Recent Environmental Changes on the Antarctic Peninsula as Recorded in an Ice Core from the Bruce Plateau

Dissertation

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By

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Abstract

Dramatic changes in net accumulation and temperature on the Antarctic Peninsula have been observed over the last century. Ice core-derived proxies provide histories of these variables extending beyond the relatively short instrumental records available for this region. Histories of net annual accumulation and δ¹⁸O from an ice core drilled to bedrock on the Bruce Plateau in the Antarctic Peninsula provide an additional multi-century record of climate variability in this region. Time series for the period 1750-2009 CE are generated, evaluated for changes over time and compared to other relevant environmental data. Comparisons to satellite and station data are conducted over the relatively short duration of these records. Large scale atmospheric oscillations, such as the Southern Annular Mode and El Niño-Southern Oscillation, are investigated as potential drivers of the observed changes from 1900-2009 CE. Ice core records from other locations are analyzed to provide a larger spatial context for the changes observed on the Antarctic Peninsula from 1750 to 2009 CE. Sulfate in ice cores can originate from sea salt, oceanic biological productivity, volcanic activity, and anthropogenic influences. Different methods are used to distinguish the sulfate contributed by each of these sources and thereby to generate source specific sulfate histories.
Dedication

to my parents
Tom and Janet Goodwin
and my brother
Matthew Goodwin
Acknowledgments

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**Fields of Study**

Major Field: Atmospheric Sciences
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Chapter 1: Introduction

The Antarctic Peninsula (AP) region is a complex environment which includes open ocean, sea ice, land ice, and significant topographic relief within a relatively small area. Understanding the climate of such a complex region would be challenging enough if the local climate were stable. Unfortunately, the climate of the AP has been changing rapidly over the last century [Thomas et al., 2008, 2009] further complicating analyses of climatic processes in the region where climate changes may be manifested in a variety of ways. Two major climatic variables, temperature and accumulation rate, can be reconstructed over long time periods from ice core-derived proxy records and provide an opportunity to better clarify the forces driving the changes observed in the region. In addition to temperature and accumulation, there are many complex interactions and feedbacks that result in changes in other components of the local environment. For example, increasing temperatures may lead to a melting of land ice which increases sea level or to a reduction in sea ice extent or ice shelf decay which allows outlet glaciers to accelerate and discharge more ice further contributing to sea level rise. It has been suggested [Krinner et al., 2007, Ligtenberg et al., 2013] that warmer sea surface temperatures may increase the accumulation rate and hence the mass balance over East
Antarctica, partially offsetting the rise of sea level. The balance among such competing influences is important to understand so they can be better incorporated into climate models used to predict the Earth System’s response to future emissions scenarios [Intergovernmental Panel on Climate Change, 2007].

The AP has long been postulated to be one of Earth’s most sensitive regions where the effects of a warming climate would be evident earlier than in many other areas. Mercer [1978] predicted that warming caused by anthropogenic emission of greenhouse gases could lead to the systematic collapse of ice shelves southward along the AP. This prediction proved to be remarkably prescient with the collapse of the Larsen A ice shelf in 1995 and the collapse of the Larsen B ice shelf in 2002 [Scambos et al., 2000, 2004]. The changes observed over the last century in the AP may be indicative of larger scale climate changes likely to impact other regions across the globe in the future. A new ice core from the Bruce Plateau (BP) on the AP provides an excellent opportunity to deepen our understanding of climate changes in the region and potentially identify the processes most likely to affect humans and ecosystems over the next century.

This core, henceforth called the BP core, is 448.12 meters long and was drilled in 2010 through the BP ice field (66.03°S; 64.07°W; 1975.5 masl). Here the high annual mass accumulation (2.56 m of water equivalent (w.e.) per year from 1990 to 2009) ensures a high resolution climate history for this region. The bottom age of the BP core is currently undetermined as the analysis is in progress. This research project focuses on the period
from 1750 through 2009 CE while the accumulation rate reconstruction covers the period from 1400 through 2009 CE.

Reconstructed accumulation rates and stable isotopic composition provide proxy-based histories of precipitation and temperature. Chemical analyses yield high resolution records of sea salt and non-sea salt derived ionic content that help elucidate the relative importance of different sources of ionic species at this site. These ice core-derived histories, when combined with observational data from meteorological stations in the region, data characterizing changes in atmospheric circulation patterns, and reanalysis data, help clarify the changes that have occurred over time, and point to their causal mechanisms.

Over the last century there has been a significant increase in the variety and volume of climatological data available from the AP region and these have drastically expanded the type of analyses possible using ice core-derived proxy data from the BP. There are three major transition points in climatological data availability for the AP region: the development of reconstructions of large scale atmospheric oscillation indices (such as the Southern Annular Mode, Southern Oscillation Index, and Pacific Decadal Oscillation) extending back to approximately 1900, the establishment of a permanent meteorological observing station at Faraday in 1947, and the advent of satellite borne observations in 1979. These time periods are useful guides for organizing the analyses presented in this dissertation.
Chapter 2 presents a review of previously published scientific literature that provides essential context for the current work. Chapter 3 provides an overview of the data used in this analysis as well as the analytical and mathematical methods employed. Chapter 4 discusses the reconstruction of the accumulation rate record since 1400 CE. Chapter 5 covers both the satellite era, 1979-2009 CE, and the instrumental era on the AP, 1947-2009 CE, over which observational and ice core-derived proxy-data are compared. Also included are in situ meteorological measurements collected for several months at the drill site which are compared with AP station data to determine how representative those data are of meteorological conditions at the drill site. In Chapter 6, the relationships between several important atmospheric oscillations and the ice core-derived proxy records are examined for their period of overlap, 1900 to 2009 CE. Chapter 7 discusses the temperature, accumulation, and chemical species histories extracted from the BP ice core from 1750 to 2009 CE, a period encompassing both pre-anthropogenic and anthropogenic conditions. Included are comparisons of the BP records to those from other ice cores from the region. Chapter 8 presents a new method for assessing the anthropogenic contribution to the total sulfate measured in the BP core. This requires separating the total sulfate into its major components: sea salt, biogenic, volcanic, and anthropogenic. Chapter 9 summarizes the major conclusions of this research and discusses future plans for work to expand on the topics covered in this dissertation.
Chapter 2: Literature Review

The following sections provide background information and a summary of previous research relevant to the goals of this dissertation. Topics covered in this section include (1) the history of ice core drilling in Antarctica, (2) climate records available from the Antarctic Peninsula (AP) region, and (3) Southern Hemisphere (SH) large scale atmospheric oscillations. Two maps are provided, Figure 2.1 shows locations of the major ice cores discussed herein while Figure 2.2 shows the locations of the ice cores and established stations located in the AP region that are discussed in this dissertation. Note that the Dallinger Ice Cap in Figure 2.2 is the location of the ice cores from James Ross Island (JRI).
Figure 2.1 Locations of Antarctic ice cores discussed in this work.
2.1 Antarctic Ice Core Drilling History

A detailed history of early ice core drilling, in both Greenland and Antarctica, is provided by Langway [2008]. A condensed history is presented here highlighting the early and important ice cores collected from Antarctica and further emphasizing the more recent cores. The first ice core drilling attempted in Antarctica was by an international team of British, Norwegian, and Swedish scientists between 1949 and 1952 [Langway, 2008]. Their expedition drilled an ice core to approximately 100 meters depth in a coastal ice
shelf in Droning Maud Land. Challenges with the drilling led to poor core quality and a somewhat discontinuous record which created difficulty in interpreting data from this core, but the results of the expedition are summarized in two volumes. The first volume [Swithinbank, 1957] discusses surface deformation and ice flow from stake measurements on the ice shelf. The second volume [Schytt, 1958] contains the results of both snow studies and the ice core drilling with information about density variation with depth, crystallography, and observations of the air bubbles trapped in the ice.

The International Geophysical Year (IGY) in 1957-1958 was the impetus for the first large scale ice core drilling operations. The Antarctic portion of IGY involved 48 locations on the Antarctic continent with participation by 12 nations [Langway, 2008]. Two ice cores were drilled in Antarctica as part of the IGY research program. In 1957-1958 a core was drilled to a depth of 307 m at Byrd Station [Patenaude et al., 1959; Marshall, 1962]. This core was of high quality such that a stratigraphic sequence was determined. Differences in grain size were observed and found to be related to summer and winter accumulation. The depth between the layers of differing density was measured and used to estimate annual accumulation rates [Patenaude et al., 1959]. In addition to stratigraphy, density measurements were conducted throughout the core and borehole temperatures were measured down to approximately 300 m [Patenaude et al., 1959]. Additionally, concentrations of particulate matter were measured and used to identify summer and winter accumulation and to determine annual accumulation rates [Marshall, 1962]. The second Antarctic core collected as part of the IGY was drilled
from the Ross Ice Shelf at a site named Little America V. It extended through the entire 264 m thick ice shelf to the ocean surface [Ragle et al., 1960; Gow, 1961, 1968].

Following the successful drilling of ice cores in Antarctica and Greenland during the IGY, ice core research received greater attention and numerous coring projects were initiated. Initially, Greenland was the focus of deep coring efforts where new deep drilling capabilities were tested and evaluated at Camp Century in 1966 [Langway, 2008]. These new tools were used over two field seasons at Byrd Station in Antarctica to extract a core 2,164 m long to bedrock in 1968 [Ueda and Garfield, 1969]. Measurements from this core include a density profile, identification of stratigraphic markers, crystallography, and electrical conductivity [Gow, 1970]. Borehole temperatures were also measured. The isotopic composition of the deep Byrd Station core was measured by two groups [Epstein et al., 1970; Johnsen et al., 1972] and provided a proxy-temperature history over roughly the last 80,000 years for this location [Johnsen et al., 1972]. Initially there was some controversy over the time scale of the Byrd core but recent entrapped gas analysis has confirmed that the core dates back roughly 80,000 years [Ahn and Brook, 2008].

Subsequently, numerous ice cores have been recovered from Antarctica. The most important cores are the long cores from Vostok and Dome C, both on the East Antarctic ice sheet. These cores provide the longest climate histories yet obtained from any polar ice cores and are discussed in more detail here.
A climate record covering 420,000 years from the 3,623-m long Vostok core was completed in 1998 [Petit et al., 1999]. Drilling was terminated approximately 120 meters above the bottom of the ice to prevent contamination of the subglacial Lake Vostok. Data obtained from this core include isotopic composition, major ion chemistry, dust concentration, and records of greenhouse gas concentrations from entrapped gas bubbles. These data cover four glacial-interglacial cycles and show a strong relationship between greenhouse gas concentrations and isotopically derived temperatures over this time period. The greenhouse gas history shows similar glacial-interglacial changes in carbon dioxide and methane concentrations. Carbon dioxide concentrations vary from 180 parts per million (ppm) during glacial stages to 280-300 ppm during interglacials [Petit et al., 1999]. The Vostok core results revealed the important relationship between atmospheric composition and temperature changes over glacial and interglacial periods and confirmed that current greenhouse gas concentrations are unprecedented in the context of the last several hundred thousand years. In February 2012, the Russian drilling team breached the bottom of the ice to reach subglacial Lake Vostok [Herszenhorn and Gorman, 2012].

The oldest ice yet recovered is from the European Project for Ice Coring in Antarctica (EPICA) Dome C ice core which extends back 800,000 years. Data from the recent part of the record agree well with those from the Vostok core for their period of overlap [EPICA Community, 2004]. Data from EPICA Dome C are similar to those obtained from Vostok with major ion chemistry, dust concentration, isotopic composition, and greenhouse gas concentrations from entrapped air bubbles all available [EPICA
Community, 2004]. The data from the EPICA Dome C core reveal the relationship between atmospheric dust concentrations and climate over eight glacial and interglacial periods [Lambert et al., 2008]. Additionally, extensive chemical analysis of the EPICA Dome C ice reveals the nature of glacial-interglacial changes in sea salt and non-sea salt ionic species [Wolff et al., 2010]. These analyses have examined relationships between ice core chemistry and both biological productivity and sea ice extent.

Recently, another very deep core was completed as part of the West Antarctic Ice Sheet (WAIS) Divide project. On December 31, 2011 the WAIS Divide team reached a depth of 3,405 meters [Dahnert and Johnson, 2012]. Analysis and interpretation of the WAIS Divide core is currently ongoing, but as of January, 2013 there have been 38 peer reviewed publications related to the WAIS Divide core. A wide variety of analyses have been performed on the WAIS Divide core including entrapped gas analysis [Ahn et al., 2012], black carbon measurements [Bisiaux et al., 2012], and borehole temperatures [Orsi et al., 2012]. The WAIS Divide site experiences much higher accumulation rates than the deep ice core sites from East Antarctica. This higher accumulation rate results in a higher resolution ice core but a shorter record than from East Antarctic cores. The WAIS Divide core provides a 62,000 year climate record for West Antarctica that is comparable in resolution to cores collected from Greenland. Publication of this full climate record is expected to occur in the near future.
In addition to these very deep cores, several intermediate depth cores have been collected from the AP region, closer to the location of the Bruce Plateau (BP) core. These cores are discussed in detail in the next section.

2.2 Antarctic Peninsula Climate Records

There is a large amount of evidence from ice cores [Mosley-Thompson et al., 1990; Thompson et al., 1994; Raymond et al., 1996; Thomas et al., 2008, 2009; Abram et al., 2010], model simulations [Hansen et al., 1999; Dethloff et al., 2010] and reanalysis data [Sime et al., 2009; Thomas and Bracegirdle, 2009] that the climate of the AP has been changing rapidly over the last century. Annual accumulation rates measured in ice cores from the AP have increased over the last century with rapid, dramatic increases since about 1950 [Thomas et al., 2008]. The rapid increase in accumulation rate has been accompanied by a similar change in the isotopic composition of the ice [Thomas et al., 2009], likely indicating a contemporaneous increase in temperature.

Two main types of climate records are available from the AP: proxy records from ice cores and instrumental records from occupied or unoccupied weather stations along the AP. Ice core proxy records have the potential to provide much longer time series of climate data than are available from the instrumental record. Station data from the AP are only available over approximately the last 60 years but provide high resolution climate and weather information which is very accurately dated.
2.2.1 Antarctic Peninsula Ice Core Records

The AP has been the site of several ice core drilling expeditions with a number of intermediate to deep cores retrieved from the area. While the time periods covered by these records are much shorter than those from East Antarctica, the AP presents some advantages for more recent climate records. The accumulation rate on the AP is much higher (ranging from 0.45 meters of w.e. at Dyer Plateau (DP) [Dai et al., 1997] to 1 meter of w.e. at Gomez [Thomas et al., 2008]) than that in East Antarctica (15 to 30 mm w.e. at Vostok [Ekaykin et al., 2004]). Thus many of the AP records are annually resolved for at least several hundred years. In addition, the proximity of AP core sites to the ocean and their more northward location provide an opportunity to examine some processes and conduct analyses that are not possible using continental Antarctic ice cores.

Four cores providing relatively long annually resolved climate histories from the AP are Siple Station, Gomez, DP, and JRI. These four cores are discussed briefly below.

The Siple Station core is from West Antarctica at the base of the AP and provides an annually resolved climate history back to 1410 CE [Mosley-Thompson et al., 1990]. Over this time period the average annual accumulation rate is approximately 0.55 meters of w.e. [Dai et al., 1997]. Isotopic composition and dust analyses from this core show that, during the Little Ice Age period in the Northern Hemisphere (≈1600 to 1830 CE), conditions were warmer and less dusty than average [Mosley-Thompson et al., 1990]. The oxygen isotopic (δ$^{18}$O) record from Siple Station was compared to that from shorter ice core records (James Ross Island [Aristarain et al., 1986], Dolleman Island, and
Palmer Land Plateau [Peel et al., 1988]) and instrumental temperature records from around the AP, and the Antarctic continent. Significant differences were found between the observed temperature records from Siple Station and South Pole Station in East Antarctica [Mosley-Thompson et al., 1990]. These comparisons confirm that there is significant climate variability across Antarctica and that a widely distributed set of records are required to accurately characterize the spatial distribution of climatic variables on the continent.

Following the collection of the Siple Station core, a core from further north on the AP was collected from the DP in 1989/90 [Thompson et al., 1994]. The DP core yields an average annual accumulation rate of 0.45 meters of w.e. and extends back to 1505 CE [Dai et al., 1997]. Extensive chemical analysis of the DP core was performed to determine potential sources of and trends in ionic composition. Sodium to chloride ratios were found to be relatively constant and consistent with that of sea salt. No trends in species concentration over time were observed, indicating little anthropogenic influence on the site in recent decades [Dai et al., 1995]. Data from the Siple Station and DP cores were combined to generate a SH volcanic history based on non-sea salt sulfate (NSSS) fluxes [Dai et al., 1997]. For this analysis, time series of NSSS from both cores were analyzed to identify periods with elevated concentrations. Fifteen events with high sulfate content were identified in the DP core while 18 were found in the Siple core. Of these events, two of the Siple events occurred prior to the start of the DP record and there was excellent agreement between the two records on the timing of the remaining events
with only one event, in 1622 CE, identified in the Siple core not found in the DP core [Dai et al., 1997]. The identification of volcanic events relied on estimating the background NSSS concentration from marine biogenic sources. To account for the varying background concentration of NSSS, only years in which the NSSS concentration exceeded the mean by two standard deviations were considered potential volcanic events [Dai et al., 1997]. This analysis provided a solid volcanic history for the SH identifying all major known eruptions as well a previously unrecognized large eruption in 1809 CE of unknown origin.

A third relatively long core from the AP is the Gomez core collected by researchers from the British Antarctic Survey in 2007. This 136-m core provides a 150-year climate history from the southwestern AP [Thomas et al., 2009]. The δ¹⁸O analysis of this core indicates a warming trend over the last half century [Thomas et al., 2009]. In addition, accumulation at the Gomez site increased from approximately 0.5 meters of w.e. per year in 1850 CE to more than 1 meter of w.e. per year in 2006 [Thomas et al., 2008]. These changes in δ¹⁸O (a temperature proxy) and accumulation are consistent with the large and rapid environmental changes observed on the AP over the last half century. The Gomez core is potentially the core most similar to the BP core due to their relative proximity (see Figure 2.2) and the fact that both sites experience very high annual accumulation rates.

In February, 2008 an ice core was collected from JRI at the northern tip of the AP. Several short cores had previously been collected from the ice cap on this island.
[Aristarain et al., 1982, 1990; Aristarain and Delmas, 2002; Aristarain et al., 2004; McConnell et al., 2007] but this core was the first to be drilled to bedrock from JRI. The core was collected from 57°41.10'W, 64°12.10'S at 1542 m above sea level and extends 363.9 m to bedrock [Abram et al., 2011]. Records obtained from this ice core include accumulation, isotopic, and chemical histories. An analysis of the most recent section of the core (1967-2008) shows that accumulation at the JRI site is higher during years with decreased westerly winds (corresponding to negative Southern Annular Mode [SAM] conditions) [Abram et al., 2011]. Abram et al. [2011] also conducted an investigation into seasonal patterns of sea-salt and marine biogenic compounds and found that sea-salt concentrations were highest during the winter and that the marine compounds (methyl sulfonic acid) were derived from summer productivity in the region just to the south of the site. The history extracted from the JRI core was extended through the entire Holocene [Mulvaney et al., 2012]. The temperature history over the Holocene showed an early Holocene warm period from approximately 12,000 to 9,200 years before present. This was followed by relatively stable conditions that were similar to the present from 9,200 to 2,500 years before present. The last 2,500 years have been characterized by a cooling trend until 600 years before present and a rapid warming trend over the last 600 years.

2.2.2 Antarctic Peninsula Instrumental Records

There are several occupied research stations along the AP. Most stations are located along the western coast of the AP either on the continent or on islands just off of the
coast. The three stations closest to the BP core location (66.03°S, 64.07°W) are Rothera (67.57°S, 68.13°W), Faraday/Vernadsky (65.25°S, 64.26°W), and Palmer (64.77°S, 64.05°W). These stations serve as bases for scientific operations in Antarctica and are situated to allow easy access during the SH summer. During the austral summer sea ice along the western edge of the AP is at a minimum allowing access to these stations by sea. There are no occupied stations along the east coast largely due to the much colder conditions and persistent sea ice or ice shelves.

The first occupied station near the AP was Orcadas Base established in 1904. Orcadas is operated by Argentina and located in the South Orkney Islands just off the northern tip of the AP. While not located on the AP, it is the only station in the vicinity with any meteorological records that predate the 1940s. Data availability for the early part of the record is limited but it does provide some information on the climate of the AP region over the last century.

Faraday Station, established in 1947, was initially operated by the United Kingdom. In 1996, operation of the station was transferred to Ukraine and the station was renamed as Vernadsky. Meteorological data from this station are available continuously from 1947 to the present. Rothera Station was established in 1975 by the British Antarctic Survey and serves as the primary British base of operations on the AP. Continuous meteorological data are available from 1975 to the present. These two stations bracket the BP drill site with Rothera Station slightly to the south (~250 km) and
Faraday/Vernadsky Station slightly to the north (~90 km). The third occupied station in the area is Palmer Station which is operated by the United States. Palmer Station, established in 1968, lies further north and is a Long Term Ecological Research site with available meteorological records extending back to 1968. These three stations provide the most relevant instrumental records from the AP for comparison with the ice core record from the BP site.

In addition to these occupied stations, several automatic weather stations (AWS) have been deployed along the AP for various periods of time. These AWS, operated with limited maintenance, provide meteorological data for remote locations, and thereby provide improved spatial coverage for meteorological observations. Data from AWS are less reliable than from occupied stations as instrument failures or sensor drift cannot be immediately rectified. A disadvantage of AWS data is that they provide only a limited suite of climatic variables. In Antarctica, they are typically not able to provide high quality records of precipitation or atmospheric moisture due to the cold temperatures and the fact that much of the precipitation falls as snow. AWS provide the only longer-term (multiple year) records from the eastern side of the AP. Data from these sites have been critical to understanding the differences in circulation patterns that impact both coasts of the AP [Turner et al., 2002].

The west coast is dominated by westerly maritime flow from the Southern Ocean or South Pacific. The east coast is dominated by continental flow from the Antarctic
interior [Turner et al., 2002]. The difference in air mass source regions results in much warmer temperatures and more precipitation along the west coast and lower temperatures and less precipitation along the east coast. The crest of the AP acts as a topographic divide separating the two regions. The BP ice core was drilled slightly east (≈ 1 to 2 km) of the topographic divide. However, the site is in close proximity to the west coast and likely experiences predominantly westerly maritime flow.

2.3 Large Scale Atmospheric Oscillations

While the climate histories derived from the BP ice core are the main thrust of this research, large scale atmospheric oscillations are useful in interpreting these data (temperature, accumulation, and sea ice extent) and thus are discussed briefly here. Large scale atmospheric oscillations play an important role in determining both weather and climate in many regions around the globe. In the SH, the SAM is one of the primary oscillators [Gong and Wang, 1999]. In addition to its primary variability, the SAM is also modulated by changes in the El Niño-Southern Oscillation (ENSO), a Pacific Basin oscillator with teleconnections (linkages) to many parts of the globe.

The SAM has also been called the high latitude mode or the Antarctic Oscillation Index, defined by Gong and Wang [1999] as the difference of zonal mean sea level pressure between 40°S and 65°S. The term SAM is generally the preferred nomenclature for the observed behavior and will be used exclusively throughout this work to describe this mode of climate variability.
The SAM has two phases; a positive phase when pressures over Antarctica are lower than pressures in the SH mid-latitudes and a negative phase when pressures over Antarctica are higher than pressures in the SH mid-latitudes. Both the sign and magnitude of the SAM exert control on SH climate variations. The SAM has drawn recent attention as it has been trending toward more positive values in recent decades [Marshall, 2003]. Positive SAM values are correlated with cold temperatures over the majority of Antarctica, with the exception of the AP, which exhibits warmer temperatures.

While the positive trend in the SAM over time is relatively clear cut, the cause or causes of the observed changes are not as clear. Modeling studies have examined the potential impacts of different forcing mechanisms, both anthropogenic and natural, on the SAM over time. Arblaster and Meehl [2006] examined potential impacts of natural solar variability, volcanic activity, changes in sulfate aerosol concentration, stratospheric ozone loss, and increases in greenhouse gas concentrations on the SAM. The results indicated that stratospheric ozone loss and increases in greenhouse gas concentrations are the most likely factors contributing to increases in 20\textsuperscript{th} century SAM values. They proposed that while stratospheric ozone loss is likely diminishing in importance at this time, increasing greenhouse gas concentrations over the 21\textsuperscript{st} century will more than compensate for the reduction in stratospheric ozone forcing and thus the SAM is likely to continue to increase over time. Arblaster and Meehl [2006] concluded that the recent local cooling observed over parts of Antarctica due to the increased SAM will be more than offset by
increased radiative forcing due to increases in greenhouse gas concentrations anticipated to warm the entire SH.

In the AP, models do not predict climate changes well as a result of changes in the SAM. Thus, studies have focused on this region to gain a better understanding of the climate response that can be expected here with changes in the SAM [Karpechko, 2009; Carril et al., 2005]. Carril et al. [2005] used a multimodel ensemble approach to focus on the response of the SAM to increasing concentrations of greenhouse gases. Seven models were evaluated for the periods from 1970-1999 and 2070-2099 to compare present and future climate conditions, respectively. With comparison to present climate, the models underestimated sea level pressure in the mid-latitudes, and predicted approximately 20% less sea ice than observed. Surface temperatures were typically 2 to 6 K warmer in the model than in observations for the Southern Ocean but deviations of more than 10 K were found in the Ross Sea region. These future climate simulations were performed using the A2 scenario from IPCC AR4 and focused on October to December, which exhibits the strongest SAM. The future climate scenario model output showed significant warming over the AP with cooling over the rest of Antarctica. Additionally, sea ice extent in the Amundsen and Weddell Seas was predicted to undergo the largest reduction of any area around Antarctica. The predicted seasonal pattern showed warming in the AP during austral spring and summer, but not fall and winter. This modeling study highlights the need for improved climate models at high latitudes, especially in the SH. The model predictions do not agree with observations of climate variables over
Antarctica or the Southern Ocean. The problem is compounded by the relative lack of observational data in this region, as well as the short length of available observations. Due to the relatively large errors observed when attempting to model the present climate, predictions for the magnitude and spatial distribution of future climate changes in the Antarctic must be viewed cautiously.

In addition to the effects of the SAM on temperature and precipitation patterns, other studies have focused exclusively on the impact of the SAM on sea ice extent in the SH. *Lefebvre et al.* [2004] attempted to determine the response of ocean circulation and sea ice to changes in the SAM. They found strong correlations (on a zonal average basis) between the SAM index and the zonal, meridional, and vertical currents predicted by the model as well as strong correlations between the SAM index and the zonal and meridional wind stresses. Intuitively, the stronger westerly winds associated with the positive phase of the SAM should correspond to stronger ocean currents. However, *Lefebvre et al.* [2004] found only weak correlations between the SAM index and the zonal average surface air temperature, zonal average sea surface temperatures, or zonal average sea ice cover in this region.

*Lefebvre et al.* [2004] also investigated the portion of the SAM that is not zonally symmetric. It is these components of the SAM, and of the climate impacts induced by the SAM, that allow differential responses at different longitudes around Antarctica. These local impacts can be substantial and can either enhance or counteract the zonal
mean behavior associated with the SAM. One example of a local effect is the enhanced cooling of sea surface temperatures in the Ross Sea during years with high SAM indices and concomitant warming of sea surface temperatures in the Weddell Sea. Similar anomalies in sea ice cover are found in the Ross (positive anomaly) and Weddell (negative anomaly) Seas during high SAM years. It is important to account for not only the zonal mean behavior associated with the SAM, but to recognize that the differential distribution of land and sea area in the region results in localized impacts that can vary significantly at different locations situated along the same latitude.

ENSO teleconnections between the tropics and the South Pacific-Drake Passage region were examined by Fogt and Bromwich [2006]. They found little correlation between the SAM and tropical sea surface temperatures during the 1980s, but a significant relationship during the 1990s. The SAM appears to weaken during El Niño conditions. The teleconnection is temporally variable throughout the year, with the strongest teleconnection between September and February. The difference in the relationship between the SAM and ENSO between the 1980s and 1990s may indicate a shift in behavior, or reflect several periods during the 1990s with very strong ENSO signals. Thus natural climate variability (e.g., shifts from El Niño to La Niña conditions) may also play an important role in oscillations of the SAM complementing anthropogenic forcing due to ozone depletion and/or increasing greenhouse gas concentrations.
Modeling studies examining the relationship between the SAM and ENSO found that variability in sea surface temperatures in the tropical Pacific can have a driving influence on the SAM [Fogt and Bromwich, 2006; Cai et al., 2011]. The relationship may vary over different time scales from seasonal, to annual, to decadal. It appears that variability in the SAM is not caused exclusively by anthropogenic forcing factors, but also by natural climate variability. Thus, the combination of these multiple forcing factors must be incorporated when attempting to make model predictions for the SAM in the future.

In addition to the SAM and ENSO, there are other large climate oscillations that may be of importance for understanding the climate of the AP. These oscillations may be centered on regions far afield from the AP but nevertheless have impacts that are felt in the region. One such oscillation is the Pacific Decadal Oscillation (PDO). Zhang et al. [1997] noted what they referred to as “ENSO-like” variability in the North Pacific with variability on decadal timescales. The term PDO was coined by Mantua et al. [1997] to describe a pattern observed in sea surface temperatures (SSTs) in the North Pacific that had strong impacts on salmon populations. The PDO was observed to have undergone changes in polarity in 1925, 1947, and 1977 with the periods prior to 1925 and after 1977 characterized as “warm” PDO and the period between 1947 and 1977 as “cool” PDO. The 1977 transition showed a shift to warmer tropical eastern Pacific SSTs and cooler central North Pacific SSTs [Zhang et al., 1997]. The “cool” phase observed between 1947 and 1977 resulted in La Niña-like teleconnection patterns while the warm phase enhances El Niño-like patterns. The interplay between the phase of the PDO and ENSO
may be important for determining the characteristics and strength of the impacts of teleconnections between the northern and tropical Pacific Ocean with the Southern Ocean, and specifically the AP.
Chapter 3: Data and Methods

The primary data used for this dissertation research are from the analysis of a new ice core collected from the Antarctic Peninsula (AP) in 2010. In addition to these primary data, complementary data from two ice cores previously collected from the AP by the Ohio State University Ice Core Paleoclimatology Research Group (OSU ICPRG) were also used in the analysis along with published records from ice cores collected by other groups. Instrumental and reanalysis based meteorological data, and atmospheric oscillation index data were examined to aid in the analysis and interpretation of the ice core results. Each of these data sets is discussed in more detail below. Also discussed in this chapter are the methods employed to obtain a density depth relationship, age scale, and species fluxes for the Bruce Plateau (BP) core.

3.1 Bruce Plateau Ice Core Data

The primary data underpinning this dissertation are the chemical and physical analyses of an ice core collected between December 2009 and February 2010 from the BP on the AP as part of the LARsen Ice Shelf System Antarctica (LARISSA) Project. The LARISSA
Project is an interdisciplinary project designed to investigate the history of ice shelves along the AP. Two ice cores were collected during the field campaign, one short core (≈ 143-m depth), and one deep core (≈ 449-m depth). Three distinct analyses were conducted on co-registered samples over the entire length of the deep core: one sample was analyzed for total insoluble dust concentration as well as particle size distribution; one sample was analyzed for major ion chemistry; and one sample was analyzed for oxygen and hydrogen isotopic ratios. In addition, selected sections of the short core, drilled a few meters from the long core, were analyzed in the same manner to test for reproducibility. A fourth constituent, beta radioactivity, has been measured for selected sections of both cores. Each of these four data sets is discussed in more detail in the following sections.

3.1.1 Sample Handling

Routine analysis of the core over its entire length consists of three co-registered analyses: insoluble dust and particle size distributions, major ion chemistry, and stable water isotope measurements. Samples were cut using a band saw in a cold room at -5°C and the sample size varied with depth with surface samples 10 cm long and 4 cm long samples for the section dated to 1750 CE. Sample dimensions were chosen so that each subsample consisted of at least 5 mL of water. The outer surface (the portion that was in contact with the drill during collection, the core handlers during processing, and then exposed to the plastic tubing during shipment and storage) of each sample was removed with the band saw and samples were taken from the interior of the core. The upper 76
meters of the core were processed as firn and below this depth samples were handled as ice. For firn processing, an additional thickness of core was cut and used as a base to ensure that the samples for dust and major ion chemistry did not come into contact with potentially “dirty” surfaces (e.g., band saw cutting platform, counter top). Additionally, cut firn samples were only handled using specially cleaned gloves to minimize the potential for contamination during the cutting process. Once cut, firn samples were placed in individual pre-cleaned plastic cups and melted for analysis. For ice processing, no base was used, and normal plastic gloves were worn. The samples for insoluble dust and major ion chemistry were transported to a Class 100 Clean Room facility where they were rinsed with ultra-pure 18.2 MΩ MilliQ water prior to melting so that the outer layer of the sample that was in contact with the saw blade, cutting table, or gloves was removed. This rinsing could not be performed for firn samples because of the highly porous nature of firn. Any water that would have come into contact with firn would be absorbed by the sample rather than contributing to melt on the outside. These handling procedures apply to both dust and major ion chemistry samples. Stable isotope samples were melted in plastic cups without rinsing for both firn and ice samples. Beta radioactivity samples were only collected in the ice section of the core, so no additional firn processing steps were required for these samples.

3.1.2 Dust Data

Dust analysis of the BP core was conducted using a Multisizer 4 Coulter Counter configured with a 30 µm diameter aperture tube to provide size distribution information
for 14 particle size ranges, from 0.63 to 16 μm, as well as a count of the total number of particles in a sample. Dust samples were allowed to melt and rise to room temperature prior to analysis. Additionally, a sodium chloride electrolyte solution was added to each dust sample (0.075 ml per ml of melted ice) prior to analysis so that an electrical charge could be measured in the sample. Blanks for dust analysis were analyzed both before and after every batch of samples.

3.1.3 Major Ion Chemistry Data

Major ion chemistry analysis was conducted using a Dionex ICS-3000 Ion Chromatograph (IC). The IC was configured with parallel paths for cation and anion analysis. Cation analysis was conducted using an IonPac® CS12A column with 11.75 mM sulfuric acid eluent. Anion analysis was conducted with an IonPac® AS11 column using sodium hydroxide eluent with a gradient (0.375 – 10.375 mM) method. Calibration standards, blanks, and quality control standards were included with every batch of samples analyzed. Cations measured were sodium (Na\(^+\)), ammonium (NH\(_4\)^+), potassium (K\(^+\)), magnesium (Mg\(^{2+}\)), and calcium (Ca\(^{2+}\)). Anions measured were, methyl sulfonate [MSA] (CH\(_3\)O\(_3\)S\(^-\)), chloride (Cl\(^-\)), nitrate (NO\(_3\)^-), and sulfate (SO\(_4\)^{2-}\).

3.1.4 Stable Isotope Data

Isotopic analysis was conducted using a Picarro L2120-i δD and δ\(^{18}\)O Analyzer. Data obtained from the isotope analysis included the ratio of water with \(^{18}\)O to water with \(^{16}\)O compared to standard mean ocean water (SMOW) as well as the ratio of deuterated water
(HDO) to H$_2$O. These two data sets were combined to determine the deuterium excess (d) of each sample using $d = \delta D - 8\delta^{18}O$ [Dansgaard, 1964].

3.1.5 Beta Radioactivity Data

Beta radioactivity measurements were conducted on selected sections of the core to determine the depth at which atomic bomb horizons could be observed in the ice [Crozaz et al., 1966]. This method estimates the emission of beta particles from the decay of $^{90}$Sr by counting all beta particles and is a reliable measure of anthropogenic radioactivity provided background beta emissions are subtracted. Beta samples were cut from the outside of the core in long, thin strips with one beta sample per tube (approximately 1 meter of depth per sample). The beta samples were weighed using a Mettler PM15-K balance and were then acidified using 4 N hydrochloric acid. Following acidification the samples were allowed to melt overnight. The following day, each melted sample was pumped through two filters (Macherey-Nagel cation and anion exchange filter paper) for two hours so that particulate and ionic material was collected on the filters. The two filters were then removed from the filter holder and allowed to dry before being placed in labeled bags and stored for beta radioactivity analysis. A Tennelec LB 1000 Series Low Background Alpha/Beta Counting System was used for beta radioactivity counting. Beta counts were collected over 500 minutes for each sample and the total counts were recorded on a dedicated logsheet. Blank beta radioactivity measurements were made every three days. The beta radioactivity count for each sample was corrected using the blanks measured prior to and following the sample analysis. The mass of each sample
and the duration of the counting interval were also used to convert the background corrected beta radioactivity measurement to units of decays per hour per kilogram of sample.

3.2 Other Ice Core Data

Two other ice cores from the AP region, Dyer Plateau (DP) and Siple Station, have been collected and analyzed by the OSU ICPRG. Both core sites are further south than the BP ice core location. Data from these cores were used to provide a larger spatial perspective on climate in the AP. The data available from each of these cores are discussed in more detail below.

3.2.1 Dyer Plateau Ice Core

During the 1989-1990 Antarctic field season, two ice cores were collected from the DP. A full description of the cores and the climate history they contain is provided by Thompson et al. [1994] but a brief summary is provided below. The drill site on the DP (70°40’16”S, 64°52’30”W) was located at 2002 m above sea level with a mean annual temperature of -21°C. Here, two cores of approximately 235 m depth (core 1 – 233.8 m, core 2 – 235.2 m) were collected from the ice divide [Thompson et al., 1994]. Samples from the DP core were analyzed for particulate concentration, oxygen isotopic ratios, liquid conductivity, and anion concentrations (Cl⁻, SO₄²⁻, and NO₃⁻). No cation analysis was performed and MSA was not one of the anions selected for analysis. Particulate concentrations, oxygen isotopic ratios, and anion concentrations were determined using
similar methods to those described above for the BP core. Beta radioactivity was also measured on the core and borehole temperatures were logged. The DP core was dated using annual cycles in both $\delta^{18}$O and $\text{SO}_4^{2-}$ and provided a climate history extending back to 1510 CE. The primary DP data of interest for this research project are the accumulation history determined from the thickness of the annual layers, the $\delta^{18}$O history used as a proxy for temperature, and the non-sea salt sulfate (NSSS) history. In addition to the Thompson et al. [1994] paper describing the DP cores, additional discussion of the cores are provided by Dai et al. [1995, 1997] and Raymond et al. [1996].

3.2.2 Siple Station Ice Core

During the 1985-1986 Antarctic field season an ice core was collected from Siple Station ($75^\circ55'S$, $84^\circ15'W$; 1054 masl) located in West Antarctica at the base of the AP [Mosley-Thompson et al., 1990]. The 302-m long core provides a climate history dating back to 1410 CE. The Siple Station core was analyzed for particulate concentration, anion concentrations, and oxygen isotopic ratios. The core was dated using seasonal variations in $\delta^{18}$O and $\text{SO}_4^{2-}$ [Mosley-Thompson et al., 1990; Dai et al., 1997]. Particulate concentrations, oxygen isotopic ratios, and anion concentrations were determined using similar methods to those described for the BP core. As with the DP ice core, ice core-derived histories of interest from the Siple Station core are accumulation, $\delta^{18}$O (temperature proxy), and NSSS.
3.2.3 Other Antarctic Peninsula Ice Core Data

For cores not collected by the OSU ICPRG, only accumulation and isotope data are available for comparison. Accumulation records from the Gomez, ITASE 01_05, and short JRI cores were published by Thomas et al. [2008]. Annual accumulation data from these cores was provided by Elizabeth Thomas for comparison with the BP record. Gomez δ\(^{18}\)O data were obtained from the World Data Center for Paleoclimatology website [Thomas, 2012] and were originally published in Thomas et al. [2009]. ITASE d18O data were obtained from the National Snow and Ice Data Center website [Steig, 2009] and were originally published in Steig et al. [2005], Schneider et al., [2006], and Schneider and Steig [2008]. An isotopic temperature proxy using δD was published for the more recent JRI core by Mulvaney et al. [2012]. The δD data span the entire period covered by the BP core and were provided in the supplementary material for Mulvaney et al. [2012].

3.3 Atmospheric Oscillation Index Data

To examine the relationship of BP climate histories with large scale atmospheric circulation patterns, oscillation indices were used. The Southern Annular Mode (SAM) index has typically been calculated over approximately the last 50 years using reanalysis data from either the National Center for Environmental Prediction/National Center for Atmospheric Research or the European Center for Medium-Range Weather Forecasts. These datasets are very sparse prior to 1979 when satellite data became available and thus, significant uncertainty exists in calculation of the SAM prior to 1979. The
reanalysis data are based on an Empirical Orthogonal Function analysis of 700 mb pressure in the Southern Hemisphere (SH). The SAM is defined as the leading mode of this analysis using data between 1979 and 2000 and is obtained from the National Oceanic and Atmospheric Administration (NOAA) [NOAA, 2012a].

Using station pressure data the SAM index has been reconstructed back to 1865 for austral summer and autumn and to 1905 for austral winter and spring [Jones et al., 2009]. Two of these reconstructions are described by Jones et al. [2009] and Fogt et al. [2009] while a third station-based reconstruction is described by Visbeck [2008]. These reconstructions represent an additional application of SAM modeling other than prediction of future trends in the SAM. Being able to model the SAM back in time could provide valuable insight to past climate and provide context for the recent changes in the SAM which have been observed. Data for SAM reconstructions by Fogt [2009], Visbeck [2009], and Marshall [2012] were used to extend the SAM record back in time before 1979.

A number of El Niño-Southern Oscillation (ENSO) histories are available based on sea surface temperature (El Niño-La Niña) [e.g. Smith et al., 2008] or surface pressure differences between Tahiti and Darwin (Southern Oscillation Index) [e.g. NOAA, 2012a; Ropelewski and Jones, 1987]. Data were obtained from the Climatic Research Unit, University of East Anglia website [Climatic Research Unit, 2013] using the method outlined in Ropelewski and Jones [1987] for comparison with the BP climate histories.
These SAM and ENSO reconstructions and reanalysis-based indices were integrated with the BP climate histories to determine relationships between the ice core data and the atmospheric oscillations. Pacific Decadal Oscillation data from 1900 through 2009 were obtained from the University of Washington’s Joint Institute for the Study of the Atmosphere and Ocean website [Mantua, 2012].

3.4 Antarctic Peninsula Station Data

To examine how well the BP ice core-derived climate histories reflect climate trends and variability in the AP, station-based and automatic weather station (AWS)-based data were used. Hourly meteorological observations for a number of AP stations are available from NOAA’s National Climatic Data Center website [NOAA, 2012b]. Additional data are available from the British Antarctic Survey for permanent stations [British Antarctic Survey, 2012a] as well as for AWS [British Antarctic Survey, 2012b]. The time resolution of these data exceeds that needed for comparison with the annually and possibly seasonally resolved ice core records. Hourly or daily data were averaged to monthly resolution for comparison with ice core data. Further averaging to seasonal and annual averages was required for some analyses so that the ice core proxy records could be accurately compared to the instrumental record.

3.5 Depth-Density Relationship

Freshly fallen snow and ice found at depth in glaciers vary substantially in their physical properties. Over time, snow is converted to ice through a combination of physical
processes by which the density of the snow increases until it reaches the density of ice. This process occurs in stages which are governed by the specific nature of the process as well as the environment. Freshly fallen snow is first converted to firn which is then converted to ice under the pressure of the overlying material.

Models have been developed [Herron and Langway, 1980] to establish the depth-density relationship for the snow-firn-ice series using inputs such as annual average temperature, accumulation rate, density of freshly fallen snow, and density of glacier ice. These models attempt to predict the density at depth using knowledge of the physical processes which convert snow to ice. Models such as these are useful when the environment is well characterized (i.e. temperature and accumulation rate are known) but are less useful when the temperature and accumulation rate are unknown. This can be the case at some remote sites where ice cores are drilled but there is no long term environmental monitoring in place. In cases where the physical environment is not well known, empirical depth-density relationships must be determined.

Depth-density models are used to determine the water equivalent depth of ice cores. Water equivalent depth is the conversion of the depth of snow or ice to an equivalent depth of water using the density of the ice and the density of water. The water equivalent depth will always be less than the actual depth of snow or ice. An understanding of the water equivalent depth is necessary to convert depth in core to annual accumulation (usually measured in meters of water equivalent). As the snow is compacted, the density
increases, and thus the thickness of any layer decreases. The water equivalent depth, however, is not affected by changes in density and will remain constant for a given layer of snow or ice.

For the two ice cores collected from the BP an empirical depth-density relationship was determined based on density measurements conducted at the time the cores were recovered. The dimensions (diameter and length) of the core segments were recorded as they were collected and the mass of each core segment was then obtained so that a density could be calculated. The density calculation relies on the assumption that the ice core segment is a perfect cylinder. This assumption is not completely true for real ice cores as they have uneven edges on the top and bottom as well as the potential to lose some chips along their length due to cracks in the ice. Masses were obtained for ice core sections of approximately 1 meter in length. These cores could be made up either of one continuous piece or of several broken pieces. The density calculation is more accurate for single sections of ice with increasing uncertainty as the core is broken into additional pieces.

For the two cores, dimensions and masses were recorded in the field for the upper 140 meters of the cores. Density calculations were performed for each section but only sections made up of two or fewer pieces without longitudinal fractures were included in the empirical depth-density model development. In addition to the density measurements from the cores, shallow pit samples were collected to determine the density near the surface (in the upper 1 meter of the snow). Pit samples were collected at 10 centimeter
intervals. With increasing depth, the act of collecting a pit sample alters the density of the snow so pit samples deeper than 50 cm were not included in the depth-density modeling. Density measurements from snow pit samples and the two cores were combined to determine the density depth relationship for the BP core.

Figure 3.1 displays the individual density measurements collected from the ice cores and pit samples as well as the empirical depth-density relationship that was determined from the density measurements. The empirical relationship used here is a combination of two separate relationships which were determined independently for the upper section (depth less than 12.2 m) and lower section of the core. The upper section model was developed using data from 0-15 m depth while the lower section model was developed using data from 10-140 m depth. Using a two section model allows for better agreement with the observations and acknowledges that the processes governing densification differ in the upper and lower sections of an ice core. A glacier ice density of 922 kg/m$^3$ was assumed for deep ice. Once the empirical model reached this density, all samples below that depth were assigned a density of 922 kg/m$^3$. The intercept for the upper section of the depth-density relationship gives the density of surface snow to be 293.3 kg/m$^3$ which is within the range of expected fresh snow density [Bohren and Beschta, 1979]. The depth at which the lower section empirical model reaches a density of 922 kg/m$^3$ is approximately 147.75 m. This three stage density depth relationship allows for the conversion of layer thickness at any point in the core to water equivalent thickness given only the depth of the layer.
Figure 3.1 Depth-density model for the Bruce Plateau ice core.
3.6 Time Scale Development

The first step in processing data from the chemical and physical analyses of the BP ice core is to identify annual layers and develop a time scale for the core. Depending on the core, different species may provide the best annual signal to allow for determining the age-depth relationship in the ice. Based on modeling performed prior to the collection of the BP core and observation of the conditions in the field while collecting the ice, a recent accumulation rate of approximately 2 m w.e. per year was estimated. Upon initial analysis of the BP core, methyl sulfonic acid (MSA) was found to show the most pronounced seasonal cycle and provided the most robust basis for age estimation. In addition to MSA, NSSS was also found to exhibit a seasonal cycle and was used to confirm the time scale developed using MSA. It should be noted that the δ¹⁸O profile from the BP core did not show a strong seasonal cycle, rather there were significant variations in δ¹⁸O of much shorter duration than the seasonal fluctuations observed in MSA and NSSS. Because of this shortcoming in the isotopic record, variations in δ¹⁸O were not used to determine the age-depth relationship in the BP core.

The seasonality in MSA measured in the BP core is due to changes in oceanic biological productivity over the course of a year. During the austral summer, centered in January, phytoplankton in the Southern Ocean are actively growing. One result of this growth is
that they emit dimethyl sulfide (DMS) into the atmosphere. Once emitted, DMS is oxidized relatively rapidly following one of two oxidation pathways. One pathway leads to formation of sulfate while the other pathway forms MSA. During the austral winter, the Southern Ocean is typically covered by sea ice and growth of these oceanic phytoplankton is limited. This period of relative biological inactivity results in near zero emissions of DMS and thus also of MSA. While there are other sources of sulfate that may be important (e.g. sea salt, volcanic emissions, anthropogenic emissions), there are no other sources of MSA that are important for remote locations such as the AP. Thus, all of the MSA deposited and measured in the BP core is likely due to biological productivity in the nearby Southern Ocean. This seasonal cycle in MSA emissions has previously been detected in both aerosol measurements [Saltzman et al., 1986; Bates et al., 1992; Jourdain and Legrand, 2001; Gondwe et al., 2004] and ice cores [Legrand et al., 1991; Abram et al., 2010].

Confirmation of this annual cycle is given by the concentrations of MSA measured near the surface of the BP core. The core was collected starting on December 31, 2009 during the austral summer in a period that would be expected to have relatively high MSA concentrations. Figure 3.2 shows a profile of MSA concentrations in the upper 25 m of the BP core. This profile shows that the surface does have high concentrations of MSA that decrease with depth and then increase at a depth that is close to the estimate for annual accumulation at this site. This behavior confirms the annual cycle in MSA is present and useful for annual dating of the BP core.
This cycle in MSA was used to determine the age and thickness of each annual layer. Years were assigned from peak to peak in the annual MSA cycle. This dating scheme approximates a calendar year from January to January as the MSA peak in austral summer likely occurs sometime in January. Drilling of the BP core commenced on December 31, 2009 and thus the surface is approximately dated to be January 2010. The section of core from the surface to the first peak in MSA is then assigned to the year 2009. Layers are counted continuing down the core in this manner. When combined with the depth-density information discussed previously, the thickness of each annual layer can be converted to water equivalent thickness to determine how much accumulation occurred during a given year.
Figure 3.2 MSA Profile in upper 25 m of Bruce Plateau ice core.
In addition to counting of annual layers, the time scale of the BP ice core can be constrained by identification of stratigraphic horizons such as fallout from nuclear bomb testing or volcanic events. Each of these dating methods provides well dated markers that have been confirmed in multiple locations around the globe that can be used to confirm or alter the timescale developed for the BP core through annual layer counting.

Beta counting is a useful dating technique for the period between approximately 1950 and 1970 during which significant nuclear testing was being conducted in the atmosphere. Radioactive material from these tests was dispersed in the atmosphere and became well mixed around the globe. Increased levels of radioactive material in the atmosphere led to increased deposition of those particles on surfaces all over the world. In Antarctica, the radioactive material deposited on the snow was trapped in place as additional precipitation fell over time. This pattern of periodic radioactivity deposition can be used to establish a rough age scale for ice cores using the known dates of major nuclear tests that result in radioactivity horizons in the ice.

Prior to the initial large scale nuclear testing conducted in the atmosphere there was some small background level of radioactivity found in the natural environment. This background level was likely close to the concentration found in the present environment since no atmospheric nuclear testing has been conducted over the last 40 years. Beta radioactivity counting can be used to determine at what depth in an ice core the first measurable quantities of radioactivity, above the natural background levels, are found.
and to assign an age to that depth in the core. The largest beta radioactivity spike is expected to be found around 1964 following large scale nuclear testing in 1962-1963. The estimated accumulation rate obtained from observations of accumulation in the field was used to determine the appropriate depths to analyze for beta radioactivity in the BP cores. Analysis was conducted initially on Core A which is the shorter of the two cores (approximately 143 meters total depth). One sample was taken from each tube between 105 and 143 meters depth (38 total samples). The procedure for collection and analysis of these samples was described previously.

Figure 3.3 shows the beta radioactivity profiles obtained from both the short and long BP cores. The largest peak in radioactivity observed in Core A (short) occurs at approximately 120 meters depth in the core. All of the beta radioactivity values for these cores are relatively low compared to values obtained from previously collected Antarctic cores. This is likely due to the extremely high accumulation rate observed at this site. Nearly the same amount of radioactive material would be expected to be deposited at any Antarctic site as a result of a nuclear test. The high accumulation rate at the BP site means that this deposition is diluted more than it would be at a site with a lower accumulation rate. The radioactivity peak at 120 meters likely corresponds to large nuclear tests in 1962-1963. Data from Core A alone do not provide enough information to identify the initial rise in radioactivity due to 1951 nuclear testing. Core A only extends to a depth of 143 meters and with the high accumulation rate at the site this depth
may capture the potential rise in radioactivity associated with the onset of nuclear testing in 1951, but does not provide any indication of the pre-nuclear era baseline.

Figure 3.3 Beta radioactivity profiles for Core A (black) and Core B (red).
Beta radioactivity measurements from Core B (long) provide confirmation of the radioactivity profile generated from Core A as well as extending the measurements to the pre-nuclear era. Beta samples were analyzed from Core B between 119 and 148 meters of depth again using approximately 1 meter sample sizes. The beta radioactivity profile obtained from Core B is also shown in Figure 3.3. The beta profile for Core B shows that the peak observed in Core A at approximately 141 meters depth is likely the initial 1951 radiation peak with a baseline very close to zero observed for the six samples at the bottom of the depth range analyzed for Core B. The two profiles also show similar structure overall with the 1951 peak near 141 meters depth and then a gradual increase in radioactivity up to the large peak around 120 meters. The depths of some of the peaks is slightly offset between the two cores, but the difference in depth is smaller than the 1 meter size resolution used for the samples. The depth of the upper radioactivity peak at 120 meters falls within the 1964 year (119.490 – 121.549 m) based on annual layer counting. The lower peak, between 140.42 and 141.49 meters depth in Core B, does overlap with the 1951 year (139-179 – 140.528 m) based on annual layer counting. The synchronization of these radioactivity events with the layer counting time scale provides confidence in the robustness of the age scale developed for the BP core.

In addition to beta radioactivity horizons, large volcanic eruptions also provide an opportunity to confirm the time scale generated from the counting of annual layers. These large eruptions produce signals that can be measured in the ice that are unlikely to
be missed. In the BP core, large volcanic eruptions produce a significant signal in NSSS. Over the time period covered in this dissertation there are several major eruptions that are useful for dating. *Gao et al.* [2008, 2009] developed a chronology of volcanic events over 1500 years using ice cores from both Antarctica and Greenland. Ice cores from Antarctica, such as the BP core, are likely to record only those volcanic events which inject sulfate into the atmosphere of the SH. The chronology of *Gao et al.* [2008, 2009] resolves sulfate injection by hemisphere so only those eruptions which have SH impacts need to be considered for the BP core. Table 3.1 displays the dates of the major (>10 Tg stratospheric sulfate injection) SH impacting eruptions since 1750 CE as well as the magnitude (as indicated by stratospheric sulfate injection) and location based on *Gao et al.* [2008, 2009].

**Table 3.1 Dates, magnitudes, and locations of major Southern Hemisphere volcanic eruptions between 1750 and 2000 CE, based on *Gao et al.* [2008, 2009].**

<table>
<thead>
<tr>
<th>Date</th>
<th>SH Stratospheric Sulfate Injection (Tg)</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>1991</td>
<td>15.05</td>
<td>Pinatubo</td>
</tr>
<tr>
<td>1963</td>
<td>11.00</td>
<td>Agung</td>
</tr>
<tr>
<td>1883</td>
<td>10.71</td>
<td>Krakatau</td>
</tr>
<tr>
<td>1835</td>
<td>13.80</td>
<td>Cosigüina</td>
</tr>
<tr>
<td>1815</td>
<td>51.03</td>
<td>Tambora</td>
</tr>
<tr>
<td>1809</td>
<td>26.18</td>
<td>Unknown</td>
</tr>
</tbody>
</table>

Depending on the location of and eruption, and its timing within a year, the impacts in the chemistry of the BP ice core could be delayed by a few months to one year. Typically,
the maximum impact from an eruption will be observed in the year following the event. This delay results in volcanic age controls in 1810, 1816, 1836, 1884, 1964, and 1992 using the major eruptions shown in Table 3.1. A detailed discussion of volcanic event identification is provided in Chapter 8 but a brief outline of the timing of these major eruptions is provided here to assist with time scale verification.

Identification of the most recent eruption, Pinatubo in 1991, is complicated slightly by a recent increase in sulfate likely due to anthropogenic influences. Even with these changing background conditions, the Pinatubo eruption is detected in the layer assigned to 1992 (66.788 – 69.948 m) in the BP core. The next eruption, Agung in 1963, is easily identified in the 1964 (119.490 – 121.549 m) annual layer and is contemporaneous with the beta radioactivity peak from the 1962-1963 atmospheric nuclear tests. The eruption of Krakatau in 1883 is identified within the 1884 (209.183 – 210.068 m) annual layer. The 1835 eruption of Cosigüina is detected in both the 1836 (253.621 – 254.555 m) and 1837 (252.783 – 253.621 m) layers in the BP core. Each of these eruptions from 1835 to the present was found in the expected year using annual layer counting and thus provide strong confidence in the time scale developed for the BP core back to 1835.

The eruption of Tambora in 1815 is well dated in a number of other ice core records and also documented in historical records. The peak impact from the Tambora eruption should be found in 1816 for an ice core from the AP. Initial annual layer counting placed the peak impact from Tambora (270.4 m depth in core) in 1815 rather than 1816. Upon
further examination, one year was removed from the annual layer counting timescale in the 1830s (257.3 m depth in core). This year had been identified due to a decrease in MSA for only one sample as well as three samples with low NSSS concentrations. The short length of the decreases in MSA and NSSS concentrations was highly anomalous for this section of core. The winter above had seven MSA samples below 1.5 ppb and the winter below had six samples below 1.5 ppb. The annual accumulation for the deleted year was the lowest in the period from 1750-2009. For these reasons, and due to the mismatch in time scale between the Tambora eruption and the annual layer counts, this year was deleted from the timescale. Following this adjustment the maximum impact of the Tambora eruption was seen in 1816 (270.276 – 270.848 m), as expected. The 1809 unknown eruption was also identified in the expected year, 1810 (274.423 – 275.542 m), following this time scale adjustment. There were no large volcanic eruptions that could be used for dating control between 1750 and 1809. The time scale developed using counting of annual layers has an uncertainty of ±1 year prior to 1832. The next point in the core where there is uncertainty in the annual layer counting is 1732, prior to the period covered in this work. Thus, the uncertainty in age for the BP core is ±1 year between 1750 and 1832, and < 1 year after 1833.

3.7 Sub-Annual Resolution

Because of the extremely high accumulation rate at the BP site, there is the potential to conduct some analyses, over recent time periods, at sub-annual resolution. This requires more detailed age constraint than is possible using the simple annual layer counting
procedure described in the previous section. Sub-annual resolution at three different scales has been attempted for this ice core. The process of determining these time scales is discussed individually in the following sections.

3.7.1 Summer-Winter Resolution

The annual cycle in MSA which is used to delineate years in the BP core can also be used to differentiate between summer and winter accumulation in the core. During the summer, MSA concentrations are elevated and samples with high MSA can be assumed to have been deposited during the summer. The same procedure can be used to assign low MSA samples to the winter season. There are two challenges associated with this procedure. The first is to determine at what concentration to divide summer and winter accumulation. The second is to decide which months constitute the summer and winter periods. For this work, it was assumed that, over the long term, there is an equal amount of summer and winter precipitation and thus the split between summer and winter should be at the median MSA concentration for the 1750-2009 period. The median MSA concentration for the 4,442 chemistry samples in this interval is 1.63 ppb. Thus, samples with MSA concentrations greater than 1.63 ppb were assigned to summer and samples with concentrations less than 1.63 ppb were assigned to winter. In some cases, smoothing was applied to the data to avoid samples alternating between summer and winter during the seasonal transition. A discussion of which months constitute each season is included within the analysis sections using comparisons to station and reanalysis data.
3.7.2 Seasonal Resolution

Building on the summer-winter split, MSA can further be used to divide each year into four periods corresponding to the seasons. The peak in MSA during the summer was used to determine annual averages and the 1.63 ppb threshold was used to separate summer from winter. Using a process similar to the peak identification method, the minimum MSA concentration was also determined. These points were then used to divide the year into four sections for seasonal resolution. The first season stretches from the summer peak in MSA to the point where MSA drops below 1.63 ppb and corresponds to summer. Next, autumn is defined as the period from 1.63 ppb to the winter MSA minimum. Winter is the period from the minimum to 1.63 ppb and spring is from 1.63 ppb to the next summer maximum. These definitions allow for relatively easy identification of four seasons over the entire time period. These seasonal definitions may vary slightly from traditional seasonal definitions as the summer maximum and winter minimum are used to separate seasons rather than being midpoints within a season. This mismatch in timing can be accounted for in the definition of which months constitute each of the four seasons within this framework. An examination of that partitioning is provided in the analysis chapters.
Chapter 4: Accumulation Rate Reconstruction

Counting of annual layers provides a good estimate of age with depth in the Bruce Plateau (BP) core but does not accurately reflect the net annual accumulation rate history. In order to reconstruct accumulation through time, the thinning of annual layers as they are buried must be taken into account. Changes in annual layer thickness may be due to real changes in accumulation rate or to ice flow within the ice sheet. Several models and assumptions about accumulation rate and ice flow are evaluated to determine a best fit for reconstructing the original thicknesses of the annual layers in the BP core.

4.1 Annual Layer Thinning

The time scales developed in Chapter 3 determine the age of ice at depth within the BP core but additional information is required to determine the original thickness of each annual layer. Individual layers of snow deposited on the surface experience changes in density as they are buried and subjected to pressure of overlying layers. The effect of densification on the annual layers is accounted for using the depth-density relationship developed earlier to convert the layer thicknesses to water equivalent. A second process
that affects the thickness of layers preserved in the core is deformation due to ice flow which further thins the layers in the core.

Ice flow is a complex process that is still a topic of active research [e.g. Martin and Gudmundsson, 2012]. The primary factors driving flow within the ice mass include the slope of the surface and bedrock, environmental conditions at the ice-bedrock interface, the vertical temperature profile, and the vertical and horizontal strain rates. For ice core sites that are visited only once, such as the BP, many of these parameters are not measured at all, or are measured but lack the accuracy required for use in complex two or three dimensional ice flow models. Fortunately, there are a number of simpler one dimensional ice flow models that make some assumptions about ice flow and environmental conditions that can be used to approximate the effects of thinning with depth in an ice core. Often, these models are used to determine age scales for ice cores when clearly identifiable annual layers are not present and the assumption of long term steady state accumulation is valid. In the case of the BP core, the annual layer counting provides an independent age scale for the upper 89% (e.g., 400 meters) of the core so an ice flow model is only needed to constrain thinning of annual layers with depth. These models can be used to determine degree to which a layer has thinned given its depth below the ice sheet surface.
4.2 Strain Rate Based Models

One commonly used ice flow model was developed by Dansgaard and Johnsen [1969] which is an extension of work by Nye [1963]. In Nye’s [1963] original work, the vertical strain rate is assumed to be constant throughout the thickness of the ice, the ice is assumed to be frozen to the bed (i.e., no sliding), and the thickness of the ice sheet is assumed to be constant over time. Using this model, a simple relationship between depth and thinning is developed as:

\[
\lambda_H = \frac{H}{y} \lambda
\]

Equation 4.1

where

\( \lambda \) = reduced layer thickness

\( \lambda_H \) = initial layer thickness

\( y \) = present height above bed

\( H \) = initial height above bed (ice thickness)

The units for each of these parameters are meters of water equivalent (m w.e.). This equation can be applied to every annual layer or every sample to determine its original thickness. The reduced thickness is just the measured length of the sample as it was analyzed or the reconstructed annual layer. The initial height above the bed is the thickness of the ice sheet, converted to m w.e. using the depth-density relationship. The 448 meter BP core contains 388.288 m w.e. depth. The present height above bed is the
ice sheet thickness minus the depth of a sample in m w.e. All these parameters are known and thus the original thickness of each sample can be calculated.

The assumptions inherent in this model vary in their validity for the BP drill site. The ice at this site is below the melting point throughout the thickness of the ice sheet [Zagorodnov et al., 2012] so the assumption that the ice is frozen to the bed is likely valid. The assumption of constant thickness, and thus long term steady-state accumulation, is of unknown validity, but changes in ice thickness are likely to be small over scales of a few hundred years. Evidence from some other ice cores in the region suggests that there has been a recent large increase in accumulation [Thomas et al., 2008]. If the accumulation rate at the BP site has increased similarly the assumption of constant ice thickness may not be valid over the last two centuries. The most troubling assumption is that of a constant strain rate throughout the ice thickness which is modified within the Dansgaard and Johnsen [1969] ice flow model.

The Dansgaard and Johnsen [1969] model assumes that there is a constant strain rate over the upper part of the ice sheet and that the strain rate decreases linearly below a specified depth. This modification to the Nye [1963] model results in a two stage relationship between depth and thinning with a transition point above which the strain rate is assumed to be constant. While the depth of this transition is not known for the BP site, it is a function of distance from the ice divide. At the ice divide, the transition to a constant strain rate occurs at ~70% of the ice thickness, near the divide (within one ice
thickness) the transition occurs at ~50% of ice thickness, and on the flank (more than two ice thicknesses from the divide) the transition occurs at ~20% of the ice thickness [Morse et al., 2002; MacGregor et al., 2012]. The Nye [1969] model assumes that this transition occurs in a very narrow layer at the base of the ice sheet, or 0% of the ice thickness. The BP core site is located approximately 2 km, or four ice thicknesses, east of the ice divide, so flank flow with a transition to constant strain at 20% of the ice thickness may be appropriate for this site.

The relationship between depth and thinning using the Dansgaard and Johnsen [1969] model is:

\[
\lambda_H = \frac{2H-h}{2y-h} \lambda \quad \text{Equation 4.2}
\]

for \(h \leq y \leq H\) and

\[
\lambda_H = \left(\frac{2H-h}{h}\right) \left(\frac{h^2}{y^2}\right) \lambda \quad \text{Equation 4.3}
\]

for \(0 \leq y \leq h\)

where

\(h = \) height of transition to constant strain rate

and all other parameters are the same as in the Nye [1963] model.
For the BP ice core, using $h = 0.2H$ gives a transition point from a linear strain rate to a constant strain rate at 77,658 meters above the bed. The profile differs from the Nye [1963] profile, even above this depth, because the bottom depths experience different strain rates.

While the Dansgaard and Johnsen [1969] model has been used extensively to model ice flow (e.g., Schøtt et al., 1992; Dansgaard et al., 1993; Siegert and Payne, 2004; van Ommen et al., 2004), there are other viable models. Bolzan [1984] used different classes of strain rate profiles to determine thinning of layers at a given depth within the ice sheet. Two different classes of strain rate were evaluated with slightly different forms. Class A flow has a strain rate profile of the form

$$\dot{\varepsilon}(z) = \dot{\varepsilon}_a \left(1 - \left(\frac{z}{H}\right)^p\right)$$  \hspace{1cm} \text{Equation 4.4}

where $\dot{\varepsilon}(z)$ is the strain rate at depth $z$, $\dot{\varepsilon}_a$ is the strain rate at the surface, $z$ is the depth within the ice sheet, $H$ is the ice sheet thickness, and $p$ is a fitting parameter. Class B flow has a strain rate profile of the form

$$\dot{\varepsilon}(z) = \dot{\varepsilon}_a \left(1 - \frac{z}{H}\right)^p$$  \hspace{1cm} \text{Equation 4.5}
where all parameters are the same as in the class A strain rate equation. These strain rate profiles can be integrated and then used to determine the vertical velocity and thinning at depth within the ice sheet.

For class A flow, the vertical velocity with depth is given by

\[ v_z(z) = a(1 - \zeta)^{p+1} \quad \text{Equation 4.6} \]

\[ \zeta = \frac{z}{H} \quad \text{Equation 4.7} \]

where \( a \) is the long term average accumulation rate.

For class B flow, the vertical velocity with depth is given by

\[ v_z(z) = a - a \frac{\zeta}{p} [(p + 1) - \zeta^p] \quad \text{Equation 4.8} \]

for \( p \neq 0 \) and where all parameters are the same as for class A flow.

These vertical velocity profiles can be compared to the annual layer thickness data from the BP core to determine how well each model describes thinning of the annual layers.

With a long term steady state accumulation rate, the original layer thickness in the
Dansgaard and Johnsen (DJ) model is equal to the accumulation rate. This accumulation rate is then used as a fitting parameter in each model. Each model then has two fitting parameters, average accumulation rate and transition height for DJ, or average accumulation rate and exponent for the Bolzan models.

Two methods for fitting these models to the layer thickness data were used. One method minimized the sum of squared differences between observed layer thickness and model layer thickness with depth from the surface to 343 meters depth (2009-1400 CE). The second method allowed for a possible recent increase in accumulation by only fitting the models between 120 meters and 343 meters (1934-1400 CE). The average accumulation rates giving the best fit along with their associated fitting parameters for the different models are shown in Table 4.1 for both time intervals.

<table>
<thead>
<tr>
<th>Interval</th>
<th>Model</th>
<th>Accumulation (m w.e.)</th>
<th>Fitting parameter (h or p)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1400-2009 CE</td>
<td>DJ</td>
<td>1.95</td>
<td>0.18</td>
</tr>
<tr>
<td></td>
<td>Class A</td>
<td>2.14</td>
<td>0.38</td>
</tr>
<tr>
<td></td>
<td>Class B</td>
<td>2.31</td>
<td>0.49</td>
</tr>
<tr>
<td>1400-1934 CE</td>
<td>DJ</td>
<td>1.74</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>Class A</td>
<td>1.80</td>
<td>0.17</td>
</tr>
<tr>
<td></td>
<td>Class B</td>
<td>1.87</td>
<td>0.74</td>
</tr>
</tbody>
</table>

The reconstructed accumulation rates from each model using the parameters given in Table 4.1 are shown in Figure 4.1 (reconstruction based on 1400-2009 CE) and in Figure 4.2 (reconstruction based on 1400-1934 CE).
Figure 4.1 Accumulation reconstructions for three models using fitting parameters determined over 1400-2009 CE interval.
Figure 4.2 Accumulation reconstructions for three models using fitting parameters determined over 1400-1934 CE interval.

Each of these reconstructions has a similar shape with an increasing accumulation trend from ~1500 to 1770 CE, a decreasing trend from ~1770 to 1930 CE, and an increasing trend after 1930 CE. They differ in slope during the early periods and in the magnitude of the accumulation maximum around 1770 CE. All models display a rapid accumulation increase after 1930 CE. As the three models give similar results, further discussion is restricted to the DJ model as it has been widely used to model ice flow in other cores.
4.3 Empirical Models

An alternative method for determining thinning is to fit an empirical thinning curve to the layer thickness data. An empirical relationship can be developed to describe the entire core or a piecewise relationship can be used for different depth intervals. A single relationship empirical fit was employed by Thompson et al. [1985] to model ice flow on the Quelccaya ice cap in southern Peru. Two empirical models are evaluated here, one which was fit to the entire 1400-2009 CE interval and a two stage relationship allowing for the recent increase in accumulation.

4.3.1 Single Stage Relationship

Fitting a polynomial to the measured layer thickness data over the depth interval discussed here (0-343 meters) is one method to account for thinning of annual layers with depth in an ice core. This method assumes that all long-term changes in layer thickness are due to changes in ice flow and not to changes in accumulation rate. A constraint on this type of fitting approach is that the vertical velocity at the bed must equal zero such that thinning approaches infinity. A sixth order polynomial fit to the annual layer thickness data with height above bed provides a good fit to the data \((R^2 > 0.9)\). Table 4.2 presents the coefficients for this sixth order fitting function. Thinning at depth \(z\) can then be calculated as the ratio between the fitting function at the surface and the thinning function at depth \(z\).
Table 4.2 Parameters for sixth order fit of layer thickness with height above bed.

<table>
<thead>
<tr>
<th>Order</th>
<th>Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>-2.5541*10^{-14}</td>
</tr>
<tr>
<td>5</td>
<td>2.53968*10^{-11}</td>
</tr>
<tr>
<td>4</td>
<td>-8.24078*10^{-9}</td>
</tr>
<tr>
<td>3</td>
<td>8.78772*10^{-7}</td>
</tr>
<tr>
<td>2</td>
<td>1.39946*10^{-3}</td>
</tr>
<tr>
<td>1</td>
<td>-1.7838*10^{-4}</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.912</td>
</tr>
</tbody>
</table>

Figure 4.3 shows the accumulation profile obtained from the single stage fitting process. This accumulation history does not show a rapid increase in accumulation over the last 50 years. Because of the fitting procedure used, any long term changes in accumulation are removed and only short term variations are preserved. This accumulation reconstruction assumes a steady-state accumulation that persists over the entire interval. This result probably does not accurately reflect the magnitude of accumulation on the BP, but it may be useful to evaluate relative changes in accumulation over short (decadal to century) time scales.
<table>
<thead>
<tr>
<th>Year (CE)</th>
<th>Accumulation (m w.e.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1400</td>
<td>2.0</td>
</tr>
<tr>
<td>1500</td>
<td>2.5</td>
</tr>
<tr>
<td>1600</td>
<td>3.0</td>
</tr>
<tr>
<td>1700</td>
<td>3.5</td>
</tr>
<tr>
<td>1800</td>
<td>4.0</td>
</tr>
<tr>
<td>1900</td>
<td>4.5</td>
</tr>
<tr>
<td>2000</td>
<td></td>
</tr>
</tbody>
</table>

Figure 4.3 Accumulation rate reconstruction using sixth order polynomial fit (Table 4.2) to the annual layer thicknesses from the Bruce Plateau ice core.
4.3.2 Two Stage Relationship

An alternative to fitting one relationship over the entire core, is to apply a multi-stage relationship between layer thickness and depth. In order to determine the depths to include in each stage, a closer examination of the layer thickness data is required. Figure 4.4 displays the 11-year running average of layer thicknesses with depth. The 11-year smoothing removes some of the year to year variability in accumulation at the BP site. The layer thicknesses can be broken into three distinct sections. Over the upper ~110 meters there is a rapid linear decrease, while there is a slower linear decrease from ~110-240 meters, and below ~240 meters the layer thickness decreases non-linearly.
If the rapid change in layer thickness at the top of the core is assumed to result from a recent increase in accumulation rate to which the ice sheet has not yet equilibrated then a two stage thinning model can be applied. With this assumption, a linear relationship between depth and thinning can be developed between 110 and 240 meters depth and then extrapolated over the upper 110 meters of the core. A separate non-linear thinning relationship can then be developed for the deep section of the core. Using this procedure, and fitting the two pieces together, gives a linear thinning relationship for the upper 242
m of core and a fourth order polynomial relationship for the bottom 146.29 m of core. Thinning from these relationships is calculated as for the single stage relationship discussed previously. Figure 4.5 shows the accumulation reconstruction obtained using this two stage relationship. This reconstruction gives a steady state accumulation rate prior to 1930 CE and shows a large accumulation increase over the last 80 years. There are still short term variations in accumulation over the early part of the record, but this model forces the long term accumulation rate, prior to the recent increase, to be constant.

Figure 4.5 Accumulation reconstruction using a two stage piecewise empirical fit to annual layer thickness data and allowing for a recent increase in accumulation.
4.4 Model Selection

The models described here present a range of options for reconstructing the actual accumulation rate history on the BP. Selecting one model requires comparing their relative merits and features. Six potential accumulation reconstructions (Figure 4.6) were considered in making this decision. The raw layer thicknesses are included to represent an extreme case where there is no thinning of layers with depth in the core and all changes in layer thickness are due to changes in accumulation. Four DJ models were included with transitions to constant accumulation at 0, 0.12, 0.18, and 0.2 times the ice thickness. The first (transition at 0 ice thickness) is the Nye model. The second and third (transitions at 0.12 and 0.18 ice thicknesses, respectively) represent the best fit models over the time periods presented in Section 4.2. The fourth DJ model (transition at 0.2 ice thicknesses) is a standard model for ice flow on the flank of an ice cap. Also included are the two empirical fits discussed in the previous section.
The layer thickness and single stage polynomial reconstructions provide useful end-members for possible accumulation histories. The layer thickness profile assumes that all
variations in layer thickness result from changes in accumulation rate while the polynomial profile assumes that all long-term trends result from changes in ice flow and that the current accumulation rate has been constant over the reconstructed period. Neither of these assumptions is likely to be valid and thus these reconstructions are rejected for use on the BP ice core. These profiles may, however, provide guidance as to which of the other profiles is most realistic for the site.

Of the remaining profiles, the Nye and DJ 0.12 reconstructions are rejected as they do not adequately account for thinning in the lower section of the core. The accumulation profiles obtained from these two reconstructions show behavior similar to that for the layer thickness profile prior to 1780 CE and likely reflects inaccurately modeling the thinning in this section. The DJ 0.18 and DJ 0.2 profiles are very similar, as expected given the similarity in their forms, so only the DJ 0.2 profile is considered further as it has been used as a basis for other ice flow modeling. This screening leaves only the DJ 0.2 reconstruction and the two stage empirical reconstruction for consideration. Figure 4.7 shows these two reconstructions with the others removed so their similarities and differences can be seen more easily.
Figure 4.7 Bruce Plateau accumulation reconstructions using the Dansgaard and Johnsen model with $h=0.2H$ (black) and the two stage empirical model (red).

From 1400-1550 CE both reconstructions show relatively steady-state accumulation but the DJ model gives a higher accumulation rate. From 1550-1775 CE there is a slow increase in accumulation using the DJ model that is absent with the empirical model. Accumulation then decreases from 1775 CE through the 1930s in the DJ reconstruction but remains constant in the empirical model. From the 1930s on, both models show similar trends in accumulation with a rapid increase to present conditions. All short term fluctuations are similar between the two models as it is only the long term accumulation
pattern that changes. The choice between these two models requires deciding if the high accumulation rate from 1400-1775 CE and decrease from 1775 through the 1930s is real or if the steady state conditions provided by the empirical fit model are more realistic.

Accumulation records from other ice cores on the Antarctic Peninsula (AP) may be useful in making this assessment. Many of these records are from cores which penetrate through less than half of the thickness of the ice sheet at the location where they were drilled with some of them penetrating only about one third of the ice thickness. These short cores provide accumulation records that are not as strongly affected by the choice of ice flow model because the importance of ice flow increases with depth in the ice sheet. For example, if the BP core had been drilled through only one third of the ice thickness at the site, the record would extend back to only 1943 CE. The reconstruction of accumulation rate over this upper section of the core is largely independent of the thinning model selected. Ice core records from lower accumulation sites at Dyer Plateau (DP), Siple Station, Gomez, ITASE 01_05, and James Ross Island (JRI) have longer records in the upper third of the ice sheet and provide a regional accumulation history that is not greatly influenced by annual layer thinning. Eleven-year average accumulation records from these cores are shown in Figure 4.8 over the periods available for each record (or back to 1700 CE for DP and Siple Station). Of these records, the DP, Siple Station, and Gomez cores show no indication of a decreasing trend in accumulation between 1770 and 1930 CE. The Gomez record begins in the 1850s so it provides no information for the early period, but accumulation increases from the beginning of the
record to the present. The DP and Siple Station records both show relatively steady state accumulation with a slight increase since 1700 CE. The JRI core shows an increasing trend over this interval as well, with a possible slight decrease in accumulation between the start of the record in 1832 CE and 1900 CE. Identification of this trend in the JRI core is complicated as the accumulation oscillates with an ~35 year periodicity. The ITASE core shows a decreasing trend in accumulation between 1890 CE and 1970 CE and the relative magnitude is similar to that in the DJ reconstruction of the BP core but the timing is off. The decrease in ITASE accumulation starts ~120 years after that at the BP. In general, the accumulation histories from other AP ice cores do not provide evidence for decreasing accumulation in this region from 1770 CE to 1930 CE. These records suggest that the empirical reconstruction may be a better estimate of BP accumulation than the DJ reconstruction unless some local influence is dominating accumulation at the BP site during this time period. It is difficult to envision a century scale shift in atmospheric circulation that would drastically impact accumulation at the BP site and not the other AP ice core sites. This comparison with other ice core records from the AP suggests that the DJ model may not accurately reflect long term changes in accumulation at the BP site.
Figure 4.8 Accumulation reconstructions from other Antarctic Peninsula ice cores.

Isotope data from the BP core may provide additional guidance as to which accumulation reconstruction to use. The annual isotope history does not depend on the thinning model employed so it should provide independent verification of the accumulation reconstruction. Theoretically, warmer temperatures (less depleted isotope values) should
result in an increase in atmospheric moisture, and thus an increase in precipitation which should be reflected by an increase in the accumulation. In practice this is not always the case. Running 11-year average $\delta^{18}O$ and accumulation data for the DP and Siple Station cores (Figure 4.9 and Figure 4.10, respectively) show that short term fluctuations in $\delta^{18}O$ and accumulation are often correlated, this is not consistently the case for longer term trends.

Figure 4.9 Accumulation (red) and $\delta^{18}O$ (black) for the Dyer Plateau ice core.
Figure 4.10 Accumulation (red) and $\delta^{18}$O (black) from the Siple Station ice core.

Similar profiles for the two BP accumulation reconstructions are shown in Figure 4.11 (DJ model) and Figure 4.12 (empirical model). The DJ model shows good long term agreement between changes in $\delta^{18}$O and accumulation, with the accumulation minimum in the 1930s coinciding with the $\delta^{18}$O minimum. The empirical accumulation reconstruction does not show the same degree of agreement between $\delta^{18}$O and accumulation. This comparison suggests that the DJ model may provide a better accumulation reconstruction than the empirical model. Data from the other cores
however, show that there is not always a long term correlation between $\delta^{18}O$ and accumulation at a specific site.

Figure 4.11 Accumulation reconstructed using the Dansgaard and Johnsen [1969] flow model (red) and $\delta^{18}O$ for the Bruce Plateau ice core.
Figure 4.12 Accumulation reconstructed using the empirical flow model (red) and $\delta^{18}$O for the Bruce Plateau ice core.

This evidence, coupled with the accumulation records from other cores in the region, does not definitively identify one reconstruction method as superior to the other. A final piece of evidence that may help distinguish the better reconstruction comes from the fact that annual layers are still distinguishable at 343 meters, which exceeds 75% of the ice thickness (448 meters). This implies that the accumulation rate is very large and suggests that the DJ model is more likely to be correct as it predicts a larger steady state accumulation rate than the empirical model. Additionally, the DJ model has been used extensively to model ice flow in other cores and is physically based. For these reasons,
the DJ model with $h=0.2H$ is used to reconstruct accumulation for the analyses presented in this work. In some instances, the empirical accumulation reconstruction is also included to show how using that model would affect the analyses presented here.

This discussion highlights the challenges encountered in reconstructing annual layer thicknesses, particularly in the lower portion of an ice sheet. Ice flow, which has relatively little effect on layer thinning in the upper third of an ice sheet, becomes increasingly important with depth. The BP core presents a unique opportunity to gain a better understanding of this flow because the very high accumulation rate allows annual layers to be counted over a large fraction of the ice thickness. This tight age control means that thinning models can be evaluated for their performance rather than being used to determine the age of ice at a given depth. The uncertainty in thinning with depth can only be determined because of the independent age scale obtained from annual layer counting. Reconstructed accumulation rates from cores that extend beyond the upper third of the ice thickness have inherent uncertainty as ice flow dynamics can vary with time and depend on complex topography both on the surface and in the bedrock. Over long time scales, reconstructed isotopic histories are likely to have less uncertainty than accumulation reconstructions over the same intervals, especially for high accumulation sites such as the BP where the effects of isotopic diffusion are likely to be small.
Chapter 5: Comparison to Satellite and Instrumental Records

The establishment of Faraday Station on the west coast of the Antarctic Peninsula (AP) (Figure 5.1) in 1947 CE provided the first long-term meteorological monitoring in the vicinity of the Bruce Plateau (BP) drill site. Since then, several other surface monitoring stations have been installed in the region. These stations provide the best data for comparison with the proxy records extracted from the BP ice core. The advent of routine satellite-borne measurements in 1979 CE dramatically increased the availability of data from remote regions such as the AP, provided the first information about long term conditions over the Southern Ocean and is critical for placing proxy records, such as those from the BP ice core, in a larger spatial context. The following sections discuss the proxy data from the BP ice core, as well as short term meteorological measurements collected from the drill site, in the context of nearby surface observations and satellite-based measurements.
5.1 Meteorological Measurement Comparison

Immediately after completion of ice core drilling on the BP, an automated meteorology ice geophysics station (AMIGOS) was deployed at the drill site and collected data from February 9, 2010 until July 20, 2010, often intermittently near the end. The station
collected temperature, relative humidity, pressure, dew point, wind speed, and wind
direction. Data were recorded and transmitted at hourly resolution over 162 days. Data
completeness degraded significantly over the life of the station with very few
measurements during June and July. The degradation in data quality resulted from the
high accumulation rate at the site which progressively elevated the snow surface and
completely submerged the station in July. The wind variables were not measured as
reliably as the temperature, moisture, and pressure variables due to ice formation (riming)
on the instrumentation. Table 5.1 summarizes the available data for the wind and other
(moisture, pressure, and temperature) variables by month and includes the number of
valid measurements and the fraction of possible measurements collected. Data
completeness drops off significantly in June and July so that this analysis extends only
from February 9 to May 25, 2010.

<table>
<thead>
<tr>
<th>Month</th>
<th>Possible Hours</th>
<th>Wind Count</th>
<th>Wind Fraction</th>
<th>Other Count</th>
<th>Other Fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td>February</td>
<td>480</td>
<td>233</td>
<td>0.485</td>
<td>478</td>
<td>0.996</td>
</tr>
<tr>
<td>March</td>
<td>744</td>
<td>199</td>
<td>0.267</td>
<td>696</td>
<td>0.935</td>
</tr>
<tr>
<td>April</td>
<td>720</td>
<td>49</td>
<td>0.068</td>
<td>568</td>
<td>0.789</td>
</tr>
<tr>
<td>May</td>
<td>744</td>
<td>461</td>
<td>0.620</td>
<td>562</td>
<td>0.755</td>
</tr>
<tr>
<td>June</td>
<td>720</td>
<td>138</td>
<td>0.192</td>
<td>176</td>
<td>0.244</td>
</tr>
<tr>
<td>July</td>
<td>480</td>
<td>24</td>
<td>0.050</td>
<td>92</td>
<td>0.192</td>
</tr>
</tbody>
</table>
To determine how representative other meteorological stations in the area are of conditions on the BP the temperature and pressure data from the BP station were compared to those from other meteorological observing stations in the region. From February 9 to May 25, 2010, there were 12 meteorological stations operating in the AP region with data available through the National Oceanic and Atmospheric Administration’s (NOAA) Climate Data Online website [NOAA, 2013]. Locations of the stations and the BP site are shown in Figure 5.1.
Table 5.2 shows the name, latitude, longitude, and elevation of each of these stations as well as the BP drill site. The table is organized from northernmost to southernmost with stations grouped by location. Four stations are grouped relatively closely at the northern tip of the AP, three are along the west coast to the north of the BP, and five are south of the drill site (two of which are on the east side of the AP). Most sites are near sea level and none are higher than 200 m elevation (with the exception of the BP drill site at 1975.5 m).
Table 5.2 Meteorological stations operating in the vicinity of the Bruce Plateau between February and June, 2010.

<table>
<thead>
<tr>
<th>Region</th>
<th>Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Peninsula</td>
<td>Joinville Island</td>
<td>-63.183</td>
<td>-55.4</td>
<td>75</td>
</tr>
<tr>
<td></td>
<td>Bernardo O’Higgins</td>
<td>-63.317</td>
<td>-57.9</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>Base Esperanza</td>
<td>-63.4</td>
<td>-56.983</td>
<td>24</td>
</tr>
<tr>
<td></td>
<td>Base Marambio</td>
<td>-64.233</td>
<td>-56.717</td>
<td>198</td>
</tr>
<tr>
<td>Northwest Coast</td>
<td>Palmer Station</td>
<td>-64.766</td>
<td>-64.083</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>Bonaparte Parkpoint</td>
<td>-64.783</td>
<td>-63.067</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>Vernadsky Station</td>
<td>-65.25</td>
<td>-64.266</td>
<td>11</td>
</tr>
<tr>
<td>Central Peninsula</td>
<td>Bruce Plateau</td>
<td>-66.03</td>
<td>-64.07</td>
<td>1975.5</td>
</tr>
<tr>
<td>East Coast</td>
<td>Larsen Ice Shelf AWS</td>
<td>-67.013</td>
<td>-61.47</td>
<td>45</td>
</tr>
<tr>
<td>Southwest Coast</td>
<td>Rothera Station</td>
<td>-67.57</td>
<td>-68.124</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>Base San Martin</td>
<td>-68.117</td>
<td>-67.133</td>
<td>4</td>
</tr>
<tr>
<td>Southern Peninsula</td>
<td>Fossil Bluff</td>
<td>-71.317</td>
<td>-68.283</td>
<td>55</td>
</tr>
<tr>
<td>Southeast Peninsula</td>
<td>Butler Island AWS</td>
<td>-72.206</td>
<td>-60.17</td>
<td>115</td>
</tr>
</tbody>
</table>

Data for temperature and sea level pressure were obtained for each station at the highest resolution available and then averaged to daily resolution for ease of comparison among the sites. Temperature, moisture, and pressure variables at each of the stations were correlated with those at the BP site from February 9 to May 25, 2010 (Table 5.3). The pressure correlations are higher than those for temperature at all sites. Temperature and pressure at Rothera Station are most strongly correlated with observations on the BP. Base San Martin, which is close to Rothera, shows very similar results with a slightly weaker correlation for temperature. As might be expected, the correlations weaken with increasing distance northward from the BP site. The eastern and southern stations are more difficult to characterize as a group since they are not as tightly spaced as the northern stations. Pressure correlations with the Larsen Ice Shelf and Fossil Bluff stations are strong while those with the Butler Island automatic weather station (AWS),
much further south along the east coast of the AP, are weaker. Temperature correlations show a similar pattern except for the Larsen Ice Shelf AWS which is relatively close to the drill site (~150 km to the ESE). Temperatures here are more weakly correlated with those on the BP than any site except for the Butler Island site far to the south. This weak correlation indicates that the BP, despite being slightly east of the topographic divide, is influenced more by conditions to the west of the AP than by conditions to the east, at least during this relatively short period of comparison.

Table 5.3 Correlation between temperature and pressure measured on the Bruce Plateau and at other meteorological monitoring stations in the Antarctic Peninsula region. All correlations are significant at the 99% level.

<table>
<thead>
<tr>
<th>Region</th>
<th>Name</th>
<th>Temperature Correlation</th>
<th>Pressure Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Peninsula</td>
<td>Joinville Island</td>
<td>0.68</td>
<td>0.78</td>
</tr>
<tr>
<td></td>
<td>Bernardo O’Higgins</td>
<td>0.64</td>
<td>0.78</td>
</tr>
<tr>
<td></td>
<td>Base Esperanza</td>
<td>0.69</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>Base Marambio</td>
<td>0.68</td>
<td>0.88</td>
</tr>
<tr>
<td>Northwest Coast</td>
<td>Palmer Station</td>
<td>0.77</td>
<td>0.90</td>
</tr>
<tr>
<td></td>
<td>Bonaparte Parkpoint</td>
<td>0.76</td>
<td>0.92</td>
</tr>
<tr>
<td></td>
<td>Vernadsky Station</td>
<td>0.69</td>
<td>0.92</td>
</tr>
<tr>
<td>East Coast</td>
<td>Larsen Ice Shelf AWS</td>
<td>0.57</td>
<td>0.91</td>
</tr>
<tr>
<td>Southwest Coast</td>
<td>Rothera Station</td>
<td>0.81</td>
<td>0.97</td>
</tr>
<tr>
<td></td>
<td>Base San Martin</td>
<td>0.79</td>
<td>0.97</td>
</tr>
<tr>
<td>Southern Peninsula</td>
<td>Fossil Bluff</td>
<td>0.69</td>
<td>0.92</td>
</tr>
<tr>
<td>Southeast Peninsula</td>
<td>Butler Island AWS</td>
<td>0.50</td>
<td>0.81</td>
</tr>
</tbody>
</table>

To examine the stability of these relationships over time, correlations over running seven day periods for temperature (Figure 5.2) and pressure (Figure 5.3) were examined for a subset of the stations. To reduce the number of plots per figure, one station from each of
the six regions (Table 5.3) is shown. This comparison reveals that there is significantly less variability in the pressure relationship than the temperature relationship between the BP and the other stations. The temperature relationship also shows that, at different times, different stations are more strongly correlated with the BP. During the 100 averaging periods, each of the six regions has the highest correlation with the BP at some point. There are also periods with significant negative correlation ($r < -0.50$) for each region except Rothera. Overall, meteorological conditions at Rothera are most strongly correlated with those over the BP, although the significant variability over the course of the week means that the BP is influenced by air masses originating from different directions around the AP.
Figure 5.2 Running seven day temperature correlation coefficients between the Bruce Plateau and other Antarctic Peninsula meteorological stations.
Figure 5.3 Running seven day pressure correlation coefficients between the Bruce Plateau and other Antarctic Peninsula meteorological stations.
The overall relatively strong correlation between the BP meteorological observations, particularly temperature, and those from other regional stations provides some guidelines for interpreting the proxy records of temperature and accumulation extracted from the BP core. These data indicate that, over a three month period (from February through May), the correlation coefficient between temperature on the BP and temperature at any of the meteorological stations in the region does not exceed 0.81. This provides an upper bound for correlation between the ice core-derived temperature proxy ($\delta^{18}O$) and surface temperature observations. Actual correlations are expected to be lower than this due to the variable nature of the relationship (Figure 5.2) and the fact that air masses from different regions are likely to influence the BP over the annual time resolution of the ice core data. The meteorological data suggest that the BP $\delta^{18}O$ record should be most closely related to temperatures measured at Rothera Station. However, the short duration of meteorological measurements available from the drill site prohibit a robust analysis as this relationship may vary from season to season. Conditions at Rothera are most similar to those on the BP, but the degree of correlation with the other monitoring stations is sufficiently high that there may be value in comparing the BP $\delta^{18}O$ history to temperature observations from other sites in the region.

Table 5.4 shows the average and standard deviation of temperature for each station from February 9 through May 25, 2010. The data show that there is not a coherent trend in temperature with warmer temperatures in the north and colder temperatures in the south.
Other factors, including station elevation and east-west location relative to the spine of the AP are important as well. The warmest temperatures are observed at the Northwest Coast stations with average temperatures around 0°C and the smallest standard deviations (~2.6°C). The Southwest Coast stations are slightly colder, around -2°C with slightly larger standard deviations (~3 to 4°C). The Northern Peninsula stations experience highly variable temperatures that are all colder than the coastal stations farther to the south. The colder temperature at these stations likely reflects more continental flow moving northward from the Antarctica interior along the east coast of the AP. This atmospheric flow regime effect also affects the Larsen Ice Shelf and Butler Island sites located further south on the east side of the AP where temperatures are cold and highly variable. Fossil Bluff, located nearly as far south as Butler Island but on the east side of the AP is 10°C warmer on average, likely due to the dominance of westerly maritime flow over southerly continental flow. As expected, the BP drill site is 11-13°C colder and more highly variable on average relative to the stations along the west coast although temperatures are even more variable at the sites east of the topographic divide. The temperature on the BP is colder than the coastal stations because of the relatively high elevation of the site as well as the potential for occasional continental Antarctic influences from air masses moving north along the east side of the AP which is reflected in the larger standard deviation of temperature on the BP relative to western coastal
stations. Mean temperature offsets between the BP and Rothera Station and Faraday/Vernadsky Station are 11°C and 13°C, respectively.

Table 5.4 Average and standard deviation of temperature observations from the Antarctic Peninsula between February 19 and May 25, 2010.

<table>
<thead>
<tr>
<th>Region</th>
<th>Name</th>
<th>Average Temperature (°C)</th>
<th>Temperature Std. Dev. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Peninsula</td>
<td>Joinville Island</td>
<td>-7.31</td>
<td>7.19</td>
</tr>
<tr>
<td></td>
<td>Bernardo O’Higgins</td>
<td>-3.02</td>
<td>4.14</td>
</tr>
<tr>
<td></td>
<td>Base Esperanza</td>
<td>-5.97</td>
<td>7.36</td>
</tr>
<tr>
<td></td>
<td>Base Marambio</td>
<td>-10.97</td>
<td>8.64</td>
</tr>
<tr>
<td>Northwest Coast</td>
<td>Palmer Station</td>
<td>0.01</td>
<td>2.66</td>
</tr>
<tr>
<td></td>
<td>Bonaparte Parkpoint</td>
<td>-0.31</td>
<td>2.59</td>
</tr>
<tr>
<td></td>
<td>Vernadsky Station</td>
<td>-0.94</td>
<td>2.70</td>
</tr>
<tr>
<td>Central Peninsula</td>
<td>Bruce Plateau</td>
<td>-13.23</td>
<td>4.49</td>
</tr>
<tr>
<td>East Coast</td>
<td>Larsen Ice Shelf AWS</td>
<td>-18.27</td>
<td>10.68</td>
</tr>
<tr>
<td>Southwest Coast</td>
<td>Rothera Station</td>
<td>-2.08</td>
<td>3.36</td>
</tr>
<tr>
<td></td>
<td>Base San Martin</td>
<td>-2.20</td>
<td>4.25</td>
</tr>
<tr>
<td>Southern Peninsula</td>
<td>Fossil Bluff</td>
<td>-9.56</td>
<td>7.36</td>
</tr>
<tr>
<td>Southeast Peninsula</td>
<td>Butler Island AWS</td>
<td>-19.56</td>
<td>8.28</td>
</tr>
</tbody>
</table>

5.2 Isotope Comparison to Station Data

Comparisons of the BP ice core-derived δ¹⁸O record and station based temperature data are essential to determine whether δ¹⁸O can be used as a reliable proxy for regional temperature. For the BP core, this comparison can be performed at time resolutions ranging from annual to seasonal. The high resolution nature of the record may allow for determination of seasonal differences in the relationship between conditions on the BP and at the coastal stations if such differences exist. Two long term stations, Rothera (RS)
and Faraday/Vernadsky (FVS) are used for these comparisons since their records are most strongly correlated with those on the BP and the other station records are much shorter. Monthly temperature measurements for both RS and FVS were aggregated into seasonal or annual averages for comparison with the isotope data. All data were obtained from the British Antarctic Survey READER website [British Antarctic Survey, 2012a].

5.2.1 Annual Average Comparisons

The Rothera dataset begins in March, 1976 but data are flagged as incomplete until April, 1977. July and August of 1999 are also flagged as incomplete. Data completeness for the flagged months ranges from 19% to 89%. For this comparison, all months were included regardless of the data completeness for the monthly average. Annual average isotope values correspond approximately to calendar years so this comparison was conducted from 1977 through 2009 CE as those are the years with data available for all months from RS.

The annual average δ¹⁸O from the BP ice core and the annual average temperature from RS from 1977 through 2009 CE (Figure 5.4) are well correlated (r = 0.47, p = 0.008). Significant correlations are evident in both the year to year variability and the long term trend. As expected, this correlation is much smaller than that for ~four months of temperature observations from the BP and RS (r = 0.82). Over this same interval the correlation between the BP core δD record and RS temperatures is slightly lower (r =
0.43, \( p = 0.012 \) and thus \( \delta^{18}O \), which is slightly better correlated with RS temperature is used as the temperature proxy for all further comparisons.

**Figure 5.4** Annual average Bruce Plateau \( \delta^{18}O \) (black) and Rothera Station temperature (red) from 1977 through 2009 CE.

Figure 5.5 shows a similar plot of \( \delta^{18}O \) and temperature from FVS from 1947 through 2009 CE. The correlation over the entire record is \( r = 0.46 \) (\( p < 0.0001 \)). For the period of overlap with the RS record, (1977-2009 CE), the correlation is \( r = 0.41 (p = 0.009) \). Correlation coefficients between FVS temperature and \( \delta D \) are 0.48 and 0.36 for the two
time periods, respectively. As expected, the correlations are lower than that for observed temperatures at BP and FVS, but are still significant over the entire interval.

![Figure 5.5 Annual average Bruce Plateau $\delta^{18}O$ (black) and Faraday/Vernadsky Station temperature (red) from 1947 through 2009 CE.](image)

These correlations are based on calendar year data from the stations and annually reconstructed $\delta^{18}O$ or $\delta D$ records. To explore the impact of temporal differences in the records, correlations were calculated for 12-month periods sequentially offset by one month throughout the calendar year (November-October through May-April) with similar results for all species. For FVS, correlation coefficients for all twelve month periods
were between 0.45 and 0.47. The lower correlations between the isotope and temperature histories versus both temperature histories undoubtedly reflects some uncertainty in the timing of the split between years in the ice core and the fact that accumulation on the BP is unlikely to be evenly distributed throughout the year.

5.2.2 Sub-Annual Comparisons

Sub-annual comparisons of $\delta^{18}O$ with temperature at RS and FVS were conducted on both semi-annual (6 months) and seasonal (3 months) time scales. Data from FVS, the longer record, were used to determine which three month periods showed the strongest correlation with the seasonal BP $\delta^{18}O$ record. Using this procedure, the best seasonal periods were January-March (summer), April-June (autumn), July-September (winter), and October-December (spring). Since the divisions between seasons, Summer-Winter, and annual averages use the same split points (based on methyl sulfonic acid [MSA] concentrations) a single timeline was used for consistency. For Summer-Winter comparisons, Summer was the sum of spring and summer seasons (October-March) and Winter was the sum of autumn and winter seasons (April-September). Annual averages were then equal to calendar years (summer-autumn-winter-spring). It is important to note that these seasonal definitions do not match the standard seasons for Antarctica. This is due to their definition with respect to MSA concentrations measured in the ice core and dividing seasons based on peaks and valleys rather than having seasons centered on the peak or valley of MSA. Regardless of the sub-annual divisions selected, there is some uncertainty in the establishment of a sub-annual time scale. The procedure used here
defines four equal length seasons which maximize correlation with station temperature over the entire year.

The six sub-annual comparisons of BP $\delta^{18}$O and RS temperature (Figure 5.6) confirm their relationship on sub-annual timescales. The seasonal difference in $\delta^{18}$O is not as marked as that in temperature, but summer $\delta^{18}$O is typically less depleted than winter $\delta^{18}$O. There is more noise in the $\delta^{18}$O data than the temperature data and more noise in the three month data than the six month data as would be expected. Correlation coefficients for BP $\delta^{18}$O and RS temperature (Table 5.5) for annual and sub-annual periods demonstrate that $\delta^{18}$O from the BP provides a reasonably reliable proxy for regional air temperature. The reliability of the proxy decreases as the time scale is reduced, i.e. annual averages are more reliable than seasonal averages. This is expected due to uncertainty in assigning a sub-annual timescale as well as potential year to year shifts in the timing of events (e.g., the summer MSA peak) used to establish the timescale.
Figure 5.6 Sub-annual Rothera Station temperature (red) and Bruce Plateau $\delta^{18}$O (black).
Figure 5.7 is similar to Figure 5.6 but for FVS instead of RS. Similar patterns are seen in these plots over a longer period of time. There is not as much of a difference in the $\delta^{18}$O between the seasons as seen in the temperature data and there is more year to year variability in the $\delta^{18}$O data than the temperature data. There is also more year to year variability in the winter temperature (winter, seasonal autumn, and seasonal winter) than the summer temperature. Year to year variability in $\delta^{18}$O does not vary significantly between seasons. Table 5.5 shows the correlation coefficients between BP $\delta^{18}$O and FVS temperature for annual averages as well as each of the seasonal periods. In order to compare FVS with RS, separate correlation coefficients are shown for 1947-2009 and 1977-2009 CE for the FVS data.
Figure 5.7 Sub-annual Faraday/Vernadsky Station temperature (red) and Bruce Plateau $\delta^{18}$O (black).
Table 5.5 Correlation coefficients between Bruce Plateau $\delta^{18}$O and station temperatures. Correlation coefficients significant at the 95% level are shown in bold text.

<table>
<thead>
<tr>
<th></th>
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<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Annual</td>
<td>Jan-Dec</td>
<td>0.466</td>
<td>0.421</td>
<td>0.457</td>
</tr>
<tr>
<td>Summer</td>
<td>Oct-Mar</td>
<td>0.300</td>
<td>0.246</td>
<td>0.361</td>
</tr>
<tr>
<td>Winter</td>
<td>Apr-Sep</td>
<td>0.315</td>
<td>0.323</td>
<td>0.401</td>
</tr>
<tr>
<td>Seasonal-Summer</td>
<td>Jan-Mar</td>
<td>0.158</td>
<td>-0.029</td>
<td>0.274</td>
</tr>
<tr>
<td>Seasonal-Autumn</td>
<td>Apr-Jun</td>
<td>0.477</td>
<td>0.348</td>
<td>0.365</td>
</tr>
<tr>
<td>Seasonal-Winter</td>
<td>Jul-Sep</td>
<td>0.195</td>
<td>0.143</td>
<td>0.345</td>
</tr>
<tr>
<td>Seasonal-Spring</td>
<td>Oct-Dec</td>
<td><strong>0.419</strong></td>
<td><strong>0.317</strong></td>
<td><strong>0.321</strong></td>
</tr>
</tbody>
</table>

These correlation coefficients show a significant relationship between BP $\delta^{18}$O and temperature at both RS and FVS on an annual average basis. When the time scale is reduced to six months, correlations are generally lower but remain significant except FVS summers from 1977-2009 CE. Further reducing the period of comparison to three months results in significant correlations in autumn and spring but not summer and winter for both sites during the 1977-2009 CE interval. Significant correlations are found for all sub-annual time periods when using the entire 63 year FVS record. The poor correlation at finer time scales may reflect mismatches in assigning the sub-annual timescale. Errors could be introduced by changes in the timing of the summer peak/winter minimum in MSA or the transition from high to low MSA concentrations during autumn and spring. Another potential source of error is in the definition of the seasons. Here, each season was defined to be three months and seasons had to fill all
twelve months. If there is significant month to month variability in accumulation, it is possible that periods with lower accumulation will not correlate as well with temperature records. A concurrent project estimating the monthly accumulation at the BP site has the potential to provide additional insight to these comparisons and possibly alter the sub-anual timescale. Although not yet complete, the initial findings indicate that there is a minimum in accumulation from December through February.

5.2.3 Conclusions

The analysis presented here indicates that $\delta^{18}O$ can be used as a reliable proxy of annual average regional air temperature. This allows a long term reconstruction of air temperature from the $\delta^{18}O$ history obtained from the BP core dating back to at least 1750 CE. When extended to shorter time scales the reliability of the $\delta^{18}O$ proxy is somewhat diminished but there is still potential to evaluate changes in summer-winter temperatures over longer time periods and thereby extend the instrumental record back over the time interval for which annual resolution is available in the BP ice core. Comparisons at seasonal (3 month) resolution are less certain, but correlations with FVS temperature over 63 years indicate the potential to extract some information from $\delta^{18}O$ at seasonal resolution.

5.3 Trends in $\delta^{18}O$

The correlation between $\delta^{18}O$ and temperature allows the BP $\delta^{18}O$ record to serve as a proxy for regional air temperature. Trends in annual and sub-annual $\delta^{18}O$ can be
examined for temperature changes at the site over time. These trends can be compared to
the trends in measured station temperatures over the length of the instrumental record to
determine how well the observed temperature changes are recorded in the $\delta^{18}O$ record.
These trends can be compared on both annual and sub-annual time scales.

Table 5.6 displays the trends in temperature from both RS and FVS and $\delta^{18}O$ from the BP
on annual and sub-annual timescales. Trends are shown for both 1977-2009 CE and
1947-2009 CE, corresponding to the length of the observational records at RS and FVS,
respectively. All trends are shown on a per decade basis. The largest warming trends at
both RS and FVS occur during the winter months (April-September) while the $\delta^{18}O$ data
indicate more warming on the BP during summer than winter. The trends in $\delta^{18}O$ from
1977-2009 CE for the winter periods are close to zero with autumn (July-September)
showing a slight decrease in $\delta^{18}O$ over this time. From 1947-2009 CE, the $\delta^{18}O$ trends
during winter are similar to those in summer and for the annual average. The temperature
trends at FVS over this longer interval still show more warming during the winter than
during the summer.
Table 5.6 Trends in station temperature and $\delta^{18}$O on annual and sub-annual time scales.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rothera</td>
<td>Faraday</td>
</tr>
<tr>
<td></td>
<td>Trend</td>
<td>Trend</td>
</tr>
<tr>
<td></td>
<td>°C/decade</td>
<td>°C/decade</td>
</tr>
<tr>
<td>Annual</td>
<td>0.688</td>
<td>0.779</td>
</tr>
<tr>
<td>Summer</td>
<td>0.392</td>
<td>0.370</td>
</tr>
<tr>
<td>Winter</td>
<td>0.967</td>
<td>0.980</td>
</tr>
<tr>
<td>Seasonal-Summer</td>
<td>0.320</td>
<td>0.231</td>
</tr>
<tr>
<td>Seasonal-Autumn</td>
<td>0.787</td>
<td>0.553</td>
</tr>
<tr>
<td>Seasonal-Winter</td>
<td>1.146</td>
<td>1.769</td>
</tr>
<tr>
<td>Seasonal-Spring</td>
<td>0.496</td>
<td>0.565</td>
</tr>
</tbody>
</table>

The difference in the temperature and $\delta^{18}$O trends on sub-annual time scales raises questions as to why these two temperature indicators do not show similar changes over the last 33 years. One possible explanation is that different circulation patterns bring air masses to the BP at different times of year. Instrumental observations from the BP extended only from February through May so if there are differences in circulation patterns, and therefore correlations with the coastal stations to the west, over the course of a year, those changes would not be evident in the instrumental data used to compare conditions on the BP with the other stations from the region. Reanalysis data providing greater spatial coverage may reveal the mechanism(s) responsible for the observed differences between BP $\delta^{18}$O and temperatures measured at coastal monitoring stations. A spatial analysis of $\delta^{18}$O and surface air temperature is provided in the next section.
5.4 Correlation with Reanalysis Data

Single point correlation fields can be calculated using time series of data from the BP core and spatially resolved reanalysis data sets. For the analysis presented here, NCEP/NCAR reanalysis data [Kalnay et al., 1996] were used to generate the correlation plots. All plots were generated using an online tool provided by the NOAA/ESRL Physical Sciences Division, Boulder, Colorado on their website at http://www.esrl.noaa.gov/psd/data/correlation. This tool allows users to generate plots of correlation fields between pre-selected single variable time series (e.g., Southern Oscillation Index, Pacific Decadal Oscillation, Southern Annular Mode) or user supplied time series and reanalysis field data. The reanalysis data may be analyzed on different time scales (annual, seasonal, monthly) and the start and end dates for the correlation plots can be selected. Reanalysis fields containing temperature, pressure, atmospheric moisture, and wind variables can be evaluated at several atmospheric levels.

The time series of δ¹⁸O on annual, summer-winter, and seasonal time scales were compared with surface air temperature and correlation maps were used to identify likely source regions for air masses bringing precipitation to the BP. These plots were generated down to seasonal scale to evaluate if there are differences in these source regions over the course of the year. Plots were also generated for running ten year periods to identify changes in source regions over time. Interpretation of correlations with reanalysis data are restricted to 1979-2009 CE as the reanalysis data may be unreliable in the vicinity of Antarctica prior to the satellite era [Hines et al., 2000].
5.4.1 Annual Correlation Fields

The annual correlation of BP δ18O and surface air temperature (SAT) between 1979 and 2009 CE (Figure 5.8) reveals a region of high correlation ($r > 0.7$, $p < 0.00001$) to the west of the AP in the Bellingshausen Sea which confirms that the predominant circulation pattern in this region brings air from the Bellingshausen Sea to the BP where precipitation is deposited with δ18O values that are highly correlated with surface air temperatures over the Bellingshausen Sea. There are no other areas of high correlation (Figure 5.8) which strongly indicates that region is the dominant source of air masses bringing precipitation to the BP.
Figure 5.8 Annual correlation between Bruce Plateau $\delta^{18}$O and surface air temperature from the NCEP/NCAR reanalysis between 1979 and 2009 CE.

Further analysis of annual correlations between $\delta^{18}$O and surface air temperature was conducted to evaluate changes in spatial correlation patterns over time. Correlation fields similar to Figure 5.8 were generated for all ten year periods between 1979 and 2009 (1979-1988, 1980-1989, …, 2000-2009) which results in 22 separate correlation fields that show changes in the region most highly correlated with BP $\delta^{18}$O. Four examples (1980-1989, 1990-1999, 1997-2006, and 2000-2009) are shown in Figure 5.9. The 1980-1989 correlation field is representative of the early years of the comparison with an area
of high correlation centered just west of the AP but also extending slightly to the east of the AP and covering a large spatial extent. The size of the strong correlation ($r > 0.7$) gradually shrinks over time until it is much smaller and centered further west, as shown in the 1990-1999 correlation field. The region of strong correlation continues to shrink until a transition period shown by the 1997-2006 correlation field. For this ten year period, there is an irregularly shaped area of strong correlation that extends along the west coast of the AP, but also encompasses the northern AP (including the BP drill site) and a significant area to the east of the AP over the Weddell Sea. As shown for 2000 to 2009, this transition culminates with three correlation maxima to the west, north, and east of the AP. This transition suggests that there may have been a transition in atmospheric flow bringing moisture to the BP with westerly flow dominating from the 1980s through the mid-1990s with increasingly diverse flow during the late-1990s and 2000s.
Figure 5.9 Annual correlation fields between Bruce Plateau $\delta^{18}$O and NCEP/NCAR reanalysis surface temperature for (A) 1980-1989 CE, (B) 1990-1999 CE, (C) 1997-2006 CE, and (D) 2000-2009 CE.
5.4.2 Seasonal Correlation Fields

To identify changes in the regions of strongest correlation between BP δ¹⁸O and SAT over the course of the year, similar correlation fields (Figure 5.10) were calculated for summer (Jan-Mar), autumn (Apr-Jun), winter (Jul-Sep), and spring (Oct-Dec) from 1979 to 2009 CE. The maximum correlations for these seasonal periods are all weaker than those for an annual average basis but regions of strong correlation are present. The correlation in summer is greatest to the west of the AP along the coast of Antarctica. Autumn shows the strongest correlation with the region of strong correlation to the north and west of the AP. Winter correlations are generally weak and no regions have correlation greater than $r = 0.5$. The maximum correlation is over the Bellingshausen Sea to the west of the AP. Spring exhibits a distinctly different correlation pattern with the maximum to the east of the AP over the Weddell Sea. These correlation plots strongly indicate that large differences in the atmospheric circulation, and hence in the moisture sources, prevail on the BP throughout the year.
Figure 5.10 Seasonal correlation fields between Bruce Plateau $\delta^{18}$O and NCEP/NCAR reanalysis surface temperature from 1979-2009 CE for (A) Summer (Jan-Mar), (B) Autumn (Apr-Jun), (C) Winter (Jul-Sep), and (D) Spring (Oct-Dec).
To identify temporal trends as in the annual correlation field seasonal correlations are also presented as ten year running correlations. Figure 5.11 shows the summer (Jan-Mar) correlation fields over the same time periods as for the annual average (Figure 5.9). For the 1980 to 1989 CE period, there is a large area of strong correlation over the Bellingshausen Sea to the west of the AP but over time, this region shrinks and moves eastward closer to the AP. The 1990-1999 CE correlation field shows that the region of strongest correlation to the west of the AP transitions to the northern tip of the AP and to the east. By 1997 to 2006 CE this transition is complete with BP δ¹⁸O and most strongly correlated with SAT in a band extending from the base of the AP along the east coast past its northern tip. This pattern persists through the 2000-2009 CE period with increasing correlations east of the AP that exceeds r = 0.7. This spatial transition in the BP δ¹⁸O-SAT relationship during the Austral summer is striking and suggests that the Bellingshausen Sea region was the primary moisture source during the early part of the record, but not during the most recent decade. Similarly, the Weddell Sea region appears to become the dominant moisture source during summer over the last decade.
Figure 5.11 Seasonal correlation fields for Summer (Jan-Mar) between Bruce Plateau $\delta^{18}O$ and NCEP/NCAR reanalysis surface temperature for (A) 1980-1989 CE, (B) 1990-1999 CE, (C) 1997-2006 CE, and (D) 2000-2009 CE.
Similar plots for autumn (Apr-Jun) show little variability in the spatial distribution of maximum correlation (Figure 5.12). The region of maximum correlation lies generally to the west of the AP over the Bellingshausen Sea. The region of maximum correlation is centered over the AP for 1980-1989 CE but is smaller and further west for 1990-1999 CE. By 1997-2006 CE the maximum correlation extends along the length of the west coast of the AP and north to southern tip of South America. This pattern remains in the 2000-2009 CE correlation field although with somewhat diminished strength. For autumn, the main difference in the correlation fields over time is the large negative correlation in recent years over the South Pacific, Southern Ocean, and much of the Antarctic continent. This large area of negative correlation is absent in the two early time periods and the magnitude is greatest in the 2000-2009 CE correlation plot.
Figure 5.12 Seasonal correlation fields for Autumn (Apr-Jun) between Bruce Plateau $\delta^{18}$O and NCEP/NCAR reanalysis surface temperature for (A) 1980-1989 CE, (B) 1990-1999 CE, (C) 1997-2006 CE, and (D) 2000-2009 CE.
Similar correlation fields for winter (Jul-Sep) show a region of maximum correlation to the northwest of the AP, just off the west coast of South America for 1980-1989 CE (Figure 5.13). The region of strongest correlation gradually moves southward until it is centered over the Bellingshausen/Amundsen Sea for 1990-1999 CE. Simultaneously, correlations over West Antarctica reach a maximum. For the 1997-2006 CE interval strong correlations exist with many regions including the Amundsen Sea, the northern tip of the AP, the Weddell Sea, and over a large portion of East Antarctica. The widespread nature of these strong correlations may suggest that atmospheric flow is increasingly diverse during winter over this time period. The most recent decade shows a markedly different correlation field with no positive correlation to the west of the AP and only relatively weak positive correlation to the east over the Weddell Sea. This correlation plot for 2000-2009 CE is much different from those observed for the other time periods investigated and suggests the probability of a recent shift in winter circulation patterns.
Figure 5.13 Seasonal correlation fields for Winter (Jul-Sep) between Bruce Plateau $\delta^{18}$O and NCEP/NCAR reanalysis surface temperature for (A) 1980-1989 CE, (B) 1990-1999 CE, (C) 1997-2006 CE, and (D) 2000-2009 CE.
Correlation fields for spring (Oct-Dec) reveal a significant shift in region of maximum BP $\delta^{18}$O-SAT correlation (Figure 5.14). From 1980-1989 CE, the region of maximum correlation lies along the west coast of the AP over the Bellingshausen Sea but shifts to the northern tip of the AP, with a secondary region of strong correlation to the east of the AP during 1990-1999 CE. The shift continues through the 1997-2006 CE period when the strongest correlations are over the Weddell Sea with a secondary maximum at the northern tip of the AP. The transition is complete by the 2000-2009 CE period when a strong maximum lies over the Weddell Sea. This shift in the location of the correlation fields is more striking in spring than in the other seasons. The BP $\delta^{18}$O-SAT correlation over the Bellingshausen Sea transitions from strongly positive to neutral/weakly negative while the correlation over the Weddell Sea transitions from strongly negative/neutral to strongly positive.
Figure 5.14 Seasonal correlation fields for Spring (Oct-Dec) between Bruce Plateau $\delta^{18}O$ and NCEP/NCAR reanalysis surface temperature for (A) 1980-1989 CE, (B) 1990-1999 CE, (C) 1997-2006 CE, and (D) 2000-2009 CE.
These seasonal comparisons confirm that there have been significant changes in the regions supplying moisture to the BP. One possible explanation is a shift in atmospheric circulation patterns at different times of year. There is a strong correlation between BP $\delta^{18}O$ and SAT over the Bellingshausen Sea for all seasons from 1980-1989 CE but this correlation pattern has changed in recent years, with only autumn continuing to show a strong positive correlation between $\delta^{18}O$ and SAT over the Bellingshausen Sea. More recently the region of strongest correlation in spring and summer has shifted eastward over the Weddell Sea. Over the 31 year period, the correlation between $\delta^{18}O$ and SAT in winter weakens and changes sign in the Bellingshausen Sea while a moderately strong positive correlation develops over the Weddell Sea.
Chapter 6: Large Scale Atmospheric Oscillations

Reconstructions of the major large scale atmospheric oscillations based on station data begin around 1900 CE and characterize changes in circulation patterns over periods of years to decades. This variability may be recorded also in ice cores as large scale circulation changes will affect temperature and accumulation at specific locations. The Southern Annular Mode (SAM) is the most important mode of variability in Southern Hemisphere (SH) circulation and plays a dominant role in controlling conditions on the Antarctic Peninsula (AP). The SAM is also modulated by oscillations centered in other regions, such as the El Niño-Southern Oscillation (ENSO) or Southern Oscillation Index (SOI) in the tropical Pacific Ocean, and the Pacific Decadal Oscillation (PDO) in the northern Pacific Ocean. There are complex interactions among these three oscillating systems that determine weather patterns on the Bruce Plateau (BP). This chapter examines the relationships between temperature and accumulation reconstructions from the BP core, as well as from other AP ice cores, and these three oscillations.
6.1 Accumulation Record

These large scale oscillations may be compared with the BP accumulation record from 1900 to 2009 CE which is very well dated with time scale uncertainty of less than one year. As this period is contained in 194.23 m of core, or roughly the upper half of the BP ice sheet, potential errors in the accumulation reconstruction arising from the choice of ice flow model should be minimized. The annual accumulation record (Figure 6.1) is nearly constant from 1900 CE until 1940 CE followed by a slight increase through the 1970s and a rapid increase thereafter. Accumulation exhibits large interannual variability around an average value of 1.84 m w.e. with a standard deviation of 0.56 m w.e. The precise timescale coupled with high interannual variability makes this site ideal for exploring how large scale circulation patterns affect the accumulation on the AP. The longer term (decadal scale) increase in accumulation may also be a product of changes in atmospheric circulation. Influences on both interannual and decadal time scales are examined in the following sections.
Figure 6.1 Annual and 11-year running mean accumulation history from the Bruce Plateau ice core from 1900 to 2009 CE.

6.2 Atmospheric Oscillations

The three oscillations examined here, the SAM, SOI, and PDO, have their individual characteristic variability and trends on annual to decadal time scales. The following sections provide information on the data used to describe each oscillation, the time period covered, and the important features of each oscillation since 1900 CE.
6.2.1 Southern Annular Mode

There are several different SAM indices that cover different time periods and are derived using different methods. This research uses the SAM index reconstructed by Fogt et al. [2009] that provides annual and seasonal resolution between 1905 and 2005 CE. Other SAM indices (Marshall [2012], NOAA [2012a], Jones [2009], and Visbeck [2008]) were evaluated but are not discussed here because their overall characteristics were found to be similar to those of the Fogt et al. [2009] reconstruction. Moreover, the temporal extent of the Fogt et al. [2009] reconstruction makes it better suited for the analysis of trends over the time period covered here. Figure 6.2 shows the annual average and the 11-year running mean of Fogt et al.’s [2009] reconstructed SAM index. Annual averages were calculated using four seasonal SAM reconstructions (DJF, MAM, JJA, SON) and thus the SAM average represents December through November rather than a calendar year. Annual average data from the BP core are assumed to correspond approximately to calendar years but uncertainty in the timing of summer maxima means that there is uncertainty in the split between years of one to two months. The SAM shows large interannual variability over the 20th Century as well as a possible ~30 year periodicity over much of the century followed by an increasing trend toward a positive SAM after ~1980.
6.2.2 El Niño-Southern Oscillation

The Southern Oscillation is the atmospheric component of ENSO and is measured as the difference in sea level pressure between Tahiti and Darwin. These longer term station records provide a long duration observational measure of ENSO activity. SOI data are available at monthly resolution from the Climatic Research Unit, University of East Anglia [Climatic Research Unit, 2013] from 1866 CE to the present and were averaged to annual resolution for this comparison. Annual averages were calculated on a calendar year basis. Negative SOI values correspond to El Niño conditions and positive SOI
values correspond to La Niña conditions. Figure 6.3 shows the annual average and 11-year running mean SOI from 1900 to 2009 CE. Over the early part of the record there is no long term trend in the SOI but since 1980 CE it has trended toward more negative values. There is significant year to year variability in the SOI over this interval.

![Graph showing annual average and 11-year running mean SOI from 1900 to 2000 CE. The graph indicates that before 1980 CE there was no long-term trend in the SOI, but since then it has trended toward more negative values. There is significant year-to-year variability in the SOI over this interval.]

Figure 6.3 Annual average and 11-year running mean SOI.
6.2.3 Pacific Decadal Oscillation

The PDO describes sea surface temperatures in the northern Pacific Ocean and shows significant decadal scale variability. Data for the PDO were obtained from Mantua [2012] for the period between 1900 and 2009 CE at monthly resolution and averaged to annual resolution on a calendar year basis. Figure 6.4 shows the annual average PDO and 11-year running mean PDO between 1900 and 2009 CE. The PDO shows a 40-50 year cycle with positive PDO peaks around 1940 and 1980 CE and a negative PDO period centered near 1960 CE. This periodicity is the major defining feature of the PDO with modest year to year variability.
Figure 6.4 Annual average and 11-year running mean PDO.

6.3 Accumulation - Oscillation Relationships

The nature of the relationships between accumulation and each of these three oscillations were examined over their periods of overlap as well as for shorter intervals. Anecdotal evidence of possible relationships is also considered with emphasis placed on years with extreme values of accumulation or the oscillation indices.
6.3.1 Accumulation - Southern Annular Mode Relationship

The recent increasing trend in the SAM index has been used to explain the increased precipitation on the AP as a result of increased westerly winds and cyclonic activity in this region that correspond to positive SAM conditions. Fogt [2007] used outgoing longwave radiation and atmospheric moisture measurements between 1971 and 2000 CE to identify the location of the predominant SH storm tracks for different SAM and ENSO conditions. Years were grouped according to oscillation index values and atmospheric moisture anomalies were used to identify areas with above or below average precipitation to determine where the predominant storm tracks were located. Figure 6.5, adapted from Fogt [2007] shows the location of both enhanced storm tracks (green) and weakened storm tracks (brown) for positive SAM conditions (upper panel) and negative SAM conditions (lower panel). The BP ice core site is represented by a black dot. This schematic shows that, with positive SAM conditions, there is enhanced atmospheric moisture in the vicinity of the BP with drier conditions present to the north. Conversely, under negative SAM conditions drier conditions dominate near the BP and wetter conditions occur further north. This relationship underpins the argument that the recent trend toward more positive SAM conditions has resulted in an increase in accumulation on the AP.
Figure 6.5 Enhanced and weakened storm track positions (1971-2000 CE) for positive SAM conditions (upper panel) and negative SAM conditions, adapted from Fogt [2007].

Figure 6.6 shows annual and 11-year running mean accumulation and SAM values together. One interesting feature of this plot is the similar increase seen in both the 11-year average accumulation and SAM index since the late-1970s. This longer term correlation between accumulation and the SAM is not evident in the early part of the record when the mean values fluctuate independently (e.g., peak in SAM during the 1930s is not accompanied by a similar increase in accumulation). Additionally, many, but not all, years with high accumulation also have high SAM index values and years with low accumulation have low SAM index values. Since 1980, the three years with SAM values less than -1 (1980, 1991, and 2002) have annual accumulation rates less than
the 11-year running mean centered on those years. Similarly, the six years with SAM values greater than +1 (1989, 1993, 1998, 1999, 2001, and 2003) all have annual accumulation rates greater than the 11-year running mean centered on those years. However, this is not always the case. For example, accumulation in 1997 was much lower than the 11-year running mean accumulation but the SAM index was +0.71. Visually, there appears to be a relationship between accumulation and the SAM, but it is not always consistent.

![Figure 6.6 Annual average and 11-year running mean accumulation (black) and SAM (red).](image-url)
The correlation coefficient, $r$, between annual accumulation from the BP core and the SAM index is 0.323 with a p-value of 0.001 for 1905-2005 CE. Although statistically significant, SAM only explains ~10.5% of the variance in accumulation. An initial analysis of this relationship for the post 1970 period produced a much stronger relationship between accumulation and the SAM but as the record was extended back in time the relationship weakened. In order to investigate this, running 11-year correlations (Figure 6.7) were calculated to determine if the relationship has been temporally stable or if it has changed over time. The early part of the record, until 1948 CE, is characterized by generally strong positive correlation between accumulation and the SAM but this is followed by a fairly rapid transition to a negative correlation that is sustained from ~1948 to 1974 CE. Note that the absolute values for the periods of positive and negative correlation are nearly identical and that the early (1905-1948 CE) and late (1974-2005 CE) periods of positive correlations have nearly identical $r$ values. As a reference, for 11-year running correlations, the 95% significance level is $r = 0.602$, the 90% significance level is $r = 0.521$, and the 80% significance level is $r = 0.419$. These correlations generally fall in the 80-90% significance range with some periods greater than 90% significance. While the significance levels are not impressive, the dramatic shifts between positive and negative correlations in the late 1940s and mid 1970s are compelling evidence that some type of transition is occurring during these periods. The early and late periods show the expected relationship between accumulation and the SAM while the relationship during the middle period is opposite that predicted by the storm track analysis of Fogt [2007].
Figure 6.7 Running 11-year correlation coefficient between annual accumulation from the Bruce Plateau ice core and the SAM index.

6.3.2 Accumulation - Southern Oscillation Relationship

Fogt [2007] performed a storm track analysis for ENSO conditions similar to that for SAM conditions. Figure 6.8 shows the locations of enhanced and weakened storm tracks with positive (top) and negative (bottom) SOI conditions. This plot shows that the BP region is likely to be wetter than average under positive SOI conditions and drier than average under negative SOI conditions. The storm track identified by Fogt [2007] does not extend the wetter or drier conditions over the AP, but the location of the storm tracks
suggests that the pattern to the west of the AP would likely lead to changes in precipitation over the BP.

![Figure 6.8 Enhanced and weakened storm track positions for positive SOI conditions (upper panel) and negative SOI conditions, adapted from Fogt [2007].](image)

Figure 6.9 shows the annual and 11-year running mean accumulation and SOI values, similar to Figure 6.6 for the SAM. Accumulation is not as well correlated with the SOI as with the SAM index. The recent trend toward more negative SOI conditions is not accompanied by reduced accumulation as predicted from the storm track analysis. There is however, anecdotal evidence of year to year variability in the SOI having affecting accumulation. Accumulation in 1987 and 1997 is much lower than the 11-year averages.
centered on those years and both years have large negative SOI values. Such year to year variability is evident throughout the record but the long term trends in accumulation do not mimic those in the SOI.

![Figure 6.9 Annual average and 11-year running mean accumulation (black) and SOI (red).](image)

A correlation analysis between accumulation and the SOI, performed identically to that for the SAM over the same interval (1905-2005 CE) to facilitate comparison, yields an r value of 0.124 (p < 0.217). This correlation with accumulation is much weaker than that for the SAM (r = 0.323; p < 0.001). Similarly to examine the stability of the relationship...
over time, the running 11-year correlations were calculated (Figure 6.10). Overall the BP accumulation and the SOI are positively correlated, in agreement with the storm track analysis of Fogt [2007]. The correlation during the early part of the record is relatively weak, with typical values between 0.1 and 0.4 until the late 1940s when correlation values increase, ranging between 0.5 and 0.9. In the early 1970s the correlation drops rapidly with a short interval of negative correlation centered on 1980 CE that is followed by another period of relatively strong positive correlation that is stronger than the early period, but not as strong as the period between the late 1940s and 1970. Overall, the correlation between accumulation and the SOI appears to be generally positive but differs in strength on decadal time scales.
6.3.3 Accumulation - Pacific Decadal Oscillation Relationship

While the PDO varies on much longer time scales than the SAM or the SOI, it may exert an influence on the accumulation observed on the BP. No analysis of the effects of the PDO on SH storm tracks was included by Fogt [2007] as the PDO was in a persistent positive phase for nearly the entire analysis window (1971-2000 CE). The time period covered in this work does allow some analysis of the accumulation-PDO relationship because there are both long term positive and negative PDO phases between 1905 and 2005. Figure 6.11 shows the annual average and 11-year running mean accumulation and
PDO and reveals little correlation on either long time scales or in the year to year variability. The PDO trends and extreme values do not correspond to the trends in the accumulation record or to unusual accumulation years.

The correlation over the entire time period is -0.047 with a p-value of 0.640 which is effectively zero and not statistically significant over the entire time period. Eleven-year running correlations were calculated for the accumulation-PDO relationship as for the other two oscillations. Figure 6.12 shows the running 11-year correlation from
1905-2005 CE which exhibits a significant periodicity with positive correlation peaks around 1930 and 1980 and negative correlation peaks around 1910, 1960, and 2000. At these peaks, the correlation coefficients approach, or slightly exceed the significance threshold. It is interesting to note that the periods of negative correlation between accumulation and the PDO occur when the PDO is negative and positive correlations occur when the PDO is positive; however, the accumulation-PDO relationship is not stable over this interval.

Figure 6.12 Running 11-year correlation coefficient between annual accumulation from the Bruce Plateau ice core and the PDO.
6.4 Interaction Effects

The preceding analysis evaluated the relationship between accumulation and each of the large scale oscillations. It is very likely, however, that the relationship between accumulation and these large scale circulation patterns is not so simple and that the SAM, SOI, and PDO interact to determine how accumulation on the BP varies. Fogt [2007] identified interactions between the SAM and the SOI that resulted in slightly different storm tracks than if only one oscillation was considered. These were typically small modifications that resulted in tracks that were deflected slightly to the north or south depending on the interaction.

The single oscillation analysis discussed above reveals there is no simple relationship among the three oscillations and accumulation on the BP. The most striking change is that between the SAM and BP accumulation between the late-1940s and the mid-1970s when the relationship becomes negative in contrast to the positive relationship that dominates the rest of the record. Moreover, the strongest correlation between BP accumulation and the SOI occurs during this same period. Figure 6.13 shows the correlation coefficients for BP accumulation and these two oscillations to highlight the synchronicity of the changes in correlation.
The timing of the two transitions in both correlation regimes coincides with the long term negative phase of the PDO from the late-1940s through the mid-1970s as shown in Figure 6.14. The timing of the shift in correlation between accumulation and the oscillations does not exactly match the transition from positive to negative PDO, but is more closely aligned with the period during which the PDO index falls below -0.5. This could explain why similar behavior in the correlation relationships is not observed for other negative PDO periods (1914-1921 and since 2003) when the PDO is only slightly negative.
The data suggest that the relationship between accumulation and the SAM/SOI differs during periods of negative versus positive PDO conditions. Zhang et al. [1997] showed that the negative phase of the PDO (initially referred to as ENSO-like interdecadal variability) produces La Niña-like teleconnections. In the vicinity of the AP, this would
lead to a southward shift in the storm track bringing moisture into the region. This shift would likely not have been detected by Fogt’s [2007] analysis because there were very few negative PDO years included in the interval analyzed. Thus those storm tracks likely represent neutral to positive PDO conditions but not negative PDO conditions. Assuming a simple southward shift of storm tracks under negative PDO conditions provides a framework for understanding the change in correlation between BP accumulation and the SAM during negative PDO periods.

Fogt [2007] developed enhanced storm tracks for differing SAM conditions (Figure 6.15, panel A) that likely reflect the positive or neutral PDO conditions under which they were developed (1971-2000 CE). The same storm tracks are shifted southward (Figure 6.15, panel B) to represent the dominant storm tracks under negative PDO conditions. The area of enhanced precipitation under positive PDO and positive SAM conditions lies close to the BP site while the area of wetter conditions shifts northward under positive PDO and negative SAM conditions resulting in a positive correlation between BP accumulation and the SAM. However, when the storm tracks are shifted southward under negative PDO conditions (Figure 6.15, panel B), the accumulation maximum is near the BP site under negative SAM conditions while it is pushed further south under positive SAM conditions. This set of conditions would result in a negative correlation between BP accumulation and the SAM under negative PDO conditions. This mechanism would appear to account for the observed BP accumulation-SAM relationship under both positive and negative PDO conditions.
6.5 Isotopically Derived Temperature Record

To this point, the analysis of large scale oscillation impacts on the BP has been restricted to the accumulation record and storm track location. Another climate parameter measured in the BP core that may provide information about these large scale oscillations is the oxygen isotopic ratio ($\delta^{18}O$) that serves as a proxy for temperature [Dansgaard, 1964]. Figure 6.16 shows this history, using $\delta^{18}O$ as the proxy, since 1900 CE. The $\delta^{18}O$ and accumulation histories from 1900 CE forward share many similar features and exhibit relatively steady state conditions from 1900 through 1940 CE and both trend
upward thereafter. The $\delta^{18}O$ history differs, however, in that the cool period during the 1970s, inferred from $^{18}O$ depletion, does not correspond to a concomitant reduction in accumulation. The year to year variability in $\delta^{18}O$ exhibits different patterns than those in the accumulation. From 1900 to 2009 CE, the average $\delta^{18}O$ value is -19.62 ‰ with a standard deviation of 0.87 ‰.

Figure 6.16 Annual and 11-year running mean $\delta^{18}O$ history from the Bruce Plateau ice core from 1900 to 2009 CE.
6.6 \( \delta^{18}O \) - Oscillation Relationships

Accumulation and \( \delta^{18}O \) are relatively highly correlated between 1900 and 2009 CE with \( r = 0.583 \) (\( p < 0.0001 \)). There are, however, some differences in the two histories so their relationship was examined using a similar analysis as that performed for accumulation and the three atmospheric oscillations. The similarities and differences between the accumulation-oscillation relationships and the \( \delta^{18}O \)-oscillation relationships are discussed in the following sections.

6.6.1 \( \delta^{18}O \) - Southern Annular Mode Relationship

From 1905 to 2005 CE, the correlation between \( \delta^{18}O \) from the BP core and the SAM is 0.168, weaker than that (\( r = 0.323 \)) between BP accumulation and the SAM. The running 11-year correlations for both accumulation (black) and \( \delta^{18}O \) (red) with the SAM show a similar overall structure (Figure 6.17), but with a few key differences. Between 1920 and 1950 CE, the \( \delta^{18}O \)-SAM correlation is close to zero while that between accumulation and the SAM is close to 0.4. The sharp transition from positive to negative correlation in the late 1940s for accumulation is not present in the \( \delta^{18}O \) correlation but the period of negative correlation is present for both parameters. The subsequent transition back to positive correlation values is not as sharp as for \( \delta^{18}O \) although both parameters show similar correlations with the SAM since 1980 CE.
Figure 6.17 Running 11-year correlation for BP accumulation (black) and BP $\delta^{18}$O (red) with the SAM.

6.6.2 $\delta^{18}$O - Southern Oscillation Index Relationship

From 1905 to 2005 CE, the correlation between BP $\delta^{18}$O and the SOI is 0.210, higher than that ($r = 0.124$) between BP accumulation and the SOI. The correlations for accumulation and $\delta^{18}$O with the SOI appear similar over the majority of the record (Figure 6.18). Two notable exceptions are a short period centered around 1940 where the SOI is more strongly correlated with $\delta^{18}$O than accumulation and the absence of any
δ¹⁸O - SOI correlation from 1950 to 1971 CE when accumulation and SOI were correlated.

![Graph showing correlation coefficients](image)

**Figure 6.18** Running 11-year correlation for BP accumulation (black) and BP δ¹⁸O (red) with the SOI.

### 6.6.3 δ¹⁸O - Pacific Decadal Oscillation Relationship

From 1905 to 2005 CE, the correlation between BP δ¹⁸O and the PDO is -0.038, similar to that (r = -0.047) between BP accumulation and the PDO. The correlations for
accumulation and $\delta^{18}O$ with the PDO (Figure 6.19) are similar over the entire period of the comparison with only minor differences in the short term behavior.

![Figure 6.19 Running 11-year correlation for BP accumulation (black) and BP $\delta^{18}O$ (red) with the PDO.]

6.7 Conclusions from Accumulation and $\delta^{18}O$ Relationships

The differences in the correlations for accumulation and $\delta^{18}O$ with the large scale oscillations provide insight to the linkages between climate on the BP and large scale oscillations. Overall, the strongest relationship is between accumulation and the SAM
and although the $\delta^{18}O$ appears similarly correlated with the SAM, the sharp transitions from positive to negative correlation are not as pronounced. Correlations with the SOI are similar for both parameters except that during a negative PDO the correlation between accumulation and the SOI increases while the correlation between $\delta^{18}O$ and the SOI remains relatively constant. Correlations with the PDO are relatively weak and highly variable. In general, the SAM appears to exert a stronger influence on accumulation than temperature although this is modulated by the phase of the PDO. The SOI appears to influence accumulation and $\delta^{18}O$ equally while phase of the PDO modulates accumulation but not temperature (inferred from $\delta^{18}O$). The PDO does not exert a direct influence on either accumulation of $\delta^{18}O$, but rather influences the relationship between these parameters and the other oscillations evaluated.

If a long term PDO reconstruction were available, the accumulation record reconstructed from the BP may provide a longer term record of SAM variability by virtue of the different relationships between the SAM and accumulation during positive and negative PDO phases. In the absence of this long term PDO record, the change in behavior over time makes any long term SAM reconstruction from the BP data difficult. The most stable relationship over time is between the SOI and $\delta^{18}O$ as it remains relatively strong and constant through both positive and negative PDO phases with only short term deviations from this mean behavior over the length of the record presented here. When the BP core has been fully analyzed, a further analysis of this SOI relationship over a longer time period may provide a proxy for ENSO conditions over the time interval for
which annual resolution is available. This analysis will likely be confined to decadal scale estimates of ENSO variability given the increased dating uncertainty with increasing age in the BP core.

6.8 Evaluation of Other Ice Cores

The relationships between accumulation (and $\delta^{18}$O) and the large scale atmospheric oscillations shown here were developed using only data from the BP core. Data from other AP cores were used to evaluate the plausibility of the mechanism proposed to explain the changes in the relationship between the BP ice core proxy records and the large scale atmospheric oscillations. The two cores closest to the BP site are from James Ross Island (JRI) to the north and the Dyer Plateau (DP) to the south (Figure 2.2).

Annual accumulation data from the JRI core published as decadal averages [Thomas et al., 2008] were provided by Elizabeth Thomas of the British Antarctic Survey. Annual accumulation data from the DP core [Thompson et al., 1994] were readily available as that core was collected and analyzed by the Ohio State University Ice Core Paleoclimatology Research Group.

6.8.1 James Ross Island Core

The annual and 11-year running mean accumulation rates from the JRI core are shown in Figure 6.20. The average accumulation from 1900 to 1997 is 0.613 m w.e. ($\sigma = 0.158$ m w.e.) and there is a generally increasing trend in accumulation with a roughly thirty year oscillation. The JRI accumulation record ends in 1997 when the core was collected.
Both the average and standard deviation in accumulation for JRI core are significantly smaller than for the BP core.

Figure 6.20 Annual and 11-year running mean accumulation from the James Ross Island ice core.

Over the 1905-1997 CE period the correlation between JRI accumulation and the SAM is 0.115 and the correlation between JRI accumulation and the SOI is 0.129. Running 11-year correlations between accumulation and the SAM (Figure 6.21) and the SOI (Figure 6.22) are generally weaker than those with accumulation in the BP core. Additionally,
the rapid shifts between positive and negative correlation are not present for the JRI core although there appears to be some structure to the accumulation-SAM correlation coefficient with stronger positive values in the 1910s, 1930s, and 1990s while, negative correlations dominate from the 1940s through the 1970s. A strong correlation between JRI accumulation and the SOI is found only in the 1940s where the correlation coefficient rises above 0.5 for a brief period.

![Figure 6.21 Running 11-year correlation between James Ross Island accumulation and the SAM.](image-url)
Figure 6.22 Running 11-year correlation between James Ross Island accumulation and the SOI.

6.8.2 Dyer Plateau Core

The annual and 11-year running mean accumulations from the DP core are shown in Figure 6.23. The average accumulation from 1900 to 1989 CE is 0.486 m w.e. ($\sigma = 0.103$ m w.e.). Accumulation from the DP core shows no temporal trend with the possible exception of a slight increase over the last few years of the record which ends in 1989 when the core was collected. The average and standard deviation in accumulation for the DP core are significantly smaller than for the BP core.
From 1905 to 1989 CE the correlations between DP accumulation and the SAM is -0.015 and DP accumulation and the SOI is 0.017. Running 11-year correlations between accumulation and the SAM (Figure 6.24) and the SOI (Figure 6.25) do not show the same structure as those between BP accumulation and the oscillations. There is more variability in the correlation coefficients for DP accumulation than for JRI accumulation. Correlations for both the SAM and the SOI with DP accumulation are negative in the 1940s but otherwise there is little coherence.

Figure 6.23 Annual and 11-year running mean accumulation from the Dyer Plateau ice core.
Figure 6.24 Running 11-year correlation between Dyer Plateau accumulation and the SAM.
Figure 6.25 Running 11-year correlation between Dyer Plateau accumulation and the SOI.

6.9 Conclusions from Multiple Cores

The relationships between accumulation and the large scale oscillations observed in the BP core do not appear to be present in cores located either to the north or south. One possible explanation for this lack of a coherent regional pattern is that the BP core is uniquely positioned to provide a record of the strength of westerly flow in the region and may be more strongly influenced by conditions in the tropical Pacific. Sites located further north or south may be further removed from the influences that large scale oscillations impose on the position and intensity of the SH circumpolar storm track.
Additionally, the extremely high accumulation rate at the BP site may allow for easier identification of the influence of large scale oscillations than is possible at lower accumulation sites. The average annual accumulation at the BP site (1.686 m w.e.) is 2.7 times larger than the accumulation on JRI (0.616 m w.e.) and almost 3.5 times larger than that on the DP (0.486 m w.e.) over an identical time period (1900 to 1989 CE). The year to year variability in BP accumulation is also much larger than on JRI or the DP, on both absolute and relative scales. The standard deviation of accumulation at the BP site (0.426 m w.e.) is 2.8 times that at the JRI site (0.150 m w.e.) and 4.1 times that at the DP site (0.103 m w.e.). On a relative basis, the standard deviation at the BP site is 25 percent of the mean accumulation while it is 24 and 21 percent of the mean at JRI and DP, respectively.

The large interannual variability in accumulation at the BP site makes it an ideal location to identify changes in accumulation due to large scale oscillations. All ice core-derived accumulation records will contain some noise introduced by factors such as sample size and delineation of annual layers as well as noise not related to large scale climatology, such as blowing or drifting of snow. At sites with lower accumulation, attempts at determining precise year to year accumulation rates for comparison with climatological variables may be greatly impacted by even minimal drifting. Such errors are likely to be much smaller, on a relative basis, at sites with very high accumulation rates. For example, a 5 cm error in the determination of accumulation from blowing or drifting
would result in error of 10.3 percent of the average accumulation at DP, 8.1 percent of the average accumulation at JRI, and only 3.0 percent of the average accumulation at BP. The large accumulation rate on the BP minimizes the impact of drifting snow and errors associated with delineation of annual layers and results in a record more strongly reflecting changes in large scale oscillations rather than micro-scale processes or analysis induced uncertainty.
Chapter 7: The Industrial Era, 1750-2009 CE

Human influences on the environment have been present at least since the Industrial Revolution (Vitousek et al., 1997). The ice core record from the Bruce Plateau (BP) provides a record of environmental conditions on the Antarctic Peninsula (AP) over the entire interval of potential human influence and may provide insight on how human activities have impacted the environment as well as when these influences may have started. The early part of this time period is more difficult to interpret as there are no observational data from the region with which to compare the ice core proxy records. Thus, the relationships developed over the recent observational period must be extended back in time and assumed to remain constant over the length of the record. Fortunately, there are a few other ice core records available from the AP which can be used to place the proxy histories obtained from the BP core in a regional perspective.

This chapter examines the annual histories of accumulation, isotopic temperature proxies, chemical species, and dust with the goal of characterizing and understanding the changes in these parameters over the industrial era.
7.1 Accumulation

Annual accumulation layers were determined using variations in methyl sulfonic acid (MSA) measured in the BP core. Years were delineated from summer peak to summer peak which approximates accumulation over a calendar year (Section 3.6). Annual layer thinning was determined using the Dansgaard and Johnsen [1969] model for a flank position with h=0.2*H (Section 4.2). The annual accumulation history from 1750 through 2009 is shown in Figure 7.1 along with the 11-year running average accumulation for this time period. Two distinctly different periods are evident in the accumulation history, one from 1750 through 1940, and a second from 1940 to the present. The early time period is characterized by a decreasing trend in accumulation from 2.5 meters of water equivalent (m w.e.) per year in 1750 to 1.4 m w.e. per year in 1940. The more recent time period exhibits an increasing trend from 1.4 m w.e. in 1940 to more than 2.7 m w.e. per year since 2000. The mean accumulation for the entire time period is 1.98 m w.e. with a standard deviation of 0.55 m w.e. Over this 260-year record there appears to be a 30-40 year cycle in accumulation that is superimposed upon the longer term trends. The high accumulation in 1869 is due to a section of poor quality core with two chip samples representing 70 cm of depth. One of those samples is likely fully within 1869 while the other is likely split between 1868 and 1869. Since no other samples were split between years, both chip samples were assigned to 1869 giving a high accumulation rate for that year and a low accumulation rate for 1868. The accumulation measured over the time period (1750-2009 CE) ranges from 0.82 to 3.75 m w.e. per year.
Figure 7.1 Annual accumulation (black) and 11-year running mean accumulation (red) for the Bruce Plateau ice core from 1750 to 2009 CE.

7.2 Isotopes

Isotopic histories of $\delta^{18}$O, $\delta$D, and deuterium excess are available for the BP core and provide proxy measurements of temperature over the entire length of the ice core record. Figure 7.2 shows both the annual average and 11-year running mean $\delta^{18}$O history for the BP core. This temperature proxy shows relatively constant conditions from 1750 until 1890 followed by a cold period from 1900 through the mid-1970s with warming
thereafter. The isotope-derived temperature proxy shows that the temperature on the BP is as warm as it has been since 1750, but that temperatures in the late 1700s and early 1800s were comparable to contemporary levels. The average $\delta^{18}O$ over this interval is $-19.41\%$ with a standard deviation of $0.80\%$.

![Graph showing $\delta^{18}O$ history from 1750 to 2009 for the Bruce Plateau ice core.](image)

Figure 7.2 Annual (black) and 11-year running average (red) $\delta^{18}O$ history from 1750 to 2009 for the Bruce Plateau ice core.

Figure 7.3 presents similar data for $\delta D$. The structure of the $\delta D$ history is similar to $\delta^{18}O$ over the entire record. The average $\delta D$ is $-147.32\%$ with a standard deviation of $6.37\%$.  

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Again, the recent $\delta$D indicates that the current temperature on the BP is about as warm as it has been since 1750, but that the late 1700s and early 1800s were approximately as warm as current conditions. One difference between the two profiles is that the local minimum observed in the 1930s in the $\delta^{18}$O profile is not as pronounced in the $\delta$D record. The borehole temperature reconstruction from the BP site [Zagorodnov et al., 2012] shows a minimum temperature in the 1930s and thus agrees more closely with $\delta^{18}$O as a temperature proxy than with $\delta$D as a temperature proxy. Using $\delta^{18}$O as a temperature proxy, the periods of minimum temperature are the 1930s and the 1970s. Following the minimum in the 1970s there has been a rapid increase in $\delta^{18}$O indicating rapid regional warming over this time period.
In addition to $\delta^{18}$O and $\delta$D, the isotopic analysis also provides deuterium excess that is a measure of the difference between the two isotope measurements. Deuterium excess can be difficult to interpret physically, but has been suggested to indicate changes in either moisture (Merlivat and Jouzel, 1979; Ciais, et al., 1995) or wind (Petit et al., 1991) conditions at a site. Figure 7.4 shows the annual average and 11-year running mean deuterium excess history for the BP site from 1750-2009 CE. Deuterium excess was relatively constant at around 8 ‰ from 1750 through 1900 CE. Since 1900, the
deuterium excess values have experienced large oscillations that may be indicative of changes in the source region of the moisture transported to the BP. Chapter 5 discussed possible changes in moisture source over the last thirty years as suggested by reanalysis data. These data indicate that conditions since 1900 CE may have been significantly more variable than over the previous 150 years.

Figure 7.4 Annual (black) and 11-year running average (red) deuterium excess history from 1750 through 2009 CE for the Bruce Plateau ice core.
7.3 Chemical Species

A number of cation and anion species were quantified in samples from the BP ice core. These species were measured on individual samples and these individual sample concentrations were then used to determine annual species fluxes. The flux calculations require the sample sizes in w.e. and thus depend on the thinning model applied. Therefore, the histories presented here should be viewed in light of the thinning model discussion presented in Chapter 4. Species measured include sodium, ammonium, potassium, magnesium, calcium, methyl sulfonate, chloride, nitrate, and sulfate. Only the anion species (chloride, sulfate, nitrate, and methyl sulfonate) are discussed in detail here.

7.3.1 Chloride

Chloride measured in remote ice cores can usually be attributed to sea salt aerosols. In the case of the BP core, this is definitely the case as the ice core site is located in close proximity to the Southern Ocean. This proximity results in extremely high concentrations of sea salt ions, including chloride, in the BP core. Chloride is the most abundant ion in sea salt and thus chloride concentrations measured in the BP core are typically higher than for all other ions. Transport of sea salt aerosols to the BP site is highly dependent on wind speed and direction as well as sea ice conditions in the region surrounding the AP. Figure 7.5 shows both the annual and 11-year running mean chloride flux from the BP ice core. The average chloride flux is 715 kg/km² with a standard deviation of 463 kg/km². There is large interannual variability in chloride flux with some years experiencing extremely high fluxes. The flux history shows a
decreasing trend from 1750 through 1950 followed by an increase since 1950. This trend mimics the trend in accumulation and may result in part from the choice of thinning model. Years with large chloride fluxes may be attributed to years with stronger westerly winds which are capable of transporting more sea salt from the Southern Ocean to the BP site, which may also correspond to years with increased accumulation. This same profile is observed for other ions for which sea salt is an important source (sodium and sulfate).

![Graph showing annual and 11-year running mean chloride flux](image)

**Figure 7.5** Annual (black) and 11-year running mean (red) flux of chloride measured in the Bruce Plateau ice core.
7.3.2 Sulfate

Sulfate in ice cores can originate from a variety of sources. In the case of the BP core, the dominant source of sulfate is likely to be sea salt aerosols. Unlike chloride however, there are other sources that may contribute significant amounts of sulfate to the total sulfate burden. Separation of these different sources is discussed in detail in Chapter 8 so this discussion is limited to total sulfate. Figure 7.6 shows the annual and 11-year running mean fluxes of total sulfate from the BP ice core. The sulfate flux history is quite similar to the chloride flux history over the entire period with two significant differences. The first is the large sulfate flux in 1816 relative to the chloride flux. This increase is due to the large eruption of Tambora in 1815 which contributed significant sulfate to the total flux this year. The other major difference is the relative enhancement in sulfate over the last 50 years. This recent enhancement of sulfate is likely due to the increasing importance of anthropogenic influences. This enhancement is evident in Figure 7.6, but is more easily identified when sea salt sulfate (SSS) is removed as discussed in Chapter 8. The average sulfate flux over this time period is 130 kg/km$^2$ with a standard deviation of 70 kg/km$^2$. As with the chloride data, the majority of the variability in total sulfate is due to changes in transport of sea salt to the BP site with some additional variability due to volcanic eruptions and/or anthropogenic influences.
Figure 7.6 Annual (black) and 11-year running mean (red) flux of sulfate measured in the Bruce Plateau ice core.

7.3.3 Nitrate

Figure 7.7 displays the annual and 11-year running average flux of nitrate measured in the BP core. Nitrate shows a slightly different profile than chloride and sulfate with smaller magnitude and less variability. Nitrate is not a significant component of sea salt so the variability in sea salt transport does not impact year to year nitrate fluxes. The average nitrate flux is 25 kg/km² with a standard deviation of 9 kg/km². The average and standard deviation of nitrate fluxes are much smaller than for the sea salt species. There
also appears to be a cycle in nitrate fluxes of about 25 years. The nitrate flux for 2009, at the surface, is extremely high, likely due to contamination of the upper few centimeters of the core. There appears to be an increase in nitrate flux over the last 50 years, perhaps indicating an anthropogenic influence. Individual sample concentrations, however, do not increase over this period and dividing flux by accumulation (see Section 8.5 for a description of this method that was used to analyze the sulfate history) does not show an increase in nitrate since 1950 CE.

Figure 7.7 Annual (black) and 11-year running mean (red) flux of nitrate measured in the Bruce Plateau ice core.
7.3.4 Methyl Sulfonic Acid

MSA is measured in the BP core in its ionic form as methyl sulfonate. MSA is produced via oxidation of dimethyl sulfide that is emitted by oceanic phytoplankton as they grow. MSA exhibits a strong seasonal cycle that corresponds to phytoplankton growth patterns with high concentrations during the austral summer and low concentrations during the austral winter. These seasonal variations were used to delineate years within the BP core. The analysis presented here looks at longer term trends in MSA concentrations by using annual average fluxes as well as 11-year running mean fluxes.

Figure 7.8 shows both the annual and 11-year running mean MSA flux since 1750 CE. The MSA flux history does not show the same profile as many of the other species which mimic the accumulation history. There is some year to year variability, but no long term decreasing trend in MSA since 1750 CE as in the profiles for chloride and sulfate. There is, however, an increase in MSA since the 1970s that results in the highest MSA concentrations for the entire time period over the last twenty years. This increase may be due to increased biological productivity associated with warmer temperatures, or with a shift in source region for MSA production from further afield to closer to shore. The average MSA flux over the 260-year time period is 7.0 kg/km$^2$ with a standard deviation of 2.4 kg/km$^2$. Both the average and standard deviation in MSA flux are the smallest of all of the anions measured indicating less year to year variability in MSA compared to the
other ions. MSA is not associated with sea salt aerosol so the major source of other ions should have no impact on the measured MSA fluxes in the BP core.

![Figure 7.8 Annual (black) and 11-year running mean (red) flux of MSA measured in the Bruce Plateau ice core.](image)

**Figure 7.8** Annual (black) and 11-year running mean (red) flux of MSA measured in the Bruce Plateau ice core.

### 7.4 Dust

Dust analysis of the BP core provides information on the insoluble components present in the ice core. Figure 7.9 shows the dust concentration history from the BP core. The average annual dust concentration is 2094 particles/ml with a standard deviation of 1096.
particles/ml. There is significant year to year variability in dust concentration and a slight increase in dust is observed since 1950 CE. The average dust concentration prior to 1950 CE is 1972 particles/ml while the average since 1950 CE is 2504 particles/ml (a 27 percent increase in dust concentration). Some samples with very high dust concentrations (> 11,000 particles/ml) may be contaminated. When these samples are removed from the calculation the average concentration is 1669 particles/ml prior to 1950 CE and 2332 particles/ml thereafter. This is a 40 percent increase in dust concentration. The recent increase in dust may be attributed to enhanced anthropogenic particulate emissions over the latter half of the 20\textsuperscript{th} century. The upper part of a core which is composed of lower density firn, may be contaminated by core handling and drilling operations in the field. However, the increase in the average background concentrations of insoluble dust in the BP core since 1950 CE is unlikely to result from contamination. The years from 2009 to 1950 are from the upper 140.5 meters of the core while the firn to ice transition in the core is at roughly 76 meters. The average dust concentrations from the firn, 1989-2009 CE, (2583 particles/ml) and post-1950 ice, 1950-1988 CE (2462 particles/ml) are similar with a small (five percent) increase in the firn samples.
Figure 7.9 Annual (black) and 11-year running mean (red) average dust concentration measured in the Bruce Plateau ice core.

7.5 Comparisons to Other Ice Cores

The histories presented to this point give a good picture of how conditions have changed on the BP since 1750 CE but do not show how these changing conditions fit with regional observations of change from nearby locations. Data from other AP ice cores provide the best comparison with the BP data in order to place the BP histories in a larger spatial context.
7.5.1 Accumulation

The accumulation record from the BP core differs in a number of ways from previous records obtained from the AP region. The most striking difference is that there is significantly higher accumulation at the BP site than at other AP sites over this time period. Other sites from the AP show accumulation rates of approximately 0.5 m w.e. per year [Mosley-Thompson et al., 1990; Thompson et al., 1994] with the Gomez site increasing to 1 m w.e. over the last 50 years [Thomas et al., 2008]. The BP accumulation rate is 2.5 times greater than any other accumulation rate measured on the AP. This is likely due to a number of different factors including local influences, position relative to the topographic divide, and location of the drill site relative to the dominant regional storm track. The second difference between the BP accumulation record and others from the region is the decreasing trend in accumulation observed from 1750 through 1940. This phenomenon has not been seen in other ice core records from the region. To illustrate this point, Figure 7.10 shows 11-year running mean accumulation rates from the BP core as well as several other cores from the region. Note that the accumulation scale is different for each core. Since 1940, the trends in the accumulation record from the BP agree well with the Gomez core and show an inverse relationship with the 1998 James Ross Island (JRI) core. The other cores do not exhibit the same magnitude of change in accumulation as seen in the BP core. The discrepancy among the records is most evident over the early part of the record where only the BP core exhibits a decreasing accumulation trend. Of the other cores, only the Dyer and Siple cores provide annual accumulation histories reaching back to 1750 CE, the other records all extend back to at
least the 1860s. This discrepancy among the accumulation records is interesting, and potentially reflects the thinning model selected to describe ice flow in the BP core as discussed in Chapter 4.
Figure 7.10 Eleven-Year running average accumulation rates for Antarctic Peninsula ice cores.
7.5.2 Isotopes

The isotopic history from the BP core shows rapid warming over the last 50 years with current conditions being as warm as, but not significantly warmer than, any time since 1750 CE. Isotopic temperature proxy data are available from five other cores from the AP for comparison. $\delta^{18}$O data are available from the ITASE_0105 (ITASE), Siple Station, Gomez, and Dyer Plateau (DP) cores while $\delta$D data are available from the recently drilled 2008 JRI core. These five cores, when combined with the isotope data from the BP core provide a north-south transect of temperature changes along the AP that can be used to evaluate changes at any one location. Figure 7.11 shows 11-year running mean isotope temperature proxies for these six ice cores, arranged from northernmost at the top to southernmost at the bottom (see locations in Figure 2.2). The isotope ranges in Figure 7.11 cover 2‰ for each core ($\delta^{18}$O equivalent). The six temperature proxies do not provide a coherent picture of regional warming since 1750 CE. There are periods when two or more of the cores show similar trends, but also periods where they show opposite warming/cooling behavior. One such period is the 1970s when only the BP core shows a $\delta^{18}$O minimum while the other cores exhibit local $\delta^{18}$O maxima. One consistent aspect among the temperature proxy records is the general warming trend seen at all sites since 1930-1950 CE with the onset of warming differing between sites.
Figure 7.11 Eleven-year running mean isotope temperature proxies for James Ross Island (JRI, green, $\delta^1$D), Bruce Plateau (BP, red, $\delta^{18}$O), Dyer Plateau (Dyer, black, $\delta^{18}$O), Gomez (Gomez, dark red, $\delta^{18}$O), Siple Station (Siple, blue, $\delta^{18}$O), and ITASE_0105 (ITASE, pink, $\delta^{18}$O) ice cores.
The warming trend at the Siple Station and ITASE sites is less pronounced than at the other four sites. Two factors may contribute to this difference. The Siple Station core was collected in the mid-1980s and no new $\delta^{18}$O data from the site are available. The Siple and ITASE cores are also located much further south than the other four cores at the base of the AP in West Antarctica. The timing of the isotopic minima at Siple (1953), ITASE (1957), and Gomez (1961) are delayed approximately twenty to thirty years compared to the other sites. This could indicate that changes in conditions further to the north did not begin to impact the southern AP/West Antarctica region until two decades after the impacts were first felt at the northern sites. The isotopic minima for each of the other three ice cores are found between 1927 and 1937. The later minimum for the BP core (1972) is slightly lower than the minimum in the 1930s, but the two periods are roughly equivalent in terms of 11-year average isotopic values. Recent warming at DP and JRI has resulted in the warmest isotopically-derived temperatures observed since 1750 CE at both sites: 1981 CE at DP and 1996 CE at JRI. Warming at the BP site has not exceeded the maximum isotopic values observed in the 1780s and 1800s, but recent isotope values are nearly equivalent to these maximum values. In contrast, the Siple Station $\delta^{18}$O history, extending only to 1985 CE, reveals a longer term cooling trend. The ITASE core shows a similar cooling trend prior to 1900 CE and onset of warming in the 1950s with generally stable $\delta^{18}$O since the 1970s.
The ITASE site, furthest to the south, has the most depleted $\delta^{18}$O (coldest temperatures). Moving north, the Siple Station site is next most depleted, but the $\delta^{18}$O from the Gomez site is less depleted than $\delta^{18}$O from DP which is further north. This difference between Gomez and DP indicates that the Gomez site is likely more strongly impacted by maritime flow from the west while the DP site experiences flow from the south along the east side of the AP topographic divide. This is similar to the spatial pattern of temperature observed at surface stations and discussed in Chapter 5. The $\delta^{18}$O from the BP core is less depleted than the cores further south indicating warmer temperatures. $\delta D$ from the JRI core is approximately equal to $\delta D$ from the BP suggesting that temperatures are roughly equivalent at these sites. JRI is further north than the BP but is one the east side of the AP and likely affected by different atmospheric circulation patterns than the BP core.

When viewed together, the three long duration cores from the AP (JRI, BP, and DP) present a coherent picture of regional temperature trends since 1750 CE. A general cooling trend prevailed from 1750-1920 CE at all three sites with the greatest magnitude cooling observed at the BP site. Between 1927 and 1937, all three records show a transition from a cooling trend to a warming trend that continues to the present at all three locations with some variation among sites. Some periods of cooling at the BP site correspond with periods of warming at the other two sites, and vice versa. The most obvious examples of this behavior occur during the 1970s when the BP isotope record shows a minimum while those from the other two sites show maxima. These differences
may reflect the position of the ice core sites relative to the topographic divide of the AP. Both the JRI and DP cores sites lie east of the topographic divide and are likely influenced by different atmospheric circulation patterns than those that dominate at the BP site, which experiences primarily westerly flow. The three cores show warming over the last 70-80 years resulting in record, or near-record, warmth since 1750 CE, confirming that this warming trend is indeed regional. The Gomez core also shows warming over this period with record warmth since 1990 over its shorter record (1857-2005 CE).

These isotope data provide the best indicator of the temperature history at the ice core sites examined here but there are other potential sources of temperature data that can be obtained from ice cores. When conditions on a glacier warm sufficiently the surface may experience melt which can be preserved within the ice core as thin crusts if the melting is brief, or as ice layers if the melting is prolonged. Such melt layers eventually refreeze, usually at night, and are easily identified as they are very clear and devoid of bubbles. Presence of this type of feature is a good indicator that temperatures have exceeded freezing and allowed some ice to melt and then refreeze. Recent work on the new JRI core by Abram et al. [2013] shows a significant increase in melting at this site since the late 1400s with the largest increase over the latter half of the twentieth century. This enhancement of melting serves as further evidence of a rapidly warming regional climate with temperatures likely unprecedented over the last several hundred years.
The BP core, however, shows very little evidence of melting with only three small melt layers identified between 24 and 25 meters depth in the core. These melt layers were dated to the austral summer of 2004-2005 (December 2004). The three melt layers were each approximately 3 mm thick and were spread over two samples. The melt layers were situated just below samples with some of the least depleted isotopic values measured and likely represent surface melting at the BP site. The presence of these layers is an indication of unusual warming as there is no evidence of melt deeper in the core, but as a single event it is not appropriate to draw firm conclusions about increasing melt at the BP site. Moreover, it is expected that with a regional warming there would be significantly more melt at the JRI site which is located further north and at a lower elevation than the BP site.

7.5.3 Chemistry

Chemical analysis data for comparison with the BP core are available from the DP and Siple Station ice cores. Analysis of the previous cores was confined to anions and did not include MSA as a species of interest. Thus comparisons are possible only for chloride, sulfate (total and non-sea salt), and nitrate.

Figure 7.12 shows the 11-year running mean chloride fluxes for the three cores, arranged from north to south. The scale for each core is adjusted to show the variability in species values. For chloride, the difference in the magnitude of the fluxes among the three sites is very large. Cl⁻ fluxes on the BP range between 300 and 1200 kg/km² per year while at
Siple Station Cl⁻ fluxes range between 20 and 40 kg/km² per year and at DP the Cl⁻ flux is very low ranging between 10 and 35 kg/km². Despite the large differences in magnitude, the flux histories do have several features in common. Maxima (e.g., 1870, 1895, 1920, and 1950) and minima (e.g., 1880, 1900, 1930, and 1965) in chloride flux are largely contemporaneous among the three cores. However, the chloride flux on the BP decreases between 1750 and 1850 but not at the other two sites. This decreasing trend in chloride is concomitant with the decreasing trend in accumulation (Figure 7.1). The close proximity of the BP site to the Southern Ocean explains the significantly higher fluxes of chloride, which arrives primarily as sea salt aerosol. The recent increase in chloride flux at the BP is also likely linked to the recent increase in accumulation, but whether the recent increases is a regional signal cannot be determined given the older end dates for those records. There is, however, some indication of an upward trend in chloride flux over the last two decades of data available from the DP.
Figure 7.12 Eleven-year running mean chloride fluxes for the Bruce Plateau (red), Dyer Plateau (black), and Siple Station (blue) ice cores.
Figure 7.13 shows a similar plot for sulfate flux which differs markedly among the cores. The fluxes on the BP are highest (80-200 kg/km$^2$) with those at DP and Siple typically in the range of 10-25 kg/km$^2$. Again, the BP core shows a decreasing trend in sulfate between 1750 and 1850 reflective of the accumulation rate history. The sulfate flux histories from the DP and Siple Station ice cores are very similar over the interval shown here. The large spike in sulfate in the early 1800s originates from two large volcanic eruptions in 1809 and 1815 CE and many of the other sulfate peaks also correspond with volcanic eruptions (Gao et al., 2008, 2009). The sulfate history from the BP core is quite different. For example, the early 1800s volcanic peak is less prominent due to the dominance of sea salt, which also contains some sulfate, on the BP where it can contribute as much as 95 percent of the total sulfate deposited there (Section 8.2). In contrast, the other two sites are not as strongly impacted by sea salt aerosol where SSS typically accounts for only 10-20 percent of the total sulfate. A prominent feature of the BP sulfate history is the significant increase in sulfate over the last thirty years that, unfortunately, is not covered by the two older cores.
Figure 7.13 Eleven-year running mean sulfate fluxes for the Bruce Plateau (red), Dyer Plateau (black), and Siple Station (blue) ice cores.
Figure 7.14 shows flux histories for non-sea salt sulfate (NSSS) from the three ice cores. As might be expected given the small contribution of sea salt aerosols to those sites, the NSSS flux histories for the DP and Siple Station ice cores look very similar to the total sulfate histories (Figure 7.13) with slightly reduced magnitudes. However, the BP NSSS flux history bears little resemblance to its total sulfate history given the dominance of sea salt on the BP. With the removal of the sea salt component the decreasing trend from 1750 to 1850 is minimized, the volcanic peaks in the early 1800s are much more prominent and the three NSSS histories are similar over their period of overlap with concomitant peaks coinciding with known volcanic events. There is still a large increase at BP in NSSS flux over the last thirty years that likely reflects the onset of deposition of anthropogenic sulfate emissions in the AP. This increase in NSSS persists (albeit with a smaller magnitude) when flux is divided by annual accumulation (see Section 8.5 for a discussion of this procedure). Previous Antarctic cores, including the DP and Siple Station cores discussed here, do not contain evidence of anthropogenic sulfate deposition (Dai et al., 1997; Abram et al., 2011). Chapter 8 examines the sulfate history in greater detail and describes a method developed to account for other non-anthropogenic sources of sulfate to determine whether this observed increase can indeed be attributed to anthropogenic influences.
Figure 7.14 Eleven-year running mean non-sea salt sulfate fluxes for the Bruce Plateau (red), Dyer Plateau (black), and Siple Station (blue) ice cores.
Finally, Figure 7.15 displays the flux history of nitrate from the three cores discussed here. The magnitude of the fluxes for nitrate are much more similar among the three cores with only slightly larger fluxes on the BP (15-30 kg/km²) than measured at the other two drill sites (8-16 kg/km²). The decreasing trend in the early part of the record from the BP, suspected to reflect the accumulation rate, is still evident while the other two cores show slowly increasing nitrate fluxes over the entire record. Nitrate in the BP core closely follows the changes in accumulation observed over the entire record. Nitrate is not a significant component of sea salt so the large sea salt influence over the BP should not affect the comparison of nitrate fluxes among core locations. The apparent increase in BP nitrate flux since 1970 CE disappears when the flux is divided by the annual accumulation rate (or when evaluating average nitrate concentrations), likely indicating that this increased nitrate flux is due to increased accumulation rather than an anthropogenic influence.
Figure 7.15 Eleven-year running mean nitrate fluxes for the Bruce Plateau (red), Dyer Plateau (black), and Siple Station (blue) ice cores.
7.6 Dust

Figure 7.16 shows the dust concentrations from the BP and DP cores. The BP dust history shown here has individual samples with dust concentrations greater than 11,000 particles/ml removed. The two sites have similar dust concentrations with the BP core slightly dustier than the DP core. They also show similar recent increases in dust concentrations with the DP increase starting slightly earlier than the BP increase. The peak in dust in the 1920s is present in both cores but other large peaks (e.g. 1780s, 1880s, 1950s) are present in only the BP core. These peaks may be contamination artifacts that were not adequately removed by the screening process.
Figure 7.16 Eleven-year running mean dust concentrations for the Bruce Plateau (red) and Dyer Plateau (black) ice cores.

7.7 Conclusions

The combination of accumulation, isotopic, chemical, and dust histories from the BP core provides an interesting picture of the changing conditions over the Bruce Plateau during the last 260 years. The most dramatic changes have occurred in the last 60 years with rapid increases in accumulation, isotopic temperature proxies, and nitrate, NSSS, and
dust. These changes point to significant changes in circulation patterns as well as the likelihood of increasing anthropogenic influences on the Antarctic Peninsula. Changing atmospheric circulation patterns are likely responsible for the increase in accumulation on the BP as well as the warming trend. Anthropogenic emissions from industrialization appear to have resulted in increases in sulfate and dust concentrations measured in BP ice since 1950 CE. The increase in nitrate over this period appears to be due to increased accumulation rather than an anthropogenic influence. Over the latter half of the 20\textsuperscript{th} century, other ice core records from the AP confirm the wetter and warmer conditions recorded in the BP core. Chemical and dust analyses from these cores are either not available or do not extend far enough into the present to confirm whether the anthropogenic influences observed on the BP are recorded elsewhere. The combination of accumulation, isotopic, and sulfate increases measured in the BP ice core make it clear that, over the last 60 years, anthropogenic influences are now recorded on the AP and suggest that further changes should be expected in the future.
Chapter 8: Sulfate Apportionment

8.1 Introduction

The sulfate measured in ice cores can originate from a variety of sources which may contribute different amounts depending on the location of the core, strength of the source reservoir, and time of year. At the Bruce Plateau (BP) site there is sufficient information to attempt to distinguish the different sulfate sources. The major source of sulfate is almost certainly associated with sea salt aerosol. Sea salt has a relatively constant composition of ionic constituents and the major components (chloride and sodium) have few other atmospheric sources that are relevant for this remote location. If all of the chloride (or sodium) is assumed to come from sea salt, then the sea salt contribution of the other ions present in sea salt can be calculated.

Once the sea salt component is subtracted from the total measured concentration for each ion (chloride, sulfate, sodium, potassium, magnesium, and calcium), the remainder is the non-sea salt component for that ion. Rarely, but in some cases, this remainder can be negative. The meaning of such negative values is uncertain, and it may be that negative
values should either be set to zero or averaged with the values of the adjacent samples to provide a somewhat courser resolution record with reduced noise.

The second major source of sulfate at the BP site is biogenic and originates from the oxidation of dimethyl sulfide (DMS) emitted from oceanic phytoplankton. This DMS can be oxidized by two pathways to either sulfate or methyl sulfonic acid (MSA) and the partitioning between these two end products is a function of temperature [Welch et al., 1993]. Just as sea salt can be assumed to be the only source of chloride and/or sodium, biological productivity can be assumed to be the only source of MSA. Using the measured MSA concentration and a temperature derived from the δ^{18}O data, it is possible to estimate the amount sulfate originating from oxidation of biogenic DMS.

The remaining sulfate is likely either volcanic or anthropogenic in origin. Volcanic sulfate should be highly variable in time and correspond to known volcanic eruptions [Dai et al., 1997]. The timing of volcanic influences may be much easier to determine than the magnitude of the influence. Calculating the sulfate contribution from volcanoes begins with identification of the specific events. Fortunately, most explosive eruptions are documented or have been identified in other ice core sulfate records [Dai et al., 1997]. The greater challenge is to calculate the volcanic fraction of the remaining sulfate. An approach to determine the volcanic contribution to the sulfate load is described and evaluated in the following sections.
The remaining sulfate, after removal of sea salt sulfate (SSS), biogenic sulfate, and volcanic sulfate is very likely of anthropogenic origin. The time series of this anthropogenic sulfate provides a high resolution record of human influence on sulfate deposition in the Antarctic Peninsula (AP) region. The methodology for determining each sulfate fraction and analysis of the resulting time series are presented in detail in the following sections.

8.2 Sea Salt Sulfate

The bulk ionic composition of seawater is well-characterized and largely invariant over time and space. The major ionic components in seawater are chloride (Cl\(^-\)), sodium (Na\(^+\)), sulfate (SO\(_4^{2-}\)), magnesium (Mg\(^{2+}\)), calcium (Ca\(^{2+}\)), potassium (K\(^+\)), and bromine (Br\(^-\)). Table 8.1 displays the weight percent of each of these major sea salt constituents. Concentrations of each of these ions, except for bromine, were measured for each sample from the Bruce Plateau ice core. Assuming a constant composition, and that there are no sources of a constituent other than sea salt, it is possible to use one constituent to calculate the sea salt fraction of each of the other sea salt components. Typically, this sea salt determination has used either sodium or chloride for polar ice cores but other methods have been used as well. For example, Dixon et al. [2005] used a combination of all the sea salt ions to determine which ion is limiting and then assigned the remainder of the other species as the non-sea salt fraction. The constant composition assumption employed in this approach may not be valid if aerosol deposited at the site has been generated from frost flowers [Rankin et al., 2000]. Frost flowers may form in the thin
layer of brine that remains on the surface of fresh sea ice and they are depleted in sulfate and sodium relative to bulk sea water. Once frost flowers are formed they may be weathered by the wind and form aerosols with a different ionic composition than aerosols generated from sea spray. The ratio of sodium to sulfate in aerosol generated from frost flowers will be higher than the ratio in aerosol generated from open water [Rankin et al., 2000]. This fractionation may produce the low (negative) values obtained for non-sea salt sulfate (NSSS) in winter months for regions with significant sea ice in close proximity (coastal sites).

Table 8.1 Ionic Composition of Sea Salt from [Seinfeld and Pandis, 1998]

<table>
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<tr>
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<th>Weight Percent</th>
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</tr>
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</tr>
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</tbody>
</table>

SSS is calculated by multiplying the reference ion (sodium or chloride) concentration by the ratio of sulfate to the reference ion in sea salt (7.68:55.04 for chloride or 7.68:30.61 for sodium). This calculation yields a SSS concentration for each sample analyzed using both sodium and chloride as the reference. Figure 8.1 shows a scatter plot of the SSS calculated using sodium versus SSS calculated using chloride. This plot displays all 4,441 samples between 1750 and 2009 CE except the surface sample which was clearly
contaminated. There appears to be very little scatter in this plot which accommodates the entire range of SSS concentrations ($R^2 = 0.9913$, slope = 1.11, and intercept = -2.97). This high degree of correlation could be misleading as the majority of the data are in the range of 0-200 ppb of SSS. Figure 8.2 shows a similar scatter plot of the 4,236 samples with SSS concentrations less than 200 ppb. Clearly, the correlation remains high ($R^2 = 0.983$, slope = 1.06, and intercept = -0.71), but there are some samples which give higher SSS when chloride is used as the reference ion (seen as scatter below the bulk of the data). These samples have high chloride relative to sodium which could reflect either short duration chloride sources or analytical error in the ion chromatography. In analyzing the distribution of these points there is no obvious pattern and there are never more than two samples in any tube that fall significantly off of the regression line. It is likely that these errors arise from analytical error in determining either sodium or chloride.

Figure 8.3 compares the sea salt component calculated using chloride and sodium for a section of the BP core from 0 to 25 meters depth. The main conclusion drawn from these data is that there is little difference in the amount of SSS calculated using sodium or chloride. The slope of the correlation between SSS by sodium and by chloride is slightly greater than unity which means that using sodium as the reference ion results in a slightly higher sea salt contribution than using chloride (and thus more samples with negative NSSS concentrations).
Figure 8.1 Sea salt sulfate concentration calculated using sodium (y axis) and chloride (x axis) as the reference ion.

$R^2 = 0.9913$
Figure 8.2 Sea salt sulfate concentration calculated using sodium (y axis) and chloride (x axis) as the reference ion for samples with less than 200 ppb of sea salt sulfate.
Figure 8.3 Sea salt sulfate concentration using chloride (black) and sodium (red) as the reference ion for the upper 25 m of the Bruce Plateau ice core.
For this project, chloride has been used as the reference ion to determine the sea salt contribution to the total sulfate measured in each sample. The time series of the annual fluxes of total measured sulfate and SSS from 1750 through 2009 CE in the BP core (Figure 8.4) reveals that the sea salt component dominates the sulfate measured at the BP site. Over the 1750-2009 CE interval, the SSS fraction varies from 0.27 to 1.09 with a median of 0.75. There are two years, 1911 and 1943, for which the SSS flux exceeds the measured sulfate flux. These two years are characterized by multiple samples with extremely high chloride (and other sea salt ion) concentrations and has at least one sample with a chloride concentration exceeding 4000 ppb (the concentration of the highest calibration standard). The high concentrations and potential non-linearity in the relationship between sulfate and chloride at such levels likely results in the calculated SSS exceeding the total measured sulfate for these years.
Figure 8.4 Time series of annual fluxes of total measured sulfate (black) and calculated sea salt sulfate (red) between 1750 and 2009.

8.3 Non-Sea Salt Sulfate

Once SSS has been calculated, NSSS can be estimated by subtracting SSS from the measured sulfate for each sample. In some instances, the NSSS may be negative for a given sample. This happens when the amount of sea salt-derived sulfate that would be associated with the chloride concentration measured in a sample exceeds the sulfate concentration measured in that sample. In most cases where the NSSS component is negative it is only slightly negative. However, there are a few samples with large
negative NSSS concentrations (<-50 ppb). A deeper examination of the data shows that the majority of these samples have very high concentrations of both sodium and chloride (as noted above during 1911 and 1943). There are also some instances where consecutive samples show large positive and large negative NSSS concentrations. To account for these negative NSSS concentrations, an algorithm was developed (Section 8.3.1) to ensure that all NSSS contributions were positive as it is physically unrealistic to have negative source contributions from the other sources of sulfate considered in this work.

8.3.1 Negative Non-Sea Salt Sulfate Processing

Several methods for dealing with negative NSSS concentrations were considered and combined to develop the algorithm for processing the data. The final algorithm used a combination of screening and linear interpolation for samples with negative NSSS concentrations and samples adjacent to them. The first step was to retain all positive NSSS concentrations that were not adjacent to negative NSSS samples. These samples were assumed to be unaffected by the processes leading to the negative concentrations. Next, samples adjacent to a negative sample but with a calculated NSSS concentration of less than 10 ppb were retained. After this, a linear interpolation process was used to assign NSSS concentrations to the remaining samples (i.e., those with negative NSSS and those with an adjacent sample with positive NSSS exceeding 10 ppb). This process accounted for both types of negative values well. For instances when NSSS concentration went slightly below zero, this procedure allowed for interpolation with the
nearby values providing additional guidance. When there were large fluctuations between positive and negative NSSS concentrations in adjacent samples, all of the samples were removed when performing the interpolation. Figure 8.5 shows that most of the sample concentrations are unchanged by the screening process and that the screened data set contains no samples with negative NSSS and no large fluctuations between positive and negative NSSS concentrations in adjacent samples.

Figure 8.5 Depth series of original (black) and screened (red) non-sea salt sulfate for a section of the Bruce Plateau core.
8.3.2 Non-Sea Salt Sulfate History

Following this screening procedure, annual fluxes of NSSS were determined over the entire time period. Figure 8.6 shows the annual NSSS fluxes for the BP core since 1750 CE based on the screened data discussed in Section 8.3.1. This NSSS is likely composed of three components, biogenic sulfate, volcanic sulfate, and anthropogenic sulfate. Using other information from the ice core it is possible to estimate the magnitude of the contributions from each of these sources of sulfate over time.

![Figure 8.6 Annual non-sea salt sulfate flux from 1750 through 2009 CE in the Bruce Plateau ice core.](image-url)
8.4 Biogenic Sulfate

One component of the NSSS shown in Figure 8.6 is sulfate that originates from biological activity in the Southern Ocean. Phytoplankton in the ocean emit DMS as they grow. In the atmosphere, this DMS is oxidized relatively rapidly to either sulfate or MSA. The proportion of DMS which is oxidized to sulfate (the biogenic sulfate component) rather than MSA depends on the temperature in the atmosphere where the reaction takes place. Bates *et al.* [1992] determined a relationship between temperature and the MSA to NSSS ratio. They found the relationship to be:

\[
\frac{MSA}{nssSO_4^2} = -1.5 \times T + 42.2
\]

Equation 8.1

where the MSA/NSSS ratio is measured as a percentage and T is the temperature in degrees Celsius.

The MSA concentration was measured for each sample in the BP ice core so that the only information needed to estimate NSSS from biogenic sources is the air temperature in the region where the reaction is occurring. This temperature is not the temperature recorded in the isotopic record as that proxy record is a function of several factors (e.g., moisture source, evaporation/condensation history, temperature at final condensation) and provides a measure of the temperature at the site of the precipitation (i.e. at 2000 m elevation on the AP). The temperature that determines this relationship is likely the surface air
temperature from the biologically productive region of the Southern Ocean. While the isotopically-derived temperature cannot be used directly in this equation, it may be used to determine an estimate of the appropriate temperature. An analysis of surface air temperatures using NCEP/NCAR reanalysis data shows that the Bellingshausen Sea and Southern Ocean to the west of the AP had temperatures between 0 and -7.5°C from 2000 through 2009 with a strong north-south gradient. The BP site is close to the -2.5°C isotherm that extends to the west of the AP and so this temperature is used as guidance for determining the partitioning between MSA and NSSS through the oxidation of biogenic DMS.

Using the borehole temperature profile from the BP ice core presented in Zagorodnov et al. [2012], a site-specific temperature-isotope relationship was developed for the BP site. The relationship was developed using the recent warm period and the minimum temperature estimated to have occurred around the 1930s as control points. This gives

\[ \delta^{18}O = 0.5 \times T - 11.8 \]  \hspace{1cm} \text{Equation 8.2}

where T is measured in degrees Celsius. Using this relationship, the annual average temperature between 2000 and 2009 is -14.04°C on the BP.
In Chapter 5 the relationship between the temperature on the BP and the temperature at stations on the west coast of the AP at sea level was examined. The temperature offset between the BP and the western coastal stations was 11-13°C. Using the derived isotope-temperature relationship (Equation 8.2) gives an 11.5°C offset between the BP temperature and the reanalysis based Southern Ocean surface temperature. These two estimates of temperature variation from sea level to the elevated BP agree well, increasing confidence in their accuracy. By adding 11.5°C to the isotopically-derived temperature, an equivalent temperature for the Southern Ocean can be estimated and then used in Equation 8.1 to determine the partitioning between MSA and sulfate in the oxidation of biogenic DMS. This assumes that there is a constant relationship between the isotopic temperature from the BP and the surface air temperature over the Southern Ocean.

Figure 8.7 shows the annual average biogenic sulfate fluxes calculated using the measured MSA concentrations and the temperature dependent MSA to NSSS ratio described above. There is significant year to year variability in this flux that depends on both the MSA concentration and the isotopically derived temperature proxy for the samples assigned to a given year. Over the entire time period there is a nearly constant biogenic sulfate flux with a moderate increase since the 1970s. The year to year fluctuations are likely indicative of changes in biological productivity in the Southern Ocean near the AP.
Figure 8.7 Annual biogenic sulfate flux for the Bruce Plateau ice core between 1750 and 2009 CE.

The biogenic sulfate can be subtracted from the NSSS on a sample by sample basis to remove the biogenic component from the total. Using this procedure, it is again possible to have negative values for the remainder of the sulfate. These negative values were found to be close to zero and thus all negative values for NSSS minus biogenic sulfate were set to zero. The resulting time series of annual fluxes of non-sea salt/non-biogenic sulfate is shown in Figure 8.8. This time series should be composed primarily of
volcanic and anthropogenic sulfate. Identification of volcanic events is discussed in the next section.

Figure 8.8 Annual fluxes of volcanic and anthropogenic sulfate measured in the Bruce Plateau ice core between 1750 and 2009.

8.5 Volcanic Sulfate

Volcanic eruptions can contribute significant amounts of sulfate to the total sulfate flux measured in ice cores, but the impacts of a volcanic event typically last only one to three years after the eruption. In some cases, identifying volcanic events can be easy (very
large, well dated eruptions like Tambora in 1815 CE) while in other cases it can be challenging to identify periods which are affected by volcanic sulfate. Dai et al. [1997] developed a method to identify volcanic events in the NSSS signal from Antarctic ice cores. This method entails identifying sulfate peaks that are more than two standard deviations above the mean for annual sulfate flux. The method works well when there is a stable long-term mean background sulfate concentration with constant standard deviation. The BP core does not have a constant mean background as evidenced by the increase in the baseline (31-year running mean after removal of volcanic events) since 1980 CE as shown in Figure 8.8.

As a first step for establishing a baseline, years affected by the two large eruptions in 1809 and 1815 were removed from the time series so that those large sulfate peaks do not unduly impact the mean and standard deviation of the remainder of the data. The years removed from the time series are 1809, 1810, 1811, 1816, 1817, and 1818. Removal of these six years reduces the average annual sulfate flux from 19.71 to 18.37 kg/km² and the standard deviation from 14.74 to 10.22 kg/km². In order to account for the increasing baseline, a 31-year running average was calculated for the screened time series. At the start and end of the record, this 31-year average was reduced to a shorter term average such that the 2009 baseline is an average of the 16 years between 1994 and 2009 with a similar reduction for the period just after 1750. Figure 8.9 shows the same data presented in Figure 8.8 but with the addition of this baseline. This baseline is still potentially influenced by volcanic events (see the mid-1830s) but the length of the averaging period
and removal of the impacts from the two largest eruptions minimizes the influence. A moving standard deviation was also calculated in the same manner and the mean (baseline) plus 2.2 standard deviations is also shown in Figure 8.9. The 2.2 standard deviation threshold was chosen as it was the maximum standard deviation that resulted in identification of the known volcanic events. This moving standard deviation accounts for the fact that there is increased variability over the recent portion of the record compared to the early section.

Figure 8.9 Annual flux of anthropogenic/volcanic sulfate (black), 31-year running average baseline (thick red), and volcanic event threshold (thin red) for volcanic event identification.
This first pass at volcanic event identification identified 12 years as having volcanic influences (including the 6 which were initially removed). The baseline and standard deviations were then recalculated with these volcanic years removed and the event identification repeated. This second iteration was performed in order to remove the influence of the volcanic events identified during the first round and resulted in identification of an additional 4 years as having a volcanic influence. Figure 8.10 shows the final baseline and standard deviation used to identify volcanic events.
Figure 8.10 Final volcanic event identification showing annual anthropogenic/volcanic sulfate flux (black), baseline sulfate (thick red), and volcanic event threshold (thin red) at baseline + 2.2*stdev.

Table 8.2 shows the 16 years identified as having a volcanic influence as well as the likely source of the volcanic eruption. Bold years in the table were identified as volcanic on the first pass. Many of these years agree with known, well dated volcanic eruptions but there are some that do not correspond to documented volcanic events. A closer examination of the record shows that two of these years, 1968 and 1869, experienced above average accumulation such that their enhanced sulfate flux could represent enhanced wet deposition rather than an increased volcanic sulfate aerosol burden.
Table 8.2 Years identified as volcanic events in the sulfate record of the Bruce Plateau ice core.

<table>
<thead>
<tr>
<th>Year(s)</th>
<th>Eruption</th>
</tr>
</thead>
<tbody>
<tr>
<td>1992</td>
<td>Pinatubo</td>
</tr>
<tr>
<td>1968</td>
<td>?</td>
</tr>
<tr>
<td>1964</td>
<td>Agung</td>
</tr>
<tr>
<td>1884, 1885</td>
<td>Krakatau (Krakatoa)</td>
</tr>
<tr>
<td>1869</td>
<td>?</td>
</tr>
<tr>
<td>1862</td>
<td>Makian</td>
</tr>
<tr>
<td>1835, 1836, 1837</td>
<td>Cosigüína (Cosegüína)</td>
</tr>
<tr>
<td>1816, 1817, 1818</td>
<td>Tambora</td>
</tr>
<tr>
<td>1809, 1810, 1811</td>
<td>Unknown 1809 eruption</td>
</tr>
</tbody>
</table>

The structure of the sulfate baseline also suggests that the accumulation rate may be affecting volcanic event identification. As seen in Figure 8.10, there is a generally flat baseline from 1750 until around 1940 when it increases relatively rapidly to the present. Closer examination reveals that the baseline actually decreases from 20 kg/km$^2$ in 1750 to around 10 kg/km$^2$ in 1940. Physically, this decrease is hard to explain as the baseline is interpreted as the anthropogenic (or background) contribution. There is no reason to suspect that the anthropogenic influence in Antarctica decreased from 1750 through the early part of the 20th century. The magnitude of the recent rise in the baseline is similarly troubling as there has been no indication from other AP ice core records of an anthropogenic influence of this magnitude since the 1940s. These changes do, however, coincide with changes in annual accumulation observed in the BP core.
The accumulation history, shown in Figure 7.1, shows a recent increase in accumulation as well as a minimum around 1940 which was preceded by higher accumulation in the earlier part of the record. Atmospheric sulfur dioxide (SO$_2$) can be deposited on snow surfaces through both dry and wet deposition. The dry deposition rate depends on the atmospheric concentration, surface characteristics, and turbulence in the boundary layer. Dry deposition is relatively slow, but occurs continuously. Wet deposition occurs only when there is precipitation but contributes relatively more to the total flux when it occurs. The weather conditions on the BP are such that there is some snowfall on approximately 90% of all days so wet deposition in this region clearly dominates dry deposition. The rate of wet deposition of SO$_2$ (sulfate) can be increased in one of two ways, increasing its atmospheric concentration or increasing the amount of precipitation.

Thus, the decrease in the baseline sulfate flux from 1750 to 1940 could be the result of a decrease in the atmospheric SO$_2$ concentration, a decrease in the amount of precipitation, or a combination of the two. Similarly, the recent increase in the baseline sulfate flux could result from an increase in SO$_2$ concentration in the atmosphere, an increase in precipitation, or both. One way to remove the influence of the amount of precipitation in a given year is to divide the sulfate flux by the accumulation rate. This is similar to evaluating the average concentration of sulfate in a given year’s samples while ignoring the number of samples assigned to that year. Using this procedure also removes the influence of ice flow model discussed in Chapter 4 as the water equivalent thickness of each sample is removed from the calculation of sulfate flux. Figure 8.11 shows that this
transformation of the data (from Figure 8.8) removes the decreasing trend in the baseline from 1750 through 1940. However, the increasing baseline since 1940 is still present, albeit at a smaller magnitude, which suggests that the decrease in anthropogenic (background) sulfate flux from 1750 to 1940 was due primarily to a reduction in precipitation while the recent increase in anthropogenic sulfate flux is the combined result of increased precipitation and increased atmospheric SO$_2$ concentration from anthropogenic sulfur emissions. In most ice cores this type of transformation of the data is not necessary as accumulation is generally more constant over time. The large changes in accumulation observed in the BP ice core mean that the fluxes must be evaluated in this way to assess whether they reflect changes in precipitation. The separation of sea salt and biogenic sulfate from the total measured sulfate should not be affected by this accumulation dependence because those separations were performed on a sample by sample basis and were not aggregated into annual averages prior to their calculation.
A similar analysis to that performed on the sulfate flux data was performed on the flux/accumulation data to assess how the identification of volcanic peaks was affected by this change. In the first step of this analysis, only 5 years (1810, 1811, 1816, 1817, and 1818) were removed from the baseline and standard deviation calculation and 12 years were identified as having volcanic influences. The second pass resulted in the identification of 2 additional years as having volcanic impacts. Table 8.3 provides a list of the years identified as containing volcanic signatures using the flux divided by
accumulation method of event identification. This method results in only one year, 1770, identified as having a volcanic influence for which there is no known large volcanic eruption that is likely to impact Antarctica. This year may be affected by a small unknown, possibly local, eruption or may be identified as volcanic in error. The timing of the MSA peak, used to determine dating, and the NSSS peaks is not always perfect and here the assigned split between years results in assignment of more of the NSSS to 1770 than 1769. It is possible that this slight mismatch in sulfate partitioning is responsible for the identification of 1770 as a volcanic event. Figure 8.12 shows the baseline and threshold for event identification using this method.

Table 8.3 Years identified as volcanic events in the sulfate record of the Bruce Plateau ice core using flux divided by accumulation.

<table>
<thead>
<tr>
<th>Year(s)</th>
<th>Eruption</th>
</tr>
</thead>
<tbody>
<tr>
<td>1991, 1992</td>
<td>Pinatubo, Cerro Hudson</td>
</tr>
<tr>
<td>1964</td>
<td>Agung</td>
</tr>
<tr>
<td>1884, 1885</td>
<td>Krakatau</td>
</tr>
<tr>
<td>1836, 1837</td>
<td>Cosigüina</td>
</tr>
<tr>
<td>1816, 1817, 1818</td>
<td>Tambora</td>
</tr>
<tr>
<td>1810, 1811</td>
<td>Unknown 1809 eruption</td>
</tr>
<tr>
<td>1770</td>
<td>?</td>
</tr>
<tr>
<td>1761</td>
<td>Makian</td>
</tr>
</tbody>
</table>
Figure 8.12 Final volcanic event identification showing annual anthropogenic/volcanic sulfate flux divided by annual accumulation (black), baseline sulfate (thick red), and volcanic event threshold (thin red) at baseline + 2.2*stdev.

Sulfate fluxes for years identified as containing volcanic events were removed and the baseline was interpolated for those years. This baseline was then subtracted from the anthropogenic/volcanic sulfate value to determine the magnitude of the volcanic influence for a given year. Figure 8.13 shows the magnitude of the identified volcanic eruptions (using flux divided by accumulation) over the period from 1750 to 2009. The Tambora eruption in 1815 is the largest magnitude eruption over this time period with
impacts spread over three years. The Unknown 1809 eruption shows impacts in two years with approximately half the magnitude of Tambora. The eruptions of Cosigüina, Krakatau, Agung, and Pinatubo/Cerro Hudson show similar magnitudes while the eruption of Makian and the potential 1770 eruption are of the smallest magnitude. The relative magnitudes of these eruptions using the flux/accumulation method further supports this as the more appropriate method for identifying volcanic events. The Tambora eruption should be about twice the magnitude of the 1809 eruption (Gao et al., 2008, 2009). When volcanic events are identified using only flux measurements, the 1809 eruption is slightly larger than the Tambora eruption. When flux/accumulation is used to identify volcanic events, however, the magnitude of the Tambora eruption is approximately twice that of the 1809 eruption, as would be expected.
Figure 8.13 Magnitude of volcanic influences on the sulfate record from the Bruce Plateau ice core calculated as annual flux divided by annual accumulation.

8.6 Anthropogenic Sulfate

With the volcanic component removed, the remaining sulfate is likely anthropogenic in origin. Figure 8.14 shows the anthropogenic sulfate history for the BP ice core since 1750 CE. There is a constant background from 1750 until 1940 with only minor fluctuations in the baseline which may reflect the incomplete removal of large volcanic events from the anthropogenic sulfate component. Since 1940 the increasing trend in anthropogenic sulfate suggests that human activity affected the Antarctica atmosphere for
the last 70 years. The increase is not large, but it is significant that this is the first observation of anthropogenic sulfate in ice cores from the AP. There is a minimum in anthropogenic sulfate during the 1970s that corresponds with the accumulation and temperature minima observed over the same time period. The anthropogenic sulfate influence peaked in the early 2000s and has been decreasing since that time but the short duration of the record since then makes it difficult to determine if this is a robust trend or merely a product of short term noise in the data.

Figure 8.14 Annual (black) and 11-year running mean (red) anthropogenic sulfate history from the Bruce Plateau ice core.
The data presented in Figure 8.14 are independent of the ice flow model chosen as the flux of sulfate has been divided by the annual accumulation rate to produce this plot. It is also useful, however, to look at the actual sulfate flux data using the selected ice flow model. Figure 8.15 displays the anthropogenic sulfate flux history from the BP core since 1750 CE. While the slow decline in anthropogenic sulfate flux between 1750 and 1940 CE suggests that the thinning model may influence this history, the rapid increase since 1940 is very pronounced. There is an initial rise in anthropogenic sulfate in the 1950s and 1960s followed by a slight decrease in the 1970s and then a large, rapid increase in anthropogenic sulfate through the late 1990s, more than tripling the anthropogenic sulfate flux measured in the 1940s. While some of this increase may be due to the concomitant increase in accumulation, consideration of the data in Figure 8.14 and 8.15 confirms an increase in atmospheric sulfate concentration due to anthropogenic activity since the 1940s.
Figure 8.15 Anthropogenic sulfate flux history from the Bruce Plateau ice core.

8.7 Conclusions

Sulfate in ice cores can originate from a number of diverse sources. Understanding the magnitude of the contributions from these sources can lead to a better understanding of the environment from which the ice core was collected. In many ice cores, contributions from sea salt and biogenic sources are either relatively constant, minor, or both. When this is the case, volcanic and anthropogenic sulfate contributions can be estimated relatively easily. In the case of the BP core, however, the sea-salt sulfate component is
both large and highly variable and the biogenic sulfate component experiences a dramatic 
annual cycle and is SST dependent. These two factors make identification of trends in 
the smaller and/or more sporadic sources much more challenging. The method 
developed here allows the removal of the sea salt and biogenic components, as well as 
identification of the volcanic component to estimate an anthropogenic sulfate history 
from the BP ice core.

The resulting estimate of anthropogenic sulfate deposition on the BP strongly suggests 
that anthropogenic emissions have had an impact on Antarctic sulfate concentrations 
since the 1950s, with a marked increase since the 1970s. This result contrasts with 
results from previous ice cores in the region that have not contained evidence of 
anthropogenic sulfate influence. The combination of high accumulation, resulting in 
large fluxes of sulfate, and a method to account for a highly variable biogenic sulfate 
signal at this site facilitated the development of this new anthropogenic sulfate history. 
An evaluation of the anthropogenic sulfate flux as well as the flux divided by the 
accumulation rate shows that this recent increase is not due solely to an increase in 
accumulation at the site, but to an increase in the atmospheric sulfur concentration over 
the AP.
Chapter 9: Conclusions and Future Work

The Bruce Plateau (BP) ice core provides a unique record of climate variability on the western Antarctic Peninsula (AP) over the last 260 years. Measurements of insoluble dust, major ion chemistry, and isotopic composition are used to determine annual accumulation rates, species fluxes, and a proxy for temperature. When combined with meteorological station observations, reanalysis data, large scale atmospheric circulation indices, and histories from other ice cores, the new data provide significant insight to the regional climate variability on the western AP since 1750 CE.

Seasonal variations in methyl sulfonic acid (MSA) provide a robust time scale for the BP core with high MSA concentrations corresponding to the biologically productive summer season in the Bellingshausen Sea and low MSA concentrations corresponding with the winter season. These annual layers can be counted to determine the age of ice to a depth of 400 meters (~1400 CE) after which the annual layers can no longer be clearly and unambiguously discerned. This time scale is very robust over the time covered by this work and is confirmed by both volcanic eruption signals and radioactivity horizons from atmospheric nuclear weapons testing. The layer thickness data are combined with the density profile and an ice flow model to reconstruct annual accumulation rates.
The average annual accumulation rate on the BP from 1750 to 2009 CE is 1.98 m w.e. which is more than twice the accumulation recorded in other ice cores collected on the AP. Comparison of meteorological data collected at stations in the AP as well as from the BP drill site with the ice core-derived proxy data demonstrates that conditions at the BP site most strongly reflect those on the west coast of the AP. The strongest correlation is with Rothera Station on the western side of the AP. This indicates that the BP is situated in a different climate zone than the other AP ice cores that are more likely to record conditions on the eastern side of the AP. The high BP accumulation rate also suggests that the site, while slightly east of the topographic divide, experiences predominantly westerly flow with storms from the Bellingshausen Sea bringing large amounts of precipitation to the site. On shorter time scales (days to weeks), the site may be influenced also by weather conditions to the south or east. Since the 1970s, there have been shifts in the regions for which δ¹⁸O from the BP and surface temperatures are most strongly correlated. Over this interval the correlation with conditions to the east of the AP has increased significantly for annual averages as well as during spring and summer seasons. These shifts in correlation have been accompanied by rapid increases in accumulation as well as by enrichment of the δ¹⁸O temperature proxy.

Accumulation on the BP is strongly and positively correlated with the strength of the Southern Annular Mode (SAM) since the mid-1970s. This positive accumulation-SAM correlation agrees with the predicted relationship based on increasing the strength of
westerly winds and the dependence of the Southern Hemisphere storm track on the SAM. However, this relationship is not temporally stable over the 20th century. Between the late-1940s and early-1970s, BP accumulation was negatively correlated with the SAM with positive correlations both prior to the mid-1940s and after the 1970s. These three periods are marked by rapid shifts in correlation from positive to negative and back. The period of negative correlation is characterized by long term negative Pacific Decadal Oscillation (PDO) conditions. A potential explanation is that the negative phase of the PDO results in a teleconnection that shifts the dominant storm track southward producing a negative accumulation-SAM relationship. The same behavior is not evident in the relationship between BP δ18O and the SAM or between accumulation from other AP ice cores and the SAM. The lack of relationship between the SAM and other ice cores from the AP may reflect the fact that other cores are not as strongly influenced by conditions to the west of the AP. Additionally, the strength of the relationship between the Southern Oscillation Index and accumulation increases during the same interval (late-1940s to early-1970s). These temporal changes in the accumulation-oscillation relationships indicate that they are not simple and are modulated by other factors such as the PDO.

Since 1750 CE there have been significant variations in accumulation rate and the temperature proxy, δ18O, which share several common characteristics. Both have minima in the mid-20th century followed by rapid increases over the last 50 years. The recent increases in accumulation and δ18O result in values that are either at their highest level since 1750 CE (accumulation) or nearly at their highest level (δ18O). Prior to 1930 they
differ with accumulation decreasing from 1750 through 1930 while the decrease in $\delta^{18}O$
does not show the same steady long term declining trend. The trend in accumulation
prior to 1930 may be influenced by the ice flow model that governs thinning of annual
layers with depth in the BP ice core. Since 1930 the choice of thinning model has little
impact on the accumulation history. The $\delta^{18}O$ history is not affected by the choice of
thinning model over any depth interval. The accumulation history from the BP core
differs from other AP cores from 1750 to 1930 CE with the other cores not showing a
decline in accumulation over this period. This is likely due to the different climate
regimes on the eastern and western sides of the AP but may also be influenced by the ice
flow model used to determine accumulation rates in the BP core. Many of the chemical
and dust species histories mimic the accumulation history with minima in the mid-20th
century and rapid increases thereafter. The rapid increases in dust and sulfate over the
latter half of the 20th century indicate that anthropogenic sources of these species have
recently become important contributors to the total flux to the AP.

Sulfate in ice cores can originate from sea salt, biological productivity, volcanic
eruptions, and anthropogenic emissions. To determine the anthropogenic component it is
necessary to estimate the contribution from each of the other sources. In the BP ice core
SSS is by far the largest source of sulfate but it is also quite variable from sample to
sample. The sea salt fraction is estimated by assuming that all chloride is from sea salt
and that there is a constant ratio of chloride to sulfate in sea salt. The amount of sulfate
associated with sea salt is then subtracted from the measured sulfate to determine the
NSSS. The biogenic sulfate component is then estimated using the measured concentration of MSA, a temperature estimated from δ\textsuperscript{18}O, the borehole temperature profile, a temperature offset between the surface of the Bellingshausen Sea and the BP, and the temperature dependent oxidation of dimethyl sulfide to either MSA or sulfate. Volcanic events are then identified using an event identification algorithm to separate the anthropogenic sulfate component. This procedure results in an anthropogenic sulfate history for the BP that shows a large increase in anthropogenic sulfate since 1950. The increase in anthropogenic sulfate is larger than that expected solely from increased wet deposition due to an increase in accumulation. This increase in anthropogenic sulfate deposited in the AP is important as it has not previously been measured in Antarctic ice cores and indicates that human emissions, primarily from lower latitudes, are beginning to have an impact on Antarctica.

When the BP ice core is fully analyzed there will be histories of accumulation and δ\textsuperscript{18}O that extend back over at least the last 2000 years and likely much longer. Preliminary analysis of the bottom section of the core suggests that there may be glacial stage ice present in the bottom few meters. Regardless, these records will provide the longer temporal context for assessing whether the current rate of increase and elevated accumulation and δ\textsuperscript{18}O are unique over at least the last two millennia.

Establishing an age estimate for the bottom few meters will allow a much longer, but low-resolution climate history from the BP core and may provide additional opportunities
for interpretation. The separation of sulfate sources will also be continued over the period prior to 1750 CE for which the core can be annually dated. Additional future work will revolve around continued analysis of the upper sections of the core at high time resolution (~weekly) building on work conducted by Daniel Miller (B.S., 2013, OSU Atmospheric Science Program). The ice core data from the BP core will also be combined with other proxy histories available from LARISSA collaborators. These interdisciplinary comparisons will attempt to link the atmospheric record preserved in the BP core with biological activity in the surrounding oceans, records of environmental conditions preserved in nearby ocean cores, and glaciological conditions on the AP over the last several thousand years.
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