gow (1969).
The authors conclude that "*… Our tests indicate that the model is sensitive to air temperature errors. This dependency is significant for temperature profiles inside the snow cover, but less important for snow density and grain-size modelling. The uncertainty on the infrared radiation budget has a low impact on the simulation. On the other hand, accumulation variability is an important factor for grain-size evolution. The density and grain-size of surface snow are functions of the meteorological parameters (air temperature and wind speed), which strongly condition the evolution of the snow cover. Therefore the formulation of these initial values need to be adapted to polar meteorological conditions. The results of this study are quite encouraging because, even with minor adaptations to polar conditions, Crocus simulates many important aspects of polar snow-cover evolution."*

2.2.4 Impact of snow drift on the Antarctic ice sheet surface mass balance: Possible sensitivity to snow-surface properties. (Gallée and others, 2001)

The "Modèle Atmosphérique Régional" MAR had been previously coupled to a physically-based model of the snow pack (Gallée and Duynkerke, 1997), which is now completed by adding the metamorphism laws of CROCUS. This snow-cover model is verified at Col de Porte. However, the goal is to develop a parameterization for blowing and drifting snow to be tested with data from the Byrd snow project held in west Antarctica in 1962 (Budd and others, 1966).

2.3 Greenland

2.3.1 Numerical simulations of Greenland snow-pack and comparison with passive microwave spectral signatures. (Genthon and others, 2001)

The goal is to present selected aspects of the microwave emissivity of snow in central Greenland and to show that these aspects correlate spatially with results of numerical simulations of snow. The modelling approach is similar to the one found in Dang and others (1997). Sensitivity to the forcing meteorological data is analyzed.
The authors conclude that "*… very different results are obtained when the same snow model is forced by two different meteorological datasets. In particular, grain-size, stratigraphy or*

density show strong differential sensitivity to temperature, precipitation or wind … an empirical parameterization of surface density is used, the possible deficiencies of which can have obvious consequences for the interpretation of density profiles, although grain-size and stratigraphy are probably weakly affected. …"

2.4 Sample of standard climate data

In a workshop report on glaciological data, Barry (1984) included a set of standard climatological data for the South Pole but not including radiation.

2.5 Further reading

Literature not listed under References

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Van As, D., M. van den Broeke, C. Reijmer and R. van de Wal. 2005. The summer surface energy balance of the high Antarctic Plateau. *Bound.-Layer Meteor.*, 115(2), 289-317.

Van de Berg, W. J., M. R. van den Broeke, C.H. Reijmer and E. van Meijgaard. 2005. Characteristics of the Antarctic surface mass balance, 1958-2002, using a regional atmospheric climate model. *Ann. Glaciol.*, 41, 97-104.

3 Meteorological parameters

The extreme conditions encountered in Antarctica are known to cause problems in retrieving reliable measurements of meteorological parameters. This was also one of the major challenges in this modelling study of the eastern antarctic snow cover. On the other hand, these extreme conditions make it a very interesting research area.

Both study sites (Mid Point and Dome C) are characterised by little precipitation, rather low air humidity and very low air temperatures. Due to drifting and blowing snow, the snow surface at Mid Point shows a rather rough surface. Automatic snow depth measurements are available for this site but – most probably moving – surface features such as sastrugi as well as the poor quality of the snow depth data prevented to perform any meaningful snow cover simulations. Nevertheless, input data as well as an initial snow profile were prepared for the Mid Point site (see Appendix A4, p. 24).

Below the input data for Dome C Station are described in more details.

3.1 Relative Humidity

Relative humidity is one of those parameters that are difficult to measure at low temperatures. Relative humidity values measured in 2005 and 2006 at Dome C are shown in Fig. 1.

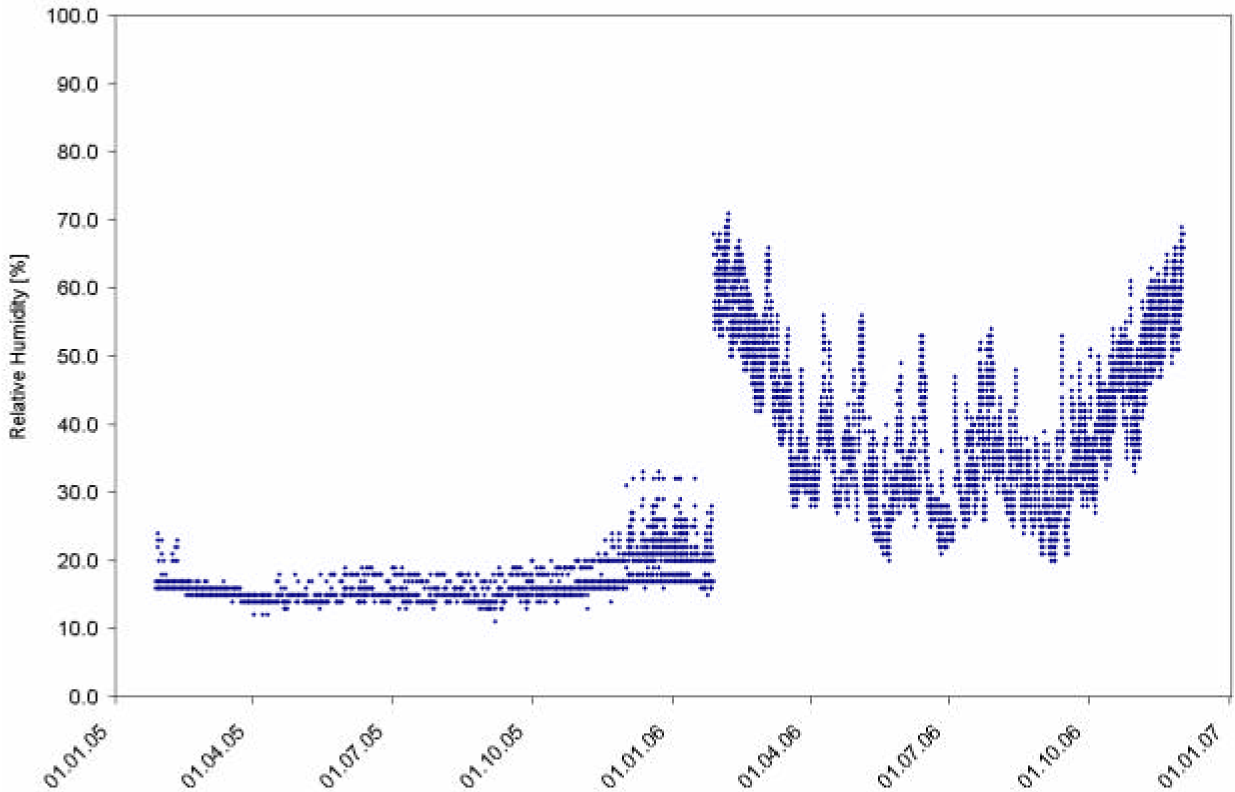

Fig. 1 Measured relative humidity (%) at Dome C in 2005/2006.

The values of relative humidity were also compared with some daily weather reports from Dome C. This showed that the supplied relative humidity values for 2005 must be too low. Model runs for 2005 account for a total sublimation of 24.4 kg m^{-2} and the total snow mass decreased by 14.9 kg m^{-2}. For 2006, however, the total sublimation was only 5.4 kg m^{-2} and the snow mass increased by 3.8 kg m^{-2}. This means that for 2005, modelled sublimation was even larger than accumulation (snowfall etc.) and modelling a developing snow cover was thus not possible (see Fig. 2). The accumulation in 2006 is also very small, especially compared to measured snow height changes and a rough estimate of mass increase – about 15 to 20 kg m^{-2}. Thus the mass balance is influenced by other factors than relative humidity only. These factors will be described in the next sections.

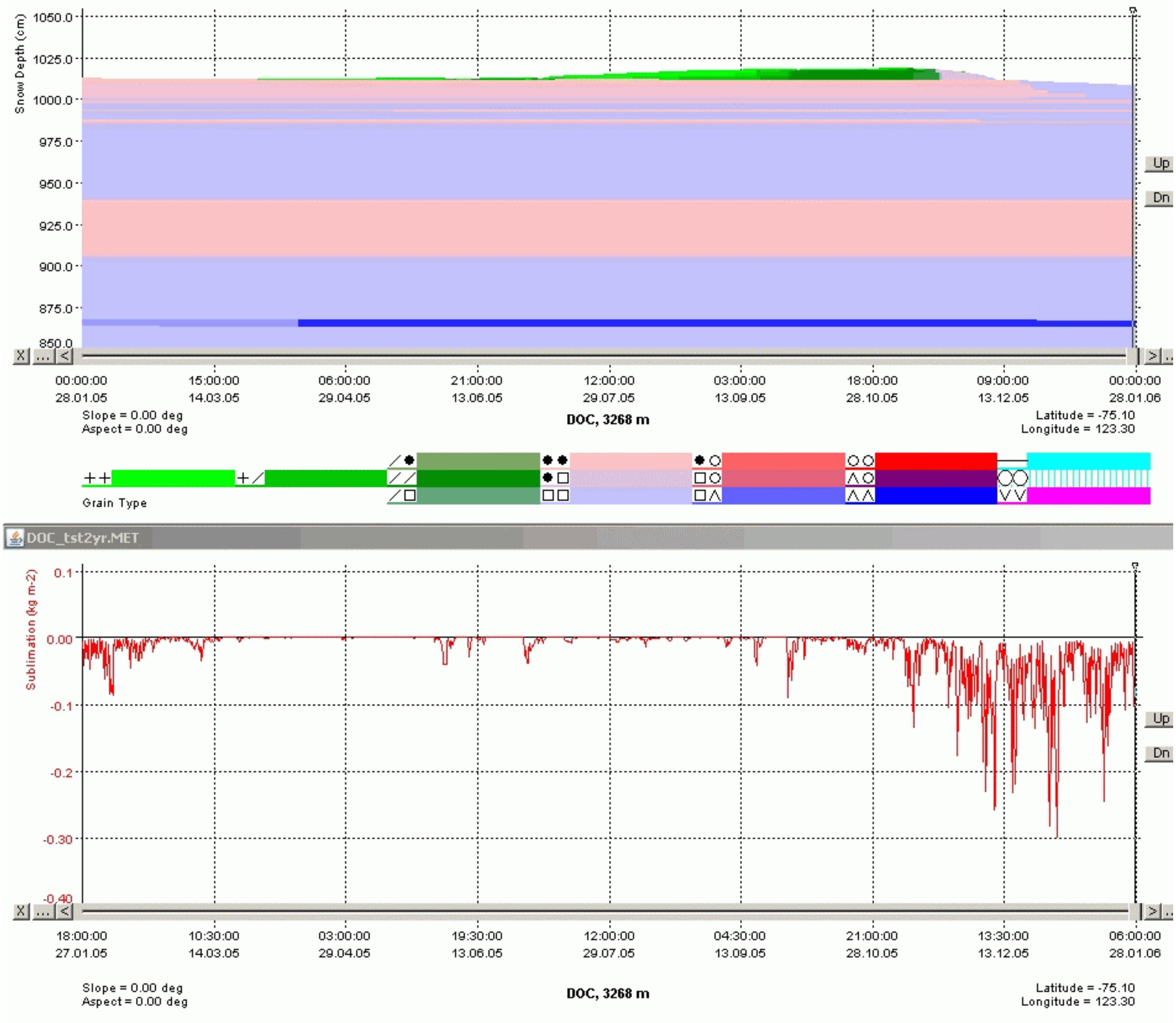

Fig. 2 Modelled grain type at Dome C (upper panel, starting at a height of 850 cm) and sublimation (lower panel, kg m^{-2}) from January 2005 to January 2006.

3.2 Longwave radiation and cloud cover

Another problem in the meteorological data was the cloud cover, *N*. Since no measurements of incoming longwave radiation were available, we needed the cloudiness to be able to make an estimate of the incoming longwave radiation. There were cloud cover observations available in the period between 19 Jan 2006 and 22 Nov 2006. The average cloud cover (N_{avg} = 0.075) observed during this period is also used for December 2006, the beginning of January 2007 and for some other sporadic days where there were no observations available. The cloud cover estimates were only done once a day and, therefore, we had to assume that the observed cloud cover was valid for the whole day. This average cloud cover is extremely low if compared to either Town and others (2007), who showed that at South Pole, the annual average cloud cover is approximately N_{avg} = 0.55, or Van den Broeke and others (2006), who observed an average cloud cover of N_{avg} = 0.33 at Kohnen.

This extremely low average cloud cover observed at Dome C could undoubtedly strongly effect the energy balance of the snow cover. Fig. 3 shows the incoming longwave radiation in 2006 using either observed or fixed (*N* = 0.5) cloud cover and the parameterization by Omstedt (1990). The two curves are very well correlated in time but, as expected, observed cloud cover mostly underestimates incoming longwave radiation.

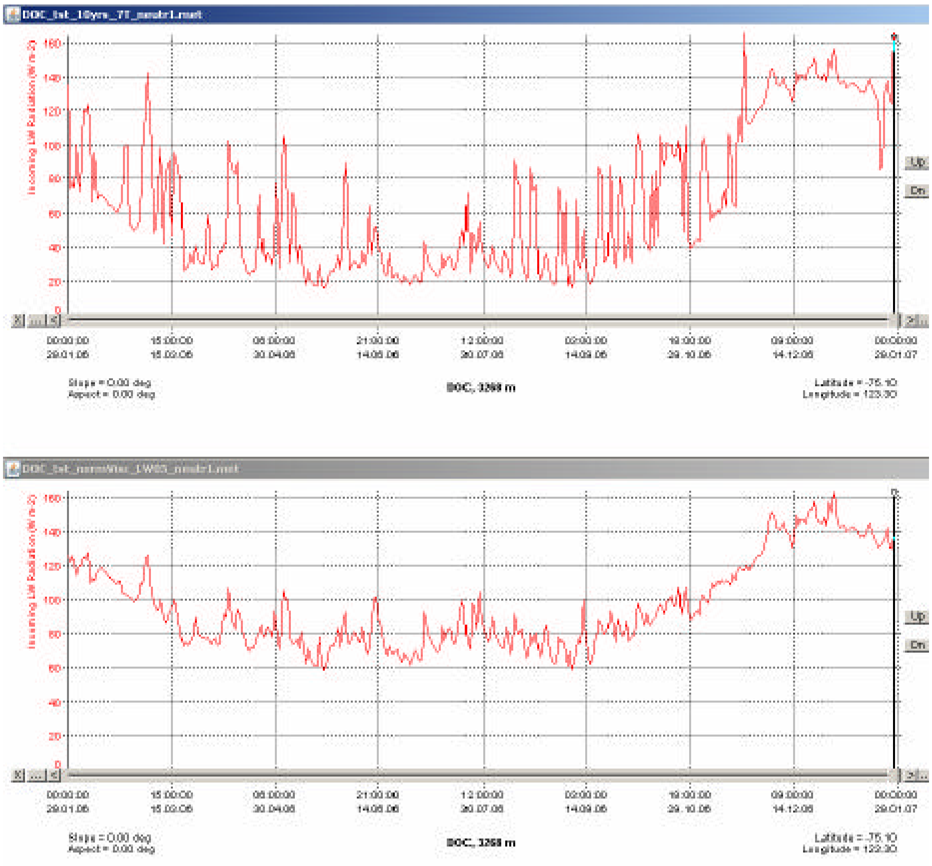

Fig. 3 Parameterized incoming longwave radiation at Dome C in 2006 using either observed cloudiness (upper panel) or a fixed cloud cover of 0.5 (lower panel).

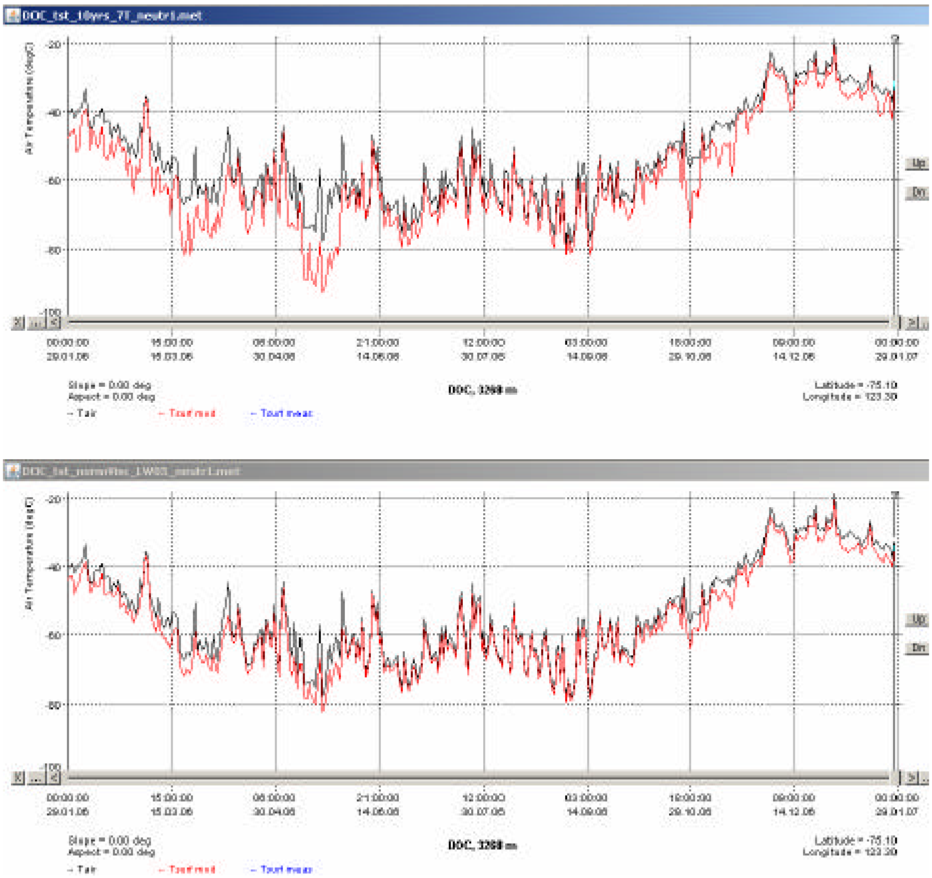

Fig. 4 Measured air (black solid line) and modelled surface temperature T_{SS} (red solid line) at Dome C in 2006 using either observed cloudiness (upper panel) or a fixed cloud cover of 0.5 (lower panel).

A further comparison was done with radiation measurements taken at Kohnen from 1998 to 2001 by Van de Broeke and others (2004). Table 1 shows an overview of yearly means of longwave radiation for both the measurements at Kohnen and our two parameterizations. Both the latter yield markedly lower values than the measurements. The energy loss (net longwave radiation) using observed cloud cover is about twice as large as measured at Kohnen on an annual basis and even larger during austral winter (July/August). Using a fixed cloud cover, discrepancies reduce at winter time and there is a better match during summer. This seems surprising but shows how difficult it is to get a good observation of cloudiness even during daytime, i.e. summer.

Table 1: Yearly means of outgoing, incoming and net longwave radiation (W m^{-2}) from parameterizations with either observed or fixed (N = 0.5) cloud cover as well as from measurements done at Kohnen by Van den Broeke and others (2004).

	Observed cloudiness	Fixed cloud cover (N = 0.5)	Measurements at Kohnen
LW out	128.4	132.2	154.9
LW in	66.6	95.5	125.4
LW net	-61.9	-36.7	-29.5

As a result of this seemingly large energy loss, the modelled surface temperature T_{ss} shown in Fig. 4 is markedly lower when using observed cloud cover. On the other hand, when keeping cloudiness fixed at N = 0.5, T_{ss} matches quite closely air temperature. Thus energy loss at Dome C may not be overestimated as much as the comparison with the measurements at Kohnen would indicate.

3.3 Precipitation

Comparison of the observed snow height changes over 2 years and the amount of precipitation were not consistent. Between 21 Jan 2005 and 10 Jan 2007, the snow height measured at the stakes increased by 12.19 cm. The density of the upper 10 cm is, on average, 316 kg m^{-3}. A rough estimate of the increase in snow mass thus yields 38.5 kg m^{-2}. In our simulations, however, the snow water equivalent increased by 27.6 kg m^{-2} only. Therefore a substantial amount of snow must be added by blowing and drifting snow. Unfortunately, SNOWPACK does not currently model snow transport by the wind on flat fields. We solved this problem by increasing measured precipitation by 40 %. Though this solves the problem of the integrated mass balance, we now add the missing snow at a wrong time and assign it the density of the measured precipitation. However, the actual density of wind transported snow might actually be quite larger, according to research of Doorschot and others (2004). Another potential error that would also explain the inconsistency is a relative movement of the snow stake to the snow at the stake basis, which cannot be excluded.

3.4 Stability effects

Atmospheric stability effects should be considered in Antarctica, but the parameterization now available in SNOWPACK doesn’t give a better result than a parameterization assuming neutral conditions. This might be related to the wind speed data and the height at which wind speed was measured.

4 Data used for model runs at Dome C

The names of standard input and initialization files are to be found in Appendix A4, p. 24 ff.

4.1 Input data (*.inp file)

As shown in 3.1 above, relative humidity data from 2005 cannot be used for meaningful simulations. In addition, cloudiness was available for 2006 only. Thus we decided to build a one year input data set for the period Jan 2006 to Jan 2007. Note, however, that other meteorological parameters available for 2005 are comparable to the 2006 data. The missing relative humidity data for December 2006 are replaced by those of December 2005 increased by +45 %, which is roughly the difference found at the break in the original data set on 27 January 2006 (see Fig. 1). Missing cloud cover data were completed as described in 3.2 above. In addition to the usual SNOWPACK driving data, measured new snow density was added as an input too.

For comparison purposes, in-snow temperatures are available for the period Jan 2005 to Jan 2006. The depths of the 7 sensors at the beginning of a run (28 Jan 2005 00:00) are 0.05, 0.10, 0.50, 1.00, 1.50, 2.00 and 4.00 m. To cover the period of in-snow temperature measurements, the 2006 input data is also used for 2005. This two years 'standard' input set may then be used cyclically as many times as wanted.

4.2 Initial snow profile (*.sno)

In January 2005, a snow profile of 10 m depth was taken at Dome C. Based on this profile, an initial profile for SNOWPACK was elaborated. See Appendix A1, p. 21 ff for details.

Usually, the ratio of bond size to grain size will not be allowed to get larger than 0.75 by the model during a run. However, initial profile values larger than this threshold will not be affected by this limitation. To stiffen the snowpack as much as possible to account for known low settling of Antarctic snow even under large overburden stress, the ratio of bond size to measured grain size was taken as 0.75 in the upper layers (till 2 m below the snow surface) and set to 0.9 in the layers beneath.

Initially, the in-snow temperature sensors are placed at a certain depth, but with time they will settle with the snowpack, assuming no differential settling takes place. Within the initial profile, we especially marked model layers – or better elements –, the depths of which are closest to the initial sensor depths. Monitoring the modelled temperature of these elements thus allows for a one to one comparison with the measurements, assuming the settling of the snow cover is correctly modelled.

5 Model adjustments

SNOWPACK is a flexible modular model with many different possible settings. Parameterizations are mainly applicable to the climate of the Alps and some may not work in the extreme antarctic climate. Obvious adjustments in settings and routines of SNOWPACK have thus been done to get the best possible representation of the snow cover at Dome C. The model is run using Neumann boundary conditions at the snow – atmosphere boundary and Dirichlet boundary conditions at the bottom of the model snowpack. Turbulent heat fluxes are calculated assuming neutral conditions in the atmospheric boundary layer (see 3.4 above).

5.1 Data control

All measured input parameters need to be within reasonable physical limits that need to be adapted to the extreme antarctic conditions, in particular air and snow temperatures. In addition, measured time series can be checked for outliers and gaps filled by a linear interpolation (Lehning and others, 2002a). Both these procedures are not applied to wind speed and radiation input is not checked for outliers. Because too many measurements would be considered as outliers and therefore variation would become too small, we choose not to check relative humidity for outliers either.

5.2 New snow

Another aspect we changed in the model was the initial density of new snow. Normally, new snow density is either parameterized or assigned a fixed density. While we do not expect the parameterization to work for Antarctica, a fixed value of 316 kg m^{-3} as measured over the top 10 cm of the snow cover at Dome C could be appropriate. At Dome C, water equivalent was measured for new snow collected about 0.8 m above the snow surface. According to these measurements, new snow density is expected to vary between 5 and 306 kg m^{-3} with a mean around 80 kg m^{-3}. To allow using that data, measured new snow densities were added to our input file. Thus, given a precipitation amount in kg m^{-2}, the model will calculate the height of new snow according to the corresponding measured density. In addition, whereas the standard model would wait with adding new snow to the snow cover until a certain minimum amount of snow is reached, it will now add the new snow directly to the snow cover.

The standard model assumes initial dendricity and sphericity to be 1 and 0.5, respectively. However, for this application, while initial dendricity remains unchanged (see 5.4 below), initial sphericity is set to 0.75. In case of wind speeds in excess of 5 m s^{-1}, both parameters are set to 0.15 and 1.0, respectively.

5.3 Albedo

The parameterization developed for alpine environments is used but without the term accounting for the age of surface snow (time since deposition). Indeed, snow in Antarctica may be regarded as very clean, making this term obsolete. It is also possible to fix the albedo at a constant value.

5.4 Snow compaction

During fall 2007 we undertook a major revision of the viscosity parameterization used in SNOWPACK. It was thus verified that this parameterization works well in an alpine environment. It is however questionable whether it can be applied in Antarctica. At Dome C, the little overburden due to accumulation will not be effective at compacting the underlying snow. SNOWPACK, however, includes settling due to snow metamorphism while dendricity is comprised between 0.9 and 0.3. To allow for this mechanism to work, initial new snow dendricity is set to 1.0 and the working range extended down to zero (see also 6.1 below).

5.5 Temperature measurements

Usually up to 5 in-snow measurements of temperature can be compared to modelled temperatures at fixed heights (from the bottom). Here we added the possibility to compare up to 7 measured and modelled temperatures, accounting for varying depth due to compaction. This assumes that the temperature sensors follow snow settlement and that the latter is well reproduced by the model. At Dome C, 7 measurements located initially 0.05, 0.10, 0.50, 1.00, 1.50, 2.00 and 4.00 m below the surface are available from Jan 2005 to Jan 2006. Note that the deepest temperature record (10 m depth) was used as the lower Dirichlet boundary condition for the model.

6 Results and discussion

Basic simulation can thus be run from Jan 2005 to Jan 2007, i.e. the period for which we have measurements. However, it is hard to verify how well the model simulates snow cover evolution at Dome C. We have an initial profile, seven in-snow temperatures and some snow height measurements (stake fields) to compare with the model.

6.1 Settlement

As the snow stakes settle with the underlying snowpack, they are not a good indication on how well overall settlement is reproduced. However, they could be used to assess the settling of the new snow relative to the surface of the initial snow cover (see Fig. 5). Unfortunately, measurements are sparse in time and erosion events are not accounted for by SNOWPACK yet. Thus it is difficult to draw more than a qualitative agreement from a comparison with these measurements. On the other hand, from measurements we expect snow density over the top 10 cm to increase to approximately 300 kg m^{-3} with time. However, as can be seen in Fig. 5, the average density of the roughly 20 cm deep layer of snow deposited over two years time is less than 200 kg m^{-3}. Thus, using the new snow densities measured above the snow surface, model settling is not strong enough to compact properly this new snow. We suspect wind to play a major role with this respect, but neither transport by the wind nor erosion nor compaction by the wind are accounted for in the model yet. It may also be that a much higher initial density needs to be assumed as has been done by the earlier studies cited above. In this case, the densities measured at the traps above the surface used here would not be representative of the snow permanently deposited on the ground, which is plausible. Clearly, the deposition process needs to be better understood.

Furthermore, we don't know if the settlement of the 10 m deep initial snow cover is realistic. To verify this, we would need a time sequence of snow profiles the layers of which are dated. Qualitatively, however, we may get an estimate by running the two years data set cyclically to get a ten years simulation (see Fig. 6). The deepest layer already starts to become firn showing densities of about 690 kg m^{-3}, i.e. an increase of roughly 230 kg m^{-3} in ten years at 10 m depth. We would have expected compaction to be much slower at this depth, being able to consider it as an almost firm base to our snow cover. Indeed, according to Gow (1975), at a depth of 10 m and a temperature of –57 °C, which is comparable to our situation, the snow density increases by only about 12 kg m^{-3} in 10 years, i.e. about 20 times less than modelled. In this context we further note that the viscosities calculated for the initial profile are presumably one order of magnitude too low compared with those resulting from Kojima's parameterization (Kojima, 1964; Morris, 1997).

In summary, we may list a few reasons why, until now, we haven't been able to correctly model the settling of the snow. First of all, the parameterization of the viscosity was developed for snow temperatures between –20 and 0 °C. Possible additional mechanisms preventing compaction at lower temperatures are thus not taken into account yet. Second, wind is expected to strongly effect snow compaction at the surface. This influence is not very well known and not accounted for in SNOWPACK yet. Third, we add extra precipitation to get the right amount of snow mass. But since some of this additional mass is due to drifting snow and not to precipitation, we most probably underestimate the initial new snow density.

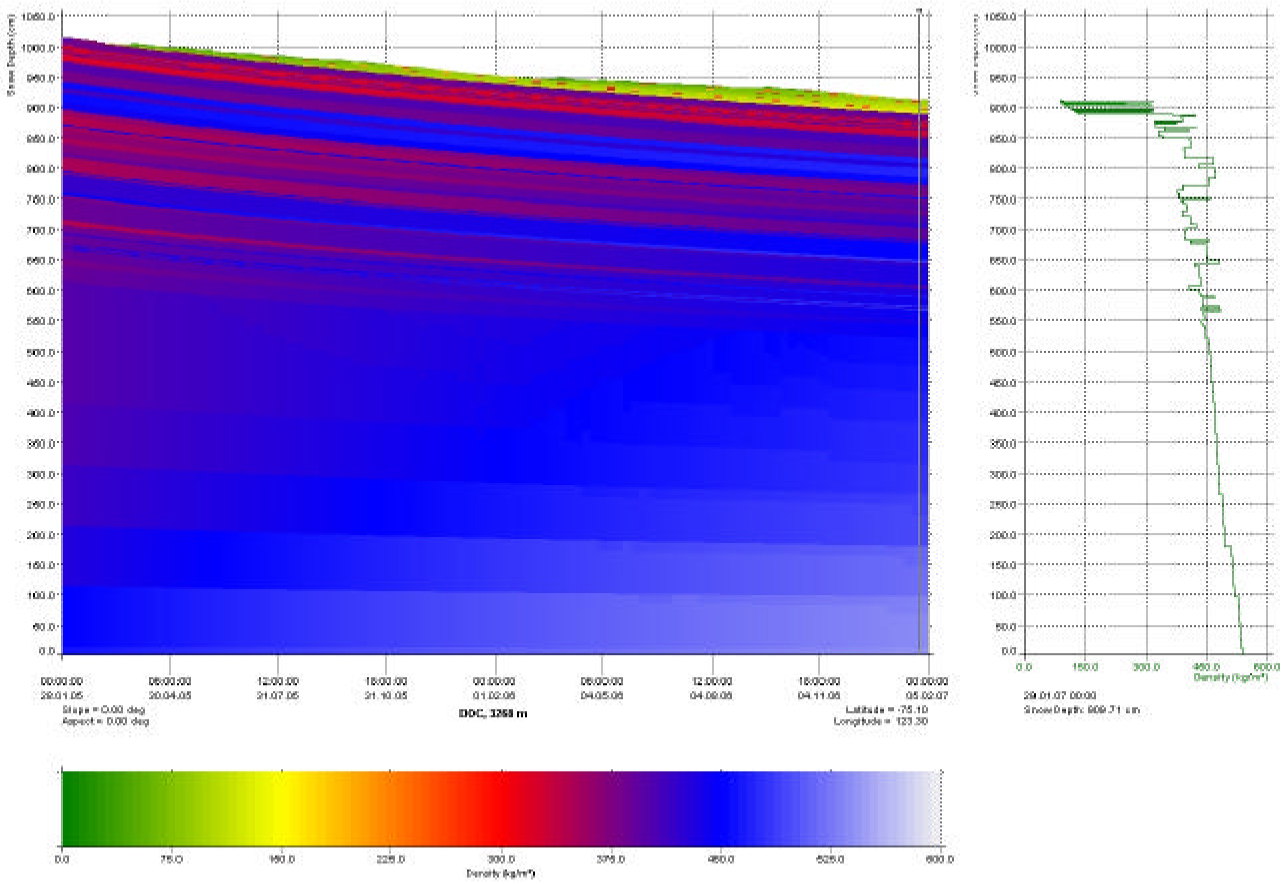

Fig. 5 Density in kg m^{-3}, two years simulation Jan 2005 to Jan 2007, Dome C.

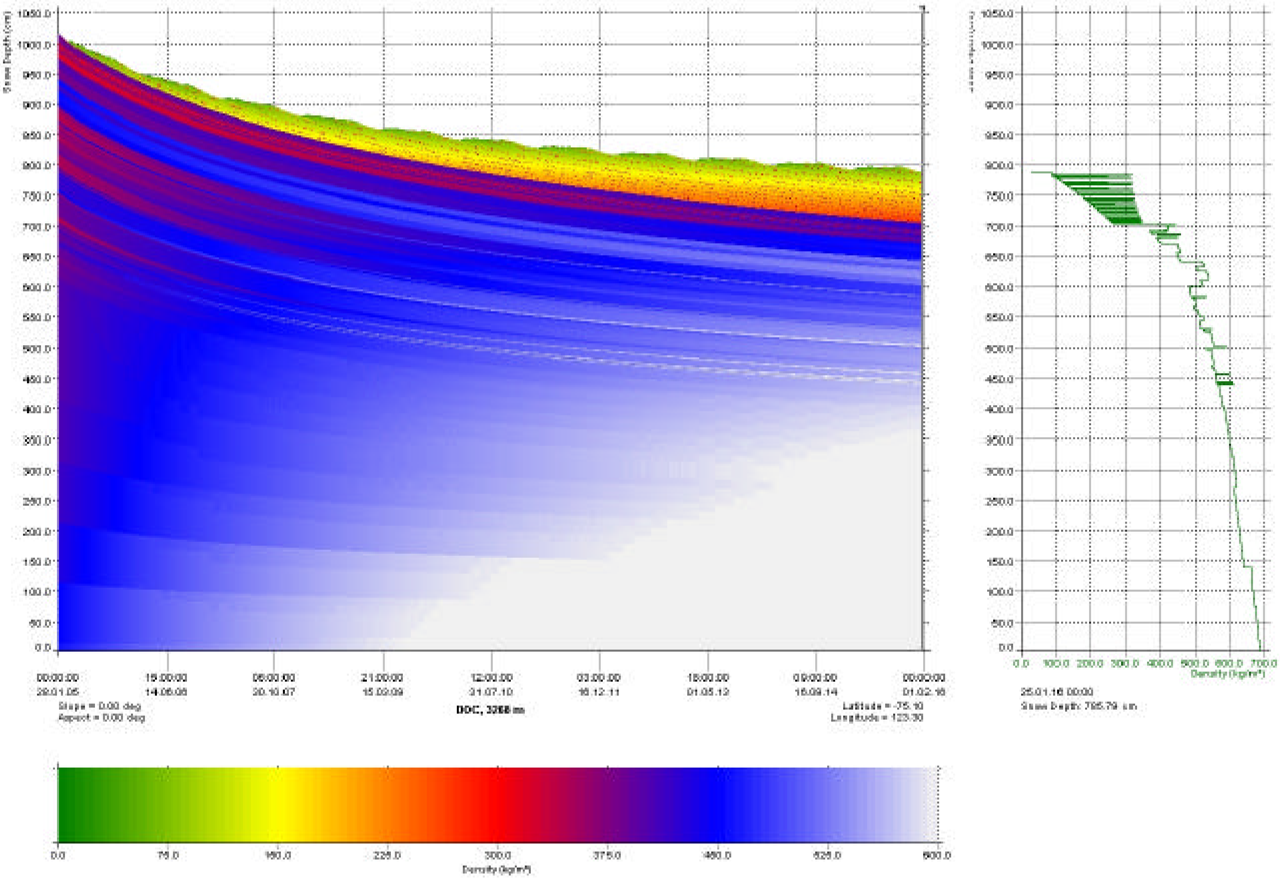

Fig. 6 Density in kg m^{-3}, ten years simulation Jan 2005 to Jan 2015, Dome C.

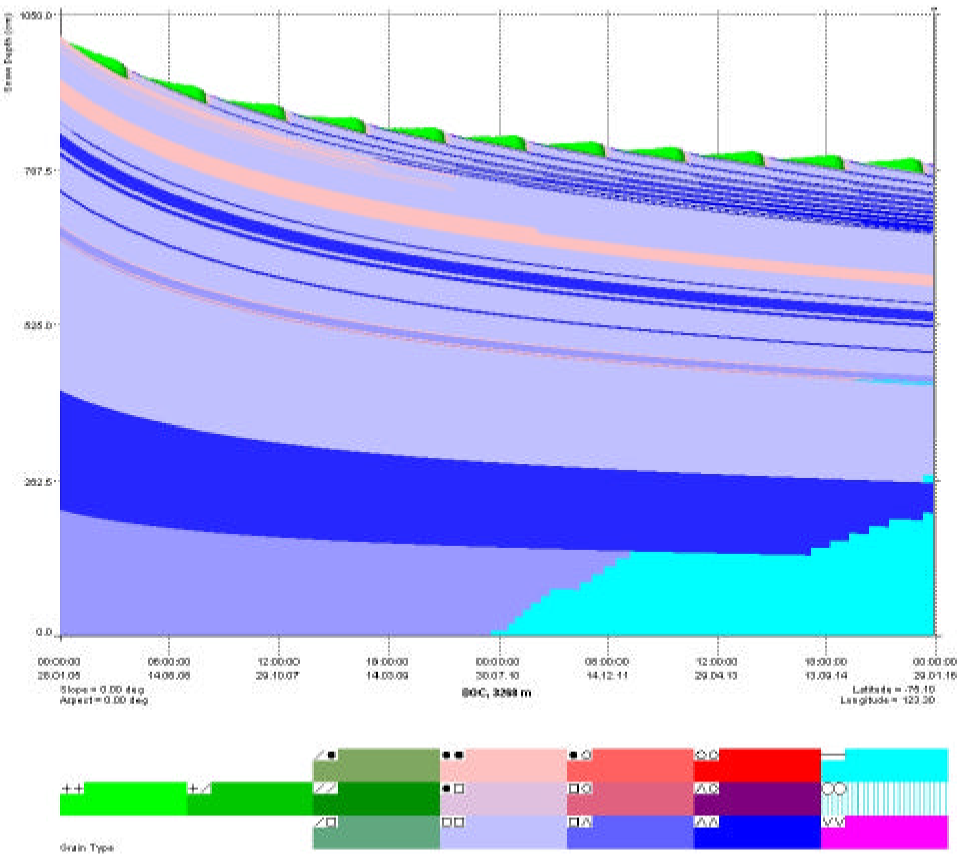

Fig. 7 Grain type, ten years simulation Jan 2005 to Jan 2015, Dome C.

6.2 Stratigraphy

None of both initial profiles of either density or grain type clearly reveals annual patterns one could use to reliably date the layers (see Appendix A, p. 21). It rather looks like we could read signatures of past – short – climatic periods in those profiles. Using repeatedly our one year data set could thus mimic such a period. In that case we would produce a depth hoar band (see Fig. 7) as found twice in the initial profiles. However, we would need many more records taken at Dome C over a longer period of time to verify and better understand the evolution of stratigraphy.

On the other hand, due to the compaction problems discussed above, the initial snow cover seems to develop fully independently from the new snow accumulating at the surface, showing almost no changes except presumably overestimated densification.

6.3 In-snow temperatures

First remember that the depths of both modelled and measured temperatures are changing with time according to settling. While settling in the uppermost part of the initial snow cover may be somewhat realistic (see also Fig. 5), we showed above that compaction is most probably overestimated for deeper lying layers.

The surface energy balance mostly influences in-snow temperatures near the surface. As our data set is based on 2006 data, we compare the three uppermost modelled temperatures with measured ones (initial depths 0.05, 0.10 and 0.50 m) for this year only (see Fig. 8). In general the simulation yields too warm temperatures with a more pronounced mismatch the deeper we go. Furthermore, temperature extremes are delayed relative to their measured counterparts. Both these discrepancies can partly be attributed to the insulation power of the low density snow lying on top of the initial snow cover (about 10 to 15 cm in January 2006).

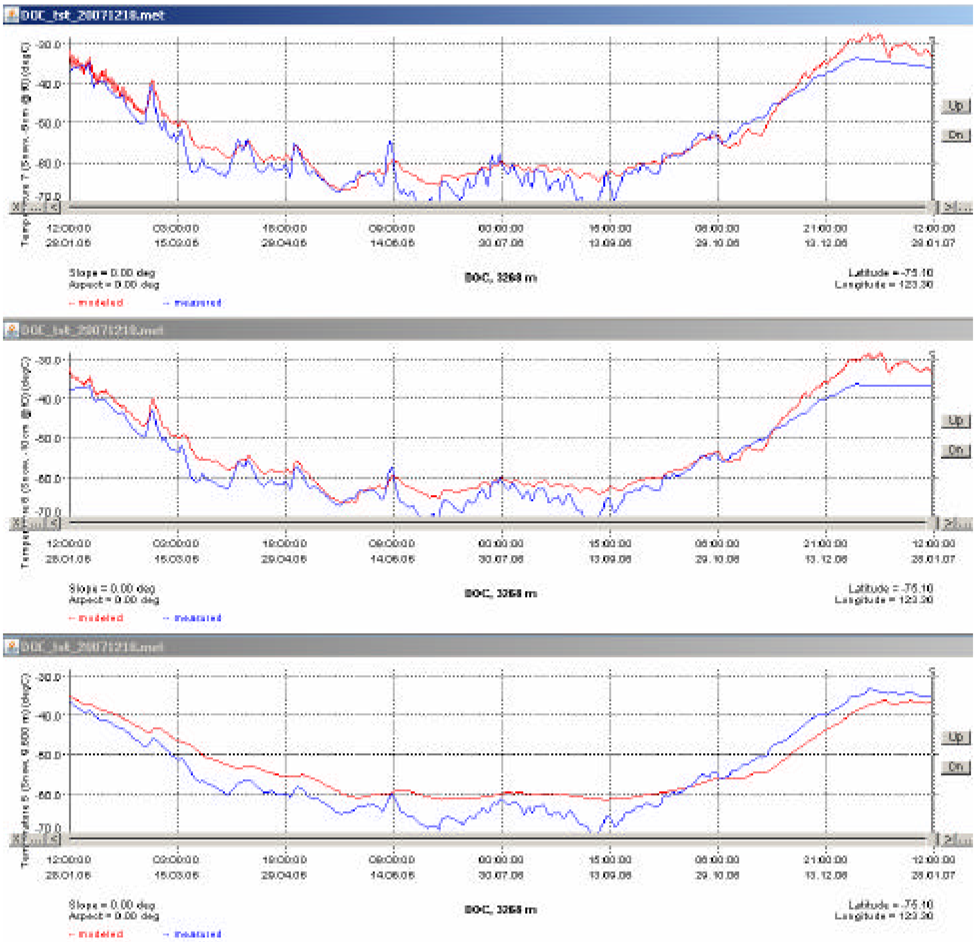

Fig. 8 Modelled (red solid line) and measured (blue solid line) in-snow temperatures at Dome C in 2006 for initial (28 Jan 2005) depths of 0.05 m (upper panel), 0.10 m (middle panel) and 0.50 m (lower panel).

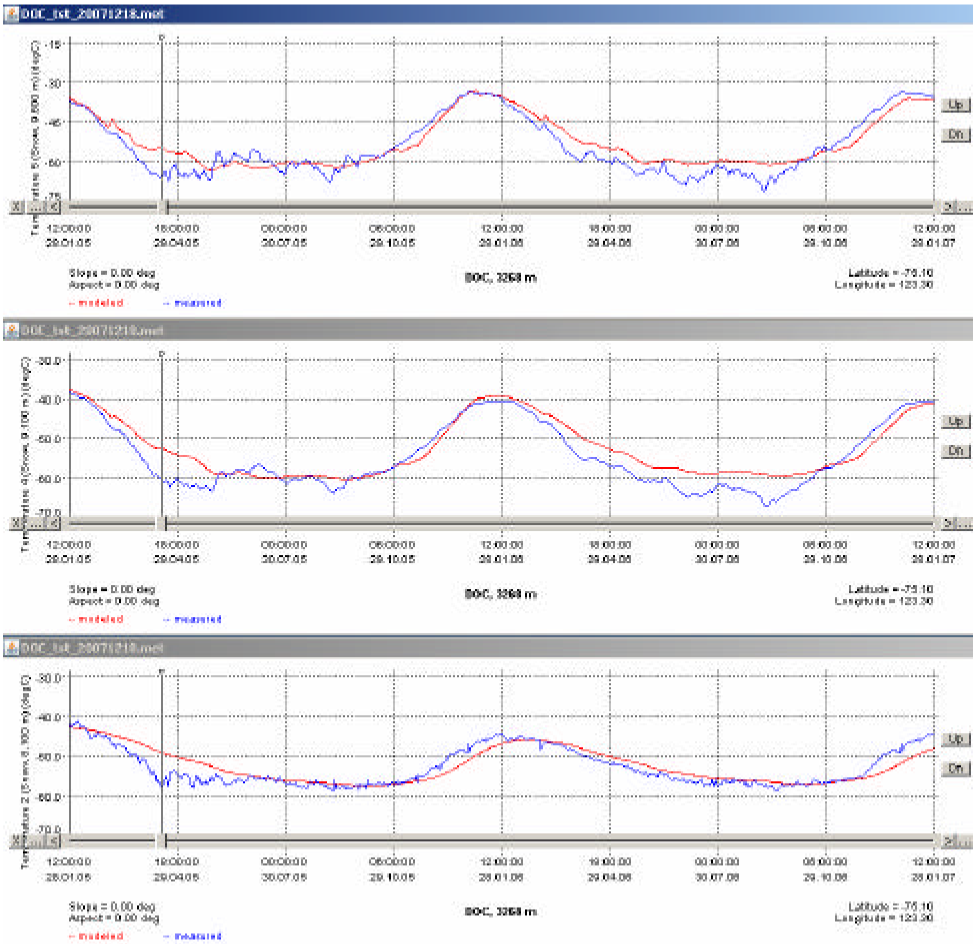

Fig. 9 Modelled (red solid line) and measured (blue solid line) in-snow temperatures at Dome C from Jan 2005 to Jan 2007 for initial (28 Jan 2005) depths of 0.50 m (upper panel), 1.00 m (middle panel) and 2.00 m (lower panel).

The insulating effect of the new snow acts down to a depth of at least 1.0 m in 2006 (see Fig. 9) while deeper in the snowpack the match is more satisfying. On the other hand, Fig. 9 shows that temperatures between 0.5 and 2.0 m depth are quite well reproduced from June 2005 to Jan 2006, a period during which the insulating surface snow is expected to be less influential as the first accumulation cycle is not over yet. However, the large discrepancy from Jan 2005 to Jun 2006 is difficult to explain. Are the two corresponding periods in 2005 and 2006 that different (remember we use 2006 data in 2005)? Are the temperature measurements reliable as there seems to be no time delay in the extremes the deeper we go into the snowpack.

In summary, simulated in-snow temperatures agree qualitatively only with measurements. Even so experimental errors may not be excluded, the main reason for this mismatch are attributable to a still poor model representation of the antarctic snow cover at Dome C.

Conclusions and Recommendations

It is difficult to retrieve an accurate meteorological data set for the extreme climate of Antarctica. Relative humidity, cloud cover and precipitations showed to be the most difficult parameters to get at and our input for cloud cover and precipitations should be further improved for future simulations. Observations indicate an extremely low cloud cover at Dome C compared to other published data. Assuming a constant cloud cover does only partly solve the problem, though. As good data are rare, one option may be to use meteorological re-analysis data sets.

Solving the precipitation problem turns to be even more difficult. Indeed, snow transport by the wind plays a major role in antarctic mass balance, both adding and eroding snow from the surface. SNOWPACK does not represent these processes yet and improvements would need both experimental data and a deeper understanding of the processes at work. By increasing the amount of observed precipitation, we had a way to add the correct mass to the snow cover. However, properties of wind blown snow (grain shape, density) are presumably quite different from those of the observed precipitation particles that are used in the simulations. This introduces yet another source of – initial – errors. Unfortunately, there seem to be no particular observations of blowing and drifting snow for the period under investigation. Nevertheless, for Concordia research station (Dome C), we managed to put together a quite reliable one year input data set that allowed to run simulations reproducing correctly the mass increase measured from Jan 2005 to Jan 2007.

Solving the input data problems, however, will not lead straightforwardly to satisfying snow cover simulations. Our first runs showed that quite a few questions regarding snow cover evolution at temperatures below –30 to –40 °C would have to be answered first. In addition, a much better understanding of surface processes in these regions of extremely low accumulation would be needed to account for the compaction of near surface snow layers. A correct representation of settling and thus density are a pre-requisite for being able to improve model representation of other physical quantities and properties such as in-snow temperatures and microstructure, respectively. To help answer these questions, more experimental data are needed. In particular, time series of dated snow profiles would prove to be very valuable.

In summary, it would be a challenging and interesting task to improve our understanding of the modelling of polar snow covers and in particular the eastern antarctic snowpack. Indeed, better simulations could most probably help answer some important questions related to polar snow cover such as snow – atmosphere interactions in Antarctica or may be even the long term evolution of polar snow into firn and finally into ice. However, this endeavour would require quite substantial additional studies to develop further our current snow cover models. To this purpose, additional verification data would also have to be found or provided, particularly stratigraphic profiles of the uppermost 10 m of the snowpack.

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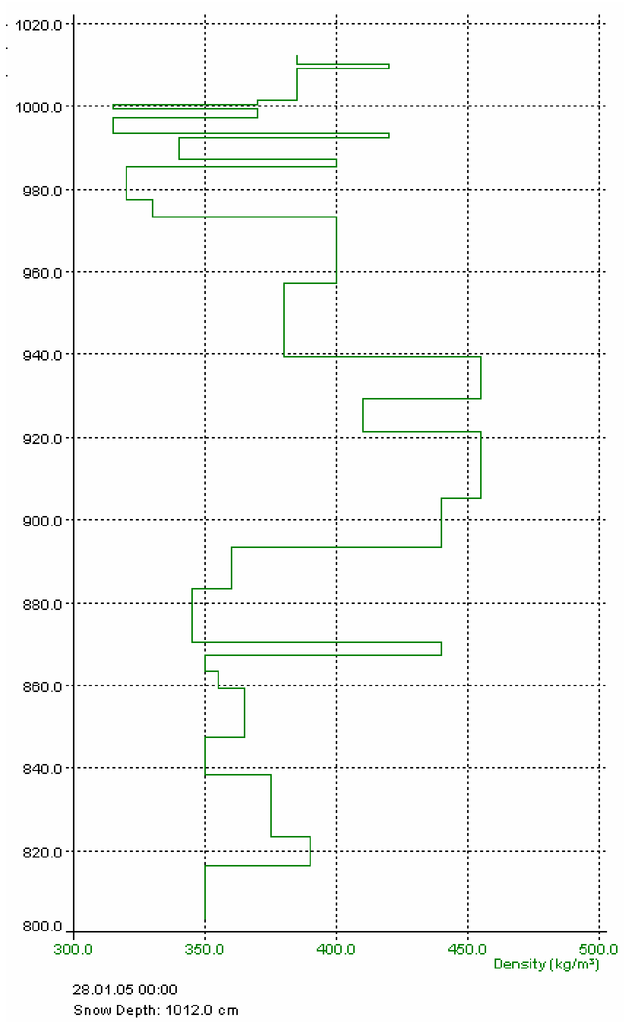
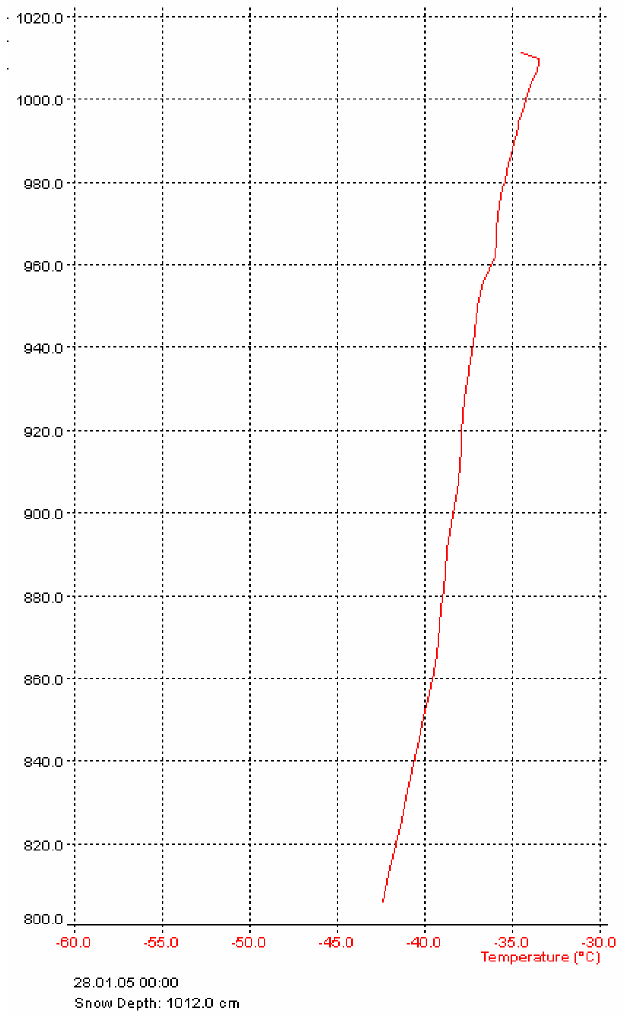
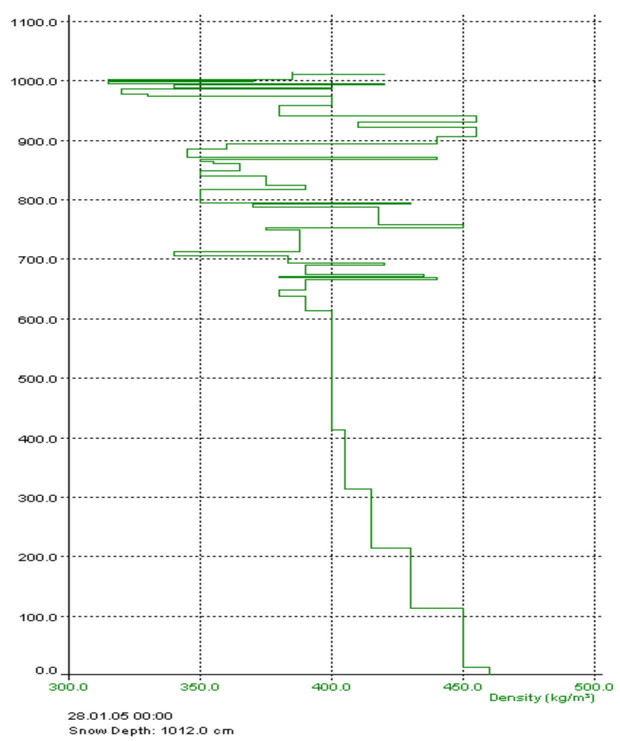
Appendix A:

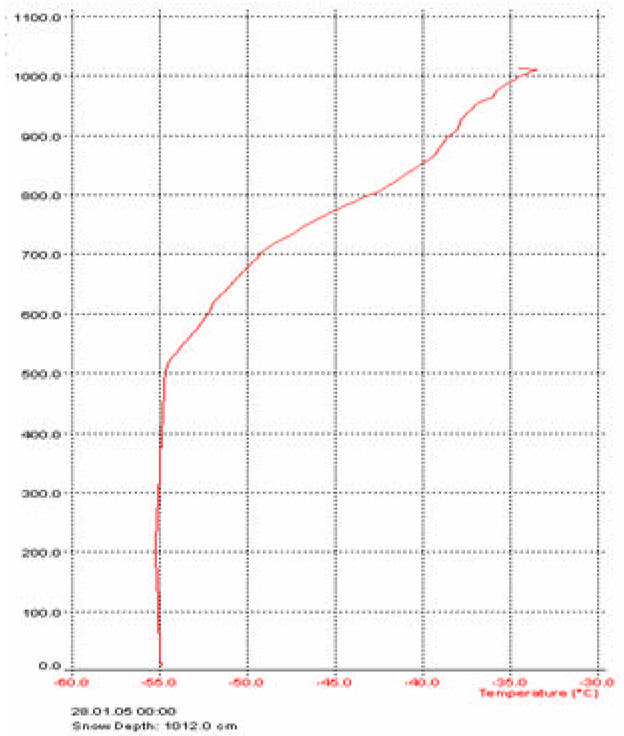
A1: Initial snow profile on 28 Jan 2005: Density and tempearture

Snow height: upper panel 800 to 1012 cm, lower panel 0 to 1012 cm









A2: Initial snow profile on 28 Jan 2005: Grain shape and grain size

Snow height: upper panel 800 to 1012 cm, lower panel 0 to 1012 cm

A3: Initial snow profile on 28 Jan 2005: Tabulated values (*.sno file)

Snow height, density and layer water equivalent *LW* are indicated but not contained in *.sno

First line in table corresponds to bottom layer (0 to 0.12 m)

Year: layers attributed to same year add to about 20 to 40 kg m^{-2} SWE (from 1984)

T: snow temperature (°C) at top of corresponding layer

θ_i : volume fraction ice (1)

θ_a : volume fraction air (1)

rg : grain radius (mm)

rb : bond radius (mm); $0.75 \leq$ (rb / rg) ≤ 0.9

dd : dendricity (1)

sp: sphericity (1)

ne: number of elements in the layer

Snow height *(m)*	*Density* *(kg m^{-3})*	*LW* *(kg m^{-2})*	Year	layer thick-ness L (m)	T (°C)	θ_i (1)	θ_a (1)	rg (mm)	rb (mm)	dd (1)	sp (1)	mass hoar (kg m^{-2})	ne
0.12	*460*	*55.20*	1982	0.120	-54.96	0.50164	0.49836	0.750	0.675	0	0.50	0	1
1.12	*450*	*450.00*	1982	1.000	-55.11	0.49073	0.50927	0.750	0.675	0	0.50	0	6
2.12	*430*	*430.00*	1982	1.000	-55.26	0.46892	0.53108	0.750	0.675	0	0.50	0	5
3.12	*415*	*415.00*	1982	1.000	-55.06	0.45256	0.54744	0.750	0.675	0	0.30	0	5
4.12	*405*	*405.00*	1982	1.000	-54.86	0.44166	0.55834	0.750	0.675	0	0.30	0	5
5.12	*400*	*400.00*	1982	1.000	-54.67	0.43621	0.56379	0.450	0.405	0	0.40	0	5
6.02	*400*	*360.00*	1982	0.900	-52.07	0.43621	0.56379	0.450	0.405	0	0.40	0	9
6.12	*400*	*40.00*	1982	0.100	-52.07	0.43621	0.56379	0.450	0.405	0	0.40	0	1
6.37	*390*	*97.50*	1982	0.250	-51.28	0.42530	0.57470	0.500	0.450	0	0.40	0	5
6.47	*380*	*38.00*	1982	0.100	-50.95	0.41439	0.58561	0.500	0.450	0	0.40	0	2
6.65	*390*	*70.20*	1982	0.180	-50.35	0.42530	0.57470	0.500	0.450	0	0.40	0	4
6.67	*440*	*8.80*	1982	0.020	-50.29	0.47983	0.52017	0.200	0.180	0	0.70	0	1
6.70	*380*	*11.40*	1982	0.030	-50.19	0.41439	0.58561	0.750	0.675	0	0.40	0	1
6.73	*435*	*13.05*	1982	0.030	-50.09	0.47437	0.52563	0.200	0.180	0	0.70	0	1
6.89	*390*	*62.40*	1982	0.160	-49.56	0.42530	0.57470	0.750	0.675	0	0.40	0	3
6.92	*420*	*12.60*	1982	0.030	-49.46	0.45802	0.54198	0.200	0.180	0	0.70	0	1
7.05	*383*	*49.83*	1982	0.130	-49.03	0.41803	0.58197	0.500	0.450	0	0.40	0	2
7.12	*340*	*23.80*	1982	0.070	-48.77	0.37077	0.62923	0.350	0.315	0	0.40	0	1
7.47	*388*	*135.80*	1982	0.350	-46.63	0.42312	0.57688	0.500	0.450	0	0.40	0	5
7.52	*375*	*18.75*	1982	0.050	-46.32	0.40894	0.59106	0.750	0.675	0	0.30	0	1
7.57	*450*	*22.50*	1982	0.050	-46.01	0.49073	0.50927	0.375	0.338	0	0.70	0	1
7.87	*418*	*12.40*	1982	0.300	-43.90	0.45583	0.54417	0.350	0.315	0	0.40	0	6
7.89	*370*	*7.40*	1982	0.020	-43.76	0.40349	0.59651	0.350	0.315	0	0.40	0	1
7.91	*370*	*7.40*	1982	0.020	-43.61	0.40349	0.59651	0.500	0.450	0	0.40	0	1
7.93	*430*	*8.60*	1982	0.020	-43.47	0.46892	0.53108	0.350	0.315	0	0.40	0	1
8.03	*350*	*35.00*	1982	0.100	-42.39	0.38168	0.61832	0.500	0.450	0	0.40	0	2
8.08	*350*	*17.50*	1982	0.050	-42.39	0.38168	0.61832	0.500	0.450	0	0.40	0	1
8.16	*350*	*28.00*	1982	0.080	-41.88	0.38168	0.61832	0.750	0.563	0	0.40	0	1
8.23	*390*	*27.30*	1982	0.070	-41.51	0.42530	0.57470	0.350	0.263	0	0.40	0	1
8.38	*375*	*56.25*	1984	0.150	-40.70	0.40894	0.59106	0.750	0.563	0	0.30	0	3
8.47	*350*	*31.50*	1985	0.090	-40.21	0.38168	0.61832	0.750	0.563	0	0.30	0	2
8.59	*365*	*43.80*	1986	0.120	-39.56	0.39804	0.60196	0.500	0.375	0	0.40	0	2
8.63	*355*	*14.20*	1987	0.040	-39.38	0.38713	0.61287	0.350	0.263	0	0.40	0	1
8.67	*350*	*14.00*	1987	0.040	-39.28	0.38168	0.61832	0.750	0.563	0	0.40	0	1
8.70	*440*	*13.20*	1988	0.030	-39.21	0.47983	0.52017	0.350	0.263	0	0.40	0	1
8.83	*345*	*44.85*	1989	0.130	-38.90	0.37623	0.62377	0.500	0.375	0	0.40	0	3

Snow height (m)	Density (kg m^{-3})	LW (kg m^{-2})	Year	layer thick-ness L (m)	T (°C)	θ_i (1)	θ_a (1)	rg (mm)	rb (mm)	dd (1)	sp (1)	mass hoar (kg m^{-2})	ne
8.93	*360*	*36.00*	1990	0.100	-38.66	0.39258	0.60742	0.500	0.375	0	0.40	0	2
9.05	*440*	*52.80*	1991	0.120	-38.24	0.47983	0.52017	0.350	0.263	0	0.40	0	3
9.09	*455*	*18.20*	1992	0.040	-37.94	0.49618	0.50382	0.150	0.113	0	1.00	0	1
9.21	*455*	*54.60*	1992	0.120	-37.94	0.49618	0.50382	0.150	0.113	0	1.00	0	3
9.29	*410*	*32.80*	1993	0.080	-37.67	0.44711	0.55289	0.150	0.113	0	1.00	0	2
9.39	*455*	*45.50*	1995	0.100	-37.33	0.49618	0.50382	0.250	0.188	0	0.80	0	2
9.57	*380*	*68.40*	1997	0.180	-36.71	0.41439	0.58561	0.500	0.375	0	0.40	0	6
9.602	*400*	*12.80*	1999	0.032	-35.88	0.43621	0.56379	0.350	0.263	0	0.40	0	1
9.73	*400*	*51.20*	1999	0.128	-35.88	0.43621	0.56379	0.350	0.263	0	0.40	0	4
9.77	*330*	*13.20*	2000	0.040	-35.64	0.35987	0.64013	0.350	0.263	0	0.40	0	2
9.85	*320*	*25.60*	2001	0.080	-35.15	0.34896	0.65104	0.350	0.263	0	0.40	0	4
9.87	*400*	*8.00*	2001	0.020	-35.03	0.43621	0.56379	0.250	0.188	0	0.80	0	1
9.92	*340*	*17.00*	2002	0.050	-34.72	0.37077	0.62923	0.350	0.263	0	0.40	0	4
9.93	*420*	*4.20*	2002	0.010	-34.66	0.45802	0.54198	0.250	0.188	0	0.80	0	1
9.97	*315*	*12.60*	2003	0.040	-34.42	0.34351	0.65649	0.350	0.263	0	0.40	0	4
9.99	*370*	*7.40*	2003	0.020	-34.29	0.40349	0.59651	0.250	0.188	0	0.80	0	1
10.00	*315*	*3.15*	2003	0.010	-34.23	0.34351	0.65649	0.350	0.263	0	0.40	0	1
10.01	*370*	*3.70*	2004	0.010	-34.17	0.40349	0.59651	0.300	0.225	0	0.70	0	1
10.02	*385*	*3.85*	2004	0.010	-34.11	0.41985	0.58015	0.250	0.188	0	0.80	0	1
10.06	*385*	*15.40*	2004	0.040	-33.64	0.41985	0.58015	0.250	0.188	0	0.80	0	2
10.07	*385*	*3.85*	2004	0.010	-33.41	0.41985	0.58015	0.250	0.188	0	0.80	0	1
10.09	*385*	*7.70*	2004	0.020	-33.18	0.41985	0.58015	0.250	0.188	0	0.80	0	1
10.10	*420*	*4.20*	2004	0.010	-32.38	0.45802	0.54198	0.150	0.113	0	1.00	0	1
10.12	*385*	*7.70*	2005	0.020	-31.70	0.41985	0.58015	0.150	0.113	0	1.00	0.0625	1

A4: Data files used for snow-cover simulations

A summary of received and elaborated data:

DATA\data-summary.xls

Input files:

DATA\input\dome-c_20050127-3.sno : initial snow cover for Dome C (station Concordia)
DATA\input\dome-c_jan05-jan16_8.inp : 10 years input for Dome C (station Concordia)
DATA\input\giulia_20030112.sno : initial snow cover for Mid Point (station Giulia)
DATA\input\giulia_03-05.inp : 2 years input for Mid Point (station Giulia)

Output files 10 years run (best viewed with the included SN_GUI files):

DOC_tst_20071218.ini : settings
DOC_tst_20071218.met : time series, 1 output every 12 hours
DOC_tst_20071218.pro : profile records, one every 7 days
DOC_tst_20071218.sno : record of snow cover at end of run

Appendix B: Modelling polar snow covers (full review)

B1: Which models have been used till now to model the polar snow cover

- Morris and others (1994; 1997) used DAISY, a physics-based snow model, that is a fully-developed computer program which is described in detail in Bader and Weilenmann (1992).
- Gallée and others (2001; 2005) used a regional climate model (MAR) driven by ERA-15. The model is especially developed for polar regions, including a drifting snow model as well as a physically based snow-pack model that is verified in the French Alps.
- Scarchilli and others (2006): Calculation of the blowing snow sublimation rate with a parameterization based on a sublimation snowdrift model PIEKTUK (Déry and others, 1998) which predict the mixing ratio of suspended snow to diffusion settling and sublimation of blowing snow particle
- Genthon and others (2001): Two series of snow simulations are carried out with the snow model CROCUS whose results are still limited due to a limited validation of the model over Greenland.

B2: Model description

B2.1 DAISY (Morris and others, 1994;1997)

1994: The model DAISY contains a heat flow equation, a densification equation – suggested by Bader (1960, (1962)) – in a highly simplified form to represent the complex snow behaviour. Cited Kojima (1964) suggests to use a ascertained activation energy, compactive viscosity and a value for E/R, with E = activation energy and R the gas constant. The mass transfer by phase change and the transport of water vapour within the snowpack are not treated. The boundary condition consists of an energy flux exchange equation across the air – snow surface. For the albedo that was estimated to vary from 0.75 for old, wet snow and 0.95 for fresh, dry snow. In DAISY the turbulent transfer of sensible and latent heat are functions of the aerodynamic roughness length and the roughness lengths for heat and water vapour transport and the coefficients are adjusted for the stability of the boundary layer. In the model it is assumed that the temperature of falling snow is always the same as the air temperature. The model was calibrated with data from 1957/58 when the conditions were similar and they also were taken as initial conditions for STABLE II.

1997: After the Royal Society's expedition in 1956 – 1958 the station was established permanent (today's British Antarctic Survey BAS) to observe the meteorological conditions and the data are therefore usable in the contemporary understanding of the climatology in the area. Together with the micrometeorological experiments of King (1990) and King and Anderson (1994) the key model parameters "aerodynamic roughness length" for turbulent transfer and albedo were estimated and the model validated using an independent data set. The description of the model is the same as in the paper from Morris 1994. The energy, produced by wind-pumping and depending on the wavelength and height of sastrugi is set to zero for this modelling exercise. Mellor (1975) remarked a discrepancy between compactive viscosity values for polar snowfields and seasonal snowfields. He supposed that the compactive viscosity depends on the interval between snowfalls and the Temperature used for calculations is higher than the average temperature over periods longer than a day and certainly higher than the mean annual average because the effective diurnal temperature that is relevant for compaction is higher. DAISY in this current stage is not able to deal with phenomena linked to vapour transport, e.g. formation of internal layers of hoar frost. The upper boundary condition is described similar in Morris (1994). She gives different values for the aerodynamic roughness length and the roughness length for heat- and water-vapour flux. For latter it is a generally accepted hypothesis that they are equal. The turbulent transfer coefficients are adjusted for the stability of the boundary layer - as mentioned above - with the Richardson number calculated at the height of 1.5 m assuming a logarithmic wind-speed profile up to 10 m height. The initial conditions were chosen from the snow pit data and 18 layers were parted.

B2.2 MAR (Gallée and others, 2001; 2005):

2001: a high resolution sensitivity study of the Antarctic mass balance to the deflation (i.e., snow erosion by the wind and sublimation, including that of blown particles) is performed. The hydrological cycle of the MAR includes a cloud microphysical model (see above) with conservation equations for cloud droplet, rain drop, cloud ice crystal and snow flake concentrations (blown snow particles) and the solar and infrared radiation schemes and the longwave radiation scheme. Cloud properties are taken into account in the radiation schemes. The MAR has been coupled with a model of snow pack which is completed by adding the metamorphism laws from CROCUS. The snow is transported in three modes: creeping, saltation and turbulent diffusion if the particles are in suspension. A threshold friction velocity of the wind speed has to be exceeded that blowing snow is allowed to occur. That threshold depends on the snow pack properties which change with the metamorphism. The potential for snow erosion is parameterized by a snow mobility index depending on the snow grain characteristics. The intensity of snow erosion is quantified by a snow drift index, depending on the wind speed and the mobility index. Both plus a drag coefficient for momentum are part of the threshold friction velocity. Due to a high snow density the threshold shear stress for erosion is assumed to be proportional to the braking shear stress in snow. The difference between the friction velocity and the threshold friction velocity is very small and therefore the assumption to set them equal is made, but not verified at high wind speed. Once eroded the snow particles in the air are considered to be snow flakes by the MAR, but they have different properties. The spatial variability from not uniform fallen blown snow is taken into account. The atmospheric layer where saltation occurs is parameterized according to Pomeroy (1989) considering an equilibrium between weight of saltating snow particles and excess shear stress responsible for ejection from the ground. The turbulent diffusion starts for particles being embedded in the turbulent eddies above the saltation layer. The turbulent scales in the ASL are corrected to take into account the stabilizing effect due to the presence of blown snow particles in the air and the subsequent damping of the turbulence. The problem of turbulence is solved with a slightly adapted Obukhov scale, the friction velocity obtained by integrating the stability correction function for momentum F_m with possibility for stable conditions. The surface blowing snow decreases as turbulence weakens, which is reflected in the expression of the surface blowing snow flux. The friction velocity requires consistency between the Obukhov length and the friction velocity, therefore the contribution from friction velocity to the turbulent scale for snow flakes is linearized and furthermore the increased fluid density due to the presence of snow particles is taken into account by modifying the representation of virtual temperature. The MAR snow model is verified in the French Alps and also tested in Greenland. For the use in Antarctica it is only modified by adding new variables describing snow properties. The snow-drift model is calibrated on Byrd data with parameterization of a new roughness length the determination of a new friction velocity and the threshold friction velocity for snow erosion. For the snow-drift model the sensitivity to snow-surface properties was tested. Therefore two experiments ("OLD" and "NEW") were performed for polar night conditions. The run "NEW" used characteristics of typical fresh fallen snow and the "OLD" the ones for old blown snow. The former conditions are probably not realistic but used as an upper bound to the model sensitivity for pure katabatic wind conditions.

2005: One component of the RCM is the SMB which is equivalent to precipitation plus snow redistribution by the wind (snow drift deposition) minus sublimation, evaporation and melt (erosion). It is a hydrostatic primitive-equation model and contains a cloud microphysical model which is based on the Kessler (1969) parameterization. Ice microphysical processes, a prognostic equation for the ice crystal number and detailed solar and infrared radiation schemes are included in the atmospheric part of MAR. It is coupled with a soil-vegetation-atmosphere transfer scheme that results in the soil-ice-snow-vegetation-atmosphere vertical one-dimensional model SISVAT. The snow model from Gallée (2001) is a multi-layer model including prognostic equations for mass, temperature, water content and snow properties like dendricity, sphericity and size. The exchange of the SISVAT with the atmosphere is modelled via radiative (solar, infrared) fluxes, turbulent fluxes of momentum, heat (sensible, latent) and blown-snow particles. The vertical stability of the surface boundary layer due to snow erosion is taken into account in the aerodynamic formulae. The surface boundary condition for snow erosion is the snow particle concentration in the saltation layer and the blown-snow turbulent flux is included in the prognostic equation for snowflakes.

The horizontal resolution of the model is 40 km and the topography is adapted from a DEM. The validation of the model was done with 1 and 3 years averages of snow accumulation for each 20 km window.

B3: Model parameters

DAISY (Morris, 1994)	
effective thermal conductivity factor K_1 = 1.8, respectively 0.1 compactive viscosity factor K_2 = 1 albedo 0.9 extinction depth 20 m^{-1} all roughness lengths = 0.01 cm, resp. 5 cm new snow density 400 kg m^{-3}	initial condition for STABLE II: temperature profile measured in 1 m depth and predicted temperature below with a lower boundary condition at 12 m depth T = -18.4 °C
DAISY (Morris, 1997)	
thermal diffusivity = 5 * 10^{-7} m^2 s^{-1} for the upper 1.5 m of the snow cover between march and September (Anderson, 1994) Bader and Weilenmann (1992): H_0 = 0.18 * 10^{-5} kg m^{-1} s^{-1} C = 0.02 m^3 kg^{-1}, E / R = 8110 K Kojima (1964) suggested E = 12 kcal mol^{-1}, = 50.2 kJ mol^{-1} for C = 0.024 m^3 kg^{-1}, H_0 = 5.38 * 10^{-3} kg m^{-1} s^{-1} → E / R = 6042 K (fit better for the polar data set) initial new snow density: 400 kg m^{-3}	

albedo a= 0.9 extinction depth ß = 20 m^{-1} aerodynamic roughness length $z_0 = 10^{-4}$ m roughness-length ratio $z_t / z_0 = z_q / z_0 = 1$ compactive viscosity multiplied by 1.8 for $?_{eff}$	
MAR (Gallée, 2001)	
ASL is assumed to exist up to 4 m above the surface. mean sea level atmospheric temp. = -15 °C	
MAR (Gallée, 2005)	
ratio turbulent diffusion coefficient for snowflakes to that of momentum = 3 roughness length for momentum and blown-snow particles = 0.1 mm initial snowpack density = 30 kg m^{-3} initial shape: small rounded grains (sphericity = 1, size = 0.3 mm) initial surface temperature from ERA-15 all initial values are constant along the vertical	

B4: Papers summaries: Antarctica

Location – Surface description – Measurements – Results

B4.1 Morris and others, (1994)

At the British Antarctic Survey Halley Station on the Brunt Ice Shelf (75.6 °S and 26.8 °W. Validation of the DAISY-model has been done with data of the STABLE II experiment.

The snow surface is very uniform and nearly level and in the direction of the prevailing wind the snow cover is uniform over a long distance.

Wind and temperature fluctuation measurements using ultrasonic anemometers / thermometers installed at 5.91 m above snow surface. At the same height humidity fluctuations were measured with a Lyman-Alpha-hygrometer. But also with a cooled-mirror frost point hygrometer and a Vaisala HMP35A in naturally-aspirated radiation shields, that also measured temperature. Further temperature measurements were available from platinum resistance thermometers in forced-ventilation radiation shields. The second wind speed measurement was done with cup anemometers. The validation of these measurements has been done with a comparison of anyway measured parameters some 400 m away at the permanent installed site. The net radiation measurement was difficult because of frost formation on the radiometers. Temperatures in the snow: Miniature platinum resistance thermometers were set up at initial depths of 0.25 m, 0.5 m and 1.0 m. One measurement was always carried out in 0.05 m snow depth and as close as possible on the surface with a moveable sensor. The net accumulation was measured with a snow stick and the density values were estimated by analogy from an experiment in 1957/58.

The snow temperatures in –1.5 m depth are well simulated except in warm winters. The measured and simulated snow surface temperatures perform well, especially in warmer periods but not in periods of low air temperature combined with low wind speeds. To solve the problem the effective thermal conductivity was lowered, but that only gave better results for surface temperatures during strongly stable periods, the rest was worse. Secondly the turbulent transfer coefficient was increased by setting the roughness lengths for heat and water vapour transport to 5 cm and this gave better results. The remaining problems are referred to a not correct recorded net radiation that is considered to be an important factor. The authors of the paper suggest that models using a general roughness length of 0.01 cm underestimate the energy and mass transfer at the air – snow boundary in polar regions. The magnitude of turbulent heat and mass transfer may be larger than assumed in modelling the response of polar snow to climate change.

B4.2 Morris and others, (1997)

The DAISY-model has been calibrated using data collected 1956 – 1958 by the Royal Society's expedition at Halley Bay during the International Geophysical Year.

The snow cover of Antarctica forms a 10 – 100 m thick layer and the density increases up to 800 kg m^{-3}. At the lower point the air is isolated and marks the transition to ice. In the summer air temperature fluctuates from 0 to –10 °C and is associated with a latent-heat flux in the order of 20 W m^{-2}. On Halley Bay the local accumulation rate is about 1m $year^{-1}$. Between May 1957 and May 1958 an accumulation of 3 m snow height was observed, either because the site is exposed to drift or because the snow stake moved downwards. Large amount of snowfall are observed in the period from July to September.

On Halley Bay the air temperature and humidity were measured during 1956 – 1958 in a Stevenson screen with an Assam psychrometer at a standard height of 1.5, resp. 1.4 m above the surface. The precision of the psychrometer for relative humidity below –15 °C was not accurate, but negligible for the simulation of mass and energy balance at the snow surface. The wind speed is measured on a 10 m high anemometer tower with cup anemometers (that were cleaned from hoar-frost deposits if necessary) linked to a recording voltmeter that reported 10 min averages. The data set contained 3 h interval data and the accumulation stakes were read daily. A solarimeter using a ventilated flux-plate radiometer mounted 1m above the snow surface measured the net all-wave radiation. The snow density was measured taking samples of known volume in snow pits. The new snow was not considered, but estimated. The summer layers were identified and used by MacDowall (1964) to look at the annual layers. The snow temperatures were measured using seven thermocouples placed at 0.16, 0.34, 0.6, 1.21, 3.05, 6.1 and 12.19 m depths. MacDowall (1964) assumed in his analysis that the thermocouples all moved downwards as the snow compacted measured with the snow-stakes. The accumulation was measured at different sites, but only one was used in the modelling.

In the winter fluctuations in air temperatures from –20 to –50 °C are associated with less than 1 W m^{-2} change in the latent-heat flux. The simulation of temperature with DAISY gave accurate results but the discussion of fixed temperature measurement equipment or whether the thermocouples are moving exactly with the layers or hanging on their cables is not completely clear. During periods without snowfall the simulated and measured curves of snow height are independent and fit. This suggests that surface settling and compaction are correctly simulated. Because the dimension and the method of erection of the Dexion mast to read the snow height are not recorded, therefore the results depend on the assumption that the mast doesn't move itself. The simulation of snow density is assumed to be correct because the new snow has developed in the simulation to a similar profile as observed in the top 1 m of the December 1958 snow pit. Ice layers are not reproduced in the current version of the model which does not allow melt water to refreeze. The simulated values for net radiation and sensible-heat flux are comparable with the measured values. But the mean annual energy flux over the year at the upper boundary counteracts with –10.7 W m^{-2} the mean rate of absorption of solar radiation in the snow, which is 10.5 W m^{-2}. A sensitivity-analysis was done to distinguish the individual effect of each parameter. In all these runs the results didn't exceed the uncertainty of about 10 % of the amplitude of the annual wave which is about 1 °C. The calibration run is called HY, HA the run with an albedo of 0.8, HB is the run an increased extinction depth for shortwave radiation, decreased new-snow density leads to run HP and a decreased compactive viscosity by a factor 10 and a decreased settling rate to run HF. HY, HP and HF shows that either reducing new-snow density or the compactive viscosity reduces the predicted snow density not by more than 50 kg m^{-3}, which is the minimum expected error in the measured initial densities. Thus DAISY's performance in simulating density is not limited by uncertainty in parameter values. The validation of the model is done in the STABLE II experiment described by Morris in 1994. "DAISY in use" is described in the second-last chapter of the paper. The temperature is simulated over a period of 49 years at Faraday to detect a systematic long term warming of 2 °C. The data have been collected in boreholes. The temperature change over 49 years will penetrate around seven times deeper than an annual wave and it was seen that the temperature in 10 m depth is not necessarily an effect of the mean annual curve but rather an effect of climate variation.

B4.3 Gallée and others, (2001)

Byrd snow project 1962 in West Antarctica.

The first 10 km of the ice sheet starting from it's margin, probably receives the most important part of the Antarctic precipitation. The snow-surface properties for Byrd station may differ significantly from those at other places in Antarctica, in particular along the Antarctic coast. Surface snow densities in Antarctica are generally larger than 330 kg m^{-3} because of strong winds and favouring sintering processes while this value is an upper bound for snow in the Alps that has never been wet. The precipitation does generally not fall as snow flakes in the Antarctica but rather and mainly formed as small ice crystals.

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During the sensitivity test the following results were simulated: temperatures 10 m above the surface, snow net accumulation including precipitation, deposition, sublimation and snow erosion, wind speed. Negative net accumulation areas were found on Lambert glacier and Reeves glacier. The net snow accumulation is expressed in mm of sea level per year for a better comparison with the different contributions to the total amount of annual precipitation. In the simulation "OLD" the annual net snow accumulation of the Antarctica is negative. In the comparison between "OLD" and "NEW" a strong impact of the snow-surface properties on snow erosion was found especially if the winds were strong. That was expected because of the much smaller threshold friction velocity in experiment "NEW". The wind speeds in the experiment "NEW" were examined and it was detected that they were lower in the first level of the model and larger in the upper part of the katabatic layer (fourth model-level). Another experiment was done in letting the properties change during the simulation "NEW" and the result was a hardening of the snow surface. These results show a relatively high sensitivity and illustrate the need of a good knowledge of them. The erosion of snow by wind may be a significant contributor to the Antarctic surface mass balance and should therefore not be excluded in model simulations.

B4.4 Gay and others, (2002)

Snow-grain samples from different locations and depths, collected during activities of French, German, Italian and Norwegian field trips in recent years, were used for snow size measurements.

Surface grains at the interior of the ice-sheet (0-0.5 m depth) are uniformly small (0.1-0.2 mm) except at sites which are affected by a very particular meteorology. They may be classified according to Colbeck et al. (1990) as very fine to fine grain. The largest variability was found along the transect TNB – DC. At one site the difference in two years was very small and on another they were different, because it's a site with erosional forms and seasonal wind crust. Two other sites showed larger grain-sizes even at the surface and were characterized by wind crust which is cemented by thin films of sublimated ice. Underneath there is a depth hoar layer from prolonged sublimation. One site shows sastrugi up to 20 cm height. The differences were explained as an effect of different snow-redistribution processes due to downwind slope at local scale. At the sites where the depth was sampled an increase of grain-size is observed with depth as expected. At Dome C grain-size remains fine in the upper meter.

The grain geometry is difficult to characterize. A simple method for the measurement of grain-size is to place a sample of snow grains on a ruled plate and to estimate the mean size. Another common definition is the greatest extension (Colbeck et al., 1990) or the smallest according to Grenfell and others (1994). Snow sieving or automatic image processing are other techniques for an approximation of grain-sizes. This paper presents an objective and reproducible method based on the analysis of digital images. High-density firn can be cut with a microtome and an image of the surface is taken. This method doesn't suit to lower-density snow. Lower-density snow can be sampled in isooctane to prevent metamorphism and pictures are taken in a cold lab afterwards. An on site method is the use of macro-photography or digital images if the adequate facilities are available. The image processing is accomplished with the method of image segmentation to define the mean convex radius based on the skeletization of each grain and computed by dint of a distance map.

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B4.5 Frezzotti and others (2004)

Surface mass balance studies were done along the transect from Terra Nova Bay to Dome C as part of the International Trans Antarctic Scientific Expedition (ITASE) and in the framework of the Franco-Italian Concordia Station collaboration.

Along the TNB – DC transect the sites with high ablation values are covered by permanent wind crust and depth hoar which indicates prolonged sublimation due to an interruption in accumulation

The authors used different methods for snow accumulation measurements at eight sites: stake farms, ice cores ($d^{18}O$ analyses), snow radar, surface morphology and remote sensing. The cores were linked by snow radar and GPS surveys to provide information on spatial variability in SMB. The snow temperature was measured on 10 or 15 m depth which gives a fairly good value for the mean annual surface temperature in dry snow. This assumption is valuable in region where the maximum temperature is below zero during summer time.

The maximum value of SMB showed an excellent correlation with firn temperature, the minimum values are linked with the change of slope along the prevalent wind direction, but didn't show any correlation to the temperature. It is suspected that the interplay of katabatic winds, storm intrusion, the winter temperature inversion and the surface albedo (wind crusts, sastrugi etc.) may be very important. The snow accumulation can be calculated on the base of the firn temperature in 15 °C depth. That was done both for results from the TNB – DC transect and for the DdU – DC transect. Along the traverse TNB – DC the authors found a mega dune having a wave and formed by variable net accumulation which is detectable on the km scale. The other sites show a morphology prevalently formed by ablation processes. As a result of the $d^{18}O$ analyses sampled every 5 km it is concluded that the snow was not blown far away. The difference between the maximum and the minimum SMB value represents the ablation value which shows a god correlation to the firn temperature. Blowing sublimation is negligible at DC with a mean wind speed of 2.8 m s^{-1}. In the inland also the turbulence is reduced. The surface is often covered by a wind crust with lower albedo, thus the solar radiation penetrates and warms the subsurface snow layer. The resulting upward transport of water vapor activate the depth hoar growth. As regards the roughness parameter the one of wind crust is smaller than the one of sastrugi, dunes or barchans. According to Pettré (1986) snow transportation by saltating starts at wind speeds less than 5 m s^{-1} and accelerate with slope. The threshold wind speed at which the sublimation starts to contribute to katabatic flows in a feedback mechanism is about 11 ms^{-1}. The threshold velocity to differentiate drifting snow and blowing snow is about 13 – 14 ms^{-1}. Wind speeds less than 15 ms^{-1} produce ripples waves and barchans (depositional features) and those greater than 15 ms^{-1} longitudinal features such as dunes and sastrugis. Erosion and snowdrift depends on the snow pack characteristics and changes with snow metamorphism. Observations showed that the largest blowing snow sublimation occur just after snow precipitation. Furthermore the authors compared their SA results with previous studies and asserted a earlier over-estimation due to the inappropriate method at low accumulation sites and an underestimation of the sublimation processes.

B4.6 Gallée and others (2005)

In the context of the International Trans-Antarctic Scientific Expedition (ITASE) measurements were made in the Queen Mary Land and Wilkies Land region close to the coastal East Antarctica (Dumont d'Urville, Casey, Mirny, Vostok, Law Dome, Dome C, Ridge B) during 1980-86. The integration domain covers the part of Antarctica containing the Amery Ice Shelf, Vilkies and Victoria land, the Ross Sea and the Ross Ice Shelf.

A relationship between the spatial snow accumulation pattern, the mesoscale topography and the surface wind field was discovered by Goodwin (1990). he also mentioned a temporal variability over the period of 1930 – 85 and a 14 – 34 % variability of the mean values over 20 km windows.

The Australian National Antarctic Research expeditions measured snow accumulation on snow stakes every 2 km across the region, they measured it in snow pits and in shallow firn cores. There are 33 levels in the vertical.

The regional climate model was used to discover the impact of snow transport by wind on the surface mass balance and to compare directly the output of a fine-resolution RCM with snow-stake measurements. The location of the ablation zones is generally well simulated, although the extent and the erosion intensity seem to be overestimated. A large zone of weak ablation is simulated over the Antarctic Plateau between the Ross Sea coast and the line from Dome C to Vostok. Strong erosion leads to a excessively negative SMB which is not realistic, probably due to crude initial snow pack properties and the non-formation of wind-crusts in the model. The quality of results is calculated with a root-mean-square error and the efficiency index by Nash and Sutcliff. the agreement of the simulation with the measurement are considered to be good even some discrepancies are found. The model simulates snow accumulation relatively well. The abrupt increase of accumulation in the northeast of Dome C as observed by Siegert (2003) is also well reproduced.

B4.7 Scarchilli and others (2006)

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The site characteristics of Middle Point (MdPt, a fuel deposit between TNB and DC) are described as follows: It's situated at 2500 m a.s.l., with an average slope of 2 $^{0}/_{00}$. The temperatures are very low and the wind velocity moderate-high with a high directional constancy. The accumulation time series data seems flat but very sensitive to snow transport. Therefore the precipitation term can be considered as unknown. Precipitation events cannot be recognized without good atmospheric characterization to distinguish from wind transported snow. The studies show a high snow accumulation variability on a local scale with the presence of surface features, like sastrugi and wind crusts.

The meteorological parameters and the surface height at MdPt. are measured on an hourly basis with an AWS.

The data sets have been used to calculate the sublimation and the turbulent latent heat flux was calculated with a profile method using a neutral atmosphere stability condition. The blowing snow sublimation rate is 0.1 – 0.5 mm / day depending on the wind speed regime. During high-speed wind events surface sublimation tends to disappear due to the increasing moisture saturation. The biggest removal effect was observed at medium wind flux, because both sublimation processes can act. 11 mm w.e. is obtained for the annual total ablation considered as the sum of surface and snowdrift sublimation. The stake mean accumulation value in MdPt. for the same period is ca. 36 mm so that a precipitation can be estimated to be 47 mm. The results need to be validated.

B4.8 Ekaykin and Lipenkov, (2006)

Studies of snow surface formation at Vostok Station.

Differentiates the mainly three forms of snow surface in Antarctica: ‘micro-relief’ (probably 101-102 m), three types of ‘meso-dunes’ (probably 102-103 m) of which is shown that they already existed in the remote past and ‘macro-dunes’ (103-104 m). All of them play their role in snow surface formation (the time-series of snow accumulation rates) through different orientation of dunes slope to wind and insolation.

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B5: Papers summaries: Greenland

Location – Surface description – Measurements – Results

B5.1 Genthon and others (2001)

The snow pack simulation for central Greenland to confirm that combining satellite microwave data and a complex snow model can extend the search and validation of surface meteorology data beyond the limited set of available in-situ observations.

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The data sets to simulate the snow pack on Greenland are extracted from the European Center for medium-range Weather Forecasts (ECMWF) and short-term forecast archives are used in the input.

The results of the snow pack simulation with two data sets extracted from the ECMWF and short-term forecast archives result in different reconstructions of the snow pack.

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