Abstract

The proxy for temperature (δ-signal) in ice cores is stored in the snow/ice during precipitation events, hence it reflects the temperature at which precipitation is formed (here approximated by the inversion temperature T_i) weighted with the accumulation. Results from a 14-year integration (1980-1993) with a regional atmospheric model (RACMO, $\Delta X = 55\,\mathrm{km}$) show that the annual mean accumulation weighted inversion temperature ($T_{i,w}$) and the annual mean T_i are not covariant in time at four out of five deep drilling sites considered, mainly due to year-to-year variations in the seasonality of precipitation. As a consequence, the surface temperature ($T_{s,\mathrm{core}}$) derived from RACMO-output, using a method analogous to the retrieval of the surface temperature from ice core δ-signals, deviates from the directly modelled surface temperature T_s on interannual time scales. Results from a 5-year sensitivity integration, forced with a 2 K temperature increase, show an 18% overestimation of the increase in $T_{s,\mathrm{core}}$ relative to the increase in T_s due to a change in the relation between the inversion strength and the surface temperature in a different climate regime. Similar errors may occur in deriving the temperature difference between last glacial maximum and present-day climate from δ-signals in ice cores.

1 Introduction

The Antarctic ice sheet contains a wealth of information on past climatic conditions (e.g. Dansgaard, 1964; Peel *et al.*, 1996). Isotope or tracer concentrations in the ice allow a reconstruction of atmospheric conditions at the time of snow deposition. A difficulty in interpretation of signals from wet deposition is that concentrations strongly depend on the temporal characteristics of precipitation. At many locations precipitation is not evenly distributed throughout the year: data from Automatic Weather Stations (AWS) in Dronning Maud Land (DML) (Figure 1) show that about four large events per year cause most of the annual net accumulation even at a site with an elevation of 3 km (Reijmer and Van den Broeke, 2001). This intermittent nature is also found in an integration with a regional atmospheric model (RACMO). As an example, Figure 2 shows the modelled precipitation and surface temperature for August 1988 at Dome-C, Dome-F and DML05. Precipitation occurs when temperatures are relatively high. As a result, the isotope or tracer signal that is carried by precipitation does not represent annual mean atmospheric conditions.

Relationships have been derived between certain climatic variables and the ratio of the heavy and the light isotope of the oxygen atom ($^{18}O/^{16}O$) and of the hydrogen atom (D/H) in ice. The deviation of $^{18}O/^{16}O$ or D/H relative to a standard isotope ratio (Standard Mean Ocean Water) is referred to as $\delta^{18}O$ and δD or generally as δ. The $\delta^{18}O$-signal in the ice is determined by the temperature difference between the cloud, where condensation took place prior to the precipitation event, and the ocean water in the source region, where the evaporation took place. Because the sea surface temperatures are more stable than the air temperature at high latitudes, the δ-value in snow has often been assumed to reflect primarily the surface temperature (T_s) at the place and time of deposition. Dansgaard *et al.*, (1973), and Lorius and Merlivat (1977) have indeed found a linear spatial relation between δ and T_s. Spatial δ/T_s relationships have been used to infer temporal variations in T_s from the δ-values measured in a vertical column of snow or ice. However, spatial δ/T_s relations can differ significantly from temporal δ/T_s relations (Jouzel *et al.*, 1997). There are several processes that might cause this deviation: a change in the difference between the temperature at which the precipitation is formed and the surface temperature, a change in the temporal variation of precipitation (Steig *et al.*, 1994), a change in the origin of precipitation (Charles *et al.*, 1994), a change in sea surface temperatures (Boyle, 1997), or changes in microphysical atmospheric processes (Fisher, 1991). In this paper, we focus on the first two processes.

The δ-signal in the ice core is not directly related to the surface temperature but rather to the temperature at which the precipitation is formed. This temperature can be estimated by the inversion temperature (Robin, 1977). Strong inversions are a common feature in Antarctica (Phillpot and Zillman, 1970; Connolley, 1996), but they are weakened when clouds move over a site (Robin, 1983). When the relation between the inversion strength and the surface temperature changes in a different climate, the interpretation of the δ-signal in the core in terms of the surface temperature is affected: for Greenland, Cuffey and Clow (1997) suggested that 25% of the surface cooling inferred between ice age and present-day atmospheric conditions could be attributed to ice age strengthening of the inversion.

Changes in the distribution of precipitation throughout the year possibly influence the δ-record in an ice core (Robin, 1983; Steig *et al.*, 1994; Werner *et al.*, 2000; Krinner *et al.* 1997; Schlosser, 1999). When the majority of precipitation occurs during summer, the isotope signal will be representative of summer atmospheric conditions and reflects a temperature higher than the annual mean temperature. When precipitation occurs primarily during winter, the δ-signal reflects temperatures which are lower than the annual mean temperature. Therefore, the δ-signal is physically related to the temperature weighted with the net accumulation. Cuffey *et al.* (1995) and Johnsen *et al.* (1995) found that for Greenland, the temperature difference between the last glacial maximum and the present-day climate derived from the δ-signal using the classical approach is half of the temperature difference derived using an isotope independent method (borehole paleothermometry). A change in the seasonality of precipitation is the most plausible explanation for the disagreement between borehole thermometry and the classical approach using the δ-signal from Greenland (Werner *et al.*, 2000; Krinner *et al.*, 1997).

In summary, changes in intra-annual variations of precipitation and changes in the relation between the temperature inversion strength and the surface temperature are processes which will generate a discrepancy between spatial and temporal δ/T_s relations. The goal of this work is to obtain more insight into these processes and to understand their effect on signals measured in Antarctic ice cores. Since measurements are sparse, we use output from a regional atmospheric model (RACMO) to obtain this information. The advantage of this model over GCMs is that it is driven from the lateral boundaries and from the sea surface by fields which are essentially based on measurements. Hence, large-scale synoptic systems, which are important for the formation of precipitation, are represented by the model in close agreement with observations. In addition, the model uses a grid spacing of 55 km, which is better than the resolution commonly used in 'state of the art' global climate models used for longer integrations of about a decade. Most of the moisture transport (83%) is by the atmospheric flow that is resolved by the model grid, whereas in most models with a coarser resolution, moisture transport by horizontal diffusion plays a dominant role. In addition, the near-surface climatology (especially temperature) of RACMO is in better agreement with measurements than the 15-year re-analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF), which suffer from a decoupling between the lowest atmospheric model layer from the overlying atmosphere (Van Lipzig *et al.*, 1999).

An integration spanning the 14-year period 1980-1993 is used to study the effect of temporal variations of precipitation and temperature inversion strength on the signals that are expected to be found in ice cores. We successively study how seasonality, daily variations, and the diurnal cycle of accumulation affect the 14-year time series of the net accumulation-weighted inversion temperature at five deep drilling sites in Antarctica. In addition, the effect of these processes on 7-year mean δ-values is studied. A sensitivity integration of 5 years, forced with a temperature increase of 2°C at the lateral boundaries and at the sea surface together with a retreat of the sea ice, is used to study the effect of changes in the relation between inversion strength and surface temperature on the temperature derived from the δ-signal in ice formed during a different climatic regime.

2 Description of the integrations

2.1 Control integration

An integration of 14 years is performed for the period 1980-1993 with the regional atmospheric model RACMO, using a grid spacing of 55 km. The grid covers the Antarctic ice sheet and a large part of the Southern Ocean with 122×130 grid points. The model uses the parameterisations of the physical processes from the ECHAM-4 model (Roeckner *et al.*, 1996) and the formulation of the dynamical processes from the HIRLAM model (Gustafsson, 1993). The model formulation is described in detail by Christensen and Van Meijgaard (1992) and by Christensen *et al.* (1996). Modifications for the Antarctic region are described by Van Lipzig (1999) and Van Lipzig *et al.* (1999).

The model is driven from the lateral boundaries by 15-year re-analyses from ECMWF (ERA-15), which are based on observations. The fields are updated every six hours. Measurements are also used to prescribe the sea surface temperature and the sea-ice extent (sea surface temperature and sea-ice extent were taken identical to the values used in ERA-15). The model output is in good agreement with measurements from several Antarctic stations (Van Lipzig *et al.*, 1999; Van Lipzig *et al.*, 2002a).

A comparison by Schlosser *et al.* (2002) using data from the coastal station Neumayer shows that the episodic nature of the precipitation is realistically represented by RACMO (Figure 3). Neumayer (8.4°W, 70.7°S) is located on the Ekströmisen Ice Shelf at about 7 km from the ice edge (Figure 1). The net accumulation at Neumayer was measured approximately once every week with a stake array of 25 stakes covering an area of 25×25 m^2. RACMO output for the land-ice grid box located closest to Neumayer, was interpolated to the same time interval as the measurements. In RACMO, the surface mass balance or net accumulation is defined as the precipitation minus the sublimation. In the model, the net accumulation during an event is somewhat smaller than the stake measurements indicate. It can not be concluded whether differences between RACMO output and Neumayer data are due to transport by wind-blown snow which is locally important but not taken into account in RACMO, or whether this is due to too little precipitation during a model event. Clearly, wind-blown snow plays a role at Neumayer since the net accumulation can be significantly negative during short time intervals. The effect of wind-blown snow on the scale of a model grid box is unknown, but is assumed to be of lesser importance than on the scale of the stake array (25×25 m^2). Generally, large precipitation events are present both in the model output and in the measured time series. A comparison of model output with data from an Automatic Weather Station near Svea, located 300 km from the coast, shows that the episodic nature of precipitation, which is also found at this site, is realistically represented by the model.

2.2 Sensitivity integration

For the 5-year period 1980-1984, a sensitivity integration is performed. In this integration, i) the lateral boundaries are warmed by 2°C, keeping the relative humidity fixed, ii) the sea surface is warmed by 2°C, and iii) a retreat of the sea-ice is prescribed that is consistent with the 2°C warming (Van Lipzig *et al.*, 2002b). The dynamics of the flow at the lateral model boundaries is identical to the control integration and changes in the large-scale dynamics in response to the temperature forcing are not taken into account. On the other hand, using this approach, the model is driven by large-scale flow dynamics inferred from observations, and the occurrence of synoptic systems is constrained to what is observed in the present-day climate.

3 Results

3.1 Variations in annual mean temperature

The δ-signal in an ice core is related to the temperature at which the precipitation is formed. We estimate this temperature by the inversion temperature T_i (Robin, 1977), defined as a local maximum in the stably stratified atmospheric boundary layer temperature profile. As an example, Figure 4 shows the calculated temperature profile for 1 August 1988, 12 UTC at Dome-C. In this case, the inversion occurs at 540 m height and T_i is 27°C higher than the surface temperature (T_s).

Having found that RACMO is capable of realistically representing the episodic nature of precipitation at Neumayer and Svea, we believe the model is suitable for studying the effect of the temporally irregular distribution of net accumulation (B) on signals measured in ice cores. Proxies for meteorological variables are stored in the ice during precipitation events. Therefore, the annual mean δ-signal in the ice core is not directly related to T_i, but rather to the annual mean inversion temperature weighted with the net accumulation:

$$T_{i,w} = \frac{\sum_{j=1,N} T_{i,j}\, B_j}{\sum_{j=1,N} B_j}, \tag{1}$$

where $T_{i,j}$ is the inversion temperature at time j, B_j is the accumulation over 6 hours at time j, and N is the number of 6-hour time intervals per year. (Model output is available every six hours.) Only positive accumulation events are taken into account and when B_j is negative during a 6-hour period, this period is ignored. We have calculated $T_{i,w}$ for five deep drilling sites: Dome-C (123°E, 75.1°S), Dome-F (40°E, 77.3°S), DML05 (0°E, 75.0°S), Byrd (120°W, 80.0°S) and Vostok (107°E, 78.5°S) (Figure 5). $T_{i,w}$ turns out higher than T_i indicating that the correlation between temperature and precipitation is positive: precipitation events occur when relatively warm air from the sea is advected towards the ice sheet (e.g. Noone, 1999; Bromwich, 1988). It has been checked that ignoring unstable conditions alters $T_{i,w}$ by 0.1 K, which is considered negligible.

The year-to-year variability of $T_{i,w}$ is 2 to 3 times larger than the year-to-year variability of T_i. This is due to year-to-year variations in modelled seasonality of accumulation. For example, at Dome-F, 40% of the net accumulation in the model year 1982 occurred during the summer months January and December. In the model year 1983, the seasonality shifted: the maximum of accumulation occurred during the winter month June (14% of the annual accumulation occurred during this month). Due to the shift in seasonality of precipitation, $T_{i,w}$ decreased by 5.3°C from 1982 to 1983, whereas T_i only slightly decreased by 0.2°C.

There is no significant correlation on the 95% confidence level between annual mean $T_{i,w}$ and T_i, with the exception of Byrd. The correlation coefficients for annual mean values are -0.26, 0.40, 0.31, 0.69, and 0.21 for Dome-C, Dome-F, DML05, Byrd and Vostok, respectively. This implies that year-to-year variations in the δ-signal in an ice core are poor indicators of year-to-year variations in the inversion temperature, except for Byrd. Byrd is the site with the highest accumulation (108 mm w.e. a^{-1}, where Dome-C, Dome-F and Vostok receive less than 30 mm w.e. a^{-1}). At Byrd, the number of precipitation events is larger than for the stations in East Antarctica, resulting in a smaller effect of changes in seasonality of precipitation on the signals that are measured in ice cores.

We investigate whether changes in seasonality of precipitation or changes in variations on a daily time scale are responsible for the low correlation between the annual mean $T_{i,w}$ and T_i. When inserting monthly mean values for the inversion temperature and accumulation in

Equation (1), the weighted inversion temperature is found to be 4°C to 8°C lower than $T_{i,w}$ calculated on the basis of 6-hourly means. The reason for this difference is that precipitation and temperature are correlated on short time scales: when monthly mean values are used, the positive correlation between temperature and precipitation on a daily time scale is no longer taken into account. For all drilling sites considered, the difference between annual mean values of $T_{i,w}$, calculated on the basis of monthly means, and annual mean values of $T_{i,w}$, calculated on the basis of 6-hourly means, turns out to be approximately constant in time. From this it is concluded that year-to-year variations in seasonality of precipitation are responsible for both the larger variability of $T_{i,w}$ compared to T_i, and the low correlation between $T_{i,w}$ and T_i. Year-to-year variations in intermittent nature of precipitation on a daily time scale have an insignificant effect. Further investigations show that the effect of changes in the daily cycle of precipitation are also small.

In ice core studies, the δ-signal that is measured along a core as a function of depth is used to derive the surface temperature history at the drilling site. In order to make this conversion, a relation (transfer function) between the δ-signal and T_s is needed. This transfer function is derived for a reference period (present-day climate), by taking snow samples at several sites. The spatial relation between δ and T_s, derived from the snow samples, is assumed to be identical to the temporal relation between δ and T_s. After Krinner *et al.* (1997), we use model output to mimic this procedure: on basis of RACMO-output, the δ-signal in an ice core is simulated and the surface temperature ($T_{s,\mathrm{core}}$) is derived from the δ-signal using a $T_{i,w}/T_s$ transfer function. As an example, we describe the procedure for the drilling site Dome-C.

First, the 14-year mean $T_{i,w}$ is calculated for the period 1980-1993. Model output at

15 grid boxes going from Dome-C towards the north is used to derive the spatial relation between $T_{i,w}$ and T_s (Figure 6) using a least squares method. This spatial relation is assumed to be identical to the temporal relation and can therefore be used to calculate $T_{s,\mathrm{core}}$ from $T_{i,w}$:

$$T_{s,\mathrm{core}} = -227.7 + 1.84\,T_{i,w}. \tag{2}$$

The coefficient in this equation is larger than one, since the inversion strength $(T_i - T_s)$ increases going from the coast into the interior (Figure 7; see also Phillpot and Zillman, 1970; Connolley 1996). The difference between $T_{i,w}$ and T_s is therefore largest in the interior.

Second, for each year $T_{s,\mathrm{core}}$ is calculated by inserting the annual mean value for $T_{i,w}$ in Equation (2). Figure 8 shows the time series of both T_s and $T_{s,\mathrm{core}}$. There is no significant correlation between T_s and $T_{s,\mathrm{core}}$. In addition, the variability of $T_{s,\mathrm{core}}$ is 2.3 times the variability of T_s. The reason for this is: i) the larger interannual variability of $T_{i,w}$ compared to T_s, and ii) the fact that the coefficient $\partial T_{s,\mathrm{core}}/\partial T_{i,w}$ is larger than one. Note that the spatial slope $\partial T_s/\partial T_{i,w}$ in the temperature regime representative for Dome-C is larger than the slope for the entire temperature range considered in Figure 6. If we restrict the evaluation of Equation (2) to the five coldest grid points, the coefficient is found 18% larger. This results in a 18% larger variability in the $T_{s,\mathrm{core}}$ time series, but the correlation between $T_{s,\mathrm{core}}$ and T_s is unaffected.

Results for all five drilling sites are summarised in Table 1, showing that the transfer functions are different for each site. At Dome-C, DML05 and Vostok, the coefficient $\partial T_{s,\mathrm{core}}/\partial T_{i,w}$ is 1.7 to 1.8, whereas this coefficient is only 1.3 at Dome-F and Byrd. This means that applying a transfer function derived for a specific region to a site in another region can result in errors in the interannual variations of the derived temperature up to about 40%. For all five drilling sites considered, the variability of $T_{s,\mathrm{core}}$ is 2 to 3 times the variability of T_s. The overestimation of the year-to-year variations in the temperature signal is not likely to be found in ice core δ measurements, since wind mixing and isotope diffusion smoothen the ice core record. More relevant for ice core studies is that only for one of the five sites considered, the correlation between modelled T_s and $T_{s,\mathrm{core}}$ is significant at the 95% confidence level (Table 1), implying that, at most drilling sited considered, the δ-signal in an ice core is a poor indicator for the interannual variations in surface temperature.

3.2 Variations in 7-year mean temperature

We now examine whether the temporal variability of precipitation affects the proxy or tracer in the ice core when averaged over longer periods than one year. We consider all grounded ice grid boxes. Instead of net accumulation, we use precipitation, since in some areas the modelled sublimation is unrealistically high due to an overestimation of the roughness length (Van Lipzig, 2002a). Note that the five stations considered in the previous paragraph are not located in regions where sublimation is overestimated. Since the integration covers a period of 14 years, we divide the time series in seven years with the highest surface temperature averaged over the entire ice sheet (1980, 1981, 1984, 1988, 1990, 1991, 1992) and seven years with the lowest surface temperature (1982, 1983, 1985, 1986, 1987, 1989, 1993). The difference $\langle \Delta T_s \rangle$ in surface temperature averaged over the grounded ice between the warm years τ_{warm} and the cold years τ_{cold} was 0.9°C. Note that the brackets indicate averaging in space.

We consider τ_{warm} as the reference years. To calculate the surface temperature during τ_{cold}, in analogue to the interpretation of δ-signals from ice cores, we follow the same procedure as described in the previous paragraph. The 7-year mean $T_{i,w}$ is calculated for the reference years τ_{warm}. For simplicity, one transfer function is derived for the entire grounded ice sheet (Figure 9), using a least squares method:

$$T_{s,\text{core}} = -160.19 + 1.59\,T_{i,w}. \tag{3}$$

Like in Equation (2), the coefficient $\partial T_{s,\text{core}}/\partial T_{i,w}$ is larger than one, since the inversion strength increases when going from the coast into the interior (see Figure 7). To study whether there are other effects that play a role, the relation between the surface temperature weighted with the precipitation ($T_{s,w}$) and T_s is calculated. The coefficient $\partial T_{s,\text{core}}/\partial T_{s,w}$ is slightly larger than one (1.1) due to the fact that in the interior the maximum in the precipitation occurs during summer, whereas the season with maximum precipitation varies along the coast. Therefore, $T_{s,w} - T_s$ is larger in the interior (at low temperatures) than near the coast. It is unclear whether the modelled summer precipitation maximum in the interior is realistic due to lack of reliable climatological precipitation measurements in this region, where the accumulation is low.

To give an indication of the regional differences in transfer function, Figure 10 shows the derived surface temperature ($T_{s,\text{core}}$) calculated from Equation (3) minus T_s for the reference years τ_{warm}. In regions where the transfer function deviates from Equation (3), $T_{s,\text{core}}$ - T_s is large. In the interior, $T_{s,\text{core}}$ is larger than T_s. This is consistent with the deviation of $T_{i,w}$ from the linear regression line for the lower temperature range (interior) (Figure 9). Other areas where $T_{s,\text{core}}$ is larger than T_s are Oates coast and the area east of the Ross Ice Shelf. Areas where $T_{s,\text{core}}$ is smaller than T_s are the interior of West Antarctica and the region east of the Amery Ice Shelf.

The value for $T_{i,w}$ is calculated for the years τ_{cold}. The surface temperature during τ_{cold}, simulating the temperature found from the δ-signal in an ice core, is derived by inserting $T_{i,w}(\tau_{\text{cold}})$ in Equation (3). The difference between $T_{s,\text{core}}$ and T_s for the years τ_{cold} is similar to the difference for the years τ_{warm}, indicating that the regional differences in transfer function during τ_{cold} are similar to the regional differences during τ_{warm}.These regional differences are mainly caused by regional differences in the relation between T_s and T_i.

The difference $\langle \Delta T_{s,\text{core}} \rangle$ between τ_{warm} and τ_{cold} derived with Equation (3) averaged over the grounded ice is 0.9°C, which is identical to the value found for $\langle \Delta T_s \rangle$. However, the spatial pattern of $\Delta T_{s,\text{core}}$ is very different from ΔT_s (Figure 11). The spatial variability of $\Delta T_{s,\text{core}}$ is much larger than the spatial variability of ΔT_s. At 30% of the grid boxes, $\Delta T_{s,\text{core}}$ is negative, whereas ΔT_s is negative at only 3% of the grid boxes. In addition, at 10% of the grid boxes $\Delta T_{s,\text{core}}$ is larger than 3°C. This temperature difference never occurs for T_s. The large spatial variability of $\Delta T_{s,\text{core}}$ is caused by differences in dynamics of the flow between τ_{warm} and τ_{cold}, affecting the 7-year mean seasonality of precipitation and consequently $T_{i,w}$.

Most of the stations considered are located in a region where $\Delta T_{s,\text{core}}$ is smaller than ΔT_s, but for Byrd $\Delta T_{s,\text{core}}$ is larger than ΔT_s (Table 1). Using the regional coefficients in the transfer function given in the first column of Table 1 instead of the coefficient in Equation (3) can result in a $\pm 20\%$ change of $\Delta T_{s,\text{core}}$. The correspondence between $\Delta T_{s,\text{core}}$ and ΔT_s improves for Byrd but deteriorates for Dome-C. The mean absolute difference between $\Delta T_{s,\text{core}}$ and ΔT_s for the five stations considered does not change using the regional coefficients. The mean absolute difference $\frac{1}{M}\sum_M |\Delta T_{s,\text{core}} - \Delta T_s|$, where M is the number of grounded land ice grid boxes, is 1.2°C, which is larger than the mean temperature difference between τ_{warm} and τ_{cold}. This indicates that, in ice core studies, an averaging period of seven years is too short to relate the ice core δ-signal to surface temperature.

3.3 Changes in the relation between inversion strength and surface temperature

A 5-year sensitivity integration is performed in which a temperature forcing of 2°C is prescribed at the lateral boundaries of the model domain and at the sea surface together with a retreat of the sea ice. We have used the results of this integration to study the effect of changes in the T_i/T_s relation on the temperature derived from the δ-signal in the ice. The T_i/T_s relation changes when the relation between inversion strength and surface temperature changes in a different climatic regime. The inversion strength is smaller in the sensitivity integration SENS than in the control integration CTL. In addition, the surface temperature averaged over the grounded ice is 3.4°C warmer in SENS than in CTL. The response of T_s is larger than the applied temperature forcing of 2°C. This is caused by the water vapour feedback: the amount of water vapour and specific liquid water increases in the sensitivity integration, resulting in an increase in downward longwave radiation (Van Lipzig *et al.*, 2002b). The amplification of the 2°C temperature forcing and the decrease in inversion strength are largest in the interior of the ice sheet, where the surface elevation is largest.

Although the forcings in the integration are very simple, the study is useful for identifying the mechanisms that can cause a difference between surface temperature and derived surface temperature using the δ versus T_s relationship (transfer function). Changes in the seasonality of precipitation between the two integrations are expected to be small, since changes in the circulation at the lateral model boundaries are not taken into account.

The weighted temperature and the transfer function are derived for CTL:

$$T_{s,\text{core}} = -154.15 + 1.56\, T_{i,w}. \tag{4}$$

This equation slightly differs from Equation (3), since a different time period (1980-1985) is considered. Inserting $T_{i,w}$ from SENS in Equation (4) yields the derived temperature $T_{s,\text{core}}(\text{SENS})$. Figure 12 shows $\Delta T_{s,\text{core}} = T_{s,\text{core}}(\text{SENS}) - T_{s,\text{core}}(\text{CTL})$ and $\Delta T_s = T_s(\text{SENS}) - T_s(\text{CTL})$. Again, the spatial variability of $\Delta T_{s,\text{core}}$ is much larger than the variability of ΔT_s, due to local changes in the 5-year mean seasonality of precipitation. Probably, the changes in seasonality averaged over the integration period decrease with the length of the integration, so the spatial variability of $\Delta T_{s,\text{core}}$ might be smaller for longer integration periods.

The increase in $T_{s,\text{core}}$ averaged over the grounded ice is 4°C; $\langle \Delta T_{s,\text{core}} \rangle$ is 18% larger than $\langle \Delta T_s \rangle$. We separate the effect of changes in the T_i/T_s relation from changes in seasonality of precipitation by multiplying the difference in inversion temperature between SENS and CTL (2.7°C) by $\partial T_{s,\text{core}}/\partial T_{i,w}$ (1.56). We find a value of 4.2°C, which corresponds closely to the value found when both changes in seasonality and inversion strength are included. This indicates that changes in the T_i/T_s relation are primarily responsible for the difference between $\langle \Delta T_{s,\text{core}} \rangle$ and $\langle \Delta T_s \rangle$.

There are two opposing effects that cause the difference between $\langle \Delta T_{s,\text{core}} \rangle$ and $\langle \Delta T_s \rangle$. First, the meridional gradient in inversion strength results in a coefficient $\partial T_{s,\text{core}}/\partial T_{i,w}$ larger than one, amplifying the difference between SENS and CTL in $T_{i,w}$. Second, the inversion strength in SENS is smaller than in CTL. Therefore, the increase in T_i is smaller than the increase in T_s. The first effect dominates.

Were the temporal relation between T_s and $T_{i,w}$ identical to the spatial relation, then the transfer function for the SENS-integration would be identical to Equation (4). Since this is not the case, the transfer function for the SENS-integration ($T_{s,\mathrm{core}} = -143.55 + 1.52\,T_{i,w}$) differs from the CTL-integration. The difference in $\partial T_{s,\mathrm{core}}/\partial T_{i,w}$ between CTL and SENS is significant on the 99% confidence level. The SENS-value for $\partial T_{s,\mathrm{core}}/\partial T_{i,w}$ is smaller than the CTL-value since the increase in inversion strength going from the coast into the interior is smaller in SENS than in CTL: in CTL, $\partial T_s/\partial T_i$ is 1.49, whereas in SENS this value is 1.43. In both integrations 96% of the variance in T_i is explained by a linear relation between T_s and T_i.

We conclude that the increase in simulated surface temperature, being derived with a method analogous to those used in ice core studies, is found to be overestimated by 18% when compared to the direct model surface temperature. This error is of the same order of magnitude as the errors found by Delaygue *et al.* (2000) for a change in seasonality of precipitation (resulting in an underestimation of 15%) and a change in the temperature of the source, where evaporation took place (resulting in an overestimation of 10% to 30%). These errors are all smaller than the 100% underestimation of the temperature difference between last glacial maximum and present-day climate when comparing the method using the spatial δ/T_s slope with borehole paleothermometry for Greenland (Cuffey *et al.*, 1995; Johnsen *et al.*, 1995).

4 Discussion

From a 14-year integration, we found no significant correlation between the modelled surface temperature and the modelled inversion temperature weighted with the surface mass balance (r varies from -0.26 to 0.40) for the stations Dome-C, Dome-F, DML05 and Vostok. Only for the station Byrd the correlation coefficient is significant at the 95% confidence level (r=0.69).

For both the Antarctic and the Greenland ice sheet, measured annual mean surface temperature (T_s) variations have been compared with δ-signals from ice cores. The results are ambiguous. The Greenland Summit δ-signals are more closely related to the accumulation-weighted temperature at Jakobshavn than to the mean annual temperature (Steig *et al.*, 1994). However, good correlation has been found between δ-signals and the unweighted surface temperature derived from satellite measurements (Shuman *et al.*, 2001) and between δ, averaged over six cores, and the local temperature over the past century (White *et al.*, 1997) ($r = 0.47$).

In the Antarctic Peninsula area, a reasonable correlation has been found between the annual mean δ-signals and measured T_s (Aristarain *et al.*, 1986; Peel *et al.*, 1988). On the other hand, at Neumayer, T_s is not directly correlated with δ (Schlosser, 1999). Isaksson and Karlén (1994) report little year-to-year correlation between the temperature record at the Antarctic station Halley and the δ record from cores drilled on the ice shelf and in the escarpment area below 2 km height in Western Dronning Maud Land, but they find a better correlation for higher altitude cores, where accumulation is low. At South Pole station, measured summer temperatures are significantly correlated to δ-maxima ($r = 0.75$) but winter temperatures are not significantly correlated to the δ-minima ($r = 0.06$). The annual mean values significantly correlate on the 97% confidence level ($r = 0.47$) (Jouzel *et al.*, 1983).

Model output is used to study the relation between $T_{s,\mathrm{core}}$ and T_s for the sites discussed above. Only at the station Faraday in the Antarctic Peninsula, the correlation is significant at the 99% confidence level ($r = 0.81$). At South Polar station, the correlation coefficient for annual mean values is 0.52, which is significant at the 90% confidence level. At the other sites considered (Neumayer and the sites near Halley), the correlation is not significant at the 95% level (r varies from -0.01 to 0.31).

In summary, at several sites the correlation between annual mean δ-signals and measured T_s is insignificant, but at other sites the signals are significantly correlated. This correlation is not directly related to the total annual amount of accumulation. The large spread in observed correlations between δ-signals from ice cores and measured surface temperatures is qualitatively represented by the model. A quantitative agreement is not expected since the length of the time series differs and snowdrift is not taken into account in the model. Generally longer time series of temperature and surface mass balance at the drilling sites at high temporal resolution (hourly) are necessary to evaluate the RACMO in better detail. This kind of observations is available for short periods only and virtually absent for longer periods of at least a decade.

An example of a study on a 1-year period is discussed by McMorrow *et al.* (2001). They measured temperature and the surface mass balance at the drilling site at high temporal resolution (several times per day) and compare the unweighted surface temperature and the surface temperature weighted with the surface mass balance to the isotope ratio measured in ice/firn that is deposited during the period when the AWS was operational. They conclude that $\delta^{18}O$-values measured during precipitation events correlate closely to the measured temperatures during that event. In addition, they find that the $\partial\delta^{18}O/\partial T_s$-slope, derived on an event-by-event basis is half of the slope treating the ice core as a continuous record. However, longer time series are necessary to conclude whether such results are significant.

In addition to studying the modelled annual mean values for $T_{s,\mathrm{core}}$ and T_s, we have analysed two sets of model output, namely τ_{warm} versus τ_{cold}, and SENS versus CTL. Seven years with high surface temperatures (τ_{warm}) are compared with seven years with low temperatures (τ_{cold}). There is no externally imposed forcing and the temperature difference between the two groups of years is caused by differences in the temperature over sea/sea-ice (the sea-ice is 0.2 ° warmer during τ_{warm} than during τ_{cold}) and by differences in large-scale atmospheric circulation. During τ_{warm} the atmospheric meridional heat exchange is more efficient than during τ_{cold}. This comparison is suitable for studying the signals that can be expected in ice formed during different atmospheric flow regimes on decadal time scales.

The other comparison is essentially different. In the sensitivity integration SENS, an external temperature forcing is prescribed. This integration is compared with the control integration CTL. The integrations SENS and CTL are relevant to study signals that can be expected in ice formed during different externally forced climate regimes. The dynamics of the flow at the lateral boundaries is unchanged, but in the interior of the model domain, the flow adjusts to the modified forcing.

For both sets of model output (τ_{warm}/ τ_{cold} and SENS/CTL) the value for $T_{s,\mathrm{core}}$ has been calculated, representing the model analogue of the surface temperature derived from the δ-signal in an ice core. A more general form of Equation (2) is given by

$$T_{s,\text{core}}(t,x,y) = C + \Big(\frac{\partial T_{i,w}}{\partial T_s}\Big)^{-1}_{t=\text{REF}} T_{i,w}(t,x,y), \tag{5}$$

where x and y are the zonal and meridional coordinates along the surface of the ice sheet, t is the time and C is a constant derived from the linear regression between $T_{i,w}$ and T_s. The variable between brackets is the spatial $T_{i,w}/T_s$ slope for the reference climate (REF), where $T_{i,w}$ is given in Equation (1). To study variations of $T_{s,\text{core}}$ in time, the partial derivative of Equation (5) to the surface temperature is taken keeping x and y fixed:

$$\Big(\frac{\partial T_{s,\text{core}}}{\partial T_s}\Big)_{x,y} = \Big(\frac{\partial T_{i,w}}{\partial T_s}\Big)^{-1}_{t=\text{REF}} \Big(\frac{\partial T_{i,w}}{\partial T_s}\Big)_{x,y}. \tag{6}$$

There are only two time groups of years for each set of model output, namely τ_{warm} and τ_{cold} or SENS and CTL. We write the temporal X/T_s slope of an arbitrary variable X as:

$$\Big(\frac{\partial X}{\partial T_s}\Big)_{x,y} = \frac{\Delta X}{\Delta T_s}, \tag{7}$$

where ΔX and ΔT_s are the differences in X and T_s between τ_{warm} and τ_{cold} or SENS and CTL.

When seasonal variations in precipitation are ignored, $\partial T_{i,w}/\partial T_s$ can be approximated by $\partial T_i/\partial T_s$ yielding:

$$\frac{\langle \Delta T_{s,\text{core}}\rangle}{\langle \Delta T_s\rangle} = \Big(\frac{\partial T_i}{\partial T_s}\Big)^{-1}_{t=\text{REF}} \frac{\langle \Delta T_i\rangle}{\langle \Delta T_s\rangle}. \tag{8}$$

In the SENS/CTL comparison, the spatial T_i/T_s slope is 0.64, whereas the temporal slope averaged over the ice sheet is 0.8. In other words, the decrease in inversion strength as a function of surface temperature is larger in space than in time. This difference between spatial and temporal slope is the primary cause for the overestimation of $\langle \Delta T_{s,\text{core}}\rangle$ compared to $\langle \Delta T_s \rangle$. This effect, combined with the smaller effect of the seasonality of precipitation results in an 18% overestimation of $\langle \Delta T_{s,\mathrm{core}} \rangle$ compared to $\langle \Delta T_s \rangle$.

Although the prescribed temperature forcing at the lateral boundary of the model domain is constant with height, we believe that the increase in meridional gradient in inversion strength is realistic. The inversion is mainly the result of the net radiative heat loss (Phillpot and Zillman, 1970) and therefore unlikely to be sensitive to conditions near the surface at 50°-60°S. In addition, Van Lipzig *et al.* (2002b) explain the mechanism for the amplification of the temperature forcing near the ice sheet surface. The humidity and liquid water in the atmosphere increases resulting in an increase in the downward long wave radiation. Therefore, the increase in near surface temperature is larger than the increase higher in the atmosphere, resulting in a weakening of the inversion strength.

Interestingly, the temporal T_i/T_s slope in the $\tau_{\mathrm{warm}}/\tau_{\mathrm{cold}}$ comparison is identical to the SENS/CTL comparison. This indicates that the weakening of the inversion strength in a warmer climate is independent of whether the forcing is internally (atmospheric flow) or externally (temperature forcing) generated. Like in the SENS/CTL comparison, the spatial T_i/T_s slope is smaller than the temporal slope, resulting in an overestimation of $\langle \Delta T_{s,\mathrm{core}} \rangle$. However, in the $\tau_{\mathrm{warm}}/\tau_{\mathrm{cold}}$ comparison, changes in the seasonality of precipitation induced by a different atmospheric flow regime play an important role. During τ_{cold}, the winter accumulation averaged over the grounded ice is 3% lower and the summer accumulation is 3% higher than during τ_{warm}, resulting in an underestimation of $\langle \Delta T_{s,\mathrm{core}} \rangle$. Averaged over the grounded ice, these two effects compensate for each other, with the result that $\langle \Delta T_{s,\mathrm{core}} \rangle / \langle \Delta T_s \rangle$ is approximately one. Locally, $\Delta T_{s,\mathrm{core}}$ deviates largely from ΔT_s due to changes in seasonality of precipitation between τ_{warm} and τ_{cold}. For example at Dome-C, the modelled surface temperature decreases with 1.1°C during τ_{cold}, whereas the simulated model surface temperature derived from the δ-signal in an ice core ($\Delta T_{s,\text{core}}$) is found to *increase* by 1.5°C during τ_{cold}.

5 Conclusions

Results from an integration with a regional atmospheric model, driven from the lateral boundaries by ERA-15 and from the sea surface by observed sea surface temperatures and sea-ice extent, are used to study the effect of temporal variability of precipitation and changes in the relation between temperature inversion strength and surface temperature on the proxy for temperature (δ) in an ice core. The δ-signal in an ice core has often been assumed to be related to the surface temperature (T_s), but it is physically better related to $T_{i,w}$: the temperature at which precipitation is formed (i.e. approximated by the inversion temperature T_i) weighted with the net accumulation (precipitation minus sublimation) at the surface. Only for Byrd there is a significant correlation at the 95% confidence level between $T_{i,w}$ and T_s, whereas for Dome-C, Dome-F, DML05, and Vostok the correlation is not significant. For all sites considered, the year-to-year variability of $T_{i,w}$ is 2 to 3 times the variability of T_i, due to year-to-year variations in the seasonality of precipitation.

In analogy to ice core studies, where a spatial δ/T_s relation is used to derive time variations in T_s from ice core δ-signals, we derive the spatial $T_{i,w}/T_s$ relation (transfer function) from 14-year mean model output. Annual mean values for $T_{i,w}$ are inserted in the transfer function to obtain time series of surface temperature ($T_{s,\mathrm{core}}$), as if found from the δ signal in an ice core (method by Krinner *et al.* (1997)). For four out of the five sites considered there is no significant correlation at the 95% confidence level between $T_{s,\mathrm{core}}$ and T_s. The temporal variability of $T_{s,\mathrm{core}}$ is found to be 2 to 3 times the variability of T_s. We therefore conclude that, at most deep drilling sites considered, annual mean δ-values derived from ice cores are poor indicators of year-to-year variations in the surface temperature, since the annual mean $T_{i,w}$ cannot be related to T_s, without detailed knowledge on the seasonality of precipitation.

The effect of variations in precipitation on the 7-year mean $T_{s,\mathrm{core}}$, is studied by dividing the 14-year integration period into warm and cold years (τ_{warm} and τ_{cold}). The spatial $T_{i,w}/T_s$ relation (transfer function) is derived for τ_{warm}. The values for $T_{s,\mathrm{core}}(\tau_{\mathrm{warm}})$ and $T_{s,\mathrm{core}}(\tau_{\mathrm{cold}})$ are calculated by inserting $T_{i,w}$ for both groups of years separately in the transfer function. It is found that the spatial pattern of $\Delta T_{s,\mathrm{core}} = T_{s,\mathrm{core}}(\tau_{\mathrm{warm}}) - T_{s,\mathrm{core}}(\tau_{\mathrm{cold}})$ and $\Delta T_s = T_s(\tau_{\mathrm{warm}}) - T_s(\tau_{\mathrm{cold}})$ are significantly different: the spatial variability of $\Delta T_{s,\mathrm{core}}$ is larger than the spatial variability of ΔT_s. The mean absolute difference between $\Delta T_{s,\mathrm{core}}$ and ΔT_s, for all grounded land ice grid boxes, is 1.2°C, which is larger than the mean temperature difference over the grounded ice between the two groups of years (0.9°C). These results indicate that ice core δ-signals averaged over periods of seven years or shorter are poorly related to surface temperature variations. Longer integrations are necessary to study the relation between $T_{s,\mathrm{core}}$ and T_s on time scales beyond seven years.

The effect of changes in the relation between temperature inversion strength and surface temperature in a different climatic regime is studied with results from a sensitivity integration (SENS) in which a temperature forcing of 2°C is prescribed. It is found that $\langle \Delta T_{s,\mathrm{core}} \rangle = \langle T_{s,\mathrm{core}}(\mathrm{SENS}) - T_{s,\mathrm{core}}(\mathrm{CTL}) \rangle$ is 18% larger than $\langle \Delta T_s \rangle = \langle T_s(\mathrm{SENS}) - T_s(\mathrm{CTL}) \rangle$. There are two opposing effects that cause a discrepancy between $\langle \Delta T_{s,\mathrm{core}} \rangle$ and $\langle \Delta T_s \rangle$. First of all, the inversion strength decreases in space as a function of the surface temperature resulting in a spatial T_i/T_s slope smaller than one (0.64). Therefore the coefficient in the transfer function $\partial T_{s,\mathrm{core}}/\partial T_{i,w}$ is larger than one. Second, the inversion is weaker in SENS than in CTL, resulting in a temporal T_i/T_s slope smaller than one. In our integration, the former effect dominates.

Although the temperature forcing in the integration is very simple, the mechanisms behind the discrepancy between $\langle \Delta T_{s,\mathrm{core}} \rangle$ and $\langle \Delta T_s \rangle$ might also play a role in deriving temperatures for a different climatic regime from the δ-signals in ice cores. Not taking into account ice age strengthening of the inversion during the last glacial maximum has been identified as a possible cause for an underestimation of the surface temperature difference between last glacial maximum and present-day climate derived from the δ signal. In our integrations, the spatial T_i/T_s slope (0.64) is smaller than the temporal slope (0.80) and the spatial $T_{i,w}/T_s$ slope (0.64) is smaller than the temporal slope (0.76). This difference in spatial and temporal slope would result in an 18% *overestimation* of the surface temperature difference between last glacial maximum and present-day climate derived from the δ signal from Antarctic regions.

Acknowledgements

We would like to thank scientists from KNMI and IMAU, especially Michiel van den Broeke, Roderik van der Wal and Carleen Reijmer for useful discussions and Prof. E. Steig and an anonymous reviewer for valuable comments which substantially improved the manuscript. This work was supported by the Netherlands Earth and Life Sciences Foundation (ALW) and the National Computing Facilities Foundation (NCF) for the use of supercomputer facilities with financial aid from the Netherlands Organisation for Scientific Research (NWO).

This work is a contribution to the "European Project for Ice Coring in Antarctica"

(EPICA), a joint ESF (European Science Foundation)/EC scientific programme, funded by the European Commission and by national contributions from Belgium, Denmark, France, Germany, Italy, the Netherlands, Norway, Sweden, Switzerland and the United Kingdom. This is EPICA publication no. NN.

REFERENCES

Aristarain, A.J., J. Jouzel, M. Pourchet, 1986. Past Antarctic Peninsula climate (1850-1980) deduced from an ice core isotope record. *Climate Change*, **8**(1), 69-89.

Boyle, E.A., 1997. Cool tropical temperatures shift the global $\delta^{18}O - T$ relationship: An explanation for the ice core $\delta^{18}O$-borehole thermometry conflict? *Geophys. Res. Lett.*, **24(3)**, 273-276.

Bromwich, D.H., 1988. Snowfall in high southern latitudes. *Rev. Geophys.*, **26**(1), 149-168.

Charles, C., D. Rind, J. Jouzel, R. Koster, and R. Fairbanks, 1994. Glacial interglacial changes in moisture sources for Greenland: Influences on the ice core record of climate. *Science*, **261**, 508-511.

Christensen J.H. and E. Van Meijgaard, 1992. On the construction of a Regional Atmospheric Climate Model. KNMI Tech. Rep. 147, Royal Netherlands Meteorological Institute, de Bilt, The Netherlands, 22 pp.

Christensen, J.H., O.B. Christensen, P. Lopez, E. Van Meijgaard and M. Botzet, 1996. The HIRHAM4 Regional Atmospheric Climate Model. DMI Scientific Rep. 96-4, Danish Meteorological Institute, Copenhagen, Denmark, 51 pp.

Connolley, W.M., 1996. The Antarctic temperature inversion. *Int. J. Climatol.*, **16**(12), 1333-1342.

Cuffey, K.M., G.D. Clow, R.B. Alley, M. Stuiver, E.D. Waddington and R.W. Saltus, 1995. Large Arctic temperature change at the Wisconsin-Holocene glacial transition. *Science*,

270, 455-458.

Cuffey, K.M., and G.D. Clow, 1997. Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition. *J. Geophys. Res.*, **102(C12)**, 26383-26396.

Dansgaard, W., 1964. Stable isotopes in precipitation. *Tellus*, **16 (4)**, 436-468.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and N. Gundestrup, 1973. Stable isotope glaciology. *Medd. Grønl.*, **197 (2)**.

Delaygue, G., J. Jouzel, V. Masson, R.D. Koster, E. Bard, 2000 Validity of the isotopic thermometer in central Antarctica: limited impact of glacial precipitation seasonality and moisture origin. *Geophys. Res. Lett.*, **27 (17)**, 2677-2680.

Fisher, D.A., 1991. Remarks on the deuterium excess in precipitation in cold regions. *Tellus*, **43B**, 401-407.

Gustafsson, N., 1993. The HIRLAM 2 final report. HIRLAM Tech. Rep. 9, Swedish Meteorological and Hydrological Institute, Norrköping, Sweden.

Isaksson, E. and W. Karlén, 1994. High resolution climatic information from short firn cores, western Dronning Maud Land, Antarctica. *Climate Change*, **26**, 421-424.

Johnsen, S.J., D. Dahl-Jensen, W. Dansgaard and N. Gundestrup, 1995. Greenland paleotemperatures derived from GRIB bore hole temperature and ice core isotope profiles. *Tellus*, **47B**, 624-629.

Jouzel, J., R.B. Alley, K.M. Cuffey, W. Dansgaard, P. Grootes, G. Hoffmann, S.J. Johnsen, R.D. Koster, D. Peel, C.A. Shuman, M. Stievenard, M. Stuiver and J. White, 1997. Validity of the temperature reconstruction from water isotopes in ice cores. *J. Geophys. Res.*, **102 (C12)**, 26471-26487.

Jouzel, J., L. Merlivat, J.R. Petit and C. Lorius, 1983. Climatic information over the last century deduced from a detailed isotopic record in the south pole snow. *J. Geophys. Res.*, **88**(C4), 2693-2703.

Krinner, G., C. Genthon and J. Jouzel, 1997. GCM analysis of local influences on ice core δ signals. *Geophys. Res. Lett.*, **24 (22)**, 2825-2828.

Lorius, C. and L. Merlivat, 1977. Distribution of mean surface stable isotope values in East Antarctica: Observed changes with depth in a coastal area. In: *Isotopes and Impurities in Snow and Ice, IAHS Publ.*, **118**, 125-137.

McMorrow A.J., M.A.J. Curran, T.D. Van Ommen, V. Morgan, M.J. Pook and I. Allison, 2001. Intercomparison of firn core and meteorological data. *Antarct. Sci.*, **13**(3), 329-337.

Noone, D., J. Turner and R. Mulvaney, 1999. Atmospheric signals and characteristics of accumulation in Dronning Maud Land, Antarctica. *J. Geophys. Res.*, **104(D16)**, 19191-19211.

Peel, D.A., R. Mulvaney and M. Davison, 1988. Stable-isotope/air-temperature relationships in ice cores from Dolleman Island and the Palmer Land plateau, Antarctic Peninsula. *Annals of Glaciol.*, **10**, 130-136.

Peel, D.A., R. Mulvaney, E.C. Pasteur, and C. Chenery, 1996. Climate changes in the Atlantic sector of Antarctica over the past 500 years from ice-core and other evidence. In: *Climatic variations and forcing mechanisms of the last 2000 years*, P.D. Jones, R.S. Bradley, and J. Jouzel, Eds., Springer-Verlag, Berlin Heidelberg, 243-262.

Phillpot, H.R., J.W. Zillman, 1970. The surface temperature inversion over the Antarctic continent. *J. Geophys. Res.*, **75**(21), 4161-4169.

Reijmer and Van den Broeke, 2001. Moisture sources of precipitation in Western Dronning Maud Land, Antarctica. *Antarctic Science*, **13 (2)**, 210-220.

Robin, G.d.Q., 1977. Ice cores and climatic changes. *Philos. Trans. R. Soc. London, Ser. B.*, **280**, 143-168.

Robin, G.d.Q., 1983. Isotopic temperature $(d-q)$ noise. In: *Climatic Record in Polar Ice Sheets*, G.d.Q. Robin, Ed., Cambridge Univ. Press, New York, 184-189.

Roeckner, E., K. Arpe, L. Bengtsson, M. Christoph, M. Claussen, L. Dümenil, M. Esch, M. Giorgetta, U. Schlese and U. Schulzweida, 1996. The atmospheric general circulation model ECHAM-4: Model description and simulation of present-day climate. MPI Rep. 218, Max-Planck Institut für Meteorologie, Hamburg, Germany, 90 pp.

Schlosser, 1999. Effects of seasonal variability of accumulation on yearly mean $\delta^{18}O$ values in Antarctic snow. *J. Glaciol.*, **45(151)**, 463-468.

Schlosser, E., N.P.M. van Lipzig and H. Oerter, 2002. Temporal variability of accumulation at Neumayer Station, Antarctica from stake array measurements and a regional atmospheric model. *J. Glac.*, **48(160)**, 87-94.

Shuman C.A., D.H. Bromwich, J. Kipfstuhl and M. Schwager, 2001. Multiyear accumulation and temperature history near the North Greenland Ice Core Project site, north central Greenland. *J. Geophys. Res.*, **106**(D24), 33853-33866.

Steig, E.J., P.M. Grootes, and M. Stuiver, 1994. Seasonal precipitation timing and ice core records, *Science*, **266**, 1885-1886.

Van Lipzig, N.P.M., 1999. The surface mass balance of the Antarctic ice sheet: a study with a regional atmospheric model. PhD thesis, Institute for Marine and Atmospheric research Utrecht, Utrecht University, The Netherlands, 154 pp.

Van Lipzig N.P.M., E. Van Meijgaard and J. Oerlemans 1999. Evaluation of a regional atmospheric model using measurements of surface heat exchange processes from a site in Antarctica. *Mon. Wea. Rev.*, **127**, 1994-2011.

Van Lipzig, N.P.M., E. van Meijgaard and J. Oerlemans, 2002a. The spatial and temporal variability of the surface mass balance in Antarctica: results from a regional climate model. *Int. J. Climatol.*, **22**, 1197-1217.

Van Lipzig, N.P.M., E. van Meijgaard and J. Oerlemans, 2002b. Temperature sensitivity of the Antarctic surface mass balance in a regional atmospheric climate model. *J. Clim.*, **15(19)**, 2758-2774.

Werner, M., U. Mikolajewicz, M. Heimann, and G. Hoffmann, 2000. Borehole versus isotope temperatures on Greenland: Seasonality does matter. *Geophys. Res. Lett.*, **27(5)**, 723-276.

White, J.W.C., L.K. Barlow, D. Fisher, P. Grootes, J. Jouzel, S.J. Johnsen, M. Stuiver, and H. Clausen, 1997. The climate signal in the stable isotopes of snow from Summit, Greenland: Results of comparisons with modern climate observations. *J. Geophys. Res.*, **102**(C12), 26425-26439.

ILLUSTRATIONS AND TABLES

Figure 1: *Regions and stations mentioned in the text.*

Figure 2: *Temperature (dashed line) and precipitation (solid line) at (a) Dome-C, (b) Dome-F, and (c) DML05, calculated with* RACMO *for August 1988.*

Figure 3: *Net accumulation rate during the year 1987 at Neumayer station measured with a stake array (Schlosser, 1999) (solid line) and calculated with* RACMO *(dashed line) (see also Schlosser et al. (2002)). The times at which the measurements were made are indicated with a dot near the upper axis. Model output is interpolated to the time interval of the measurements.*

Figure 4: *The temperature profile at Dome C for 1 August 1988, 12 UTC. The inversion temperature is indicated with* T_i *and the surface temperature with* T_s*. The dotted line shows the extrapolated free atmospheric temperature profile. In the atmospheric boundary layer, the temperature deviates substantially from the extrapolated free atmospheric temperature. The squares refer to model levels.*

Figure 5: *Time series of annual mean inversion temperature (solid line) and annual mean inversion temperature weighted with the net accumulation (*$T_{i,w}$*) (dotted line) at (a) Dome-C, (b) Dome-F, (c) DML05, (d) Byrd, and (e) Vostok.*

Figure 6: *Fourteen-year mean surface temperature (T_s) versus the 14-year mean inversion temperature weighted with the net accumulation ($T_{i,w}$) for 15 grid boxes on a line going from Dome-C towards the north. The solid line in the graph shows the transfer function (Equation (2)) calculated with the least squares method.*

Figure 7: *Fourteen-year mean temperature inversion strength ($T_i - T_s$), derived from model output.*

Figure 8: *Annual mean T_s (solid line) and $T_{s,\mathrm{core}}$ (dotted line) for Dome-C. $T_{s,\mathrm{core}}$ is derived from model output with Equation (2) using a method (Krinner et al., 1997) analogous to the retrieval of T_s from the δ-signal observed in ice cores.*

Figure 9: *Seven-year mean T_s versus 7-year mean inversion temperature weighted with the precipitation ($T_{i,w}$) for all grounded ice points for the years τ_{warm}. The solid line shows the transfer function (Equation (3)) calculated using the least squares method.*

Figure 10: *Difference between the 7-year mean $T_{s,\mathrm{core}}$, derived with Equation (3), and the 7-year mean T_s for the years τ_{warm}.*

Figure 11: *Difference between τ_{warm} and τ_{cold} in (a) $T_{s,\mathrm{core}}$, derived using Equation (3), and (b) T_s.*

Figure 12: *Difference between* SENS *and* CTL *in (a) $T_{s,\mathrm{core}}$, derived using Equation (4), and (b) T_s.*

Table 1: *Characteristics that are relevant for the interpretation of ice core signals are given for five drilling sites. The meaning of the columns is* **i)** $\partial T_{s,\mathrm{core}}/\partial T_{\mathrm{i,w}}$: *the slope of the linear relation between* $T_{s,\mathrm{core}}$ *and* $T_{\mathrm{i,w}}$. *To calculate* $\partial T_{s,\mathrm{core}}/\partial T_{\mathrm{i,w}}$, *15 grid boxes are selected going from the drilling site towards the north for Dome-C, Dome-F and Vostok. For DML05, grid boxes along a line going from the point 23° E; 74.6° S towards 0° E; -74.8° S are selected. For Byrd, grid boxes along a line going from the point 119° W; -79.7° S towards 106° W; 75.5° S are selected. These directions are chosen in relation with the local orography. Grid boxes where the roughness length* z_0 *is larger than 1.2 mm are excluded;* **ii)** $r_{spatial}(T_s, T_{\mathrm{i,w}})$: *the spatial correlation between 14-year mean* T_s *and* $T_{\mathrm{i,w}}$ *for the selected grid boxes along the same line as described above;* **iii)** $\sigma(T_{s,\mathrm{core}})/\sigma(T_s)$: *the ratio between the standard deviations of annual mean* $T_{s,\mathrm{core}}$ *and annual mean* T_s; **iv)** $r_{temporal}(T_s, T_{s,\mathrm{core}})$: *temporal correlation between annual mean values of modelled* T_s *and* $T_{s,\mathrm{core}}$; **v)** ΔT_s: *difference in* T_s *between the warmest seven years and coldest seven years;* **vi)** $\Delta T_{s,\mathrm{core}}$: *difference in* $T_{s,\mathrm{core}}$, *derived using Equation (3), between the warmest seven years and coldest seven years;* **vii)** ΔT_s: *difference in* T_s *between the* SENS *and* CTL*-integration;* **viii)** $\Delta T_{s,\mathrm{core}}$: *difference in* $T_{s,\mathrm{core}}$, *derived using Equation (4), between the* SENS *and* CTL*-integration.*