Atmospheric Change in Antarctica since the 1957–1958 International Geophysical Year

Dissertation

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By

Julien Pierre Nicolas, M.S.

Graduate Program in Atmospheric Sciences

The Ohio State University

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Dissertation Committee:
David H. Bromwich, Advisor
Jay S. Hobgood
Jeffrey C. Rogers
Jialin Lin
Abstract

The Antarctic Ice Sheet holds a volume of ice and snow equivalent to 55 meters of sea level. The melting of only a relatively small fraction of this volume could have dramatic consequences for populations around the world. With this in mind, the research presented here focuses on two atmospheric variables that are key controls of the state of the ice sheet: its surface mass balance (or net snowfall) and its near-surface air temperature. The analysis aims to understand how these two parameters have changed—if at all—since the 1957–1958 International Geophysical Year (IGY), the start of the instrumental era in Antarctica. Particular attention is given to the part of the continent known as West Antarctica, the most vulnerable to atmospheric and oceanic warming, and the one where rapid glacial change is currently taking place. The research is divided into three parts.

The first part uses a set of seven global reanalyses to investigate the changes in Antarctic surface mass balance and Southern Ocean precipitation since 1979 (start of the reanalyses). This investigation is also intended to shed light on the reliability of these reanalyses, which often contained artifacts caused by changes in the observing system, particularly in high southern latitudes. Spurious changes in precipitation are found to various degrees in all data sets but with varying characteristics and origins. According to the two reanalyses deemed most reliable, neither Antarctic surface mass balance nor Southern Ocean precipitation have changed significantly over the past three decades.
The second part consists of a multifaceted investigation of the near-surface temperature record from Byrd Station, in central West Antarctica. As the only meteorological record in this region to extend back to the IGY, it is a critical data set, but also one with a complicated history and substantial data gaps. A comprehensive revision of the record is undertaken and a novel approach is used to estimate the missing observations. The complete Byrd record reveals a marked increase in the annual mean temperature since the late 1950s. This warming is not only stronger than previously estimated by other studies, but also establishes central West Antarctica as one of the fastest-warming regions on Earth. A review of the atmospheric and oceanic drivers of the temperature trends highlights their strong seasonal dependence and the complex interplay between low-latitude sea surface temperature forcing and high-latitude atmospheric variability.

The third and final part of the research builds upon the new Byrd record and the records from 14 other stations to generate an Antarctic-wide temperature reconstruction spanning the IGY to the present time. The spatial interpolation method is adapted from, and improves upon, a kriging technique previously employed for the same purpose. The reconstruction is then used to re-examine the relationship between the Southern Annular Mode (the dominant mode of high southern latitude atmospheric variability) and Antarctic temperatures. The analysis shows how the strengthening of the SAM in austral summer and fall seen in recent decades has mitigated an otherwise stronger background warming of Antarctica.
Dedication

This is dedicated to my family.
Acknowledgments

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Vita

November 22, 1979 ..................... Born, Nancy, France

2006 ................................... B.S. Physics, University of Grenoble, France

2008 ................................... M.S. Atmospheric Sciences and Oceanography,
University of Toulouse, France

2008 to present ........................ Graduate Research Associate,
The Ohio State University

Publications


**Fields of Study**

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B.3 Monthly mean near-surface temperature anomalies at the 15 locations identified in Fig. 4.2d (page 121) and based on the following data sets: our reconstruction (RECON; black), ERA-Interim forecast and analysis 2-meter temperature ($T_{2m}$) (red and orange, respectively), MERRA $T_{2m}$ (purple), and CFSR $T_{2m}$ (blue). The time series are smoothed with a 36-month moving average and the anomalies are calculated with respect to 1980–1983. Note that the temperature range along the vertical axis is identical in all plots except for “Amery Ice Shelf.”
Chapter 1: General Introduction

From a meteorological standpoint, Antarctica is a place of superlatives. It is the coldest, driest, windiest continent on Earth. The shrinking of its ice sheet in response to atmospheric and ocean warming, and its resulting impact on global eustatic sea level rise, is also one of the greatest challenges posed to human societies in the near and more distant future. Understanding Antarctic climate change in recent decades is a prerequisite for anticipating its future trajectory. The goal of this dissertation is to contribute to this understanding by focusing on two atmospheric variables that are key controls of the state of the ice sheet: the surface mass balance (net annual snowfall), and the near-surface air temperature.

1.1 Antarctic Surface Mass Balance and Global Reanalyses

The contribution of Antarctica to sea level is determined by its mass balance, i.e., the difference between the mass gained primarily through snowfall at the surface, and the mass lost at the continent’s margins through the calving of icebergs and the formation of floating ice shelves. Atmospheric modeling is currently the main approach used to assess the spatio-temporal variability of the surface mass balance (mass input at the surface). This dissertation will be mostly concerned with one particular class of atmospheric models: global atmospheric reanalyses. These reanalyses are produced by combining the fields from a numerical weather prediction model with past observations in order to “reconstruct” the
state of the atmosphere globally and at fine temporal resolution. Because of the observa-
tional constraint, it is not uncommon to see these reanalyses referred to as “observations”
in the literature. On the one hand, global reanalyses have proved extremely valuable for
investigating the climate of such data-sparse regions as high southern latitudes. On the
other hand, the data sparsity also has reduced—sometimes dramatically—their reliability
in the very same regions, particularly with regard to multi-decadal trends in precipitation,
temperature, etc. Regional atmospheric models, with finer spatial resolutions and better
representation of the physical processes relevant to the polar environment, are often pre-
ferred over global reanalyses to study the spatio-temporal variability of Antarctic surface
mass balance. However, these regional models are forced to their lateral boundaries by
global reanalyses, and therefore issues present in the “parent” reanalysis can make their
way into the regional model.

1.2 Antarctic Near-Surface Temperatures

The near-surface air temperature is not only a fundamental indicator of the long-term
changes of the Antarctic atmosphere; it is also one of the parameters that can—once it
reaches the melting point—directly affect the ice sheet’s mass balance. Indeed, numerous
studies from Antarctica and Greenland have shown that the effect of surface melting is
not limited to the loss of meltwater through runoff, but that, by lubricating the base of
the ice or contributing to the fracturing of coastal ice shelves, it can also significantly
impact the ice dynamics. Current knowledge of Antarctic temperature trends relies on
the meteorological records from the small number of research stations scattered around
the continent. Most of these stations were established around the 1957–1958 International
Geophysical Year, often considered the start of the instrumental era for Antarctica. The
rapid warming of the Antarctic Peninsula and little temperature change in East Antarctica. One of the most obvious shortcomings of this station network is the vast data void left in West Antarctica, that is, the very part of Antarctica most vulnerable to climate warming. Until the late 2000s, little was known about the temperature trends in West Antarctica. It is only recently that temperature reconstruction efforts have tried to address this important issue. Yet, the differences and contradictions in their results have left major sources of uncertainty unresolved.

1.3 Outline of the Dissertation

The dissertation is divided into three main chapters, each one consisting of a compilation of manuscripts that have been published in the peer-reviewed literature or—in case of Chapter 4—has been submitted to a scientific journal and is currently in review. Each chapter begins with an introduction describing its context, reviewing the relevant literature, and stating its specific motivations and goals.

Chapter 2 provides an assessment of the changes in Antarctic surface mass balance and Southern Ocean precipitation since 1979 (start of most current global reanalyses). This assessment was initially motivated by the release of a new generation of reanalyses between 2008 and 2013. The chapter reviews the strengths and weaknesses of these data sets and, in so doing, provides background information useful for the two following chapters.

Chapter 3 revisits the temperature record from Byrd Station. As the single meteorological archive from West Antarctica going back to the International Geophysical Year, it

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Whether first or second author, my contribution to these papers is no less than the contribution expected of a doctoral candidate to his/her dissertation. I have been personally deeply involved in the writing of all of the papers and in the research presented in them. My advisor, Dr. David Bromwich, has provided guidance on the research and reviewed the manuscripts. Dr. Matthew Lazzara, Ms. Linda Keller, and Mr. George Weidner have been responsible for the maintenance of the Byrd AWS, and for the collection, correction, and distribution of its temperature readings, which I use in Chapter 3. Dr. Andrew Mongahan is acknowledged for laying the groundwork of various aspects of my research during his own Ph.D. and for providing helpful advice on many occasions.
is a critical data set, but also a challenging one because of its extensive data gaps. The chapter describes the various steps necessary to obtain a complete 50-plus-year record for Byrd and what it revealed about the long-term temperature changes over a vast portion of West Antarctica.

Chapter 4 sets out to reconstruct Antarctic near-surface temperatures from 1958 onward through spatial interpolation of a small number of Antarctic meteorological records (including the new Byrd record obtained in Chapter 3). The reconstruction sheds light on the temperature variability and trends over vast, unsampled areas of Antarctica. It is then used to further understand the influence of the Southern Annular Mode (leading mode of extratropical atmospheric variability in the Southern Hemisphere) on Antarctic temperature changes.

Chapter 5 provides a general conclusion, with a summary of the key findings and future research questions.

2.1 Introduction

2.1.1 Climatic Role of High Southern Latitude Precipitation

Over the last decade, there has been increasing evidence of a positive contribution of the Antarctic Ice Sheet to global sea level rise (Shepherd and Wingham, 2007; Allison et al., 2009; Shepherd et al., 2012). The corresponding ice mass loss has been mainly driven by enhanced ice discharge into the ocean from West Antarctica and the Antarctic Peninsula (Rignot et al., 2008; Pritchard et al., 2009) (see map in Fig. 2.1). There is however great uncertainty as to how the net snowfall accumulation at ice sheet’s surface, a.k.a. surface mass balance (SMB), has been responding to recent climate change. This uncertainty is the combined result of sparse in situ measurements (Eisen et al., 2008) and large spatial and temporal variability (e.g., Kaspari et al., 2004; Frezzotti et al., 2005; Monaghan and Bromwich, 2008). Previous studies have reported insignificant trends in the overall Antarctic SMB since the late 1950s (Monaghan et al., 2006b) as well as since the early 1980s (Van de Berg et al., 2005; Monaghan et al., 2006a). With the equivalent of 6 mm global mean sea

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level falling each year as snowfall on the ice sheet (Church et al., 2001), the evolution of the
Antarctic SMB remains an important determinant of present and future sea level changes.

Owing to extremely low atmospheric moisture content over the continent, Antarctic
snowfall—the main component of the SMB—largely results from moisture advection from
the surrounding ocean areas (Tietäväinen and Vihma, 2008; Sodemann and Stohl, 2009).
Thus, variations in Antarctic SMB are intimately linked to changes in Southern Ocean
precipitation, both of which are examined in this chapter. The assessment of precipitation
trends over the Southern Ocean is further motivated by the determinant impact of its
freshwater budget on the stability of the water column, which in turn affects the extent of
the seasonal sea-ice cover, the meridional overturning circulation, and the ocean’s ability
to absorb CO₂ (e.g., Le Quéré et al., 2007; Böning et al., 2008; Bintanja et al., 2013). The
Southern Ocean is already known to be marginally stable (Martinson, 1990), which implies
that relatively small changes in the amount of annual precipitation could have far-reaching
climatic consequences.

2.1.2 Global Reanalyses: Strengths, Weaknesses, and Recent Developments

Global atmospheric reanalyses potentially constitute valuable resources for investigating
climate change during recent decades. These data sets are produced with a weather
forecasting model anchored with a variety of meteorological observations. Four new global
reanalyses have been produced since 2008: the European Centre for Medium-Range Weather
Forecasts (ECMWF) ‘Interim’ reanalysis (ERA-Interim) (Dee et al., 2011); the National
Aeronautics and Space Administration (NASA) Modern Era Retrospective-analysis for Re-
search and Applications (MERRA) (Rienecker et al., 2011); the National Centers for En-
vironmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) (Saha et al.,
2010); and very recently (late 2013), the Japan Meteorological Agency (JMA) 55-year Re-
analysis (JRA-55) (Ebita et al., 2011). These four data sets are intended to improve upon, and extend, earlier-generation reanalyses: the NCEP-Department of Energy Atmospheric Model Intercomparison Project 2 reanalysis (NCEP-DOE AMIP-2, or NCEP-2) (Kana-
mitsu et al., 2002); ECMWF’s 40-Year Reanalysis (ERA-40) (Uppala et al., 2005); and JMA’s 25-year Reanalysis (JRA-25) (Onogi et al., 2007).

There are nevertheless known important caveats to using reanalyses for climate change assessment. While the use of a fixed assimilation system and forecasting model eliminates spurious shifts in the data caused by model upgrades (Bengtsson and Shukla, 1988), the reanalyses remain sensitive to changes in the observing system (Bengtsson et al., 2004a,b). This sensitivity is exacerbated in high southern latitudes due to limited observational con-
straints (Hines et al., 2000; Marshall, 2002; Bromwich and Fogt, 2004; Bromwich et al., 2007). For example, the introduction of satellite atmospheric sounding data in late 1978 produces a jump in the Antarctic SMB simulated by ERA-40 (Van de Berg et al., 2005; Bromwich et al., 2007). In mid-1987, JRA-25 precipitation exhibits a sudden drop over the Southern Ocean following the assimilation of observations from the Special Sensor Mi-
crowave/Imager (SSM/I) (Bosilovich et al., 2006). Furthermore, reanalyses may be con-
taminated by artificial trends in the observations, such as caused by a temporal drift of satellite radiances (Dee and Uppala, 2009).

The four most recent reanalyses have certainly benefited from the continuing advances in numerical weather prediction and guidance from previous reanalysis efforts (Bengtsson et al., 2007). A variational bias correction, allowing for an automatic and adaptive handling of biases in satellite radiances, has been implemented in ERA-Interim, MERRA and CFSR.
Improved analysis of moisture has been made possible through a four-dimensional variational (4D-Var) assimilation system in ERA-Interim and JRA-55 (Andersson et al., 2007; Simmons et al., 2010), or a nudging technique in MERRA (Cullather and Bosilovich, 2011). Data mining efforts and the use of error statistics from previous reanalysis experiments have led to more complete and higher-quality observation data sets (Haimberger, 2007). Finally, increased computational power has allowed for higher horizontal and vertical model resolution, a critical aspect for modeling Antarctic SMB given the importance of orographic precipitation on the ice sheet’s steep coastal slopes.

2.1.3 Transformation of the Global Observing System Since the 1980s

The onset of the modern satellite era in 1979 represents a major transition towards a mostly satellite-based observing system (Bengtsson et al., 2004a). This system has also profoundly evolved in the more recent decades. In particular, a number of polar-orbiting satellites have been launched in the late 1990s and early 2000s, providing enhanced data coverage of the high latitudes on daily and sub-daily timescales. The operation of these new satellites has brought about a dramatic increase in the volume of satellite observations available for data assimilation (see Fig. 2.2) to the point that some parallel can be drawn with the late 1970s “shock” in the global observing system (Bromwich and Fogt, 2004). It is noteworthy that these transformations have been largely confined to areas outside continental Antarctica, because the difficult detection of clouds and uncertainties in surface emissivity still largely preclude the assimilation of satellite radiances over ice sheets (Bouchard et al., 2010; Guedj et al., 2010). Over Antarctica, the reanalyses still rely primarily on a sparse network of conventional observations (Andersson, 2007).
2.1.4 Purpose and Outline of the Analysis

The purpose of this chapter is to assess the realism of the changes in the hydrological cycle in high southern latitudes as depicted by the global reanalyses, focusing particularly on their response to the massive input of satellite observations in the late 1990s/early 2000s. The paper is organized as follows. Section 2.2 presents the data and methodology. In Section 2.3, the annual mean Antarctic net precipitation \( P-S \) simulated by the reanalyses during 1989-2009 is examined and evaluated against long-term observations. Sections 2.4 and 2.5 investigate the temporal variability and trends of Antarctic \( P-S \) and Southern Ocean precipitation. Section 2.6 describes the changes in the atmospheric circulation in high southern latitudes in an effort to explain some of the trends in Antarctic \( P-S \). Concluding remarks are given in Section 2.7.

2.2 Data and Methods

2.2.1 Background on Reanalysis Data Sets

Seven global reanalysis data sets are examined: NCEP-2, ERA-40, JRA-25, ERA-Interim, MERRA, CFSR, and JRA-55. Their main characteristics (time span, horizontal and vertical resolution, data assimilation system) are listed in Table 2.1. Because ERA-40 does not extend beyond August 2002, it is not included in all figures and tables. Brief background information about each data set is provided in the following paragraphs.

NCEP-2 and ERA-40 have been widely used in the literature and are not be redescribed here. Worth noting, however, is that NCEP-2 has the coarsest horizontal and vertical resolutions among the six reanalyses and makes limited use of satellite observations (Kanamitsu et al., 2002).
JRA-25 is based on JMA’s operational forecasting model and assimilation system as of April 2004. As such, it has been the first reanalysis to take advantage of some observations from the constellation of polar-orbiting satellites launched in the late 1990s-early 2000s, especially as part of NASA’s Earth Observing System (EOS) Program (e.g., QuikSCAT, Terra and Aqua satellites). Evaluation of JRA-25 precipitation showed higher correlation with global precipitation analyses than NCEP-2 and ERA-40 (Onogi et al., 2007; Bosilovich et al., 2008).

ERA-Interim has been the first reanalysis to implement a 4D-Var data assimilation system (Dee and Uppala, 2009). Biases in satellite radiances are corrected via a variational bias correction scheme (Dee and Uppala, 2009). A similar technique is also used to handle biases in surface pressure observations (Vasiljevic et al., 2006). Thanks to improved model physics and moisture analysis, ERA-Interim has eliminated some of the problems with the representation of the hydrological cycle in ERA-40 (Andersson et al., 2005; Uppala et al., 2008).

MERRA is the second reanalysis produced by NASA’s Global Modeling and Assimilation Office (GMAO). It uses an advanced form of 3D-Var data assimilation scheme known as the Grid-point Statistical Interpolation (GSI) (Kleist et al., 2009); a variational bias correction of satellite radiances; and the Incremental Analysis Updates (IAU), a nudging technique allowing for a smooth transition from the model states toward the observed state (Rienecker et al., 2008; Cullather and Bosilovich, 2011). Test evaluations of MERRA precipitation for two months in 2004 revealed performance equal to or greater than other reanalyses (Bosilovich et al., 2008).
CFSR was completed in January 2010 with data currently spanning 1979–2009\(^3\). It brings major improvements to NCEP-2, including higher horizontal and vertical resolutions (highest among the reanalyses used here) and intensive use of satellite observations (assimilating satellite radiances instead of satellite retrievals). Similar to MERRA, atmospheric observations are assimilated via a 3D-Var GSI algorithm. Unique among current global reanalyses, CFSR uses a coupled atmosphere-ocean-sea-ice-land model for its short-term forecasts.

JRA-55, released in late 2013, is the latest Japanese reanalysis, coming after JRA-25. It features a 4D-Var data assimilation system and a variational bias correction of satellite radiances. It is the only recent reanalysis experiment to extend back prior to 1979. Preliminary assessment carried out by Ebita et al. (2011) has already shown significant improvement in the representation of hydrological cycle in JRA-55 compared to JRA-25.

### 2.2.2 Estimation of Antarctic Surface Mass Balance

Over Antarctica, precipitation minus surface sublimation (\(P-S\)) is used as an approximation of the ice sheet’s SMB. \(P-S\) neglects some ablation processes such as melt/runoff, horizontal snow transport and blowing-snow sublimation. Melt/runoff is mostly confined to the coastal ice shelves. On the continent, the meltwater generally refreezes within the snowpack, thereby contributing negligibly to the overall mass loss from the ice sheet (Liston and Winther, 2005). The two wind-induced processes (not represented in the reanalysis models) are most consequential in the escarpment zone, where the katabatic flow is strongest, and can lead locally to complete removal of the annual snow layer (Frezzotti et al., 2004; Vanden Broeke et al., 2004; Frezzotti et al., 2007; Genthon et al., 2007; Lenaerts et al., 2010; Scarchilli et al., 2010). Averaged over the entire ice sheet, the snow transport becomes a

\(^3\)CFSR has been extended through 2013 but access to the 2010–2013 data is limited.
2nd-order contributor to the SMB (Déry and Yau, 2002; Van den Broeke et al., 2006a). On the other hand, the average blowing-snow sublimation is of the same order as surface sublimation, accounting for 10-20% of precipitation (Bintanja, 1998; Déry and Yau, 2002; Van den Broeke et al., 2006a). With regards to the positive contributions to the SMB, none of the reanalysis models includes a representation of clear-sky precipitation, a potentially important, yet uncertain, component on the East Antarctic plateau (Bromwich, 1988; Fujita and Abe, 2006).

As will be demonstrated hereafter, the differences in \( P \) and \( S \) among the reanalyses are of first order compared to their average values and interannual variability. Therefore, a refined SMB calculation (as, e.g., in Lenaerts et al., 2012) is deemed unnecessary at this point, as it would further complicate the understanding of the discrepancies between reanalyses.

Monthly estimates of total precipitation (\( P \)) and surface sublimation (\( S \)) are taken from the short-term forecast fields of the reanalyses (as opposed to their analysis fields) and, as such, are more influenced by the model physics than standard variables like atmospheric temperature and pressure. \( S \) is not available in NCEP-2, JRA-25, and CFSR and is estimated from the surface latent heat flux (\( F_{LH} \)) as \( l_s F_{LH} \), where \( l_s \) is the latent heat of sublimation, taken here as \( 2.838 \times 10^6 \) J kg\(^{-1}\) (Rogers and Yau, 1989).

### 2.2.3 Observation-based Data Sets

Over Antarctica, \( P - S \) is evaluated against the observation-based accumulation map from Arthern et al. (2006). This data set is based on the same compilation of in-situ measurements as the earlier work from Vaughan et al. (1999), but employs an improved background field to guide the interpolation: satellite passive microwave and thermal infrared (TIR) observations for Arthern et al. (2006) versus TIR-only for Vaughan et al. (1999).
The map from Arthern et al. (2006) incorporates glaciological observations collected since the 1950s, but heterogeneously distributed in time. This data set can thus be considered as a depiction of the long-term Antarctic SMB, assuming that no significant change in the long-term accumulation has occurred in Antarctica during the second half of the 20th century. This assumption, which is also required to compare Arthern et al. (2006) with the reanalysis data, is supported by the results from Monaghan et al. (2006b) and Monaghan and Bromwich (2008).

The accuracy of the Vaughan et al. (1999) and Arthern et al. (2006) maps in some areas has been questioned because of the use of unreliable accumulation measurements (Frezzotti et al., 2004; Magand et al., 2007; Genthon et al., 2009; Favier et al., 2013). In addition, the passive microwave background used by Arthern et al. (2006) was shown to be unreliable in coastal areas affected by melt/refreeze during summer (Magand et al., 2008). Nonetheless, Arthern et al. (2006) showed that their data set alleviates some of the biases present in Vaughan et al. (1999), producing especially lower accumulation values on the East Antarctic plateau.

Over the Southern Ocean, two satellite-gauge precipitation analysis data sets are used to evaluate the reanalysis precipitation: the Version 2 Global Precipitation Climatology Project (GPCP) merged precipitation data (Adler et al., 2003), and the Climate Prediction Center Merged Analysis of Precipitation (CMAP) (Xie and Arkin, 1997). These data sets provide gridded monthly precipitation data at a 2.5°×2.5° resolution. CMAP has missing data poleward of 60° latitude. GPCP data provide a global coverage but its precipitation estimates over ice-covered areas are deemed unreliable (Serreze et al., 2005).

As the short descriptions of the Arthern et al. (2006), GPCP, and CMAP data sets already suggest, and as will become more obvious in the analysis that follows, there are
significant errors and uncertainties in the very data sets that should in theory serve as references for the evaluation. Therefore, the “realism” of the reanalyses, or their consensus (or lack thereof), will often be used as a more subjective metric to assess the reliability of their data.

2.3 Representation of Long-term Antarctic Surface Mass Balance

An accurate representation of Antarctica’s long-term mean SMB is the first challenge faced by the reanalyses whose model physics is generally developed, and more adapted, for lower-latitude climate than for polar environment. Monaghan et al. (2006a) and Van de Berg et al. (2006) summarized several Antarctic SMB estimates from recent studies, showing a significant scatter among the results (84 to 180 mm year\(^{-1}\) on average for the grounded ice sheet). The new set of reanalyses provide the opportunity to determine whether this scatter is still an issue.

2.3.1 Spatial distribution

The spatial distribution of mean total annual \(P-S\) from six reanalyses during 1979–2009 is displayed in Fig. 2.3, along with the accumulation map from Arthern et al. (2006). NCEP-2 notwithstanding, the reanalyses correctly reproduce the known contrast between the high-accumulation coast and the dry interior. NCEP-2 on the other hand shows small accumulation rates over most of coastal East Antarctica. As already noted by Monaghan et al. (2006a), these low values are due to unrealistically high surface sublimation fluxes (see also Section 2.3.3), linked to overestimated latent heat fluxes in stable boundary layer conditions (Hines et al., 1999). In coastal areas, ERA-Interim, MERRA and CFSR capture the leeward/windward precipitation gradient, e.g., across Law Dome (110°E; Van Ommen et al., 2004). At least two studies (Magand et al., 2007, 2008) have suggested that the
accumulation rates from Arthern et al. (2006) are excessively low in coastal areas (Fig. 2.3g).

Over the East Antarctic interior, which receives substantially less snowfall than the coast, the differences between the reanalyses are accentuated by the non-linear scale. Over most of the plateau, JRA-25 exhibits $P-S$ values $>50$ mm year$^{-1}$, with maxima $>100$ mm year$^{-1}$, whereas ERA-Interim exhibits values $<10$ mm year$^{-1}$. Intermediate accumulation rates are found in MERRA, CFSR, and JRA-55. The locations of negative $P-S$ in CFSR are consistent with existing blue-ice areas (Winther et al., 2001). But because these ablation areas result from wind processes not included in the reanalysis, such agreement would rather suggest insufficient $P$ or excessive $S$ locally in CFSR. In West Antarctica, relatively good agreement is found between the data sets. The model resolution (Table 2.1) greatly affects the precipitation gradient in coastal areas, with enhanced orographic precipitation in the higher-resolution models, e.g. in Marie Byrd Land.

### 2.3.2 Evaluation Against Accumulation Observations

Figure 2.4 presents, for each reanalysis, the ratio between the 31-year averaged $P-S$ (previously shown in Fig. 2.3) and the accumulation estimates from Arthern et al. (2006). Of note, $P-S$ better approximates the “true” SMB on the plateau (as opposed to the coast) because of weaker katabatic winds, and thus lower contribution of wind-driven ablation processes.

The large positive biases found in all six reanalyses over Marie Byrd Land, West Antarctica, are in fact due to strongly underestimated accumulation rates in Arthern et al. (2006), which in this area relies on poorly constrained stratigraphic observations (Van de Berg et al., 2006; Van den Broeke et al., 2006b). Figure 2.4a confirms the excessively low $P-S$ values in NCEP-2 seen in Fig. 2.3a. Over a large fraction of the East Antarctic plateau, especially
its highest-elevated portion, the overestimation of accumulation in JRA-25 exceeds 60% compared to Arthern et al. (2006). Onogi et al. (2007) indicate that such excessive snowfall may be related to the spectral truncation of model variables in regions with extremely low saturated water pressure vapor. This problem is no longer present in JRA-55.

On the East Antarctic plateau, mainly for elevations >2500 m a.s.l., $P-S$ in ERA-Interim represents less than 20% of the accumulation values from Arthern et al. (2006). Magand et al. (2007) applied a strict quality-control filtering to the observations used by Arthern et al. (2006) in the 90°E-180°E quadrant. They find long-term accumulation rates of $>40$ mm year$^{-1}$ both in the 3000-3500m and 3500-4000m elevation ranges, which confirms the dry bias in ERA-Interim. Other recent accumulation measurements in the hinterland of Dronning Maud Land generally produce values $>20$ mm year$^{-1}$ (e.g., Kameda et al., 2008; Anschütz et al., 2009). It is noteworthy that the dry bias is significantly reduced when ERA-Interim $P-S$ is estimated from the atmospheric moisture flux budget method rather than from the reanalysis forecast fields (Lennart Bengtsson, personal communication 2010).

MERRA, CFSR, and JRA-55 best agree with Arthern et al. (2006) overall. Because $P-S$ does not account for wind-induced ablation, one would expect ratios >1 not only in the immediate coastal band (as seen in Fig. 2.4d and 2.4e) but also farther inland. Therefore, ratios within 0.6-1.0 suggest insufficient precipitation amounts in these three reanalyses.

### 2.3.3 Antarctic-wide Averaged Surface Mass Balance

The 1979–2009 mean total annual values for $P$, $S$ and $P-S$ spatially averaged over the Antarctic Ice Sheet (grounded and with ice shelves) are reported in Table 2.2. Because of sublimation fluxes representing roughly 40% of the annual precipitation amount, NCEP-2 exhibits $P-S$ values substantially lower than the other data sets. JRA-25, MERRA and CFSR show close $P-S$ values, within 155-165 mm year$^{-1}$ averaged over the grounded
ice sheet, although the similarity between JRA-25 and CFSR masks contrasting spatial distributions. These estimates compare well with 157±9 mm year$^{-1}$ obtained by Monaghan et al. (2006a) for 1985–2001 with a regional climate model forced at the boundaries by NCEP-2. The observation-based data set from Arthern et al. (2006) yields a somewhat lower value (143±4 mm year$^{-1}$) as a result of lower coastal accumulation. Close to the 155-160 mm year$^{-1}$ range are the results from Van de Berg et al. (2006), 166 mm year$^{-1}$ for 1980-2002, based on a regional climate model simulation forced by ERA-40, and from Krinner et al. (2007), 151 mm year$^{-1}$ for 1980-1999, based on a global coupled ocean-atmosphere model simulation. With the lowest precipitation amounts, ERA-40 and ERA-Interim have $P-S$ values markedly smaller than the other reanalyses (except for NCEP-2). Nevertheless, it is noteworthy that ERA-Interim has reduced the dry bias in ERA-40, primarily thanks to smaller surface sublimation fluxes.

2.4 Temporal Variability and Trends in Antarctic Surface Mass Balance

The mean annual values and spatial distributions of the Antarctic $P-S$ described above largely depend upon the physics and configuration of the reanalysis models. The time series and trends of Antarctic $P-S$, which are examined in this section, reflect not only actual changes in the Antarctic climate but also the potentially spurious impact of the evolving observing system.

2.4.1 Time Series of Antarctic Surface Mass Balance

The time series of total annual $P$, $S$ and $P-S$ spatially averaged over the grounded Antarctic Ice Sheet are shown in Fig. 2.5 and presented as anomalies from their respective 1990–1995 means. Despite some common interannual variability, marked differences are clearly visible between the data sets. NCEP-2 displays a pronounced upward trend in
precipitation through the entire period, which is not seen elsewhere (Fig. 2.5a). The reasons for such evolution are unclear. MERRA departs markedly from the other reanalyses in 1999; during 1999-2009, its precipitation is on average 15 mm year\(^{-1}\) (\(\sim 10\%\)) larger than during 1979-1998. The upward shift in MERRA in the late 1990s/early 2000s is related to the assimilation of radiances from the Advanced Microwave Sounding Unit (AMSU) (Cullather and Bosilovich, 2011; Rienecker et al., 2011). This instrument, which is part of the Advanced TOVS (ATOVS) instrument suite, started being flown on the polar-orbiting NOAA-15 satellite in 1998. The impact of AMSU data in MERRA is further discussed in Sections 2.5 and 2.6.

From 2006 onward, a widening gap occurs between ERA-Interim and CFSR on the one hand, and the three other reanalyses on the other hand. It is not certain whether this is related to changes in the assimilated observations. It is noteworthy that, in 2006, ERA-Interim and CFSR start assimilating Global Positioning System radio-occultation data from the Constellation Observing System for Meteorology Ionosphere and Climate (COSMIC) mission (Poli et al., 2010; Saha et al., 2010). These observations, which can be assimilated without bias correction, can be used over Antarctica and have been shown to have a positive impact on the quality of analyses and forecasts in the Antarctic (Cucurull et al., 2006; Healy, 2007).

The interannual variations in S in NCEP-2 greatly exceed those in the other data sets (Fig. 2.5b). These variations are commensurate with the total sublimation fluxes in NCEP-2, shown to be unrealistically high (Table 2.2). With the exception of NCEP-2, the variations of \(P-S\) largely mimic those of P. Therefore, the description of Fig. 2.5a can, to a large extent, be applied to Fig. 2.5c.
Table 2.3 presents the linear trends associated with the P–S time series shown in Fig. 2.5c for the 1979–2009 and 1989–2009 periods, respectively. The statistical significance of the trends is estimated from the $p$-value of a two-tailed Student’s $t$-test. During neither period are the P–S trends in CFSR and ERA-Interim significantly different from zero. On the contrary, the trends in NCEP-2, JRA-25, MERRA, and JRA-55 are all positive and statistically significant at the 95% confidence level during 1979–2009. The significance of these trends even exceeds 99% during 1989–2009. Overall, Table 2.3 serves as a good example of the difficulty that one faces when trying to assess global reanalyses in high southern latitudes. As the rest of the chapter will show, the fact that the majority of the reanalyses exhibit a significant increase in Antarctic P–S is by no means a reliable indicator.

### 2.4.2 Spatial Distribution of Trends Across Antarctica

The temporal trends in average Antarctic P–S mask important regional differences, with again contrasting pictures among the reanalyses (Fig. 2.6). Validating these trends against in situ observations is challenging. Few ice-core accumulation records extend beyond 2000. Other in-situ observations (e.g., stake farms) only provide a partial coverage of the 1979–2009 period.

ERA-Interim displays overall smaller and less significant trends than the other reanalyses. NCEP-2, JRA-25 and MERRA reveal large areas with highly significant positive trends in Dronning Maud Land (DML), albeit with rather dissimilar patterns. The analysis of shallow cores from coastal DML (7–9°W) by Fernandoy et al. (2010) rather suggests decreasing accumulation through 1989-2007. A similar analysis conducted farther east (~12°E) by Anschütz et al. (2007) does not suggest any significant trend through 1989-2004.

Some similarity is found among the six maps in the 70°-170°E sector, with positive trends over Lambert Glacier and central Wilkes Land, and downward trends over western
Wilkes Land and Victoria Land. This alternation of patterns of opposite signs is suggestive of changes in the atmospheric circulation and the associated moisture advection towards/off the coast (see Section 2.6). The magnitude and spatial extent of the patterns differ substantially among the reanalyses. NCEP-2 and JRA-25 exhibit trends in excess of 200 mm year$^{-1}$ decade$^{-1}$ over extensive areas of central Wilkes Land. Such large values cast doubts on the realism of the two reanalyses in this sector. These large increase also contrast with zero or negative trends in ERA-Interim. All data sets except ERA-Interim have statistically significant (CL>90%) positive trends over Law Dome, whereas a ice-core measurements indicate decreasing accumulation during 1989-2004 (Van Ommen and Morgan, 2010). Farther east, the patterns of large negative trends in $P-S$ over Victoria Land—a common feature among the reanalyses—are consistent with increasing sea-ice cover in the neighboring Ross Sea (Stammerjohn et al., 2008; Turner et al., 2009), both resulting from enhanced off-continent winds.

Over West Antarctica, the trends in $P-S$ are generally not statistically significant; this corroborates the conclusions from a recent analysis of airborne-radar snow accumulation measurements (Medley et al., 2013). The positive trends near the southern portion of the Antarctic Peninsula are consistent with the marked increase in accumulation found at the nearby Gomez ice-core site from the 1970s through 2006 (Thomas et al., 2008).

### 2.5 Beyond Antarctica: Precipitation Changes over the Southern Ocean and Lower Latitudes

In this section, the analysis is extended spatially to encompass precipitation changes over the ocean areas surrounding Antarctica. These changes are also placed in a broader context by considering global and Southern Hemisphere precipitation. In this case, the temporal variability of precipitation over the Southern Ocean can be—tentatively—assessed against
satellite-gauge precipitation analyses. In addition, inhomogeneities in the time series related to the assimilation of satellite observations are expected to appear more clearly over the ocean than over Antarctica due to the paramount role of satellite observations over ocean areas.

2.5.1 Time Series of Southern Ocean Precipitation

Figure 2.7 shows the time series of mean annual precipitation spatially averaged over four domains. Precipitation averaged between 60°S and 60°N (Fig. 2.7a) largely reflects the global mean. The precipitation estimates from the reanalyses—with the exception of MERRA prior to 1998—systematically exceed those from GPCP and CMAP, because of overestimated tropical rainfall in the reanalyses (Bosilovich et al., 2008). The origin of the global precipitation increase in NCEP-2 during the first part of the 1990s is unclear. The erratic behavior of ERA-40 results from deficiencies in its moisture analysis, particularly following the eruption of Mt. Pinatubo in 1991 (Andersson et al., 2005; Uppala et al., 2005). In ERA-Interim, artificial shifts in global mean precipitation have been linked to changes in the volume of SSM/I radiances assimilated into the reanalysis (Dee et al., 2011). This sensitivity explains the drop in 1992 and the positive trend after 2005. As already shown over Antarctica for MERRA, the impact of AMSU is visible globally not only in MERRA, but also in JRA-25 and CFSR (Onogi et al., 2007; Sakamoto and Christy, 2009; Saha et al., 2010; Cullather and Bosilovich, 2011). Further assessment of the global hydrological cycle in recent global reanalyses can be found in Trenberth et al. (2011).

In the 50–60°S latitude band (Fig. 2.7b), GPCP estimates are approximately 40% higher than those from CMAP. The reanalysis values (with the exception of CFSR) lie between the two curves. The large discrepancy between GPCP and CMAP is related to the use of TOVS precipitation retrievals in GPCP, but not in CMAP (Adler et al., 2003; Yin
et al., 2004). The drop in JRA-25 precipitation from 1987 to 1988 is a known issue related to the assimilation of SSM/I precipitable water observations (Bosilovich et al., 2006; Onogi et al., 2007).

In the two plots pertaining to the Southern Ocean (Figs. 2.7b and 2.7c), a jump in MERRA precipitation occurs between 1998 and 2001, resulting in a $\sim 25\%$ increase in the mean annual value. As mentioned above, this artifact is caused by the assimilation of AMSU radiances into the reanalysis. Cullather and Bosilovich (2011) report an even greater jump ($\sim 40\%$) in $P-S$ (instead of $P$ considered here). The timing of the precipitation jump in MERRA is further examined in Fig. 2.8. This figure shows the monthly precipitation differences between MERRA and ERA-Interim over the $50^\circ$-$60^\circ$S latitude band. The jump actually consists of two successive steps: the first one in late 1998, concurrent with the assimilation of AMSU data from the NOAA-15 satellite; the second one in early 2001, concurrent with the assimilation of AMSU data from the NOAA-16 satellite.

Over the ocean south of $60^\circ$S (Fig. 2.7c), CFSR clearly stands out, with mean annual precipitation about $25\%$ greater than the other data sets. Relatively good agreement is found between NCEP-2, JRA-25, ERA-Interim and GPCP during the 1990s. However, the curves show contrasting evolution from 2001 onward, with a marked increase in NCEP-2 (in line with MERRA), and a sharp decline in ERA-Interim and GPCP.

The comments made previously about Antarctic precipitation (Fig. 2.5a) largely apply to Fig. 2.7d, which is simply shown here with different units and with absolute values instead of anomalies. A striking feature in Fig. 2.7d is the sudden drop in GPCP precipitation after 1999 and continuing throughout the 2000s, making this data set unreliable over Antarctica during this period. This sharp decrease may be caused by a reduction in the amount of TOVS data used in GPCP from 1999 onward (Adler et al., 2003).
Table 2.4 summarizes the trends in average precipitation over the Southern Ocean (south of 50°S) from the different data sets. The trends from the reanalyses bear much similarity with those calculated for Antarctic $P-S$ (Table 2.3), with the notable exception of JRA-25, as a result of the SSM/I-related drop in precipitation in 1987 (Fig. 2.7b). Large positive trends are found in NCEP-2, MERRA and CMAP. Trends are not significantly different from zero in CFSR during either period.

### 2.5.2 Spatial Distribution of Precipitation Trends Over the Southern Hemisphere

Figure 2.9 shows the 1989–2009 linear trends in precipitation over the entire Southern Hemisphere. MERRA stands out with a pattern of large positive and highly significant trends over the Southern Ocean south of 40°S, an area in which the reanalyses are almost exclusively reliant on satellite data. Figures 2.5 and 2.6 strongly suggest that these inhomogeneities have contaminated the trends in $P-S$ over Antarctica in MERRA.

This pattern likely accounts for the predominant positive trends in $P-S$ over Antarctica in MERRA (Fig. 2.6). Large positive precipitation trends are also found in the Pacific sector of the Southern Ocean in NCEP-2. In contrast, small and insignificant trends prevail over the Southern Ocean in ERA-Interim, CFSR, and JRA-55. The widespread positive trends in CFSR over the Atlantic and eastern Pacific basins have been—here again—attributed to the spurious effect of the assimilation of AMSU data (Zhang et al., 2012).

Some overall agreement is found between GPCP and CMAP (Figs. 2.9e and 2.9f), despite the low temporal correlations between the two data sets noted by Yin et al. (2004). The precipitation trends in ERA-Interim and JRA-55 are those that best agree with the trends from GPCP and CMAP in low and mid-latitudes. Despite strong dissimilarities, all maps in Fig. 2.9 share some common features: downward precipitation trends in the
western equatorial Pacific; upward (and generally significant) trends over the South Pacific Convergence Zone (SPCZ); and southeast-northwest contrast over Australia.

2.6 Contribution of Atmospheric Circulation Changes to Apparent Trends in Antarctic Surface Mass Balance

This section investigates the contribution of the atmospheric circulation to the trends in Antarctic P–S simulated by the reanalyses. Because of the large contrast in moisture content between the polar continental atmosphere and the surrounding maritime air masses (Tietäväinen and Vihma, 2008), changes (either real or spurious) in the meridional wind component can greatly affect the moisture transport and the related precipitation (e.g., Schlosser et al., 2010).

2.6.1 Trends in 500-hPa Geopotential Height and Surface Pressure

The spatial distribution of the 1989–2009 linear trends in 500-hPa geopotential height ($Z_{500}$) and surface pressure poleward of 40°S are displayed in Figs. 2.10 and 2.11, respectively. A common characteristic among the reanalyses is a pattern of negative trends in the South Pacific sector of the Southern Ocean (90°-180°W). Again, the contours and magnitude of the pattern vary greatly among the data sets. Over Antarctica, NCEP-2 and JRA-25 exhibit the largest absolute trend magnitudes, but mostly with opposite signs. CFSR and JRA-55 exhibit large positive trends in $Z_{500}$ and surface pressure over East Antarctica. In ERA-Interim, the negative pattern over the South Pacific Ocean extends over the Ross Sea and Ross Ice Shelf and across West Antarctica to Dronning Maud Land. A relatively similar pattern is present in MERRA.
2.6.2 Time Series of 500-hPa Geopotential Height

Possible explanations for some of the discrepancies between the data sets are given in Figs. 2.12 and 2.13. Figure 2.12 shows spatially averaged $Z_{500}$ differences between the reanalyses and ERA-Interim. Over Antarctica (Fig. 2.12a), the $Z_{500}$ discrepancy between NCEP-2 and ERA-Interim (and the other reanalyses) becomes substantially reduced from 1998 onward. This discontinuity may come from the transition between the assimilation of TOVS retrievals to the direct assimilation of TOVS radiances, which was to be implemented in 1998 in the NCEP-National Center for Atmospheric Research (NCAR) Reanalysis (Kalnay et al., 1996) and thus, supposedly, in NCEP-2.

For JRA-25, the $Z_{500}$ difference exhibits a marked decreased both over Antarctica (Fig. 2.12a) and over the surrounding ocean (Fig. 2.12b) beginning around 1998-1999. This decline may be related to the transition from TOVS to Advanced TOVS (ATOVS) in October 1998, which also coincided with a change in the bias correction technique applied to these sounding radiances (Onogi et al., 2007; Sakamoto and Christy, 2009). Both CFSR and JRA-55 exhibit a pronounced increase in the $Z_{500}$ difference over the continent. Comparison with radiosonde observations from Amundsen-Scott South Pole Station indicates an positive shift in $Z_{500}$ between 1995 and 2000 (Bracegirdle and Marshall, 2012). Until 1996, CFSR underestimates $Z_{500}$ by $\sim 20$ gpm whereas, after 2000, this bias is considerably reduced and comparable to the values seen in the other data sets.

2.6.3 Case Study: Casey Station, Coastal East Antarctica

The marked downward trends in $Z_{500}$ and surface pressure in NCEP-2 and JRA-25 around $120^\circ$E contrast drastically with the upward trends in the three other reanalyses (Figs. 2.10 and 2.11). The greatest discrepancy is seen over the coast, in the vicinity of the
Australian station, Casey, whose mean sea level pressure (MSLP) and $Z_{500}$ observations are compared with the reanalysis data in Fig. 2.13. These observations are obtained from the Antarctic READER database (Turner and Pendlebury, 2004).

A prominent feature in Fig. 2.13 is a drop in MSLP bias in NCEP-2 between 1991 and 1994. Some similarity is seen in the evolution of the $Z_{500}$ bias, although the bias becomes negative during 1994-1996. Marshall (2002) showed that a sudden decrease in monthly mean $Z_{500}$ values in 1993 in the NCEP-NCAR Reanalysis coincides with the assimilation of pressure observations from automatic weather stations with erroneous model heights. The early-1990s drop (also seen in JRA-25) is likely related to the end of the radiosonde program at Vostok (south of Casey) in early 1992.

In JRA-25, the large negative trends in surface pressure over Casey and the adjacent ocean sector (Fig. 2.11) result from decreasing pressure values in the early 1990s and after the year 2000 (Fig. 2.13). As mentioned in Section 2.6.6.2, the transition to ATOVS is one possible explanation for the post-2000 evolution in JRA-25 pressure/geopotential height fields.

### 2.6.4 Overall Impact and Reliability of Simulated Changes in the Atmospheric Circulation

Overall, a clear correspondence can be established between the trends in $Z_{500}$ and surface pressure (especially the high-gradient zones) and the trends in $P-S$ from Fig. 2.6. In NCEP-2 and JRA-25, the changes in $Z_{500}$ and surface pressure act to greatly enhance the poleward wind flow over coastal central Wilkes Land. In CFSR, the large positive trends in $Z_{500}$ and surface pressure over the East Antarctic interior and over the coast near 130°E account for enhanced positive and negative trends in $P-S$ in the 45°-160°E sector, as compared

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4http://www.antarctica.ac.uk/met/READER/.
to ERA-Interim (Fig. 2.6c and 2.6e). There is however no apparent net impact on the Antarctic $P-S$ simulated by CFSR. A similar reasoning can be applied to other Antarctic regions such as Dronning Maud Land or West Antarctica. From the spurious origin of the trends in NCEP-2 and JRA-25, one may conclude that the trends in $P-S$ in these data sets are also spurious or overestimated.

2.7 Conclusions

Four main conclusions can be drawn from the analysis presented in this chapter. First, because of the absence of obvious artifacts in its $P-S$ time series, ERA-Interim likely provides the most realistic depiction of the interannual variability and trends in Antarctic $P-S$ during 1989–2009. This conclusion, first formulated by Bromwich et al. (2011) and Nicolas and Bromwich (2011b), was later confirmed for a vast sector of West Antarctica by Medley et al. (2013). With regard to the absolute value of $P-S$, ERA-Interim exhibits a dry bias overall, particularly pronounced in the interior of East Antarctica.

Second, there is strong evidence that the large positive trends in Antarctic $P-S$ during 1989–2009 in NCEP-2, JRA-25 and MERRA are exaggerated; the non-significant trends found in ERA-Interim and CFSR are likely closer to the “truth”. This conclusion corroborates the absence of significant long-term change in Antarctic $P-S$ previously reported by Van de Berg et al. (2005), Monaghan et al. (2006a) and Monaghan et al. (2006b). The large interannual variability and the relatively short time period investigated here emphasize the need for longer time series.

Third, the “shock” represented by the dramatic increase in the volume of satellite observations in the 1990s-2000s (Dee et al., 2009; Saha et al., 2010) has proved of a different nature from what was initially conjectured. The increase in the volume of observational
data results to a large extent from the new data provided by high-resolution atmospheric sounders launched in the 2000s, such as the Atmospheric Infrared Sounder (AIRS). No disruption in the reanalysis time series has been found to be associated with the introduction of these sounding radiances. Surprisingly, the assimilation of AMSU observations, with comparably few channels, appears to have a much greater impact on some of the reanalyses.

The fourth and final conclusion is that the high southern latitudes remain a challenging place for retrospective analysis experiments. The issue of artificial trends and jumps warrants caution when using these data sets for climate change assessment. The range of trends in Antarctic $P-S$ among the five data sets analyzed here—between +1 and +26 mm year$^{-1}$ of change in 20 years (Table 2.3)—is a difference equivalent to 1 mm of sea level. This discrepancy underscores the importance of improving the reanalyses simulations of Antarctic snowfall so that the contribution of the Antarctic ice sheet to sea level rise can be better understood.
Table 2.1. Characteristics of the global atmospheric reanalyses used in this study, presented in their chronological order of release.

<table>
<thead>
<tr>
<th>Reanalysis</th>
<th>Organization</th>
<th>Time coverage</th>
<th>Horizontal grid</th>
<th>Vertical levels</th>
<th>Assimilation system</th>
</tr>
</thead>
<tbody>
<tr>
<td>ERA-15*</td>
<td>ECMWF</td>
<td>1979–1993</td>
<td>T106, ~125 km</td>
<td>31</td>
<td>1D-Var</td>
</tr>
<tr>
<td>NCEP-NCAR*</td>
<td>NCEP/NCAR</td>
<td>1948–present</td>
<td>T62, ~210 km</td>
<td>28</td>
<td>3D-Var</td>
</tr>
<tr>
<td>NCEP-2</td>
<td>NCEP/DOE</td>
<td>1979–present</td>
<td>T62, ~210 km</td>
<td>28</td>
<td>3D-Var</td>
</tr>
<tr>
<td>ERA-40</td>
<td>ECMWF</td>
<td>1957–08/2002</td>
<td>T159, ~125 km</td>
<td>60</td>
<td>3D-Var</td>
</tr>
<tr>
<td>JRA-25</td>
<td>JMA/CRIEPI</td>
<td>1979–present</td>
<td>T106, ~125 km</td>
<td>40</td>
<td>3D-Var</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>ECMWF</td>
<td>1989–present</td>
<td>T255, ~80 km</td>
<td>60</td>
<td>4D-Var</td>
</tr>
<tr>
<td>MERRA</td>
<td>NASA GMAO</td>
<td>1979–present</td>
<td>1/2°×2/3°, ~55 km</td>
<td>72</td>
<td>3D-Var</td>
</tr>
<tr>
<td>CFSR</td>
<td>NCEP</td>
<td>1979–2009</td>
<td>T382, ~38 km</td>
<td>64</td>
<td>3D-Var</td>
</tr>
<tr>
<td>JRA-55</td>
<td>JMA</td>
<td>1958–present</td>
<td>T319, ~60 km</td>
<td>60</td>
<td>4D-Var</td>
</tr>
</tbody>
</table>

* Data set not used in this chapter.
Table 2.2. Mean total annual precipitation (P), sublimation (S), and net precipitation (P–S) during 1979–2009 (1979–2001 for ERA-40) in mm year$^{-1}$ spatially averaged over the grounded ice sheet (12.35 $\times$ 10$^6$ km$^2$) and over the total ice sheet (13.82 $\times$ 10$^6$ km$^2$).

<table>
<thead>
<tr>
<th></th>
<th>Grounded Ice Sheet</th>
<th>Total Ice Sheet</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>P</td>
<td>S</td>
</tr>
<tr>
<td>NCEP-2</td>
<td>165</td>
<td>69</td>
</tr>
<tr>
<td>JRA-25</td>
<td>202</td>
<td>37</td>
</tr>
<tr>
<td>ERA-40</td>
<td>145</td>
<td>25</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>148</td>
<td>17</td>
</tr>
<tr>
<td>MERRA</td>
<td>168</td>
<td>12</td>
</tr>
<tr>
<td>CFSR</td>
<td>201</td>
<td>40</td>
</tr>
<tr>
<td>JRA-55</td>
<td>165</td>
<td>21</td>
</tr>
</tbody>
</table>
Table 2.3. Trends in total annual net precipitation ($P - S$) over the grounded Antarctic Ice Sheet in mm year$^{-1}$ decade$^{-1}$ for the 1979–2009 and 1989–2009 periods. The trend uncertainties correspond to two standard errors of the linear regression slope. Bold font indicates trends statistically significant at the 95% confidence level. Asterisks are used where this level exceeds 99%.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>NCEP-2</td>
<td>7.9 ± 2.9*</td>
<td>8.4 ± 4.7*</td>
</tr>
<tr>
<td>JRA-25</td>
<td>4.2 ± 3.9</td>
<td>11.8 ± 6.8*</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>−1.3 ± 3.2</td>
<td>0.5 ± 5.4</td>
</tr>
<tr>
<td>MERRA</td>
<td>5.5 ± 3.9*</td>
<td>13.3 ± 6.1*</td>
</tr>
<tr>
<td>CFSR</td>
<td>0.4 ± 3.6</td>
<td>2.6 ± 6.0</td>
</tr>
<tr>
<td>JRA-55</td>
<td>6.1 ± 3.3*</td>
<td>7.9 ± 6.2*</td>
</tr>
</tbody>
</table>
Table 2.4. Trends in daily average precipitation over the Southern Ocean (50–60°S) in mm day$^{-1}$ decade$^{-1}$ for the 1979–2009 and 1989–2009 periods. The trend uncertainties correspond to two standard errors of the slope of the linear regression. Bold font indicates trends statistically significant at the 95% confidence level. Asterisks are used where this level exceeds 99%.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>NCEP-2</td>
<td>0.14 ± 0.03*</td>
<td>0.20 ± 0.06*</td>
</tr>
<tr>
<td>JRA-25</td>
<td>−0.14 ± 0.07*</td>
<td>0.10 ± 0.05*</td>
</tr>
<tr>
<td>ERA-Interim</td>
<td>−0.03 ± 0.02</td>
<td>−0.02 ± 0.04</td>
</tr>
<tr>
<td>MERRA</td>
<td>0.19 ± 0.07*</td>
<td>0.36 ± 0.11*</td>
</tr>
<tr>
<td>CFSR</td>
<td>0.00 ± 0.03</td>
<td>0.02 ± 0.07</td>
</tr>
<tr>
<td>JRA-55</td>
<td>0.02 ± 0.02</td>
<td>0.04 ± 0.05</td>
</tr>
<tr>
<td>GPCP</td>
<td>0.03 ± 0.04</td>
<td>0.03 ± 0.08</td>
</tr>
<tr>
<td>CMAP</td>
<td>0.04 ± 0.03</td>
<td>0.13 ± 0.04*</td>
</tr>
</tbody>
</table>
Figure 2.1. Map of Antarctica.
Figure 2.2. Time series of the number and types of observations (a) considered for assimilation in the MERRA reanalysis during a 6-h window, and (b) actually assimilated in this reanalysis for the period spanning 1979–2011. Reprint from Fig. 4 of Rienecker et al. (2011).
Figure 2.3. Long-term average total annual snow accumulation over Antarctica in mm w.e. year\(^{-1}\) based on 1979–2009 \(P-S\) estimates from six reanalyses (a to f) and the observation compilation from Arthern et al. (2006) (g).
Figure 2.4. Comparison (ratio) between the 1979–2009 averaged $P-S$ fields from global reanalyses and the observation compilation from Arthern et al. (2006). All the data sets are interpolated to a $1^\circ \times 1^\circ$ latitude-longitude grid.
Figure 2.5. Time series of total annual precipitation \( P \) (a), sublimation \( S \) (b), and net precipitation \( P - S \) (c) in mm w.e. year\(^{-1}\) spatially averaged over the grounded Antarctic Ice Sheet. The time series are presented as departures from their respective 1990–1995 means.
Figure 2.6. Linear trends in $P-S$ during 1989–2009 in mm year$^{-1}$ decade$^{-1}$. Hatching and stippling denote trends statistically significant at the 90% and 95% confidence levels, respectively.
Figure 2.7. Mean annual precipitation in mm day$^{-1}$ spatially averaged over four domains: (a) 60°S–60°N; (b) 50–60°S; (c) the Southern Ocean poleward of 60°S; and (d) the Antarctic continent, including the floating ice shelves. The precipitation estimates are from six global reanalyses plus the GPCP and CMAP data sets. CMAP does not provide data poleward of 60°S and is therefore not included in plots (c) and (d).
Figure 2.8. Monthly precipitation differences between MERRA and ERA-Interim in mm day$^{-1}$ spatially averaged over the Southern Ocean (50–60°S), highlighting the jump in MERRA precipitation in the late 1990s/early 2000s.
Figure 2.9. Linear trends in total annual precipitation during 1989–2009 in mm year$^{-1}$ decade$^{-1}$ over the Southern Hemisphere from the global reanalyses (a to f) and from the GPCP and CMAP data sets (g and h). For lack of data or reliability, the trends from GPCP and CMAP are not shown poleward of 60°S. Hatching (stippling) denotes trends statistically significant at the 90% (95%) confidence level.
Figure 2.10. Linear trends in 500-hPa geopotential height poleward of 40°S during 1989–2009 in gpm decade$^{-1}$. Stippling denotes trends statistically significant at the 95% confidence level.
Figure 2.11. Linear trends in surface pressure poleward of 40 °S during 1989–2009 in hPa decade$^{-1}$. Stippling denotes trends statistically significant at the 95% confidence level.
Figure 2.12. Time series of the annual mean 500-hPa geopotential height ($Z_{500}$) differences between the six reanalyses listed at the bottom of the plot and ERA-Interim. The differences are spatially averaged (area-weighted) over Antarctica (a) and the Southern Ocean poleward of 50$^\circ$S (b).
Figure 2.13. Difference between reanalysis and observed values at Casey Station for (a) mean annual mean sea level pressure in hPa, and (b) mean annual 500-hPa geopotential height in meters. The station’s location is shown in the inset map in the top panel.
Chapter 3: The Byrd Temperature Record and the Warming of Central West Antarctica

3.1 Introduction

3.1.1 Climate Change Uncertainties in West Antarctica

Glacier acceleration along the Amundsen Sea coast (Rignot, 2008) (see location in Fig. 3.1) has been responsible for the increasing mass loss from the West Antarctic Ice Sheet (WAIS) in recent years (King et al., 2012). This has raised concerns about the present and future state of the WAIS, given its known potential instability in a warmer climate (Joughin and Alley, 2011). Some of the key mechanisms behind this acceleration have been identified as the melting and thinning of the floating ice shelves triggered by a warmer ocean (Jacobs et al., 2011; Pritchard et al., 2012). In comparison, it is still a matter of debate whether the atmosphere above the WAIS has warmed significantly over the last few decades, especially since the 1957–58 International Geophysical Year (IGY), the start of the instrumental period in Antarctica (Steig et al., 2009; O’Donnell et al., 2011; Schneider et al., 2012a). Unlike Greenland, where the extent of surface melting has grown dramatically (Tedesco et al., 2011), West Antarctica has not shown any unequivocal signs of atmospheric warming.

5This chapter is adapted from the main text and supplementary material of Bromwich, D. H., J. P. Nicolas, A. J. Monaghan, M. A. Lazzara, L. M. Keller, G. A. Weidner, and A. B. Wilson, 2013: Central West Antarctica among the most rapidly warming regions on Earth. Nature Geoscience, 6, 139–145, and from the Corrigendum to the paper published in January 2014 in the same journal.
warming (Tedesco and Monaghan, 2009; Kuipers Munneke et al., 2012). One important question is, therefore, whether West Antarctic temperatures have, indeed, not changed significantly (or even decreased) since the 1950s; or whether they have increased but not so much as to reach the melting point at the surface. In other words, could the WAIS be on the verge of becoming like Greenland? If so, is the exceptionally warm summer month of January 2005, when widespread surface melting occurred over a large portion of the WAIS (Nghiem et al., 2005) (see Fig. 3.2), an early manifestation of this transition?

3.1.2 A Key yet Challenging Temperature Record: Byrd

Assessing Antarctic climate change on time scales of a few decades is a well-recognized challenge due to the paucity of surface observations. Accordingly, statistical methods have been utilized to reconstruct Antarctic near-surface temperatures by interpolating the sparse meteorological records available since the IGY (Chapman and Walsh, 2007; Monaghan et al., 2008a; Steig et al., 2009; O’Donnell et al., 2011). These reconstructions have produced contrasting, and sometimes contradictory, temperature trends over West Antarctica (this issue will be further discussed in Chapter 4). This is not surprising as, in this region, the reconstructions can only rely on incomplete observations from a single site: Byrd Station (80°S, 120°W; see Fig. 3.1). Furthermore, West Antarctica is to a large extent climatologically distinct from the rest of the continent, especially with greater influence from the Tropics (Guo et al., 2004; Bindschadler, 2006; Ding et al., 2011), so that its climate variability and trends are not necessarily well reflected in peripheral temperature records. On the other hand, Byrd’s central location in West Antarctica, combined with the relatively flat terrain of the polar plateau (see photograph in Fig. 3.3b), account for the broad spatial footprint of its temperature observations illustrated in Fig. 3.1 and emphasizes the critical importance of this record for this data-sparse region.
3.1.3 Purpose and Outline of the Analysis

In this chapter, a new reconstruction of the Byrd temperature record is presented. The main goals are to document important revisions to the original observations, reduce the uncertainty of previously-used infilling methods (Shuman and Stearns, 2001; Reusch and Alley, 2004; Monaghan et al., 2008a; Küttel et al., 2012), and ultimately provide new insight into temperature changes over a significant portion of West Antarctica. The chapter begins with the two preliminary steps that were necessary to arrive at a complete temperature data set: correcting the observations (Section 3.2) and filling in the missing data (Section 3.3), combined with an uncertainty analysis (Section 3.4). Section 3.5 describes the temperature variability and trends inferred from the reconstructed record, places them in the context of regional and global temperature changes, and discusses their consistency with the recent literature. Section 3.6 follows with a discussion of the causal mechanisms (or “forcings”) behind the temperature changes seen at Byrd. Concluding remarks are given in Section 3.7.

3.2 Overhaul of the Byrd Temperature Record: History, Corrections, and Reconstruction

The Byrd record itself is a composite of observations from the initial staffed station (1957–1975), and from the subsequent automatic weather station (AWS) (1980–present). The reliability of the temperature observations is of course a central premise before attempting to “reconstruct” of the data gaps. Therefore, the first step was to verify the quality and temporal consistency of the reported temperatures throughout the five decades spanned by the Byrd record. This work involved a fair amount of detective work and resulted in various types of corrections, which are described in this section and in Appendix A.
3.2.1 Data Sources and Historical Background

3.2.1.1 Source of Temperature Observations

Monthly mean temperature observations from Byrd Station (1957–1975) and Byrd AWS (1980–2011) are obtained from the Antarctic READER archive (Turner et al., 2004). For Byrd AWS, the original source of the observations is the AMRC, which provides 3-hourly quality-controlled data to READER. READER’s monthly temperatures are utilized in the reconstruction provided that, for each month, ≥90% of the 6-hourly data are available and no block of data is missing for more than two consecutive days (these criteria are identical to those used by Turner et al. (2004)).

3.2.1.2 Byrd Since 1957: From the Staffed to the Automatic Weather Station

A historical overview helps understand the characteristics of the Byrd temperature record and the challenges involved in the reconstruction. The first occupied Byrd Station was established in January 1957 as a year-round research facility. In February 1962, a second station (“New Byrd”) was commissioned, 10 km from the original site (“Old Byrd”, abandoned) and slightly higher in elevation (1533 m versus 1511 m) (Schwerdtfeger, 1970; Stewart, 1990). Old and New Byrd provided uninterrupted near-surface temperature observations from 1957 to 1969. In the early 1970s, the facility transitioned into a summer-only station, known as Byrd Surface Camp (BSC), which has since been occupied intermittently during the Antarctic field season. Aiming to continue the existing temperature record, a redesigned Stanford AWS was installed in 1979, 1.4 km from BSC, 1530 m a.s.l., with the qualified Byrd AWS data record beginning as of January 1980 (Stearns and Savage, 1981; Schwerdtfeger, 1970; Stewart, 1990; Stearns and Savage, 1981).

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6http://www.antarctica.ac.uk/met/READER/
7ftp://amrc.ssec.wisc.edu/pub/aws/antrdr/
Lazzara et al., 2012). Since then, the AWS has been maintained by the Antarctic Meteorological Research Center (AMRC) at the University of Wisconsin-Madison, as part of the U.S. Antarctic AWS Program (Lazzara et al., 2012). A GPS survey of the area (Hamilton and Spikes, 2004) carried out in the early 2000s did not suggest any significant change in the AWS elevation since 1980. The proximity of the successive locations and the small elevation differences between them suggest little impact of the moving of the measurement site on the long-term temperature trends.

The challenges of operating autonomous systems in the Antarctic (long polar night, harsh climate, limited window for maintenance during austral summer) have resulted in numerous interruptions in the AWS observations. Their remarkable continuity through most of the 1980s is attributed to the radioisotope thermo-electric generator (RTG) originally used as a power source for the AWS (Lazzara et al., 2012). The RTG was subsequently replaced with batteries charged (only during the polar day) by solar panels, making the AWS more prone to power outages, particularly during austral winter and spring (see photographs of the Byrd AWS in Fig. 3.3).

3.2.2 Revision of the Temperature Observations

3.2.2.1 Corrections to Byrd Station’s Observations (1957–1975)

As mentioned above, the monthly mean temperatures compiled by the Antarctic READER Project constitute the 1957–1975 portion of the Byrd record. Whenever possible, READER uses 6-hourly data to compute the monthly means in an effort to produce consistent meteorological time series for Antarctica. Because six-hourly data were not found initially for Byrd for 1957–1975, the monthly means reported by READER for this period were taken from the previous compilations of Antarctic temperatures from Jacka et al. (1984)\(^8\) and

\(^8\)http://staff.acecrc.org.au/~jacka/temperature.html
Jones and Reid (2001)\(^9\), itself based upon monthly reports published in the *Monthly Climatic Data for the World* or the *World Weather Records*. One concern about these reports was the lack of information regarding the methodology used to derive the monthly means.

Through my investigations, I came across previously unused sub-daily (3-hourly and/or 6-hourly) meteorological observations from Byrd Station from the 1957–1975 period on the website of the National Climatic Data Center\(^10\). This data set has been available for an unknown number of years but has—to my knowledge—not been used in any study published over the past two decades. It was in particular overlooked in version 1.0 of the reconstructed Byrd record that provided the basis for the analysis presented in Bromwich et al. (2013).

The new set of observations made it possible to recalculate the monthly mean temperatures from 1957–1975 in a manner consistent with the more recent portion of the Byrd record. For October through March, the recalculation yielded higher temperatures (by 0.3 °C on average) than those previously published by READER\(^11\), with virtually no effect (< 0.1 °C) during the rest of the year. These discrepancies likely stem from the use of daily minima and maxima for the monthly reports published in the 1950s–1970s, in conjunction with the effects of the diurnal cycle of air temperature during the sunlit part of the year. The work on the 6-hourly data revealed complicated stories (e.g., error in units) behind some of the original monthly temperature values. Details are given in Appendix A.

### 3.2.2.2 Corrections to Byrd AWS’s Observations (1989–2011)

Another type of correction—this time for the data recorded by the Byrd AWS—resulted from the work of Matthew Lazzara, Linda Keller, and George Weidner carried out at the

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\(^9\)http://cdiac.esd.ornl.gov/epubs/ndp/ndp032/ndp032.html

\(^{10}\)http://www.ncdc.noaa.gov/

\(^{11}\)READER was informed of this discovery in March 2013 and its database was updated shortly thereafter.
AMRC in Madison, Wi. During the 2010–2011 Antarctic field season, a new CR1000 datalogger (used to record and disseminate the readings from the various AWS sensors) was installed on the Byrd AWS in replacement of the AWS-2B electronic system used since 1989 (the replacement occurred on 18 January 2011). Upon inspection of the old system at the AMRC, a calibration error of 1.5 °C (in excess) was identified for the temperature observations recorded since 2002. In addition, subsequent testing in a newly available cold chamber at the AMRC provided more accurate measurements of the temperature sensitivity of the AWS-2B system, which results in a negative temperature drift as the temperature decreases. As a result, corrections were made to the temperature observations recorded by the Byrd AWS between 1989 and 17 January 2011. The release of the corrected observations on the AMRC’s ftp server in December 2011 was followed by an update of the Byrd AWS data set on the READER archive.

The effect of the corrections on the reconstructed temperatures is illustrated in Fig. 3.4. In the lower temperature range (typically -50 to -30 °C), the temperature drift largely compensated for the 1.5 °C error in the 2002–2010 observations, whereas no compensation occurred at higher temperatures (-20 to 0 °C). This explains the differing impact of the corrections between summer and winter, and between the 1989–2001 and 2002-2010 periods. It is noteworthy that the temperature drift problem did not affect significantly the AWS observations from 1980–1988, and this for two reasons: (i) Excess power from the RTG was used to keep the internal temperature of the electronics above -20 °C. This extra power was no longer available when the AWS started relying on batteries charged by solar panels. And (ii) The central processing unit of the AWS was (paradoxically) a newer version than the one subsequently used from 1989 onward.
3.2.2.3 Other Potential Sources of Error

Two other possible types of error are not accounted for in the Byrd AWS observations. One stems from the non-adjustment of the temperatures to height changes due to snow accumulation. This can be an important issue because of the strong surface temperature inversion characteristic of the Antarctic boundary layer (Ma et al., 2011). The main reason why this problem is likely inconsequential here is that it would actually cause the observations to underestimate the temperatures compared to a properly calibrated record. In addition, the sensors are periodically raised on the AWS mast during maintenance visits.

A second concern has been suggested by a recent study (Genthon et al., 2011) showing AWS temperature measurements in the East Antarctic interior to be biased warm in summer because of intense solar radiation and insufficient ventilation of the sensor, particularly during low-wind conditions. Again, several lines of evidence suggest that this problem is likely of secondary importance for the Byrd AWS: (i) During summer, the near-surface wind speed at Byrd is typically 1.5 times higher than on the adjacent East Antarctic plateau (based upon monthly mean wind speeds from the READER archive). (ii) Cloudiness is greater at Byrd Station (Turner and Pendlebury, 2004), reducing the amount of incoming solar radiation at the surface. And (iii) The occurrence of warmer-than-normal summer months in the 1990s and 2000s was corroborated by surface melt episodes in the vicinity of Byrd (Tedesco et al., 2007), most notably in January 2005 (Nghiem et al., 2005). (iv) Finally, the marked increase in the mean summer temperature at Byrd during the 1980s is consistent with changes in the atmospheric circulation, as shown in Figs. 3.16 and 3.17.
3.3 Reconstruction of the Byrd temperature record (1957–2012)

Global atmospheric reanalyses are, in principle, uniquely suited for filling in the data gaps in the Byrd record. By synthesizing a wide range of historical observations with the fields from a state-of-the-art atmospheric model, these reanalyses provide a “best possible” representation of the state of the atmosphere, complete both in space and time. Yet, the quality of their temperature estimates has long remained inadequate in Antarctica (Reusch and Alley, 2004; Marshall, 2002; Bromwich and Fogt, 2004), prompting the use of alternative reconstruction methods (Shuman and Stearns, 2001; Reusch and Alley, 2004; Monaghan et al., 2008a; Küttel et al., 2012). Compared to other reanalyses, the ERA-Interim Reanalysis (Dee et al., 2011) turned out to be predicting the near-surface temperature at Byrd with markedly greater skill—even without the constraint of Byrd AWS observations. Therefore, despite the caveats generally associated with reanalysis data (see Chapters 2 and 4 of this dissertation), ERA-Interim was deemed appropriate for the data infilling (with certain important caveats). The uncertainties of the reconstruction are admittedly larger during 1970-1978 owing to the lower reliability of the reanalysis data sets available for this period and the almost complete absence of Byrd observations.

3.3.1 Infilling of the Data Gaps from 1979 Onward

3.3.1.1 Primary Infilling Method

For the period starting in January 1979, monthly mean 2-meter temperature ($T_{2m}$) data from ERA-Interim are used to fill in the gaps in the Byrd record. To avoid a circularity problem between predictor and predicted variable (ERA-Interim does assimilate observations from Byrd AWS), the $T_{2m}$ data are taken from ERA-Interim forecast fields rather than from the analysis fields (see Section 3.3.1.3 for details about this distinction). For months
without any observations, ERA-Interim monthly $T_{2m}$ is adjusted to the monthly mean temperature observations available before and after the months in question (see the schematic description of the algorithm in Fig. 3.5). This is done by estimating the reanalysis-versus-observations bias in a multiyear moving window (separately for each month of the year). The width of the window (5–9 years) depends on the number of observations available (a minimum of two observed monthly mean temperatures is required). This empirical adjustment technique accounts for the model-versus-observed elevation difference (or any other systematic bias), and prevents the reconstruction from being contaminated by spurious trends in ERA-Interim.

### 3.3.1.2 Secondary Infilling Method

A variant of the method is used when observations are partially available for a given month. In this case, the infilling relies upon 6-hourly temperature observations from the AWS (resampled from the 3-hourly data obtained from the AMRC ftp server); and ERA-Interim 6-hourly $T_{2m}$ data (from the forecast fields). The reconstructed monthly mean temperature at Byrd ($T_{rec}$) is estimated, for each month, by (i) computing the difference ($\Delta T$) between all available 6-hourly AWS observations and the corresponding 6-hourly $T_{2m}$ data from ERA-Interim; and (ii) using $\Delta T$ to adjust ERA-Interim monthly mean $T_{2m}$. Unsurprisingly, the error of $T_{rec}$ increases as the percentage of observations available per month decreases (Fig. 3.6). In other words, the fewer the observations available, the less $\Delta T$ becomes representative of the ERA-Interim temperature bias for the entire month. The cut-off percentage for this adjustment method to be used is set to 40% as it ensures
RMSE $\leq 1^\circ$C. This alternative method is used only for 11% of the months with incomplete/missing observations during 1979–2011. It is primarily designed to take advantage of the few September–October months with observations in the late 1990s and 2000s\textsuperscript{12}.

3.3.1.3 ERA-Interim Forecast versus Analysis Temperatures

When using ERA-Interim $T_{2m}$ estimates, one caveat arises from the use of temperature observations from Byrd AWS in the reanalysis itself through its data assimilation system (Uppala et al., 2005; Dee et al., 2011). While this forces the reanalysis model to remain close to the observations, it also leads to an underestimation of the model error in the absence of observations, i.e., precisely when ERA-Interim data are needed for the temperature reconstruction. As a result, we used ERA-Interim forecast $T_{2m}$ data (the products of the reanalysis model short-term forecasts) were preferred over analysis $T_{2m}$ data (produced by combining the forecast (a.k.a. “first-guess”) fields with all other available observations). The forecast $T_{2m}$ time series are constructed from the 6th and 12th hours of the forecasts initialized at 0000 UTC and 1200 UTC.

The use of the forecast temperature is further supported by the fact that, in ERA-Interim, the assimilation of near-surface observations is performed separately from the main (4D-Var) data assimilation system, using an Optimal Interpolation (OI) scheme (Simmons et al., 2010; Dee et al., 2011). The analysis $T_{2m}$ generated by the OI is not used to modify the temperature field which is carried on to initialize the short-term forecasts (Simmons et al., 2010). In other words, ERA-Interim forecast $T_{2m}$ estimates at Byrd are not affected by the AWS observations, even when these are available. This condition of independence between model estimates and observations is essential for estimating, and correcting for,

\textsuperscript{12}Austral spring is the season with the largest data gaps in the Byrd AWS record
the “true” reanalysis bias in the infilling method. As one would expect, the RMSE of ERA-Interim forecast $T_{2m}$ (1.27°C) is greater than that of the analysis $T_{2m}$ (1.01°C). However, this difference is small, and the forecast $T_{2m}$ RMSE is substantially lower for ERA-Interim than for the other reanalyses (Fig. 3.7).

### 3.3.2 Infilling of the Data Gaps prior to 1979

For 1970-1978, two methods were employed to estimate Byrd temperatures:

1. Monthly mean temperature observations from Antarctic research stations with records going back to the 1950s were interpolated to the location of Byrd using a kriging method previously employed to reconstruct Antarctic temperatures (Monaghan et al., 2008a).

2. The monthly $T_{2m}$ fields from the NCEP-NCAR Reanalysis (Kalnay et al., 1996) (NNR) and the ERA-40 Reanalysis (Uppala et al., 2005) were used: NNR for the fall-winter-spring months, and ERA-40 for the summer months, because of unrealistic temporal changes in the NNR temperatures in summer during the 1970s. The reanalyses’ temperatures were adjusted to the observations (separately for each month) based upon the mean bias during 1957-1975.

The final temperature values used in the reconstruction are taken as the average of the results from methods 1 and 2. The curves generated with the two methods are plotted separately in Fig. 3.4.

### 3.4 Uncertainty Calculation

#### 3.4.1 Uncertainties of the Reconstructed Temperatures

The uncertainties of the temperature reconstruction are assessed separately for the 1970–1978 and 1979–2011 periods. For 1979–2011, monthly mean temperatures ($T_{rec}$) predicted
by the ERA-Interim-based infilling method (see Section 3.3.1) can be directly compared against AWS observations. As shown in Fig. 3.8a, the $T_{rec}$ estimates from 1979–2011 are characterized by an RMSE of 0.87°C. The smallness of the RMSE can be appreciated when compared against the standard deviations of the monthly temperatures at Byrd (Fig. 3.8d). The biases of $T_{rec}$ with respect to the observations do not exhibit any particular temperature dependence. Their relatively uniform distribution around zero (Fig. 3.8c) explains the negligible overall bias (0.01°C). This also results in a smaller RMSE (0.62°C) for the mean seasonal temperatures (Fig. 3.8b). It must be noted that the derivation of such RMSE statistics is possible because the temporal gaps in the observations are limited in extent and the skill of ERA-Interim relatively homogeneous throughout the period.

Such conditions are not present during the 1970-1978 period. Indeed, with the exception of a few summer months, temperature observations are completely absent during this time interval (see Fig. 3.9, histograms). In addition, the quality of the reanalyses (ERA-40, NNR) is highly variable in time. It increases most significantly with the introduction of abundant satellite observations in 1979 (Bromwich and Fogt, 2004). Thus, meaningful RMSE statistics cannot be derived for this period. As a result, the envelope of uncertainties around $T_{rec}$ is defined by the two temperature curves generated with the two infilling methods employed for this portion of the Byrd record (see Section 3.3.2). For each month without observations, the error of $T_{rec}$ is taken as half the difference between the temperature estimates suggested by the two methods. As shown in Table 3.1, the temperatures from 1970–1978 have a minimal impact on the 1958–2010 temperatures trends.

3.4.2 Uncertainties of the Temperature Trends

The uncertainties of the temperature trends include both the standard errors of the regressions and the uncertainties of the infilling methods. For the standard error, the sample
size, \( n \), used is adjusted for autocorrelations in the time series as described in Santer et al. (2000). The resulting effective sample size, \( n_{\text{eff}} \), is defined as \( n_{\text{eff}} = n(1 - r_1)/(1 + r_1) \), where \( r_1 \) is the lag-1 autocorrelation.

The trend uncertainties due to the infilling are estimated via a Monte Carlo procedure (Von Storch and Zwiers, 1999). A large number (\( N = 10^5 \)) of realizations of the temperature reconstruction are generated by randomly varying the temperature estimates within plus or minus two RMSE around \( T_{\text{rec}} \) for months without observations (for 1970-1978, temperatures are randomly selected within twice the range of uncertainties defined above). The linear trends that are derived from each realization are normally distributed around a mean trend value. The uncertainties of the trends for the monthly, seasonal and annual mean temperatures are taken as two standard deviations of the respective normal distributions. Note that this second type of uncertainty is one order of magnitude smaller than the standard errors of the regressions.

3.5 Analysis of the Reconstructed Byrd Temperature Record: Strong Warming and Seasonal Contrasts

Sections 3.2 to 3.3 have described the various steps taken to obtain a robust and complete temperature record for Byrd. The analysis that follows is based on version 2.0 of the data set\(^{13}\), which was presented in Bromwich et al. (2014). I discuss what the reconstructed time series tell us about the temperature variability and trends in central West Antarctica and how these compare to what was reported in previous studies.

3.5.1 Temperature Trends at Byrd Station

The reconstructed Byrd record is presented in Fig. 3.9 as annual and seasonal mean temperature time series from 1957 to 2011. For the annual mean temperature, the linear

\(^{13}\)The complete temperature data set is available online at http://polarmet.osu.edu/Byrd_recon/.
trend calculated for 1958–2010 reveals an average warming of 0.42±0.24 °C per decade, statistically significant at the 99% confidence level (CL) (Fig. 3.10a), which translates into a temperature increase of 2.18±1.25 °C in 52 years. This warming is close to that measured at Faraday/Vernadsky, on the western coast of the Antarctic Peninsula (0.58±0.31 °C per decade), a site already known for its rapid atmospheric warming (Vaughan et al., 2003). The temperature trends at Byrd and Faraday are substantially greater than the global average (0.13±0.03 °C per decade; Hansen et al. (2010)) and comparable in magnitude to the warming observed over land in the Northern Hemisphere high latitudes (Fig. 3.10c).

Seasonally, the temperature trends at Byrd are positive throughout the year, yet only statistically significant in austral winter (JJA) and spring (SON) (Fig. 3.10a). SON exhibits the largest temperature trend (0.75±0.40 °C per decade) and highest significance level (99%), followed by JJA (0.52±0.50 °C per decade, 95% CL). The trend in DJF (0.30±0.27 °C per decade) is only marginally significant (89%). However, and importantly, the summertime warming is maximum and exceeds the 95% CL in December–January, the two climatologically warmest months of the year at Byrd and peak of the melting season in Antarctica.

Signs of interdecadal variability are evident in Byrd’s annual and seasonal temperature time series. On the annual scale, most of the warming at Byrd appears to have occurred during the mid- to late 1980s, with temperatures seemingly leveling off since the early 1990s (Fig. 3.9a). Both as a result of slower temperature increases and large interannual variability (particularly marked in SON in the latter part of the record), none of the trends attain statistical significance during 1980-2010 (Fig. 3.10b).

3.5.2 Comparison with Other Temperature Reconstructions

The temperature trends estimated from the reconstructed Byrd record (referred to below as the “new record”) are contrasted with those derived from four sets of Antarctic
temperature reconstructions (Chapman and Walsh, 2007; Monaghan et al., 2008a; Steig et al., 2009; O’Donnell et al., 2011) for 1958–2005 (Fig. 3.11) and 1958–2001 (Table 3.2). The new record shows (almost systematically) stronger warming than all other datasets although, because of the error bars, the various trend estimates are not always statistically distinguishable from one another.

There is overall agreement among the reconstructions on greatest seasonal warming occurring in SON (statistically significant only in the new record and Steig et al. (2009); Monaghan et al. (2008a)), which corroborates the conclusions from a recent investigation of the West Antarctic warming during this season (Schneider et al., 2012a). However, the new record and Monaghan et al. (2008a) indicate temperature trends at least twice as large as the other reconstructions. There is overall agreement on the lack of statistically significant warming in DJF. It is noteworthy that the positive trends in this season present in all datasets are at odds with the marked tropospheric cooling seen in Microwave Sounding Unit (MSU) observations available since 1979 (Johanson and Fu, 2007). In MAM and JJA, the new record agrees relatively well with Steig et al. (2009) for the 1958–2005 period. Both find JJA the second fastest-warming season (yet without statistical significance), which confirms other evidence of surface and tropospheric winter warming over West Antarctica (Turner et al., 2006; Johanson and Fu, 2007; Ding et al., 2011); and both show very similar temperature trends in MAM (significant only in Steig et al. (2009)). In these two seasons (MAM and JJA), the other reconstructions (Chapman and Walsh, 2007; Monaghan et al., 2008a; O’Donnell et al., 2011) have substantially smaller trends (or even negative values in O’Donnell et al. (2011)).

Annually, a pronounced warming in West Antarctica in recent decades has also been detected in recent borehole temperature measurements (Barrett et al., 2009; Orsi et al.,
2012), especially at the WAIS Divide drilling site, 160 km northeast of Byrd. The WAIS Divide record (Orsi et al., 2012), in particular, suggests a warming occurring later than at Byrd (early to mid-1990s) and continuing into the 2000s, instead of flattening out. Although this may reveal greater spatial heterogeneity in the temperature changes than inferred from Byrd temperature footprint (Fig. 3.1), the discrepancy may also be related to the non-linear temporal smoothing inherent to borehole temperature retrievals.

3.6 Mechanisms Responsible for the Seasonal Temperature Trends at Byrd

An entire dissertation could certainly be dedicated to investigating the mechanisms responsible for the warming, and more generally the temperature variability, of West Antarctica. One could point out that addressing the “why” may still be premature as there is still large uncertainty about the “what happened” (i.e., how much temperature change? where? etc.) in this extremely data-sparse region. Nonetheless, the primary purpose of this section is to examine whether the temperature changes seen in the reconstructed Byrd record are consistent with recently proposed mechanisms for the West Antarctic warming.

3.6.1 Investigation of the Winter and Spring Warming

The causes of the West Antarctic warming in JJA and SON have been investigated in two recent studies (Ding et al., 2011; Schneider et al., 2012a) that have highlighted, in particular, its linkage to lower-latitude sea surface temperature (SST) changes. Is the temperature variability observed at Byrd consistent with these findings? And can they also explain the warming in austral summer?

In JJA, the warming has been associated with an increase in geopotential heights over West Antarctica (Ding et al., 2011) (Fig. 3.12a). This pattern has promoted onshore winds (warm advection) to Marie Byrd Land and is consistent with the spatial correlations between
Byrd temperature and the 500 hPa geopotential height ($Z_{500}$) field during 1979–2009 (Fig. 3.12d). The absence of statistically significant trends in $Z_{500}$ over the area can be explained by relatively little change since the early 1990s (Fig. 3.12g). The higher geopotential heights observed over West Antarctica have been described as being part of an atmospheric Rossby wave train forced by higher SST in the central tropical Pacific (Ding et al., 2011). The signature of this wave train is clearly apparent in the correlations with Byrd temperature (Fig. 3.12d), more in JJA than in any other season. This atmospheric teleconnection is manifested in the second mode of covariability between tropical SST ($20^\circ$S–$20^\circ$N) and Southern Hemisphere atmospheric circulation (Ding et al., 2011). Notably, it involves an SST forcing distinct from the traditional eastern equatorial Pacific ENSO region (Ding et al., 2011), yet consistent with the increasing frequency of El Niño events with SST anomalies in the central Pacific (Lee and McPhaden, 2010).

In SON (and DJF), the trends in $Z_{500}$ project more poorly onto the spatial correlations than JJA (Figs. 3.12b and 3.12e), suggesting a more complex causality of the warming. In other words, the mechanisms accounting for the secular temperature trends may differ (in part) from those responsible for the interannual temperature variability. In SON, the West Antarctic warming has been primarily attributed to lower geopotential heights in the South Pacific that have enhanced northerly warm air advection toward West Antarctica (Schneider et al., 2012a) (Fig. 3.12b). This change in the atmospheric circulation is congruent with the trends in the two modes of high-latitude atmospheric variability often associated with ENSO, and known as the Pacific South American (PSA) modes (Mo, 2000; Schneider et al., 2012a). These trends, however, have been linked to positive SST anomalies in the tropical branch of the South Pacific Convergence Zone (SPCZ) (Ding et al., 2011) and appear to be more clearly distinct from ENSO variability (either of the eastern or central Pacific.
type) compared to JJA (Ding et al., 2011). The close parallel between SST anomalies in the subtropical SPCZ region and West Antarctic temperatures over the past 50 years (Schneider et al., 2012a) suggests a more southern location of the relevant SST forcing than indicated in Ding et al. (2011). This hypothesis is confirmed by conducting with a covariance analysis between SST and 200 hPa geopotential heights \( (Z_{200}) \) similar to that presented in Ding et al. (2011), but encompassing SSTs beyond the tropical latitudes (Fig. 3.13 and 3.14).

The link between West Antarctic warming and (sub)tropical SST anomalies has not been established convincingly with model sensitivity experiments in SON (Ding et al., 2011; Schneider et al., 2012a), in contrast to JJA (Ding et al., 2011), suggesting again that other mechanisms may be at play. Remarkably, the record-high temperatures at Byrd in the mid-1990s through late 2000s occurred along with a marked increase in the interannual temperature variability, which is also well reflected in the \( Z_{500} \) time series over the Bellingshausen Sea sector (Fig. 3.12h). This increased variability can be related to the greater in-phase behavior between the SAM and ENSO observed since the early 1990s (Stammerjohn et al., 2008; Fogt et al., 2011). It also explains the small \( Z_{500} \) trends over the Bellingshausen Sea region (Fig. 3.12b), by compensation of large anomalies of opposite signs. Importantly, the warmest SON at Byrd (in 2002) coincided with an exceptional and well-documented sudden stratospheric warming over Antarctica, following an early breakdown of the polar vortex (Shepherd et al., 2005). Yet, there is no evidence of such phenomenon in other abnormally warm springs at Byrd (e.g., 2000, 2001, 2005). Thus, one can assume that this mechanism is likely not a significant driver of the long-term SON warming, which is further supported by the fact that the Antarctic polar vortex tended to break up later in the 1990s than in the 1960s (Haigh and Roscoe, 2009).
3.6.2 Hypotheses for the Summer Warming

As mentioned above and shown in Fig. 3.10, the 1958–2010 temperature trend in DJF is only marginally significant. It was, however, significant above the 95% CL in version 1.0 of the reconstruction used by Bromwich et al. (2013) and, as such, was among the findings highlighted in the study. I have intentionally kept the discussion below as it was in the paper as it provides some useful background and potential research directions about the temperature variability in a season that is rarely discussed in association with West Antarctica.

Because of the relative novelty of a West Antarctic warming in DJF, little has been said about its possible attributions. On the contrary, studies have generally emphasized the cooling effect of recent atmospheric circulation changes for West Antarctica in austral summer (Kwok and Comiso, 2002; Gillett and Thompson, 2003; Van den Broeke and van Lipzig, 2004). Largely dominated by the positive trend in the SAM index (Marshall et al., 2004), these changes have been characterized by lower geopotential heights over the continent (Thompson and Solomon, 2002) (Fig. 3.12c) and stronger circumpolar westerlies, reducing the meridional heat exchange. This negative relationship between the strength of the SAM and West Antarctic temperatures is apparent in the Byrd temperature-$Z_{500}$ correlations calculated for 1989–2011 (Fig. 3.15), but mostly vanishes when the period is extended to 1979 (Fig. 3.12f). This supports the fact that the observed strengthening of SAM is inconsistent with a summer warming at Byrd. Moreover, it is clear from Fig. 3.12i that the changes in $Z_{500}$ over the Bellingshausen Sea region during the 1980s fail to explain the quasi-stepwise increase in Byrd DJF temperature around 1986–1989.

This late 1980s warming turns out to be consistent with a westward shift and deepening of the persistent center of low pressure over the Amundsen/Ross Sea sector (see Section 3.6.3
and Figs. 3.16 and 3.17). The resulting anomalous northerly warm advection toward Byrd was further enhanced by the adiabatic warming of air masses descending onto the lee side of Marie Byrd Land’s coastal mountain ranges (foehn effect) (Nicolas and Bromwich, 2011a). It appears, however, that the position of the low—known to be influenced by ENSO (Bertler et al., 2004)—does not exhibit any significant trend over the post-1979 period (Turner et al., 2013) and, therefore, cannot account by itself for the long-term summer warming at Byrd. It is also noteworthy that the warming at Byrd coincided with a sharp decrease in summer sea-ice concentrations in the Bellingshausen Sea in 1989, which resulted from the permanent loss of multiyear sea ice (Jacobs and Comiso, 1997). There is, however, no clear evidence of a linkage between the two phenomena.

The SST region potentially linked to Byrd summer warming cannot be identified in a straightforward manner, as exemplified by the two strongly contrasting SST anomaly patterns associated with peaks in Byrd temperature in DJF 1997–1998 and DJF 2005–2006 (Fig. 3.9 and Fig. 3.18). As in the other seasons, the second mode of covariability between (sub)tropical SST and Southern Hemisphere atmospheric circulation best captures the observed SST trends, in particular the warming of the subtropical SPCZ region (Ding et al., 2012) (Figs. 3.13 and 3.14). This mode, which is separate from ENSO variability (as in SON), is thus likely to reflect the SST forcing responsible for long-term temperature trends at Byrd. The patterns of anomalies associated with this second mode show anomalous northerly winds over the Byrd region occurring in conjunction with higher SSTs over the subtropical SPCZ region (Fig. 3.13). With this impact on the atmospheric circulation, the warming of the subtropical SPCZ region may have at least mitigated the cooling induced by a stronger SAM, at most contributed to abnormally high temperatures at Byrd as seen in DJF 2005–2006 (Fig. 3.18).
3.6.3 Warming at Byrd in the 1980s: Contribution of the Atmospheric Circulation over the Ross-Amundsen Seas

The rapid 1980s warming at Byrd in DJF coincided with a change in the position of the persistent center of low pressure off the coast of West Antarctica (Fig. 3.16). Often termed Amundsen-Sea low (Turner et al., 2013), it resides predominantly over the Ross Sea in the non-summer months and tends to migrate towards the Bellingshausen Sea during austral summer (Bromwich and Wang, 2008). The variability of its position is markedly greater in DJF than during the rest of year, which is apparent in the time-averaged geopotential height field (Fig. 3.16a). A low positioned in the Ross Sea favors greater meridional heat transport toward Marie Byrd Land, and therefore warmer conditions at Byrd (Nicolas and Bromwich, 2011a). A transition occurred in 1989 when the low started residing predominantly in the Ross Sea in DJF, until roughly the mid-1990s (Fig. 3.16b). This transition coincided with the SAM starting to shift toward its high polarity and with a slight increase (6%) in the number of cyclones in the broad Amundsen-Bellingshausen Sea region (Fogt et al., 2012b). Surprisingly, further (and sharper) increase in the SAM index in the mid-1990s did not have any clear effect on the position of the low. Moreover, this position has not displayed any significant longitudinal trend in DJF since 1979 (Turner et al., 2013) (Fig. 3.16b). As a result, this mechanism alone is unlikely to account for the long-term summer warming. These results highlight the important role of the regional atmospheric circulation in the Amundsen-Bellingshausen Sea sector. This issue would certainly deserve further investigation that is beyond the scope of this dissertation.

3.7 Conclusions

The new reconstructed Byrd temperature record presented in this chapter reveals one of the most rapidly warming places on the planet since the 1950s, and its spatial footprint (Fig. 3.16b).
3.1) indicates that similar change has likely occurred over a broad area of West Antarctica. These results underscore the importance of maintaining a robust observational network in the region (the area is currently devoid of upper-air observations), first and foremost at Byrd. The analysis confirms the strong seasonality in the characteristics of the warming previously found and suggests that not all causal mechanisms are yet fully identified or understood. The significant warming during austral summer reported here is a matter of concern because of its implications for the state of the West Antarctic Ice Sheet and its coastal ice shelves. Despite no evidence of greater frequency or intensity of surface melting in this region (Tedesco and Monaghan, 2009; Kuipers Munneke et al., 2012), the increase in the background summer temperature has, in effect, enhanced the probability of extensive melting events such as seen in January 2005 (Fig. 3.2). Although Byrd (1530 m a.s.l.) still has mean January temperatures around -10°C (at most), one must bear in mind that the vast portion of West Antarctica lies at lower elevation and is therefore more exposed to temperatures reaching 0°C.
Table 3.1. Linear temperature trends at Byrd computed for: the 1958–2010 and 1980–2010 periods using the fully reconstructed Byrd record (columns 2 and 3); the 1958–2010 period using the reconstructed Byrd record with no infilling of the missing data between 1970 and 1978 (column 4); and the 1958–2010 period using the uncorrected Byrd AWS observations available prior to December 2011 (column 5). Italic and bold fonts denote trends statistically significant at the 90% and 90% confidence level, respectively.

<table>
<thead>
<tr>
<th>Season or month</th>
<th>Linear temperature trends (°C per decade)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ANN</td>
<td>0.42±0.24</td>
</tr>
<tr>
<td>DJF</td>
<td>0.22±0.26</td>
</tr>
<tr>
<td>MAM</td>
<td>0.21±0.46</td>
</tr>
<tr>
<td>JJA</td>
<td>0.52±0.50</td>
</tr>
<tr>
<td>SON</td>
<td>0.75±0.40</td>
</tr>
<tr>
<td>DJ</td>
<td>0.26±0.24</td>
</tr>
<tr>
<td>Jan</td>
<td>0.33±0.32</td>
</tr>
<tr>
<td>Feb</td>
<td>0.08±0.58</td>
</tr>
<tr>
<td>Mar</td>
<td>0.40±0.70</td>
</tr>
<tr>
<td>Apr</td>
<td>-0.05±0.82</td>
</tr>
<tr>
<td>May</td>
<td>0.28±0.68</td>
</tr>
<tr>
<td>Jun</td>
<td>0.67±0.71</td>
</tr>
<tr>
<td>Jul</td>
<td>0.52±0.84</td>
</tr>
<tr>
<td>Aug</td>
<td>0.36±0.92</td>
</tr>
<tr>
<td>Sep</td>
<td>1.12±0.81</td>
</tr>
<tr>
<td>Oct</td>
<td>0.72±0.69</td>
</tr>
<tr>
<td>Nov</td>
<td>0.41±0.44</td>
</tr>
<tr>
<td>Dec</td>
<td>0.21±0.25</td>
</tr>
</tbody>
</table>
Table 3.2. Linear temperature trends at Byrd in °C per decade from the reconstructed record presented here (“This study”) and from several other Antarctic temperature datasets: Chapman and Walsh (2007); Monaghan et al. (2008a); Steig et al. (2009); and O’Donnell et al. (2011). The trends are computed for the 1958–2005 and 1958–2001 periods, respectively, to account for the different time spans of the datasets. Italic and bold fonts denote trends statistically significant at the 90% and 90% confidence level, respectively.

<table>
<thead>
<tr>
<th>Period</th>
<th>Season</th>
<th>Linear temperature trends (°C per decade)</th>
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</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>v1</td>
</tr>
<tr>
<td>ANN</td>
<td>-</td>
<td>0.31±0.25</td>
</tr>
<tr>
<td>1958</td>
<td>DJF</td>
<td>0.22±0.30</td>
</tr>
<tr>
<td></td>
<td>to MAM</td>
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</tr>
<tr>
<td>2005</td>
<td>JJA</td>
<td>0.47±0.58</td>
</tr>
<tr>
<td></td>
<td>SON</td>
<td>0.84±0.47</td>
</tr>
<tr>
<td>ANN</td>
<td>0.48±0.33</td>
<td>0.07±0.19</td>
</tr>
<tr>
<td>1958</td>
<td>DJF</td>
<td>0.23±0.34</td>
</tr>
<tr>
<td></td>
<td>to MAM</td>
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</tr>
<tr>
<td>2001</td>
<td>JJA</td>
<td>0.62±0.66</td>
</tr>
<tr>
<td></td>
<td>SON</td>
<td>0.84±0.47</td>
</tr>
</tbody>
</table>

*Steig et al. (2009): The ‘v1’ dataset is the main reconstruction of Steig et al. (2009), based on non-detrended satellite infrared observations. The ‘v2’ dataset is a second reconstruction presented in their Supplementary Information and based on detrended satellite infrared observations. The two datasets span 1957–2006 and are courtesy of E. Steig at http://faculty.washington.edu/steig/nature09data/data/.

*O’Donnell et al. (2011): The two temperature reconstructions are based, respectively, upon the regularized least square (RLS) and the eigenvector-weighted (E-W) methods. The two datasets span 1957–2006 and are courtesy of R. O’Donnell at http://www.climateaudit.info/data/odonnell/.


*Monaghan et al. (2008a): The revised version of their reconstructed Byrd temperature record is used here (see Küttel et al. (2012)). It relies on temperature observations from the Byrd AWS prior to the 2011 corrections.
Figure 3.1. Map of Antarctica and spatial footprint of the annual mean temperature at Byrd. The color shadings represent the Pearson’s coefficient of correlation ($r$) between the annual mean temperatures at Byrd and the annual mean temperatures at every other grid point in Antarctica. The correlations are computed using ERA-Interim 2-meter temperature time series from 1979–2011. The filled black circles denote the locations of permanent research stations with long-term temperature records.
Figure 3.2. Number of melting days over Antarctica in January 2005 derived from satellite passive microwave observations (Picard and Fily, 2006). The black star denotes the location of Byrd Station. The melt data are courtesy of Ghislain Picard (LGGE) at http://www-lgge.obs.ujf-grenoble.fr/~epicard/melting/. 
Figure 3.3. Photographs of the Byrd Station/AWS site at various stages of its history. (a) Staffed Byrd Station in September 1957 (courtesy of Wesley Morris). The thermometer shelter can be seen in the foreground. (b) Aerial view of Byrd Surface Camp in January 2010 (courtesy of Lee Thomas). (c) Byrd AWS in February 1980 with the radio-isotope thermoelectric generator (RTG) standing next to it (courtesy of AMRC-UW). (d) Byrd AWS in January 2011 (courtesy of Lee Welhouse/AMRC-UW).
Figure 3.4. Annual and seasonal mean temperature time series from the final reconstructed Byrd record (black curve) and from the following datasets: interpolated temperature observations from Antarctic research stations (blue curve); adjusted 2-meter temperature estimates from the ERA-40 (for DJF) and NNR (for the other seasons) reanalyses (green curve); and reconstructed Byrd record based upon uncorrected AWS observations (red curve).
Figure 3.5. Schematic describing the procedure used to adjust the monthly mean 2-meter temperatures from ERA-Interim and fill in the data gaps in the Byrd AWS record (see Section 3.3.1.1 for further details). The figure shows how the monthly mean temperature at Byrd in January 2005 (missing) is estimated based on the January mean temperatures from the two preceding and two following years.
Figure 3.6. Root mean square error (RMSE) of the monthly mean temperatures predicted by the secondary infilling method (see Section 3.3.1.2) (y-axis) as a function of the percentage of 6-hourly observations available per month from the Byrd AWS record (x-axis). The percentage of observations available each month is varied through random selection of 6-hourly data. Only the data from 1980–1988 (when the Byrd AWS record is most complete) are used to ensure a uniform sampling of the months.
Figure 3.7. Monthly mean 2-meter temperature ($T_{2m}$) estimates from five global reanalyses interpolated to Byrd’s location (y-axis) versus observed monthly mean temperatures from Byrd AWS (x-axis). The root mean square error (RMSE) of the reanalysis temperatures with respect to the observations, in units of $^\circ$C, is shown in the bottom-right corner of each plot. The reanalysis datasets shown are ERA-Interim (a,b); MERRA (c); CFSR (d); the NCEP-NCAR Reanalysis (e); and ERA-40 (f). For ERA-Interim, the analysis $T_{2m}$ (a) and the forecast $T_{2m}$ (b) are shown on separate plots. The reanalysis data are adjusted for the model-versus-observed elevation differences using a dry adiabatic lapse rate of 0.01 $^\circ$C m$^{-1}$.
Figure 3.8. (a) Monthly mean temperatures ($T_{rec}$) predicted by the primary infilling method (see Section 3.3.1.1) (y-axis) versus observed monthly mean temperatures from Byrd AWS (x-axis). The temperature data are from the time period covered by ERA-Interim (1979–2011). The comparison between predicted and observed temperatures is done only for months with ≥90% of observations available. The root mean square error (RMSE) of $T_{rec}$ with respect to the observations is shown in the bottom-right corner of the plot. (b) Same as (a) but for the seasonal mean temperatures. (c) Bias of the monthly $T_{rec}$ estimates with respect to the observations (y-axis) as a function of the observed temperature (x-axis). (d) Histograms comparing the RMSE of $T_{rec}$ (red bars) and the standard deviation of the monthly mean temperatures (gray bars) for each month of the year.
Figure 3.9. Temperature time series from the reconstructed Byrd record. Annual (a) and seasonal (b-e) mean temperature time series. Red markers denote the portions of the record for which $\frac{1}{3}$ of the observations are missing; black markers are used otherwise. The solid gray line represents the centered 5-year moving average temperature. The histograms (right vertical axis) show the number of monthly mean temperature observations available per year or per season. For summer (DJF), the year refers to January.
Figure 3.10. Linear temperature trends of the annual and seasonal mean temperature at Byrd during 1958–2010 (a) and 1980–2010 (b). The error bars denote the 95% confidence interval. Trends significant above (below) the 95% level are shown in red (gray). The trend values are given in Table 3.1. (c) Linear change in the annual mean temperature (i.e., trend × number of years) during 1958–2009 from the reconstructed Byrd record (red-filled black circle) and from the CRUTEM4 dataset (Jones et al., 2012) (background map).
Figure 3.11. Comparison of the linear temperature trends at Byrd during 1958–2005 from several reconstructions: the reconstructed Byrd record presented here ("This study"); Monaghan et al. (2008a); Steig et al. (2009); and O’Donnell et al. (2011). For Monaghan et al. (2008a), the revised version of their reconstructed Byrd record (Küttel et al., 2012) is used here. The trend values, along with details about each dataset, are given in Table 3.2. The error bars denote the 95% confidence interval.
Figure 3.12. Relationships between Byrd temperature and the atmospheric circulation during the three warming seasons. **Top row:** Linear trends of the seasonal mean 500 hPa geopotential heights ($Z_{500}$) from ERA-Interim during 1979–2009. The thick dashed black lines denotes the 95% significance level of the trends. **Middle row:** Correlations ($r$) between reconstructed Byrd temperatures and $Z_{500}$ calculated for 1979–2009. The correlations are based on detrended time series. Areas with $r > 0.5$ over West Antarctica are denoted with a thick red line. **Bottom row:** Times series of seasonal mean temperature at Byrd (black line) and $Z_{500}$ averaged over boxes 1, 2, and 3 in (d), (e), and (f), respectively (red line). The black arrows in (a) through (f) denote the sector/direction of the warm air advection.
Figure 3.13. Maximum covariance analysis (MCA) between seasonal mean estimates of sea surface temperatures (SST) and 200 hPa geopotential heights ($Z_{200}$) for 1979–2009. The MCA is performed as in Ding et al. (2011, 2012), except that the SST data span 50°S to 20°N (instead of 20°S to 20°N in Ding et al.) and the $Z_{200}$ data are taken from ERA-Interim (instead of a composite ERA-40/ERA-Interim in Ding et al.). The figure shows the spatial patterns of SST (50°S to 20°N) and $Z_{200}$ (equator to 87.5°S) associated with the first (left column) and second (right column) modes of covariability between these two fields. The number in the upper-right corner indicates the percent covariance explained by each mode. As in Ding et al., the amplitudes are scaled by one standard deviation of their respective time series. These time series are shown in Fig. 3.14. Units are °C for SST and meters for $Z_{200}$. The SST data are from the ERSST version 3 data set (Smith et al., 2008).
Figure 3.14. Continuation of the maximum covariance analysis (MCA) described in Fig. 3.13. Shown are the expansion coefficients of the first and second modes of SST-$Z_{200}$ covariability (the corresponding spatial patterns are shown in Fig. 3.13). The number in the upper-right corner indicates the correlation between the SST and $Z_{200}$ time series.
Figure 3.15. Correlations between Byrd temperatures and ERA-Interim 500 hPa geopotential height calculated for 1989–2011 during the three warming seasons: winter (JJA), spring (SON), and summer (DJF). The correlations are based on detrended time series.
Figure 3.16. (a) Time-averaged 700 hPa geopotential height (Z700) field in austral summer (DJF) during 1979–2011. The star symbol denotes the location of Byrd. (b) Hovmueller diagram showing the difference in Z700 relative to longitude 120°W as a function of the longitude (y-axis) and time (x-axis). The Z700 field is meridionally averaged over the area shown with a black box in (a). (c) Temperature anomalies at Byrd in DJF. The dashed gray line in (b) and (c) highlights the marked temperature increase at Byrd in the late 1980s.
Figure 3.17. **Top:** Linear trends in ERA-Interim 500 hPa geopotential heights ($Z_{500}$) in austral summer (DJF) during 1979–1993. The thick dashed black lines denotes the 95% significance level of the trends. **Bottom:** Mean DJF temperature time series at Byrd (black line) and $Z_{500}$ averaged over the Ross Sea (‘Box 1’). As in Fig. 3.16, the dashed gray line highlights the marked temperature increase at Byrd in the late 1980s.
Figure 3.18. Sea surface temperature (SST) and 500 hPa geopotential height ($Z_{500}$) anomalies in austral summer 1997–1998 (top) and 2005-2006 (bottom). These two summers have been the two warmest at Byrd since 1957 (see Fig. 3.9). SST anomalies, in °C, are color-shaded. $Z_{500}$ anomalies, in meters, are shown with black line contours. The star symbol denotes the location of Byrd.
Chapter 4: Reconstruction of Antarctic Near-surface Temperatures since the late 1950s: Trends, Variability, and Influence of the Southern Annular Mode

4.1 Introduction

4.1.1 Antarctic Warming and Global Sea Level

Long-term changes in Antarctic near-surface air temperature constitute a critical parameter controlling, directly or indirectly, the ice sheet’s mass balance and its contribution to sea level rise. An increase in temperature has two opposite effects for Antarctica. The first effect is surface melting, which starts occurring if the temperature reaches 0 °C. This can lead to mass loss directly, through meltwater runoff, or indirectly, by contributing to ice-shelf break-up, the ensuing glacier acceleration, and ultimately increasing ice discharge into the ocean. The first mechanism (runoff), prominent in Greenland, is still relatively limited in Antarctica (Tedesco et al., 2007; Kuipers Munneke et al., 2012) and is expected to remain largely so in the decades to come (Krinner et al., 2007; Ligtenberg et al., 2013). The second mechanism (ice-shelf break-up) is potentially far more significant given the buttressing effect of the large coastal ice shelves for the structurally unstable West Antarctic Ice Sheet. It has already been observed in the Antarctic Peninsula (Rignot et al., 2004;
Scambos et al., 2004; Pritchard and Vaughan, 2007) but seems to have played a minor role in the recent acceleration of glaciers in the Amundsen Sea sector of West Antarctica, linked primarily to warm ocean water and sub-ice-shelf melting (Jacobs et al., 2011; Pritchard et al., 2012; Depoorter et al., 2013; Stanton et al., 2013).

The second effect of higher temperature results from the positive exponential relationship between atmospheric moisture and temperature, and is manifested through greater snowfall over Antarctica (at least on a multi-year time scale) and thus the removal of water from the ocean. This phenomenon accounts for the negative sea-level contribution of Antarctica projected for the 21st century (Gregory and Huybrechts, 2006; Bengtsson et al., 2011; Ligtenberg et al., 2013). The magnitude of the precipitation-temperature sensitivity varies between studies, however, and depends in part on how well temperature changes near the surface reflect those occurring throughout the troposphere. The extent to which the positive “precipitation effect” will offset the dynamical loss of ice is one of the great unknowns of Antarctica’s future climate trajectory.

4.1.2 Sparse Observations and Need for Statistical Reconstructions

Given the large interannual and decadal variability of Antarctic climate, placing recent temperature changes in a longer-term context is important to determine their significance and the potential contribution of anthropogenic forcing (Schneider et al., 2006; Gillett et al., 2008; Goosse et al., 2012; Abram et al., 2013; Steig et al., 2013; Thomas et al., 2013). Data sparsity is the primary challenge that one faces when trying to piece together the temperature history of Antarctica. With regard to direct meteorological observations, the “long term” in Antarctica starts with the 1957–1958 International Geophysical Year (IGY). Poleward of 60°S, 15 research stations have temperature records extending from the IGY (or immediately thereafter) to the present. Despite their small number, these records have been
used to reveal and investigate important aspects of Antarctic climate change, especially the contrasting temperature trend pattern between the Antarctic Peninsula and East Antarctica (Thompson and Solomon, 2002; Vaughan et al., 2003; Turner et al., 2005; Marshall et al., 2013; Richard et al., 2013). The main limitation of these investigations is that 1) most stations are located on the coast, thus providing little insight into the vast Antarctic interior; and 2) they leave aside the entire West Antarctic sector, owing to the lack of long-term continuous temperature records in the region. Global reanalyses, which combine various types of meteorological observations with the fields from an atmospheric model, represent a promising alternative for reconstructing the Antarctic climate from the past decades. Unfortunately, despite significant progress made in recent reanalysis experiments, their trends have proved largely unreliable in high southern latitudes (Marshall, 2002; Bromwich and Fogt, 2004; Bromwich et al., 2007, 2011).

The spatial homogeneity of Antarctic monthly mean temperature anomalies over hundreds of kilometers (Comiso, 2000; Monaghan et al., 2008a) has fortunately made it possible to exploit the sparse observation network to reconstruct Antarctic temperatures from the late 1950s onward. Since 2007, several temperature reconstructions have been published, based on various spatial interpolation techniques. These include natural neighbor interpolation modulated by a correlation length scale parameter (Chapman and Walsh, 2007); kriging-based method with kriging field derived from global reanalysis temperature estimates (Monaghan et al., 2008a); principal component analysis combined with spatio-temporal covariances derived from satellite infrared skin temperatures (Steig et al., 2009; O’Donnell et al., 2011). Some of these reconstructions have been key in drawing attention to the warming of West Antarctica since the IGY (Steig et al., 2009; Schneider et al.,
2012a), but their overall lack of agreement on the magnitude, pattern, and seasonality of this warming undermines our current understanding of the phenomenon.

4.1.3 The “West Antarctic Challenge”

These discrepancies can be attributed, in part, to different approaches or assumptions regarding the use of temperature observations from Byrd Station, in the middle of West Antarctica. This data set fills a vast data void between the western Ross Ice Shelf, South Pole, and the northern Antarctic Peninsula (see Fig. 4.1), but it is unfortunately only partially complete. Repeated efforts have been made to fill in the temporal gaps of the Byrd record using various ancillary data sets (Shuman and Stearns, 2001; Reusch and Alley, 2004; Monaghan et al., 2008a; Küttel et al., 2012; Bromwich et al., 2013). The importance of the Byrd record for Antarctic temperature reconstructions is highlighted in Monaghan et al. (2008a) by the marked sensitivity of their results to the inclusion of Byrd in the spatial interpolation. This comes from the fact that temperature variability in central West Antarctica is poorly captured by the surrounding long-term temperature records (Bromwich et al., 2013), which in turn stems from the distinct climatic nature of West Antarctica compared to the rest of the continent (e.g., Guo et al., 2004; Bromwich et al., 2004; Bindschadler, 2006). The infilling technique originally used by Monaghan et al. (2008a) was revised in 2010 and resulted in a significant change in their temperature trends over West Antarctica (from insignificant cooling to highly significant warming), in better agreement with those found by Steig et al. (2009). The results from the revised Monaghan et al. (2008a) reconstruction (from now on referred to as M10) were first used by Schneider et al. (2012a), with details of the new infilling method later described in Küttel et al. (2012).
4.1.4 Purpose and Outline of the Analysis

The comprehensive re-examination of the Byrd temperature record presented in Chapter 3 of this dissertation contributed to ascertain long-term temperature changes in West Antarctica. It resulted in more temporally consistent observations and a more reliable infilling of the data gaps. The motivations for the new (two-dimensional) temperature reconstruction presented here come from: the importance of the Byrd record mentioned above; the fact that none of the previous reconstructions includes its recent revisions; the current lack of consensus on temperature changes in West Antarctica; and the multiple examples of ongoing rapid climate change in the region. The data and methods used to produce and analyze this reconstruction are described in section 4.2. The skill of the reconstruction is then evaluated (section 4.3), followed by a discussion of the variability and trends in Antarctic temperature during 1957–2012 (section 4.4). The reconstruction is then used to investigate the relationship between the Southern Annular Mode, the main mode of extratropical Southern Hemisphere atmospheric variability, and Antarctic temperatures since 1957. Finally, section 4.6 highlights some implications of the results for future Antarctic mass balance.

4.2 Data and Methods

4.2.1 Temperature Data Sets

The reconstruction presented here (hereafter RECON) aims to produce estimates of monthly mean temperature anomalies\textsuperscript{15} for all Antarctica from 1958 to 2012 based on 15 Antarctic stations with observations available almost continuously from the late 1950s or early 1960s onward. The station locations are shown in Fig. 4.1 and their names are listed in

\textsuperscript{15}Monthly anomalies refer to the monthly departures from the mean annual cycle.
Table 4.1. The monthly mean temperature observations (including those used for independent verification in section 4.3.1) are obtained from the SCAR READER database (Turner et al., 2004)\textsuperscript{16}, except for Byrd, for which I used the reconstructed record\textsuperscript{17} presented in Chapter 3 of this dissertation as well as in Bromwich et al. (2013, 2014).

Monthly gridded 2-meter temperature ($T_{2m}$) data from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim Reanalysis (Dee et al., 2011) spanning 1979–2012 are used to determine the spatio-temporal correlation fields required in the spatial interpolation. The choice of ERA-Interim is motivated by its better overall performance and reliability in high southern latitudes compared to other global reanalyses (Bromwich et al., 2011, 2013; Bracegirdle and Marshall, 2012; Medley et al., 2013; Sanz Rodrigo et al., 2013). ECMWF provides two $T_{2m}$ data sets for ERA-Interim on its data server. The \textit{analysis} $T_{2m}$ is generated by the surface analysis (Simmons et al., 2010; Dee et al., 2011), whereby surface observations are combined with $T_{2m}$ estimates from a short-term model forecast (a.k.a. \textit{forecast} $T_{2m}$). In other words, the forecast $T_{2m}$ corresponds to the $T_{2m}$ field prior to the assimilation of surface observations. Due to its greater observational constraint, the analysis $T_{2m}$ is the product more commonly used in the literature. However, as demonstrated in section 4.3.3, the changing availability of observations is responsible for spurious shifts and trends in the analysis $T_{2m}$ that significantly distort its interannual and decadal variability. This prompted us to use the forecast $T_{2m}$ to compute the correlations, and reserve the analysis $T_{2m}$ for the validation stage. ERA-Interim data, initially available at a resolution of $\sim0.7^\circ$ in latitude-longitude, were bilinearly interpolated onto a 60$\times$60 km Cartesian grid to allow uniform sampling at all latitudes.

\textsuperscript{16}http://www.antarctica.ac.uk/met/READER/

\textsuperscript{17}The reconstructed Byrd temperature record is available at http://polarmet.osu.edu/Byrd_recon/.
Additional temperature data sets are used for comparison. They consist of three Antarctic temperature reconstructions, namely the M10 reconstruction (A. Monaghan, personal communication); the main reconstruction of Steig et al. (2009), based on trended AVHRR data and available at http://faculty.washington.edu/steig/nature09data/; and the regularized least square (RLS) reconstruction of O’Donnell et al. (2011), available at http://www.climateaudit.info/data/odonnell/. Monthly mean $T_{2m}$ data from two other global reanalyses are also used: the National Aeronautics and Space Administration (NASA) Modern Era Retrospective-Analysis for Research and Applications (MERRA) (Rienecker et al., 2011), and the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) (Saha et al., 2010).

4.2.2 Spatial Interpolation Method

The interpolation method largely builds upon the kriging technique employed by M10 and originally developed to reconstruct Antarctic snowfall (Monaghan et al., 2006b). Here, the original approach is modified (primarily regarding the definition of the kriging weights) to more fully exploit the properties of kriging. A brief description of the two main equations is provided below, with additional details given in Appendix B.1.

For a certain month $t$, the temperature anomaly $\Delta \hat{T}$ predicted at a certain location or grid point $x$ is estimated as the linear combination of the temperature anomalies $\Delta T_k$, $k = 1, \ldots, N$, observed at the $N$ stations considered in the interpolation (here, $N=15$):

$$\Delta \hat{T}(x, t) = \sum_{k=1}^{N} \eta_k(x) \lambda_k(x) \Delta T_k(t)$$

(4.1)

$\lambda_k(x)$ represents the weight ($\geq 0$) assigned to the $k$th station at location $x$. $\eta_k(x)$ is either 1 or -1 depending on the sign of the temperature correlation between the $k$th station and location $x$. To account for spatial differences in variance (typically higher in the Antarctic
interior and lower near the coast), the observations are normalized by their respective standard deviations during the relevant reference period (see details below). Thus, Eq. 4.1 produces a normalized value, which is then multiplied by ERA-Interim standard deviation at location $x$ to yield the final temperature anomaly.

The kriging weights ($\lambda$) characterize the spatial representativeness (or footprint) of the observations. As in M10, these footprints are derived from the temperature field of an atmospheric model (here, ERA-Interim) and defined as the squared Pearson’s coefficients of correlation ($r^2$) between the model temperature at the grid point closest to the observations and the model temperature at every other Antarctic grid point. $r^2_k(x)$ denotes the spatial footprint associated with the $k$th observation.

One important aspect of kriging introduced here and absent from M10 is the use of optimized weighting coefficients. In M10, the contributions of all observations in the interpolation are considered, regardless of multicollinearity between some of the records. This contradicts the status of $\Delta T_k$ as independent variables assumed in Eq. 4.1. This problem is fortunately specifically addressed in Ordinary Kriging by requiring the minimization of the estimation error (Cressie, 1993; Olea, 1999). This condition is satisfied by the following matrix equation defining the optimal weights, noted $\lambda = (\lambda_1, \ldots, \lambda_N)^T$:

$$
\begin{bmatrix}
A & 1 \\
1^T & 0
\end{bmatrix}
\begin{bmatrix}
\lambda \\
\alpha
\end{bmatrix} =
\begin{bmatrix}
b_1 \\
1
\end{bmatrix}
$$

(4.2)

$A$ is an $N \times N$ symmetric matrix describing the variance shared within the observation network. $b$ is an $N$-element column vector containing the spatial footprints evaluated at location $x$ where the temperature anomaly must be estimated. $A^*$, $\lambda^*$, and $b^*$ are identical to $A$, $\lambda$, and $b$, except for one additional element in each dimension. The entries of $A$ and
\( b \) are defined as:

\[
A_{ij} = r_i^2(x_j), \quad b_j = r_i^2(x)
\]

with \( i = 1, \ldots, N \) and \( j = 1, \ldots, N \) (\( i \) and \( j \) are interchangeable with \( k \) used previously). \( x_j \) denotes the location (nearest grid point) of the \( j \)th observation. \( \alpha \) is a parameter known as the Lagrange multiplier, required in the error minimization but without further bearing on the method. Equation 4.2 can be solved for \( \lambda^* \) (and therefore \( \lambda \)) by inverting matrix \( A^* \). This inversion yields weighting coefficients that are normalized to one and can be supplied to Eq. 4.1 to predict the temperature anomaly. A more detailed description of the content of Eq. 4.2 is provided in Appendix B.1.

One error inherent to the interpolation method is the use of fixed (time-invariant) kriging weights, which does not account for interannual or decadal variability related to changes in the atmospheric circulation. To address this issue, an ensemble of 10 reconstructions was generated using 10 different sets of weighting coefficients. Each set is associated with a time period consisting of 15 randomly-selected years within the 34 years covered by ERA-Interim. One important assumption of this approach is that the range of spatial relationships during this era also captures the conditions that prevailed prior to 1979. By virtue of kriging, the reconstructed temperatures at grid points coinciding with the stations used in the interpolation remain unchanged across all versions of the reconstruction.

### 4.2.3 Estimation of the Temperature Trends and Contribution of the Southern Annular Mode

The reconstruction provides the basis for an analysis of Antarctic temperature changes and their linkages to the SAM. Trends derived from least-square linear regression are the main indicators used here to characterize the changes in temperature. In the calculation of their standard error (\( \sigma_{reg} \)), the number of degrees of freedom is adjusted for autocorrelation.
in the time series as in Santer et al. (2000) (the same applies for the correlations and regres-
sions below). The standard deviation of the trends obtained from the 10 reconstructions is
used as a measure of the uncertainty ($\sigma_{\text{recon}}$) due to the interpolation. The total error of
the trends is then estimated as $\sigma_{\text{trend}} = \sqrt{\sigma_{\text{reg}}^2 + \sigma_{\text{recon}}^2}$.

The investigation of the SAM-temperature relationships presented in section 4.6 uses the
met/gjma/sam.html, and relies on least-square linear correlations and regressions between
this index and the reconstructed temperatures. Both correlations and regressions are based
on linearly detrended time series. The SAM contribution to Antarctic temperature trends
is estimated as in Thompson et al. (2000) by calculating the temperature trends that are
linearly congruent with the SAM; that is, by first linearly regressing Antarctic temperatures
onto the SAM index, and then by multiplying the resulting regression coefficients ($b$) by the
linear trend in the SAM index ($b'$). Using $\sigma_b$ and $\sigma_{b'}$ to denote their respective standard er-
ors, the error of the SAM-congruent trend is calculated as $\sigma_{\text{cong}} = \sqrt{b^2 \sigma_{b'}^2 + b'^2 \sigma_b^2 + \sigma_b^2 \sigma_{b'}^2}$,
similar to the product of the variances of two independent variables (Goodman, 1960).
Subtracting the SAM-congruent trends from the total trends yields residual trends that
are linearly independent of the SAM; their error is calculated as $\sigma_{\text{resid}} = \sqrt{\sigma_{\text{trend}}^2 + \sigma_{\text{cong}}^2}$.
Further, the contribution of the SAM to the temperature anomaly of a given month, season,
or year can be estimated by multiplying the regression coefficient, $b$, by the corresponding
value of the SAM index. Subtracting this contribution from the original (reconstructed)
temperature anomaly yields an anomaly “with the SAM removed”.

Throughout the chapter, a trend, correlation, or regression coefficient is deemed “sta-
tistically significant” if it exceeds the 95% confidence interval, defined as $\pm 1.96 \sigma$. 


4.3 Assessment of the Temperature Reconstruction

The scarcity of temperature observations in Antarctica inevitably poses a challenge for evaluating RECON. Accordingly, to perform this evaluation, I decided to rely not only on independent observations (section 4.3.1) as in other reconstructions, but also on $T_{2m}$ estimates from ERA-Interim (section 4.3.2). The latter comes with important caveats that are discussed in section 4.3.3. Three statistical parameters are assessed for the reconstructed annual mean temperature anomalies: detrended correlations, detrended root mean square differences (RMSDs), and trend differences (see Fig. 4.2). The reference periods used to compute the anomalies are tailored to the temporal coverage of each observation or reanalysis record. Because the length of these periods varies, the trend differences are normalized by standard errors. The interpretation of the correlations (Fig. 4.2a) and RMSDs (Fig. 4.2b) is relatively straightforward. As for the trend differences (Fig. 4.2c), positive (negative) values indicate that RECON is warming more (less) or cooling less (more) than the observations or the reanalysis.

4.3.1 Reconstructed Temperatures versus In Situ Observations

RECON is first evaluated against direct temperature observations from staffed and automatic stations not used in the spatial interpolation (filled circles in Fig. 4.2; see station information in Table B.1 in Appendix B). To ensure meaningful statistics, the comparison is restricted to records with at least 12 complete years between 1958 and 2012 and exclude those nearly collocated with stations used in the reconstruction (e.g., on Ross Island or King George Island). As expected, the 20 selected records provide only limited spatial and temporal coverage. Most of them are concentrated in the northern Antarctic Peninsula or
the western Ross Ice Shelf regions, and, outside of the Peninsula, only three records extend
back prior to 1980.

All correlations are statistically significant at 99% (Fig. 4.2a). They exceed 0.7, except
at Marambio (0.69), Neumayer (0.64) and Belgrano I (0.59). All RMSDs are <0.8°C,
and only slightly higher at Belgrano I (0.83°C) and Marambio (0.83°C) (Fig. 4.2b). The
normalized trend differences are ≤ 1 at 13 sites distributed across the various regions of
Antarctica, and ≥ 2 at only 2 sites (Ferrel AWS and Belgrano II) (Fig. 4.2c). Care
must be taken when interpreting some of the results as the reliability of the observations
themselves is sometimes open to question (e.g., at Ferrel AWS). Neverthele, as a whole,
these statistics provide a first indication of the high skill of RECON at capturing both the
phase and magnitude of the interannual and decadal variability in temperature.

4.3.2 Reconstructed Temperatures versus ERA-Interim Reanalysis Temperatures

To enhance the spatial coverage of the evaluation, the reconstruction is compared against
ERA-Interim during their period of overlap (1979–2012). There are two obvious caveats to
this comparison. First, since ERA-Interim supplies the kriging field for the interpolation,
the two data sets are not independent from each other. However, because of its atmospheric
forecast model and advanced data assimilation system, the reanalysis uses a more advanced
form of interpolation than the kriging method and, as a result, is expected to better capture
the interannual variability of the temperature. The second caveat (which qualifies the pre-
vious statement) lies in the sensitivity of ERA-Interim to changes in the observing system,
which makes it prone to artifacts and reduces the reliability of its decadal trends. The fact
that RECON, on the other hand, relies on a fixed network of observations makes it more
temporally consistent. Thus, while ERA-Interim may shed some light on the reconstruction’s skill, the reverse (the reconstruction shedding light on the quality of ERA-Interim) is true as well and both aspects are documented here.

The comparison confirms, to a large extent, the skill of the reconstruction, especially away from the 15 input stations. Correlations are significant at the 99% level over all but 2% of Antarctica; they are >0.7 and >0.8 over (respectively) 77% and 52% of the continent (Fig. 4.2a). RMSDs are <1.0 °C and <0.5 °C over (respectively) 90% and 37% of Antarctica (Fig. 4.2b). The largest RMSDs (>1.5 °C) are concentrated in four main areas: Ross Ice Shelf, eastern Ronne-Filchner Ice Shelf, coastal Dronning Maud Land, and Amery Ice Shelf plus the nearby inland sector. The normalized trend differences reveal good agreement between reconstruction and reanalysis in the 90°E–180° quadrant of Antarctica, but more conflicting results over the rest of the continent (Fig. 4.2c). I show in the next section that the large RMSDs and trend differences are, in most cases, attributable to problems in ERA-Interim data rather than to the lower skill of RECON.

4.3.3 Issues with ERA-Interim Reanalysis Temperatures

Because there is some uncertainty both in RECON and in ERA-Interim, it is desirable to understand the origin of their differences and possibly determine which one is closer to the “truth”. For this purpose, temperature time series are extracted for various locations characterized by low correlations, large RMSDs, and/or large trend differences in Fig. 4.2. These locations are marked with numerical labels in Fig. 4.2d and the corresponding time series are shown in Fig. 4.3 (for four representative sites) and Figs. B.2 and B.3 of Appendix B (for all other sites).

We first consider the difference between ERA-Interim analysis and forecast $T_{2m}$, known as the analysis increment (Fig. 4.3, top panels, and Fig. B.2 in Appendix B). This increment
denotes the corrective impact of surface observations on the reanalysis forecasts and usually reflects the quality of the latter. In Antarctica, however, zero or near-zero increments often signify the absence of observations rather than the good performance of the model forecasts. The non-stationary nature of the increment appears clearly at most sites. The largest corrections (>1°C) and most prominent shifts concern locations in the interior of East Antarctica (Figs. 4.3a and 4.3b). The shifts can sometimes be attributed to the known beginning or end of certain observations records, e.g., the installation of Relay Station AWS in 1995 on the East Antarctic plateau (Fig. 4.3b), or the closing of the coastal Soviet station, Leningradskaya, in early 1992 (Fig. 4.3c). In other cases, the reason for ERA-Interim’s behavior cannot be readily established. The case of South Pole is worth noting as Fig. 4.3a indicates that surface observations, although available (since 1957), were not assimilated into ERA-Interim prior to 1985, for reasons that are unclear. Overall (with a few exceptions), the changes in the analysis increment have produced artificial cooling (warming) trends in ERA-Interim in the Antarctic interior (on the coast). This lends further credence to the trends from RECON, which tend to be above (below) those from ERA-Interim in the interior (coastal) regions (see Fig. 4.2c).

Little or no change in the analysis increment does not necessarily entail the absence of spurious variations in the reanalysis data. This is shown in Fig. 4.3 (bottom panels) and in Fig. B.3 (Appendix B), where ERA-Interim analysis and forecast \( T_{2m} \) time series are compared with those from RECON and two other global reanalyses (MERRA and CFSR). In some places (e.g., South Pole and Relay Station; Figs. 4.3e and 4.3f), the shifts described above only affect the analysis \( T_{2m} \). In other places (e.g., Leningradskaya and Belgrano; Figs. 4.3g and 4.3h), analysis and forecast vary in tandem and exhibit temporal variations that 1) strongly differ from the three other data sets (including RECON), and 2) display
clear signs of non-stationarity. In this case, the problem is likely tied to upper-air (e.g., radiosoundings, satellites) rather than surface observations, but further investigation of this issue remains beyond the scope of this dissertation. Again, this comparison suggests greater realism in the reconstructed temperatures than in ERA-Interim data.

4.4 Analysis of Antarctic Temperature Variability and Trends

4.4.1 Annual Mean Temperature Time Series

The annual mean temperature time series from RECON spatially averaged over all Antarctica and its three subregions are shown in Fig. 4.4, along with similar time series from the reconstructions of M10, Steig et al. (2009), and O’Donnell et al. (2011). In addition, Table 4.2 presents a quantitative comparison of the four data sets, namely the detrended correlations with respect to RECON, and the linear trends during their period of overlap (1960–2006). For Antarctica as a whole (Fig. 4.4a), the annual time series of the four reconstructions display a high degree of resemblance, which is reflected in high correlations coefficients (>0.79). RECON correlates best with O’Donnell et al. (2011), and only slightly less with M10 (the order is reversed in the Peninsula). The good agreement between RECON and M10 is somewhat expected given their common methodological basis; the same good agreement with O’Donnell et al. (2011) is more surprising and proves that quite different interpolation methods can produce very similar results over vast portions of Antarctica. The correlations with Steig et al. (2009) are consistently and markedly lower, particularly in the Peninsula. In this region, the excessively low variance of the Steig et al. (2009) time series seen in Fig. 4.4d is a problem previously noted by O’Donnell et al. (2011).

Regarding the 1960–2006 temperature trends, none of the reconstructions produces a statistically significant change in East Antarctica. It is noteworthy that the trend estimates from RECON and O’Donnell et al. (2011) are virtually identical and about half those of
M10 and Steig et al. (2009). In West Antarctica, O’Donnell et al. (2011) stands out as the only data set with insignificant change, similar to what Bromwich et al. (2013) reported when comparing the reconstructions at Byrd Station. The other three reconstructions agree to within 0.04 °C decade$^{-1}$, with a mid-range estimate from RECON. The similarity of the West Antarctic trends in RECON and Steig et al. (2009) despite the relatively low correlation of their time series (0.64) suggests that the two data sets may agree for the “wrong” reason.

The temperature trends from RECON for 1958–2012 are presented in Table 4.3. Antarctica as a whole exhibits a statistically significant warming of 0.12 ± 0.08 °C decade$^{-1}$. This estimate primarily reflects the average temperature change in East Antarctica (~70% of the continent), which is somewhat smaller than the continent-wide value and not statistically significant (0.08 ± 0.10 °C decade$^{-1}$). West Antarctica and the Antarctic Peninsula exhibit substantially larger trends (0.21 ± 0.10 °C decade$^{-1}$ and 0.42 ± 0.21 °C decade$^{-1}$, respectively), as expected from the mounting evidence of rapid warming of these two regions (see next section). RECON further suggests that 2002–2011 was the warmest decade in Antarctica since 1958. This was due, to a large extent, to higher-than-average temperatures in East Antarctica during the 2000s, coming after a markedly colder 1990s decade (1992–2001 was the third coldest decade in East Antarctica since 1958).

### 4.4.2 Annual and Seasonal Temperature Trend Patterns

The spatial distribution of the temperature trends derived from RECON for the 1958–2012 period are shown in Fig. 4.5. The rapid warming of the Antarctic Peninsula seen in the figure is already a well-documented phenomenon, at least in the northern part of the spine (Vaughan et al., 2003; Turner et al., 2005). The fact that this feature, along with its seasonality (e.g., maximum in JJA), is well captured by RECON is worth underscoring.
since that is not the case of all reconstructions (see section 4.4.1). In addition, Fig. 4.5 suggests that the strong warming is also present in the southern portion of the Peninsula, which is consistent (at least on the annual scale) with an ice-core temperature proxy record from the area (Thomas et al., 2009).

RECON shows significant warming trends over most of West Antarctica, as has been suggested by borehole measurements (Barrett et al., 2009; Orsi et al., 2012), ice-core proxy records (Steig et al., 2013; Thomas et al., 2013) and, to various extents, by other temperature reconstructions (Steig et al., 2009; O’Donnell et al., 2011; Schneider et al., 2012a). In this part of the continent, the seasonality and magnitude of the trends largely reproduce those reported by Bromwich et al. (2013) for Byrd Station. The strongest and most widespread warming occurs in austral spring (SON), in agreement with Schneider et al. (2012a). In austral winter (JJA) and summer (DJF), the areas of significant warming look much alike and are limited to the eastern (Bellingshausen Sea) sector of West Antarctica. Importantly, this sector encompasses the catchments of the Pine Island and Thwaites Glaciers (outlined in Fig. 4.1), whose acceleration over the past two decades has been the main factor responsible for the net mass loss from West Antarctica (Rignot and Thomas, 2002; Wingham et al., 2009; Shepherd et al., 2012). Summer warming in this area, by enhancing the exposure of the coastal ice shelves to surface melting, could have important implications for the stability of the West Antarctic Ice Sheet (cf. Introduction and section 4.6). No significant changes are observed in austral fall (MAM), the season during which West Antarctic temperatures have likely been most impacted by the strengthening of the SAM (see section 4.5.1).

In West Antarctica, a common feature across the seasons is the tongue-shaped pattern assumed by the largest trends, stretching from the Amundsen Sea coast to the Transantarctic Mountains and the western Ross Ice Shelf. This pattern corresponds to the atmospheric
signature typically associated with warm marine air intrusions (Nicolas and Bromwich, 2011a), thereby reflecting the main atmospheric mechanism thought to be responsible for the West Antarctic warming in JJA and SON (Ding et al., 2011; Schneider et al., 2012a; Bromwich et al., 2013).

In East Antarctica, positive temperature trends prevail overall but are smaller and less significant than in the other two regions. The largest patches of significant warming occur in SON and are primarily concentrated in the 90–180°E quadrant. Cooling, on the other hand, is more widespread in MAM and most pronounced at the western (30°W–0°E) and eastern (120–170°E) edges of East Antarctica. These regional coolings are likely related to atmospheric circulation changes that have promoted offshore (cold) tropospheric flow over these areas (Turner et al., 2009; Bromwich et al., 2011; Marshall et al., 2013). Other regional features are also worth noting. For example, the transition from negative to positive trends near the Greenwich meridian in the annual mean is consistent with the temperature changes seen in several borehole records from the area (Muto et al., 2011). The plateau-versus-coast contrast in DJF is consistent with the more negative impact of a stronger SAM on temperatures in coastal, katabatic-prone areas of Antarctica (Van den Broeke and van Lipzig, 2004). Finally, the more pronounced warming in the 80–100°E sector is consistent with the temperature signal of the coastal borehole record analyzed by Roberts et al. (2013).

4.5 Insight into the SAM Contribution to Antarctic Temperature Change

The strengthening of the Southern Annular Mode (SAM) in recent decades (e.g., Marshall et al., 2004) is considered one of the main contributors to Antarctic temperature trends, most notably by explaining the contrasting pattern of little change or cooling in East Antarctica and rapid warming in the Antarctic Peninsula in austral summer and fall.
The changes in the SAM are believed to have masked, at least partially, a “background warming” of East Antarctica, i.e., a warming that would have occurred without changes in the atmospheric circulation (Marshall, 2007; Gillett et al., 2008; Screen and Simmonds, 2012). In this section, RECON is used to re-examine the SAM-temperature relationship in Antarctica by expanding, either spatially (particularly over West Antarctica) or temporally, upon previous studies (e.g., Kwok and Comiso, 2002; Thompson and Solomon, 2002; Schneider et al., 2004, 2012a,b; Van den Broeke and van Lipzig, 2004; Marshall et al., 2006, 2013; Marshall, 2007; Fogt et al., 2012a). In the process, we reassess the SAM contribution to Antarctic temperature trends and infer a “best” estimate of the background warming of the continent since the late 1950s.

4.5.1 SAM-Temperature Correlations

The SAM influence on Antarctic temperatures is illustrated with correlation maps in Fig. 4.6. These correlations are calculated for a single period (1958–2012); as such, they do not capture the non-stationarity of the SAM-temperature relationship in certain areas (see, e.g., Marshall et al., 2011, 2013) and should therefore only be regarded as the characteristics that prevailed during these 54 years.

4.5.1.1 East Antarctica and Antarctic Peninsula

A strong SAM, or positive SAM index, is typically associated with lower (higher) temperatures in East Antarctica (the Antarctic Peninsula). The cooling effect in East Antarctica stems both from reduced meridional heat exchange within the troposphere, and from reduced downward turbulent heat flux near the ice sheet’s surface (Van den Broeke and van Lipzig, 2004). The correlations are most negative in DJF within 500–1000 km from the
coast (the katabatic zone), and are generally lower, yet still statistically significant, in SON. The western edge of East Antarctica stands out with insignificant correlations in all seasons but DJF. This feature is consistent with the non-stationarity of the SAM-temperature relationship in the area (Marshall et al., 2011).

In the Antarctic Peninsula, the stronger circumpolar flow associated with a positive SAM index enhances warm air advection across the region, which in turn favors positive temperature anomalies, especially on the eastern side (Marshall et al., 2006; van Lipzig et al., 2008). However, significant positive correlations are restricted to the very northern tip of the Peninsula; and temperatures in the rest of the region are positively correlated with the SAM only during MAM and JJA, and never in a statistically significant manner. This is largely due to the prominent role of the tropics, especially the El Niño–Southern Oscillation (ENSO), in driving atmospheric variability in and around the Peninsula (Kwok and Comiso, 2002; Schneider et al., 2012a; Ding and Steig, 2013; Clem and Fogt, 2013).

4.5.1.2 West Antarctica

In West Antarctica, the correlations are negative everywhere throughout the year but are smaller overall and less significant than in East Antarctica. The difference between the two regions can be traced, again, to a greater imprint of ENSO in West Antarctica (e.g., Bromwich et al., 2000, 2004; Guo et al., 2004; Gregory and Noone, 2008; Schneider and Steig, 2008; Okumura et al., 2012). On the annual scale, the area of significant correlations encompasses the western two thirds of West Antarctica, leaving out the Bellingshausen Sea sector, adjacent to the southern Peninsula. The seasonal correlations show that the annual pattern is primarily reflective of MAM, which has some implications for the interpretation of temperature records from the area that do not resolve the seasonal cycle (e.g., ice-core proxies).
Furthermore, a contrast appears clearly in the magnitude and significance of West Antarctic correlations between the two halves of the year (DJF-MAM versus JJA-SON). Several factors may contribute to the weak linear SAM-temperature relationship in JJA and SON, such as: 1) a stronger ENSO teleconnection in the southeastern South Pacific in these two seasons (Bromwich et al., 2004; Harangozo, 2004; Lachlan-Cope and Connolley, 2006; Schneider et al., 2012a), and thus greater modulation of the atmospheric impact of the SAM in West Antarctica (e.g., Stammerjohn et al., 2008); 2) a more pronounced non-annular component of the SAM over the same oceanic sector also in JJA and SON (Fogt et al., 2012a); and 3) decadal variability in the characteristics of ENSO itself, its teleconnection pattern, and the SAM-ENSO interaction (Bromwich et al., 2004; Fogt and Bromwich, 2006; Fogt et al., 2011; Lee and McPhaden, 2010; Ding et al., 2011; Okumura et al., 2012).

In the other two seasons, it is worth noting the longitudinal (westward) shift of the area of significant correlations between DJF and MAM. This shift may be related to the seasonal change in the position of the climatological center of low pressure off the coast of West Antarctica, known as the “Amundsen Sea Low” (ASL) (Fogt et al., 2012b; Hosking et al., 2013; Turner et al., 2013). The ASL is generally centered over the Amundsen/Ross Seas except in DJF when it migrates toward the Bellingshausen Sea. Interestingly, Hosking et al. (2013) found little sensitivity of this longitudinal shift of the ASL to the background pressure, which is influenced, e.g., by the phase of the SAM.

4.5.2 SAM-congruent and Residual Temperature Trends

Given the prevailing negative SAM-temperature correlations in Fig. 4.6, it is not surprising to see from the SAM-congruent trends that the strengthening of the SAM has acted to cool down most of Antarctica (Table 4.3 and Fig. 4.7, top panels). The actual SAM impact in each season also depends on the trend in the SAM index; during 1958–2012, it is
maximum in DJF and MAM, positive but not statistically significant in JJA, and virtually zero in SON (see trend values in Fig. 4.7, top panels). Once the SAM contribution is removed from the total trends, the residual trends (Table 4.3 and Fig. 4.7, bottom panels) are for the most part positive and larger than the total trends, with greater statistical significance, particularly for the annual mean. This is expected given the fact that most of Antarctica has warmed since 1958 (albeit not always significantly) despite the SAM-related cooling. In other words, as suggested by several other studies (Marshall, 2007; Gillett et al., 2008; Screen and Simmonds, 2012), the strengthening of the SAM has only mitigated an otherwise stronger Antarctic warming.

The residual trends should not be mistaken for the background warming that I seek to quantify here. They may simply reflect temperature changes caused by atmospheric circulation changes not related (at least not linearly) to the SAM. That is certainly the case in West Antarctica and in the Peninsula where the regional warmings have been thus far primarily linked to tropically forced pressure anomalies (Ding et al., 2011; Schneider et al., 2012a; Ding and Steig, 2013). When spatially averaged over the three Antarctic regions, the residual trends show less seasonal variation in East Antarctica (Table 4.3). For this reason, the annual residual trend for this region (0.17±0.12 °C decade⁻¹) is deemed a closer estimate of the background warming of the continent than the annual residual trend averaged over all Antarctica. This estimate of Antarctic background warming during 1958–2012 turns out to be close to the residual trends of each region in austral summer (~0.20 °C decade⁻¹), and amounts to a 0.86±0.59 °C increase in the annual mean temperature per 50 years.
4.5.3 Sensitivity to the Study Period

The assessment of Antarctic temperature trends and the respective contributions of various climate modes is often complicated by large interannual and decadal variability, and thus by the sensitivity of the results to the choice of the time period (see, e.g., Chapman and Walsh, 2007; Monaghan et al., 2008b; O’Donnell et al., 2011). Figure 4.8 is an attempt to quantify this issue and define lower and upper bounds for the background warming of Antarctica. The figure shows the total trends in the annual mean temperature of all Antarctica and East Antarctica, respectively, their SAM-congruent portion, and the residual trends calculated for different time periods. The beginning and end of each period are taken within two 15-year windows at both ends of RECON (1958–1972 for the start and 1998–2012 for the end).

The total trends (Figs. 4.8a and 4.8d) show maximum cooling of all Antarctica and East Antarctica concentrated during some of the shortest periods of the ensemble (starting in 1970–1972 and ending in 1999–2001). This time span corresponds to the one for which the “cooling continent-warming Peninsula” pattern was first identified (Doran et al., 2002; Thompson and Solomon, 2002; Vaughan et al., 2003). When the periods considered are extended backward or forward in time, the trends become less negative/more positive. Most of the largest warming rates, both continent-wide and for East Antarctica only, are associated with the longest periods (starting in 1958–1962 and ending in 2006–2012).

The SAM-effect on Antarctic temperatures (Figs. 4.8b and 4.8e) is always negative, but it varies by a factor of three across the periods sampled in the figure. Not surprisingly, the largest SAM-related cooling coincides with the most negative total trends described in the previous paragraph; it occurs primarily during periods that maximize the upward trend in the SAM index. The smallest SAM-induced cooling are for periods starting prior to 1962;
these encompass a brief interval of positive annual SAM indices, before a prolonged span
of negative values during the rest of 1960s and 1970s (see Fig. 4.10b discussed in the next
section).

The residual trends (Figs. 4.8c and 4.8f) are positive regardless of the time period and
the region considered. They display greater statistical significance than the total trends
(Fig. 4.8a and 4.8d). As in section 4.5.2, the background warming is estimated based on
the residual trends for East Antarctica rather than the whole continent. In doing so, we
exclude the shortest periods (start in 1970–1972, end in 1998–2001) to reduce the skewness
of an otherwise near-normal distribution of the trend values. Averaged over all other time
periods (213 out of 225), the residual temperature increase amounts to +0.80±1.11 °C per 50
years, with both the standard error of the trends and the sampling error being factored into
the error interval. This value is remarkably close to the 0.78±0.29 °C temperature increase
per 50 years of the entire Southern Hemisphere during 1958–2012, which was estimated
from the GISS Surface Temperature Analysis (Hansen et al., 2010)\textsuperscript{18}.

4.6 Implications for West Antarctic Mass Balance

4.6.1 Current Exposure of West Antarctic Glaciers to Surface Melting

What implications do the temperature trends and the SAM-temperature relationships
described above entail for the mass balance of Antarctica? Here, we address this ques-
tion by focusing on the region of the Pine Island (PIG) and Thwaites (THW) Glaciers in
West Antarctica, and this for three main reasons: 1) as mentioned in section 4.4.2, these
outlet glaciers are currently the largest contributors to the negative mass balance of West
Antarctica; 2) they are also particularly vulnerable to further climate warming (Vaughan
et al., 2001; Joughin and Alley, 2011); and 3) this sector of West Antarctica exhibits, in

\textsuperscript{18}http://data.giss.nasa.gov/gistemp/.
austral summer, both statistically significant warming and statistically significant SAM-temperature correlations (Figs. 4.5b and 4.6b). For the ice sheet, the most direct potential effect of an increase in temperature is the occurrence of surface melting, that is, assuming warm enough conditions. This phenomenon can also be especially consequential because of its indirect, potentially abrupt, and long-lasting effect on the ice discharge (cf. Introduction). In the case of PIG and THW, one may argue that the temperature trends in summer are relatively modest compared to those seen in spring and winter (Fig. 4.5). However, as far as melting is concerned, one must keep in mind that the magnitude of the temperature change is less important than the proximity of the temperature to 0°C.

This proximity (or lack thereof) to the melting point in Antarctica at the peak of austral summer (December–January, or DJ) is illustrated in Fig. 4.9a, based on ERA-Interim mean DJ $T_{2m}$ data averaged over the 1990–2012 period. The figure highlights areas with temperatures $>-5^\circ$C (yellow/orange shades) since this is the monthly mean threshold above which surface melting is typically observed in Antarctica during the course of a month (Liu et al., 2006). The inset map in Fig. 4.9a indicates temperatures $>-5^\circ$C over the ice shelves and coastal margins of PIG and THW, already suggesting some degree of exposure to surface melting. Satellite passive microwave observations (Fig. 4.9b) confirm that the phenomenon is not unusual in the area. Melt data from 1990–2012 show that parts of PIG Ice Shelf experience, on average, up to 15 days of melting each summer, and up to 9 days for THW Ice Shelf.

4.6.2 Mitigating Impact of the SAM on West Antarctic Temperatures

The summer temperatures and melt observations described above are characteristic of two decades dominated by a positive SAM index (Fig. 4.10b, histogram). The impact of this stronger SAM on mean DJ temperatures over the PIG catchment area is examined
in Fig. 4.10a by comparing the original temperature anomaly time series from RECON (black curve) with those obtained after removing the contribution of the SAM (red curve). The actual quantification of the SAM effect (difference between black and red curves) for a given year is contingent upon the period selected to define the “normal” state of the SAM. Here, we use 1958–1979 (the “pre-ozone hole era”) as the reference period since the SAM strengthening in austral summer has been attributed primarily to post-1980 stratospheric ozone depletion (Thompson et al., 2011, and references therein). Thus, we estimate that the positive SAM index that has prevailed in DJ since 1994 has, on average, kept the temperature over PIG $0.49 \pm 0.80^\circ C$ lower than it would have been without any change in the SAM. This estimate is based on the SAM-temperature regression coefficient calculated for 1959–2012 (Fig. 4.10b, black curve). As this figure shows, the magnitude of this coefficient varies in time; it is notably greater when the regression period starts in 1975 or later. If we use the regression coefficient for 1975–2012, the average SAM contribution to the 1994–2012 temperature anomalies increases (in absolute terms) to $-0.63 \pm 0.88^\circ C$. In either case, the error bars denote a large degree of uncertainty, owing both to the imperfect knowledge of the temperature history (the reconstruction uncertainty) and to the imperfect fit of the regression line.

### 4.6.3 Looking into the Future: Risks and Uncertainties

The cooling effect induced by a strong SAM can also be interpreted as the additional warming that might occur if the SAM were to revert to its average pre-1980 state. This scenario would increase the probability of occurrence of surface melting over the PIG and THW Ice Shelves. The actual trajectory of the temperature in the PIG area will depend, in part, upon the future evolution of both the SAM and the SAM-temperature relationship. The latter is largely unknown, as are the factors that influence it. For instance, is the
increase in the magnitude of the regression coefficient seen in Fig. 4.10b linked to the strengthening of the SAM itself? Or is it related to another mode of climate variability, such as ENSO or the Pacific Decadal Oscillation? (see, e.g., Okumura et al., 2012). With the ongoing reduction of ozone-depleting substances, future SAM changes will be determined by the competing effect of ozone recovery and increasing greenhouse gas concentrations in the atmosphere. The two phenomena are projected to largely offset each other during the twenty-first century (Perlwitz et al., 2008; Son et al., 2009; Simpkins and Karpechko, 2012; Gillett and Fyfe, 2013; Zheng et al., 2013), thus maintaining a strong SAM environment (with respect to the pre-ozone hole era), and reducing the likelihood of a SAM-induced enhanced warming of Antarctica.

Important sources of uncertainty remain, however, regarding this projection. First, the ability of global climate models to simulate the SAM and, more generally, various aspects of Antarctic climate is one of them (Fogt et al., 2009; Bracegirdle, 2013; Simpson et al., 2013a,b). Second, natural variability has played an important role in the SAM trends in the recent past, particularly in MAM (Marshall et al., 2004; Fogt et al., 2009). Accordingly, future SAM changes could well deviate, if only temporarily, from the expected climate response to anthropogenic forcings. Third, some individual models do suggest a weakening of the SAM during the twenty-first century, and thus the possibility of it returning to its pre-ozone hole state (Perlwitz et al., 2008; Zheng et al., 2013). Finally, while the chapter focuses on the contribution of the SAM to Antarctic temperature trends, other modes of climate variability, especially emanating from the tropics, may also contribute significantly to future Antarctic temperature changes in austral summer. In short, the prospect of more frequent surface melting adding to warm ocean water to further weaken the PIG and THW
Ice Shelves in the coming decades cannot be ruled out, and with it greater mass loss from West Antarctica.
Table 4.1. List of stations whose temperature records are used in the reconstruction. The numbers in column 1 refer to those in Fig. 4.1.

<table>
<thead>
<tr>
<th>Nb.</th>
<th>Station Name</th>
<th>Lat. (°S)</th>
<th>Lon. (°E)</th>
<th>Start Year</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Faraday/Vernadsky</td>
<td>65.2</td>
<td>-64.3</td>
<td>1950</td>
</tr>
<tr>
<td>2</td>
<td>Bellingshausen</td>
<td>62.2</td>
<td>-59.0</td>
<td>1968</td>
</tr>
<tr>
<td>3</td>
<td>Orcadas</td>
<td>60.7</td>
<td>-44.7</td>
<td>1903</td>
</tr>
<tr>
<td>4</td>
<td>Halley</td>
<td>75.6</td>
<td>-26.6</td>
<td>1957</td>
</tr>
<tr>
<td>5</td>
<td>Novolazarevskaya</td>
<td>70.8</td>
<td>11.8</td>
<td>1961</td>
</tr>
<tr>
<td>6</td>
<td>Syowa</td>
<td>69.0</td>
<td>39.6</td>
<td>1957</td>
</tr>
<tr>
<td>7</td>
<td>Mawson</td>
<td>67.6</td>
<td>62.9</td>
<td>1954</td>
</tr>
<tr>
<td>8</td>
<td>Davis</td>
<td>68.6</td>
<td>78.0</td>
<td>1957</td>
</tr>
<tr>
<td>9</td>
<td>Mirny</td>
<td>66.6</td>
<td>93.0</td>
<td>1956</td>
</tr>
<tr>
<td>10</td>
<td>Vostok</td>
<td>78.5</td>
<td>106.8</td>
<td>1958</td>
</tr>
<tr>
<td>11</td>
<td>Casey</td>
<td>66.3</td>
<td>110.5</td>
<td>1959</td>
</tr>
<tr>
<td>12</td>
<td>Dumont d’Urville</td>
<td>66.7</td>
<td>140.0</td>
<td>1956</td>
</tr>
<tr>
<td>13</td>
<td>Scott Base</td>
<td>77.9</td>
<td>166.8</td>
<td>1957</td>
</tr>
<tr>
<td>14</td>
<td>Byrd</td>
<td>80.0</td>
<td>-119.5</td>
<td>1957</td>
</tr>
<tr>
<td>15</td>
<td>Amundsen-Scott</td>
<td>90.0</td>
<td>0.0</td>
<td>1957</td>
</tr>
</tbody>
</table>
Table 4.2. Comparison statistics per Antarctic region for four temperature reconstructions: our reconstruction (RECON); Monaghan et al. (2008a) as revised in 2010 (M10); Steig et al. (2009) (S09); and O’Donnell et al. (2011) (O11). Correlations and trends are based on annual mean temperature time series spanning 1960–2006. The correlations are computed from linearly detrended time series. Boldface denotes trends statistically significant at the 95% level.

<table>
<thead>
<tr>
<th>Data set</th>
<th>Antarctica</th>
<th>East Ant.</th>
<th>West Ant.</th>
<th>Ant. Penin.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td><strong>Correlations with RECON</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>M10</td>
<td>0.97</td>
<td>0.97</td>
<td>0.88</td>
<td>0.93</td>
</tr>
<tr>
<td>S09</td>
<td>0.79</td>
<td>0.81</td>
<td>0.64</td>
<td>0.41</td>
</tr>
<tr>
<td>O11</td>
<td>0.98</td>
<td>0.98</td>
<td>0.90</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td><strong>1960–2006 trends (°C decade⁻¹)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RECON</td>
<td>0.10±0.10</td>
<td>0.05±0.12</td>
<td>0.19±0.13</td>
<td>0.33±0.23</td>
</tr>
<tr>
<td>M10</td>
<td>0.14±0.13</td>
<td>0.10±0.16</td>
<td>0.20±0.16</td>
<td>0.39±0.22</td>
</tr>
<tr>
<td>S09</td>
<td>0.12±0.11</td>
<td>0.11±0.12</td>
<td>0.16±0.10</td>
<td>0.11±0.07</td>
</tr>
<tr>
<td>O11</td>
<td>0.07±0.09</td>
<td>0.05±0.11</td>
<td>0.08±0.09</td>
<td>0.30±0.13</td>
</tr>
</tbody>
</table>
Table 4.3. Decomposition of the total temperature trends during 1958–2012 (1959–2012 for DJF) between the SAM-congruent trends and the residual trends. All trends are in °C decade⁻¹. Boldface denotes trends statistically significant at the 95% level. The four 3-month seasons are abbreviated as follows: December–January (DJF), March–May (MAM), June–August (JJA), September–November (SON).

<table>
<thead>
<tr>
<th>Season</th>
<th>Antarctica</th>
<th>East Ant.</th>
<th>West Ant.</th>
<th>Ant. Penin.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Total trends</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.12±0.08</td>
<td>0.08±0.10</td>
<td>0.21±0.10</td>
<td>0.42±0.21</td>
</tr>
<tr>
<td>DJF</td>
<td>0.08±0.15</td>
<td>0.06±0.18</td>
<td>0.13±0.14</td>
<td>0.18±0.09</td>
</tr>
<tr>
<td>MAM</td>
<td>0.02±0.17</td>
<td>-0.02±0.19</td>
<td>0.08±0.18</td>
<td>0.42±0.26</td>
</tr>
<tr>
<td>JJA</td>
<td>0.18±0.21</td>
<td>0.12±0.25</td>
<td>0.24±0.26</td>
<td>0.70±0.42</td>
</tr>
<tr>
<td>SON</td>
<td>0.22±0.12</td>
<td>0.16±0.13</td>
<td>0.40±0.18</td>
<td>0.36±0.25</td>
</tr>
<tr>
<td><strong>SAM-congruent trends</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>-0.08±0.06</td>
<td>-0.10±0.07</td>
<td>-0.07±0.06</td>
<td>0.03±0.11</td>
</tr>
<tr>
<td>DJF</td>
<td>-0.11±0.10</td>
<td>-0.13±0.11</td>
<td>-0.07±0.07</td>
<td>-0.03±0.05</td>
</tr>
<tr>
<td>MAM</td>
<td>-0.16±0.13</td>
<td>-0.18±0.15</td>
<td>-0.14±0.13</td>
<td>0.06±0.12</td>
</tr>
<tr>
<td>JJA</td>
<td>-0.09±0.14</td>
<td>-0.11±0.17</td>
<td>-0.05±0.10</td>
<td>0.06±0.13</td>
</tr>
<tr>
<td>SON</td>
<td>0.00±0.08</td>
<td>0.00±0.08</td>
<td>0.00±0.07</td>
<td>0.00±0.04</td>
</tr>
<tr>
<td><strong>Residual trends</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Annual</td>
<td>0.21±0.10</td>
<td>0.17±0.12</td>
<td>0.28±0.11</td>
<td>0.38±0.23</td>
</tr>
<tr>
<td>DJF</td>
<td>0.19±0.18</td>
<td>0.19±0.21</td>
<td>0.20±0.15</td>
<td>0.21±0.10</td>
</tr>
<tr>
<td>MAM</td>
<td>0.18±0.22</td>
<td>0.16±0.25</td>
<td>0.22±0.22</td>
<td>0.36±0.29</td>
</tr>
<tr>
<td>JJA</td>
<td>0.26±0.25</td>
<td>0.23±0.30</td>
<td>0.29±0.28</td>
<td>0.64±0.44</td>
</tr>
<tr>
<td>SON</td>
<td>0.22±0.15</td>
<td>0.15±0.15</td>
<td>0.40±0.19</td>
<td>0.36±0.25</td>
</tr>
</tbody>
</table>
Figure 4.1. Map of Antarctica showing the locations of the stations used in the spatial interpolation (filled red circles). The station names and coordinates are given in Table 4.1. The thick green lines mark the boundaries of the three main Antarctic regions (in boldface). The thin green lines in West Antarctica outline the catchment areas of the Pine Island (PIG) and Thwaites (THW) Glaciers [catchment boundaries from the SCAR Antarctic Digital Database (http://www.add.scar.org)].
**Figure 4.2.** Statistical comparison of the annual mean temperature anomalies from RE-CON versus those from ERA-Interim (background map) and independent observations (filled circles). The statistics shown are (a) the detrended correlations; (b) the detrended root mean square differences (RMSD); and (c) the normalized trend differences (either RE-CON minus ERA-Interim, or RECON minus observations). The time period for which the three statistics are computed is 1979–2012 for ERA-Interim but varies for the observations (see details in Table S1). Panel (d) identifies 15 locations characterized by low correlations, high RMSDs, and/or large trend differences, and whose temperature time series are shown in Fig. 4.3 (for circled locations) and Figs. B.2 and B.3 of Appendix B (for all other sites).
Figure 4.3. Temperature time series at four locations (one per column) where reconstruction and ERA-Interim disagree with each other in Fig. 4.2. The number in parentheses next to the site name refers to the numerical labels in Fig. 4.2d. Top panels: Monthly mean differences between ERA-Interim analysis and forecast $T_{2m}$, reflecting the corrective impact of surface observations on the reanalysis forecasts. Blue (red) filling denotes a cooling (warming) effect of the observations. Bottom panels: Monthly mean temperature anomalies from five data sets: our reconstruction (RECON; black), ERA-Interim forecast and analysis $T_{2m}$ (red and orange, respectively), MERRA $T_{2m}$ (purple), and CFSR $T_{2m}$ (blue). The anomalies are calculated with respect to 1979–1983 time series and are smoothed with a 36-month moving average.
Figure 4.4. Annual temperature anomaly times series spatially averaged over (a) all Antarctica, (b) East Antarctica, (c) West Antarctica, and (d) the Antarctic Peninsula (the region boundaries are outlined in Fig. 4.1). The time series are from: our reconstruction (RECON; thick black line); Monaghan et al. (2008a) revised in 2010 (M10; dashed blue); Steig et al. (2009) (dashed green); and O’Donnell et al. (2011) (dashed red). For RECON, the black curve represents the average of the 10 realizations (see section 4.2.3 for details) whereas the gray-shaded envelope denotes their spread (two standard deviations). All anomalies are calculated with respect to 1960–1969.
Figure 4.5. Temperature trends during 1958–2012 (1959–2012 for DJF). The thick solid (thin dashed) black line encompasses areas with trends statistically significant at the 95% (90%) confidence level. The squares denote the locations of the 15 stations used in the spatial interpolation.
Figure 4.6. Detrended temporal correlations between Antarctic temperatures and the Marshall (2003) SAM index during 1958–2012 (1959–2012 for DJF). The thick black line encompasses areas with correlations statistically significant at 95%.
Figure 4.7. SAM-congruent (top panels) and residual (bottom panels) trends in temperature during 1958–2012 (1959–2012 for DJF). The residual trends are obtained by subtracting the SAM-congruent trends from the total trends shown in Fig. 4.5. The trends (in degree decade$^{-1}$) in the annual or seasonal SAM index from Marshall (2003) are shown in the bottom left corner of the top panels. The thick black line encompasses areas with trends statistically significant at the 95% level.
Figure 4.8. Partitioning of the total trends in annual mean temperature (a,d) between their SAM-congruent (b,e) and residual (c,f) portions. The trends are averaged over all Antarctica (top panels) and East Antarctica (bottom panels). The vertical and horizontal axes represent the start and end years (respectively) of the periods for which the trends are calculated. Stippling indicates trends statistically significant at the 95% level. Note that the trends are expressed as temperature change per 50 years.
Figure 4.9. Mean temperature (a) and average number of days with surface melting (b) in December–January during 1990–2012. The inset maps zoom in on the Pine Island (PIG) and Thwaites (THW) Glaciers area. The temperature estimates are based on ERA-Interim $T_{2m}$ data regridded onto a 10×10 km grid and adjusted for differences between ERA-Interim model topography and the Antarctic digital elevation model of Liu et al. (2001). The melt data are based on satellite passive microwave observations (Picard and Fily, 2006), courtesy of G. Picard (http://lgge.osug.fr/%7epicard/melting/).
Figure 4.10. (a) Mean December–January (DJ) temperature anomalies averaged over the Pine Island Glacier (PIG) catchment area. The plot shows the original temperature anomalies from RECON (black curve) and the temperature anomalies obtained after removing the impact of the SAM (red curve). The estimation of the SAM effect is based on the SAM-temperature regression coefficient for 1959–2012 (see details in section 4.2.3). The anomalies are calculated with respect to 1959–1979 (the “pre-ozone hole era”) and are smoothed with a 5-year moving average. The red and gray shadings represent the envelopes of uncertainty around the two curves. (b) Histogram: DJ SAM index from Marshall (2003) (the year refers to January). Black curve: Regression coefficients between the DJ SAM index and the mean DJ temperature anomalies over the PIG area in units of °C per standard deviation of the SAM index. The regression period starts from the corresponding year on the x-axis and extends through 2012. For example, the value for 1975 represents the regression coefficient for 1975–2012. The two dashed curves denote the 95% confidence interval of the regression coefficients.
Chapter 5: General conclusion

The research presented in this dissertation has helped advance Antarctic climate science on five different fronts: (i) assessment of the quality and reliability of climate data sets; (ii) addition of one long-term record to the limited Antarctic meteorological archive; (iii) insight into climate change over vast, unsampled stretches of Antarctica; (iv) evidence of rapid atmospheric changes in West Antarctica; and (v) understanding of (some of) the causal mechanisms behind observed atmospheric changes. The following paragraphs provide a summary of the key findings, followed by a list of research questions that would deserve further consideration.

5.1 Key Findings

Chapter 2 pursues two main goals: assess the reliability of several global reanalyses (including four released in recent years) in high southern latitudes; and determine the presence of trends in Antarctic surface mass balance and Southern Ocean precipitation. Spurious changes in precipitation are found to various degrees in all data sets but their characteristics (e.g., magnitude, region affected) and origins (e.g., observations involved) vary. The reanalyses with the most problematic precipitation fields are NCEP-2, JRA-25, and MERRA. The last, in particular, exhibits a large precipitation jump over the Southern Ocean in the late 1990s, concurrent with the start of AMSU satellite radiance
assimilation. Overall, ERA-Interim and CFSR provide the most realistic depiction of the precipitation variability and trends in high southern latitudes. According to these two reanalyses, neither Antarctic surface mass balance nor Southern Ocean precipitation have changed significantly over the past three decades. The use of reanalysis data is recurring throughout the dissertation, therefore, to some extent, this chapter lays the groundwork for what follows.

Chapter 3 consists of a comprehensive investigation of the temperature record from Byrd Station (1957–present), a critical yet incomplete meteorological archive from West Antarctica. The entire data set is first subject to thorough scrutiny; several types of errors are identified; a substantial portion of the original observations are corrected as a result; and a novel approach (primarily based on ERA-Interim temperatures) is applied to fill in the missing data. The complete Byrd record reveals a $2.2 \pm 1.3^\circ$C increase in the annual mean temperature between 1958 and 2010, establishing central West Antarctica as one of the fastest-warming regions on Earth. It confirms other reports of West Antarctic warming, in the annual average and in austral spring and winter, but suggests that the temperature trends were previously underestimated. In addition, the analysis shows statistically significant warming in December–January, the peak of austral summer and peak of the melting season in Antarctica. This last result is important since a continued rise in summer temperatures could lead to more frequent and extensive episodes of surface melting in the low-lying parts of West Antarctica. The review of the atmospheric and oceanic drivers of the temperature trends highlights their strong seasonal dependence and the complex interplay between low-latitude sea surface temperature forcing and high-latitude atmospheric variability.

Chapter 4 presents a 55-year (1958–2012) reconstruction of near-surface temperatures over all Antarctica. It is based on the spatial interpolation of 15 meteorological records,
including the Byrd record reconstructed in Chapter 3. The method builds upon a kriging technique previously employed for a similar purpose, but also introduces various improvements to the original approach, including the use of ERA-Interim temperature field as the kriging field. Apart from the strong warming of central West Antarctica, the reconstruction reveals warming in the climatically sensitive Amundsen-Bellingshausen Sea sector, which has been experiencing glacier acceleration in recent years. Although East Antarctica does not exhibit any significant long-term change, its temperature trend shows a clear reversal in the early 2000s after marked cooling in the 1990s. The reconstruction is then used to re-examine the relationships between the Southern Annular Mode (dominant mode of Antarctic atmospheric variability) and Antarctic temperatures. The analysis shows that the strengthening of the SAM in austral summer and fall seen in recent decades has mitigated an otherwise stronger background warming of Antarctica. Further, a future weakening of the SAM in austral summer could expose glaciers along the Amundsen Sea coast to enhanced surface melting, which in turn could lead to greater mass loss from West Antarctica.

5.2 Research Perspectives

As has been shown throughout this dissertation, global reanalyses play an essential role in our ability to detect and interpret climate change in Antarctica. It is noteworthy that century-long reanalyses have been recently produced (Compo et al., 2011) or are in preparation (e.g., at ECMWF). With regard to high southern latitudes, however, such efforts seem premature as it is clear that major improvements are still needed during the post-1979 era alone, and that having reliable atmospheric fields that extend back to the 1957–1958 International Geophysical Year would be an important milestone. Considering
atmospheric reanalysis as an “iterative process” (Uppala et al., 2007), it is hoped that the next generation of global reanalyses will address most of the current issues.

One of the main achievements of this dissertation is to have established a robust and complete temperature record for Byrd Station. The data set is now publicly available online and kept up-to-date to encourage further work on West Antarctic temperature variability and trends. Here are related issues that would deserve further investigation:

- The marked increase in the annual mean temperature at Byrd during the 1980s turns out to be close to the start of the reanalysis and sea surface temperature data sets that are used to investigate the causality of the warming. Accordingly, the investigation lacks a proper baseline (reference period) that would capture the dominant atmospheric and oceanic regimes before the warming occurs (i.e., pre-1979). Being able to establish such baseline is obviously largely contingent upon the availability of reliable reanalysis data.

- The analysis of the SAM-temperature relationships presented in Chapter 4 is an attempt to shed light on the pre-1979 period. However, the use of an atmospheric index (SAM Index) allows only a very partial (mostly zonally averaged) description of the changes in the atmospheric circulation. Knowledge of the regional atmospheric patterns, particularly in the West Antarctic sector of the Southern Ocean, would be necessary to more fully understand the role of the atmosphere in the temperature changes.

- In very recent years, the temperature observations from Byrd have revealed interesting features. For example, since 2009, both the magnitude and variance of the mean temperature in austral spring have been markedly reduced compared to what they
were during the previous decade. Could this be the sign of a transition into a new temperature/atmospheric regime?

- The investigation into the mechanisms responsible for the temperature trends at Byrd shows that both the large scale (e.g., influence from the tropics) and the regional/local scale should be taken into account. For example, comparison between the Byrd temperature record and borehole temperatures from WAIS Divide, only 160 km to the east, suggests that the strong warming of the 1980s seen at Byrd may have been delayed by 5–10 years at WAIS. Was this indeed the case? And why?

- No statistically significant trends has been found in the surface mass balance of Antarctica for the period since 1979, but that is also true for the near-surface temperature over most of the continent (the changes are significant only when calculated from 1957–1958 onward). The lack of significance is primarily due to large interannual variability.

As a final thought, it is worth pointing out that current meteorological observations available from central West Antarctica are far less comprehensive than they were 50 years ago in the wake of the International Geophysical Year, and this is despite growing awareness in the scientific community of the vulnerability of the ice sheet and mounting evidence of rapid glacier changes in the region. Continued research into West Antarctic meteorology and climate change is certainly needed to better comprehend how rapidly and how soon this ice sheet could collapse.


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Steig, E. J., et al., 2013: Recent climate and ice-sheet changes in West Antarctica compared with the past 2,000 years. *Nature Geosci.*, 6, 372–375.


Appendix A: Supplementary Material for Chapter 3

A.1 Byrd Temperature Record in READER

The READER database (Turner et al., 2004) is an important resource for high-quality, long-term (>10 years) surface and upper-air meteorological observations from Antarctic stations. This database is the primary source of monthly temperature observations used in the reconstruction of the Byrd temperature record presented in Chapter 3. In READER, the temperature values for the early part of the record (1957–1975) were taken from the compilation of monthly temperature and pressure observations produced by Jones and Reid (2001). This compilation relied on monthly reports published in the *Monthly Climatic Data for the World* (MCDW) or *World Weather Records* (WWR). Jones and Reid (2001) themselves built upon the work of Jacka et al. (1984) and Jones and Limbert (1987).

Of concern for the temporal consistency of the Byrd temperature record was the uncertainty about the method used to calculate the monthly means. My analysis suggested that the monthly temperatures reported for Byrd Station in MCDW and WWR were likely based on daily minimum and maximum temperatures, a method responsible for a cold bias in the results from October to March. As explained below, the monthly reports suffer from additional issues, such as errors in temperature units or partial temporal coverage.
A.2 Recalculation of the Monthly Mean Temperatures

The monthly mean temperatures from 1957–1975 were recalculated using the 6-hourly observations from the Integrated Surface Hourly Data set (DS 3505) found on the website of the National Climatic Data Center (NCDC). Note that observations are available every 3 hours for most months until 1965, but only during austral summer thereafter. For consistency, only 6-hourly data were used in the calculation, as is done by READER (see Turner et al. (2004) for details).

In addition to the 6-hourly observations, NCDC also provides daily minimum and maximum temperatures ($T_{\text{min}}$, $T_{\text{max}}$) from Byrd Station as part of the Global Historical Climatology Network-Daily Data Set (GHCN-Daily). The temporal availability of the $T_{\text{min}}$ and $T_{\text{max}}$ data is generally greater than that of the 6-hourly data, that is, for certain days, only $T_{\text{min}}$ and $T_{\text{max}}$ are available. In such instances, a daily average temperature computed as $(T_{\text{min}} + T_{\text{max}})/2$ was used to complement to the 6-hourly data. As explained in the main text, the “min/max method” introduces a cold bias but the error is smaller than the uncertainty associated with the infilling method used by Bromwich et al. (2013) for the 1970s (uncertainty conservatively estimated at ±2.5°C, but difficult to actually quantify because of the very small number of observations from Byrd during the entire 1970s decade).

Despite a revision of the Byrd temperature data set on READER’s website on 7 March 2013, some (small) differences now exist (as of January 2014) between my version and READER’s. The differences pertain to the months discussed in the next section.

19 The temporal frequency of the observations varies within the ISH/DS 3505 data set. For months with 3-hourly observationsthe data were resampled to 6-hourly time series.

20 The reconstructed Byrd record is available online at http://polarmet.osu.edu/Byrd_recon/.
A.3 Details about Specific Months

Details about the temperature recalculation for a few specific months are given below. In particular, I explain why some monthly values have been changed, removed, or added compared to the Byrd temperature data set previously available on READER’s website (prior to 7 March 2013).

- **January 1957 (changed).** Old value: -15.6 °C. New value: -15.1 °C. Meteorological observations at Byrd Station began on 10 January 1957. The monthly mean temperature previously (and still currently) reported by READER for that month is therefore based on observations from only 22 days. The 9-day data gap was filled in with the multi-year daily average temperatures for the first 9 days of January during 1958-1970. This data infilling resulted in a +0.5 °C adjustment, reflecting the fact that the beginning of January is almost always warmer than the latter part of the month.

- **February 1970 (changed).** Old value: -19.0 °C. New value: -22.4 °C. The value reported in READER was based on observations ($T_{min}, T_{max}$) from only 16 days (1–16 February 1970), as I concluded from an analysis of the GHCN-Daily data set. It was replaced with the monthly value found in MCDW.

- **October 1970 (added).** Old value: none. New value: -26.3 °C. The monthly mean temperature reported in MCDW (-15.3 °C) was substantially higher than the multi-year average for October (-30.6 °C for 1957–1969). As a result, it was not reported by Jacka et al. (1984), Jones and Limbert (1987), nor Jones and Reid (2001).
I believe that the temperature in MCDW was mistakenly reported in degrees Fahrenheit instead of degrees Celsius. The monthly value recalculated from the 6-hourly observations (-26.3°C) is more consistent with the climatology.

- **November 1970** (added). Old value: none. New value: -22.3°C. Same problem as for October 1970 (value reported in MCDW: -7.0, likely in °F instead of °C). The monthly value was recalculated from the 6-hourly observations.

- **February 1971** (removed). Old value: -20.7°C. New value: none. The value reported by READER was based on observations from only 17 days (1–17 February 1970), as I concluded from an analysis of the GHCN-Daily data set. Since no monthly temperature was found in MCDW, no monthly value is provided for this particular month.

- **January 1973** (added). Old value: none. New value: -10.8°C. The monthly mean temperature was computed using a complete set of 6-hourly (94%) and $T_{min}/T_{max}$ (6%) observations.

- **November 1974** (removed). Old value: -15.9°C. New value: none. The GHCN-Daily data set indicates that only 18 days of observations (13-30 November 1974) are available for this month. Since no monthly temperature was found in MCDW, no monthly value is provided for this particular month.

### A.4 Sources of temperature observations

- 6-hourly temperatures:
  
  Data set: Integrated Surface Hourly (ISH) (DS 3505)
  
  Website: [http://www.ncdc.noaa.gov/doclib/](http://www.ncdc.noaa.gov/doclib/)
• Daily minimum/maximum temperatures:

  Data set: Global Historical Climatology Network (GHCN)–Daily

  Website: http://www.ncdc.noaa.gov/oa/climate/ghcn-daily/

• Monthly mean temperature reports:

  Data set: Monthly Climatic Data for the World (MCDW)

  Website: http://www.ncdc.noaa.gov/IPS/mcdw/mcdw.html
Table A.1. Monthly mean temperatures from Byrd Station (1957-1975) recalculated based on sub-daily observations.

<table>
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<th>Jan</th>
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<th>Mar</th>
<th>Apr</th>
<th>May</th>
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Legend:

- **Black**: All 6-hourly observations are available.
- **Black underlined**: 6-hourly observations are partially available. For days without 6-hourly data, the daily mean temperature is computed as \((T_{\text{min}} + T_{\text{max}}) / 2\), using \(T_{\text{min}}\) and \(T_{\text{max}}\) data from the GHCN-Daily data set. The combination of 6-hourly and \(T_{\text{min}}/T_{\text{max}}\) data provides observations for \(\geq 90\%\) of the days.
- **Blue**: The monthly mean temperature is taken from MCDW except for August 1970, for which the temperature in MCDW is missing. Here, the monthly value for this month is as reported in Jones and Reid (2001).
- **Red**: Observations (either 6-hourly or \(T_{\text{min}}/T_{\text{max}}\)) are available for fewer than 90% of the days. The monthly mean temperature is *not used* in our reconstruction. The only exception is January 1957.
Appendix B: Supplementary Material for Chapter 4

B.1 Equations for Optimal Kriging Weights

This Appendix details the content of Eq. 4.2 (page 96) used to determine the weights of the observations in the spatial interpolation. Using the same notation as in Section 4.2.2, we define the following variables and parameters of the equation (with \( i = 1, \ldots, N \) and \( j = 1, \ldots, N \)):

- \( N \), the number of observations;
- \( x_i \), the location of the \( i \)th observation;
- \( x \), the location where the temperature anomalies must be estimated;
- \( r^2_{ij}(x_j) \), the squared correlation between the \( i \)th and the \( j \)th observation records;
- \( r^2_i(x) \), the squared correlation between the \( i \)th observation record and the temperature at location \( x \);
- \( \alpha \), a parameter known as the Lagrange multiplier.
- \( \lambda_i \), the weight assigned to the \( i \)th observation (i.e., the unknown);

The theory of Ordinary Kriging (e.g., Cressie, 1993; Olea, 1999) demonstrates that the weighting coefficients \( \lambda_i \) are optimal if they minimize the estimation error (a.k.a. kriging
This condition requires the coefficients to be solutions of the following system of \( N+1 \) linear equations:

\[
\begin{align*}
\sum_{i=1}^{N} \lambda_i \ r_i^2(x_i) + \alpha &= r_1^2(x) \\
\sum_{i=1}^{N} \lambda_i \ r_i^2(x_i) + \alpha &= r_2^2(x) \\
&\quad \ldots \quad \ldots \\
\sum_{i=1}^{N} \lambda_i \ r_i^2(x_i) + \alpha &= r_N^2(x) \\
\sum_{i=1}^{N} \lambda_i &= 1
\end{align*}
\]

In matrix notation, the system becomes:

\[
\begin{bmatrix}
  r_1^2(x_1) & \ldots & r_N^2(x_1) & 1 \\
  \vdots & \ddots & \vdots & 1 \\
  r_1^2(x_N) & \ldots & r_N^2(x_N) & 1 \\
  1 & 1 & 1 & 0
\end{bmatrix}
\begin{bmatrix}
  \lambda_1 \\
  \vdots \\
  \lambda_N \\
  \alpha
\end{bmatrix}
=
\begin{bmatrix}
  r_1^2(x) \\
  \vdots \\
  r_N^2(x) \\
  1
\end{bmatrix}
\]
B.2 Supplementary Tables and Figures

Table B.1. List of stations used for independent verification in Fig. 4.2 (page 121). The first column refers to the numbers shown in Fig. B.1. The table below is based on station information and temperature data obtained from the Antarctic READER website.

<table>
<thead>
<tr>
<th>Station Number</th>
<th>Name</th>
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<th>Longitude</th>
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<td>2</td>
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<tr>
<td>3</td>
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Figure B.1. Map showing the locations of the 20 stations (staffed or automatic) used for independent verification in Fig. 4.2 (page 121). The name, coordinates, and period of operation of each station are listed in Table B.1.
Figure B.2. Monthly mean differences between ERA-Interim analysis and forecast 2-meter temperature at the 15 locations identified in Fig. 4.2d (page 121). These differences reflect the corrective impact of surface temperature observations on the reanalysis forecasts. Blue (red) filling denotes a cooling (warming) effect of the observations on the forecasts.
Figure B.3. Monthly mean near-surface temperature anomalies at the 15 locations identified in Fig. 4.2d (page 121) and based on the following data sets: our reconstruction (RECON; black), ERA-Interim forecast and analysis 2-meter temperature ($T_{2m}$) (red and orange, respectively), MERRA $T_{2m}$ (purple), and CFSR $T_{2m}$ (blue). The time series are smoothed with a 36-month moving average and the anomalies are calculated with respect to 1980–1983. Note that the temperature range along the vertical axis is identical in all plots except for “Amery Ice Shelf”.
Appendix C: List of Acronyms

<table>
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<tr>
<th>Acronym</th>
<th>Definition</th>
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<tr>
<td>ATOVS</td>
<td>Advanced TIROS Operational Vertical Sounder</td>
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<tr>
<td>AWS</td>
<td>Automatic Weather Station</td>
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<tr>
<td>CFSR</td>
<td>Climate Forecast System Reanalysis</td>
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<tr>
<td>CL</td>
<td>Confidence level</td>
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<td>CMAP</td>
<td>Climate Prediction Center Merged Analysis of Precipitation</td>
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<td>December-January-February (austral summer)</td>
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<td>ECMWF</td>
<td>European Centre for Medium-Range Weather Forecasts</td>
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<tr>
<td>ENSO</td>
<td>El Niño Southern Oscillation</td>
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<tr>
<td>GHCN</td>
<td>Global Historical Climatology Network</td>
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<td>GPCP</td>
<td>Global Precipitation Climatology Project</td>
</tr>
<tr>
<td>IGY</td>
<td>International Geophysical Year (1957-1958)</td>
</tr>
<tr>
<td>JJA</td>
<td>June-July-August (austral winter)</td>
</tr>
<tr>
<td>JRA-25</td>
<td>Japanese 25-year Reanalysis</td>
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<td>JRA-55</td>
<td>Japanese 55-year Reanalysis</td>
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<tr>
<td>M10</td>
<td>Refers to the reconstruction of Monaghan et al. (2008a) revised in 2010</td>
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<tr>
<td>MAM</td>
<td>March-April-May (austral fall)</td>
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<tr>
<td>MCDW</td>
<td>Monthly Climatic Data for the World</td>
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<tr>
<td>MERRA</td>
<td>Modern Era Retrospective-Analysis for Research and Application</td>
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<td>NCDC</td>
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<td>NCEP-2</td>
<td>NCEP-DOE AMIP-II Reanalysis</td>
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<tr>
<td>P-S</td>
<td>Precipitation-minus-surface sublimation</td>
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<tr>
<td>PIG</td>
<td>Pine Island Glacier (West Antarctica)</td>
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<tr>
<td>READER</td>
<td>REference Antarctic Data for Environmental Research</td>
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<tr>
<td>RECON</td>
<td>Refers to the new temperature presented in Chapter 4</td>
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<tr>
<td>RMSE</td>
<td>Root mean square error</td>
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<td>SAM</td>
<td>Southern Annular Mode</td>
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<tr>
<td>SMB</td>
<td>Surface mass balance</td>
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<tr>
<td>SON</td>
<td>September-October-November (austral spring)</td>
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<td>SPCZ</td>
<td>South Pacific Convergence Zone</td>
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<tr>
<td>SST</td>
<td>Sea surface temperature</td>
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<td>T2m</td>
<td>2-meter air temperature</td>
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<td>THW</td>
<td>Thwaites Glacier (West Antarctica)</td>
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<tr>
<td>TOVS</td>
<td>TIROS Operational Vertical Sounder</td>
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<tr>
<td>WAIS</td>
<td>West Antarctic Ice Sheet</td>
</tr>
<tr>
<td>Z_{500}</td>
<td>Geopotential height of the 500-hPa pressure surface</td>
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