RECENT VARIABILITY AND TRENDS IN ANTARCTIC SNOWFALL ACCUMULATION AND NEAR-SURFACE AIR TEMPERATURE

A Dissertation

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By

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Approximately 6 mm of sea level equivalent falls as snow on Antarctica each year, comprising the dominant term in the surface mass balance - the net accumulation of moisture – over the continent. Therefore, short- or long-term fluctuations in the annual amount of snowfall may have a substantial impact on the surface mass balance and eustatic global sea level change. Due to the continent’s inaccessibility and the inherent shortage of observational data, Antarctic snowfall accounts for substantial uncertainty in global sea level change estimates. Taken over the grounded ice sheet, even the sign of the snowfall trends – whether they are contributing to, or mitigating sea level rise – is a matter of intense debate. The work presented here uses a variety of methods to better understand how Antarctic snowfall, as well as near-surface air temperature (an important modulator of snowfall), have been changing in the later decades of the 20th century.

Polar MM5, a mesoscale atmospheric model optimized for use over polar ice sheets, is employed to simulate Antarctic snowfall over the past two decades. Two sets of simulations, each with different initial and boundary conditions, are evaluated for the 17-y period spanning 1985-2001. The initial and boundary conditions for the two sets of runs are provided by the (1) European Centre for Medium-Range Weather Forecasts 40-year Reanalysis, and (2) National Centers for Environmental Prediction – Department of
Energy Atmospheric Model Intercomparison Project Reanalysis II, an approach used so that uncertainty can be assessed by comparing the resulting datasets.

The simulated snowfall changes are in general agreement with ice core and snow stake accumulation records at various locations across the continent, indicating broad areas of both positive and negative trends. Averaged over the continent, the annual snowfall trends in both Polar MM5 datasets are not statistically different from zero, suggesting that recent Antarctic snowfall changes do not mitigate currently observed sea level rise. However, the lack of a continent-averaged annual trend does not suggest that Antarctica is isolated from the recent climate changes occurring elsewhere on Earth. Rather, snowfall variability is expressed by strong seasonal and regional changes.

The Polar MM5 simulations are useful for assessing snowfall variability from the mid-1980s onward, but because the accuracy of the model initial conditions relies on the quality and volume of satellite data over the otherwise data sparse Antarctic, atmospheric models and reanalyses are known to be unreliable over Antarctica prior to the modern satellite era, which begins around 1980. However, by blending the spatial information provided by the model fields during the post-1980 era with scores of new ice core measurements in the pre-1980 era, it has been possible to reconstruct Antarctic snowfall with spatial and temporal continuity over 1955-2004. The resulting dataset is consistent with the results from the shorter, 1985-2001 Polar MM5 assessment, indicating that there has been no significant net change in Antarctic snowfall since the 1950s, and thus Antarctic snowfall is not mitigating observed global sea level rise as expected, despite recent warming of the overlying atmosphere.
The observational analysis technique devised to ‘fill in the gaps’ between the ice core records to reconstruct snowfall can be applied to other fields in Antarctica for which few observations exist. A new Antarctic near-surface temperature dataset spanning 1960-2005 is constructed using the same methodology used for snowfall. The new dataset is compared with other observationally-based Antarctic near-surface temperature datasets for the past ~45 years and all are in good agreement at annual and seasonal timescales. The new snowfall and near-surface temperature records, which are representative of the entire continent, are useful for assessing global climate model (GCM) simulations of Antarctic near-surface temperature during the 20th century, and thus the reliability of 21st century Antarctic climate change scenarios. Five 20th century GCM ensembles run in support of the Intergovernmental Panel on Climate Change Fourth Assessment Report are evaluated. It is found that the GCMs overestimate annual Antarctic near-surface temperature trends on average by a factor of 3 during the 20th century, and by more than 5 times during the latter half of the 20th century. The positively-biased Antarctic temperature increases in the GCMs appear to be due to an increase in total column water vapor that enhances the downward longwave radiation incident at the surface. The relationship between Antarctic near-surface temperature changes and the Southern Hemisphere Annular Mode, the primary mode of Antarctic atmospheric variability, is of secondary importance in the GCMs compared to the water vapor feedback.

Snowfall variability in the GCMs is compared to the 5-decade record of Antarctic snowfall described above. The positive GCM snowfall trends agree well with the positive trends in the observational record that occur into the 1990s. However most of the GCM runs end in the late 1990s, when the observed snowfall sharply decreases, so
uncertainty remains as to whether the models would have captured the downward fluctuation. The GCMs are able to accurately simulate the observed sensitivity of Antarctic snowfall to near-surface temperature of about $+5-6 \, \% \, K^{-1}$, suggesting that if Antarctic near-surface temperature increases by about 2-3.5 K by the end of the 21st century as the GCMs predict, snowfall will increase by about 10-20%, having a negative impact on sea level of about -0.5 to -1.0 mm y$^{-1}$ by 2100. Properly simulating Antarctic temperature in GCMs is thus critical for understanding how Antarctic snowfall will change and consequently affect global eustatic sea level rise.
To my parents, whose hard work, integrity, sacrifice and selflessness are inspirational.
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CHAPTER 1

INTRODUCTION

1.1 Project Summary

Snowfall accumulation is the driving mass input to the Antarctic ice sheet, and is the net result of precipitation, sublimation/vapor deposition, drifting snow processes, and melt (Bromwich 1988). Summing all of the components that influence snowfall accumulation results in surface mass balance, B:

\[ B = P - S_{sfc} - S_{blsn} - D_{blsn} - M \]  \[1\]

where \( P \) is precipitation (mainly in the form of snowfall), \( S_{sfc} \) is sublimation/deposition at the surface, \( S_{blsn} \) is sublimation of blowing snow particles, \( D_{blsn} \) is divergence of blowing snow, and \( M \) is melt water runoff. The closure of the Antarctic ice sheet mass budget is not well constrained, and thus accounts for the largest single uncertainty in global sea level budget change estimates (Vaughan 2005). Taken over the grounded ice sheet, even the sign of the surface mass balance trends – whether they are contributing to, or mitigating, sea level rise – is a matter of intense debate (e.g., Van de Berg et al. 2005, Davis et al. 2005).

The work presented herein uses a variety of techniques to ascertain how the Antarctic surface mass balance has been changing in recent decades. In Chapter 2 Polar MM5, a mesoscale atmospheric model optimized for use over polar ice sheets (e.g., Bromwich et al. 2001, Cassano et al. 2001), is employed to simulate the largest components of the...
Antarctic surface mass balance, that is, precipitation and sublimation. The results demonstrate that Polar MM5 is able to reproduce Antarctic snowfall accumulation variability and change that is comparable to the observational records over the past two decades. However, the Polar MM5 simulations are dependent on the accuracy of which large-scale reanalysis that is used to provide the initial and boundary conditions. Due to the poor quality of both of the existing long-term reanalysis datasets in simulating surface and mid-troposphere atmospheric variability over Antarctic and the Southern Ocean prior to the modern satellite era (e.g., Bromwich and Fogt 2004) the Polar MM5 results are likely to be unreliable before 1979.

In Chapter 3, a second method is devised to extend the record of Antarctic snowfall accumulation to the decades prior to the modern satellite era. The technique exploits the relatively large number of ice core accumulation records that have become available in recent years due to glaciological field programs, mainly the International Transantarctic Scientific Expedition (ITASE; Mayewski and Goodwin 1997). Combining the accumulation changes indicated by these ice core records with contemporary model simulations of snowfall accumulation provides a means by which to spatially extrapolate the ice core records back in time. By this technique, a high resolution depiction of Antarctic surface mass balance back to the 1950s can be obtained. The end result is the best estimate yet of surface mass balance changes over Antarctica for the past ½ century, including the first reliable estimate prior to the modern satellite era.

In Chapter 4, the same technique is used to extend the Antarctic near-surface temperature record back to 1960, just after the permanent observing network was established. Together, these two long-term records are used to evaluate simulations of
Antarctic snowfall (P) and temperature from global climate models (GCMs) for the 20th century. The results provide a proxy of the reliability of the model projections for the end of the 21st century, which in turn has important implications for ice sheet stability and global sea level rise.

1.2 Components of the Antarctic Surface Mass Balance

1.2.1 Total Surface Mass Balance (B)

In recent decades, estimates of surface mass balance over the Antarctic ice sheets have been made by three techniques: in-situ observations, remote sensing, and atmospheric modeling. Constructing a reliable data set of snowfall accumulation over Antarctica for a long time period from these methods has been difficult for numerous reasons, including for example a sparse surface observational network (e.g., Giovinetto and Bentley 1985); difficulties distinguishing between clouds and the Antarctic ice surface in satellite radiances (Xie and Arkin 1998); and incomplete parameterizations of polar cloud microphysics and precipitation in atmospheric models (Guo et al. 2003). Considering the limitations of the techniques, it is not surprising that the long-term-averaged continent-wide maps of surface mass balance over Antarctica yield a broad envelope of results. The long-term estimates of B from several studies range from +119 mm yr\(^{-1}\) (van de Berg et al. 2005) to +197 mm yr\(^{-1}\) (Ohmura et al. 1996) water equivalent (weq) for the grounded ice sheet (estimates for the conterminous ice sheet, which includes the ice shelves, are generally ~10% higher). In general, the studies employing glaciological data are considered the most reliable; the study of Vaughan et al. (1999) approximates B=149 mm yr\(^{-1}\) for the grounded ice sheet, although a recent study (van de Berg et al. 2006) shows
evidence that the Vaughan et al. (1999) dataset may underestimate coastal accumulation, and gives an updated value of 171 mm yr\(^{-1}\). Considering the large spread between estimates, it is not surprising that calculated temporal trends vary widely (Table 1.1). Uncertainties in B provide strong motivation for this work.

1.2.2 Precipitation (P)

Precipitation, nearly all of which is snowfall over Antarctica, is the dominant term of the surface mass balance at regional and larger scales (Bromwich 1988) and establishing its spatial and temporal variability is essential. Comprehensive studies of precipitation characteristics over Antarctica include Bromwich (1988), Turner et al. (1999), Genthon and Krinner (2001), van Lipzig et al. (2002) and Bromwich et al. (2004a). Antarctic precipitation is influenced to first order by the surface topography. Most of the snow falls along the steep coastal margins and is caused by orographic lifting of moisture-laden air associated with the many transient, synoptic-scale cyclones that encircle the continent, gathering moisture from the surrounding Southern Ocean. The influence of synoptic activity decreases with distance from the coast. Over the highest, coldest reaches of the continent the primary mode of precipitation is due to cooling of moist air just above the surface-based temperature inversion. The air has little moisture-bearing capacity due to its extremely low temperature year-round, and thus the interior of East Antarctica is a so-called polar desert, having a surface area of about 2.5x10\(^6\) km\(^2\) (larger than Greenland) that receives less than 50 mm yr\(^{-1}\) weq (e.g., Vaughan et al. 1999, Giovinetto and Zwally 2000).
Precipitation recycling (due to locally sublimated snow) is a term that in the future should be considered in equation [1], as the recycled component of Antarctic snowfall does not have an impact on eustatic sea level change. Trenberth (1998) evaluates moisture recycling over the high and middle latitudes of the globe and finds that it comprises nearly 20% of precipitation at 1000-km length scales on average. Delaygue et al. (2000) assess the source regions of Antarctic precipitation with a global climate model and report that only ~2% is recycled due to the low evaporation rates over Antarctica compared to other regions of the globe. Delaygue et al. (2000) do not include estimates of blowing snow sublimation, so they likely underestimate Antarctic precipitation recycling.

In the following chapters, precipitation will be derived primarily from model simulations that include global reanalyses, as well as Polar MM5. A description of the Polar MM5 precipitation scheme can be found Grell et al. (1994). Large- (grid-) scale precipitation processes are represented by the explicit microphysics parameterization of Reisner et al. (1998), while sub-grid-scale clouds and precipitation are parameterized with the Grell cumulus parameterization (Grell et al., 1994). Large-scale processes dominate precipitation over Antarctica in the Polar MM5 simulations.

1.2.3 Surface Sublimation (SU_{sfc})

Estimates of SU_{sfc} in Antarctica have been made based on observations from automatic weather stations (AWSs; e.g., Stearns and Weidner 1993, van den Broeke et al. 2004, 2005, van As et al. 2005), and for larger scales, from atmospheric models (e.g., van den Broeke 1997, Déry and Yau 2002, Bromwich et al. 2004a). Stearns and Weidner
(1993) report that $SU_{sfc}$ (both deposition and sublimation) amounts to 20-80% of the annual accumulation on the Ross Ice Shelf, but caution that the results are highly uncertain due to instrumental problems measuring relative humidity at low temperatures, unknown instrument heights, and uncertain snowfall observations. Estimates from four weather stations in Dronning Maud Land indicate that 3-7% of annual snowfall is removed by $SU_{sfc}$, with the largest values occurring in summer and during strong katabatic wind events (van den Broeke et al. 2004). A study by van As et al. (2005) at Kohnen base (75°S, 0°E), a site that is representative of the vast high plateau of East Antarctica, shows that $SU_{sfc}$ is a negligible component of the diurnal surface energy balance fluctuation during summer, the season for which sublimation is typically largest. A study of the annual cycle of surface energy balance for a transect of weather stations from a coastal ice shelf to the East Antarctic plateau shows that the stations at lower elevations on the coastal slopes exhibit significant sublimation in summer due to an increase in net radiation and from occasional katabatic winds (van den Broeke et al. 2005). Modeling studies indicate that surface sublimation removes about 10-15% of the annual precipitation over Antarctica (e.g., van den Broeke 1997, Déry and Yau 2002, Bromwich et al. 2004a).

The Polar MM5 $SU_{sfc}$ is employed in the figures discussed in section 1.3, and in Chapter 2 (denoted there as ‘E’). It is derived from the modeled latent heat flux, which in turn is parameterized with the 1.5-order turbulence closure scheme used in the National Centers for Environmental Prediction Eta model (Janjic, 1994). The parameterization calculates exchange coefficients using similarity theory, and calculates the vertical fluxes
with an implicit diffusion scheme. For example, the moisture flux at the lowest model near the surface is specified as (Janjic, 1994; eqn 4.17):

\[
\left( \frac{\lambda}{z_q} \right)(q_0 - q_s) = K_{Hsfc} \left( \frac{(q_{lm} - q_0)}{(z_{lm} - z_0)} \right) \tag{2}
\]

where \( q \) is specific humidity (kg kg\(^{-1}\)), \( z \) is height above the surface (m), \( K_{Hsfc} \) is the exchange coefficient (m\(^2\) s\(^{-1}\); identical to that for heat), \( \lambda \) is the molecular diffusivity for water vapor (m\(^2\) s\(^{-1}\)), \( z_q \) is the roughness length for moisture (m), the subscript “lm” specifies the lowest model level, and the subscript “0” specifies the boundary between the thin viscous sublayer in contact with the surface and the turbulent layer above it.

1.2.4 Blowing Snow Processes (SU\(_{blsn}\) and D\(_{blsn}\))

Blowing snow is known to have a substantial - in some cases first-order – impact on surface mass balance at local space scales over Antarctica (Bintanja 1998, Gallée 1998, Mann 1998, Bintanja 2001, Déry and Yau 2002, Frezzotti et al. 2005). Three modes of snow transport by wind are (1) creep, (2) saltation, and (3) suspension, with suspension (D\(_{blsn}\)) being by far the most important in terms of mass transport (Pomeroy and Gray 1995). The sublimation of blowing snow particles (SU\(_{blsn}\)) removes moisture from the surface (e.g., Bintanja and Reijmer 2001). SU\(_{blsn}\) is a function of ice particle size, water vapor diffusivity, turbulence near the particle surface, and the water vapor density of ambient air and air near the particle surface (e.g., Landine and Gray 1989).

Several investigators have measured, or estimated via empirical and physically-based models, wind-driven snow transport over Antarctica. One early and still valuable study with the goal of quantifying D\(_{blsn}\) was the Byrd Snow Drift Project (Budd et al. 1966).
Measurements taken near Byrd Station (80°S, 120°W) led to the development of a simple empirical function of wind speed for $D_{\text{blsn}}$ between 1 mm and 300 m above the surface. Despite its simplicity, Budd’s relationship has endured and was recently applied to Polar MM5 wind fields to estimate annual $D_{\text{blsn}}$ over Antarctica by Bromwich et al. (2004a). Bromwich et al. (2004a) demonstrated that $D_{\text{blsn}}$ occurs mainly in coastal escarpment areas and is localized, with regions of convergence adjacent to regions of divergence having space scales of $1 \times 10^2$ km, a finding that is consistent with studies using more complex models (e.g., Déry and Yau 2002; a map demonstrating the space scales is presented later). Loewe (1970) made detailed measurements in the layer below 10 m where most transport occurs, and combined his own algorithm with that of Budd et al. (1966) for heights above 10 m to refine calculations of $D_{\text{blsn}}$ over Antarctica. Loewe (1970) used the equations to estimate blowing snow transport across the edge of the Antarctic ice sheet, and found it to be about 5% of the annual snow accumulation, a number that compares well with recent studies (Giovinetto et al. 1992, Déry and Yau 2002, Bromwich et al. 2004a, van den Broeke et al. 2006b). Wendler (1989) measured blowing particle size and distribution in one of the windiest areas of the continent, Adelie Land. The mean annual wind speed at weather station D47 (67.4°S 138.7°E), where the study was conducted, is 12.8 m s$^{-1}$; instantaneous values commonly exceed 20 m s$^{-1}$ (Wendler 1989). Wendler estimated $D_{\text{blsn}}$ in coastal Wilkes Land on the order of $6.3 \times 10^6$ kg m$^{-1}$ yr$^{-1}$. Wendler’s estimate is comparable to Loewe (1970) for the same area, and about an order of magnitude greater than Kobayashi (1978) found at Mizuho (77.6°S, 44.3°E), an area with weaker winds ($9.9$ m s$^{-1}$ annual average; Kawaguchi 1979).
The potential importance of the sublimation of blowing snow particles, and the consequence for the Antarctic surface mass balance, is emphasized by a number of recent observational and modeling investigations. A field program in Antarctica that emphasized measuring the heat and vapor fluxes due to blowing snow was the British Antarctic Survey Second Stable Antarctic Boundary Layer Experiment (STABLE2), at Halley Station (75.6°S, 24.6°W; Dover 1993, Mann 1998) between March and November 1991. Measurements of wind speed, air temperature, humidity, air pressure, and snow particle size and distribution were taken at various heights on a 30 m mast (15 m was the upper limit of the $SU_{blsn}$ observations). One important finding was that during blowing snow events, $SU_{blsn}$ is strongly self-limited by the formation of a layer of near 100% relative humidity in at least the lowest 12 m of the boundary layer, and this restricts its magnitude during winter to about 10% of total snow accumulation for the same period. Modeling studies of annual $SU_{blsn}$ over the entire continent are in general agreement with this conclusion; they indicate $SU_{blsn}$ removes about 10% of the annual snowfall from the Antarctic ice sheet (Bintanja 1998, Déry and Yau 2002, van den Broeke et al. 2006b).

In section 1.3 the relative components of the Antarctic surface mass balance are evaluated. Unlike $P$ and $SU_{sfc}$, which are simulated by Polar MM5, $SU_{blsn}$ and $D_{blsn}$ must be estimated with ‘offline’ parameterizations (i.e., Bromwich et al. 2004a, Box et al. 2004). Eventually a more complex blowing snow model will be coupled to Polar MM5 (or the successor to Polar MM5, Polar WRF). Calculating $SU_{blsn}$ ‘inline’ will allow two-way interaction of momentum, heat, and moisture between the blowing snow model and atmospheric model. The calculation of $SU_{blsn}$ employed here is based on the same ‘bulk’ methodology typically used to derive $SU_{sfc}$, which employs aerodynamic roughness
lengths as a means of controlling the rate of the turbulent exchange of moisture between the surface and overlying atmosphere. During blowing snow episodes the aerodynamic roughness lengths for momentum and moisture are increased as a power function of wind speed to match observed values for blowing snow conditions, as characterized by Bintanja and Reijmer (2001). The result is the total sublimation, $S_{U_{\text{tot}}} (S_{U_{\text{sfc}}} + S_{U_{\text{blsn}}})$. In theory $S_{U_{\text{sfc}}}$ and $S_{U_{\text{blsn}}}$ can be separated since the near-surface layer becomes quickly saturated during blowing snow events and effectively shuts down the surface sublimation. Therefore, under a threshold wind speed (or friction velocity), only surface sublimation occurs, and above it, only blowing snow sublimation occurs. In practice, the specified threshold wind speed does not partition the surface and blowing snow contributions well, but overall, the resulting $S_{U_{\text{tot}}}$ compares reasonably with similar studies (discussed below). The technique is formulated following van den Broeke et al. (2004) as:

$$S_{U_{\text{tot}}} = S_{U_{\text{blsn}}} + S_{U_{\text{sfc}}} = \rho u_* q_*$$  \[3\]

where $\rho$ is the density of the near surface air, and $u_*$ and $q_*$ are the turbulent scaling parameters of momentum and moisture, given by the ‘bulk’ method (Denby and Greuell 2000):

$$u_* \cong k \frac{[V(z_v) - V(z_{o,v})]}{\ln \frac{z_v}{z_{o,v}} - \Psi_m \left( \frac{z_v}{L_{mo}} \right)}$$  \[4, 5\]

$$q_* \cong k \frac{[q(z_q) - q(z_{o,q})]}{\ln \frac{z_q}{z_{o,q}} - \Psi_k \left( \frac{z_q}{L_{mo}} \right)}$$
where, \( V(z_v) \) is the wind speed at height \( z_v \) (10 m in this case), \( V(z_{o,v}) \) is the wind speed at the surface (assumed to be zero), \( z_{o,v} \) is the roughness length for momentum, \( q(z_q) \) is the specific humidity at height \( z_q \) (2 m in this case), \( q(z_{o,q}) \) is the specific humidity at the surface (assumed to be saturated with respect to ice or water), \( z_{o,q} \) is the roughness length for moisture, \( L_{mo} \) is the Monin-Obukhov length scale, and \( \psi_m \) and \( \psi_h \) are stability corrections for momentum and heat/moisture. The stability corrections are neglected here following van den Broeke et al. (2004), who note that the profile functions that account for stable/unstable conditions are no longer valid in the presence of a moisture source and heat sink (i.e., blowing snow) because the new source/sink terms are not accounted for. The equations for \( z_{o,v} \) and \( z_{o,q} \) are based on observations and are given by Bintanja and Reijmer (2001):

\[
\begin{align*}
    z_{o,v} &= 0.0039202 \times u_*^{2.1968} \\
    z_{o,q} &= 0.50324 \times u_*^{0.1141}
\end{align*}
\]

An upper bound of 0.1 m is imposed for \( z_{o,v} \) and \( z_{o,q} \) per van den Broeke et al. (2004), and a lower bound (for calm conditions) of 0.0002 m is imposed per Bintanja and Reijmer (2001) when \( u_* \) falls below 0.3 m s\(^{-1}\). An iterative procedure is used to solve for \( z_{o,v} \) and \( u_* \), and then the other terms are calculated in a strait-forward manner.

The drift snow divergence is solved using the relationship of Budd et al. (1966) and employed previously in Polar MM5 by Bromwich et al. (2004a):

\[
\log \left| Q_{10^{-3}}^{300} \right| = 1.1812 + 0.0887 \left| V_{10} \right| 
\]

\[
D_{blw} = \nabla \cdot \left( \frac{Q_{10^{-3}}^{300}}{V_{10}} V_{10} \right)
\]
where $Q_{10}^{300}$ (g m$^{-1}$ s$^{-1}$) is the drift snow in the layer from 1 mm to 300 m above the surface and $V_{10}$ is the 10 m wind speed. It is noteworthy that this methodology assumes an infinite supply of snow is available and neglects the effects of sublimating particles, and thus is likely to overestimate $D_{blsn}$. However, the results for both $D_{blsn}$ and $SU_{blsn}$ agree well with those obtained with a bulk blowing snow model coupled to an atmospheric model (Déry and Yau 1999, Déry and Yau 2002).

### 1.2.5 Surface Meltwater Runoff (M)

Previous modeling studies indicate that melt water production and runoff plays a minor role in the overall surface mass balance of Antarctica at present (e.g., van de Berg et al. 2005). For the northern Antarctic Peninsula, however, $M$ is an important term (Fahnestock et al. 2002, Torinesi et al. 2003) during summer, but the northern Peninsula is a small region compared to the ice sheet. Liston and Winther (2005) modeled the surface and subsurface meltwater production and found that it comprises a considerable portion of the total surface mass balance, about 31 mm yr$^{-1}$. However, most of the water refreezes near where it formed, and thus is inconsequential to the surface mass budget, although there are obvious implications for the interpretation of ice core stratigraphic records. The bulk of the melt occurs beneath the surface; only about 3.5 mm yr$^{-1}$ occurs at the surface. Van den Broeke et al. (2006b) estimate similar values of surface melt of about 1-6 mm yr$^{-1}$ from 1980-2004. Considering this is the total surface melt, much of which refreezes, runoff (M) is presumably much less. Meltwater runoff is neglected in this study; the previous results suggest the omission of M will have little impact on the overall results.
1.3 Examination of the Antarctic Surface Mass Balance

Figure 1.1 shows the sublimation and drift snow components of the surface mass balance for Antarctica, averaged for 1985-2001. SU_sfc (Fig. 1.1a) is $<\pm 5$ mm y$^{-1}$ in the interior where temperatures are extremely low year round (e.g., Bromwich and Parish 1998). Over much of the East Antarctic plateau and the higher-elevation regions of West Antarctica, there is net annual deposition (Kameda et al. 1997). Along the coastal margins, where temperatures are highest and winds are strongest due to steeper slopes, SU_sfc is much greater, approaching 300 mm y$^{-1}$. The calculated SU_tot (Fig. 1.1b) is similar to SU_sfc, but is on average about twice the magnitude (SU_tot=32 mm y$^{-1}$ vs. SU_sfc=17 mm y$^{-1}$ averaged annual over Antarctica). As SU_tot is the sum of SU_sfc+SU_blso, SU_blso (Fig. 1.1c; the difference of 1.1b minus 1.1a) is roughly the same magnitude as SU_sfc (SU_blso=32-17=15 mm y$^{-1}$ averaged annually over Antarctica). Other studies have also noted that the two terms are approximately equal (Déry and Yau 2002, van den Broeke et al. 2006b). It is noteworthy that the calculated SU_tot has greater values of deposition (compared to SU_sfc) over the plateau regions of East and West Antarctica, as well as on the eastern Ross Ice Shelf. This is because the minimum roughness length for SU_tot is larger than that used in Polar MM5 to calculate SU_sfc, even during calm periods with strong inversions (when the moisture gradient is directed downward). D_blso (Fig. 1.1d; here multiplied by -1 so that convergence is positive) indicates that snow on average is transported from the interior plateau downwind (and downslope) toward the coast. Along the coastal margins are alternating regions of convergence and divergence, suggesting that much of the drift snow transport occurs in regions adjacent to each other.
This map broadly agrees with the maps produced by Déry and Yau (2002), and van den Broeke et al. (2006b), although regional differences exist. For example, Fig. 1.1d indicates a broad region of divergence in the Amery Ice Shelf basin (60°-75°E), in agreement with Déry and Yau (2002), but in disagreement with van den Broeke et al. (2006b). Much of the Amery Ice Shelf Basin surface is characterized by blue ice (Winther et al. 2001), suggesting that divergence may more accurately describe the blowing snow processes in the region. The algorithm used by van den Broeke et al. (2006b) uses annual mean wind vectors as a proxy of $D_{blsn}$, whereas the algorithm used here (equation [9]), and in Déry and Yau (2002) uses 6-hourly wind vectors to calculate $D_{blsn}$, then sums the results for each year. The relationship between $Q_{10^{-3}}^{300}$ and wind speed is logarithmic (equation [8]), indicating that strong wind episodes during the year play an important role in the annual total $D_{blsn}$. The use of annual mean wind speed to calculate $D_{blsn}$ by van den Broeke et al. (2006b) neglects strong wind episodes and thus may not properly characterize $D_{blsn}$ in the Amery Ice Shelf Basin.

Figure 1.2 shows sublimation and drift snow components of the surface mass balance as a fraction of the annual snowfall (P). $SU_{sfc}$ is less than 5% of P over interior East Antarctica, and nearly all of West Antarctica and the Peninsula. $SU_{sfc}$ becomes larger in the katabatic wind zones of coastal East Antarctica with a few small pockets in mountainous regions where it exceeds P. As expected, the fraction comprised by $SU_{tot}$ is roughly double that of $SU_{sfc}$, with a similar spatial distribution and maxima in the same regions. Interestingly, as a fraction of precipitation $D_{blsn}$ is largest over the interior of the continent, in contrast to $SU_{sfc}$ and $SU_{tot}$, which are limited due to extremely low temperatures and periods of deposition during winter (Bromwich and Parish 1998).
Combined, $SU_{tot}$ and $D_{blsn}$ (Fig. 1.2d) remove/redistribute $>50\%-200\%$ of snowfall in some coastal areas of the ice sheet where high snowfall rates, strong winds, and favorable topography are found.

Figure 1.3 shows various total mass balance estimates, ranging from $P$ to $P+SU_{tot}+D_{blsn}$. The inclusion of $SU_{sfc}$ (Fig. 1.3b) does not alter the map substantially (compared to Fig. 1.3a). Accounting additionally for the blowing snow components (Figs. 1.3c and d) removes up to $200\%$ of $P$ in some coastal pockets, for the reasons described in the previous paragraph. The gray areas in the Figs. 1.3c and d indicate regions where a net loss of snow/ice occurs annually. The locations coincide with many blue ice areas observed in satellite imagery (Winther et al. 2001). It is noteworthy that the surface mass balance estimate shown in Fig. 1.3d ($P+SU_{tot}+D_{blsn}$) has smaller differences along the coastal margins when compared to the glaciological compilation of Vaughan et al. (1999) than the estimate using $P+SU_{sfc}$ (Fig. 1.3b). A figure of the latter versus the Vaughan et al. (1999) compilation is shown in Chapter 2 (Fig. 2.3). The inclusion of blowing snow processes ($P+SU_{tot}+D_{blsn}$) reduces the modeled surface mass balance along the steep coastal slopes on average by a factor of two compared to $P+SU_{sfc}$.

Figure 1.4 shows time-series of the areal-averages of the surface mass balance and associated components over the grounded ice sheet from 1985-2001. The total surface mass balance is best represented by \( "p-su(calc)" \), which is $P+SU_{tot}$ (and $SU_{tot}$ is $SU_{sfc}+SU_{blsn}$). $D_{blsn}$ is negligible when integrated over the ice sheet (<1 mm y$^{-1}$), in agreement with the results of Déry and Yau (2002), Bromwich et al. (2004a), and van den Broeke et al. (2006b). Given this, Fig. 1.4a suggests that snowfall ($P(mod)$) dominates the annual surface mass balance variability over Antarctica, which is confirmed by taking
the anomalies of each component with respect to the long-term mean and plotting them in Fig. 1.4b. Precipitation (solid black with diamonds; $\sigma=8.5$ mm y$^{-1}$) explains nearly all of the variability of the surface mass balance (small dashed with triangles; $\sigma=8.4$ mm y$^{-1}$). Thus, considering mainly $P$, or $P+SU_{\text{sf}}$ (sometimes denoted $P-E$), is justified when assessing the variability and trends of the Antarctic surface mass balance at the continental scale, consistent with earlier assessments (Genthon 2004). Therefore, in the following chapters blowing snow processes are not considered when investigating the variability of the surface mass balance. However, blowing snow processes are implicitly considered in the Antarctic snowfall reconstruction in Chapter 3 (the methodology is described in the Appendix), as the modeled precipitation field is bias-adjusted using long-term ice core accumulation records in order to provide the most realistic background field possible.

1.4 Summary

In this chapter the motivation and background for the dissertation has been described, including a description of the components of the Antarctic surface mass balance. Chapters 2, 3, and 4, described in section 1.1, comprise the main body of work. Each chapter is self-contained, having been written for publication in refereed journals, and having a detailed literature review that exceeds that presented in this chapter. A summary of the results and recommendations for future research are given in Chapter 5. An appendix outlining the methods employed for the snowfall reconstruction in Chapter 3 is presented in Chapter 6. Chapter 7 contains a bibliography.
At the time of this writing Chapters 2 and 3 have been published (Monaghan et al. 2006a, 2006b), and Chapter 4 is in preparation for publication. By publishing each chapter in its entirety, this work will serve those who compile estimates of global sea level change, those who model Antarctic ice sheets and ice shelves, and those who endeavor to understand Antarctic climate in a changing global climate system.
### 1.5 Tables and Figures

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Table 1.1. Summary of surface mass balance trends based on estimates from various studies, including the hindcast work completed in Chapter 2 (“This Study”). Uncertainties about the trends in this study are based on the 90% confidence interval. Bromwich et al. (2004a) uncertainties were based on the 95% confidence interval, and those from van de Berg et al. (2005) were computed as one standard deviation about the trend. The contribution to global sea level is calculated by multiplying the trend by the ratio of the grounded ice sheet surface area (1.19 x 10⁷ km²) to that of the global ocean (36.13 x 10⁷ km²). The calculation for percentage of current GSL rise is based on a GSL increase of 3 mm yr⁻¹ (Hansen et al. 2005). Note that the contribution to global sea level rise becomes increasingly important each year, as the trend indicated is the additional amount of water each year compared to the previous year. *Grounded ice sheet only.
Fig. 1.1. Components of the surface mass balance (mm y\(^{-1}\) weq) averaged for 1985-2001: a) Polar MM5 \(S_{\text{fsc}}\); b) calculated \(S_{\text{fsc}}+S_{\text{blsn}}\); c) b-a (a proxy of the contribution of \(S_{\text{blsn}}\)); and d) calculated \(D_{\text{blsn}}*-1\) (convergence is positive).
Fig. 1.2. Components of the surface mass balance (averaged for 1985-2001) as a fraction of Polar MM5 precipitation (P): a) Polar MM5 SU_sfc/P; b) Calculated SU_tot/P; c) Calculated D_blkn/P; and d) Calculated (SU_tot+D_blkn)/P.
Fig. 1.3. Estimates of total surface mass balance (mm y\(^{-1}\) weq) averaged for 1985-2001: a) Polar MM5 P; b) Polar MM5 P plus calculated SU\(_{\text{sf}}\); c) Polar MM5 P plus calculated SU\(_{\text{tot}}\); and d) Polar MM5 P plus calculated SU\(_{\text{tot}}\)+D\(_{\text{blsn}}\). D\(_{\text{blsn}}\) is multiplied by negative one so that positive values represent snow convergence.
Fig. 1.4. a) 1985-2001 Antarctic surface mass balance components, expressed in absolute terms. b) The same components expressed as anomalies from the long-term mean. Units are mm yr$^{-1}$ weq.
CHAPTER 2

RECENT TRENDS IN ANTARCTIC SNOW ACCUMULATION FROM POLAR MM5 SIMULATIONS

2.1 Introduction

2.1.1 General spatial and temporal characteristics of Antarctic snowfall

The accumulation term is the primary mass input to the Antarctic ice sheets, and is the net result of precipitation, sublimation/vapor deposition, drifting snow processes, and melt. Precipitation, which primarily occurs as snowfall, is dominant among these components (Bromwich 1988) and establishing its spatial and temporal variability is necessary to assess ice sheet surface mass balance. Comprehensive studies of snowfall characteristics over Antarctica are given by Bromwich (1988), Turner et al. (1999), Genthon and Krinner (2001), van Lipzig et al. (2002) and Bromwich et al. (2004a). Snowfall is influenced to first order by the Antarctic topography. Most of the snowfall occurs along the steep coastal margins and is caused by orographic lifting of relatively warm, moist air associated with the many transient, synoptic-scale cyclones that encircle the continent. The influence of synoptic activity decreases inward from the coast, and over the highest, coldest reaches of the continent the primary mode of snowfall is due to cooling of moist air just above the surface-based temperature inversion. This extremely cold air has little capacity to hold moisture, and thus the interior of the East Antarctic Ice
Sheet is a polar desert, with a large area that receives less than 5 cm water equivalent each year (e.g., Vaughan et al. 1999, Giovinetto and Zwally 2000).

2.1.2 Long-term Antarctic accumulation estimates

In recent decades, estimates of snowfall and snowfall accumulation (surface mass balance) over the Antarctic ice sheets have been made by three techniques: surface-based observations, remote sensing, and atmospheric modeling. Constructing a reliable data set of measurements over Antarctica for a long time period from these methods has been difficult for numerous reasons. These include a sparse surface observational network (e.g., Giovinetto and Bentley 1985), difficulties distinguishing between clouds and the Antarctic ice surface in satellite radiances (Xie and Arkin 1998), and incomplete parameterizations of polar cloud microphysics and precipitation in atmospheric models (Guo et al. 2003), as well as errors in the large-scale circulation (e.g., Bromwich and Fogt 2004). Considering the limitations of these techniques, it is not surprising that the long-term-averaged continent-wide maps of snow accumulation over Antarctica yield a broad spectrum of results. The long-term estimates from several studies range from 119 mm yr\(^{-1}\) (van de Berg et al. 2005) to 197 mm yr\(^{-1}\) (Ohmura et al. 1996) for the grounded ice sheet (estimates for the conterminous ice sheet, which includes the ice shelves, are generally ~10% higher). In general, the studies employing glaciological data are considered the most reliable; the study of Vaughan et al. (1999) represents the current best approximation of 149 mm yr\(^{-1}\) for the grounded ice sheet, although a recent study (van de Berg et al. 2006) shows evidence that the Vaughan et al. (1999) dataset may
underestimate coastal accumulation. It is anticipated that advancements in all three techniques will reduce the uncertainties in the future.

2.1.3 Recent trends in Antarctic snowfall

On average, about 6 mm global sea level equivalent falls as snow on Antarctica each year (Budd and Simmonds 1991). Thus, it is important to assess trends in Antarctic snowfall, as even small changes can have considerable impacts on the global sea level budget. Vaughan (2005) notes that the greatest uncertainty in future predictions of sea level rise lies in the contribution of the Antarctic ice sheet, and considering the disagreement between various estimates of long-term accumulation discussed above, the ability to resolve recent trends has been limited. Bromwich et al. (2004a) inferred an upward trend of snowfall of about 1.5 mm yr$^{-2}$ (water equivalent) occurred from 1979-1999 from large-scale global models. This suggests that the trend in Antarctic snowfall removes an additional $\sim$0.05 mm yr$^{-1}$ from the global ocean. This is an important effect considering the current rate of sea level rise of $+2.8 +/- 0.4$ mm yr$^{-1}$ (Leuliette et al. 2004). Davis et al. (2005) suggest a much larger negative contribution to global sea level rise based on observations of thickening over East Antarctica from satellite altimetry. They suggest that most of this thickening is due to increases in snowfall, and that taken collectively, snowfall over East Antarctica to $\sim$81$^\circ$S mitigates sea level rise by about 0.12 mm yr$^{-1}$ between 1992-2003. This result is surprising considering it suggests a significant increase of snowfall over East Antarctica during one decade. There is little evidence of this magnitude of change in the observational records from ice cores and snow stakes examined in this study. The results of van de Berg et al. (2005) for a similar period,
1980-2001, are in contrast. RACMO2/ANT, an atmospheric mesoscale model, shows a statistically insignificant upward trend about an order of magnitude less than Bromwich et al. (2004a). This may be partly because the initial and boundary conditions are provided by the recently-completed European Centre for Medium-Range Weather Forecasts 40-year Reanalysis (ERA-40), which for the same period shows a downward trend of -0.42 mm yr$^{-2}$. It is difficult to assess model snowfall trends prior to ~1979 due to artificial trends or jumps in model fields related to the change in the volume and type of data available at the onset of the modern satellite era (e.g., Hines et al. 2000, van de Berg et al. 2005).

2.1.4 Description of the current study

This study employs Polar MM5 to simulate Antarctic accumulation in recent decades. Polar MM5 is a version of the Pennsylvania State University/NCAR fifth generation mesoscale model (MM5; Grell et al. 1994) optimized for the environment of polar ice sheets by the Polar Meteorology Group of the Byrd Polar Research Center at Ohio State University (Bromwich et al. 2001, Cassano et al. 2001). The two sets of simulations, each with different initial and boundary conditions, are evaluated for the 17-y period spanning 1985-2001. The initial and boundary conditions for the two sets of runs are provided respectively by (1) ERA-40, and (2) the National Centers for Environmental Prediction – Department of Energy Atmospheric Model Intercomparison
Project Reanalysis II (NCEP-II). The two-model approach is used so that uncertainty can be assessed by comparing the two resulting data sets. If the results from the two sets of simulations are similar, this increases confidence that the trends are robust in regions where long-term observational data are absent.

One motivation for this work is that most existing long-term global reanalyses have been performed with models tuned to provide optimal results for lower-latitudes. Therefore, they have traditionally had problems that limit their use for assessing climate trends at high-latitudes. A thorough review of the problems associated with the various reanalysis data sets is given by Bromwich and Fogt (2004). Here results are optimized by employing a model that has proven good skill over the Antarctic (Guo et al. 2003), and by using the most up-to-date topography at a horizontal resolution that exceeds the reanalyses (60 km in Polar MM5 versus ~125 km in the highest-resolution reanalysis available). Due to the computational expense, only recently have mesoscale models been used to explore Antarctic snowfall variability at decadal time scales (van Lipzig et al. 2002, van de Berg et al. 2005). This is the first time Polar MM5 has been used to do so.

In Section 2, the model formulation and related data sets are described. In Section 3, the simulated mean accumulation and trends are compared to observationally-based data sets to assess their quality. In Section 4, the spatial distribution of the temporal trends from 1985-2001 is evaluated. In Section 5, the mean snowfall and surface mass balance trends averaged over the whole of Antarctica from the simulations and several other data sets are compared. Conclusions are drawn in Section 6.
2.2 Polar MM5 Configuration

A full description of the standard MM5 modeling system is given by Grell et al. (1994). Bromwich et al. (2001) and Cassano et al. (2001) give a detailed description of the important changes to MM5 to optimize the model for use over ice sheets as Polar MM5. These include a modified parameterization for the prediction of ice cloud fraction, improved cloud-radiation interactions, an optimized stable boundary layer treatment, improved calculation of heat transfer through snow and ice surfaces, and the addition of a fractional sea ice surface type. Guo et al. (2003) evaluate Polar MM5 performance over Antarctica for a one-year period (1993) for a similar model configuration and show that the intraseasonal and interseasonal variability in pressure, temperature, wind, and moisture are well-resolved.

A 60-km polar stereographic domain with 121 grid points in the x and y directions covers Antarctica and the surrounding ocean (Fig. 2.1). The vertical formulation consists of 31 terrain-following half-sigma levels, with 11 levels in the lowest 1000 m to capture the complex interactions in the planetary boundary layer. The lowest half-sigma level is about 13 m above the surface at sea level. The model top is set at a pressure of 10 hPa with a rigid lid upper boundary. The model topography is interpolated from the 1-km resolution RADARSAT Antarctic Mapping Project Digital Elevation Model Version 2 (Liu et al., 2001), the so-called RAMP DEM. The regions spanned by the Ronne/Filchner Ice Shelf and Ross Ice Shelf are specified as land ice.

The Polar MM5 is initialized once-daily at 0000 UTC for each day from 1 January 1979 through 31 August 2002 (the ERA-40 dataset ends at this time). Each forecast
extends out to 42 hours. Hours 24 (00Z day+1), 30 (06Z, day+1), 36 (12Z, day+1), and
42 (18Z, day+1) of each of the ~8,600 forecasts are joined together to form a continuous
6-hour data set spanning a ~24-year period. The first 24 hours of the forecast are
discarded to allow for spin-up of the model hydrologic cycle.

As mentioned earlier, 2 sets of ~24-year simulations are performed, one with initial
and boundary conditions from ERA-40 (Uppala et al. 2005, henceforth, E40) and the
other with initial and boundary conditions from NCEP-II (Kanamitsu et al. 2002,
henceforth, NN2). A description of each of these data sets with respect to its skill over
Antarctica is given below. The resolution of the data provided to Polar MM5 from both
E40 and NN2 is 2.5° latitude x 2.5° longitude. The variables provided for the initial and
lateral boundary conditions are 3-dimensional temperature, pressure, meridional and
zonal winds, vertical winds, and specific humidity. Sea ice fraction and land surface/sea
surface temperatures are provided at the lower model boundary. The boundary
conditions are updated at 6-hourly intervals throughout the forecast. Daily polar gridded
sea ice concentration data with 25-km horizontal resolution derived from satellite passive
microwave data and obtained from the National Snow and Ice Data Center are used to
identify the sea ice surface type and its fractional coverage at each model grid point in
both sets of runs. Sea surface temperatures in both sets of runs are provided by the
1.125° latitude by 1.125° longitude sea surface temperature data set used in E40 (Fiorino
2004). This data set uses 1) the monthly mean HadISST data set from the United
Kingdom Meteorological Office Hadley Center for 1956-1981; and 2) the weekly NCEP
2DVAR data for 1982-present, and is thought to be of better quality than previous
datasets (Fiorino 2004).
Bromwich and Fogt (2004) evaluated mean sea level pressure (MSLP), 2-m temperature, and 500 hPa geopotential height in E40 for 1958-2001. They found that the assimilation system is strongly constrained by satellite data, and thus performs relatively poorly over Antarctica until the modern satellite era, about 1979, and performs with good skill thereafter. Van de Berg et al. (2005) find similar results for E40 snowfall – their results indicate a large jump in mean annual snowfall of ~30 mm in about 1979, with the magnitude thereafter matching other estimates more closely.

NN2 was intended to be an improvement over the NCEP/National Center for Atmospheric Research (NCAR) Reanalysis (Kalnay et al. 1996). Over Antarctica this meant correcting a problem with the assimilation of the Australian Surface Pressure Bogus Data for the Southern Hemisphere (PAOBS) from 1979-1982 (Kistler et al. 2001), and using improved sea ice and sea-surface temperature fields (Kanamitsu et al. 2002). The NCEP/NCAR Reanalysis was subject to spurious trends in MSLP and geopotential height into the 1990s in the Antarctic coastal regions due to the increase in amount of observations assimilated into the system with time (Hines et al. 2000, Marshall and Harangozo 2000). Although Guo et al. (2004) note a 3-4 hPa drop in MSLP and a corresponding 30-40 m drop in 500-hPa geopotential height at Leningradskaya in NN2 during 1988 that may be related to a station height error, overall the spurious trends are reduced in NN2 in the period of overlap with the NCEP/NCAR Reanalysis after 1979 (Hines et al. 2000; Fig. 5). Hines et al. (2000) also show the NCEP/NCAR Reanalysis captures the interannual variability of MSLP with good skill after 1979, and that the skill of NN2 is similar.
The 1985-2001 period is chosen to evaluate snowfall trends, rather than the entire 1979-2001 period, because agreement between the PMM5_E40 and PMM5_NN2 snowfall for 1979-2001 is not as good as for other climatological variables. The two sets of simulations appear to be especially sensitive to moisture fluxes from E40 and NN2 through the lateral boundaries of Polar MM5 between about 45°S and 57°S (Fig. 2.1). There are differences in the trends of the poleward moisture fluxes at these latitudes that appear to be largest before ~1985. The reason for this may be related to adjustments in the ERA-40 assimilation system after the onset of the modern satellite era and prior to the mid-1980s. There is evidence of such issues in ERA-40 prior to the mid-1980s. Turner et al. (2005a) note that good agreement between the number of precipitation days observed at Faraday/Vernadsky station on the Antarctic Peninsula compared to the number of precipitation days from ERA-40 does not occur until 1984 onwards. They attribute this to a lack of humidity data being assimilated into the analysis system in ERA-40. Simmons et al. (2004) found that the near-surface air temperature over the Southern Hemisphere in ERA-40 did not coincide well with that from an observationally-based dataset until ~1987 (however, for Antarctica, agreement was good from 1979 onward). Presumably, much of the discrepancy was due to the data-sparse regions in the Southern Ocean, where the Polar MM5 boundaries are located.

In summary, the E40 and NN2 datasets appear to have adequate representations of climatological fields (near-surface temperature, MSLP, and 500-hPa heights) over Antarctica during the January 1979-August 2002 period. However, the hydrologic cycle in the two data sets does not converge until the mid-1980s; thus the period 1985-2001 is chosen to evaluate snowfall trends. It is noteworthy that E40 and NN2 are both based on
similar observational data sources. Therefore, the main differences between the two are
due to differences in data assimilation techniques and model formulation and it must be
considered when comparing the PMM5_E40 and PMM5_NN2 simulations that they are
not entirely independent.

2.3 Comparison of simulated and observed snowfall accumulation

A comparison of the Polar MM5 E40 (PMM5_E40), Polar MM5 NN2 (PMM5_NN2),
E40, NN2, and the Japanese 25-year Reanalysis (JRA) snowfall-minus-sublimation (P-E)
with observed accumulation at several ice core and snow stake sites around Antarctica is
shown in Fig. 2.2. The PMM5_E40 and PMM5_NN2 P-E have been interpolated from
the four nearest grid points to the observation location from their nominal 60-km x 60-km
polar stereographic grids. The E40, NN2, and JRA P-E have been re-gridded to 1°
latitude x 1° longitude grids from their nominal output resolution, ~125 km for E40 and
JRA, and ~200 km for NN2, and then similarly interpolated to the location of the
observation. The JRA data are included in this analysis because it has been released only
recently and this is the first time its snowfall can be evaluated over Antarctica. We will
use it later in the paper to provide an additional assessment of continent-wide trends in
snowfall.

A considerable amount of noise can be present in ice cores due to small scale
perturbations caused by the interaction between the local topography and the wind, which
can have a substantial effect on annual accumulation at a given site and mask the “true”
accumulation signal for the region as a whole (Frezzotti et al. 2005, Genthon et al. 2005).
Frezzotti et al. (2005) note that the local spatial variability of ice core accumulation can
be an order of magnitude higher than the temporal variability on decadal time scales, which they attribute mainly to wind-driven processes. Thus, noise is reduced where possible by averaging cores together if they are in the same region, and also by averaging the first and last five years of the records for the comparison. The locations of the accumulation regions from Fig. 2.2 are shown in Fig. 2.1. Citations, abbreviations and temporal coverage of the observations are given in Table 2.1. The accumulation signal at the snow stake networks at Vostok and South Pole is less noisy, as the spatial variability is already removed by averaging many stakes together. At Law Dome and Siple Dome, data are available for only one core and thus regional averaging cannot be used. However, the relatively high correlation between the simulated P-E and observed accumulation at these sites suggest that their locations at the tops of ice domes may reduce the amount of small-scale perturbation in the cores, and thus they are included in the analysis. Overall, the correlation coefficients between the simulated and observed records (Fig 2.2c) indicate that the modeled and observed annual accumulation variability share common variance different from zero at the 90% confidence interval at many of the sites (especially in West Antarctica), despite the small-scale perturbations in the observational records. Though these relationships are often weak, they are as good as can be expected considering the differences between the simulated and observed accumulation signals, and thus the following comparison is considered valid. The 90% confidence interval (obtained with a 2-tailed Student’s t-test) is used to test statistical significance throughout the remainder of this paper. This value has been chosen, rather than the 95% or 99% interval, to permit substantial changes to be more easily recognized over the relatively short time intervals evaluated.
The mean annual P-E in the western and eastern Lambert basin in all simulations is lower than observed. The NN2 accumulation is particularly low; this is due to unrealistically large sublimation rate (Hines et al. 1999) in the model that occurs over most of the coastal and interior locations in East Antarctica, but is not an issue in West Antarctica. Unrealistically large sublimation is also present in the JRA record in ELAM, causing accumulation of ~0 mm y⁻¹ and a negative correlation coefficient. PMM5_E40 and E40 provide the most realistic estimates of mean accumulation in both basins and have the highest correlation coefficients, (with the exception of JRA in WLAM). All of the models capture the observed negative change in accumulation at these two sites. At Law Dome, the P-E in all of the models correlates well with the observed accumulation. Mean annual accumulation and temporal changes cannot be shown for Law Dome as the data have yet to be published (personal communication, Tas van Ommen). However, it is noteworthy that only PMM5_E40 and E40 capture the correct sign of the changes there. In Dronning Maud Land all of the models, with the exception of NN2, simulate the mean P-E well. PMM5_E40, E40, and JRA capture the temporal change, but PMM5_NN2 and NN2 do not. PMM5_E40 simulates the mean P-E at Vostok most closely, and is also closest to the near zero temporal change observed there. PMM5_NN2, NN2, and JRA predict relatively large positive changes (as a percentage of mean accumulation) that do not appear in the Vostok record. The correlation coefficients indicate that PMM5_E40 and E40 capture interannual variability at Vostok with the most skill; the negative correlation coefficient in the NN2 P-E is due to the overestimation of the sublimation fluxes. At South Pole, the models underestimate P-E, with the exception of JRA, which overestimates accumulation over the entire interior of the ice sheet. All of the models
except NN2 capture the negative temporal change at South Pole. The correlation coefficients indicate that PMM5_E40 demonstrates some skill in simulating interannual variability at South Pole, while the other data sets show almost no skill.

At the four remaining sites, all in West Antarctica, the simulated P-E is similar to the observed accumulation for all of the models. The temporal changes at all four sites are captured in PMM5_NN2 and NN2, and at three-out-of-four sites in PMM5_E40 and E40. The JRA model captures the temporal change at one of the four sites. Near Pine Island Bay (WA01) where large changes in accumulation (Kaspari et al. 2004) and ice volume (Rignot and Thomas 2002, Thomas et al. 2004) have been measured recently, PMM5_E40, E40, and JRA fail to capture temporal change despite resolving the mean accumulation and interannual variability reasonably well.

Overall, the PMM5_E40 and E40 simulated P-E capture the mean accumulation, temporal changes, and interannual variability over the continent as a whole with the most skill. PMM5_E40 predicts higher P-E than E40 at nearly every site, and in most cases this is closer to the observed accumulation. Both predict the temporal changes in nearly every instance, with the exception of the WA01 region. In general, it cannot be said that the hydrologic cycle in PMM5_E40, with its higher resolution and polar physics, substantially improves on E40, the model that provides its initial and boundary conditions. The PMM5_NN2 and NN2 tend to underestimate P-E over East Antarctica, and do not capture interannual variability as well as PMM5_E40 and E40. Over West Antarctica, the skill of PMM5_NN2 and NN2 is comparable to PMM5_E40 and E40. Unlike PMM5_E40 versus E40, the hydrologic cycle in PMM5_NN2 improves estimates of P-E compared to NN2, being closer to the observed accumulation at all sites with the
exception of RIDS. PMM5_NN2 also captures the magnitude of temporal changes better than NN2, and captures interannual variability better than NN2 at seven-out-of-ten sites. The JRA model shows promising skill in capturing variability and trends at some locations (WLAM and DML), but is inconsistent overall, incorrectly depicting the sign of the temporal changes in five of the ten regions. For the remainder of this section, and in the next section, the spatial variability and trends of accumulation are assessed using PMM5_E40 and PMM5_NN2. Based on the findings above, the use of these two simulations is a reasonable choice: PMM5_E40 and E40 capture the mean, temporal change, and interannual variability of Antarctic accumulation with similar skill, and thus it is only necessary to show one of the two; of the remaining runs, PMM5_NN2 shows the best overall skill at depicting variability and trends versus NN2 and JRA, yet it is different enough from PMM5_E40 to provide a useful comparison.

Figure 2.3 compares the Vaughan et al. (1999) accumulation compilation, which is a synthesis of accumulation observations taken over many years, with the P-E from the PMM5_E40 and PMM5_NN2 runs. The broad features of Antarctic accumulation are captured in both of the Polar MM5 runs, with maxima in the coastal areas, especially West Antarctica and the Antarctic Peninsula, and a large area within the continental interior with accumulation of less than 50 mm y\(^{-1}\). The interior accumulation in the PMM5_NN2 runs appears to be less than the Vaughan et al. (1999) or the PMM5_E40 data sets. This is confirmed by comparing Figs. 2.3d and 2.3e. Here it is seen that PMM5_NN2 is dryer than PMM5_E40 in the continental interior. Figures 2.3d and 2.3e also reveal a general characteristic of both data sets: they are dryer than the Vaughan et al. (1999) data set in the interior and wetter in the coastal regions. Bromwich et al.
(2004a) discuss the dry bias in the interior and relate this to the limited ability of Polar MM5 to represent clear sky precipitation, which Bromwich (1988) infers to form nearly continuously in the interior of Antarctica without organized synoptic processes. In Victoria Land, the dry bias in PMM5_E40 and PMM5_NN2 may be exaggerated, as ice core and snow stake evidence from recent studies indicates that the Vaughan et al. (1999) data set overestimates accumulation in this region on average by 33%, and by as much as 65% in some areas (Frezzotti et al. 2004, Magand et al. 2004). In the coastal regions, a recent study by van de Berg et al. (2006) compares the Vaughan et al. (1999) data set to several hundred observations and concludes that the compilation generally underestimates coastal snowfall. The comparison with coastal ice core records in the previous paragraph also corroborates this conclusion, as it is generally found that the PMM5 simulated accumulation is on the order of, or less than, the long-term ice core accumulation. In summary, the spatial distribution of the PMM5_E40 and PMM5_NN2 P-E matches that in the long-term accumulation compilation of Vaughan et al. (1999), and the magnitude of long-term annual mean P-E is accurate within the bounds of uncertainty.

In summary, the PMM5_E40 and PMM5_NN2 simulations largely depict the observed spatial and temporal variability of Antarctic snowfall, especially considering the small-scale perturbations that are not resolved in the models but likely exist in the observational records despite space/time filtering where possible. The PMM5_E40 record appears to be most reliable for resolving the mean accumulation, variability, and trends, particularly in East Antarctica. In the following section, the spatial distribution of the temporal trends of snowfall is examined using these two data sets. From henceforth,
snowfall (P), rather than P-E, is used. Snowfall is the dominant term in Antarctic surface mass balance over the broad space scales that are investigated here (Bromwich 1988, Genthon 2004, also see Chapter 1).

2.4 Spatial distribution of the temporal trends of snowfall for 1985-2001

Figure 2.4 shows the spatial distribution of the 1985-2001 linear trends for the PMM5_E40 and PMM5_NN2 annual snowfall. Figures 2.4a and 2.4b express the trends in mm yr$^{-2}$, and Figs. 2.4c and 2.4d express the same trends in percent-per-decade, which allows for the relative change to be assessed over high elevations where little snow falls. Areas of strong upward and downward trends are apparent. Statistically significant (at the 90% confidence interval) upward trends are occurring near the Antarctic Peninsula, in the ocean regions north of the coastline at $\sim$0° E and between 105°-150° E, and in the Transantarctic Mountains in Victoria Land. Statistically significant downward trends are occurring just inland of the Ronne-Filchner Ice Shelf. Strong downward trends are also indicated in coastal regions of East Antarctica in both simulations, although these are, for the most part, not statistically significant. Overall, there is broad agreement between the two simulations, but the regions of positive change in the PMM5_NN2 simulations are larger. A broad region of 20-50% change-per-decade over East Antarctica is indicated in the PMM5_NN2 runs (Fig. 2.4d), but considering that this model tended to overestimate positive snowfall changes at Vostok (Fig. 2.2), the smaller PMM5_E40 trends (Fig. 2.4c) are likely to be more representative of snowfall changes over East Antarctica.

Figures 2.5 and 2.6 are similar to Fig. 2.4a and 2.4b, but for the four seasons: summer (DJF, Fig. 2.5a,b), autumn (MAM, Fig. 2.5c,d), winter (JJA, Fig. 2.6a,b), and spring
(SON, Fig. 2.6c,d). In DJF, both data sets show positive snowfall trends over the Antarctic Peninsula, the Ronne-Filchner Ice Shelf, and much of the sector between 0° E and 90° W. Both data sets also indicate strong and often significant trends over the region of East Antarctica near Vostok, and in ocean areas adjacent to coastal East Antarctica. In MAM, the spatial patterns of the trends are similar to those in DJF, except in the Weddell Sea where there is a larger area of negative trends, and over coastal West Antarctica where the trends become positive. The positive trends in MAM over the Antarctic Peninsula are stronger than in DJF. In JJA, the snowfall trends are strongly negative in the Bellingshausen Sea and over the Antarctic Peninsula, although there is a region of positive trends on the east side of the Peninsula. Broad, statistically significant negative trends are also present near the coast at 45° E, and in the ocean regions of the South Pacific between 135° W and 135° E. An area of strong positive change is located along and inland of the coast near 0° E. The spatial distribution of the trends in SON is similar to JJA, except in the Weddell Sea, where the trends become positive in SON. Over the continent as a whole, the trends are more positive in SON than in JJA. Regions of consistent positive change in all four seasons are apparent over much of the interior of East Antarctica between 90° E and 180° E, and in the ocean areas north of 0° E and 120° E. There are few areas where negative trends are present in all four seasons.

Figure 2.7 summarizes the annual and seasonal trends shown in Figs. 2.4-2.6 by averaging them into broad regions. All of the averages are taken over land/ice shelf grid points only, except for the “ALL” category, which includes the entire domain (ocean + land grid points). We show the snowfall trends over the grounded ice sheet (GIS) in Fig. 2.7, but reserve a detailed discussion of this important region for the next section. The
trends over the Ronne-Filchner Ice Shelf and the Antarctic Peninsula are positive in summer and autumn, and negative in winter and spring. There are no significant trends over the Ross Ice Shelf or West Antarctica annually or in any season, and a significant positive trend is present in East Antarctica only in spring in the PMM5_NN2 snowfall. Significant positive trends are present over the entire domain in summer and autumn. It is interesting that the trends over the Peninsula, the Ronne-Fichner Ice Shelf, and the entire domain closely follow trends in the Southern Annular Mode (SAM, Marshall 2003). The SAM can be thought of as the measure of the strength of the pressure gradient between the mid-latitudes and high-latitudes, from about $40^\circ$ S to $65^\circ$ S (e.g., Thompson and Wallace 2000, Marshall 2003, Fogt and Bromwich accepted). The upward snowfall trends in the summer and autumn coincide with a strengthening of the SAM in these seasons, and the downward trends in winter coincide with a weakening of the SAM in this season (although this weakening is not present in all proxies of the SAM). Turner et al. (2005a) also infer a strong increase in snowfall over the Antarctic Peninsula since 1950 from the record of observed precipitation days at Faraday/Vernadsky ($65.25^\circ$ S, $64.27^\circ$ W), especially in summer and autumn, and link this to the strengthening of the SAM during DJF and MAM. Although there are smaller downward trends over the Peninsula and Ronne-Filchner Ice Shelf in spring, there is little trend in the SAM during this season, as it is thought that the SAM is damped by the El Niño-Southern Oscillation (ENSO) in spring (Fogt and Bromwich accepted).

In summary, the spatial distribution of the annual and seasonal snowfall trends in PMM5_E40 and PMM5_NN2 are similar despite the differences in mean accumulation and local trends discussed in Section 3. The annual trends tend to be a small residual of
seasonal trends that exhibit a large degree of interseasonal variability. Marshall et al. (2004) demonstrated that the trends in the SAM for the annual mean and austral summer are unlikely to be due to internal climate variability alone, and that anthropogenic forcing also may play a role. Considering the relationship between seasonal snowfall trends and the SAM noted here, this suggests that some of the regional trends may be related to human-induced climate changes.

2.5 Snowfall trends over all of Antarctica for 1985-2001

Figure 2.8 shows the 1985-2001 annual and seasonal Antarctic-wide snowfall (mm y$^{-1}$) averaged over the grounded ice sheet for PMM5_E40, PMM5_NN2, E40, NN2, JRA, and the composite ("COMP") of PMM5_E40, PMM5_NN2, E40, and JRA. NN2 is not included in the composite because of its relatively poor ability to capture temporal changes over much of the interior of the ice sheet (Fig. 2.2, DML, VOS, SPOL), and its tendency to substantially overestimate changes at WLAM, ELAM, WA01, and RIDS. To make the comparison uniform, all of the data sets have been interpolated to a 1° latitude x 1° longitude grid. Grounded ice sheet grid points are defined as falling within the land mask of Vaughan et al. (1999) and being above the 300 m elevation contour of the Polar MM5 topography interpolated to the same grid, which largely eliminates ice shelf grid points (the 300 m contour of the Polar MM5 topography, which is based on the RAMP DEM, is shown in Fig. 2.1). Using this technique, the surface area of the grounded ice sheet is 11,960,000 km$^2$, nearly identical to the 11,966,000 km$^2$ area calculated by Vaughan et al. (1999). This technique also includes the northern area of the Antarctic
Peninsula that is not part of the grounded ice sheet, as snow falling on this region will affect the global sea level budget.

Both the PMM5_E40 and PMM5_NN2 runs predict higher snowfall than the respective reanalyses used to drive them, E40 and NN2; this is at least partly due to higher horizontal resolution in Polar MM5, especially along the steep coastal margins where the majority of snow falls. Overall, the annual and seasonal snowfall variability among the data sets is in relatively good agreement. The amount of JRA snowfall is very similar to that in PMM5_E40, although this relationship appears to be a result of spatial averaging. JRA overestimates snowfall on the high plateau and underestimates it in coastal regions when compared to PMM5_E40.

Table 2.2 summarizes the annual and seasonal snowfall means and trends for the grounded ice sheet for the time series shown in Fig. 2.8. The percent difference between the first five years and last five years of the record is also presented as an alternate means of assessing snowfall change for this relatively short time period. The change expressed in this manner is in good agreement with the 1985-2001 linear trends. The mean annual snowfall ranges between 158 mm y⁻¹ (E40) and 200 mm y⁻¹ (PMM5_E40). A recent study by van de Berg et al. (2006) suggests that the best estimate of accumulation over the grounded ice sheet is about 171 mm y⁻¹, substantially more than the estimate of 149 mm y⁻¹ by Vaughan et al. (1999). Considering surface sublimation is roughly 10% of snowfall (Bromwich et al. 2004a), the P-E in Polar MM5 would be close to this accumulation estimate, while that in E40 would underestimate it. The maximum seasonal snowfall occurs in autumn and winter in all of the runs except NN2. Bromwich (1988)
noted that the maximum snowfall over the continent occurs in the winter months, and thus it is likely that the seasonal cycle in NN2 is inaccurate.

The PMM5_NN2 trends are more positive than the PMM5_E40 trends in all cases. Likewise, the NN2 trends are always more positive than the E40 trends. The largest positive trends are found in NN2 in all seasons except autumn, in which PMM5_NN2 has the largest trend. The PMM5_E40 trends are more positive than the E40 trends in all cases except MAM, and the PMM5_NN2 trends are smaller than the NN2 trends in all cases. This makes the annual snowfall trends closer to zero in the PMM5_E40 and PMM5_NN2 datasets with respect to the E40 and NN2 datasets, even though the mean annual snowfall in PMM5_E40 and PMM5_NN2 is much larger than that in the respective “parent” dataset of each, E40 and NN2. The JRA trends are similar to those in PMM5_E40 and E40, suggesting that the interannual variability and trends in the poleward moisture fluxes in JRA are similar to those in E40.

Collectively considered, all of the data sets predict positive trends in DJF and MAM, but there is less agreement for JJA and SON. Few trends are significantly different from zero at the 90% confidence interval. Only NN2 shows a significant positive trend for the annual snowfall, but considering the issues with this data set noted earlier, this trend is questionable. The composite trend suggests that the annual snowfall trend over the grounded ice sheet is nearly zero.

Table 2.3 summarizes the annual snowfall and accumulation trends discussed here, as well as results from different studies for various time periods, and gives their contribution to global sea-level. The estimates from the data sets in this study are presented in terms of P-E so that they most closely reflect their sea-level contribution (surface meltwater runoff
is neglected as most meltwater refreezes near where it is produced and has been shown to have little effect on surface mass balance (Liston and Winther 2005)). The number from Bromwich and Robasky (1993) is shown, although it covers a much earlier time period (through 1975), as a means of perspective. Their analysis was based on a sparse network of accumulation observations with numerous gaps, and may substantially overestimate the snowfall increase from 1955 through 1975. However, a large change in sea ice extent was noted between these two periods (de la Mare 1997), and thus such a large, continent-wide change may have been possible.

Contemporary estimates for the period after 1979 range from +1.67 mm y^{-2} (NN2, 1979-1999) to -0.50 mm y^{-2} (E40, 1980-2001). Seven of the estimates predict positive trends in snowfall/surface mass balance, while 5 predict negative trends. Of the seven that predict positive trends, two of the estimates from Bromwich et al. (2004a) are questionable, as they are based on the E15_ECT dataset, which appears to have an artificial jump in snowfall (and other fields) after 1994 that causes spurious trends. Four of the five data sets that produce negative trends are either E40 or based on E40, which have been shown above to capture variability and trends over the continent with better skill than PMM5_NN2 and NN2 when compared to ice core and snow stake records. It is noteworthy that there is a considerable difference between the mean P-E from E40 calculated by van de Berg et al. (2005) for the 1980-2001 period (119 mm y^{-1}) and that calculated in this study (135 mm y^{-1}). This appears to be largely due to the inclusion of the northern tip of the Antarctic Peninsula in this study. There is also a discrepancy between the long-term annual NN2 snowfall from Bromwich et al. (2004a) versus the NN2 P-E from this study (188 mm y^{-1} versus 84 mm y^{-1}). Although these averages are
calculated for slightly different surface areas, the difference is primarily due to the inclusion of sublimation in this approximation and infers that sublimation is about 100 mm yr\(^{-1}\) over the grounded ice sheet, about an order of magnitude too large; this problem has been noted in previous literature for the similar NCEP/NCAR Reanalysis (Hines et al. 1999). For the three high spatial resolution mesoscale atmospheric models used (PMM5_E40, PMM5_NN2, and RACMO2/ANT), recent accumulation trends are not statistically different from zero.

Overall, the results from this study suggest that, when averaged over the grounded Antarctic ice sheet, recent accumulation trends may not be significantly different from zero, and thus the role of Antarctic surface mass balance in mitigating recent sea level rise may be minor.

### 2.6 Conclusions

This study employs Polar MM5, a model optimized for use in polar regions, for the first time to evaluate recent snowfall trends over Antarctica. Two sets of simulations, each driven by a different reanalysis, yield similar spatial patterns of annual and seasonal snowfall trends over Antarctica. By comparing these two datasets with each other and with observations, the reanalyses used to drive them, and an independent reanalysis (JRA), the uncertainty can be estimated. The similarity between the trends for the two runs, as well as their agreement with observations, lends confidence that the results are robust. The PMM5_E40 hydrologic cycle is similar to that of its “parent” reanalysis, E40, although it predicts about 40 mm yr\(^{-1}\) more annual snowfall over the grounded ice sheet. The PMM5_NN2 hydrologic cycle appears to be an improvement over its “parent”
reanalysis, NN2, generally capturing mean snowfall, temporal changes, and interannual variability with more skill compared to observations from ice cores and snow stakes. Both of the Polar MM5 runs simulate snowfall trends for 1985-2001 that are closer to zero than their respective reanalyses.

The results presented indicate that there is a substantial amount of spatial and temporal variability in the Antarctic surface mass balance. The annual variability and trends over Antarctica are the small residual of larger seasonal variability and trends that appear to be related to recent climate change, particularly seasonally varying trends in the Southern Annular Mode. This complex topic will be addressed in a follow-on publication. Though considerable changes are happening regionally, averaged over Antarctica, the annual trend in Antarctic snowfall, which is important for sea level change assessments, appears to be close to zero. Although this study is over a different time period and employs different techniques, the model and observational results presented here do not support the significant upward accumulation trend suggested by Davis et al. (2005) for the 1992-2003 period over East Antarctica.
2.7 Tables and Figures

<table>
<thead>
<tr>
<th>Source of Accumulation Observation</th>
<th>Abbreviated Name in Fig. 2</th>
<th>Approximate Location (see Fig. 1)</th>
<th>Length of Record</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average of eastern Lambert Basin cores DT085 (Xiao et al. 2001) and DT001 (Wen et al. (2001)</td>
<td>ELAM</td>
<td>~72° S, ~77.5° E</td>
<td>1985-1996</td>
</tr>
<tr>
<td>Law Dome core DSS97 used for 1985-1996 and core DSS0405 used for 1997-2001 (unpublished from Tas van Ommen)</td>
<td>LDOM</td>
<td>66.8° S, 112.8° E</td>
<td>1985-2001</td>
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<tr>
<td>Average of 13 firn cores from Dronning Maud Land (Oerter et al. 2000)</td>
<td>DML</td>
<td>~75° S, ~0° E</td>
<td>1985-1996</td>
</tr>
<tr>
<td>Vostok stake network (Ekaykin et al. 2005)</td>
<td>VOS</td>
<td>78.5° S, 106.9° E</td>
<td>1985-2001</td>
</tr>
<tr>
<td>South Pole Core used for 1985-1989 (Meyerson et al. 2003) and stake network for 1995-1999 (Mosley-Thompson et al. 1999)</td>
<td>SPOL</td>
<td>90.0° S, 0.0° E</td>
<td>1985-1999</td>
</tr>
<tr>
<td>Siple Dome core (Nereson et al. 1996)</td>
<td>SDOM</td>
<td>81.7° S, 149.0° E</td>
<td>1985-1994</td>
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<td>Average from West Antarctic firn cores 00-1, 00-4, and 00-5 (Kaspari et al. 2004)</td>
<td>WA00</td>
<td>~79° S, ~115° W</td>
<td>1985-2000</td>
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<td>Average from West Antarctic firn cores 01-2, 01-3, 01-5, and 01-6 (Kaspari et al. 2004)</td>
<td>WA01</td>
<td>~78° S, ~95° W</td>
<td>1985-1999</td>
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<tr>
<td>Average from West Antarctic firn cores RIDS-A, RIDS-B, and RIDS-C (Kaspari et al. 2004)</td>
<td>RIDS</td>
<td>~79° S, ~118° W</td>
<td>1985-1995</td>
</tr>
</tbody>
</table>

Table 2.1. Ice core and snow stake records used in this study and shown in Fig. 2.2.
### Table 2.2

Summary table of snowfall means and trends for the grounded ice sheet corresponding to the time series in Fig. 2.8. "COMP" is the composite of the PMM5_E40, PMM5_NN2, E40 and JRA time series. Bold trends are significantly different from zero at the 90% confidence interval.

<table>
<thead>
<tr>
<th>Category</th>
<th>Season</th>
<th>PMM5_E40</th>
<th>PMM5_NN2</th>
<th>E40</th>
<th>NN2</th>
<th>JRA</th>
<th>COMP</th>
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<tr>
<td>1985-2001 mean (mm y⁻¹)</td>
<td>ANNUAL</td>
<td>200</td>
<td>178</td>
<td>158</td>
<td>159</td>
<td>195</td>
<td>183</td>
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<tr>
<td></td>
<td>DJF</td>
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<td>43</td>
<td>39</td>
<td>43</td>
<td>44</td>
<td>43</td>
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<tr>
<td></td>
<td>MAM</td>
<td>54</td>
<td>49</td>
<td>41</td>
<td>41</td>
<td>54</td>
<td>49</td>
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<td></td>
<td>SON</td>
<td>47</td>
<td>42</td>
<td>36</td>
<td>36</td>
<td>45</td>
<td>43</td>
</tr>
</tbody>
</table>

| 1985-2001 trend (mm y⁻²) | ANNUAL | -0.16 | 0.49 | -0.21 | **1.20** | -0.29 | -0.04 |
|                         | DJF    | 0.02  | 0.13 | 0.01  | 0.30    | 0.03  | 0.05  |
|                         | MAM    | 0.20  | 0.40 | 0.25  | **0.51**| 0.25  | 0.28  |
|                         | JJA    | -0.30 | 0.00 | **-0.34**| 0.25    | **-0.37**| -0.25 |
|                         | SON    | -0.01 | 0.05 | -0.08 | 0.20    | -0.13 | -0.04 |

<p>| % difference of mean snowfall for 1997-2001 from 1985-1989 | ANNUAL | 0 | 4 | -1 | 10 | -1 | 1 |
|                                                           | DJF    | 0 | 2 | 1 | 8 | 1 | 1 |
|                                                           | MAM    | 6 | 13 | 9 | 17 | 7 | 9 |
|                                                           | JJA    | -7 | 0 | -11 | 7 | -9 | -7 |
|                                                           | SON    | 2 | 3 | -1 | 7 | -2 | 1 |</p>
<table>
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<tr>
<th>Study</th>
<th>Method</th>
<th>Variable</th>
<th>Mean +/- sd (mm yr(^{-1}) +/(-) mm yr(^{-1}))</th>
<th>Trend (mm yr(^{-2}))</th>
<th>Uncertainty (mm yr(^{-2}))</th>
<th>Contribution to GSL rise (mm yr(^{-1}))</th>
<th>% of current GSL rise (% yr(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bromwich and Robasky (1993)</td>
<td>Moisture budget study analyzing difference of accumulation between 1955-1965 and 1965-1975</td>
<td>P-E</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>-0.200</td>
<td>-7.1</td>
</tr>
<tr>
<td>Bromwich et al. (2004a)</td>
<td>Dynamic Retrieval from ERA15 reanalyses and ECMWF/TOGA operational analyses, 1979-1999</td>
<td>P</td>
<td>188 +/-10</td>
<td>1.65</td>
<td>0.8</td>
<td>-0.054</td>
<td>-1.9</td>
</tr>
<tr>
<td>Bromwich et al. (2004a)</td>
<td>Modeled from ERA15 reanalysis and ECMWF/TOGA operational analyses, 1979-1999</td>
<td>P</td>
<td>-180 +/-15</td>
<td>1.35</td>
<td>1.1</td>
<td>-0.044</td>
<td>-1.6</td>
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<tr>
<td>Bromwich et al. (2004a)</td>
<td>Modeled from NN2, 1979-1999</td>
<td>P</td>
<td>180 +/-12</td>
<td>1.67</td>
<td>0.59</td>
<td>-0.055</td>
<td>-2.0</td>
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<td>van de Berg et al. (2005)</td>
<td>Modeled from E40, 1980-2001*</td>
<td>P-E*</td>
<td>119</td>
<td>-0.50</td>
<td>0.22</td>
<td>0.016</td>
<td>0.6</td>
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<tr>
<td>van de Berg et al. (2005)</td>
<td>Modeled from RACMO2/ANT regional model, driven by E40 1980-2002*</td>
<td>P-E-M*</td>
<td>166</td>
<td>0.15</td>
<td>0.28</td>
<td>-0.005</td>
<td>-0.2</td>
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</table>

Table 2.3. Summary of snowfall and surface mass balance trends from various studies, including this one. Uncertainties about the trends in this study are based on the 90% confidence interval. Bromwich et al. (2004a) uncertainties were based on the 95% confidence interval, and those from van de Berg et al. (2005) were computed as one standard deviation about the trend. The contribution to global sea level is calculated by multiplying the trend by the ratio of the grounded ice sheet surface area (1.19 x 10\(^7\) km\(^2\)) to that of the global ocean (36.13 x 10\(^7\) km\(^2\)). The calculation for percentage of current GSL rise is based on a GSL increase of 2.8 mm yr\(^{-1}\) (Leuliette et al. 2004). Note that the contribution to global sea level rise becomes increasingly important each year, as the trend indicated is the additional amount of water each year compared to the previous year.
Table 2.3 Continued

<table>
<thead>
<tr>
<th></th>
<th>Modeled from Polar MM5, driven by E40, 1985-2001</th>
<th>P-E*</th>
<th>180 +/- 8</th>
<th>-0.13</th>
<th>0.70</th>
<th>0.004</th>
<th>0.2</th>
</tr>
</thead>
<tbody>
<tr>
<td>This Study</td>
<td>Modeled from Polar MM5, driven by NN2, 1985-2001</td>
<td>P-E*</td>
<td>157 +/- 9</td>
<td>0.47</td>
<td>0.81</td>
<td>-0.015</td>
<td>-0.6</td>
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<td>This Study</td>
<td>Modeled from E40, 1985-2001</td>
<td>P-E*</td>
<td>135 +/- 7</td>
<td>-0.29</td>
<td>0.62</td>
<td>0.010</td>
<td>0.3</td>
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<tr>
<td>This Study</td>
<td>Modeled from NN2, 1985-2001</td>
<td>P-E*</td>
<td>84 +/-9</td>
<td>0.58</td>
<td>0.74</td>
<td>-0.019</td>
<td>-0.7</td>
</tr>
<tr>
<td>This Study</td>
<td>Modeled from JRA, 1985-2001</td>
<td>P-E*</td>
<td>155 +/-10</td>
<td>-0.47</td>
<td>0.88</td>
<td>0.015</td>
<td>0.6</td>
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</tbody>
</table>
Fig. 2.1. Polar MM5 60-km domain and topography used in this study. Topography contours are shaded. Ice core and snow stake sites are indicated by triangles (clusters of related sites are circled). Locations of common locations and features are also shown.
Fig. 2.2. Comparison of annual accumulation records from ice cores and stake networks to simulated P-E from PMM5_E40, PMM5_NN2, E40, NN2, and JRA. The simulated P-E is interpolated to the location of the observation. Where possible, ice core records have been averaged together by region to reduce noise. The regions are defined in Table 2.1 and shown in Fig. 2.1. (a) Mean accumulation (mm yr⁻¹) for the 10 years encompassing the first five years (1985-1989) and the last five years of the record (period varies depending on the length of the observational record– see Table 2.1). (b) Change (mm) between first and last five years of each record. (c) Correlation coefficient of observed versus simulated annual accumulation from 1985 to the end of the record (dashed lines indicate the correlation is significant from zero at the 90% confidence interval). The mean accumulation and changes at Law Dome are not shown because they are not published as of this writing (personal communication, Tas van Ommen).
Fig. 2.3. Long-term accumulation distribution (mm y$^{-1}$ water equivalent) from (a) Vaughan et al. (1999), (b) 1985-2001 PMM5_E40 P-E, (c) 1985-2001 PMM5_NN2 P-E, (d) Vaughan minus PMM5 E40 and (e) Vaughan minus PMM5 NN2.
Fig. 2.4. Annual snowfall trends (mm y\(^{-2}\)) for 1985-2001 derived from a linear fit through each grid point for the (a) PMM5_E40 and (b) PMM5_NN2 runs. (c) and (d) are the same as (a) and (b), but expressed as percent change per decade. Hatching indicates the trend is significant from zero at the 90% confidence interval.
Fig. 2.5. Seasonal snowfall trends (mm y\(^{-2}\)) for 1985-2001 derived from a linear fit through each grid point for the (a) PMM5 E40 DJF, (b) PMM5 NN2 DJF, (c) PMM5 E40 MAM and (d) PMM5 NN2 MAM runs. Hatching indicates the trend is significant from zero at the 90% confidence interval. The DJF trends begin in December 1985 and end in February 2002.
Fig. 2.6. Seasonal snowfall trends (mm y⁻²) for 1985-2001 derived from a linear fit through each grid point for the (a) PMM5_E40 JJA, (b) PMM5_NN2 JJA, (c) PMM5_E40 SON and (d) PMM5_NN2 SON runs. Hatching indicates the trend is significant from zero at the 90% confidence interval.
Fig. 2.7. Regionally averaged annual and seasonal snowfall trends (mm y$^{-2}$) for 1985-2001. The regions are RIS (Ross Ice Shelf), RFS (Ronne-Filchner Ice Shelf), AP (Antarctic Peninsula), WA (West Antarctica), EA (East Antarctica), GIS (grounded ice sheet), and ALL (entire model domain). Only land/ice shelf grid points are included in the averages, except for “ALL”. A diamond above a bar indicates the trend is significant from zero at the 90% confidence interval.
Fig. 2.8. Time series of annual and seasonal forecast snowfall over the grounded ice sheet (mm y\(^{-1}\)) for the two sets of Polar MM5 runs as well as E40, NN2, JRA, and the composite (“COMP”) of PMM5_E40, PMM5_NN2, E40, and JRA.
CHAPTER 3

INSIGNIFICANT CHANGE IN ANTARCTIC SNOWFALL SINCE THE
INTERNATIONAL GEOPHYSICAL YEAR

3.1 Introduction

Global sea level (GSL) has been increasing by 1.7 mm y\(^{-1}\) over the past century (Church and White 2006) and 2.8 mm y\(^{-1}\) over the past decade (Leuliette et al. 2004). One of the greatest uncertainties in predictions of GSL rise is the contribution of the Antarctic ice sheets (Vaughan 2005). The Antarctic ice budget is balanced by the buildup of snowfall in the interior, and wastage due to the melting and calving of ice along the coastal margins. Future scenarios from global climate models (GCMs) suggest that Antarctic snowfall should increase in a warming climate mainly due to the greater moisture-holding capacity of warmer air (Huybrechts et al. 2004), partially offsetting enhanced loss at the ice sheet peripheries. Perplexing temperature trends have been reported over Antarctica since continuous monitoring began with the International Geophysical Year (IGY) in 1957-1958, varying by the season, region, and the time period analyzed (Turner et al. 2005b, Comiso 2000). A recent study suggests a strong tropospheric warming signal is manifested over Antarctica during winter since the early 1970s (Turner et al. 2006), the season during which much of the continent receives its maximum snowfall (Bromwich 1988). Satellite-based ice velocity and altimetry
measurements indicate that the West Antarctic Ice Sheet (WAIS) has been thinning over the past decade with a contribution to GSL rise of 0.13-0.16 mm y\(^{-1}\) (Rignot and Thomas 2002, Zwally et al. 2005), consistent with widespread melting of ice sheet grounding lines (Rignot and Jacobs 2002). In light of these studies, it is essential to assess whether Antarctic snowfall has been increasing.

The latest studies employing global and regional atmospheric models to evaluate changes in Antarctic snowfall indicate that no statistically significant increase has occurred since ~1980 over the entire grounded ice sheet, WAIS, or the East Antarctic Ice Sheet (EAIS) (Monaghan et al. 2006a, van de Berg et al. 2005, van den Broeke et al. 2006a). A validation of the modeled-versus-observed changes (Monaghan et al. 2006a) suggests that the recent model records are more reliable than the earlier global model records that inferred an upward trend in Antarctic snowfall since 1979 (Bromwich et al. 2004a). The new studies also clearly show that interannual snowfall variability is considerable; yearly snowfall fluctuations of +/-20 mm y\(^{-1}\) water equivalent (WEQ), i.e., +/-0.69 mm y\(^{-1}\) GSL equivalent, are common (Monaghan et al. 2006a), and might easily mask underlying trends over the short record. It is necessary to extend the snowfall record back to the IGY so that (a) trends can be assessed within a longer context; (b) it can be compared with the entire instrumental temperature record over Antarctica; and (c) to make available a 50-y benchmark for GCM evaluation.

The small volume of meteorological data over the Southern Ocean and Antarctica renders modeled snowfall amounts highly questionable prior to the modern satellite era (~1979) (van de Berg et al. 2005, Bromwich and Fogt 2004). The only other records of snowfall variability before 1979 are from ice cores, snow pits, and precipitation gages.
The spatial coverage of these data has been too sparse to accurately assess snowfall accumulation over the entire continent. However, in recent years scores of new ice core records have become available, due in large part to the International Transantarctic Scientific Expedition (ITASE), a multi-national field program aimed at reconstructing the recent climate history of Antarctica through ice coring and related observations along an extensive network of traverses (Mayewski and Goodwin 1997). In this study these new records are employed together with existing ice cores, snow pit and snow stake data, meteorological observations, and validated model fields to reconstruct Antarctic snowfall accumulation over the last 5 decades.

3.2 Methodology

Each observational record is representative of an area surrounding it (a “zone”), the size of which depends on the atmospheric circulation, the interaction of wind with topography, and the time scale considered. The method uses meteorological model reanalysis fields to determine zones of snowfall coherence that correlate with the individual records at annual timescales. Assuming these zones adequately cover most of the continent given the available observational records, this information can be employed to synthesize the observations into a continent-wide record of snowfall accumulation in a self-consistent manner. The model reanalysis dataset used is the European Centre for Medium-Range Weather Forecasts 40-y Reanalysis (ERA-40; Uppala et al. 2005). Snowfall accumulation is defined from ERA-40 precipitation fields adjusted to match long-term observed accumulation records (described in the Appendix). Precipitation dominates snowfall accumulation variability over Antarctica at model grid scales.
(Bromwich 1988, Bromwich et al. 2004a). ERA-40 precipitation is compared to independent observed accumulation records for overlap periods and shown to largely reproduce the interannual snowfall accumulation variability and trends, justifying its use for this study (Appendix). Figure 3.1 shows a composite map of the maximum correlation coefficient obtained by correlating the ERA-40 simulated percentage annual precipitation anomaly at the grid point closest to each core with every other grid point. Correlations greater than 0.5 (p<0.01) occur over most of the grounded ice sheet, indicating that the zones of spatial coherence from the available observational records cover nearly the entire continent. This robust relationship is employed to synthesize the observational data into a series of continent-wide snowfall accumulation maps for the period prior to 1985, when the snowfall variability simulated by ERA-40 is questionable (Monaghan et al. 2006a). The result is a 5-decade time series of snowfall accumulation over the grounded ice sheet; the first 3 decades are inferred from observational records, and the final 2 decades from ERA-40. A detailed description of the methodology, including the technique by which uncertainty is assessed, is given in the Appendix.

3.3 Results and Discussion

The spatial distribution of the 50-y average annual snowfall accumulation (Fig. 3.2a) closely resembles the glaciological estimate of Vaughan et al. (1999). The mean for the grounded ice sheet is 182 mm y\(^{-1}\) WEQ, compared to the lower value of 149 mm y\(^{-1}\) WEQ from the Vaughan map. A subsequent analysis (van de Berg et al. 2006) suggests that the Vaughan map underestimates coastal accumulation and that a more realistic estimate is 171 mm y\(^{-1}\) WEQ. Overall, the reconstructed mean annual snowfall
accumulation is at the high end of published estimates [119-197 mm y\(^{-1}\) WEQ (van de Berg et al. 2005, Ohmura et al. 1996] but may be realistic in light of recent findings.

The percentage differences of annual snowfall accumulation for each decade with respect to the 50-y mean (Fig. 3.2a) are shown in Figs. 3.2b-f. There are regions of both positive and negative change in all 5 decades, but no continental-scale changes of either sign dominate any period. The amplitude of the changes in Figs. 3.2b-d, the decades reconstructed from ice cores, is slightly dampened compared the final two decades (Figs. 3.2e-f). This is partly due the reconstructed data having smaller interannual variability than the model data; however this does not affect the sign of the changes and has little impact on the results at basin and continental-scales (shown in the Appendix). There is no widespread signal of increased snowfall accumulation over the EAIS for 1995-2004 that would suggest a contribution to the recently reported thickening (Davis et al. 2005). The 1995-2004 changes are mostly negative over WAIS, where net ice sheet thinning is occurring (Rignot and Thomas 2002, Zwally et al. 2005). The statistical uncertainty associated with the change at each grid point (due to the decadal variability and methodology, not shown) is typically about 4-8%, enough to overwhelm the decadal changes in most places.

Figure 3.3 shows the time series of snowfall accumulation inferred from Figs. 3.2b-e averaged over EAIS, WAIS, and the entire grounded ice sheet. All three regions are characterized by a steady upward trend from the beginning of the record through the early 1990s, and then a downward trend thereafter that is most marked over WAIS (22 mm y\(^{-1}\) WEQ for the past decade compared to the prior decade). However, this change has low statistical significance (p=0.16), indicating decadal fluctuations of this magnitude (~7%
of the 50-y mean) are probably common over WAIS. The upward trend over the ice sheets prior to the most recent decade corroborates earlier studies employing regional records (Morgan et al. 1991, Mosley-Thompson et al. 1995). Over EAIS, WAIS, and the grounded ice sheet there are no statistically significant trends in snowfall accumulation over the past 5 decades, including recent years for which global mean temperatures have been warmest (Levinson 2005). Several experiments were performed to test the sensitivity of the results in Fig. 3.3 by adjusting parameters within the methodology, and employing other methods to reconstruct the accumulation, and the results were very robust (see Appendix).

The findings presented here are somewhat inconsistent with Davis et al. (2005), who inferred from satellite altimetry data that an increase in snowfall accumulation was the primary cause of net thickening over EAIS for 1992-2003. One reason for the discrepancy may be that their radar data do not extend southward of 81.6°S, a region with strong downward trends in the past decade (Fig. 3.2f). Another factor may be their methodology. Zwally et al. (2005) found a thickening over EAIS from satellite altimetry for a similar period that was a factor of 3 smaller than the Davis study, arguing that their method more accurately accounts for firn compaction and interannual variability of dH/dt. Finally, because snowfall typically adjusts to climate change on much shorter timescales than the underlying glacial ice (Thomas et al. 2004), a linear thickening trend as reported in the Davis study could be interpreted to mean that snowfall accumulation from 1992-2003 was stepwise higher than at some time in the past when the accumulation rate and the ice sheet dynamical response were in equilibrium. In that case, the results of Davis et al. (2005) may actually suggest that snowfall accumulation over
EAIS has changed little in the past decade, consistent with this assessment. Despite disagreeing as to the causality, it is not disputed that altimetry indicates a clear thickening signal over EAIS (Zwally et al. 2005, Davis et al. 2005) that mitigates sea level rise.

3.4 Conclusions

The implications of the results presented here are categorized into two general ideas.

1. **Interannual and interdecadal snowfall variability must be more seriously considered when assessing the rapid ice volume changes that are occurring over Antarctica.** With regard to interannual variability, consider a recent estimate of Antarctic ice sheet mass loss that is the equivalent of 0.4 +/-0.2 mm y\(^{-1}\) GSL rise for 3 years (2002-2005) from satellite-derived time-variable gravity measurements (Velicogna and Wahr 2006). Antarctic-wide annual snowfall accumulation decreased by about 25 mm y\(^{-1}\) WEQ, or approximately 0.86 mm y\(^{-1}\) GSL rise, between calendar year 2002 and 2003 (Fig. 3.3), suggesting that the gravity fluctuations could be heavily influenced by interannual snowfall variations.

With regard to interdecadal variability, the ERA-40 snowfall accumulation is about 22 mm y\(^{-1}\) WEQ lower over WAIS for the past decade (1995-2004) compared to the previous decade (1985-1994) (Fig. 3.3), the GSL equivalent of 0.18 mm y\(^{-1}\). This signal is of the same order as the 47 Gt (0.13 mm y\(^{-1}\) GSL equivalent) mass imbalance reported for WAIS (defined by a slightly different area) from satellite radar altimetry for roughly the past decade (Zwally et al. 2005). In neither decade is the snowfall accumulation statistically-significantly different from the 50-y WAIS mean, suggesting that such large
fluctuations are normal. The cause of the recent mass imbalance will remain unclear until a longer satellite record is available, but it may be partly related to accumulation variability.

2. Antarctic snowfall is not currently compensating for the oceanic-induced melting at the ice sheet periphery. If anything, the 50-year perspective presented here suggests that Antarctic snowfall has slightly decreased over the past decade while global mean temperatures have been warmer than at any time during the modern instrumental record (Levinson 2005). Radiosonde and ERA-40 temperature data indicate a uniform winter warming trend in the mid-troposphere over Antarctica since the early 1970s, but seasonally-averaged ERA-40 precipitation data suggest there has been no commensurate increase in winter snowfall since at least 1985 (Monaghan et al. 2006a). This suggests that atmospheric circulation variability, rather than thermodynamic moisture increases, may dominate recent Antarctic snowfall variability.

The technique of synthesizing observational records with model reanalysis has provided a coherent record of Antarctic-wide snowfall accumulation variability extending back prior to the modern satellite era. As more and improved (e.g., ground penetrating radar) accumulation records become available it will be possible to revisit this study with greater accuracy. A longer (1-2 centuries) reconstruction was not possible due to the limitations of the current dataset, but clearly is necessary to better understand the multidecadal Antarctic accumulation variability. Satellite-based techniques show great promise for precisely measuring Antarctic ice mass changes. It is critical to extend these records to distinguish thickening/thinning signals from snowfall variability.
The results presented indicate that there is not a statistically significant global warming signal of increasing snowfall over Antarctica since the IGY, inferring that GSL rise has not been mitigated by recently increased Antarctic snowfall as expected. It may be necessary to revisit GCM assessments that show increased snowfall over Antarctica by the end of this century in conjunction with projected warming (Intergovernmental Panel on Climate Change 2001). Vigorous efforts are needed to better understand this remote but important part of the planet and its role in global climate and sea level rise.
3.5 Figures

Fig. 3.1. The composite map of the maximum absolute value of the Pearson’s correlation coefficient (|r|) resulting from correlating the ERA-40 1985-2004 percentage annual snowfall accumulation change (with respect to the 1985-1994 mean) for the grid box containing each of the 16 observation sites (yellow dots/numbers) with every other 1°x1° grid box over Antarctica (i.e., this map is a composite of 16 maps). Pink/red colors have correlations at p<0.01. The black lines delineate ice drainage basins (20), which are identified alphabetically by the black letters where they intersect the grounding line. Detailed information about the observation sites is included in the Appendix.
Fig. 3.2. a) 50-year mean annual snowfall accumulation (mm y\(^{-1}\) WEQ). b)-f) difference between mean annual snowfall accumulation for decade indicated and 50-year mean, expressed as a percentage of the 50-year mean. The scale shown in b) applies to b)-f). The mean accumulation, trends, and uncertainty are quantified for each basin in the Appendix.
Fig. 3.3. Time series of decadal mean of annual snowfall accumulation (mm y$^{-1}$ WEQ) for 1955-2004 for EAIS, WAIS, and the grounded ice sheet (ALL), calculated as described in the text. The annual accumulation is also shown for the last 2 decades, the period for which ERA-40 is used. The dotted line represents the 50-y mean. The basins which define EAIS and WAIS and ALL are given in the Appendix. Uncertainty bars are +/- 1σ per the methodology (see Appendix). The uncertainty bar at the far left of each graph is for the 50-y mean.
CHAPTER 4

AN EVALUATION OF ANTARCTIC NEAR-SURFACE AIR TEMPERATURE AND SNOWFALL IN IPCC AR4 GCMS

4.1 Introduction

About 70 m of global sea level is locked within the ice of Antarctica (Huybrechts et al. 2000). Considering the potential influence of changes in near-surface temperature and snowfall accumulation to the mass balance of Antarctic ice (e.g., Huybrechts et al. 2004), gaining an understanding of the recent variability of temperature and snowfall, and how they might change in the future, is essential. Antarctic temperature and snowfall are subject to substantial interdecadal variability (Schneider et al. 2006, Bromwich et al. 2000), and therefore observational records of at least several decades are desirable for studies of climatic variability. Near-surface temperature and snowfall records spanning nearly 5-decades and representative of the entire continent have recently been published, providing the impetus for this study. The goals are three-fold:

i. To compare recently published Antarctic near-surface temperature datasets, including a new one presented here for the first time, with the intention of converging on an accurate observational record of near-surface Antarctic temperature since 1960;
ii. To assess the observed sensitivity of Antarctic snowfall to near-surface temperature changes using the temperature records and the 50-year snowfall record of Monaghan et al. (2006b);

iii. To apply the new observational records to assess simulations of Antarctic near-surface temperature and snowfall in IPCC AR4 GCMs, as well as the sensitivity of Antarctic snowfall to near-surface temperature changes in the GCMs. The sensitivity comparison provides an indication of the reliability of GCM projections of increased snowfall in the context of projected Antarctic temperature changes, including the implications for global sea level change at the end of the 21st century.

Inhomogeneous climate changes have occurred in the Antarctic since continuous monitoring began with the International Geophysical Year (IGY) in 1957. Turner et al. (2005b) examine station temperature records for the past 50 years and report statistically insignificant temperature fluctuations over the continental Antarctica excluding the Antarctic Peninsula, with the exception of Amundsen-Scott South Pole Station, which cooled by -0.17°K decade⁻¹ for 1958-2000 (p<0.10). Turner et al. (2005b) find major warming over most of the Antarctic Peninsula, including a trend of +0.5°K decade⁻¹ at Faraday/Vernadsky station for 1951-2000 (p<0.05), compared to a global trend of +0.2°K decade⁻¹ for 1975-2004 (during which global temperatures increased more rapidly than any other period in the 20th century; Hansen et al. 2006). However, Turner et al. (2005b) report that the more recent data (1971-2000) have smaller warming (greater cooling) trends than the longer record (1961-2000) at all but 2 coastal stations. The finding of increasingly negative trends in the most recent decades is corroborated by Chapman and
Walsh (in press); they perform a gridded objective analysis of Antarctic near-surface temperatures and note that prior to 1965 the continent-wide annual trends (through 2002) are slightly positive, but after 1965 they are mainly negative (despite warming over the Antarctic Peninsula). Likewise, Kwok and Comiso (2002) find a statistically insignificant cooling trend over continental Antarctica from 1982-1998, inferred from skin temperatures from Advanced Very High Resolution Radiometer (AVHRR) instruments on polar orbiting satellites. Schneider et al. (2006) reconstruct Antarctic temperatures from ice core stable isotope records and find that despite large annual and decadal variability, a slight warming of about 0.2°K century\(^{-1}\) has occurred since ~1880 which appears to be weakly in phase with the rest of the Southern Hemisphere.

The “warm-Peninsula-cold-continent” temperature trend pattern that emerges in most Antarctic temperature evaluations has been attributed mainly to a positive trend in the leading mode of Southern Hemisphere climate variability, the Southern Hemisphere Annular Mode (SAM; Rogers and van Loon 1982, Thompson and Wallace 2000, Marshall 2003, Schneider et al. 2006, Gillet et al. 2006, Marshall 2007). The SAM has steadily increased since the 1960s (Marshall 2003), although it has leveled off since ~2000 (http://www.nerc-bas.ac.uk/icd/gjma/sam.html). The cause of the increase in the SAM is still not entirely clear, although recent modeling studies suggest it may be linked to anthropogenic changes due to greenhouse gas increases and decreasing stratospheric ozone over Antarctica (e.g., Shindell and Schmidt, 2004, Arblaster and Meehl 2006, Cai and Cowan 2007). The seasons for which the positive SAM trends have been strongest are summer and autumn, and accordingly these are the seasons in which the temperature trends at many continental stations have been most strongly negative in recent decades.
Over the Peninsula, the seasonal temperature changes are complicated. The strongest warming trends are in winter on the western side of the Peninsula, a season for which the SAM has not changed much over the past several decades, but there has been a regional sea ice reduction (Zwally et al. 2002). Along the northeastern tip, the trends have the greatest statistical significance in summer. Marshall et al. (2006) attribute this to changes in the SAM that increase the frequency of air masses that are advected over the Peninsula orography. The SAM has an important influence on observed Antarctic near-surface temperature variability, but other factors also play key roles, such as regional ocean circulation variability and air-sea-ice feedbacks (Vaughan et al. 2003), and the El Nino-Southern Oscillation (Bromwich et al. 2004b).

Obtaining an accurate record of recent Antarctic snowfall variability has been more difficult than for near-surface temperature, mainly because there are fewer observations and due to the difficulty of measuring snowfall accurately (e.g., Sevruk 1982). Techniques based on satellite altimetry are promising but cannot yet decouple snowfall from changes in surface elevation due to densification and glacial dynamics (e.g., Wingham et al. 1998). Thus, numerical models have been the primary tool for assessing the temporal variability of Antarctic snowfall. An upward trend of Antarctic snowfall of \( \sim 1.5 \, \text{mm yr}^{-2} \) was inferred by Bromwich et al. (2004a) for the period 1979-1999 from large-scale global models. Likewise, Davis et al. (2005) suggest that thickening over East Antarctica from satellite altimetry is due primarily to increased snowfall, asserting that over East Antarctica to \( \sim 81^\circ \text{S} \) positive snowfall trends mitigate sea level rise linearly by about 0.12 mm y\(^{-1}\) between 1992-2003. On the contrary, a series of recent studies employing global and regional model records indicate that no statistically significant
snowfall change has occurred in the past 20-25 years over the entire grounded ice sheet, West Antarctica, or East Antarctica (Monaghan et al. 2006a, van de Berg et al. 2005, van den Broeke et al. 2006a). Additionally, validation results by Monaghan et al. (2006a) suggest that the recent model records are more reliable than the earlier model records used by Bromwich et al. (2004a) to infer the ~1.5 mm y⁻² upward trend in Antarctic snowfall. An unfortunate problem with any of the models that cover a period longer than the last ~2-1/2 decades is that their precipitation simulations are problematic prior to the modern satellite era, about 1979. To extend the Antarctic snowfall record to the years prior to 1979, Monaghan et al. (2006b) employ ice core snowfall accumulation records, along with a snowfall background field derived from 1985-2004 modeled precipitation fields, to reconstruct Antarctic snowfall accumulation back to 1955. The result for the 5-decade time period is consistent with the findings from the latest modeling studies that cover the past 2 decades: there is no statistically significant trend in overall annual Antarctic snowfall since 1955.

In summary, despite a strong global warming trend (Hansen et al. 2006), recent literature suggests there has been little overall change in annual Antarctic near-surface temperature and snowfall during the past 5 decades, notwithstanding some seasonally-dependent regional changes (e.g., Turner et al. 2005b, Monaghan et al. 2006b). Here, newly available, extended Antarctic near-surface temperature and snowfall records are used to evaluate global climate model (GCM) simulations undertaken as part of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). In Section 2, data and methods are outlined. In Section 3, an evaluation of observationally-based Antarctic near-surface temperature records is presented, including
a new record presented for the first time. In Section 4, near-surface temperature and snowfall are evaluated for a representative selection of GCMs. In Section 5, the causality of the amplified Antarctic temperature simulations in the GCMs is investigated, followed by conclusions.

4.2 Data and Methods

4.2.1 Existing observational datasets

The observational datasets (to which the GCMs will be compared) are those that are representative of the entire Antarctic continent, and include:

i. Time-series of annual and seasonal near-surface air temperature from gridded objective analysis (1°x1°) of automatic and manned station records and ocean observations (1950-2002; Chapman and Walsh in press);

ii. A time-series of annual near-surface air temperature derived by linearly regressing stable isotope records from ice cores onto a representative Antarctic temperature record from station data (1800-1999; Schneider et al. 2006);

iii. Time-series of annual and seasonal skin temperature from a gridded 12.5x12.5 km polar stereographic AVHRR dataset (1982-2003; Kwok and Comiso 2002);

iv. A time-series of annual snowfall created by blending ice core snow accumulation records and bias-adjusted atmospheric model precipitation fields from a gridded 1°x1° dataset (1955-2004; Monaghan et al. 2006a).
The three temperature datasets (i, ii, iii) have been validated within their respective citations. Below they are compared to the new temperature reconstruction presented here. As the datasets result from different data and methods, comparing them provides a means of assessing their robustness. The Antarctic snowfall dataset (iv) is validated in Monaghan et al. (2006b) and is not further evaluated here.

4.2.2 A new near-surface temperature reconstruction

Monthly mean near-surface air temperature records from manned stations have been acquired from the Reference Antarctic Data for Environmental Research (READER) database (http://www.antarctica.ac.uk/met/READER/) (Turner et al. 2004). Each temperature record is representative of an area surrounding it (a “zone”), the size of which depends on factors such as the atmospheric circulation and the topography. The kriging-like method uses multi-year meteorological model temperature reanalysis fields from the European Centre for Medium Range Weather Forecasts 40-year Reanalysis (ERA-40; Uppala et al. 2005) as a background variable to determine zones of temperature coherence that correlate with the individual records at annual and seasonal timescales. In Section 2, ERA-40 temperature is compared to other Antarctic temperature records and shown to largely reproduce the interannual variability and trends, justifying its use for this study. Given the network of available records, if the zones of temperature coherence cover most of the continent, the observational records can be synthesized into a continent-wide record of temperature in a self-consistent manner. The technique used here generates a result that has a greater physical basis than traditional objective analysis techniques, which typically rely on functions of distance as weighting schemes. Such
methods can neglect the topographic variations, atmospheric teleconnections, or other atmospheric phenomena that are inherently accounted for in the meteorological reanalysis fields. The methodology for the new temperature reconstruction is similar to that used to reconstruct snowfall in Monaghan et al. (2006b).

The generalized objective analysis technique (Cressie 1993) is specified as:

\[
\hat{Z}(i, j) = \sum_{k=1}^{n} \lambda_{i,j,k} \times Z_k ; \quad \sum_{k=1}^{n} \lambda_{i,j,k} = 1
\]

[1]

where \( \hat{Z}(i, j) \) is the predicted value of a quantity at a desired grid point with coordinates \((i, j)\), \( n \) is the number of observations (i.e., ice cores), \( Z_k \) is the known quantity at the \( k \)th observation site, and \( \lambda_{i,j,k} \) is a predictor (weighting coefficient) that must sum to 1. The predictor, \( \lambda_{i,j,k} \), is computed by exploiting the information about spatial variability provided by the 1980-2001 gridded 2-m temperature fields from ERA-40:

\[
\lambda_{i,j,k} = \frac{r_{i,j,k}^2}{\sum_{k=1}^{n} r_{i,j,k}^2}
\]

[2]

where \( r_{i,j,k} \) is the Pearson’s correlation coefficient between percentage change at any grid point and the grid point of the \( k \)th observation. Figure 4.1 shows a composite map of the maximum \( r_{i,j,k} \) at each grid point (i.e, the highest correlation obtained by correlating temperature at each grid point with the \( n \) number of grid points corresponding to the observation locations). Statistically significant correlations \((r \geq 0.4 \, p < 0.05)\) occur over 96% of the ice sheet surface area, and correlations of \( r \geq 0.6 \) occur over 90% of the surface area, indicating that the available observational records are representative of the continent-wide temperature variability. Equation [1] is next applied to interpolate the
percentage monthly and annual temperature change of the \( k \)th observation with respect to the 1980-2001 baseline period, \( \Delta_{i,k} \), to the entire grid:

\[
\Delta_{i,j} = \sum_{k=1}^{n} \Delta_{i,j,k} \times \Delta_{i,k} \times \eta_{i,j,k} \quad ; \quad \eta_{i,j,k} = \frac{r_{i,j,k}}{|r_{i,j,k}|} \quad [3]
\]

where \( \Delta_{i,j} \) is the percentage monthly temperature change at each grid point with respect to the 1980-2001 period. The operator \( \eta_{i,j,k} \) accounts for the sign of anticorrelations (it is assumed that if a core region is anticorrelated with a grid point that the relationship is just as likely to be valid as a positive correlation since it too is likely to arise due to the atmospheric circulation). Equation 3 is applied to the monthly and annual averages for each year from 1960-2005. The resulting percentage change is converted to a temperature anomaly (°K) using the 1980-2001 mean monthly temperature in ERA-40 at each grid point. The variance of the resulting time series is adjusted to match the 1980-2001 variance in ERA0-40 at each grid point by multiplying the reconstructed temperature by \( \sigma_{\text{ERA-40}}/\sigma_{\text{reconstruction}} \). Seasonal temperature anomalies are then computed from the monthly anomalies and averaged (area-weighted) over the continent, including ice shelves. Anomalies are recalculated with respect to the 1980-1999 mean to allow comparison with other datasets. For the GCM comparison in sections 4.4 and 4.5, a 5-y running mean filter is applied to smooth the data and the number of degrees of freedom are reduced to \( (n-2)/5 \) to account for autocorrelation in the uncertainty calculations.

The records obtained from the READER website are quality controlled and monthly means are calculated only if >90% of data are available. The READER data are supplemented by observations provided by Gareth Marshall (http://www.nerc-bas.ac.uk/icd/gima/) in cases where his data are more complete. In order to have
complete records for the entire 46-year period, missing months are filled in using single or multiple linear regression based on records at nearby stations. In most cases, these data outages are a few months, with the exception of Byrd Station. Byrd does not have year-round manned records after 1969, although there are scattered summer observations through January 1975. Efforts were made to fill in the missing data because it is an isolated record in West Antarctica, where data are otherwise unavailable. Automatic weather station (AWS) observations are available from 1980-2002, but the outages are frequent and data are available for only ~50% of the months during that period. A reconstruction of Byrd temperature from 1978-1997 based on passive microwave data (Shuman and Stearns 2001, 2002) was obtained from the National Snow and Ice Data Center (http://www.nsidc.org). The passive microwave record matches the AWS record closely for the months in which both are available ($r^2=0.999$, $n=150$, $p<0.0001$), and thus the passive microwave data are considered reliable. The station and passive microwave records were combined into one record, and then the remaining missing data were filled in by optimizing the multiple linear regression relationship between the Byrd Station temperature record and records from other Antarctic stations for each month, and for the annual means. The various time series of annual near-surface temperature at Byrd are shown in Fig. 4.2. The regressed temperature record matches the observed Byrd records adequately ($r^2=0.65$, $n=29$). To test the sensitivity to this record, the Antarctic temperature was reconstructed with and without the Byrd record (shown in Section 3) and there is virtually no difference in the result. Thus, at the continental-scale, the Antarctic temperature reconstruction is not sensitive to the Byrd Station record.
Including the passive microwave data at Byrd, 95.6% of station-months for 1960-2005 are available for the 15 stations shown in Fig. 4.1. If Byrd Station is omitted, 97.7% of station-months are available.

4.2.3 IPCC AR4 GCMs

Five IPCC AR4 GCMs are chosen for evaluation, described in Table 4.1. Data are obtained from the World Climate Research Program (WCRP) Climate Model Intercomparison 3 (CMIP3) multi-model database (https://esg.llnl.gov:8443/home/public/HomePage.do). The “20c3m” scenario is used for the evaluation of 19th and 20th century climate. The criteria for the models chosen are that they have at least 4 ensemble members for both the 20c3m and sresa1b (21st century; assumes mid-range greenhouse gas increases)) scenarios, although in this study only 203cm runs are analyzed. All output fields are converted to anomalies with respect to the 1980-1999 means, and then a 5-year running mean filter is applied to reduce noise. All trends are calculated with linear regression. Ninety-five percent confidence intervals are calculated for the 5-year running mean values, with reduction of the degrees of freedom to (n-2)/5 to account for autocorrelation.

4.3 Evaluation of observationally-based Antarctic near-surface temperature records

4.3.1 Pros and cons of various Antarctic temperature datasets.

In order to understand Antarctic climate variability and to diagnose GCMs, having records that are representative of near-surface temperature over the entire Antarctic continent is desirable. One method of doing this is to simply take the linear average of all
station records available (e.g., Reid and Jones 2001). These types of analyses are useful for assessing year-to-year variability, but are not reliable for evaluating trends because of the relatively sparse network of observing stations. Temporal trends calculated by linear averaging indicate spurious warming for recent decades because a disproportionate number of stations are located on the Antarctic Peninsula, a region whose ice comprises only ~5% of the total surface area of the ice sheet (Vaughan et al. 1999), where strong warming has occurred over the past 50 years (e.g., Vaughan et al. 2003). Individual station records suggest that there has not been statistically significant warming elsewhere on the continent (e.g., Turner et al. 2005b). Due to the problem cited, linearly-averaged Antarctic temperature records are not employed in this study.

Objective analysis methods (Doran et al., 2002, Chapman and Walsh in press) have reduced problems due to linear averaging, as these methods interpolate/extrapolate to voids using station data (either trends calculated from the station data, or raw station data) that is weighted as a function of inverse distance or a natural neighbor scheme (Cressie 1999), and thereby weight stations in remote areas more heavily than those that are clustered together. These analyses do not show strong warming trends and indicate that Antarctic temperatures collectively have not changed significantly since the 1960s. Statistically insignificant cooling over most of the continent has occurred on an annual basis since about 1970 (Chapman and Walsh in press). The annual and seasonal time series from Chapman and Walsh (in press) are used in this study, as they provide the most recent and complete analysis of Antarctic temperatures at the continental scale.

Numerical atmospheric model fields provide useful assessments of temperature over Antarctica, and they account for topography, storm activity, teleconnections, and other
natural phenomena that impact climate. However, one problem that has plagued model reanalysis fields in Antarctica is the dearth of observational data assimilated into the models prior to the modern satellite era (~1979). This leads to relatively poor simulations before ~1979, and quite good simulations thereafter (e.g., Bromwich and Fogt 2004). Thus, the evaluation and use of ERA-40 temperatures is limited to the period 1980-2001 in this study. The 1980-2001 ERA-40 annual and monthly temperature fields are used to create the background field the statistical reconstruction, allowing temperature to be interpolated/extrapolated to data voids from station observations in a physically-based manner.

Skin temperature from AVHRR instruments onboard the National Oceanic and Atmospheric Administration’s suite of polar orbiting satellites is the final Antarctic temperature dataset used. AVHRR records provide the most spatially-comprehensive observations of Antarctic temperatures. However, they are only valid for clear-sky conditions, an issue that can be problematic in the coastal Antarctic regions where conditions are more often cloudy than not (Guo et al. 2003). A thorough description of the AVHRR record and its quality over Antarctica is given by Comiso (2000). The most recent realization of the AVHRR temperature dataset is used in this study, provided by Josefino Comiso of the National Aeronautics and Space Administration (NASA). The most recent published version of the dataset for Antarctica is Kwok and Comiso (2003).

4.3.2 Validation of the new Antarctic temperature reconstruction

Monthly temperature records from sixteen independent stations were selected from the READER database to validate the new Antarctic temperature reconstruction (Table 4.2).
The stations were chosen based on completeness of record, and to provide a representative sampling of the climatic variability across Antarctica. Eight stations are located on the coast, and eight are in the interior of Antarctica, six of which are above 1000 m ASL. Five of the stations have records that begin prior to 1980, the beginning of the calibration period for the reconstruction.

Figure 4.3a shows the monthly and annual standard error and correlation between the observed and reconstructed temperature anomalies. The statistics for January, for example, are calculated for all available January observations from all 16 stations (n=231). The comparison is exact, in that the same 231 months are used from the reconstructed data (rather than using the entire reconstructed record). Thus, the results provide an estimate of the average reconstruction skill at one grid point. Correlations are $r>0.7$ during seven months and annually; the lowest correlation is $r=0.58$. The correlations are highest during winter. The standard errors range from $\sim 0.6$-2.0 K, have strong seasonality, being largest in winter and smallest in summer. The standard errors are likely largest in winter due to increased continentality at the coastal stations while sea ice is present, leading to an increase in temperature variability. Figure 4.3b shows the standard error and correlation between the observed and reconstructed temperature anomalies at each of the 16 stations. The statistics for each station are computed from all available monthly averages (column “n” in Table 4.2), and thus give an estimate of the average monthly reconstruction skill at one grid point. Correlations are $r>0.7$ at 13 of the 16 stations. Two-of-the-three records with $r<0.7$ are automatic weather stations (AWSs) in West Antarctica (7 and 10), and the other is Neumayer Station (11), situated in coastal Queen Maud Land. Standard errors range from 1.1-2.0 K. The four largest standard
errors occur in West Antarctica (7, 10, 14, and 15), where only the Byrd Station record is available to constrain the reconstruction. For the three independent records on the Antarctic Peninsula (1, 5, and 13), the reconstruction has comparatively high correlations (r=0.76, 0.73, and 0.86, respectively), and low standard errors, indicating that temperature variability on the Peninsula – where the largest positive temperature trends are occurring (e.g., Turner et al. 2005b) – is well-represented by the reconstruction.

Figure 4.4a shows the monthly and annual temporal trends for the observed and reconstructed temperature anomalies. As for Fig. 4.3a, the statistics are computed from all available observations from all 16 stations for each month, and annually. The trends are not representative of actual temperature trends, as they are based on a linear regression fit to many datasets, all of which are discontinuous. However, they do provide a proxy of the average ability of the reconstruction technique to reproduce observed trends at a grid point. The READER trends are not statistically different from zero for 10 of the 12 months. The READER trends are statistically significant (p<0.5) in February, November, and annually. The reconstructed trends are of the same sign as the READER trends in 11 of 12 months, and annually (a negligible negative trend occurs in July, when the positive READER trend is smallest). The reconstructed trends are not as positive as the as the READER trends, a condition that is partially related to the high statistical uncertainty (in no instance are the reconstructed and READER trends statistically different from each other). The smaller positive trends may also be related to the slightly lower variance in the reconstructed dataset (Table 1, last column). When data is averaged regionally, such differences typically average out (not shown).
Figure 4.4b shows the monthly and annual trends of the yearly averages of all observations. For example, all of the available January observations for each of the 16 stations are averaged together each year, and then a trend is drawn through the resulting 46-year time series (the observations are normalized before averaging in order to avoid spurious significant trends due to variations in the amount of observations available). Thus, the trends in Fig. 4.4b provide a proxy of the ability of the reconstruction to resolve area-averaged trends, which are the focus of the subsequent sections. The reconstructed and READER trends are of the same sign and similar magnitude in all months. Annually, both the READER and reconstructed trends are positive and statistically different from zero. The fact that the normalized READER and reconstructed trends are similar supports the earlier suggestion that the smaller reconstructed trends may be due to the lower variance compared to the READER data.

Figures 4.3 and 4.4 demonstrate that the reconstructed Antarctic near-surface air temperature dataset compares well with independent observations from sixteen stations that are representative of the range of climatic conditions across Antarctica. The reconstructed data adequately capture the observed variability at individual stations, and area-averaged trends are similar to those observed over all months of the year, and annually. The results shown in Figs. 4.3 and 4.4 provide quantitative evidence that the continent-averaged Antarctic temperature data presented next are accurate.

4.3.3 Comparison of Antarctic temperature datasets

Figure 4.5 shows the annual Antarctic near-surface temperature anomalies for various datasets for the 1950-2005 period (Fig. 4.5a), and the more recent period from 1980-
2005, which contains several additional datasets (Fig. 4.5b). There is close agreement between the new reconstruction and that of Chapman and Walsh (in press) for the 1960-2005 period \((r=0.96;\) Fig. 4.5a). Remarkably, the stable isotope reconstruction of Schneider et al. (2006) matches the other two station-record reconstructions quite well \((r=-0.65\) compared to either dataset) considering the small-scale noise and isotope diffusion that inherently occur in ice cores (e.g., van der Veen and Bolzan 1999). For the 1980-2005 period (Fig. 4.5b), the time series have similar interannual variability, including the reconstructions, the ERA-40 temperature data, and a ‘synthetic’ reconstruction that employs ERA-40 records from the 15 observation sites. If the reconstruction methodology were perfect, the synthetic record would exactly match the ERA-40 record. The close match indicates that the synthetic reconstruction reproduces the ERA-40 record very well \((r=0.95)\). The inclusion of the “recon_nobyr” record in Fig. 4.5b demonstrates that the new reconstruction is insensitive to the omission of the problematic Byrd Station record, as the “recon” and “recon_nobyr” records are nearly identical \((r=0.99)\). The final record shown is the AVHRR skin-temperature average (“comiso”). Compared to the other records, the Comiso record is about 1.5 K higher than the other records (the Comiso temperature anomaly is 2.78 K in 1988, but is not plotted on the graph). Filtering the data for outliers at each grid point using the “1.5 interquartile range rule” (e.g., Moore and McCabe 2006), does not change the result, and therefore it cannot be determined whether the Comiso record has an outlier in 1988. The Comiso record has higher interannual variability than the other records; which may be because skin temperatures are only available during clear sky conditions (Comiso 2000).
Figure 4.6 is similar to Fig. 4.5a, but for seasonal Antarctic near-surface temperature anomalies. The Schneider et al. (2006) record is available for annual timescales only, so it is not included. The Comiso AVHRR skin temperature record from 1982-onward is included. The station-record based reconstructions are in close agreement in all four seasons, lending confidence that they are robust despite the different statistical techniques employed to create them. There are no statistically significant trends for 1960-2002 in any season (see the next paragraph). Similar to the annual case, the Comiso record has the greatest interannual variability in all seasons. The anomalously warm conditions in 1988 in the Comiso dataset are manifested in all seasons except DJF. Despite differences in amplitude, the yearly fluctuations in the Comiso dataset are similar to those in the two reconstructions based on station records.

The annual and seasonal Antarctic near-surface temperature trends are calculated for 1960-2002 and 1982-2001 (Table 4.3). The difference in end years (2002 vs. 2001) between the two periods is due to the ERA-40 records ending in 2001. The 1960-2002 annual and seasonal trends are statistically insignificant in all of the available datasets, and the 95% confidence intervals are at least twice as large as the trends in nearly every instance. The trends are of similar magnitude in the CHAPMAN and RECON datasets, suggesting that at the continental scale the results are insensitive to which technique is employed.

The annual and seasonal trends are stronger over the 1982-2001 period, but they are statistically insignificant in all but four cases. The 95% confidence intervals are larger than the 1982-2001 annual temperature trends by a factor of two or more in all six datasets, indicating the annual trends are highly insignificant. The seasonal trends are
stronger than the annual trends, and in all of the datasets are of the same sign in all four seasons, indicating good agreement and suggesting robust results. The RECON and CHAPMAN near-surface air temperature trends are significantly (p<0.5) negative in MAM (-1.1 and -0.78 K decade\(^{-1}\), respectively), although the confidence intervals underestimate the uncertainty as they do not account for uncertainty due to the methodology. The negative trends in DJF and MAM are consistent with the strong downward trend in the SAM during summer and autumn (Marshall 2003, 2007). In JJA, the positive trends are consistent with middle- and upper-tropospheric warming (1979-2001) over Antarctica in winter based on weather balloon observations (Turner et al. 2006). The SAM has not been strengthening during the winter months, raising the question of whether the JJA warming is an analog how Antarctic climate may change in other seasons in the absence of a strengthening SAM. Marshall (2007) notes that over East Antarctica the surface temperature response to SAM forcing displays little seasonality; i.e., if SAM forcing in other seasons were similar to winter, the temperature response in those seasons might also be similar. One GCM study (Shindell and Schmidt 2004) suggests the trends in the SAM might level off by mid-century if the Antarctic ozone hole mends itself. Other studies of IPCC AR4 projections suggest the SAM will continue to strengthen throughout this century (e.g., Lynch et al. 2006, Fyfe and Saenko 2006). The ERA-40 (“MODEL”) and ERA-40 synthetic trends are both statistically-significantly upward in JJA. Johanson and Fu (submitted) suggest that ERA-40 wintertime tropospheric temperature trends are too large in winter by a factor of about two; thus the statistical significance of the model-based trends is questionable.
In summary, the two station based near-surface temperature reconstructions (the new one presented here and that of Chapman and Walsh in press) correlate strongly for annual and seasonal timescales for 1960-2005, and they agree with the Schneider et al. (2006) stable isotope reconstruction for annual timescales. The new near-surface temperature reconstruction is representative of the entire continent, as indicated by the similar trends and strong correlation between the ERA-40 and “synthetic” ERA-40 datasets. All records correlate significantly with all other records during all seasons from 1982-2001 (not shown). Near-surface temperature trends are statistically insignificant (p>0.05) on annual timescales within every dataset analyzed, for both the longer (1960-2002) and shorter (1982-2001) periods. In recent decades negative temperature trends in DJF and MAM are likely related to a strengthening of the SAM. Likewise, a positive temperature trend in JJA, consistent with recently reported winter tropospheric warming, may be related to the absence of a trend in the SAM during these months. Although the uncertainty may be underestimated, the reconstructed datasets suggest the MAM trends may be statistically significant (p<0.5). In the next section, the new near-surface air temperature reconstruction and that of Schneider et al. (2006) are used to evaluate temperature simulations in GCMs during the 20th century. The snowfall dataset from Monaghan et al. (2006b) is used to assess GCM snowfall simulations during the latter half of the 20th century.

4.4 Evaluation of Antarctic near-surface temperature and snowfall in GCMs

Antarctic climate in the IPCC AR4 GCMs is evaluated in recent studies. To characterize the broad scale atmospheric circulation patterns around Antarctica, Lynch et
al. (2006) applied a self-organizing map technique to a selection of GCMs and the ERA-40 and National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) reanalyses. Lynch et al. (2006) found that some models individually do not simulate the late 20th century circulation adequately, but that the grand ensemble mean provides a reasonable representation of the key features of the Antarctic atmosphere. The models consistently predict increasing cyclones and stronger zonal winds (i.e., an upward trend in the SAM) throughout the 21st century. The IPCC AR4 GCMs are able to reproduce the Peninsula warming that is observed, which is a marked improvement over the IPCC Third Assessment GCMs (IPCC 2001, Vaughan et al. 2001). Chapman and Walsh (in press) compare the spatial patterns of near-surface temperature trends from their reconstruction to an 11-model ensemble of IPCC AR4 GCMs and find that the models are able to simulate observed warming over the Antarctic Peninsula and Ross Ice Shelf for 1958-2002, although warming in the GCMs is generally too strong over the interior of the continent. Chapman and Walsh (2006) conclude that the models may be useful, in a qualitative sense, for assessing the spatial trends of change in the 21st century. The models predict a homogenous warming over Antarctica both annually and in the four seasons. Averaged from 60°-90°S, the models project a temperature increase of 2-3.5 K for the end of the 21st century compared to the end of the 20th century.

The following analysis extends that of Chapman and Walsh (in press) by providing a more detailed and quantitative assessment of the 20th century Antarctic temperature simulations from 5 IPCC AR4 GCM ensembles (Table 4.1). One particularly useful record employed here is a two-century reconstruction of annual Antarctic temperature
(Schneider et al. 2006) which allows the GCMs to be validated over a longer (~120 year) period. Additionally, an assessment of Antarctic snowfall simulations in the GCMs is presented. The sensitivity of Antarctic snowfall to Antarctic and Southern Hemisphere temperature changes, which has important implications for Antarctica’s role in global sea level change during this century, is quantified in the GCMs and compared to the observed sensitivity.

The time series of the 5-y running mean of annual Antarctic temperature anomalies for observations (Schneider et al. 2006) versus GCM ensembles are compared in Fig. 4.7. Each of the GCM ensembles, and especially the grand ensemble, projects a steady increase in temperatures throughout the late 19th and 20th centuries that is larger and less variable than the subtle increase in the observed dataset. The trends are quantified below.

The time series of the 5-y running mean of annual Antarctic snowfall anomalies for observations (Monaghan et al. 2006b) versus GCM ensembles are compared in Fig. 4.8. The grand ensemble shows a steady rise in snowfall throughout the late 19th and 20th centuries, similar to that observed for temperature (Fig. 4.7). The time series for the individual GCM ensembles are more variable than for temperature, although all of them display increases between the beginning and end of the period. Compared to the observations, the GCMs simulate a similar increase in snowfall throughout the latter ~4 decades of the 20th century. It is not possible to assess whether the GCM snowfall continues to increase during the most recent decade, or decreases as seen in the observed dataset. Monaghan et al. (2006b) found that the change in Antarctic snowfall from 1955-2004 is not statistically significant (although an upward trend is evident from 1960-1990).
The annual temperature and snowfall trends for the time-series shown in Figs. 4.7 and 4.8 are presented in Table 4.4 (top) for two time periods: 1882-1997 and 1962-1997. The GCM grand ensemble temperature trends are 3.4 times greater (range=2.3-4.3) than the observed trend (+0.02±0.01 K decade\(^{-1}\)) for the 1882-1997 period. Over the recent 1962-1997 period, the Antarctic increases are 5.9 times greater (range=2.5-10.1) than the observed statistically insignificant trend (0.02±0.10 K decade\(^{-1}\)). The annual GCM time series (Fig. 4.7) are steadily upward throughout the ~120 year period, suggesting that they are responding monotonically to the gradually increasing greenhouse gas forcing. The ensemble mean Antarctic temperature increase is 0.75 K (p<0.05) over the 20\(^{th}\) century (The trend is rounded to 0.08 K decade\(^{-1}\) in Table 4.4), similar to the GCM and observed global average increase of about 0.6 K (IPCC 2001), suggesting that the Antarctic climate system in the GCMs might be coupled more closely with middle- and low-latitudes than observed. The causality of the larger-than-observed Antarctic temperature increases in the GCMs is examined below.

The annual snowfall trends over the 1882-1997 period, for which there is no observational dataset for comparison, increase at the rate of about +10±1.0 mm century\(^{-2}\) (note the snowfall trends are expressed as 10 mm decade\(^{-2}\) in Table 4.4). A 10 mm century\(^{-2}\) increase in Antarctic snowfall is the equivalent of an additional 0.33 mm y\(^{-1}\) of global sea level rise by the end of the 20\(^{th}\) century, compared to the beginning of the century, or about 1/5\(^{th}\) of the average annual sea level rise over the 20\(^{th}\) century of 1.7 mm y\(^{-1}\) (Church and White 2006). Over the recent 1962-1997 period, the observed and simulated GCM snowfall trends are of similar magnitude, about twice the rate of the 1882-1997 snowfall trends (~ +20±5.0 mm century\(^{-1}\)). Thus, in the GCMs, Antarctic
temperature and snowfall trends accelerate in the latter half of the 20th century. It is noteworthy that the observed snowfall trends, when calculated for 1962-2002 (rather than 1962-1997, as done for the comparison with the GCMs), are less than half the 1962-1997 trends, and are not statistically different from zero at the 95% confidence interval. As mentioned in the discussion above, it is not clear whether the GCM snowfall trends would also have decreased had longer time series been available.

In Table 4.4 (bottom), the seasonal temperature trends are shown for the 1962-1999 period. The observed seasonal trends (from the new reconstruction) are slightly different from those shown in Table 4.3 because they are calculated for a different period and they employ 5-year running means rather than annual means. On average, the seasonal trends in the GCMs are stronger than observed, except in SON. The ensemble mean temperature trends are 1.4, 1.3, and 0.9 times larger than the observed positive trends in DJF, JJA, and SON respectively, and in MAM they are all positive whereas the observed trends are negative. In MAM the absolute differences between the GCM and observed trends are largest, thus MAM accounts for the largest contribution to the larger-than-observed annual trends. The winter (JJA) trends are largest in both the GCMs and the observations, suggesting that the GCMs may be able to simulate the seasonality of the trends. The ‘GIS’ model has smaller positive trends than the other models in three-of-four seasons that most closely resemble the observations, including a statistically insignificant negative trend in DJF. In DJF and MAM, when the observed SAM trends are largest (Marshall et al. 2003), the ‘GIS’ and ‘MPI’ models, which have time-variable ozone forcing (Table 4.1) have the smallest trends. Conversely, the ‘CCC’ and ‘MRI’ models, which do not have time-variable ozone forcing, have the largest temperature trends in
MAM. In DJF, ‘CCC’ has the largest trend, and ‘MRI’ has the 3rd largest trend behind ‘NCA’ (which does have time-variable ozone forcing). Aside from ‘NCA’, the different magnitude of trends between the models with and without time-variable ozone forcing is consistent with recent studies suggesting that stratospheric ozone depletion has contributed to the DJF and MAM trends in the SAM (e.g., Cai and Cowan 2007). In turn, the strengthening SAM has had an overall negative feedback on Antarctic temperature trends in DJF and MAM (e.g., Marshall 2007).

The observed and modeled sensitivity of annual Antarctic snowfall to near-surface temperature, based on the detrended 5-y running mean time series, are presented in Fig. 4.9 for three periods: 1882-1997 (GCMs only), 1962-1997, and 1980-1997. The observed sensitivity (“ANT_OBS”) is about +5% K\(^{-1}\) on average, but is statistically significant only for the 1962-1997 period (+3.8±2.9% K\(^{-1}\)). The sensitivity in the GCMs has a broad range (0.5-7.2% for 1962-1999), but the grand ensemble the sensitivity is similar to observed, about 6±2% K\(^{-1}\) on average. The good agreement between the observed and model sensitivity suggests that if the GCMs are able to accurately project Antarctic temperature changes for the end of this century, projections for increased snowfall will be approximately correct. Chapman and Walsh (in press) noted that the range of Antarctic temperature increases in the sresa1b GCM runs is 2-3.5 K by the end of this century, corresponding to a 10-17.5% increase in snowfall (for a 5% snowfall sensitivity). Considering the average annual Antarctic snowfall accumulation rate is about 150 mm y\(^{-1}\) for the grounded ice sheet (Vaughan et al. 1999; note that estimates vary widely), the increase in accumulation rate would be about 15-26 mm y\(^{-1}\), or about 0.5-0.86 mm y\(^{-1}\) global sea level equivalent, by the end of the 21\(^{st}\) century.
A noteworthy observation from Fig. 4.9 is that the snowfall sensitivity relationship remains nearly the same for Antarctic temperature variability regardless in both time periods analyzed, but not for Southern Hemisphere temperature variability. The Southern Hemisphere sensitivity for 1962-1997, when Southern Hemisphere temperature variability is in phase with Antarctic temperature variability for the first ~20 years of the period, is similar to the “ANT_OBS” sensitivity. However, over the more recent period (1980-1997), during which overall annual Antarctic temperatures decrease slightly while Southern Hemisphere temperatures continue to increase, the sensitivity of Antarctic snowfall to Southern Hemisphere temperature changes is about zero. Additionally, the statistical uncertainty for the Southern Hemisphere sensitivity relationship is large for both periods. Thus, even though warming is occurring over a large portion of the moisture source region for Antarctic snowfall -- the Southern Ocean (e.g., Delagoe et al. 1999) -- Antarctic snowfall is sensitive to changes in local temperature. It is imperative that Antarctic temperature changes are well-simulated in GCMs if Antarctic snowfall changes, and their subsequent impact on global sea level, are to be accurately projected.

In summary, the IPCC AR4 GCMs appear to overestimate annual Antarctic near-surface temperature trends by a factor of about 3 over the past century, and by more than 5 times during the latter half of the century. The temperature rises in the GCMs are steady and exhibit generally less interdecadal variability than the observations, suggesting that a systematic feedback directly or indirectly related to the steady rise in greenhouse gases may be causing the problem. The GCMs simulate recent increases in Antarctic snowfall but due to the termination of the GCM time series in 1999 uncertainty remains as to whether they would have simulated a decrease in snowfall since the mid-
1990s. The GCMs can accurately simulate the sensitivity of Antarctic snowfall to near-surface temperature, suggesting that if the projected temperature increases of 2-3.5 K over Antarctica for end of the 21st century are reasonable estimates, that the corresponding simulated increases in snowfall of about 10-20% will be approximately correct. In the next section the possible causes of the overly-strong temperature increases in the GCMs are discussed, followed by conclusions.

4.5 Discussion and Conclusions

4.5.1 Toward diagnosing why the Antarctic temperature changes in the GCMs are too strong in the 20th century

As implied from the observed Antarctic temperature variability examined in this study, and discussed widely in the literature (e.g., Thompson and Solomon 2000, Schneider and Steig 2002), the SAM is an important modulator of Antarctic temperature, altering the large-scale atmospheric circulation (e.g., van den Broeke et al. 2004). Thus, comparing the observed SAM trends to those in the GCMs may offer evidence of why the temperature trends are too strong in the GCMs. Scatter plots of annual Antarctic temperature anomalies versus anomalies of the SAM are shown in Fig. 4.10, for the observations and the ensembles of each of the five GCMs. The SAM in the GCMs is calculated from the difference of the normalized monthly zonal mean sea level pressure anomalies at 40° and 65° S, following Gong and Wang (1999) and Marshall (2003); the observed SAM is that from Marshall (2003), available at http://www.nerc-bas.ac.uk/icd/gjma/sam.html. The observed air temperature is negatively correlated with the SAM, consistent with literature that suggests approximately 90% of Antarctica (the
portion that is not on or near the Antarctic Peninsula) has a cooling response to an upward trend in the SAM (e.g., Schneider and Steig 2002, van den Broeke and van Lipzig 2004b), as has been observed in DJF, MAM, and annually over the past several decades (Marshall 2007). All five of the GCMs are positively correlated with the SAM, indicating that they either have the wrong response to trends, or that other factors are overwhelming the influence of the SAM.

Aside from the SAM, increasing or decreasing downward longwave radiation at the surface will also impact Antarctic near-surface temperature variability (e.g., Hines et al. 1999). Longwave radiation in turn can be modulated by increasing/decreasing anthropogenic greenhouse gases, water vapor, or clouds (IPCC 2001). As a condition of the IPCC AR4 20c3m runs, the greenhouse gases in the 19th and 20th centuries increase at the rate observed (e.g., IPCC 2007), so it is already known that the greenhouse gases are positively correlated with the Antarctic temperature increases. The simulated changes in water vapor and clouds in the GCMs are investigated here to determine if water vapor and clouds are influencing the amplified Antarctic temperature rises in the models.

Figure 4.11 shows scatter plots of Antarctic near-surface temperature anomalies versus all-sky (clear + cloudy; Fig. 4.11a) and cloudy-sky (Fig. 4.11b) downward longwave radiation at the surface, as well as versus precipitable water vapor in the atmospheric column (Fig. 4.11c), for the grand ensemble of the five GCMs, for the period 1882-1997. A robust positive correlation exists between temperature and downward all-sky longwave radiation ($r^2=0.99$), indicating that changes in the radiative forcing are likely to be the cause of the strong changes in Antarctic temperature in the GCMs throughout the 20th century. The longwave radiation increase does not appear to be related so much to
changes in clouds, indicated by the much weaker correlation between temperature and cloudy-sky longwave radiation in Fig. 4.11b ($r^2=0.26$), but rather to increases in water vapor, as suggested by the strong relationship between temperature and precipitable water vapor in Fig. 4.11c ($r^2=0.98$). Unfortunately, there are no observations available to determine whether water vapor over Antarctica has actually increased over the past century. However, Monaghan et al. (2006b) did not find a statistically significant change in snowfall over the latter half of the century, which suggests that precipitable water may not have changed much over the past 50 years. In summary, the monotonic, overly-vigorous near-surface warming that occurs over Antarctica in the GCMs appears to be related to an increase in water vapor. A detailed analysis is reserved for a follow-on study, but the results presented here suggest an overly active greenhouse gas – temperature – water vapor feedback loop.

4.5.2 Conclusions

At annual timescales, three long-term continentally representative Antarctic near-surface temperature reconstructions are in good agreement (the one presented here, that of Chapman and Walsh in press, and that of Schneider et al. 2006). The two reconstructions that have sub-annual resolution (the one presented here and that of Chapman and Walsh) are also in excellent agreement at seasonal timescales. That the variability and trends are similar, although different techniques and data were employed to create them, lends confidence that the temperature reconstructions are accurate. The close agreement between the annual Antarctic near-surface temperatures from ERA-40 compared to a “synthetic” ERA-40 dataset constructed from the technique used for the
new reconstruction indicates that the new reconstruction is representative of the entire continent, even though relatively few (fifteen) records were used to create it. Near-surface temperature trends are statistically insignificant on annual and seasonal timescales within every dataset analyzed, for both the longer (1960-2002) and shorter (1982-2001) periods analyzed. Although their statistical significance is low, it is noteworthy that negative temperature trends in DJF and MAM in recent decades are consistent with a strengthening of the SAM in those seasons. Likewise, a positive temperature trend in JJA, consistent with recently reported winter tropospheric warming, may be related to the absence of a strengthening of the SAM during the winter months.

The five IPCC AR4 GCMs analyzed here overestimate annual Antarctic near-surface temperature trends by a factor of ~3 over the past century, and by more than 5 times during the latter half of the century. The GCM trends in DJF, MAM, and JJA all contribute to the amplified annual trends, especially MAM. The GCM snowfall variability compares reasonably to the observations for the period of overlap, for which there are generally upward trends. However the GCM runs end in the late 1990s, while the snowfall sharply decreases during the past decade-or-so, so it cannot be concluded that the models would have captured this downward fluctuation. The GCMs are able to accurately simulate the sensitivity of Antarctic snowfall to near-surface temperature, suggesting that if Antarctic near-surface temperature does increase as predicted by the GCMs for the end of the 21st century (2-3.5 K according to Chapman and Walsh in press), snowfall will also increase by about 10-20%. However, substantial improvements to simulations of Antarctic near-surface temperature in the GCMs will be required in
order to increase confidence in snowfall predictions, and the consequences for global sea level.

The causality of the amplified Antarctic temperature increases in the GCMs is investigated and it is found that an increase in column precipitable water vapor appears to be the main reason for an increase in downward longwave radiation incident at the surface. In turn the downward longwave radiation is closely correlated with the temperature increases. The problem may be related to a greenhouse gas – temperature – water vapor feedback loop that is too strong in the models. Within the GCMs, the role of the SAM in the temperature increases is still unclear. However, if the SAM is having an important impact, it is doing so indirectly because the sign of the Antarctic temperature trends (upward) is the opposite of what is expected over the continent as a whole for an increasing SAM scenario (e.g., Marshall 2007). Additional work is necessary to elucidate the cause of the strong water vapor feedback in the GCMs, and the role of the SAM in modulating the simulated temperature variability.
4.6 Tables and Figures

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*resolution of atmospheric component  
**A=atmosphere; O=ocean; S=sea ice; L=land

Table 4.1. Description of the GCMs used in this study.
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<th>Elevation (m)</th>
<th>Type</th>
<th>Country</th>
<th>Duration</th>
<th>n</th>
<th>$\sigma_{\text{RECON}}/\sigma_{\text{READER}}$</th>
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<td>1960-1979</td>
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*Harry Station data is suspicious after 1998; data used in analysis is thru 1998.

Table 4.2. Description of the independent READER temperature observations used to validate the reconstruction. The number of monthly means available for each station is denoted by “n”. Monthly means were computed if >90% of possible observations were available for a given month. The ratios of the reconstructed-to-observed standard deviations (of the monthly temperature anomalies) are shown in the last column.
Table 4.3. Temporal trends and 95% confidence intervals of average annual and seasonal Antarctic near-surface air temperature (K decade⁻¹) from various datasets, for two time periods. Trends different from zero (p<0.5) are underlined. Schneider et al. (2006) data are annual only, and end in 1999. The short-term trends are calculated from 1982 because that is when the “COMISO” data begin.
Table 4.4. (top): Annual trends and 95% confidence intervals for the 5-y running mean of the near surface temperature anomalies (“TEMP”; K decade$^{-1}$) and snowfall anomalies (“SNOW”; mm decade$^{-2}$) for two periods: 1882-1997 and 1962-1997. (bottom): Seasonal trends and 95% confidence intervals for the 5-y running means of the near surface temperature anomalies (K y$^{-1}$). Notations follow those described in Fig. 4.7. The observational datasets are as follows: annual temperature, 1882-1997 – Schneider et al. (2006); annual and seasonal temperature, 1962-1997 – the new reconstruction from this study; annual snowfall, 1962-1997 -- Monaghan et al. (2006b). Trends different from zero (p<0.5) are underlined.
Fig. 4.1. The composite map of the maximum absolute value of the Pearson’s correlation coefficient ($|r|$) resulting from correlating the ERA-40 1980-2001 annual temperature change (with respect to the 1980-2001 mean) for the grid box containing each of the 15 observation sites with every other 1°x1° grid box over Antarctica (i.e., this map is a composite of 15 maps). Pink/red colors have correlations at approximately p<0.05. Stations: 1=Faraday/Vernadsky; 2=Bellingshausen; 3=Orcadas; 4=Halley; 5=Novolarevskaja; 6=Syowa; 7=Mawson; 8=Davis; 9=Mirny; 10=Casey; 11=Dumont D’Urville; 12=Vostok; 13=Scott; 14=Amundsen-Scott South Pole; 15=Byrd.
Fig. 4.2. Various time series of Byrd Station annual near-surface temperature records (°C), as described in the text. The record used in the new Antarctic temperature reconstruction is a combination of “Byrd_Station” and “Byrd_Shuman”, with any missing years filled in with the using the regression relationship, “Byrd_Regress”. Time series for monthly data were constructed in a similar manner. Correlation statistics are provided in the text.
Fig. 4.3. The standard error (SE; gray; units=K) and correlation (r; white) between the observed and reconstructed temperature anomalies for a) all available observations for all 16 independent stations placed into monthly and annual (“Y”) bins; and b) all available observations for all months at each of the 16 stations. The stations are listed in Table 4.2. Confidence intervals (p<0.05) for the correlations are indicated by the error bars.
Fig. 4.4. The temporal trends for the observed (READER=gray) and reconstructed (RECON=white) temperature anomalies for (a) all available observations for all 16 independent stations placed into monthly and annual (“Y”) bins (K y⁻¹); and (b) the annual average of all available monthly and annual observations from all 16 stations (unitless). The time series used to construct the graph in “b)” were normalized before averaging in order to avoid spurious trends due to variations in the annual amount of observations available. The stations are listed in Table 4.2. Confidence intervals (p<0.05) for the trends are indicated by the error bars.
Fig. 4.5. Annual Antarctic near surface temperature (K) anomalies (with respect to the 1980-1999 mean**) for various datasets for a) 1950-2005 and b) 1980-2005. Notations are as follows: “ann_recon_nobyrd”= the reconstruction with Byrd Station record omitted; “ann_recon”= the reconstruction based on all 15 records; “ann_synth”= the reconstruction using ‘synthetic’ temperature records extracted from the 15 ERA-40 grid points that correspond to the observation sites; “ann_model”=the ERA-40 near-surface temperature; “ann_chapman”=the reconstruction of Chapman and Walsh (in press); “ann_comiso”=the AVHRR skin temperature, similar to that published in Kwok and Comiso (2003); “ann_schnd”=the reconstruction of Schneider et al. (2006) based on stable isotopes. **The ‘ann_comiso” data begins in 1982, thus the anomalies are with respect to the 1982-1999 mean. The value of the ann_comiso temperature anomaly in 1988 is 2.78 K.
Fig. 4.6. Similar to Fig. 4.5, but for the 1950-2005 a) DJF, b) MAM, c) JJA, and d) SON Antarctic near-surface temperature (K) for the new reconstruction, the Chapman and Walsh (in press) reconstruction, and (for 1980-onward) the AVHRR skin temperature dataset (e.g., Kwok and Comiso 2003). Notations are the same as in Fig. 4.5.
Figure 4.7. Time series of 5-y running mean of annual Antarctic near-surface temperature anomalies (K) from 1882-1997 for the observations (“obs”), the grand ensemble of the 5 model ensemble means (“gra”), and the 5 GCM ensembles (“ccc”, “gis”, “mpi”, “mri”, and “nca”). Table 4.1 describes each of the GCMs. The Schneider et al. (2006) time series is used for the observational dataset. The temperature anomalies are with respect to the 1980-1999 mean prior to smoothing.
Figure 4.8. Time series of 5-y running mean of annual Antarctic snowfall anomalies (mm) from 1882-1997 for the observations and GCMs, with notations following those in Fig. 4.7. The observational dataset is that of Monaghan et al. (2006b), and spans 1957-2002. The snowfall anomalies are with respect to the 1980-1999 mean prior to smoothing.
Figure 4.9. Sensitivity of annual anomalies of Antarctic snowfall to annual anomalies of near-surface temperature. Notations are as follows: “ANT_OBS” is the sensitivity of the observed snowfall (Monaghan et al. 2006b) to the mean of the three observational temperature reconstructions (Schneider et al. 2006, Chapman and Walsh in press, and the one presented here); “S_HEM” is the sensitivity of the observed snowfall to Southern Hemisphere temperatures from the Hadley Centre Climatic Research Center (HAD-CRU) record (Brohan et al. 2006); and all other categories are for the grand and individual ensemble GCM annual snowfall versus corresponding GCM annual temperature. The numbers are expressed as a percentage of the mean 1980-1999 snowfall. Confidence intervals (p<0.05) are indicated by the error bars.
Figure 4.10. Five-year running means of annual Antarctic near-surface temperature anomalies (K) plotted against annual SAM anomalies (units are standard deviations) for the observations (a; 1962-2003) and the GCMs (b-f; 1882-1997). The observations are based on the new temperature reconstruction versus the Marshall (2003) SAM. All anomalies are with respect to the 1980-1999 means.
Figure 4.11. Five-year running means of annual Antarctic near-surface temperature anomalies (K) versus anomalies of a) all-sky downward surface longwave radiation (W m\(^{-2}\)), b) cloudy-sky downward surface longwave radiation (W m\(^{-2}\)), and c) precipitable water vapor (mm). All of the data are based on the grand ensemble mean, which is the average of the ensemble means of all 5 GCMs. All anomalies are with respect to the 1980-1999 base period.
CHAPTER 5

CONCLUSIONS

5.1 Overview

Through this dissertation research, a more thorough understanding of Antarctic snowfall and temperature variability over the past \( \frac{1}{2} \) century has been achieved. Additionally, an innovative statistical technique has been developed to overcome the temporal constraints of numerical models, which unfortunately do not simulate Antarctic climate with confidence prior to the modern satellite era. With this technique, the length of the existing Antarctic snowfall record has been doubled from 25 to 50 years. It worked well enough to warrant its use for a similar reconstruction of near-surface Antarctic temperature for a 46-year period. Employing these two records, the most comprehensive evaluation to date of Antarctic snowfall and near-surface temperature has been performed for 20\textsuperscript{th} century global climate model (GCM) control scenarios, run in conjunction with the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4), due out in 2007. The key findings are summarized below.
5.2 Key Findings

Chapter 2: Polar MM5, a model optimized for use in polar regions, was used for the first time to evaluate recent snowfall trends over Antarctica. Two sets of simulations, each driven by a different reanalysis, yield similar spatial patterns of annual and seasonal snowfall trends over Antarctica. The similarity between the trends for the two runs, as well as their agreement with observations, lends confidence that the results are robust.

The results indicate that there is substantial spatial and temporal variability of Antarctic snowfall. The annual variability and trends over Antarctica are the small residual of larger seasonal variability and trends that may be related to recent climate change, particularly seasonally varying trends in the Southern Annular Mode. Though considerable changes are happening regionally, averaged over Antarctica, the annual trend in Antarctic snowfall from 1985-2001 is close to zero. The model and observational results do not support the significant upward accumulation trend over East Antarctica for 1992-2003 suggested by Davis et al. (2005). Due to strong interannual variability, this short (1985-2001) model record of Antarctic snowfall is not ideal for understanding longer-term changes. The goal of Chapter 3 is to establish a longer record.

Chapter 3: A statistical technique of synthesizing observational records with model reanalysis is used to construct a coherent record of Antarctic-wide snowfall accumulation variability extending back prior to the modern satellite era, to 1955. The results are consistent with the 1985-2001 results presented in Chapter 2, indicating that there has been no statistically significant change in net Antarctic snowfall over the past 50 years, inferring that global sea level rise has not been mitigated by recently increased Antarctic
snowfall. This is somewhat contrary to what might be expected as climate warms over the Southern Ocean surrounding Antarctica (e.g., Levinson 2005), which is the moisture source for Antarctic snowfall. Secondary messages arise from these results:

1. Interannual and interdecadal snowfall variability must be more seriously considered when assessing the rapid ice volume changes that are occurring over Antarctica. Satellite-based techniques show great promise for precisely measuring Antarctic ice mass changes, but this research suggests the records are still too short.

2. Antarctic snowfall is not currently compensating for the oceanic-induced melting at the ice sheet periphery.

This new 50-year Antarctic snowfall record is used to assess the IPCC AR4 GCM simulations, the main topic of Chapter 4.

Chapter 4: The statistical technique devised and used to extend the Antarctic snowfall record in Chapter 3 is used to construct a new Antarctic near-surface temperature dataset spanning 1960-2005. It is compared with other observationally-based Antarctic temperature datasets for the past ~45 years and all are in good agreement at annual and seasonal timescales. It is found that near-surface temperature trends are statistically insignificant on annual and seasonal timescales within every dataset analyzed, for both the longer (1960-2002) and shorter (1982-2001) periods. Even though they are not statistically significant, it is noteworthy that negative temperature trends in DJF and
MAM in recent decades are consistent with a strengthening of the SAM in those seasons. Likewise, a positive temperature trend in JJA, consistent with recently reported winter tropospheric warming, may be related to the absence of a strengthening of the SAM during the winter months.

The Antarctic near-surface temperature and snowfall reconstructions are employed, along with an additional >100 year water-stable isotope reconstruction of Antarctic near-surface temperature (Schneider et al. 2006), to evaluate Antarctic climate in the IPCC AR4 GCMs. The GCMs overestimate annual Antarctic near-surface temperature trends on average by a factor of 3 over the past century, and by more than 5 times during the latter half of the century. The GCM snowfall variability compares reasonably to the observations over the period of overlap, for which there are generally upward trends. However most of the GCM runs terminate in the late 1990s, while the snowfall sharply decreases during the past decade-or-so, so it cannot be concluded that the models would have captured this downward fluctuation. The GCMs are able to accurately simulate the sensitivity of Antarctic snowfall to near-surface temperature, suggesting that if Antarctic near-surface temperature does increase as predicted by the GCMs for the end of the 21st century (2-3.5 K according to Chapman and Walsh in press), snowfall will also increase by about 10-20%. However, substantial improvements to simulations of Antarctic near-surface temperature in the GCMs will be required in order to increase confidence in snowfall predictions, and their consequences on global sea level.

The causality of the overly-vigorous Antarctic temperature increases in the GCMs is investigated and it is found that an increase in column precipitable water vapor appears to be the main reason for an increase in downward longwave radiation incident at the
surface. In turn the downward longwave radiation is closely correlated with the temperature increases. The problem may be related to a greenhouse gas – temperature – water vapor feedback loop that is too strong in the models. Within the GCMs, the role of the SAM in the temperature increases is still unclear, but appears to be secondary compared to the water vapor feedback.

5.3 Recommendations for future research

Numerical modeling shows great promise as a technique for simulating Antarctic climate variability, especially as improvements in model physics and computational efficiency emerge. The main limitation is mainly due to data sparseness prior to the modern satellite era, as well as data discontinuities due to the occasional introduction of new data sources with time. To some extent, this problem cannot be addressed over Antarctica. However, it is incumbent upon the Antarctic meteorological community to make the most complete and highest quality datasets available to the national modeling centers for use in future global reanalyses, as these reanalyses will surely be employed as the initial conditions for future regional re-analyses over Antarctica, which are already in the planning stages. Statistical reconstructions employing the newest objective analysis techniques should be attempted over the Antarctic, as they have the potential to provide large-scale, homogenous fields to ingest into numerical models.

Extending on the recommendation of greater use of statistical reconstruction techniques, it is recommended that attempts be made to create upper air data sets from such methods; for example, 500-hPa geopotential height, temperature, and winds. The likelihood of success is promising considering the larger space scales that are represented
by upper air data compared to surface data, providing a greater probability of accurate results given a handful of observational data to perform the reconstruction. Such records will help us to understand how climate is changing over important regions that are otherwise void of data, such as the West Antarctic Ice Sheet.

The tremendous importance of having a diverse array of earth-observing satellites was underscored in Chapter 2 (satellites have led to great improvements in model simulations) and Chapter 3 (satellites are the most promising means of measuring ice sheet changes). In particular, newer satellites that precisely measure surface elevation or gravity are on the verge of establishing records of sufficient length to assess how the earth’s major ice sheets are changing. The continuation of such programs for observing and modeling polar climate is strongly endorsed.

Finally, the results of Chapter 4 clearly indicate that more attention must be paid to simulations of Antarctic climate in GCMs. The results show that the role of Antarctic snowfall in 21st century global sea level rise cannot begin to be understood without first improving simulations of Antarctic temperature. The problems are related to more than just the depiction of the Southern Hemisphere Annular Mode within the models, which is often implicated when GCMs perform poorly in the Antarctic. To improve the GCMs it will be necessary to understand how greenhouse gases, water vapor, and temperature interact over Antarctica.
APPENDIX

SUPPORTING MATERIALS FOR CHAPTER 3

A.1 Overview

Figure A.1 is a schematic of the technique employed to reconstruct the past 50 years of Antarctic snowfall accumulation (SA). For the reader who desires a more technical treatment of the methodology, a detailed section is provided at the end of this document entitled “Calculation of reconstructed SA”. To reconstruct SA for the most recent two decades (1985-2004), the ERA-40 precipitation fields are used, calibrating them to remove biases based on measured SA from overlapping ice core records. Below is a comparison of the ERA-40 snowfall data to ice core and snow stake records, adapted from Monaghan et al. (2006a), which shows that ERA-40 reliably reproduces interannual variability and trends in Antarctic SA. To reconstruct SA for the former 3 decades (1955-1984) a technique similar to ordinary kriging (Cressie 1993) is applied to the ice core records. The kriging-like technique is a method of spatial interpolation that predicts unknown values at desired locations from data at known locations. In this manner, a spatially homogenous gridded data set can be created from ice core SA records. The great advantage of this technique is that the 20-y of ERA-40 gridded precipitation fields can be used (as discussed later and in Chapter 1, precipitation, primarily in the form of
snowfall, dominates Antarctic accumulation) to create a robust relationship to describe the spatial variability of SA changes, rather than relying on assumptions about the spatial autocorrelation of snowfall over Antarctica to calculate a variogram, as would be otherwise be required. Table A.1 summarizes the basin-scale results from the reconstruction.

A.2 Ice core and other observational data

Figure A.2 and Table A.2 summarize the observational records used in this study. Most are ice cores, but there are also snow pits from Vostok, a precipitation day record at Faraday/Vernadsky, and a composite of ice core and snow stake measurements at South Pole. The locations of each are shown in Fig. 3.1 of Chapter 3, indicated by their number. These records are of varying lengths, but most go back to the mid-1950s and extend through the mid-1990s. Based on this limitation, a 4-decade period from 1955-1994 is chosen for which to analyze the records. They are averaged spatially if several exist in a given region (Table A.2). A few of the records are missing years on either end. These include LGB00 and MGA in region #3, which start in 1957, so the first 2 years are missing. Core LGB16 in region #4 ends in 1992, so the final 2 years are missing. Core LGB70 in region #5 is missing the final year, 1994. In order to apply the methodology consistently, each gap was filled in using the average of the available years within its respective decadal bin (i.e., 1955-1964, 1965-1974, 1975-1984, 1985-1994). The impact of the missing data is expected to be small as they represent <1% of the total observations. All other records spanned the entire 40-y period, with one notable exception. Cores GD03, GD06, and GD15 in region #8 only span 30-y, 1955-1984.
However, it was undesirable to omit these valuable data from a region that is otherwise void of SA records. A comparison between the composite of these three cores and other nearby East Antarctic records was made, and a strong correlation was found with the composite of the records from regions #6 and #7 and #10 for the 3-year running mean of percentage SA change (r=0.73; Fig. A.3a), a relationship that explains over 50% of the variance. The final 10 years (1985-1994) of the stack for region #8 was then reconstructed by multiplying the 30-y mean SA for region #8 by 1 +/- the fraction of the SA deviation of #6-#7-#10 from its 30-y mean for each year of 1985-1994 (Fig. A.3b).

There are three records that use observations exclusive of, or in addition to, ice cores: Vostok, Faraday/Vernadsky, and South Pole. A stack of 8 snow pit records are employed at Vostok. Ekaykin et al. (Ekaykin et al. 2004) compare this record with a more recent snow stake record for the 1970-1998 period and find good agreement. This suggests that the data are reliable for use in this study.

A precipitation day record is used for Faraday/Vernadsky, described in detail by Turner et al. (2005a). The authors present a ~50-y record of precipitation days (1951-1999) and use model data to show that precipitation days are directly proportional to the amount of precipitation occurring at Faraday with high confidence. This linear relationship infers that for a given percentage change in precipitation days (with respect to some baseline period) there is an identical percentage change in precipitation amount, the metric desired for the snowfall reconstruction. This makes the record – which is the only known contemporary observational record for Antarctic Peninsula snowfall variability that spans the entire 1955-1994 period - suitable for use in this study. Although it is not necessary, for consistency the precipitation day measurements are
converted to annual precipitation values based on the linear relationship described by Turner et al. (2005a) by using the only published measurement of annual precipitation at Faraday/Vernadsky (1045 mm y\(^{-1}\) in 1994), by the same authors. Although more than one year of data is desirable in order to account for long term variability, it is of little consequence to this study.

Multiple records are employed to reconstruct the SA record at South Pole (Fig. A.4). Mosley-Thompson et al. (1995, 1999) provide a comprehensive evaluation of South Pole SA. The following records shown in Fig. A.4 are from their work: BETA-1 (10-y average of 11 ice cores), BETA-2 (20-y average of 11 ice cores), BETA-3 (10-y average of six ice cores), EMT-FARM (annual averages from a 235 stake network), SP-Pentagon (7-y average from a 42-pole stake network), OSU-1978 (12-y average from a 23-stake farm), and 7-mile cross (8-y average from a 140 stake farm). They provide convincing evidence based on these records that a SA increase occurred at South Pole between the 1950s and 1990s and the reader is referred to those publications for a comprehensive description. Some proxy of annual variability was desired for this study in order to apply the method uniformly (although the ultimately concern is with decadal means); this could not be achieved by using the results of Mosley-Thompson et al. (1995, 1999) alone. Two other annual time series were also introduced for this purpose. This first is the annual SA from an ice core drilled by the U.S. ITASE team (Meyerson et al. 2003, Kaspari et al. 2004). This record is smoothed by applying a 3-y running mean (“ITASE_3-y” in Fig. A.3a). The record has suspiciously high SA in the early 1960s, and anomalously low SA in the late 1970’s compared to the other records, but it is still included as it gives an estimate of interannual variability and it cannot be proven to be grossly erroneous. The
second is an annual SA record from a 50-stake network near South Pole that was installed in the 1950s, although it has only been monitored continuously since 1978 ("SP-50-stake” in Fig. A.4a). The data from this network are archived by the University of Wisconsin Madison as monthly changes in snow depth averaged over the grid. The annual SA for the 50-stake network was calculated by compiling the monthly data and assuming the density of snow is 0.3 kg m\(^{-3}\). Further discussion of this stake network, and a comparison to the ITASE core, is given by others (Genthon et al. 2005). All of the records in Fig. A.4a are averaged for each year, yielding the “AVG_ALL” record which is employed in this study. Fig. A.4b shows that when decadally averaged, “AVG_ALL” is in good agreement with the results of Mosley-Thompson et al. (1995, 1999), and thus it is a valid proxy of South Pole SA over the 1955-1994 period.

A.3 The ERA-40 precipitation data

The European Centre for Medium-Range Weather Forecasts (ECMWF) 40-year reanalysis (ERA-40; Uppala et al. 2005) provides the simulated precipitation fields used in this study. ERA-40 spans the period September 1957-August 2002. This dataset is chosen based on its relatively good skill over Antarctica compared to similar reanalyses (Bromwich and Fogt 2004), and because it can be extended through 2004 by appending the ECMWF operational analysis, which employs a similar data assimilation system. Thus, the final 28 months of the precipitation record, from September 2002-December 2004 are provided by the ECMWF operational analysis. The monthly precipitation fields for this period are adjusted for the average monthly bias between ERA-40 and the ECMWF operational analysis for a 3 year period of overlap (September 1999-August
2002). This reduces the possibility that artificial trends will be introduced due to differences in the amount and/or type of data that was assimilated into either.

The large-scale precipitation fields from ERA-40 are interpolated from an N80 reduced Gaussian grid with T159 spectral truncation (approximately 1.125°), to a 1° latitude x 1° longitude grid. The precipitation fields from the ECMWF operational analysis are interpolated from an N80 Gaussian grid (also approximately 1.125°) to the same 1°x1° grid. Henceforth, the combined dataset will be referred to as ERA-40. It is noteworthy that at the time of this writing the popular 2.5°x2.5° monthly precipitation product available from ECWMF greatly underestimates precipitation over the portions of the continent that receive little annual snowfall. This is apparently due to a truncation error in the algorithm used to interpolate the data from its native resolution to the 2.5° grid. It is not clear if this problem may have affected the study of Davis et al. (2005), but they did note stronger negative biases over the interior of the continent than were found here.

The entire analysis was performed using the ERA-40 precipitation fields, and again using the precipitation-minus-evaporation (P-E) fields, as proxy of SA. Precipitation, rather than P-E, was chosen for the final results presented in this paper because using P-E caused unrealistically low mean SA in several of the basins, especially in the Transantarctic Mountains along the Victoria Land coast; it was decided that this may cause impractical estimates of the GSL contribution of these basins. Despite this problem it is noteworthy that using P-E yields a nearly identical result in terms of percentage change in each individual basin, and averaged overall. The use of precipitation (primarily in the form of snowfall) is consistent with previous studies indicate it as the
dominant term in the Antarctic surface mass balance (Bromwich 1988; see also Chapter
1). Other SA terms are typically small compared to precipitation at the basin-and-larger
scales considered in this work and recent studies indicate there are no statistically
significant trends in sublimation, melt, or blowing snow divergence over Antarctica since
1980 (van de Berg et al. 2005, van den Broeke et al. 2006b). As discussed below, the
ERA-40 precipitation field is adjusted for its bias with the SA observations in order to
ensure a quantitatively accurate proxy of SA, and thus implicitly accounts for other SA
components. Henceforth, and in the paper, it is referred to as ERA-40 SA.

The period from 1985 onward, rather than the entire ERA-40 period, is chosen for this
analysis because the ERA-40 mean sea level pressure, 2-m temperature, and 500-hPa
géopotential height are of questionable quality prior to the modern satellite era, about
1979 (Bromwich and Fogt 2004). Van de Berg et al. (2005) find similar results for ERA-
40 precipitation – their results indicate a large jump in mean annual precipitation of ~30
mm in about 1979, with the magnitude of precipitation thereafter matching other
estimates more closely. Monaghan et al. (2006a) note that the quality of ERA-40
compares best with the observational record, as well as other reanalyses, from 1985
onward. Similarly, Turner et al. (2005a) note that there is good agreement between the
number of precipitation days observed at Faraday/Vernadsky station on the Antarctic
Peninsula compared to the number of precipitation days simulated by ERA-40 from 1984
onwards. They attribute the marginal skill prior to 1984 to a lack of humidity data being
assimilated into the analysis system in ERA-40.

Figure A.5 is adapted from Monaghan et al. (2006a) and gives a comparison of the
ERA-40 precipitation to the observational records used in their study, most of which are
employed here. The records shown in Fig. A.5 are described in Table A.3, also from Monaghan et al. (2006a). They averaged the first-five years (1985-1989) and last-five years (varies depending on the record) of each record in order to smooth the noise in the ice cores. There is a tendency to simulate lower-than-observed SA by ERA-40 precipitation, especially over the interior of the ice sheet (Fig. A.5a). The biases are compensated in this study by making an adjustment for the biases between ERA-40 precipitation and the observed records (the method is discussed below). Most importantly, the temporal changes are correctly simulated at 9/10 sites (Fig. A.5b); this indicates that ERA-40 is sufficiently capturing the SA trends over the continent. The correlation coefficients between the simulated and observed records (Fig. A.5c) indicate that the modeled precipitation and observed annual SA variability (as opposed to the 5-y averaging applied to smooth the results for Figs. A.5a and A.3b) share common variance different from zero at the 90% confidence interval at 8/10 of the sites, despite the small-scale perturbations (SSPs) due to sastrugi, etc. in the observational records (van der Veen and Bolzan 1999) that are not present in model fields. Considering the differences between simulated precipitation and observed SA signals (i.e., SSPs are present in the observed signal), and that temporal smoothing is not used, the relationships in Fig. A.5c are quite robust. This result indicates that ERA-40 reasonably reproduces the observed interannual variability of SA. In summary, the ERA-40 precipitation fields are representative of the mean SA, temporal changes, and interannual variability over Antarctica and are suitable for use in this study.
A.4 Sensitivity Studies

A.4.1 Sensitivity of results to methodology

To examine the robustness of the reconstructed SA to the methodology, the continental-scale SA was reconstructed using two additional, simple techniques. The first technique (“AVG_R”) employs the mean of the percentage SA change from the 16 observational records, with each record weighted by the total area of grid boxes having a percentage SA change that correlates most closely with the percentage SA change in the grid box where the observation is located (inferred from the 20-y interannual ERA-40 SA). The second technique (“AVG_NN”) is similar to the first, but each observational record is instead weighted by the total area of grid boxes for which it is the nearest observation site. The results for both techniques yield reconstructions for the grounded ice sheet that are nearly identical to the results from the methodology (Fig. A.6). This provides confidence that the reconstruction is robust and insensitive to the methodology used.

A.4.2 Sensitivity of results to data availability

Due to the relatively small number of observational sites employed in this study (sixteen), it is necessary to test the sensitivity of the results if some of the observations are removed. This was done by running four experiments. In each experiment, four observational sites were randomly removed (each one only once), and then the SA was recalculated using the methodology. The sites removed from each experiment were: RAND1 (2,8,10,11); RAND2 (5,6,7,14); RAND3 (1,3,4,12); and RAND4 (9,13,15,16). Fig. A.7 compares the four experiments to the control SA (CTRL; that which is presented
in Ch. 3). In all cases over WAIS, EAIS, and the entire grounded ice sheet, the sensitivity of the reconstructed SA to the removal of ¼ of the observations is small – less than 2 percent in all instances but one. The impact of these differences on the 50-y trends or the change of the final decade with respect to the 50-y mean is small. Therefore, the methodology is considered insensitive to the removal of data.

To examine if the 16 observational sites are representative of accumulation over the entire continent, the methodology was used to reconstruct a 20-y (1985-2004) record of SA based on the percentage SA change from 16 “synthetic cores”. The synthetic cores were extracted from the ERA-40 SA record from the grid boxes nearest each of the 16 observational sites. The reconstructed interannual accumulation variability matches the actual ERA-40 SA over the grounded ice sheet with remarkable skill (r=0.90), providing confidence that these 16 sites are representative of accumulation variability over the entire grounded ice sheet.

**A.4.3 Other Sensitivity Tests**

Two other sensitivity tests were performed to check the robustness of the results. One was to change the predictor (equation 5 below) from \( r^2 \) to \( r^3 \), which, as expected, amplified the interannual/interdecadal variability in the basins for the prior 3 decades, making them appear more like the latter 2 decades. The end results were nearly identical; the decadal SA using \( r^3 \) was within 2% of the original results using \( r^2 \). The trends over EAIS, WAIS, and the entire grounded ice sheet were also nearly identical. The predictor chosen was \( r^2 \) because it has physical meaning (\( r^2 \) is the variance), and because the spatial
results using $r^2$ were more homogeneous than those using $r^3$, which in some cases led to unrealistic patterns of change over small space scales.

The other test was to first average the percent change in the 16 observational records for each decade and apply the kriging-like technique (rather than apply the kriging-like technique to the annual change first, and then averaging the results by decade). The results for this test were also nearly identical to the results presented in Ch. 3. Ultimately, the method of first applying the kriging-like technique to the annual data and then averaging by decade was chosen, because it allows a more physically-based method of determining uncertainty.

**A.5 Calculation of reconstructed SA**

This section describes the technical details of how the SA over Antarctica for the past 5-decades is reconstructed by blending ice core records with simulated precipitation from ERA-40. The generalized kriging technique (Cressie 1993) is specified as:

$$\hat{Z}(i, j) = \sum_{k=1}^{n} \lambda_{i,j,k} \times Z_k \quad ; \quad \sum_{k=1}^{n} \lambda_{i,j,k} = 1$$

[1]

where $\hat{Z}(i, j)$ is the predicted value of a quantity at a desired grid point with coordinates $(i,j)$, $n$ is the number of observations (i.e., ice cores), $Z_k$ is the known quantity at the observation site, and $\lambda_{i,j,k}$ is a predictor (weighting coefficient) that must sum to 1. The predictor, $\lambda_{i,j,k}$, is calculated from a variogram. A variogram is a function that is derived from the relationships between pairs of observations and describes the similarity between a measurement and a desired grid point based on the distance between the points. In this application it is desired to avoid using the distance between two points to assess their
relationship, as there may be a strong covariance between SA variability at two points that are very far apart due to teleconnections that arise from the time-mean large scale atmospheric circulation. Instead, the predictor can be more directly and accurately computed by exploiting the information about spatial variability provided by the 20-y of gridded precipitation data from ERA-40. Even though this may not be considered ‘kriging’ in the strictest sense, the final method of interpolating the data is the same (i.e., equation [1] is employed, only the weighting function is not calculated from a variogram); thus, the technique is referred to as “kriging-like”. Details are given below.

Henceforth, “core site” will mean the site of an individual record (regardless of whether it is from a core, snow pit, or precipitation day observation). A considerable amount of noise can be present in ice cores due to SSPs caused by local topography, which can have a substantial effect on annual SA at a given site and mask the “true” SA signal for the region as a whole (Frezotti et al. 2005, Mosley-Thompson et al. 2001). Spatial and temporal averaging can be applied to reduce SSP noise. First, if several cores exist in a given region (Table A.2), these are averaged spatially:

\[ B_c = \frac{\sum_{i=1}^{n_c} B_{i_c}}{n_{cr}} \]  

[2]

where \( B_c \) is the annual core SA averaged in space (\( n_c \) = number of cores in region), and \( B_{i_c} \) is the annual SA at a given core site, in mm y\(^{-1}\). Henceforth, “core region” will refer to one of the 16 regions given in Table A.2, after averaging the cores in that area. Now a baseline decade can be defined for which all other decades can be compared to assess how SA has changed. It is convenient to choose 1985-1994 because the period provides
a 10-y overlap with the modeled precipitation record from ERA-40, which covers 1985-2004. Later, this will allow the ERA-40 precipitation to be calibrated against the core SA and adjusted for biases. It will also allow the ERA-40 fields from the most recent decade, 1995-2004, to be compared to the prior 4 decades, extending the final record to 50-years. The difference in core SA (expressed as a percentage of the baseline SA) for the 3 decades prior to 1985-1994, \( \Delta_{c_{i}, o_{0}} \), is given by

\[
\Delta_{c} = \frac{B_{c_{i}} - \overline{B}_{c_{o}}}{\overline{B}_{c_{o}}} \times 100
\]  

[3]

where the subscript ‘i’ refers to each year of the 3 decades from 1955-1984, and the subscript ‘o’ refers to 1985-1994, the baseline (the bar over \( \overline{B}_{c_{o}} \) infers that it is the 10-y mean baseline SA). The percentage, rather than the absolute difference, is used so that the kriging technique (eqn. [1]) can be consistently applied for all regions, considering that annual SA spatially varies by about 2 orders of magnitude (from about 20-2000 mm y\(^{-1}\)) over the continent (see Fig. 3.2a in Chapter 3).

Next, the predictor, \( \lambda_{i,j,k} \), can be calculated from the 20-y (1985-2004) precipitation record from ERA-40. Similar to the ice cores, the baseline simulated ERA-40 precipitation for 1985-1994 is calculated. Then, the percentage change at each grid point for each of the 20 individual years is calculated with respect to the 10-y baseline. Finally, the annual percentage precipitation change for the grid point corresponding to each observation site is correlated with that at every other grid point for the 20-y model record, creating a series of maps that indicate how representative the percentage precipitation change at each observation site is with respect to all other grid points over Antarctica.
Figure 3.1 in Chapter 3 summarizes this by showing a composite of the maximum correlation/anticorrelation at each grid point, calculated as

\[ r_{G_{i,j},\text{max}} = \max \left| r_{G_{i,j},k} \right| ; \ k=1,n_r \]  \hspace{1cm} \text{[4]} \]

where \( r_{G_{i,j},k} \) is the Pearson’s correlation coefficient between percentage change at any grid point and the core grid point, and \( n_r \) is the number of core regions. The variance, \( r^2 \), gives more information than a variogram regarding the relationship between the observations and the interpolated grid points, and thus is the basis of the predictor, \( \lambda_{i,j,k} \):

\[ \lambda_{i,j,k} = \frac{r^2_{G_{i,j},k}}{\sum_{k=1}^{n_r} r^2_{G_{i,j},k}} \]  \hspace{1cm} \text{[5].} \]

Finally, the kriging-like technique (eqn. [1]) can be applied to interpolate percent change for each core region to the entire grid as follows:

\[ \Delta_{R_{i,j}} = \sum_{k=1}^{n_r} \lambda_{i,j,k} \times \Delta_{\varepsilon_k} \times \eta_{i,j,k} \hspace{1cm} ; \hspace{1cm} \eta_{i,j,k} = \frac{r_{G_{i,j},k}}{r_{G_{i,j},\text{max}}} \]  \hspace{1cm} \text{[6]} \]

where \( \Delta_{R_{i,j}} \) is the annual SA change at each grid point with respect to the baseline period for each year during the 3 decades prior to the baseline, expressed as a percentage. The operator \( \eta_{i,j,k} \) accounts for the sign of anticorrelations (it is assumed that if a core region is anticorrelated with a grid point that the relationship is just as likely to be valid as a positive correlation since it too is likely to arise due to the atmospheric circulation). It is noteworthy that no threshold is applied to eliminate weak correlations. This is due to the use of \( r^2 \) as the predictor, which makes weak correlations negligible in most cases, except in the few regions where all cores are weakly correlated with a given grid point (see Fig.
3.1, Ch. 3). In these instances, the calculated uncertainty (discussed below) will be larger-than-normal and thus the weak correlations are accounted for. As noted above, another advantage of using $r^2$ as the predictor is that it allows points that are far apart but may have a strong covariance (due to teleconnections) to be weighted accordingly, whereas weighting by distance does not; this is an important consideration considering the relatively few observation sites available.

Before reconstructing the continent-wide SA for prior decades based on the change calculated in equation [6] and the ERA-40 precipitation, the ERA-40 record must be calibrated to account for the bias between the SA for the core regions and the simulated precipitation for the 10-y overlap (baseline) period, 1985-1994. The bias for each core region, $\delta_c$, is expressed as a percentage of the ERA-40 precipitation and is given by

$$
\delta_c = \frac{\overline{B}_{mc} - \overline{B}_c}{\overline{B}_{mc}} \times 100
$$

[7]

where $\overline{B}_{mc}$ is the average ERA-40 precipitation for 1985-1994 at the core site, and $\overline{B}_c$ is the SA for the corresponding core region calculated in equation [2] temporally averaged over the same 10-y period. Next, the bias for each core region can be kriged to the entire grid using equation [1], but in this case the bias is assumed to be a function of distance since SA biases tend to be regional in nature (Fig. 2.3 of Chapter 2). Therefore, “true” ordinary kriging is employed by fitting an optimum variogram to the data at each observation point and then calculating the interpolated grid. This was done on an x-y grid in the Surfer (v 8.0) software and the resulting grid was interpolated to the same 1°x1° grid the SA data is on. The methodology of generating and optimizing a variogram is discussed by Cressie (1993). The linear model has been fitted to the experimental
variogram and optimized it by accounting for anisotropy at 45° intervals and assuming a small nugget effect to smooth the results. This variogram was employed by the kriging procedure and the resulting bias at each grid point, \( \delta_{g,i,j} \), is expressed as a percentage of the ERA-40 precipitation. A map of the biases is shown in Fig. A.8, expressed as a percentage of the observed precipitation. As discussed above (Fig. A.5a), ERA-40 tends to underestimate SA over much of the ice sheet. Errors due to kriging the bias correction are neglected, as there is still considerable uncertainty in the absolute value of Antarctic surface mass balance (Monaghan et al. (2006a) note a range of 119 mm y\(^{-1}\) – 197 mm y\(^{-1}\) averaged over the continent). Rather, the technique is intended to bring the SA closer to a realistic value than it would be without a correction.

Finally, the calibrated baseline (1985-1994) SA, \( B_{g0,i,j} \), for every grid point can be calculated:

\[
B_{g0,i,j} = B_{m,i,j} \times \left(1 - \frac{\delta_{g,i,j}}{100}\right)
\]  

where \( B_{m,i,j} \) is the 10-y averaged baseline ERA-40 precipitation at each grid point. The SA for each year of the most recent decade, \( B_{g,i,j} \) (1995-2004) is calibrated in an identical manner to equation [8], assuming that the bias that applies to the 1985-1994 period is the same for the 1995-2004 period. Then the percent SA change at each grid point for each year of the 1995-2004 decade is calculated as:

\[
\Delta_{g,i,j} = \frac{B_{g,i,j} - B_{g0,i,j}}{B_{g0,i,j}} \times 100
\]
where $B_{g,i,j}$ is the adjusted precipitation for each of the 10-y during 1995-2004. Note that the resulting expression, $\Delta_{g,i,j}$ is identical to the resulting expression in equation [6], but [6] was used to calculate the percentage change for the 3 decades prior to the baseline. Now the percent SA change with respect to the baseline for each of the 40 non-baseline years is known. Before reconstructing the SA the percentage change at each grid point is averaged temporally to further reduce noise from the small-scale perturbations in the cores, and to reduce the effects of measurement uncertainty (van der Veen and Bolzan 1999). Temporal averaging of ice core SA records is recommended by most authors (Genthon et al. 2005, Frezotti et al. 2005, Mosley-Thompson et al. 2001). Decadal averaging is employed based on their findings and recommendations:

$$\Delta g = \frac{\sum_{y=1}^{10} \Delta g_{i,j,y}}{10} \quad [10]$$

where $\overline{\Delta}_{g,i,j}$ is the 10-y average percentage annual SA change for each decade, and $y_1$ and $y_{10}$ are the first and last years of 4 decadal bins (1955-1964, 1965-1974, 1975-1984, 1985-1994).

The decadal average annual SA, $\overline{B}_{g,i,j}$ can now be calculated for each of the 4 decades with respect to the baseline decade:

$$\overline{B}_{g,i,j} = \overline{B}_{g0,i,j} \times \left(1 + \frac{\overline{\Delta}_{g,i,j}}{100}\right) \quad [11].$$
Now the 5-decade average of the reconstructed annual SA, $\overline{g}^{50}_{g_{i,j}}$, can be calculated (Fig. 3.2a, Ch. 3) and then applied to yield the percentage change by decade again, now with respect to the 50-y period instead of the 10-y baseline:

$$\Delta g^{50}_{g_{i,j}} = \frac{\overline{g}_{g_{i,j}} - \overline{g}^{50}_{g_{i,j}}}{\overline{g}^{50}_{g_{i,j}}} \times 100$$  \[12\]

where $\Delta g^{50}_{g_{i,j}}$ is the percent SA change with respect to the 5-decade mean at a grid point. This equation is applied to each decade and the resulting maps are shown in Figs. 3.2b-f in Chapter 3.

The regional SA averages over West Antarctica, East Antarctica, and the entire continent (Fig. 3.3 in Chapter 3), and the basin averages (Table A.1) were calculated by first spatially averaging the annual SA for each grid point (from eqns. [8] and [11]) within the desired area (weighted by the area of each grid box to account for effect of latitude on the size of the grid boxes), and then temporally averaging for each decade. The boundaries of the basins and the ice sheet grounding lines were provided by David Vaughan (based on Vaughan et al. 1999). They were interpolated from a polar stereographic grid to the 1x1 degree grid used here.

Calculation of Uncertainty

The numerical uncertainty in an ice core is due to interannual variability, SSPs (i.e., sastrugi, surface roughness), large-scale perturbations (dunes), changes in density, and measurement error due to the techniques used to date the annual layers (van der Veen and Bolzan 1999). The standard deviation of the mean of the percentage change of the annual
SA at a grid point for each of the 5 decades, $\overline{\sigma}_{\Delta_{\text{SA}}}$, can be determined for each of the 5 decades as follows:

$$\overline{\sigma}_{\Delta_{\text{SA}}} = \frac{\sigma_{\Delta_{\text{SA}}}}{\sqrt{10}}$$  \[13\]

where $\sigma_{\Delta_{\text{SA}}}$ is the standard deviation of $\Delta_{\text{SA}}$ (calculated in eqns. [6] and [9]), and 10 is the number of years in a decade. This technique assumes zero autocorrelation.

In this application, the primary concern is basin- and continent-wide SA changes, and calculating the uncertainty for entire regions is desired, rather than for a single grid point as in equation [13]. At the basin-scale, [13] would become

$$\overline{\sigma}_{\text{var}} = \sqrt{\frac{\sum_{k=1}^{10} (\Delta_{b_k} - \overline{\Delta}_{b_k})^2}{\frac{10}{\sqrt{10}}}}$$  \[14\]

where the numerator is the standard deviation of a 10-year time series of the basin-averaged percentage change with respect to the baseline period. The bar over the second delta term indicates that it is the 10-year basin average change for a given decade. The denominator ($\sqrt{10}$) accounts for the fact that decadal, rather than annual means of SA are being expressed. Thus, $\overline{\sigma}_{\text{var}}$ can be considered the “standard deviation of the mean”, and is unique for each of the 5 decades considered, and the “var” subscript indicates that this uncertainty applies to temporal SA variability.

For the first 3 decades, the uncertainty due to the kriging-like technique (eqn. [6]) can be approximated by comparing the calculated percentage change of the ERA-40 SA for each year of the 1995-2004 decade using equation [9], to that using [6]. As opposed to
the change calculated from [6], the change calculated from [9] is an “exact” technique, since the precipitation is known at each grid point and does not have to be interpolated. To apply [6] in this instance, the ERA-40 percent change for each year of 1995-2004 for the grid points corresponding to the 16 core regions is used (this is analogous to $\Delta_c$ in [6]). Then the difference between the “true” and “kriged” simulated SA change ($\Delta_b$ and $\Delta_{be}$, respectively) for 1995-2004 is taken at each grid point, and the error for a basin can be estimated in the same manner as equation [14]:

$$
\sigma_{b\_krig} = \sqrt{\frac{\sum_{k=1}^{10} \left( (\Delta_b - \Delta_{be})_k - (\Delta_b - \Delta_{be}) \right)^2}{10 - 1}} / \sqrt{10}
$$

[15]

where $\sigma_{b\_krig}$ is the standard deviation of the mean of the difference between equations [6] and [9] for each basin for the 1995-2004 decade. The bar over the second term indicates that it is the 10-y mean difference. Assuming this is a representative estimate of uncertainty for any of the 5 decades, this can be applied to estimate the uncertainty due to the extrapolation technique for the first 3 decades. The total uncertainty, $\sigma_{b\_tot}$ at a grid point is then estimated by:

$$
\sigma_{b\_tot} = \sqrt{\sigma_{b\_var}^2 + \sigma_{b\_krig}^2}; \quad \{1955-1964, 1965-1974, 1975-1984\}
$$

$$
\sigma_{b\_tot} = \sqrt{\sigma_{b\_var}^2}; \quad \{1985-1994, 1995-2004\}
$$

[16]

This value is expressed as a percentage of the mean baseline SA for each basin, and thus is easily converted to a measurable value in millimeters of SA (WEQ) as shown by the error bars in Fig. 3.3 in Chapter 3, which are $\pm 1\sigma$ calculated from equation [16].
To calculate the statistical significance for the SA change for 1995-2004 versus the 50-y mean (last two columns of Table A.1), first $\sigma_{b_{tot}}$ is pooled for each of the 5 decades to calculate the standard deviation of the mean, $\sigma_{b_{50y}}$, for the 5-decade mean SA:

$$\sigma_{b_{50y}} = \sqrt{\frac{\sum_{i=1}^{5} \sigma_{b_{tot}}^2}{5^2}} \quad [17].$$

Then the result of [17] is pooled with the result if [16] for the 1995-2004 decade to calculate the standard deviation of the mean for the difference between 1995-2004 versus the 50-y mean (Panofsky and Brier 1968):

$$\sigma_{b_{diff}} = \sqrt{\frac{1 \times \sigma_{b_{tot}}^2 + 5 \times \sigma_{b_{50y}}^2 \left(1 + \frac{1}{5}\right)}{(1 + 5 - 2)^2 \left(1 + \frac{1}{5}\right)}} \quad [18]$$

where $\sigma_{b_{tot}}^2$ is the standard deviation of the mean for the 1995-2004 decade. To meet the 95% confidence interval, the difference between the 50-y mean and the 1995-2004 mean has to be approximately $2\sigma$ according to the Student’s t test.

Finally, the uncertainty for the 50-y trend for each basin in Table A.1 (the 3rd column from the right), was calculated by applying a two-tailed Student’s t-test to check the goodness-of-fit of the each trend at the 95% confidence interval. The trend in each basin is the slope of the linear regression drawn through the 5 decadal means for each basin. This method neglects the uncertainty in the means themselves, this source of error is not evaluated since statistical significance is only achieved in two cases regardless.
### A.6 Tables and Figures

<table>
<thead>
<tr>
<th>Basin</th>
<th>Area ((10^3 \text{ km}^2))</th>
<th>Mean (WEQ)</th>
<th>5-decade Trend</th>
<th>1995-2004 minus 1955-2004 change</th>
<th>GSL contribution past decade w.r.t. 50-y mean</th>
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<td>((\text{mm y}^2))</td>
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<td>-5.0</td>
<td>0.01</td>
</tr>
<tr>
<td>DD'</td>
<td>687</td>
<td>214 +/- 10</td>
<td>0.02</td>
<td>-6.2</td>
<td>0.02</td>
</tr>
<tr>
<td>DD&quot;</td>
<td>173</td>
<td>287 +/- 14</td>
<td>-0.30</td>
<td>-2.7</td>
<td>0.01</td>
</tr>
<tr>
<td>D&quot;E</td>
<td>279</td>
<td>74 +/- 3</td>
<td>0.02</td>
<td>0.8</td>
<td>0.00</td>
</tr>
<tr>
<td>EE'</td>
<td>1620</td>
<td>38 +/- 1</td>
<td>0.07</td>
<td>3.5</td>
<td>-0.01</td>
</tr>
<tr>
<td>EAIS</td>
<td>9392</td>
<td>138 +/- 1</td>
<td>0.07</td>
<td>3.2</td>
<td>-0.01</td>
</tr>
<tr>
<td>GIS</td>
<td>12427</td>
<td>182 +/- 3</td>
<td>0.04</td>
<td>-33.1</td>
<td>0.09</td>
</tr>
</tbody>
</table>

Table A.1. SA results by basin. In the last 3 columns, bold (underlined) values are statistically significant at \(p<0.05\) \((p<0.1)\). Basins (i.e., E’E’’) are shown in Fig. 3.1 in Chapter 3.
<table>
<thead>
<tr>
<th>Core #</th>
<th>Source of SA Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Core S100 (Kaczmarska et al. 2004)</td>
</tr>
<tr>
<td>2</td>
<td>Stack of 13 firn cores from Dronning Maud Land (Oerter et al. 2000, Graf et al. 2002)</td>
</tr>
<tr>
<td>4</td>
<td>Western Lambert basin inland core LGB16 (Jiawen et al. 1999, Xiao et al. 2001)</td>
</tr>
<tr>
<td>5</td>
<td>Eastern Lambert basin coastal core LGB70 (Goodwin, unpublished)</td>
</tr>
<tr>
<td>6</td>
<td>Stack of eastern Lambert Basin inland cores DT085 (Xiao et al. 2001) and DT001 (Wen et al. 2001)</td>
</tr>
<tr>
<td>8</td>
<td>Stack of Wilkes Land cores GD03, GD06, and GD15 (Goodwin 1995, Goodwin 1990, Morgan et al. 1991). 1985-94 are reconstructed as described in the text</td>
</tr>
<tr>
<td>9</td>
<td>Talos Dome core (Stenni et al. 2002)</td>
</tr>
<tr>
<td>10</td>
<td>Stack of 8 snow pits at Vostok (Ekaykin et al. 2004)</td>
</tr>
<tr>
<td>12</td>
<td>Siple Dome core (Nereson et al. 1996)</td>
</tr>
<tr>
<td>13</td>
<td>Stack of West Antarctic firn cores 99-1, 00-1, 00-4, 00-5, RIDS-A, RIDS-B, and RIDS-C (Kaspari et al. 2004)</td>
</tr>
<tr>
<td>14</td>
<td>Stack of West Antarctic firn cores 01-2, 01-3, and 01-5 (Kaspari et al. 2004)</td>
</tr>
<tr>
<td>15</td>
<td>Berkner Island South Dome core (Mulvaney et al. 2002)</td>
</tr>
<tr>
<td>16</td>
<td>Faraday/Vernadsky precipitation day observations (Turner et al. 2005a)</td>
</tr>
</tbody>
</table>

Table A.2. Observational records used in this study. The numbers correspond to Fig. 3.1 in Chapter 3.
<table>
<thead>
<tr>
<th>Source of Accumulation Observation</th>
<th>Abbreviated Name in Fig. A.5</th>
<th>Approximate Location (see Fig. 3.1)</th>
<th>Length of Record</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average of eastern Lambert Basin cores DT085 (Xiao et al. 2001) and DT001 (Wen et al. 2001)</td>
<td>ELAM</td>
<td>~72° S, ~77.5° E</td>
<td>1985-1996</td>
</tr>
<tr>
<td>Average of 13 firn cores from Dronning Maud Land (Oerter et al. 2000)</td>
<td>DML</td>
<td>~75° S, ~0° E</td>
<td>1985-1996</td>
</tr>
<tr>
<td>Vostok stake network (Ekaykin et al. 2004)</td>
<td>VOS</td>
<td>78.5° S, 106.9° E</td>
<td>1985-2001</td>
</tr>
<tr>
<td>Compiled from various records at South Pole (Mosley-Thompson et al. 1995, 1999, Meyerson et al. 2003, Kaspari et al. 2004). Same record used in this study.</td>
<td>SPOL</td>
<td>90.0° S, 0.0° E</td>
<td>1985-2001</td>
</tr>
<tr>
<td>Siple Dome core (Nereson et al. 1996)</td>
<td>SDOM</td>
<td>81.7° S, 149.0° E</td>
<td>1985-1994</td>
</tr>
<tr>
<td>Average from West Antarctic firn cores 00-1, 00-4, and 00-5 (Kaspari et al. 2004)</td>
<td>WA00</td>
<td>~79° S, ~115° W</td>
<td>1985-2000</td>
</tr>
<tr>
<td>Average from West Antarctic firn cores 01-2, 01-3, 01-5, and 01-6 (Kaspari et al. 2004)</td>
<td>WA01</td>
<td>~78° S, ~95° W</td>
<td>1985-1999</td>
</tr>
<tr>
<td>Average from West Antarctic firn cores RIDS-A, RIDS-B, and RIDS-C (Kaspari et al. 2004)</td>
<td>RIDS</td>
<td>~79° S, ~118° W</td>
<td>1985-1995</td>
</tr>
</tbody>
</table>

Table A.3. Ice core and snow stake records from Monaghan et al. (2006a) and shown in Fig. A.5
Figure A.1. Schematic of technique employed to reconstruct 50-y record of Antarctic SA.
Fig. A.2. a) The 3-y running mean of the percentage change of the observed SA with respect to the 1985-1994 mean at each of the 16 sites for the period 1955-1994. A description of each observation site is given in Table A.2.
Fig. A.3. a) 3-y running mean of SA change between the East Antarctic cores (regions #6-#7-#10) and GD0X for the 30-y period 1955-1984. b) The 40-y (1955-1994) SA record showing the last 10 years of GD0X reconstructed based on the East Antarctic cores as described in the text.
Fig. A.4. Comparison of a) annual and b) decadal averages of SA from various records at South Pole, as described in the text. The “AVG_ALL” record is the average of all records for each year and is the record used in the SA reconstruction.
Fig. A.5. Comparison of annual SA records from ice cores and stake networks to simulated precipitation (snowfall) from ERA-40 (adapted from Monaghan et al. 2006a). The simulated P is taken from the nearest gridpoint to the location of the observation. Where possible, ice core records have been averaged together by region to reduce noise. The regions are defined in Table A.3. (a) Mean accumulation (or snowfall from ERA-40) from 1985 to the end of the record (period varies depending on the length of the observational record—see Table A.3). (b) Accumulation change between first and last five years of each record. (c) Correlation coefficient of observed versus simulated annual SA from 1985 to the end of the record (dashed lines indicate the correlation is significant from zero at the 90% confidence interval).
Fig. A.6. Reconstructed decadal mean annual accumulation (mm y\(^{-1}\)) for the grounded ice sheet from three methods. The thick black line (AVG\_STUDY) is from the methodology used in this study. The dotted line (AVG\_R) and the dash-dot line (AVG\_NN) are from the alternate techniques described in the text.
Fig. A.7. Difference (% of CTRL SA) of the 4 experiments in which 4 observation sites were randomly removed and the SA was re-calculated using the methodology employed in the Ch. 3. The plots are for EAIS (top), WAIS (middle) and the entire grounded ice sheet (bottom). There are no differences in the last two decades because ERA-40 SA is used from 1985-onward.
Fig. A.8. Map showing the bias between ERA-40 annual snowfall minus observed annual SA for 1985-1994, expressed as a percentage of the observed SA. The bias is kriged from each core region to the grid as described in the text. The actual biases (expressed as percentage) are indicated by the numbers next to the black circles.


Goodwin, I.D., 1990: Snow accumulation and surface topography in the katabatic zone of eastern Wilkes Land Antarctic Science, 2, 235-242


Goodwin, I.D., unpublished Antarctic CRC report.


Landine, P. G. & Gray, D. M. 1989 Snow transport and management. Internal Report, Division of Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan. 80 pp.


