The Relationship between the Southern Hemisphere Annular Mode and Antarctic Peninsula Summer Temperatures: Analysis of a High-Resolution Model Climatology

NICOLE P. M. VAN LIPZIG
Physical and Regional Geography Research Group, Katholieke Universiteit Leuven, Heverlee, Belgium

GARETH J. MARSHALL
British Antarctic Survey, Cambridge, United Kingdom

ANDREW ORR
European Centre for Medium-Range Weather Forecasts, Reading, United Kingdom

JOHN C. KING
British Antarctic Survey, Cambridge, United Kingdom

(Manuscript received 25 September 2006, in final form 25 May 2007)

ABSTRACT
The large regional summer warming on the east coast of the northern Antarctic Peninsula (AP), which has taken place since the mid-1960s, has previously been proposed to be caused by a trend in the Southern Hemisphere Annular Mode (SAM). The authors utilize a high-resolution regional atmospheric model climatology (14-km grid spacing) to study the mechanisms that determine the response of the near-surface temperature to an increase in the SAM ($\Delta T/\Delta$SAM). Month-to-month variations in near-surface temperature and surface pressure are well represented by the model. It is found that north of $68^\circ$S, $\Delta T/\Delta$SAM is much larger on the eastern (lee) side than on the western (windward) side of the barrier. This is because of the enhanced westerly flow of relatively warm air over the barrier, which warms (and dries) further as it descends down the lee slope. The downward motion on the eastern side of the barrier causes a decrease in surface-mass balance and cloud cover. South of $68^\circ$S, vertical deflection across the barrier is greatly reduced and the contrast in $\Delta T/\Delta$SAM between the east and west sides of the barrier vanishes. In the northeastern part of the AP, the modeled $\Delta T/\Delta$SAM distribution is similar to the distribution derived from satellite infrared radiometer data. The region of strongest modeled temperature sensitivity to the SAM is where ice shelf collapse has recently taken place and does not extend farther south over the Larsen-C Ice Shelf.

1. Introduction
One of the regions on earth with the largest warming over the last five decades is the Antarctic Peninsula (AP). At Faraday station (note that this station became Vernadsky in 1996; see Fig. 1 for station locations), the mean annual near-surface temperature increased by 2.9°C during the period 1951–2004 (significant at a 1% level; Marshall et al. 2006), compared to the global average of 0.5°C (Hansen et al. 2001). The Faraday warming is largest in the winter season. In contrast, the warming on the eastern side of the AP is largest in summer and autumn. Between 1965 and 2004, the summer–autumn temperature increase at Esperanza station was almost 3 times as large as at Faraday (2.7°C and 1.0°C, respectively; Marshall et al. 2006). It is the summer warming that is most relevant in terms of the melting of snow and ice in the AP region and, as a consequence, in driving regional ecological change.

Together with the observed increase in summer temperature, a number of floating ice shelves have re-
treated and disintegrated (Vaughan and Doake 1996; Skvarca et al. 1999; Doake et al. 1998; Rott et al. 1996). The collapse of the Larsen-B Ice Shelf at the end of the 2001/02 summer season is unprecedented since the Holocene began (Domack et al. 2005). Large summer meltwater fluxes on the northeastern (Weddell Sea) side of the orographic barrier of the AP are likely to have contributed to the collapse (Van den Broeke 2005). It is probable that percolating meltwater has propagated fracturing of the ice shelf, which eventually resulted in its disintegration (Scambos et al. 2004).

Contemporaneous with the observed summer warming, the principal mode of variability in the Southern Hemisphere (SH) circulation—the SH Annular mode (SAM; also known as the Antarctic Oscillation)—began a positive phase shift in the mid-1960s (Marshall 2003). The trends in the SAM are statistically significant annually, in autumn and especially in summer (Marshall et al. 2006). Model studies have identified ozone depletion (Thompson and Solomon 2002; Gillett and Thompson 2003), an increase in greenhouse gases (Kushner et al. 2001; Cai et al. 2003), and a consequent increase in the tropical sea surface temperatures (SSTs; Grassi et al. 2005) as partial underlying causes of the positive phase shift of the SAM. However, natural forcing factors or internal feedback mechanisms in the climate system have also influenced the state of the SAM, as large trends occurred around 1960, a period before ozone-depleting chemicals were released into the atmosphere and before marked anthropogenic warming occurred (Jones and Widmann 2004). Note that the highest value for the SAM index since 1957 was in the summer of 1999/2000. Also during the 2000/01 summer season (when the Larsen-B Ice Shelf collapsed) the SAM index was above the 90% quantile of the SAM index distribution since 1957.

The positive phase shift of the SAM is strongly reflected in SH tropospheric circulation (strength of the circumpolar westerly flow) and surface temperatures over the SH high latitudes during summer and autumn:
over the last 30 yr, the trend in December–May temperature has been negative over the east Antarctic stations and positive over the AP region. Trends in the SAM are likely to have contributed significantly to the Antarctic near-surface temperature changes (Thompson and Solomon 2002; Marshall 2007). Also sea-ice extent and ocean currents vary consistently with the atmospheric flow anomalies related to the SAM (Liu et al. 2004; Lefebvre et al. 2004).

Reanalyses have been used to study the effect of SAM variability on the AP region, either directly (e.g., Lefebvre et al. 2004) or by driving a regional model with reanalysis data in hindcast mode (e.g., Van den Broeke and Van Lipzig 2004). This is appropriate when the mean sea level pressure gradient (PG) between 40° and 65°S is well represented by the reanalyses. Before 1979, deficiencies in PG were found in the reanalyses, but from 1979 onward the 15- and 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15 and ERA-40) correctly represented the PG (Marshall 2003). From an integration for the period 1980–93 with a regional model, driven by ERA-15, it was shown that northwesterly flow anomalies cause warming over the Antarctic Peninsula, the Weddell Sea, and the adjacent regions during years with high SAM (strong westerly circumpolar winds; Van den Broeke and Van Lipzig 2004). The warming was found to be larger in autumn–winter than in spring–summer. Ocean–atmosphere global models, with transient forcing from preindustrial conditions and resolutions mostly in the range of 200 to 300 km, hint at similar flow anomalies and a similar weak sensitivity of temperatures in the AP region to the SAM during late spring for the present (time slice 1970–99) and future (time slice 2070–99) climate (Carril et al. 2005).

Climatological model or (re)analyses studies for the AP region generally use a horizontal grid spacing of 50 km or more (e.g., Van Lipzig et al. 2002 and the abovementioned studies). In such models, the AP, which is around 250 km wide, is represented by—at best—five grid points. The height of the orographic barrier (H) is underestimated and the effect of the complex orography on the local meteorological conditions is not fully taken into account.

An accurate representation of H in atmospheric models is needed because H influences whether the flow passes over or around the barrier. For the high latitude and elongated shape of the AP, the flow is typically blocked when the Froude number (Fr) < 0.5 ("blocked" conditions) and it passes over the barrier when Fr > 0.5 ("flow-over" conditions) (Olaffson 2000; Orr et al. 2008, hereafter OR08). The Froude number is defined as Fr = u/(NH), where u is the layer-averaged cross-barrier wind speed and N is the Brunt–Väisälä frequency. Orr et al. (2004) stressed the importance of flow blocking in deflecting the westerly winds incident to the AP to the south, advecting warm air along its western coast. From laboratory experiments and high-resolution case studies with an atmospheric model, OR08 showed that an increase in westerlies because of the positive phase shift of the SAM is likely to result in a change in the dominant flow regime from blocked to flow-over conditions characterizing the northern AP.

To accurately model the blocking of the flow by the barrier, the transition from blocked to flow-over conditions, and the effect of local orography on (katabatic) winds, integrations with a grid spacing on the order of 10 to 20 km are needed (Smith et al. 2006), which resolve the orographic barrier by about 20 grid points. Currently, these high-resolution integrations are available only for a few cases (Pascoe 2002; Orr et al. 2004). The analysis presented in this paper is based on the model integration described by Van Lipzig et al. (2004), which is currently the only model integration from which information for the AP region can be obtained for several (seven) years at a horizontal resolution as high as 14 km × 14 km. The goal of this work is to improve our understanding of the physical mechanisms through which the positive phase shift of the SAM and associated strengthening of the westerly circumpolar winds have affected the local summer climatology of the mountain range on the AP using high-resolution model output.

In section 2 of this paper the regional atmospheric model is described and a model evaluation is presented. In section 3a, the sensitivity of the model output to the SAM is discussed and composites for months with high (SAM+) and low (SAM−) values for the SAM index are defined. The spatial distribution, vertical extent of the anomalies during SAM+ and SAM−, and the mechanisms for the warming during SAM+ are presented in the remaining part of section 3. In section 4, the modeled warming during SAM+ is compared to satellite infrared radiometer data. In addition, the sensitivity of the model results to the grid spacing is examined. In section 5, we present our major conclusions and the implications of our results.

2. Model description, methodology, and model evaluation

For our analysis, we have used the regional atmospheric climate model (RACMO14) covering the AP, the Filchner–Ronne Ice Shelf, parts of the Weddell Sea, Bellingshausen Sea, and the West Antarctic Ice Sheet (Fig. 1) at a 14-km grid spacing (Van Lipzig et al. 2004).
RACMO14 is driven at its lateral boundaries by fields from RACMO55 (at a 55-km grid spacing), which is in turn driven by ECMWF ERA-15 (see Fig. 1a for the size of the model domains). The lowest layers are centered at 7, 38, 139, 367, and 752 m. Sea surface temperature and sea-ice extent in both RACMO14 and RACMO55 are prescribed from observations using Special Sensor Microwave Imager (SSM/I) data. Adjustments have been made to the original model (described by Christensen et al. 1996) to improve it for Antarctic conditions (Van Lipzig et al. 1999, 2002), and making it suitable for integrations at the high resolution of a 14-km grid spacing (Van Lipzig et al. 2004).

Atmospheric model data from RACMO14 are available for the 7-yr period of 1987–93 (Van Lipzig et al. 2004). The monthly mean model output is, in principle, a useful dataset for studying the effect of SAM variability on the local meteorological conditions. However, to do this, the ability of the model to represent the month-to-month variations in meteorological conditions needs to be evaluated. Therefore, we compare observed and modeled time series for 21 (7 yr × 3 months) summer months (December–February) at the Antarctica stations of Bellingshausen, O’Higgins, Esperanza, Faraday, and Rothera. Because these stations are all located on land, the land-ice grid point corresponding most closely to the measuring site was selected from the nine grid points surrounding the site. At Bellingshausen, the nine grid points surrounding the site were all sea grid points; thus the closest grid point was selected.

To evaluate whether the model is able to reproduce the synoptic situation during summer, the model output is compared with the monthly mean sea level pressure at the five stations considered (Fig. 2). During some months (e.g., January 1990) modeled pressure deviates from the observed value. However, for most of the time series the correspondence is very good. The correlation between modeled and observed sea level pressures ranges from 0.87 at Esperanza to 0.96 at Bellingshausen, which are each significant at a <1% level. Also, the amplitude of the pressure variations is well represented and the bias is smaller than 0.9 hPa. It would be interesting to investigate the spectral nudging technique (Von Storch et al. 2000; Miguez-Macho et al. 2004) to improve the correspondence further.

The modeled monthly mean near-surface temperature also correlates significantly (at a <1% level) with the measured temperature at the five stations (Figs. 2f–j). The value for the correlation coefficient ranges from 0.63 at O’Higgins to 0.85 at Esperanza (Table 1). The modeled temperatures are generally lower than the measured temperatures, with a maximum cold bias of 2.4 K at O’Higgins. This bias occurs partly because the orography rises steeply going inland at the coastline where the stations are located, which is at an elevation of 10 to 16 m above sea level. The model elevation of the grid box is an average for a larger area and is generally higher than the station elevation. For example, at the O’Higgins station, located at 10 m above sea level, the elevation of the closest land-ice grid box is 305 m above sea level. Since temperature decreases with height, this causes a cold bias.

A method is developed to estimate the bias that is caused by differences in elevation between the model grid box and the station (Δz). First, the regression coefficient for a linear relation between the grid box height and the modeled temperature is calculated using 25 grid boxes surrounding the station. Second, this regression coefficient (which ranges between 4.2 and 5.8 K km⁻¹) is multiplied by the Δz to estimate the model bias that is due to Δz. At Faraday, the bias can almost entirely be explained by Δz. However, negative biases between 0.8 and 1.2 K, not related to Δz, remain for the other stations. Possibly, the effect of the albedo plays an important role. In summer, the snow close to the station melts and the exposed area has the albedo of bare soil. These local effects are not taken into account in the model, where albedo is a function of temperature: when the temperature is lower than −3°C, the albedo is 0.8. The value decreases linearly to 0.7 at 0°C.

The standard deviation (σ) of modeled monthly mean temperatures is well represented for Bellingshausen, O’Higgins, Esperanza, and Faraday (differences in σ are smaller than 0.16 K; see Table 1). At Rothera, the model overestimates the variability (difference in σ is 0.32 K).

Apart from a direct comparison between modeled and measured time series, we examine the modeled and observed sensitivities of the near-surface temperature to the SAM at the five stations mentioned above. For the SAM index, we use the definition given by the University of Washington (http://www.jisao.washington.edu/aaio/). They based their index on the first principal component of the 850-hPa extratropical height field from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Thompson et al. 2000). There are also definitions of the SAM index, which are based on station data (Marshall 2003). We have chosen the index from the University of Washington since it was found to be a good indicator for the circumpolar westerly flow in our regional model. This index explains 80% (correlation coefficient = 0.89) of the RACMO55 modeled zonal wind velocity at a distance of 600–800 km from the Antarctic coastline (Van den Broeke and
Van Lipzig 2003). The correlation between the RACMO14 zonal wind at 570 hPa at the Larsen-B cross section (see Fig. 1c) and the SAM is 0.52 (significant at a <5% level). Note that throughout this paper, we have taken autocorrelation effects into account when deriving significance levels, based on the method described by Santer et al. (2000).

Marshall et al. (2006) found a significant correlation (at a <10% level) in summer between the SAM and near-surface temperature for the stations Esperanza and Marambio. The correlation was found to be insignificant (at a <10% level) at Bellingshausen, Faraday, and O’Higgins. Since they used a longer time period (mostly 1958–2000), we repeated the analysis for the 21 summer months covered by the 7-yr integration period (Table 2). Also for the shorter time period, the only station with a significant correlation was Esperanza (note that we did not consider Marambio since many data gaps were present in the period of 1987–93). At Esperanza, the linear regression coefficient between the SAM and the near-surface temperature (dSAM/dT) was well reproduced by the model.

Correlations between the temperature and the SAM were found to be insignificant at a <10% level for
Table 1. Correlation between model-derived ($R_{\text{mod}}$) or satellite-derived ($R_{\text{sat}}$) monthly mean near-surface temperatures and in situ measured monthly mean values. Std dev of the time series of monthly mean near-surface temperatures derived from in situ measurements ($\sigma_{\text{obs}}$), model ($\sigma_{\text{mod}}$), and satellite ($\sigma_{\text{sat}}$). Significant correlations are shown by the asterisks: *** $<$1%, ** $<$5%, and * $<$10%.

<table>
<thead>
<tr>
<th></th>
<th>$R_{\text{mod}}$</th>
<th>$R_{\text{sat}}$</th>
<th>$\sigma_{\text{obs}}$</th>
<th>$\sigma_{\text{mod}}$</th>
<th>$\sigma_{\text{sat}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bellingshausen</td>
<td>0.83***</td>
<td>0.66***</td>
<td>0.89</td>
<td>0.96</td>
<td>2.75</td>
</tr>
<tr>
<td>O'Higgins</td>
<td>0.63***</td>
<td>0.66***</td>
<td>0.86</td>
<td>0.94</td>
<td>2.36</td>
</tr>
<tr>
<td>Esperanza</td>
<td>0.85***</td>
<td>0.42*</td>
<td>1.28</td>
<td>1.12</td>
<td>2.71</td>
</tr>
<tr>
<td>Faraday</td>
<td>0.64***</td>
<td>-0.16</td>
<td>0.96</td>
<td>1.01</td>
<td>1.97</td>
</tr>
<tr>
<td>Rothera</td>
<td>0.64***</td>
<td>-0.17</td>
<td>0.94</td>
<td>1.26</td>
<td>1.91</td>
</tr>
</tbody>
</table>

O'Higgins, Faraday, and Rothera, both in model and in measurements. On the other hand, the model shows a significant correlation for Bellingshausen, which was not observed. The model grid point corresponding to Bellingshausen is a sea grid point. Since no sea ice is present in this region, the SST anomalies prescribed to the model determine the modeled sensitivity. Also at O'Higgins, the correlation between the temperature and the SAM was found to be larger in the model (0.37) than in the measurements (0.29), but both were insignificant at a <10% level. Local effects, shielding the observation sites, are not represented in detail by the model. Therefore the SST anomalies probably have a stronger influence in the model than in reality.

A further evaluation of wind and surface-mass balance (SMB) in RACMO14 has been performed by Van Lipzig et al. (2004). They used in situ data of the wind vector at three coastal sites located on the northern, eastern, and western sides of the AP. For two of the three sites considered, the prevailing wind direction and bimodal wind distribution are correctly represented by the model. The comparison with measurements improved when the grid spacing was decreased from 55 to 14 km. A resolution of 14 km is sufficient to capture effects such as the barrier wind at coastal stations, like Butler Island situated a few kilometers east of the steeply sloping AP east coast (see Fig. 1). The correspondence between model and measurements from this station is very good, with a difference between measured and modeled wind speed of only 0.04 m s$^{-1}$.

Van Lipzig et al. (2004) also compared modeled surface-mass balance (here defined as the sum of precipitation minus sublimation/evaporation) with all available (about 200) in situ measurements. They found that the mean bias (where positive and negative biases partly cancel) in SMB is a $-9$-mm water equivalent yr$^{-1}$, which is only 2% of the mean area-averaged SMB. This value is smaller than the uncertainty in the measurements. The average of the absolute biases in the SMB is a 220-mm water equivalent yr$^{-1}$. Part of this bias is related to summer melt (which is not taken into account in the model) and snow drift on a scale smaller than the size of a grid box (Turner et al. 2002). To the south of Alexander Island, the SMB is underestimated because of an overestimation of the leeward precipitation shadow of the mountains of Alexander Island. At 101 of the 200 sites considered, SMB is simulated within the range of uncertainty of the measurements.

In general, the evaluation presented above shows that the synoptic situation is well represented by the model, the resolution is sufficient to capture some effects of the local orography on the wind, and the month-to-month variations of the surface temperatures in summer are well represented. In addition, the model is able to represent the regional pattern of temperature variations associated with SAM variability. The effect of SST anomalies might be overestimated at some sites, indicating that even a grid spacing of 14 km is insufficient to resolve all local effects. The model has a cold bias of about 1°C at four out of the five stations, which is likely because of an overestimation of the local albedo. However, since this bias is not significantly correlated with temperature itself, it does not affect the temperature sensitivity to the SAM in the model. Based on this evaluation, we conclude that RACMO14 is an appropriate tool for our purposes.

3. Results

a. Sensitivity of the near-surface temperature to the SAM

Following Marshall et al. (2006) and many others, we define the sensitivity of the near-surface temperature to the SAM ($dT/d\text{SAM}$) as the linear regression coefficient between the SAM and the near-surface temperature. Since no significant long-term change in the relationship between the surface temperatures and the

Table 2. The sensitivity of the temperature to the SAM, as derived from linear regression between the SAM index and observed, modeled, and satellite-derived monthly mean near-surface temperatures. Only values where the correlation is significant at a <10% level are indicated. Significant regression coefficients are shown by the asterisks: *** $<$1%, ** $<$5%, and * $<$10%.

<table>
<thead>
<tr>
<th></th>
<th>Observed</th>
<th>Modeled</th>
<th>Satellite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bellingshausen</td>
<td>—</td>
<td>0.60**</td>
<td>1.4*</td>
</tr>
<tr>
<td>O'Higgins</td>
<td>—</td>
<td>—</td>
<td>1.4**</td>
</tr>
<tr>
<td>Esperanza</td>
<td>0.75**</td>
<td>0.75**</td>
<td>1.9**</td>
</tr>
<tr>
<td>Faraday</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Rothera</td>
<td>—</td>
<td>—</td>
<td>-1.8***</td>
</tr>
</tbody>
</table>
SAM has been found over the recent half-century (Marshall 2007), even a relatively short time series (such as the 7 yr considered here) is suitable to study the processes that are responsible for the regional patterns of the temperature sensitivity to the SAM. The modeled regression coefficients for the summer months in the period of 1987–93 and the corresponding significance levels for the correlation are shown in Figs. 3a,b.

Over the northern AP, the modeled response of temperature to an increase in westerlies is clearly affected by the high mountain range: the temperature sensitivity to the SAM exhibits a clear contrast between the northeastern side and the northwestern side of the barrier. A significant positive correlation is found over the northeastern side of the AP including the Larsen-B Ice Shelf. It does not, however, extend farther south over the Larsen-C Ice Shelf. For the western side of the AP as well as for the land-ice region south of 68°S, no significant correlations are found.

The significant difference in surface temperature response to the SAM between the eastern and western sides of the AP (Fig. 3a) is not found in ERA-40 (cf. Fig. 7b; Marshall et al. 2006), which operates at a grid spacing of about 125 km (T159). These authors found the largest sensitivity across both sides of the northern tip of the AP. The detailed spatial distribution of the sensitivity of the near-surface temperatures to the SAM as modeled by RACMO14 is not represented by ERA-40. Although the resolution of ERA-40 is impressive on a global scale, it is not high enough to represent the AP accurately: the orographic barrier is too wide and too low. Over the northern section of the AP, the model height is only 50% of the actual height. This is accounted for by employing a parameterization scheme (Lott and Miller 1997) in the model that was used to produce ERA-40. This parameterization scheme determines the gravity wave drag and low-level blocking drag contributions due to subgrid-scale orography. The actual drag is estimated as the combination of parameterized and resolved drag. Recent results of OR08 suggest that the ERA-40 parameterized subgrid-scale drag is too strong and that weather systems passing over the peninsula are in actuality being prevented from doing this.

Over the ocean, $dT/dSAM$ is mostly positive (Fig. 3a), which is consistent with the results found by Kwok and Comiso (2002), who correlated monthly mean surface temperatures derived from infrared satellite data with the SAM for all months in the period of 1982–98. Figure 3a shows a larger value for $dT/dSAM$ over the ocean on the western side than on the eastern side of the AP.

As shown by Marshall et al. (2006) and confirmed by the RACMO14 output (Fig. 3c), the regional pattern of significant correlation between the temperature and SAM is different during autumn compared to summer. During autumn there is no clear contrast between the eastern and western sides to the north of the AP. Therefore it is likely that the mechanisms for the warming are different. In this paper, we focus on the summer season because it is most relevant in terms of the melting of snow and ice. It is beyond the scope of this paper to investigate the mechanisms responsible for the autumn warming.

For further analysis, we divided the 7-yr integration into two composites: the nine summer months with the highest values for the SAM index (SAM+), defined by the University of Washington, and the nine summer months with the lowest values for the SAM index (SAM−; Table 3). The motivation for using composites is that it facilitates the interpretation of the model output, especially when analyzing cross sections of the modeled temperature and flow fields. When using composites, the same months determine the flow field, temperature, surface-mass balance, and cloud-cover difference between SAM+ and SAM− at all locations and at all levels. When using linear regression, this is not the case, and different outliers might affect the regression at different levels or locations. The regions for which the average near-surface temperature of the SAM+ and SAM− composite differ significantly are similar to the regions where the near-surface temperature correlates significantly with the SAM; compare Fig. 3b with 3d. Thus we conclude that composites are suitable to study the mechanisms responsible for the relation between the temperature and the SAM; we will therefore use this approach in the rest of this paper.

b. The effect of SAM variability on the spatial distribution of meteorological variables

A circumpolar trough of low surface pressure ($p_s$) is present around the Antarctic continent. In the northern region of the AP, a poleward gradient is present in both $p_s$ and in the geopotential at 500 hPa ($\Phi_{500}$), generating the mean westerly flow. The poleward gradient in $p_s$ and $\Phi_{500}$ is larger and the westerlies are stronger during SAM+ than during SAM− (Fig. 4). North of ~70°S, the anomaly in $p_s$ and $\Phi_{500}$ is to a large extent annular; however, there is also an increase in the northerly component of the wind on the western AP during SAM+.

South of ~70°S, the anomaly is largely nonanular. During SAM+, the climatological low located west of the AP (Amundsen–Bellingshausen Sea) deepens. This favors warm air advection on the southwestern side of
FIG. 3. (a) The sensitivity of the near-surface temperature to the SAM, as defined by the regression coefficients between the SAM index and near-surface temperature for the summer months in the period 1987–93. Regions for which the correlation between the SAM and near-surface temperature (b) during summer and (c) during autumn are significant at <10%, <5%, and <1% levels. (d) Regions for which the average summer near-surface temperature of the SAM+ and SAM− composite differ significantly at <10%, <5%, and <1% levels. The Student’s t test is used for deriving the significance levels.
the AP during SAM+. At 500 hPa, an increase in continental flow from the Filchner–Ronne Ice Shelf is modeled during SAM+, which advects cold air toward the southeastern side of the AP. The large-scale response of $p_s$ to an increase in the SAM is similar in RACMO14 compared to the sensitivity in ECMWF ERA-40 as presented by Marshall et al. (2006).

Different temperature regimes are modeled on either side of the AP (Fig. 5), as also identified from observations by Martin and Peel (1978) and King and Comiso (2003). The west coast is affected by warm air masses from the Bellingshausen Sea. The eastern side is influenced by cold air masses of continental origin (King et al. 2003).

The modeled response of temperature to an increase in the SAM normalized by the difference in the SAM index between SAM+ and SAM− ($\Delta$SAM = 1.29; Figs. 5, 6a) corresponds closely to the temperature sensitivity to the SAM (Fig. 3a), with the largest values over land ice on the northeastern side of the AP. For temperature, surface-mass balance, cloud cover, and vertically integrated cloud ice–water (IWP), a contrasting response to changes in the SAM is found between the eastern and western sides of the AP (Figs. 6a–d, respectively).

c. Vertical extent of the modeled warming

To study the vertical extent of the warming, we have analyzed model output for three cross sections, as indicated in Fig. 1c. The first cross section was chosen through the Larsen-B Ice Shelf. The second cross section is through Esperanza, which has a long record of meteorological measurements and is one of the stations with the strongest summer warming in the AP. This cross section is therefore relevant to identify correspondences and contrasts between the Esperanza region and the Larsen-B Ice Shelf. The third cross section is representative of the southerly part of the AP, where the response of temperature at the eastern side of the AP is moderate compared to the northern side (Fig. 6a). The cross sections are parallel to the 500-hPa isobars and therefore parallel to the mean westerly flow incident to the AP.

The mean modeled summer season atmosphere in the northern AP is stably stratified, with colder (and more stable) conditions on the eastern side than on the western side (Fig. 7a). A contrast in the response to changes in the SAM between the eastern and western sides of the AP is found at the northern AP cross sections at Esperanza and Larsen-B (Figs. 7b,c). Here, the difference in potential temperature normalized by the difference in the SAM index between SAM+ and SAM− ($\Delta\theta$/\Delta$\text{SAM}$) is much larger on the eastern side ($\Delta\theta$/\Delta$\text{SAM}$ > 0.8°C) than on the western side of the AP, where $\Delta\theta$/\Delta$\text{SAM}$ is generally smaller than 0.4°C, down to about 0°C at the foot of the barrier in the Larsen-B cross section.

Long-term measurements are available at the Esperanza station. In the model, the warming at this station is moderate compared to the Larsen-B region, where no long-term observations of temperature exist. At Esperanza, a value for $\Delta\theta$/\Delta$\text{SAM}$ > 0.8°C is found from the surface up to 800 hPa. In the Larsen-B cross section, the modeled $\Delta\theta$/\Delta$\text{SAM}$ is larger than 0.8°C along the entire lee side from the top to the base of the mountain barrier. A region of strong warming ($\Delta\theta$/\Delta$\text{SAM}$ > 1.2°C) extends for about 250 km eastward. The contrast between the windward side and the lee side of the barrier is most pronounced for the Larsen-B cross section. At this cross section, the sensitivity of $\theta$ to the SAM ($d\theta$/d$\text{SAM}$) at the base of the mountain over the layer extending from the surface to 900 hPa was found to be 0.2°C per unit change in the SAM index at the windward side and 1.5°C per unit change in the SAM index at the lee side of the barrier.

South of ~68°S, the difference vanishes between the eastern and western sides in terms of temperature sensitivity to the SAM. At the Alexander Island cross section, $\Delta\theta$/\Delta$\text{SAM}$ is around 0.4°C for most of the atmospheric layer between the surface and 600 hPa. Since the difference between SAM+ and SAM− in this region is mostly nonannular and an increase in westerlies is not present, the behavior is very different from the northern part of the AP.

At about 200 hPa, a maximum in $\Delta\theta$/\Delta$\text{SAM}$ is modeled at all three cross sections. This is caused by the increase in strength of the polar front jet stream and its southward movement during a positive phase shift of the SAM (Gallego et al. 2005). Analysis of Bellingshausen upper-air data (Marshall et al. 2006) indeed showed a maximum correlation between northern AP free-air temperatures and the SAM at the height of the

| Table 3. The nine DJF months with the highest value for the SAM index (SAM+) and the nine DJF months with the lowest (most negative) value for the SAM index (SAM−) into which the model integration, for the period 1987–93, has been subdivided. The definition for the SAM is taken from the University of Washington. |
|---------------------------------|-----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|

Unauthenticated | Downloaded 02/20/24 08:35 AM UTC
Fig. 4. Difference in DJF (a) surface pressure and (b) 500-hPa geopotential between the composite of months with high values for the SAM (SAM+) and the composite of months with low values for the SAM (SAM−). All figures are normalized by the mean difference in the SAM index between SAM+ and SAM−.

Fig. 5. DJF near-surface (7 m) temperature during (a) SAM− and (b) SAM+.
FIG. 6. Difference in DJF (a) near-surface (7 m) temperature, (b) surface-mass balance, (c) cloud cover, and (d) vertically integrated cloud ice/water between SAM+ and SAM−. The thick white (black) line shows where the positive (negative) correlation is significant at <10% level. All figures are normalized by the mean difference in the SAM index between SAM+ and SAM−.
jet streams. Note that in the measured Bellingshausen record, a cooling trend is observed above 200 hPa because of ozone loss in the stratosphere (Marshall et al. 2006).

d. Mechanisms for the warming

A markedly stronger zonal flow during SAM+ is apparent at the Larsen-B cross section (Fig. 8). This gives rise to a large increase in mean cross-barrier flow, that is, a greater frequency of occurrence of flow-over conditions, which is consistent with the results of OR08. Also noticeable are a decrease in upstream blocking and an increase in zonal velocity above the leeward slope. The difference between SAM+ and SAM− is largest above the barrier to the 600 hPa level, probably because of the transition to the flow-over regime and the associated confluence of the flow when it passes over the barrier. Mean Froude numbers are 0.10 for SAM− and 0.24 for SAM+ (based on upstream values of $u$ and $N$ at a height of around 2000 m). This increase in Fr is mainly due to a doubling of $u$ ($N$ decreases by only 3%). At about 200 hPa, the difference in zonal wind speed is largest because of an increase in strength of the polar front jet stream and its southward movement (Gallego et al. 2005).

The large increase in cross-barrier flow near the surface, as a response to an increase in SAM, is found not only at the Larsen-B cross section but over the entire grounded land ice north of ~68°S (Fig. 9). At the Esperanza cross section, Fr increases from 0.42 during SAM− to 1.04 during SAM+, mainly from a doubling of the wind speed. South of ~68°S, the higher AP (2000 m; see Fig. 1b) and generally smaller zonal velocities mean that the flow regime is dominated by flow blocking for both SAM+ and SAM−. Representative Froude numbers for the Alexander Island cross section are about zero, during both SAM+ and SAM−. The modeled warming at the lee side is small over this re-
The most remarkable feature south of ~68°S is an increase in the wind component along the western coast, likely caused by the southwestward movement and strengthening of the climatological low in the Amundsen–Bellingshausen Sea during SAM+.

The increase in flow across the AP for SAM+ was also found by Marshall et al. (2006), who based their studies on a composite of ERA-40 summers with high and low SAM values. However, because of the relatively coarse ERA-40 horizontal resolution, many local features, such as the larger sensitivity of the wind speed over land ice north of ~68°S compared to the surrounding oceans or the channeling of the flow in George VI Sound, are not captured by ERA-40.

The increase in (north) westerlies enhances both the advection of moisture and the orographically induced ascent on the western side of the AP (Fig. 10). This forced ascent results in adiabatic expansion as the pressure drops and hence the air cools, and eventually the vapor it contains reaches saturation (Roe 2005). For this reason, an increase in modeled SMB (from an increase in precipitation), cloud cover, and vertically integrated cloud ice–water (Figs. 6b–d, respectively) is found on the western side of the barrier during SAM+ years.

When air is lifted over a barrier and it is denser than its surroundings, the air will sink down the leeward side under the influence of gravity. Both the latent heat release, which has taken place on the windward side, and the transport of relatively dry air with high potential temperature from upper-atmospheric levels result in a warming and drying of air on the lee side (Beran 1967). Consequently, a decrease in the SMB, cloud cover, and IWP and an increase in temperature are found on the lee side of the barrier during SAM+ years (Figs. 6a–d). Because of the high wind speed, the turbulent exchange might also prevent the development of an inversion layer (Beran 1967). Moreover, the warm air advection from the western side of the barrier increases when more air is able to flow over the barrier.

These mechanisms described above are responsible for a clear relation between the warming on the lee side of the barrier, as described in section 3a, and an anomalous vertical motion in the atmosphere during SAM+, with upward motion at the windward side and downward motion at the lee side (Fig. 10). The anomaly in...
the vertical velocity is much larger in the Larsen-B cross section than in that through Esperanza. Changes in the vertical velocity over the Alexander Island cross section are negligible because of the absence of flowover conditions.

The presence of clouds affects the radiative balance at the surface. Therefore, it is possible that the decrease in cloud amount on the lee side of the northern AP during SAM+ has contributed to the observed warming. We investigate this by comparing the energy balance at the surface. The downward shortwave flux at the surface (SW↓) clearly increases on the lee side during SAM+ (Fig. 11). However, the effect of clouds on the net radiative balance at the ice shelf surface is small because of the high albedo and the compensating effect of clouds on the longwave downward radiation at the surface. The ice shelf surface warming is mostly driven by an increase in the sensible heat flux from the increased föhn effect during SAM+. As stronger winds during SAM+ (Fig. 9) blow relatively warm air (Fig. 6a) over the ice shelves, the sensible heat exchange between the surface and the atmosphere is enhanced. At the foot of the mountains, the increases in downward sensible heat flux are partly compensated for by an increase in upward latent heat flux (the föhn winds are generally also very dry; therefore sublimation is enhanced).

Another factor that might have contributed to the warming is an anomaly in the sea-ice concentration. The sea-ice concentration east of the AP is higher during SAM+ than during SAM− (Fig. 11f) because of the enhanced continental flow from the Filchner–Ronne Ice Shelf (Fig. 4). This leads to a negative anomaly in the net radiation at the surface (Fig. 11c) and a positive

**Fig. 9.** DJF near-surface (7 m) wind vector during (a) SAM− and (b) SAM+. (c) Normalized difference in summer near-surface wind vector between SAM+ and SAM−.
anomaly in the sensible and latent heat fluxes [an increase in sea-ice concentration during SAM+] results in a decrease in the upward (i.e., negative) sensible and latent heat fluxes over the sea-ice region]. In the absence of other forcing, an increase in sea-ice concentrations results in a cooling, since warm open water is replaced by cold sea ice. Such an effect would oppose the observed temperature sensitivity to the SAM. This makes it clear that the temperature change is not being driven by sea-ice change in this region of the AP.

4. Discussion

a. Comparison with near-surface temperature inferred from infrared satellite measurements

In section 3a, we have compared model output to in situ measurements. Apart from model data and in situ measurements, there is another source of information for the spatial variation of Antarctic surface temperature: monthly mean satellite infrared radiometer data ($T_{IR}$) from the Advanced Very High Resolution Radiometer (AVHRR) carried on National Oceanographic and Atmospheric Administration (NOAA) polar-orbiting meteorological satellites (Comiso 2000). These data are available on a 6.25 km × 6.25 km polar stereographic grid. Since the $T_{IR}$ dataset is biased by the absence of data for days with cloud cover (Comiso 2000; Shuman and Comiso 2002), we do not take this dataset as “truth,” but we want to identify the correspondence and differences between the model dataset and the $T_{IR}$ dataset.

To do this, a composite is made of the $T_{IR}$ that is analogous to Fig. 6a (see Fig. 12a). Good correspondence is found between Fig. 12a and the work of Schneider et al. (2004), who made a regression of the $T_{IR}$ grid point data over land on the first principal component of the geopotential height at 500 hPa (a representative SAM index; Thompson and Solomon 2002). They also found the highest values for the regression coefficient over the northeastern peninsula region. (Note that they did not include the sea and sea ice in their analysis.)
The satellite-derived temperature difference between SAM+ and SAM− ($\Delta T_{IR}/\Delta SAM$) is large along the northern part of the east coast of the peninsula, extending from the tip of the AP to the Larsen-B Ice Shelf (Fig. 12a). This agrees with the model results (Fig. 6a), which support that both the behavior of the model and the analysis based on the satellite infrared radiometer data in this region are realistic. But in other regions there are large differences between the $T_{IR}$ and the model data: (i) $T_{IR}$ data show a markedly larger area of warming over the ocean along the east coast than in the model and (ii) $T_{IR}$ data show a cooling over almost the entire grounded land-ice region south of 67°S, where the model shows a warming. The land–sea contrast in the $T_{IR}$ dataset is not found in our model integrations or in the results presented by Van den Broeke and Van Lipzig (2003), who use the RACMO55 model.

At the stations presented in Table 2, the value for the satellite-derived temperature sensitivity to SAM ($dT_{IR}/dSAM$) ranges from −1.8°C at Rothera to 1.9°C at Esperanza, which is an overestimation of both the positive and the negative values compared to the in situ measurements (Table 1). This overestimation of $dT_{IR}/dSAM$ is likely to be related to a larger month-to-month variability in $T_{IR}$ compared to the measured temperatures (Table 1). This can partly be explained by the fact that the satellite sees the surface temperature, whereas the measurements are made close to the surface. However, it is also possible that conditional sampling for cloud-free conditions amplifies the warming on the lee side and cooling on the windward side, which we address below.

The $T_{IR}$ dataset is biased by the absence of data for overcast days (Comiso 2000; Shuman and Comiso 2002). The dataset can be regarded as conditionally sampled for cloud-free conditions. We therefore conditionally sampled our model output in the same way as is done in the retrieval of $T_{IR}$ from satellite infrared measurements: only the cloud-free grid points are included in the monthly averaging. As a selection criterion, we use a threshold ($T$) for IWP of 5, 10, or 20 g m$^{-2}$, or a cloud cover of 50%. We calculated the same composites of SAM+ and SAM− for the conditionally sampled model temperature (COND; Fig. 12b).
and compared this with the control composites (CTL; Fig. 6a), where all grid boxes (both cloudy and cloud free) were taken into account. The sensitivity to the definition of the thresholds is small; therefore we only present the results for $H_9270/H_11005$.

Conditional sampling for cloud-free conditions results in a cold bias (reduction of the warming or even a cooling) over the northwestern side (windward side) of the AP and a warm bias (enhancement of the warming) over the northeastern side (leeward side; cf. Figs. 6a, 12b). Therefore the temperature-gradient anomaly across the northern AP is larger in COND than in CTL. This occurs because the lee side is clearer and the windward side is cloudier during SAM (Fig. 6c). Since near-surface temperature and cloud cover are positively correlated, more warm (cloudy) days are excluded in the averaging procedure in COND on the windward side during SAM than during SAM. Therefore the COND composite has a cold bias in this region compared to the CTL composite. At the lee side the physical reasoning is similar. More warm (cloudy) days are excluded in the averaging procedure in COND during SAM than during SAM. Therefore, the COND composite has a warm bias in this region compared to the CTL composite.

Although many differences exist between the signal derived from $T_{IR}$ (Fig. 12a) and that from the conditionally sampled model output (Fig. 12b), both $T_{IR}$ and COND show a SAM response that is amplified negatively on the windward side of the peninsula and positively on the lee side compared to the CTL map (Fig. 6a). This result implies a cold bias on the windward side when a dataset is conditionally sampled for cloud-free conditions, which is the case for the $T_{IR}$ dataset. This effect might contribute to the strong negative response of the $T_{IR}$ to the SAM over land grid points (Fig. 12a), which occurs mostly on the western side of the peninsula where the difference in cloud cover between SAM+ and SAM− is the largest.

The differences that remain between the signal derived from $T_{IR}$ (Fig. 12a) and that from the conditionally sampled model output (Fig. 12b) indicate that the bias in the $T_{IR}$ dataset, which is caused by the absence of data for days with cloud cover, can explain only part of the difference between the model output and the $T_{IR}$ dataset. The derivation of the surface temperature from the satellite infrared radiometer data relies on the identification of clouds. When clouds are covering the scene but are not identified, then the retrieval will return the cloud-top temperature instead of the surface temperature. But it is also possible that model deficiencies (e.g., the cloud scheme) might be responsible for these differences between model and satellite observations.

Fig. 12. Normalized difference in DJF near-surface temperatures between SAM+ and SAM− as derived from (a) satellite infrared radiometer data and (b) conditionally sampled model output for cloud-free conditions.
b. Sensitivity to model grid spacing

In a model with a coarser resolution (namely with a grid spacing of 55 km; R55), the response of the modeled large-scale meteorological conditions to the SAM is similar to the high-resolution model output (14 km; R14) presented here (not shown). One exception, however, is the northern AP region. In the Larsen-B cross section, the potential temperature difference between SAM$^+$ and SAM$^-$ ($\Delta \theta / \Delta \text{SAM}$) is much larger for R14 than for R55 (Figs. 7c, 13a). It is likely that this is related to a smoothing of the orography in R55. The difference in vertical velocity between SAM$^+$ and SAM$^-$ is found to be much smaller in R55 than in R14 in the northern AP region (Figs. 10c, 13b), which is consistent with the smaller value for $\Delta \theta / \Delta \text{SAM}$. This result shows that a high resolution is needed to obtain the strong upward and downward motion and associated warming in full detail.

5. Conclusions

We have used a regional atmospheric model climatology for the period of 1987–93 at a 14-km grid spacing to study the mechanisms behind the spatial distribution of the sensitivity of the summer near-surface temperatures to the SAM. For this, the seven-year integration has been divided into two composites: the nine summer months with the highest values for the SAM index (SAM$^+$), defined by the University of Washington, and the nine summer months with the lowest values for the SAM index (SAM$^-$; Table 3). The model is able to represent the synoptic situation and the month-to-month variations of the surface temperatures in summer. In addition, the modeled temperature sensitivity to the SAM corresponds to the in situ measured sensitivity, although the effect of sea surface temperature anomalies is overestimated at some sites, indicating that even a grid spacing of 14 km is insufficient to resolve all local effects.

The largest difference in temperature between SAM$^+$ and SAM$^-$ is found over the eastern side of the northern AP. A contrasting response to the SAM is found between the eastern and the western sides of the barrier: at a cross section through the Larsen-B Ice Shelf, the sensitivity of the potential temperature to the SAM ($d \theta / d \text{SAM}$) is 0.2°C on the windward side, whereas it is 1.5°C per unit change of the SAM index at the lee side of the barrier.

Our work clearly supports the hypothesis that this different response between the eastern and western sides is due to the strengthening of the westerlies, as proposed by Marshall et al. (2006). Stronger summer westerly winds reduce the frequency of occurrence of blocked flow conditions. Air masses are advected eastward over the orographic barrier of the northern AP more frequently. The modeled warming on the eastern side is the result of an increase in warm-air advection from the western side of the AP and orographically forced downward motion (föhn effect). Since dry air with a high potential temperature is present at upper levels, the enhanced subsidence leads to a warming and drying of the air near the mountain base. During flow-over conditions, the downstream cold air mass that is advected from cold continental regions in the south is replaced by the warm dry air, advected from upper-atmospheric levels. A grid spacing on the order of 14 km is absolutely needed to resolve the processes associated with the leeside warming: upward and downward motions are much weaker in an integration with the same model at a 55-km grid spacing. Moreover, the
significant difference in surface temperature response to the SAM between the eastern and western sides of the AP is not found in ERA-40, which operates at a grid spacing of about 125 km (T159).

At the Esperanza station, the summer temperature has risen by 2.1°C over the period of 1965–2004, which is much larger than the global average. A large part (35%–60%) of the summer warming at Esperanza is estimated to be from changes in the SAM (Marshall et al. 2006). While our model results indicate a similar large temperature sensitivity to the SAM at the east coast station of Esperanza, the strongest sensitivity occurs in a region located somewhat farther south over the Larsen-B Ice Shelf, where no long-term measurements exist. At this location, the barrier is high enough (1.4 km) to induce a strong downward motion that results in the advection of dry air with a high potential temperature from upper levels. Any such warming would be less dramatic over the far northern tip at Esperanza where the height of the barrier is about 500 m. South of 68°S, where the AP is higher (e.g., >2000 m), the typically weaker westerlies resulted in the flow regime remaining blocked during SAM+.

The region of maximum modeled response to a positive phase shift of SAM over land ice and ice shelves is similar to the region of maximum response as derived from satellite infrared radiometer data ($\Delta T_{IR}/\Delta$SAM). However, there are also discrepancies between model- and satellite-derived responses of temperature to SAM, especially over the land ice where the satellite-derived $\Delta T_{IR}/\Delta$SAM is negative. Part of this can be explained by the fact that satellite-derived temperature is conditionally sampled for cloud-free conditions (Comiso 2000; Shuman and Comiso 2002), resulting in a cold bias (reduction of the warming or even a cooling) over the northern side of the AP and a warm bias (enhancement of the warming) over the northeastern side.

The modeled surface-mass balance, cloud cover, and vertically integrated cloud ice/water is found to increase on the windward side and decrease on the northern leeward side of the barrier during SAM+. The region of strongest temperature sensitivity to the SAM is where ice shelf collapse has taken place and does not extend farther south over the Larsen-C Ice Shelf. If SAM-induced warming did play an important role in the breakup of its more northerly neighbors, then the Larsen-C may not be under threat of imminent collapse, because our analysis shows it is not as sensitive to SAM variability. The extreme sensitivity of summer temperatures in the northeastern part of the AP to the SAM may explain why the Prince Gustav (and Larsen-A) Ice Shelves have collapsed and reformed at least once during the Holocene (Pudsey and Evans 2001).

**Acknowledgments.** We thank scientists from British Antarctic Survey (BAS) for valuable discussions on the climatology of the Antarctic Peninsula; Dr. Michiel van den Broeke (Institute for Marine and Atmospheric Research, Utrecht) for exchanging ideas on how to study the effect of the SAM on Antarctic meteorological conditions; and Dr. Jan Schween (Universität zu Köln) for discussing the statistical relevance of our results. We thank three anonymous reviewers for their valuable comments on the manuscript and Dr. J. C. Comiso for providing us with the satellite infrared temperature dataset. Part of this work was supported by the European Commission under a Marie Curie Fellowship and by the Computer Services for Academic Research (CSAR) for the use of High-Performance Computing facilities.

**REFERENCES**


