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State of the Antarctic and Southern Ocean Climate System

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STATE OF THE ANTARCTIC AND SOUTHERN OCEAN CLIMATE SYSTEM

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[1] This paper reviews developments in our understanding of the state of the Antarctic and Southern Ocean climate and its relation to the global climate system over the last few millennia. Climate over this and earlier periods has not been stable, as evidenced by the occurrence of abrupt changes in atmospheric circulation and temperature recorded in Antarctic ice core proxies for past climate. Two of the most prominent abrupt climate change events are characterized by intensification of the circumpolar westerlies (also known as the Southern Annular Mode) between ~6000 and 5000 years ago and since 1200–1000 years ago. Following the last of these is a period of major trans-Antarctic reorganization of atmospheric circulation and temperature between A.D. 1700 and 1850. The two earlier Antarctic abrupt climate change events appear linked to but predate by several centuries even more abrupt climate change in the North Atlantic, and the end of the more recent event is coincident with reorganization of atmospheric circulation in the North Pacific. Improved understanding of such events and of the associations between abrupt climate change events recorded in both hemispheres is critical to predicting the impact and timing of future abrupt climate change events potentially forced by anthropogenic changes in greenhouse gases and aerosols. Special attention is given to the climate of the past 200 years, which was recorded by a network of recently available shallow firn cores, and to that of the past 50 years, which was monitored by the continuous instrumental record. Significant regional climate changes have taken place in the Antarctic during the past 50 years. Atmospheric temperatures have increased markedly over the Antarctic Peninsula, linked to nearby ocean warming and intensification of the circumpolar westerlies. Glaciers are retreating on the peninsula, in Patagonia, on the sub-Antarctic islands, and in West Antarctica adjacent to the peninsula. The penetration of marine air masses has become more pronounced over parts of West Antarctica. Above the surface, the Antarctic troposphere has warmed during winter while the stratosphere has cooled year-round. The upper kilometer of the circumpolar Southern Ocean has warmed, Antarctic Bottom Water across a wide sector off East Antarctica has freshened, and the densest bottom water in the Weddell Sea has warmed. In contrast to these regional climate changes, over most of Antarctica, near-surface temperature and snowfall have not increased significantly during at least the past 50 years, and proxy data suggest that the atmospheric circulation over the interior has remained in a similar state for at least the past 200 years. Furthermore, the total sea ice cover around Antarctica has exhibited no significant overall change since reliable satellite monitoring began in the late 1970s, despite large but compensating regional changes. The inhomogeneity of Antarctic climate in space and time implies that recent Antarctic climate changes are due on the one hand to a combination of strong multidecadal variability and anthropogenic effects and, as demonstrated by the paleoclimate record, on the other hand to multidecadal to millennial scale and longer natural variability forced through changes in orbital insolation,

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greenhouse gases, solar variability, ice dynamics, and aerosols. Model projections suggest that over the 21st century the Antarctic interior will warm by 3.4° ± 1°C, and sea ice extent will decrease by ~30%. Ice sheet models are not yet adequate enough to answer pressing questions about the effect of projected warming on mass balance and sea level. Considering the potentially major impacts of a warming climate on Antarctica, vigorous efforts are needed to better understand all aspects of the highly coupled Antarctic climate system as well as its influence on the Earth's climate and oceans.

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1. PRELUDE TO RECENT CLIMATE

[2] In this paper we review the significant roles that Antarctica and the Southern Ocean play in the global climate system. This review is a contribution to the pan-Antarctic research on Antarctica and the global climate system, carried out under the aegis of the Scientific Committee on Antarctic Research (SCAR), which is an interdisciplinary body of the International Council for Science.

[3] By way of introduction, we show some of the main elements of the geographical and climate-related characteristics of Antarctica and the Southern Ocean in Figures 1 and 2, respectively. The processes occurring in these regions are known to play a significant role in the global climate system. The Southern Ocean is the world's most biologically productive ocean and a significant sink for both heat and CO_2, making it critical to the evolution of past, present, and future climate change. The Southern Ocean is the site for the production of the coldest, densest water that participates in global ocean circulation and so is of critical importance to climate change. The strong westerly winds that blow over the Southern Ocean drive the world's largest and strongest current system, the Antarctic Circumpolar Current (ACC), and are recognized to be the dominant driving force for the global overturning circulation [*Pickard and Emery*, 1990; *Klinck and Nowlin*, 2001].

[4] Today, Antarctica holds 90% of the world's fresh water as ice. Along with its surrounding sea ice, it plays a major role in the radiative forcing of high southern latitudes and is an important driving component for atmospheric circulation. Its unique meteorological and photochemical environment led to the atmosphere over Antarctica experiencing the most significant depletion of stratospheric ozone on the planet, in response to the stratospheric accumulation of man-made chemicals produced largely in the Northern Hemisphere. The ozone hole influences the climate locally and is itself influenced by global warming.

[5] The climate of the Antarctic region is profoundly influenced by its ice sheet, which reaches elevations of over 4000 m. This ice reduces Southern Hemisphere temperatures and stabilizes the cyclone tracks around the continent. The ice sheet is a relatively recent feature geologically, developing as Antarctic climate changed from temperate to polar, and from equable to strongly cyclic, over the last 50 million years (Ma).

[6] Modern climate over the Antarctic and the Southern Ocean results from the interplay of the ice sheet, ocean, sea ice, and atmosphere and their response to past and present climate forcing. Our review assesses the current state of the Antarctic climate, identifying key processes and cycles. One aim is to try and separate signals of human-induced change from variations with natural causes. Another is to identify areas worth special attention in considering possible future research.

[7] To fully understand the operation of this system as the basis for forecasting future change we begin with the development of the Antarctic ice sheet far back in geological time (Figures 3 and 4). In the high CO_2 world of Cretaceous and early Cenozoic times, when atmospheric CO_2 stood at between 1000 and 3000 ppm, global temperatures were 6° or 7°C warmer than at present, gradually peaking around 50 Ma ago with little or no ice on land. Superimposed on this high CO_2 world, deep-sea sediments have provided evidence of the catastrophic release of more than 2000 gigatonnes of carbon into the atmosphere from methane hydrate around 55 Ma ago, raising global temperatures a further ~4°–5°C, though they recovered after ~100,000 years [*Zachos et al.*, 2003, 2005].

[8] The first continental ice sheets formed on Antarctica around 34 Ma ago [*Zachos et al.*, 1992], when global temperature was around 4°C higher than today, as a consequence of a decline in atmospheric CO_2 levels [*DeConto and Pollard*, 2003; *Pagani et al.*, 2005]. The early ice sheets reached the edge of the Antarctic continent, although they were warmer and thinner than today's. They were dynamic, fluctuating on Milankovitch frequencies (20 ka, 41 ka, and 100 ka) in response to variations in the Earth's orbit around the Sun, causing regular variations in climate and sea level [*Naish et al.*, 2001; *Barrett*, 2007]. Recent evidence from ice-rafted debris suggests that glaciers also existed on Greenland at this time [*Eldrett et al.*, 2007].

[9] Further cooling around 14 Ma ago led to the current thicker and cooler configuration of the Antarctic ice sheet [*Flower and Kennett*, 1994], and subsequently, the first ice sheets developed in the Northern Hemisphere on Greenland around 7 Ma ago [*Larsen et al.*, 1994]. The Antarctic ice sheet is now considered to have persisted intact through the early Pliocene warming from 5 to 3 Ma [*Kennett and Hodell*, 1993; *Barrett*, 1996], though temperatures several degrees warmer than today around the Antarctic margin are implied by coastal sediments [*Harwood et al.*, 2000] and offshore cores [*Whitehead et al.*, 2005] and from diatom ooze from this period recently cored from beneath glacial sediments under the McMurdo Ice Shelf [*Naish et al.*, 2007]. Global cooling from around 3 Ma [*Ravelo et al.*, 2004] led to the first ice sheets on North America and NW Europe around 2.5 Ma ago [*Shackleton et al.*, 1984]. These ice sheets enhanced the Earth's climate response to orbital

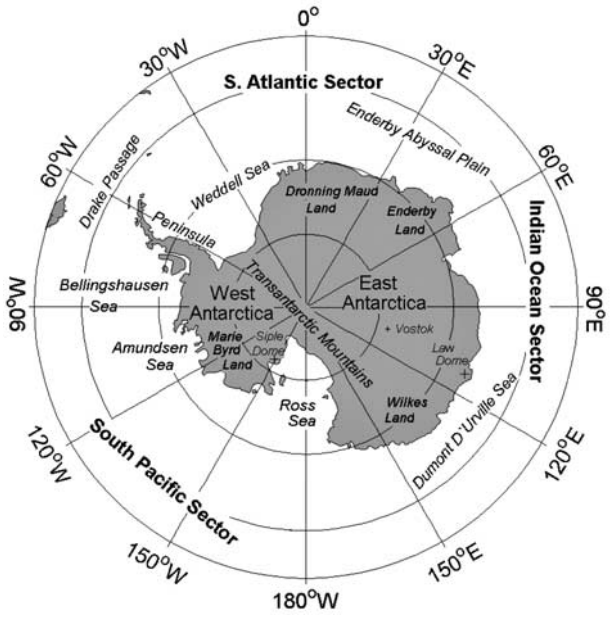

Figure 1. Geographical locations of some key places and regions in Antarctica and the Southern Ocean.

forcing, taking us to the Earth's present ''ice house'' state (Figure 4), which for the last million years (Figure 5) has been alternating over ~100,000 year long cycles including long (~90,000 years) glacials, when much of the Northern Hemisphere was ice covered, global average temperature was around 5°C colder, and sea level was 120 m lower than today, and much shorter warm interglacials like that of the last ~10,000 years, with sea levels near or slightly above those of the present.

[10] Changes in the atmospheric gases CO_2, CH_4, and N_2O and temperature through the past 650,000 years [*Siegenthaler et al.*, 2005; *Spahni et al.*, 2005] of these glacial/interglacial cycles are recorded with remarkable fidelity [e.g., *Etheridge et al.*, 1992] in deep ice cores recovered from Antarctica. The cores reveal both the responsiveness of the ice sheet to changes in orbitally induced insolation patterns and the close association between atmospheric greenhouse gases and temperature [*EPICA Community Members*, 2004]. They also demonstrate the narrow band within which the Earth's climate and its atmospheric gases have oscillated over at least the last 800,000 years through eight glacial cycles, with CO_2 values oscillating between 180 and 300 ppmv. Global average temperatures, assessed through a combination of paleoclimate records, varied through a range of 5°C (between average interglacial values of around 15°C and average glacial values of 10°C [*Severinghaus et al.*, 1998]).

[11] The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report [*Intergovernmental Panel on Climate Change (IPCC)*, 2007] reviewed the ways in which the climate worldwide has changed in response to rising levels of CO_2 in the atmosphere, which, at 380 ppm, are currently higher than at any time in the last 800,000 years [*EPICA Community Members*, 2004] and most likely in the last 25 Ma [*Royer*, 2006].

[12] Knowledge of the phasing of climate events on regional to hemispheric scales is essential to understanding the dynamics of the Earth's climate system. Correlations based on the similarities seen in Greenland and Antarctic ice core methane signals (Figure 6) suggest that climatic events of millennial to multicentennial duration are correlated between the north and south polar regions as described in the following results from *EPICA Community Members* [2006]. Antarctic warm events correlate with but precede Greenland warm events. The start of each warming signal in the Antarctic takes place when Greenland is at its coldest, the period when armadas of icebergs crossed the North Atlantic in so-called Heinrich events. Moreover, warming in the Antarctic is gradual whereas warming in the associated Greenland signal is abrupt. These relationships are interpreted as reflecting connection between the two hemispheres via the ocean's meridional overturning circulation (MOC). The lag reflects the slow speed of the MOC, with complete ocean overturning taking up to 1000 years. The data also show a strong relationship between the magnitude of each warming event in the Antarctic and the duration of the warm period that follows each abrupt warming event in Greenland [*EPICA Community Members*, 2006]. This relationship is interpreted to reflect the extent to which the MOC is reduced, with reduced overturning assumed to lead to the retention of more heat in the Southern Ocean. These associations are indirectly supported by marine sediment records off Portugal that reveal changes in deep water masses related to Antarctic Bottom Water formation and in Atlantic surface water, at the same time as the events seen in the central Greenland deep ice cores [*Shackleton et al.*, 2000]. The cause(s) of these millennial-scale climate events are not fully understood, but slowing of the MOC has been attributed to North Atlantic meltwater flood events and/or to massive iceberg discharges (Heinrich events) that slow the formation of North Atlantic Deep Water. Changes in the Antarctic ice sheet and sea ice extent can also affect Southern Ocean heat retention and ocean circulation [*Stocker and Wright*, 1991; *Knorr and Lohman*, 2003].

[13] As shown in Figures 7 and 8, over the past 12,000 years (Holocene), there have been several abrupt changes in Antarctic climate despite the fact that this period is more stable climatically than the preceding glacial period (Figure 6). These abrupt changes in Holocene Antarctic climate, as well as the abrupt changes in Holocene climate recorded in a global array of paleoclimate records covering the same period, appear to be the product of short-term fluctuations in solar variability, aerosols, and greenhouse gases superimposed on longer-term changes in insolation, greenhouse gases, and ice sheet dynamics [*Mayewski et al.*, 2004a]. There has been sufficient variability during the Holocene to cause major disruptions to ecosystems and civilizations [*Mayewski et al.*, 2004a], demonstrating that this natural variability must be taken into account in understanding modern climate and the potential for future climate change. The relation between the ice sheet and climate is not

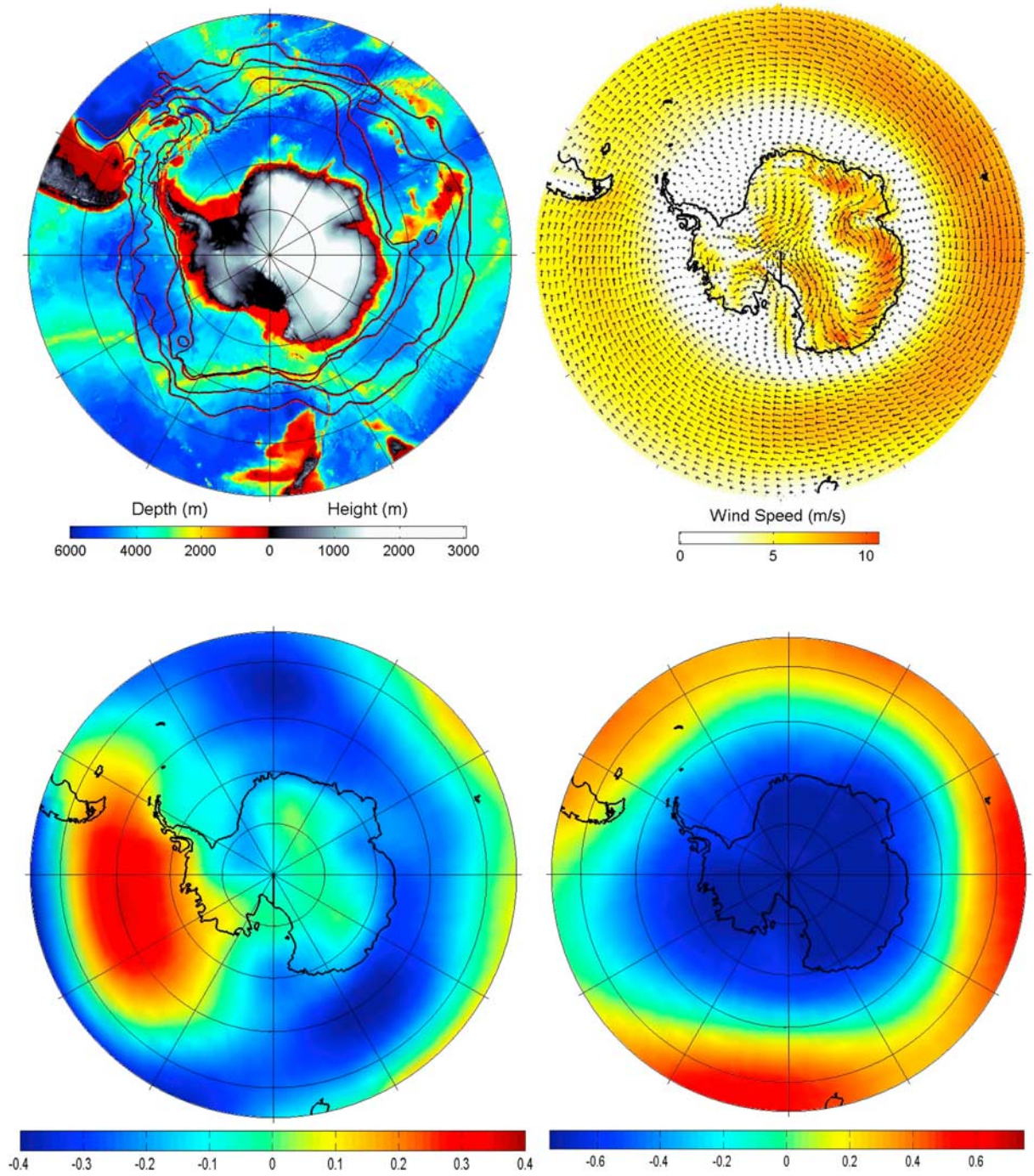

Figure 2a. Some key elements of the Antarctic and Southern Ocean climate system. (top left) The bathymetry and topography of Antarctica and the Southern Ocean, with the main fronts of the Antarctic Circumpolar Current marked over the Southern Ocean; (top right) wind vectors at 10 m height, with wind speed colored as background, where wind vector and speed are long-term means from European Centre for Medium-Range Weather Forecasts (ECMWF) 40-year reanalysis (ERA-40); (bottom left) the loading pattern of the El Niño–Southern Oscillation phenomenon over Antarctica and the Southern Ocean, defined as the correlation of the Southern Oscillation Index with surface atmospheric pressure; and (bottom right) as for Figure 2a (bottom left) but for the Southern Annular Mode index.

simple. For example, the current configuration of the Antarctic ice sheet has as its underpinning a multimillennial-scale lagged response to climate forcing. Grounding lines in the marine-based parts of the West Antarctic ice sheet, at the head of the Ross Ice Shelf, started to retreat to their current position from a position close to the edge of the current Ross Ice Shelf 7000–9000 years ago [*Conway et al.*, 1999]. Synthesis of ice core isotope proxy records for temperature

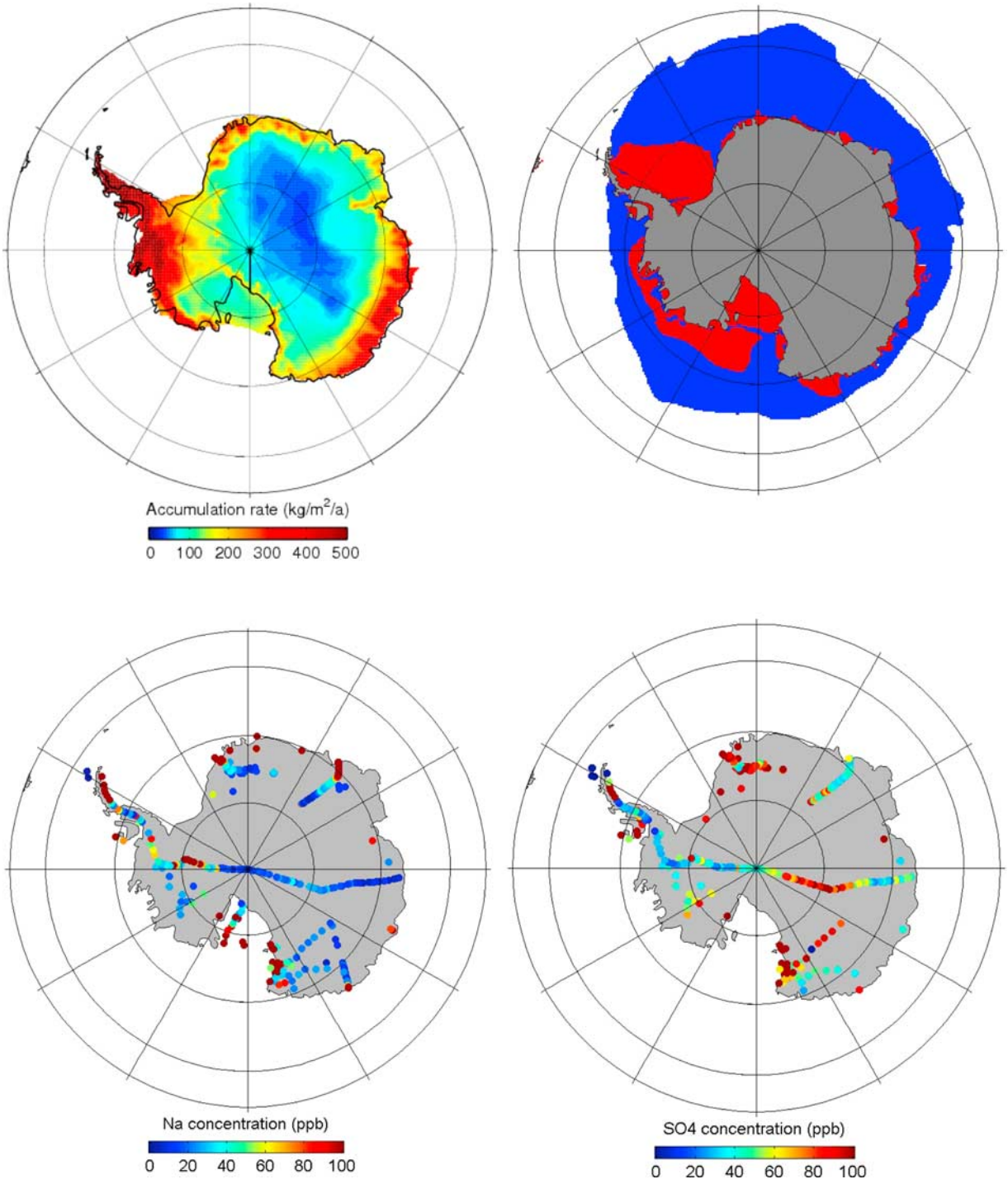

Figure 2b. Some key elements of the Antarctic climate system. (top left) Rate of accumulation of precipitation over Antarctica; (top right) the seasonal maximum and minimum of the sea ice field; (bottom left) spatial variability of sodium concentration compiled from snow samples and firn cores; and (bottom right) as for Figure 2b (bottom left) but for sulphate concentration. (Sodium is predominantly a sea-salt aerosol, while sulphate is often used to reconstruct marine bioactivity and volcanic eruptions.)

reveals that this massive retreat was preceded by an early Holocene climatic optimum between 11,500 and 9000 years ago [*Masson et al.*, 2000].

[14] Comparison of similarly resolved and analyzed ice core records from Greenland (Greenland Ice Sheet Project 2 (GISP2)) and West Antarctica (Siple Dome) reveals evidence related to phasing, magnitude, and possible forcing of changes in atmospheric circulation and temperature over the Holocene (Figures 7 and 8, core locations shown in Figure 1). Atmospheric circulation and temperature reconstructions are based on ice core proxies referenced in the following text and in Figures 7 and 9. These observations are relevant to understanding not only the forcing of climate

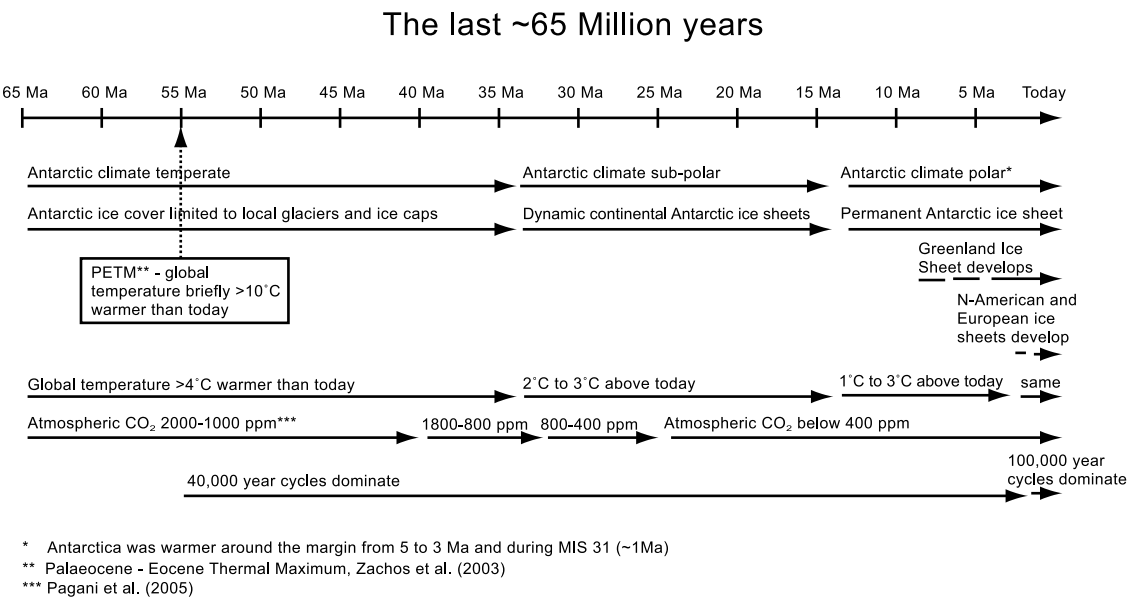

Figure 3. Main climatic events of the last 65 Ma: the Antarctic context.

change over the polar regions but also the implications of change over the polar regions for climate at the global scale.

[15] With respect to changes in atmospheric circulation, Figure 7 shows the following:

[16] 1. The North Atlantic climate record (GISP2 K^+, Na^+, and Ca^{++} proxies for Siberian High, Icelandic Low, and northern circumpolar westerlies, respectively) displays more frequent and larger shifts in atmospheric circulation than does the Antarctic climate record (Siple Dome Na^+ and Ca^{++} proxies for Amundsen Sea Low and southern circumpolar westerlies, respectively) [*Mayewski and Maasch*, 2006]. This finding is similar to that seen between Greenland and Antarctica in millennial-scale events from glacial age ice core records (Figure 6) [*EPICA Community Members*, 2006].

[17] 2. The North Atlantic climate record (GISP2 K^+, Na^+, and Ca^{++} proxies for Siberian High, Icelandic Low, and northern circumpolar westerlies, respectively) generally displays more abrupt onset and decay of multicentennial-scale events than does the Antarctic climate record (Siple Dome Na^+ and Ca^{++} proxies for Amundsen Sea Low and southern circumpolar westerlies, respectively) [*Mayewski and Maasch*, 2006] similar to the glacial age abrupt change record (Figure 6) [*EPICA Community Members*, 2006].

[18] 3. The Siple Dome and GISP2 ice core proxies for northern and southern circumpolar westerlies (Ca^{++}) show considerable similarity in event timing with major intensification periods between ~6000 and 5000 years ago and starting ~1200–600 years ago, although as noted above the Antarctic events start earlier and less abruptly than those in Greenland.

[19] 4. The most dramatic changes in atmospheric circulation during the Holocene noted in the Antarctic are (1) the abrupt weakening of the southern circumpolar westerlies (Siple Dome Ca^{++}) ~5200 years ago and (2) intensification of the westerlies and the deepening of the Amundsen Sea Low (Siple Dome Na^+) starting ~1200–1000 years ago.

[20] With respect to changes in temperature, Figure 7 shows the following:

[21] 1. The prominent temperature decrease ~8200 years ago over the North Atlantic, noted in the GISP2 $\delta^{18}O$ proxy for temperature, is subdued in the Siple Dome $\delta^{18}O$ tem-

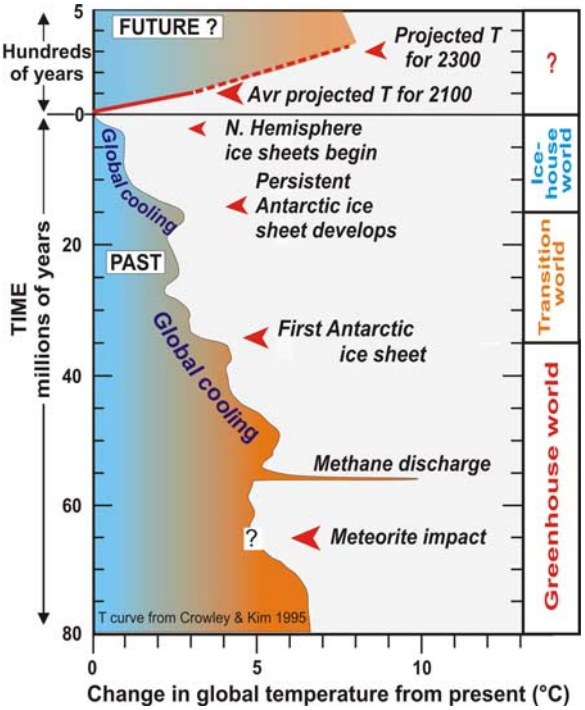

Figure 4. Change in average global temperature over the last 80 Ma, plus future rise in temperature to be expected from energy use projections and showing the Earth warming back into the "greenhouse world" typical of earlier times more than 34 Ma ago [*Barrett*, 2006]. The temperature curve of *Crowley and Kim* [1995] is modified to show the effect of the methane discharge at 55 Ma [*Zachos et al.*, 2003].

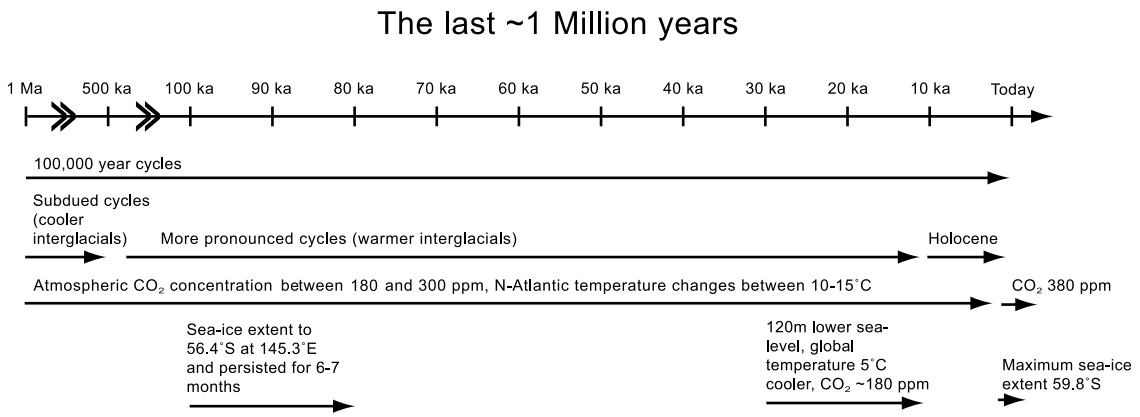

Figure 5. Main climatic events of the last 1 Ma: the Antarctic context.

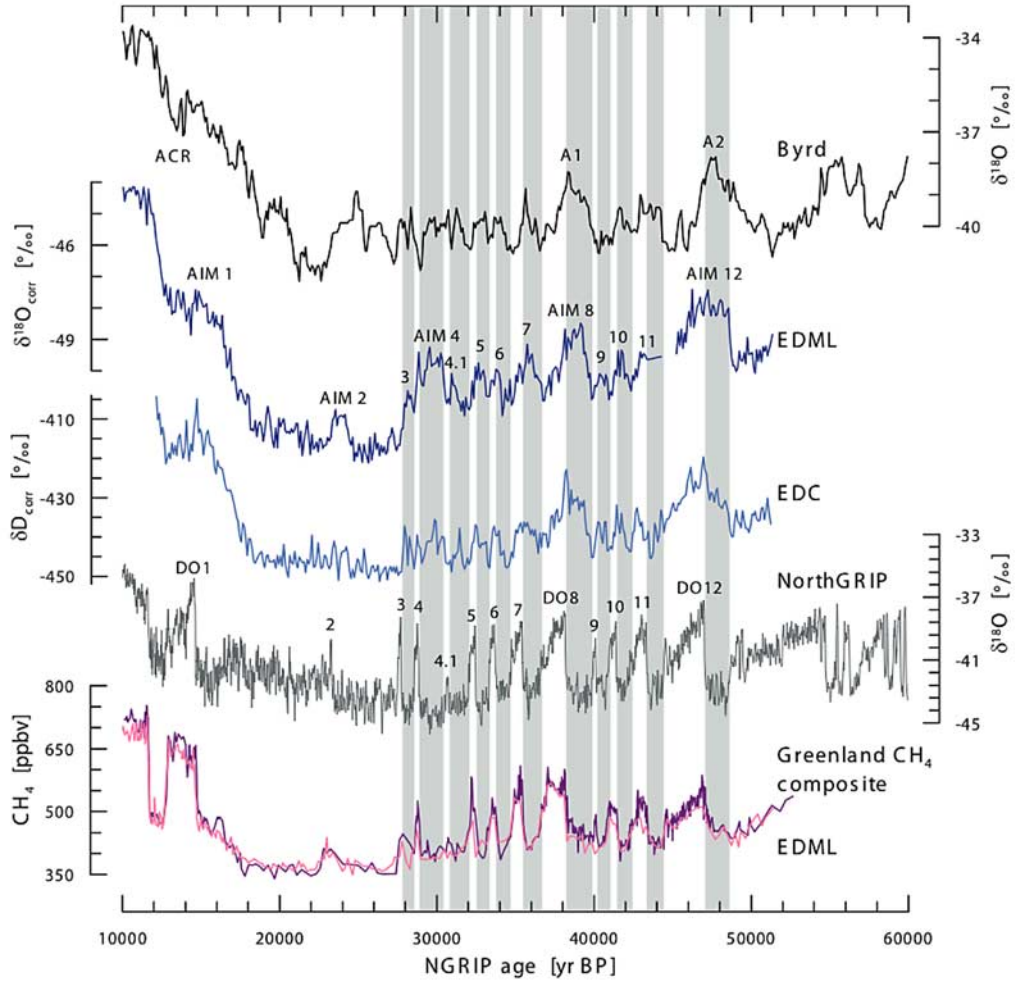

Figure 6. Methane (CH_4) synchronization of the ice core records of $\delta^{18}O$ as a proxy for temperature reveals one-to-one association of Antarctic warming (Antarctic isotope maxima (AIM)) events with corresponding Greenland cold (stadial) events (Dansgaard/Oeschger (DO)) covering the period 10,000–60,000 years ago. EDML, EPICA core from Dronning Maud Land Antarctica; Byrd, core from West Antarctica; EDC, core from East Antarctica; NGRIP, core from north Greenland. Gray bars refer to Greenland stadial periods. Greenland CH_4 composite curve is blue; EDML CH_4 signal is pink. Figure modified from *EPICA Community Members* [2006] by H. Fischer.

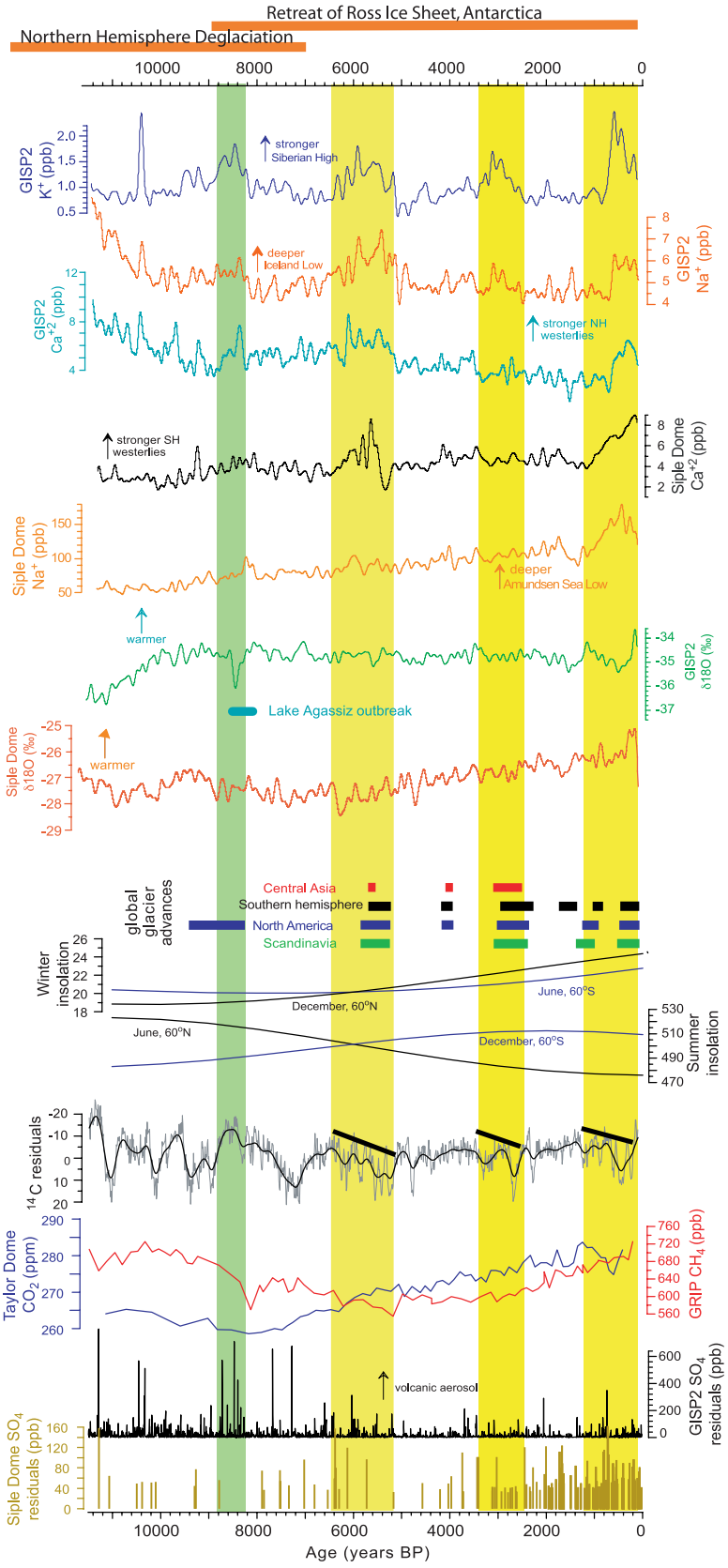
Retreat of Ross Ice Sheet, Antarctica
Northern Hemisphere Deglaciation
10000
8000
6000
4000
2000
0
GISP2 K+ (ppb)
stronger Siberian High
deeper Iceland Low
GISP2 Na+ (ppb)
GISP2 Ca+2 (ppb)
stronger NH westerlies
stronger SH westerlies
Siple Dome Ca+2 (ppb)
Siple Dome Na+ (ppb)
deeper Amundsen Sea Low
warmer
GISP2 δ18O (‰)
Lake Agassiz outbreak
Siple Dome δ18O (‰)
warmer
global glacier advances
Central Asia
Southern hemisphere
North America
Scandinavia
Winter insolation
December, 60°N
June, 60°S
June, 60°N
December, 60°S
Summer insolation
14C residuals
Taylor Dome CO2 (ppm)
GRIP CH4 (ppb)
volcanic aerosol
Siple Dome SO4 residuals (ppb)
GISP2 SO4 residuals (ppb)
Age (years BP)

Figure 7

perature proxy series, although it is suggested in a composite isotope record covering East Antarctica [*Masson et al.*, 2000], indicating that this event is not as prominent in the southern as in the northern polar regions.

[22] 2. Siple Dome $\delta^{18}O$ temperature reconstructions reveal notable cooling between ~6400 and 6200 years ago, followed by relatively milder temperatures over East Antarctica 6000–3000 years ago [*Masson et al.*, 2000], lasting until ~1200 years ago in the Siple Dome area.

[23] 3. The Siple Dome and GISP2 $\delta^{18}O$ proxies for temperature show a flattening and a decline, respectively, in temperature starting ~1200–1000 years ago, followed by warming in the last few decades; the recent warming in the Siple Dome record is the greatest of the last ~10,000 years.

[24] Several interacting factors potentially provide the climate forcing for decadal to centennial-scale and longer Holocene climate change, as demonstrated by their association in timing with climate change events. While further research is needed to go from associations in timing to mechanisms for forcing, identifying these associations is an essential first step. The most prominent associations noted from Figure 7 are the following:

[25] Intensification of atmospheric circulation in the Northern Hemisphere (stronger Siberian High and northern circumpolar circumpolar westerlies and deeper Icelandic Low) and to a lesser degree in the Southern Hemisphere (stronger circumpolar westerlies and deeper Amundsen Sea Low) occurs ~8200–8400 years ago [*Mayewski et al.*, 2004b; *Mayewski and Maasch*, 2006]. This event is associated with cooling over East Antarctica [*Masson et al.*, 2000] and, as seen in Figure 7, with a drop in CH_4, a long-term decline in CO_2, and an increase in solar energy output (based on the ^{14}C proxy for solar variability). Between ~8200 and 7800 years ago, there is a decrease in precipitation in equatorial Africa suggested to have been the consequence of an expanding polar cell and consequent displacement of moisture-bearing winds [*Stager and Mayewski*, 1997]. Intensification of the southern circumpolar westerlies ~6400–5600 years ago is preceded by cooling in West Antarctica ~6400–6200 years ago. Intensification of the Northern Hemisphere westerlies follows abruptly at ~6000–5000 years ago. All of these changes coincide with a reverse in the trend of orbitally forced insolation, a small drop in CH_4, a slight rise in CO_2, and a decrease in solar energy output and are within the period of early collapse of the Ross Sea ice sheet [*Conway et al.*, 1999]. Intensification of the Northern Hemisphere westerlies is coincident with intensification of the Icelandic Low and the Siberian High. Another period of intensification of southern circumpolar westerlies and the Amundsen Sea Low commences ~1200–1000 years ago, accompanied by relatively cooler conditions over East Antarctica [*Masson et al.*, 2000] and West Antarctica (Siple Dome). This change is associated with a decrease in solar energy output, a drop in CO_2, and increased frequency of volcanic source sulphate aerosols over Antarctica. A satisfactory explanation for the forcing of these Holocene Antarctic climate changes remains elusive, though the link to variations in solar energy output is highly suggestive. More detailed examination of forcing over the last 2000 years using the ice cores in Figure 7 and other paleoclimate records supports the close association in timing between changes in atmospheric circulation and solar energy output [*Maasch et al.*, 2005].

[26] The abrupt climate change event commencing ~1200–1000 years ago is the most significant Antarctic climate event of the last ~5000 years [*Mayewski and Maasch*, 2006]. Its onset is characterized by strengthening

Figure 7. Examination of potential controls on and sequence of Antarctic Holocene climate change compared to Greenland climate change using 200 year Gaussian smoothing of data from the following ice cores: Greenland Ice Sheet Project 2 (GISP2) ice core, K^+ proxy for the Siberian High [*Meeker and Mayewski*, 2002]; GISP2 Na^+ proxy for the Icelandic Low [*Meeker and Mayewski*, 2002]; GISP2 Ca^{++} proxy for the Northern Hemisphere westerlies [*Mayewski and Maasch*, 2006]; Siple Dome (West Antarctic) Ca^{++} ice core proxy for the Southern Hemisphere westerlies [*Yan et al.*, 2005]; Siple Dome Na^+ proxy for the Amundsen Sea Low [*Kreutz et al.*, 2000]; GISP2 $\delta^{18}O$ proxy for temperature [*Grootes and Stuiver*, 1997]; Siple Dome $\delta^{18}O$ proxy for temperature [*Mayewski et al.*, 2004a; J. White, unpublished data, 2005]; timing of the Lake Agassiz outbreak that may have initiated Northern Hemisphere cooling at ~8200 years ago [*Barber et al.*, 1999]; global glacier advances [*Denton and Karlén*, 1973; *Haug et al.*, 2001; *Hormes et al.*, 2001]; prominent Northern Hemisphere climate change events (shaded zones [*Mayewski et al.*, 2004a]); winter insolation values ($W\ m^{-2}$) at 60°N (black curve) and 60°S latitude (blue curve) [*Berger and Loutre*, 1991]; summer insolation values ($W\ m^{-2}$) at 60°N (black curve) and 60°S latitude (blue curve) [*Berger and Loutre*, 1991]; proxies for solar output ($\Delta^{14}C$ residuals [*Stuiver et al.*, 1998], raw data (light line) with 200 year Gaussian smoothing (bold line)); atmospheric CH_4 (ppbv) concentrations in the GRIP ice core, Greenland [*Chappellaz et al.*, 1993]; atmospheric CO_2 (ppmv) concentrations in the Taylor Dome, Antarctica, ice core [*Indermühle et al.*, 1999]; and volcanic events marked by SO_4^{2-} residuals (ppb) in the Siple Dome ice core, Antarctica [*Kurbatov et al.*, 2006], and by SO_4^{2-} residuals (ppb) in the GISP2 ice core [*Zielinski et al.*, 1994]. Timing of Northern Hemisphere deglaciation is from *Mayewski et al.* [1981], and retreat of Ross Sea Ice Sheet is from *Conway et al.* [1999]. Figure modified from *Mayewski et al.* [2004a, 2004b]. Green bar denotes the 8800–8200 year ago event seen in many globally distributed records associated with a negative $\Delta^{14}C$ residual [*Mayewski et al.*, 2004a]. Yellow denotes events from 6400 to 5200 years ago, events from 3400 to 2400 years ago, and events since 1200 years ago seen in many globally distributed records associated with positive $\Delta^{14}C$ residuals [*Mayewski et al.*, 2004a]. Map shows location of GISP2, Siple Dome, Icelandic Low, Siberian High, Amundsen Sea Low, Intertropical Convergence Zone, and westerlies in both hemispheres.

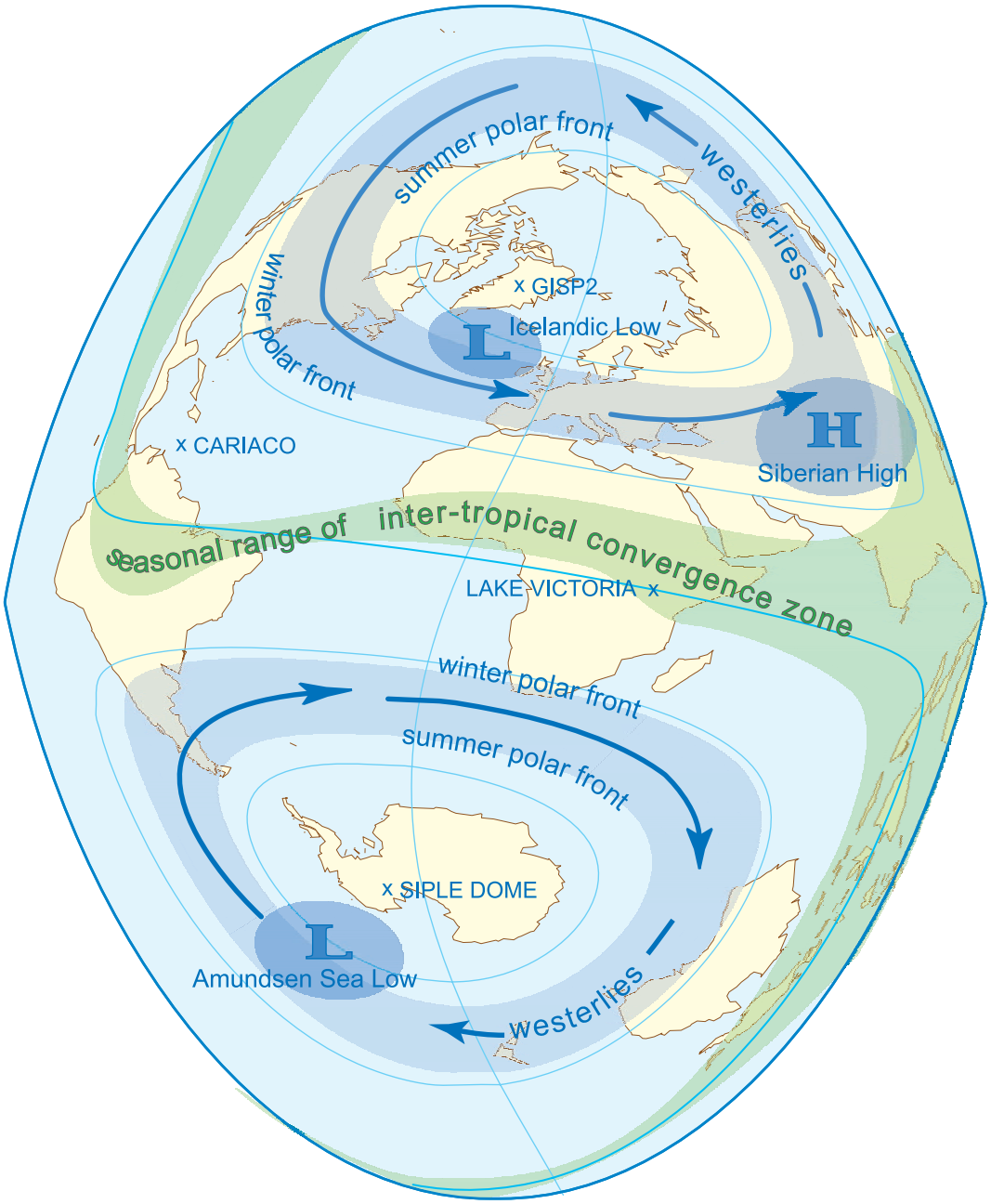

Figure 7. (continued)

of the Amundsen Sea Low (Siple Dome Na^{+}) and the southern circumpolar westerlies (Siple Dome Ca^{++}), with cooling both at Siple Dome ($\delta^{18}O$), until recent decades, and in the East Antarctic composite proxy temperature record [*Masson et al.*, 2000]. This event provides the underpinning for centennial and perhaps shorter-scale natural variability upon which future climate change over Antarctica might operate. A comparison of reconstructions of Northern and Southern Hemisphere temperature [*Mann and Jones*, 2003] and ice core proxies for Northern and Southern Hemisphere atmospheric circulation referred to in Figures 7 and 9 covering the last 2000 years, when these records are most precisely dated (±10 years), demonstrates that major changes in temperature and circulation intensity are associated, such that cooler temperatures coincide with more intense atmospheric circulation and warmer temperatures with milder circulation [*Mayewski and Maasch*, 2006]. Further, until the warming of the last few decades, major changes in temperature were preceded by or coincident with changes in atmospheric circulation. Modern warming is not preceded by or coincident with change in atmospheric circulation, suggesting that recent warming is not operating in accordance with the natural variability of the last 2000 years and that therefore, modern warming is a consequence of nonnatural (anthropogenic) forcing [*Mayewski and Maasch*, 2006].

[27] Comparison of ice core proxies for atmospheric circulation and temperature between West Antarctica (Siple

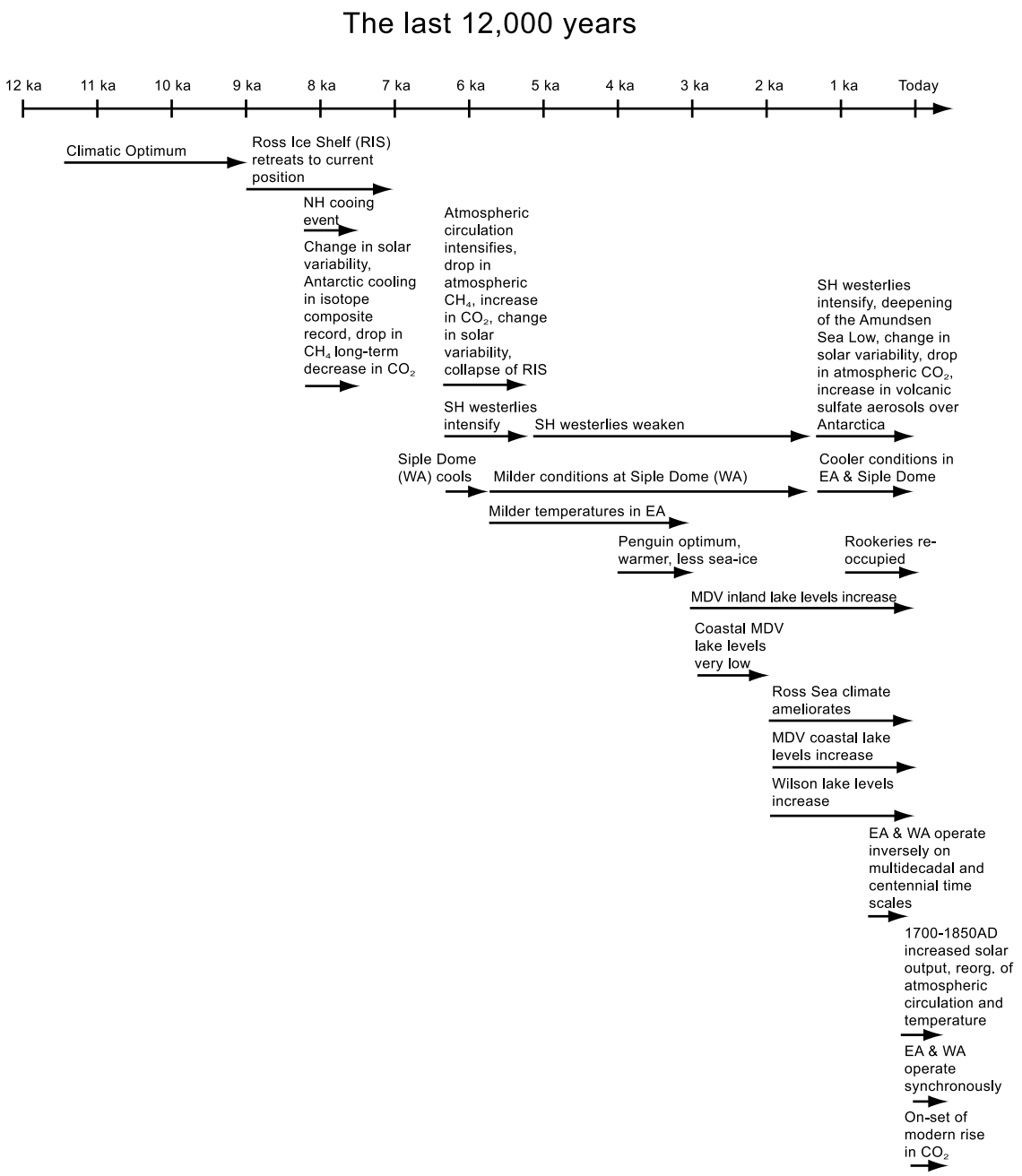

Figure 8. Main climatic events of the last 12,000 years: the Antarctic context. NH, Northern Hemisphere; SH, Southern Hemisphere; WA, West Antarctica; EA, East Antarctica; MDV, McMurdo Dry Valleys.

Dome) and East Antarctica (Law Dome) reveal that East and West Antarctica have operated inversely with respect to temperature and to strength of atmospheric circulation on multidecadal to centennial scales (Figure 9) [*Mayewski et al.*, 2004a]. The exception is a climate change event commencing ~A.D. 1700 and ending by ~A.D. 1850, during which circulation and temperature acted synchronously in both regions. This cooling period is coincident with an increase in the frequency of El Niño events impacting Antarctica as determined from the distribution of methane sulphonate indicative of a supply of methane sulphonic acid (MSA) in a South Pole ice core [*Meyerson et al.*, 2002] and with an increase in solar energy output. The close of this cooling event coincides with the onset of the modern rise in CO_2, followed by the warmest temperatures of the last >700 years in West Antarctica based on the $\delta^{18}O$ Siple Dome ice core record [*Mayewski et al.*, 2004b] and indeed of the last 10,000 years (Figure 7). The close of the cooling event is coincident with a major transition from zonal to mixed flow in the North Pacific [*Fisher et al.*, 2004], suggesting a global-scale association between Antarctic and North Pacific climate. Further investigation into this most recent abrupt climate change event to impact

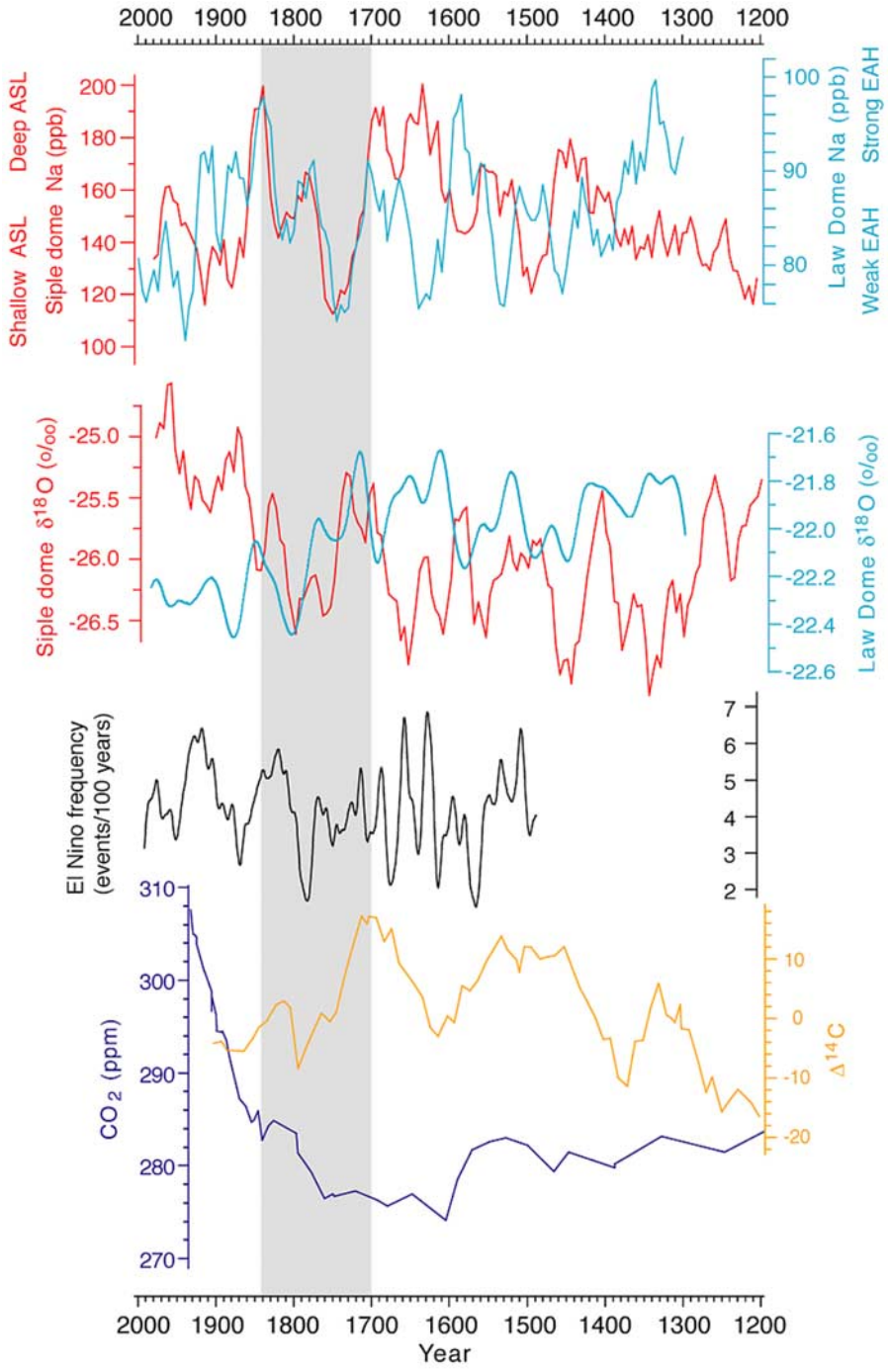

Figure 9. The 25 year running mean of Siple Dome (SD, red) and Law Dome (DSS, blue) Na (ppb) used as a proxy for the Amundsen Sea Low (ASL) and East Antarctic High (EAH), respectively, with estimated sea level pressure developed from calibration with the instrumental and National Centers for Environmental Prediction reanalysis (based on *Kreutz et al.* [2000] and *Souney et al.* [2002]). Twenty-five year running mean SD (red) and DSS (blue) $\delta^{18}O$ (‰) used as a proxy for temperature, with estimated temperature developed from calibration with instrumental mean annual and seasonal temperature values [*Van Ommen and Morgan*, 1996; *Steig et al.*, 2000]. Frequency of El Niño polar penetration (51-year Gaussian filter, black) based on calibration between the historical El Niño frequency record [*Quinn et al.*, 1987; *Quinn and Neal*, 1992] and South Pole methane sulphonate [*Meyerson et al.*, 2002]. Reprinted from *Mayewski et al.* [2004b] with permission of the International Glaciological Society. $\Delta^{14}C$ series used as an approximation for solar variability [*Stuiver and Braziunas*, 1993]; values younger than 1950 are bomb contaminated. CO_2 from DSS ice core [*Etheridge et al.*, 1996]. Darkened area shows 1700–1850 year era climate anomaly discussed in section 1.

Antarctica could have relevance to events that might occur as polar climates adjust to future warming.

[28] Information about Antarctic climate change also comes from investigations of the closed basin lakes in the McMurdo Dry Valleys region in southern Victoria Land, which provide a detailed picture of variations in the hydrological system in this region over the past 3000–4000 years (Figure 8). These lakes respond quickly and dramatically to changes in summer temperatures, which are associated with changes in the input of water from melting glaciers in the area. Lake Bonney and Lake Vida, the more inland lakes in the McMurdo region, began to refill at ~3000–2800 years ago, some ~2000 years before the refilling of more coastal lakes [*Doran et al.*, 2003; *Poreda et al.*, 2004]. The more coastal lakes, lakes Fryxell, Hoare, and Vanda, reached extremely low levels by 1200–1000 years ago [*Wilson*, 1964; *Lyons et al.*, 1998; *Poreda et al.*, 2004], then began to refill when the Ross Sea climate started to warm [*Leventer et al.*, 1993]. Lake Wilson in the Darwin Glacier area at 80°S appears to have undergone a similar drying event prior to ~1000 years ago with increased meltwater input after this time [*Webster et al.*, 1996]. The timing of these climatically induced fluctuations in lake levels can also be recognized from studies of abandoned Adélie penguin rookeries along the Victora Land coast, based on the dependence of these penguins on sea ice extent in the modern environment [*Baroni and Orombelli*, 1994]. Investigations of the distributions of Adélie penguin rookeries suggest a ''penguin optimum'' associated with a warmer climate and less sea ice between 4000 and 3000 years ago [*Baroni and Orombelli*, 1994], but this ''optimum'' ended abruptly ~3000 years ago, as the inland lakes began to fill and the coastal lakes began to decrease in size. Abandoned rookeries were reoccupied between 1200 and 600 years ago, also supporting warming along the southern Victoria Land coast [*Baroni and Orombelli*, 1994]. All these data suggest that the filling of the more inland lakes occurred at the end of a climatic optimum, and they did not decline during the subsequent cooling phase that followed or they filled during a time of climatic deterioration along the coastal areas of Victoria Land [*Poreda et al.*, 2004].

2. LAST 50–200+ YEARS

[29] Several SCAR activities have focused on understanding the climate of the last 50–200+ years, and many new data are now emerging. Here we focus on measured and estimated changes in (1) atmospheric temperature, (2) atmospheric circulation, (3) atmospheric chemistry, (4) ocean temperature and salinity, (5) ocean circulation, (6) sea ice and ice shelves, and (7) the mass balance of the Antarctic ice sheet and of glaciers in the Antarctic Peninsula, the sub-Antarctic islands, southern South America, and New Zealand, as described in section 2.7.

2.1. Changes in Atmospheric Temperature

[30] The Antarctic has undergone complex temperature changes in recent decades [*Turner et al.*, 2005a] (Figure 10). The largest annual warming trends are found on the western and northern parts of the Antarctic Peninsula, with Faraday/Vernadsky having the largest statistically significant (<5% level) trend at +0.56°C/decade from 1951 to 2000. Rothera station, some 300 km to the south of Faraday, has a larger annual warming trend, but the shortness of the record and the large interannual variability of the temperatures render the trend statistically insignificant. Although the region of marked warming extends from the southern part of the western Antarctic Peninsula north to the South Shetland Islands, the rate of warming decreases north away from Faraday/Vernadsky (50 year long record), with the long record from Orcadas (100 year long record) in the South Orkney Islands showing a warming trend of only +0.20°C/decade.

[31] The large winter season warming of 5°C over 50 years at Faraday is believed to be associated with a significant decrease in winter sea ice over the Amundsen-Bellingshausen Sea. The reason why there was more sea ice in the 1950s and 1960s is not known with certainty but may have been linked to weaker/fewer storms to the west of the peninsula and greater atmospheric blocking. A greater frequency of blocking anticyclones would have meant weaker northerly winds to the west of the Antarctic Peninsula, allowing the sea ice to advance farther north during the winter and giving colder temperatures on the western side of the Peninsula.

[32] On the eastern side of the peninsula the greatest warming is during the summer months and appears to be associated with the strengthening of the circumpolar westerlies that has taken place as the Southern Hemisphere Annular Mode has shifted into its positive phase as evidenced in the instrumental record at least since the mid-1970s [*Marshall et al.*, 2006]. Stronger winds have resulted in more relatively warm, maritime air masses crossing the peninsula and reaching the low-lying ice shelves, as well as the adiabatic descent and warming of these winds crossing the Antarctic Peninsula topography.

[33] Around the rest of the coastal region of the continent there have been few statistically significant changes in surface temperature over the last 50 years (Figure 10). However, Amundsen-Scott Station at the South Pole has shown a statistically significant cooling in recent decades that is thought to be a result of fewer maritime air masses penetrating into the interior of the continent. Unfortunately there is no long-term instrumental measurement record for the area of the Siple Dome, in West Antarctica, to compare with the warming seen in the ice core record (Figure 9). The nearest such instrumental record is at McMurdo Station, where slight warming is apparent (Figure 10).

[34] Large-scale calibrations between satellite-deduced surface temperature and ice core proxies for temperature are also now available [*Schneider and Steig*, 2002]. Reconstruction of temperatures over the past 200 years, based on eight records distributed over the ice sheet, suggests no discernible trend over recent decades, but do note a 0.2°C warming for the past century [*Schneider et al.*, 2006].

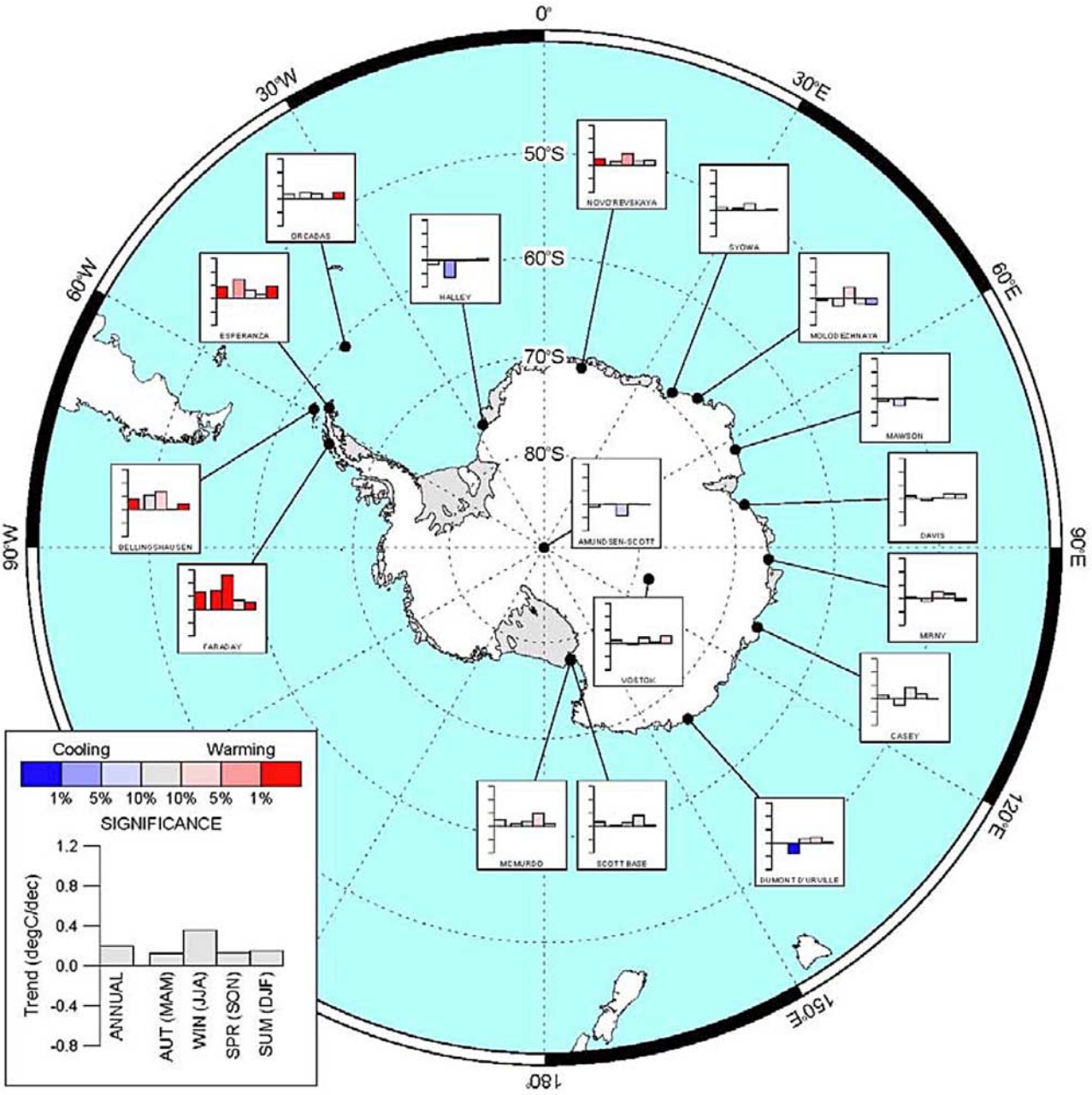

Figure 10. Annual and seasonal near-surface temperature trends for stations with long in situ records. The trends are for the full length of each record. The colors indicate the statistical significance of the trends. Trends were computed for the full length of each record. Most stations started reporting around the time of the International Geophysical Year in 1957/1958. Autocorrelation was taken into account in computing significance levels.

[35] Where ice cores are not available, lake levels provide information on climate. Because lake levels are extremely sensitive to summer temperature and albedo changes, fluctuations in their levels aid our understanding of climate change in the McMurdo Dry Valleys region. Most lakes there rose from 0.7 to 3.3 m/decade in the 1970s–1990s [*Chinn*, 1993], and Lake Wilson increased in volume by over 50% during that period [*Webster et al.*, 1996]. There is some evidence that Lake Bonney has been increasing in size since the early 1900s [*Chinn*, 1993]. A summer cooling trend during the 1990s significantly slowed this rising trend in lake levels [*Doran et al.*, 2002]. It has been suggested that the summer cooling was caused by a change in atmospheric circulation driven by the El Niño–Southern Oscillation (ENSO), resulting in a decrease in marine air mass influences reaching the dry valleys and an increased inflow to the region from West Antarctica via the Ross Ice Shelf [*Bertler et al.*, 2004].

[36] Analysis of Antarctic radiosonde temperature profiles indicates that there has been a warming of the winter troposphere and cooling of the stratosphere over the last 30 years. The data show that regional midtropospheric

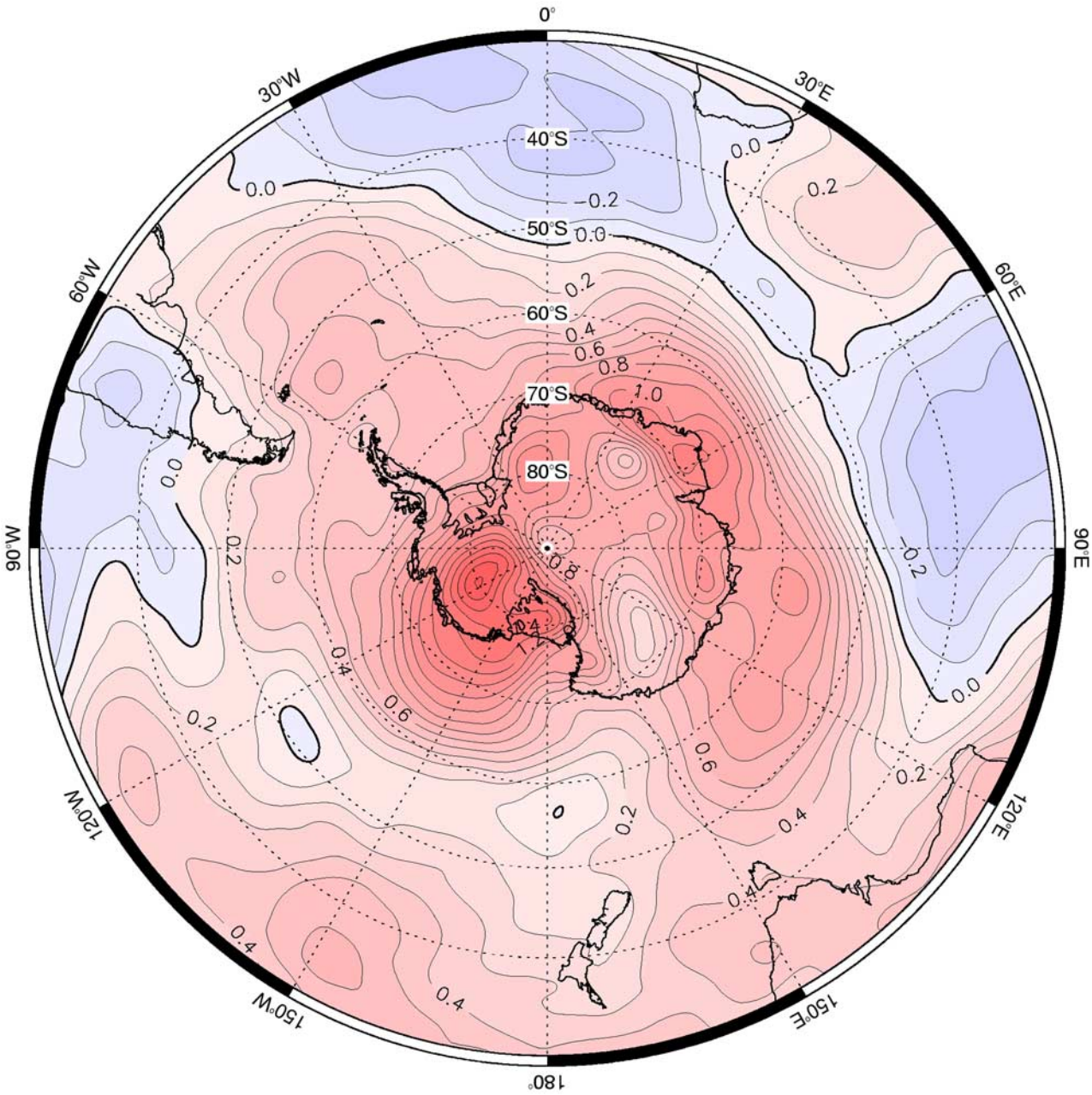

Figure 11. Trends in the 500 hPa temperature over 1979–2001 from the ECMWF 40 year reanalysis, showing tropospheric warming at ~5 km. The contours are °C/decade. From *Turner et al.* [2006]. Reprinted with permission from AAAS.

temperatures have increased most around the 500 hPa level with statistically significant changes of 0.5–0.7°C/decade (Figure 11) [*Turner et al.*, 2006]. From 1985 to 2002, the lower part of the stratosphere cooled by 10°C, and the time of decay of the polar vortex shifted from early November during the 1970s to late December in the 1990s [*Thompson and Solomon*, 2002].

2.2. Changes in Atmospheric Circulation Over Antarctica and the Southern Ocean

[37] The Antarctic atmosphere is cold and dry. There is a strong horizontal temperature gradient between the continent and the ocean and a strong vertical temperature gradient (inversion) as a result of the intense radiative cooling during the winter. For this reason, near-surface temperatures are particularly sensitive to low-level atmospheric circulation [*Van den Broeke*, 2000]. Baroclinic and depression activity peak in March and September, leading to a twice yearly contraction and expansion of the circumpolar low-pressure belt, known as the Semiannual Oscillation (SAO). Since the mid-1970s a significant weakening of the SAO has been observed, leading to cooling in coastal Antarctica in May to June [*Van den Broeke*, 2000].

[38] The principal mode of variability in the atmospheric circulation of the extratropics and high latitudes has been referred to as the Southern Annular Mode (SAM) (also

known as the high-latitude mode or the Antarctic Oscillation). The SAM has a zonally symmetric or annular structure, with synchronous anomalies of opposite sign in high latitudes and midlatitudes, although a zonal wave number three pattern is superimposed [*Lefebvre et al.*, 2004]. It can be seen in many parameters measured at high latitudes, such as surface pressure (e.g., the pressure difference between latitudes 40 and 65°S) and temperature, geopotential height, and zonal wind. Observational and modeling studies have shown that the SAM contributes a large proportion (~35%) of the Southern Hemisphere climate variability on a large range of time scales, from daily [e.g., *Baldwin*, 2001] to decadal [*Kidson*, 1999], and is also likely to drive the large-scale circulation of the Southern Ocean.

[39] Over the last 50 years the SAM has shifted more into its positive phase with decreases of surface pressure over the Antarctic and corresponding increases at midlatitudes [*Marshall*, 2003; *Thompson and Solomon*, 2002]. This has resulted in an increase in the westerlies over the Southern Ocean, with consequent oceanographic implications. These include a likely intensification of the eddy field [*Meredith and Hogg*, 2006] and a reduction of the efficiency of the Southern Ocean CO_2 sink associated with changes in upwelling and mixing [*Le Quéré et al.*, 2007]. The trend has also been linked to warming of the Antarctic Peninsula region and a general cooling of the Antarctic continent [*Marshall et al.*, 2002, 2006; *Van den Broeke and van Lipzig*, 2003]. *Van den Broeke and van Lipzig* [2004] argued that the reason for the winter cooling over East Antarctica during periods of high SAM index is greater thermal isolation of Antarctica, due to increase zonal flow, decreased meridional flow, and intensified temperature inversion on the ice sheet due to weaker near-surface winds. They showed that a strengthening circumpolar vortex leads to a pronounced deepening of the Amundsen Sea Low, cooling of most of Antarctica with the exception of the Antarctic Peninsula, as well as drier conditions over large parts of West Antarctica, the Ross Ice Shelf, and the Lambert Glacier region and wetter conditions elsewhere. Other studies have demonstrated the influence of the SAM on spatial patterns of precipitation variability in Antarctica [*Genthon et al.*, 2003] and southern South America [*Silvestri and Vera*, 2003].

[40] A number of modeling studies attribute recent positive summer changes in the SAM to ozone depletion [*Sexton*, 2001; *Thompson and Solomon*, 2002; *Gillett and Thompson*, 2003]. The polar vortex is most pronounced in the winter stratosphere when the air above the continent is extremely cold. However, the loss of springtime ozone as a result of the ''ozone hole'' has also cooled the stratosphere through the spring and summer months. This in turn has resulted in low mean sea level pressure in the Antarctic at this time of year, thereby shifting the SAM into its positive phase.

[41] Other studies have demonstrated that positive changes in the SAM may occur in response to greenhouse gas increases [e.g., *Fyfe et al.*, 1999; *Kushner et al.*, 2001; *Stone et al.*, 2001; *Cai et al.*, 2003; *Rauthe et al.*, 2004]. *Hartmann et al.* [2000] hypothesized that synergistic interactions between these anthropogenic forcing factors are responsible for the changing SAM, while *Marshall et al.* [2004] suggested that natural forcings, such as changes in shortwave radiation, may also have played a role. A number of studies [e.g., *Fogt and Bromwich*, 2006; *Mo*, 2000; *Zhou and Yu*, 2004; *L'Heureux and Thompson*, 2006] also support the role that natural forcings play in the variability of the SAM, as they have shown a linear relationship of the SAM with tropical Pacific sea surface temperatures (SSTs). However, these SST changes could be a result of the greenhouse gas increases and may not be an independent forcing mechanism on the SAM variability. Similarly, cooling of the stratosphere, which would lead to increases in the SAM, could be caused by both ozone depletion in the stratosphere and increases in greenhouse gases in the troposphere, and thus these mechanisms are likely all related [*Arblaster and Meehl*, 2006; *Cai and Cowan*, 2007].

[42] ENSO is the farthest reaching climatic cycle on Earth on decadal and subdecadal time scales. Since 1977, the Southern Oscillation Index (SOI, the atmospheric component of ENSO) has shifted toward a more negative phase, associated with more frequent and stronger El Niño events (the ocean component of ENSO). It has a profound effect not only on the weather and oceanic conditions across the tropical Pacific, where the ENSO has its origins, but also in the high-latitude areas of the Southern Hemisphere, most markedly in the South Pacific (see *Turner* [2004] for a review). In this region, a large blocking high-pressure center often forms during El Niño warm events [*Renwick and Revell*, 1999; *Renwick*, 1998; *Van Loon and Shea*, 1987]. The low-frequency variability of the atmospheric circulation is readily seen in the amplitude of this pressure center, which is part of the Pacific South American (PSA) pattern [*Mo and Ghil*, 1987]. The PSA represents a series of alternating positive and negative geopotential height anomalies extending from the west central equatorial Pacific through Australia/New Zealand, to the South Pacific near Antarctica/South America, and then bending northward toward Africa. It follows a great circle trajectory that is induced by upper level divergence initiated from tropical convection [*Revell et al.*, 2001] and is the conduit by which the ENSO anomalies in the tropics reach the high southern latitudes through the atmosphere.

[43] The decadal variability of the ENSO signal in high southern latitudes is well documented. Results from *Bromwich et al.* [2000, see also 2004a] indicate a strong shift in the correlation between West Antarctic precipitation minus evaporation (P-E) and the SOI using atmospheric reanalysis and operational analysis over the last 2 decades. The time series of P-E was positively correlated with the SOI until about 1990, after which it became strongly anticorrelated, a relationship that persisted through 2000, after which the relationship became weak, demonstrating that forecasting dependence on this association requires caution. In addition, the positive SOI–summer temperature correlation in the western Ross Sea (cooler during El Niño and warmer during La Niña time periods) was only marginally significant during the 1980s but strongly significant

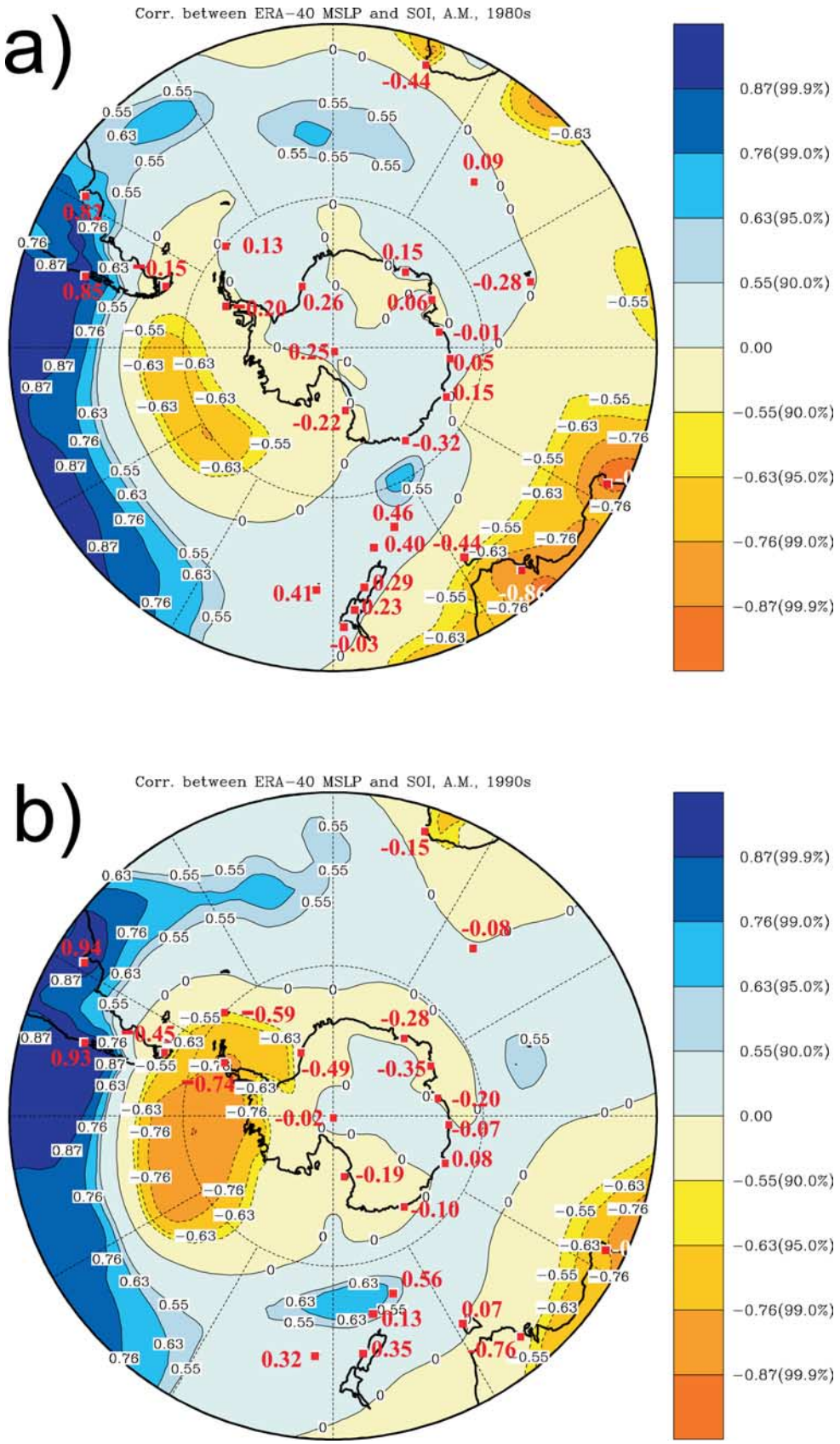

Figure 12. Colored contours denote the spatial correlations (shaded by statistical significance) of ERA-40 mean sea level pressure (MSLP) and the Southern Oscillation Index (SOI) for (a) the 1980s and (b) the 1990s. Also plotted, in red numbers, are the observed MSLP-SOI correlations for selected stations south of 30°S; at the South Pole station, pressure was used instead of MSLP. Significance levels for the correlation values are listed beside the key. Adapted from *Fogt and Bromwich* [2006].

during the 1990s [*Bertler et al.*, 2004]. *Fogt and Bromwich* [2006] similarly noted strong changes in the strength of the ENSO teleconnection between the 1980s and the 1990s. Figure 12 displays the annual mean sea level pressure (May–April) correlation of the European Centre for Medium-Range Weather Forecasts 40-year reanalysis with the SOI, a pressure-based record used to monitor ENSO variability. Figure 12a demonstrates a weak teleconnection to the South Pacific during the 1980s, which amplifies and shifts southeastward in the 1990s (Figure 12b). Expanding upon *Bertler et al.* [2004], *Fogt and Bromwich* [2006] demonstrated that the decadal variability of the high-latitude ENSO teleconnection to the South Pacific is governed by the phase of the SAM. When both are in the same phase (i.e., La Niña occurring with positive phases of the SAM and El Niño occurring with negative phases), the teleconnection is amplified; it is dramatically weakened in periods when these two main climate modes are out of phase (i.e., El Niño occurring with positive phases of the SAM). When the combined forcing of both climate drivers is considered, another decadal change is apparent. In a case study for the McMurdo Region, *Bertler et al.* [2006] showed that in the 1980s a combined SOI-SAM summer index led (+1 year) over regional summer temperatures, while it lagged (−1 year) during the 1990s, providing further evidence for a recent change or decadal oscillation in the Antarctic-tropical teleconnection.

[44] Teleconnections like the link between ENSO and the Antarctic are usually identified by statistical analyses. Much work has been done on trying to understand the underlying mechanisms using diagnostics from climate models. However, some models do not have a good representation of ENSO, which creates problems, and ocean models still lag atmospheric models in their development, so the reliance on statistics reflects the current state of research.

[45] The 21-nation consortium of the International Trans-Antarctic Scientific Expedition (ITASE) has pioneered calibration tools and reconstruction of climate indices and evidence for climate forcing using single sites through to multiple arrays of sites based on shallow ice cores covering the past 200–1000 years [*Mayewski et al.*, 2004b]. These shallow ice core records provide annually resolved accumulation rate and $\delta^{18}O$ temperature proxies used as ground truth for precipitation and temperature reanalysis products [e.g., *Genthon et al.*, 2005; *Schneider et al.*, 2005]. ITASE ice core chemistry data calibrated against modern instrumental climate data also provide climate proxies for global-scale atmospheric circulation features, for example, ENSO, by using MSA [*Meyerson et al.*, 2002; *Bertler et al.*, 2004] and for major regional atmospheric circulation features such as the Amundsen Sea Low, plus high-pressure ridging over East Antarctica, and the SAM, using Na^+, NO_3^-, and Ca^{++}, respectively [*Kreutz et al.*, 2000; *Souney et al.*, 2002; *Goodwin et al.*, 2004; *Proposito et al.*, 2002; *Shulmeister et al.*, 2004; *Kaspari et al.*, 2005; *Yan et al.*, 2005] (also see discussions in section 1 in relation to Figures 7 and 9). Ice core proxy reconstructions for the Amundsen Sea Low and the southern circumpolar westerlies indicate that these circulation features are still within the range of variability established over the last 1200 years [*Mayewski and Maasch*, 2006], despite recent impact by human-induced changes in stratospheric ozone on the strength of the circumpolar westerlies around the edge of the polar vortex [*Thompson and Solomon*, 2002]. But it is also true to say that even though the wind strengths are the same as they have been over the past 1200 years, the temperatures are different; thus in the past few decades the overall system has moved outside the range of variability established over the past 1200 years. Ice core climate proxies offer the opportunity for significantly more reanalysis testing and climate modeling.

[46] Evidence for inland penetration of summertime marine tropospheric air masses over the last few decades, relative to the last few hundred years, was detected using ice core marine sourced $SO_4^=$ inputs noted in portions of coastal West Antarctica near the Amundsen Sea [*Dixon et al.*, 2005]. Future insights into the timing and location of this and other marine air mass penetrations may prove useful in determining the cause for changes in sea ice extent and assessing the impact of greenhouse gas warming over the Southern Ocean.

[47] The impact of solar forcing (via UV induced changes in stratospheric ozone concentration) on the southern circumpolar westerlies at the edge of the polar vortex has been suggested through an association established between ice core climate proxies for the westerlies and for solar variability [*Mayewski et al.*, 2005]. This work reveals decadal-scale associations between the circumpolar westerlies, inferred from West Antarctic ice core Ca^{++}, and ^{10}Be, a proxy for solar variability in a South Pole ice core [*Raisbeck et al.*, 1990] over the last 600 years, and with annual-scale associations with solar variability inferred from the solar cycle since A.D. 1720. Increased solar irradiance is associated with increased zonal wind strength near the edge of the Antarctic polar vortex, and the winds decrease with decreasing irradiance. The association is particularly strong in the Indian and Pacific oceans and may contribute to understanding the role of natural climate forcing on drought in Australia and other Southern Hemisphere climate events.

2.3. Changes in Atmospheric Chemistry Over Antarctica

[48] Antarctic air, snow, and ice samples also provide information on changes in the chemistry of the atmosphere over time scales from storm events to hundreds of thousands of years and representing chemical variation from local to global scales.

[49] The Antarctic is particularly well known for greenhouse gas records of CO_2, CH_4, and N_2O not only from ice cores but also from in situ observations made since the onset of modern monitoring during the International Geophysical Year (IGY) of 1957–1958. Antarctic greenhouse gas monitoring has been essential in identifying the dramatic anthropogenic source increases in CO_2, CH_4, and N_2O, now close to 380, 1755, and 320 ppbv, respectively.

[50] Continent-wide Southern Hemisphere springtime depletion in another greenhouse gas, stratospheric ozone (O_3,),

was identified in the mid-1980s by continuation of a monitoring program initiated during IGY. Depletion is currently close to 60%. It is attributed to halogen-catalyzed chemical destruction, largely attributable to emissions of chlorofluorocarbons (CFCs) that provide a source for the halogen chlorine. Chlorine (along with bromine from industrial sources) accumulates throughout the winter in polar stratospheric clouds where upon contact with solar ultraviolet light during the Antarctic sunrise it acts to destroy O_3. Bromine has a similar effect. Although the total combined abundance of ozone-depleting substances continues to decline [*World Meteorological Organization (WMO)*, 2006], 2006 saw the largest Antarctic ozone hole yet. On 24 September 2006 it covered 29 million km^2, and on 8 October 2006 it reached the lowest yet measured ozone values of 85 Dobson units (http://ozonewatch.gsfc.nasa.gov/). It is not yet clear if this is part of a trend, but during 2002–2005, global mean total column ozone concentrations were 3.5% below the 1964–1980 average. There is significant year-to-year variability, which is poorly understood. For example, in the 2002 Southern Hemisphere spring, a sudden stratospheric warming caused a significant reduction in the size of the Antarctic ozone hole and led to higher ozone concentrations than usual [*WMO*, 2006]. Observations of O_3 destroying compounds demonstrate the complications associated with predicting future O_3 depletion. Between 2000 and 2004, tropospheric methyl chloroform and methyl bromide decreased by ~60 and ~45 ppt (8–9%), respectively. Since 2000, tropospheric emissions of chlorofluorocarbons (CFC-11, CFC-12, and CFC-113) decreased by 1.9 ppt/a (0.8%), 5 ppt/a (1%), and 0.8 ppt/a (1%), respectively. In contrast, hydrochlorofluorocarbons (HCFCs), in particular, HCFC-22, increased by 4.9 ppt/a (3.2%) [*WMO*, 2006].

[51] The reconstruction of changes in Pb pollution during past centuries has been achieved by analyzing snow samples collected in Coats Land, Victoria Land, and Law Dome [*Van de Velde et al.*, 2005] and from deep ice cores such as Vostok [*Hong et al.*, 2003]. These records show that Pb pollution over Antarctica started as early as the 1880s. Further, atmospheric Pb concentrations have increased twofold to threefold over Antarctica compared to average Holocene levels [*Boutron and Patterson*, 1983]. Isotopic data suggest that large anthropogenic Pb inputs to Antarctica from the 1970s to 1980s are linked to the rise and fall in the use of leaded gasoline, as already clearly observed in the Arctic [*Boutron et al.*, 1991; *Rosman et al.*, 1993]. A clear decrease in Pb levels is observed during recent years in parallel with the phasing out of Pb additives in gasoline. Antarctica also shows slight contamination with other trace metals such as Cr, Cu, Zn, Ag, Pb, Bi, and U as a consequence of long-distance transport from surrounding continents [*Wolff and Suttie*, 1994; *Wolff et al.*, 1999; *Planchon et al.*, 2002; *Vallelonga et al.*, 2002; *Van de Velde et al.*, 2005]. In general, increases in anthropogenic source pollutants, including trace elements, can be attributed to a combination of the following: Antarctic logistic activities [*Boutron and Wolff*, 1989; *Qin et al.*, 1999], industrial activities in the Southern Hemisphere, and possibly also industrial activity in the Northern Hemisphere.

[52] Persistent organic pollutants (POPs), pesticides, and industrial chemicals are found worldwide because of their propensity for long-range atmospheric transport. The distribution pattern and levels of POPs such as polychlorobiphenyls and chlorinated pesticides such as DDT and hexachloribenzene are reported in the Antarctic environment [*Fuoco et al.*, 1996; *Corsolini et al.*, 2002; *Weber and Goerke*, 2003; *Montone et al.*, 2003]. Anthropogenic source radionuclides stemming from aboveground nuclear bomb testing are also present throughout Antarctica, as is evidence of the Chernobyl nuclear accident in at least the region of the South Pole [*Dibb et al.*, 1990].

[53] Maps of the surface distribution of soluble chemical species from *Bertler et al.* [2005] indicate the potential scope for exploring chemical and climate variability (for examples, see maps of sodium and sulphate concentration in Figure 2b). As expected, the East Antarctic interior shows significantly lower values of marine source Na^+ (~2 to ~30 ppb) than coastal sites (~75 to ~15,000 ppb). The change from very low to very high concentrations occurs close to the coast. This might be caused by the dominant influence of either cyclones (Na^+ rich) or katabatic winds (Na^+ depleted). The Antarctic Peninsula shows high values overall. Marine source Cl^- variability exhibits a similar pattern to Na^+, ranging from ~1 to ~28,000 ppb, with highest values at coastal sites (~150 to ~28,000 ppb) and lower values in the interior (1 to ~150 ppb). The Antarctic Peninsula shows overall high values and no significant trend with elevation. The spatial variability of multisource (biomass, lightning, marine, and human activity) NO_3^- ranges from ~4 to ~800 ppb. Highest values are observed in Enderby Land, Dronning Maud Land, and Victoria Land, ranging from ~30 to 800 ppb. Intermediate values are reported from Marie Byrd Land, Ronne Ice Shelf, South Pole region, and Northern Victoria Land (~35 to ~100 ppb), while lowest values are observed in the Antarctic Peninsula and Kaiser Wilhelm II Land (~0 to ~20 ppb). While NO_3^- has been shown to be affected by postdepositional loss in low-accumulation sites [*Legrand and Mayewski*, 1997], the lowest values for NO_3^- are observed at sites with relatively high annual accumulation: the Antarctic Peninsula and Kaiser Wilhelm II Land. The spatial variability of multisource (marine, evaporite, volcanic, and human activity) $SO_4^=$ ranges from 0.1 to ~4000 ppb. $SO_4^=$ is particularly prone to sporadic high input through volcanic events. The Antarctic Peninsula is characterized by low values (~10 to 30 ppb), with higher values only at coastal sites (~75 to 1000 ppb). Values from Kaiser Wilhelm II Land are also relatively low (15 to 70 ppb). The spatial variability of marine phytoplankton sourced MSA ranges from 3 to ~160 ppb. With the exception of some coastal sites, the highest concentrations are at coastal sites, decreasing inland. Terrestrial and marine source Ca^{++}, Mg^{++}, and K^+ concentrations range from 0.1 to 740 ppb, from 0.2 to ~2000 ppb, and from 0.1 to 600 ppb, respectively. All three species show low concentrations across Antarctica, with a

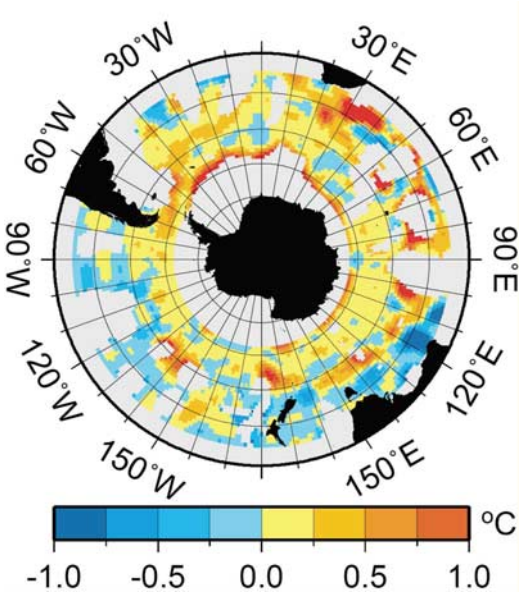

Figure 13. Temperature trends computed from float/hydrography differences bin averaged in 1° × 1° squares. For this analysis, float/hydrography pairs were used if the hydrographic measurements were collected after 1930, and they were separated from the float observations by at least 10 years in time and by less than 220 km in space. Latitude and longitude grid lines are at 10° intervals. From *Gille* [2002]. Reprinted with permission from AAAS.

few exceptions where local dust sources, such as the McMurdo Dry Valleys, or a strong marine influence, as at Terra Nova or at coastal sites at the Antarctic Peninsula, cause orders of magnitude higher concentrations.

2.4. Changes in Ocean Temperature and Salinity

[54] The Southern Ocean is one of the most poorly sampled areas on the planet, and investigations into the magnitude and causes of variability and change here are hampered by a scarcity of data from earlier eras. Little or no data exist from large areas of the Southern Ocean prior to the 1950s, and this problem persisted up to the advent of the satellite era in the 1970s and 1980s. Even today, with routine data collection from the ocean surface by satellite, obtaining information on subsurface changes over wide areas remains a huge logistical challenge, especially under the winter sea ice. Programs such as Argo (which has an emerging sub-sea-ice component) and Southern Elephant Seals as Oceanographic Samplers (the use of marine mammals for operational oceanography) are seeking to address this data void, but these are comparatively recent innovations, and it will be some years before sufficient data are accumulated to robustly reveal changes in many sectors. This lack of data severely compromises our understanding of and ability to model numerically the role of the Southern Ocean in the global climate system.

[55] These facts notwithstanding, there are indications that some very important, large-scale changes have occurred in the Southern Ocean. These most significant of the changes reported thus far are (1) a strong warming in the waters of the ACC; (2) a warming of the Antarctic Bottom Water (AABW) exported to the South Atlantic; (3) a freshening of the waters in the Indian and Pacific sectors of the Southern Ocean, including the AABW formed here; and (4) a remarkably strong warming of the upper layers of the ocean adjacent to the western Antarctic Peninsula. We here summarize each of these in turn.

2.4.1. Warming of the Circumpolar Southern Ocean

[56] Temperature records from autonomous floats drifting between 700 and 1100 m during the 1990s were collated and examined by *Gille* [2002, 2003], who made comparisons of these data with earlier temperature records collected during earlier decades (1950s onward) from ship-based instruments. A large-scale significant warming of the ACC was apparent in this depth range, of around 0.2°C (Figure 13). This was recently expanded on by *Gille* [2008], where the vertical structure of the warming was investigated, and it was found that the warming signal was vertically coherent and greatest near the surface. This pattern of warming is in accord with other studies that used large-scale compilations of in situ data, such as those of *Levitus et al.* [2000, 2005] (Figure 14), though notably of a larger magnitude.

[57] *Gille* [2008] argued that some of the warming observed could be attributable to a southward shift of the ACC current cores, essentially reflecting a redistribution of heat rather than an increase. This is in line with previous regional interpretations of change [e.g., *Aoki et al.*, 2003] and some modeling studies [e.g., *Fyfe et al.*, 2007], though other interpretations are possible, and this remains an active area of research. For example, the role of increasing air-sea heat fluxes has also been indicated in some works [e.g., *Fyfe et al.*, 2007], and it has also been shown that the known strengthening of the circumpolar westerly winds could lead to increased eddy activity in the Southern Ocean and a consequent increase in the poleward eddy heat flux [*Meredith and Hogg*, 2006; *Hogg et al.*, 2008].

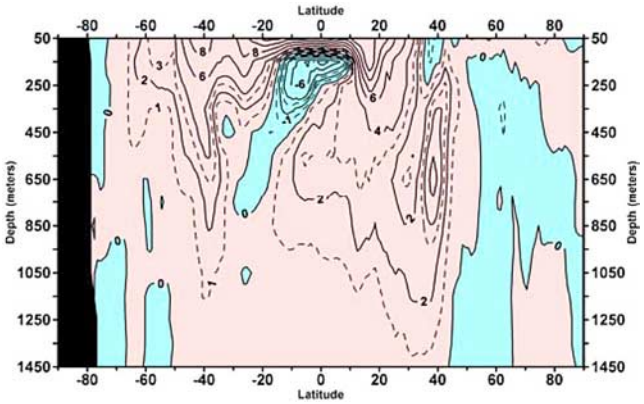

Figure 14. Zonal trend in ocean heat content during the second half of the twentieth century [from *Levitus et al.*, 2005]. Note in particular the strong warming in the region of the Antarctic Circumpolar Current. (Contour interval is 2×10^{18} J/a.)

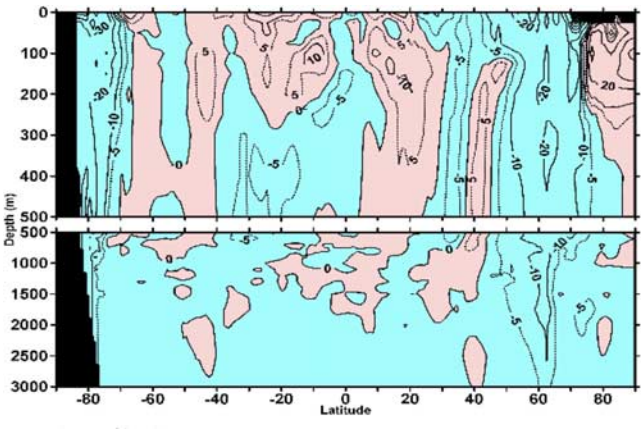

Figure 15. Zonal mean trend in ocean salinity (in units of 10^{-4}/a) during the second half of the twentieth century [from *Boyer et al.*, 2005]. Blue areas denote freshening; pink areas denote salinification. Note in particular the very strong freshening south of the region of the Antarctic Circumpolar Current.

[58] *Levitus et al.* [2005] suggested that the observed warming in the Southern Ocean was probably due to a combination of natural variability and anthropogenic forcing. *Fyfe* [2006] investigated this via coupled modeling and found that the current generation of climate models can produce a warming in the Southern Ocean comparable to that observed if anthropogenic gases and sulphate and volcanic aerosols are included. While the important role of the Southern Ocean eddy field is not well represented in these models, this study nonetheless lends weight to the argument that human activities have influenced Southern Ocean temperatures. This study also demonstrated that if the role of volcanic aerosols is neglected in climate modeling simulations, the simulated warming is nearly double, suggesting that the human impact on Southern Ocean warming is only partially realized at present because of the cooling effect of the aerosols.

[59] The circumpolar warming of the waters in the ACC is possibly impacting waters farther south in the subpolar gyres also. For example, *Robertson et al.* [2002] noted a warming of 0.3°C for the Warm Deep Water (WDW) layer of the Weddell Gyre. This is the comparatively warm, saline water mass that enters the subpolar gyre from the ACC. This change occurred over the period 1970s to 2000, and it has been argued that anomalous intrusions of water from the ACC contributed to this warming [*Fahrbach et al.*, 2004], with a recovery from the Weddell Polynya of the mid-1970s also possibly implicated [*McPhee*, 2003]. (More recently, a cooling of WDW across the Weddell Gyre was reported [*Fahrbach et al.*, 2004], though whether this is a genuine reversal of the decadal warming trend or simply a brief interruption remains unclear.)

2.4.2. Warming of the AABW Exported to the South Atlantic

[60] Intense air-sea-ice interactions in the Weddell Sea lead to the formation of the dense, cold Weddell Sea Deep Water (WSDW) and the even denser Weddell Sea Bottom Water (WSBW). Unlike WSBW, which is topographically constrained to circulate within the Weddell Gyre, WSDW is readily exported northward, whereupon it constitutes the densest layer of the AABW in the Atlantic.

[61] Various studies have demonstrated a decadal-scale warming the AABW of the South Atlantic. In the Argentine Basin, an investigation of hydrographic data collected during the 1980s showed that the coldest abyssal layer warmed significantly in that interval [*Coles et al.*, 1996]. *Johnson and Doney* [2006] demonstrated that this warming continued thereafter, up to at least 2005. Other studies have demonstrated AABW warming even farther north in the Atlantic, including at Vema Channel, the Brazil Basin to the north, and even as far as the equator [*Andrié et al.*, 2003; *Johnson and Doney*, 2006; *Zenk and Morozov*, 2007]. *Meredith et al.* [2008] demonstrated that these lower-latitude changes in AABW were due to changes in the properties of the WSDW being exported northward from the Weddell Sea.

[62] The cause of the WSDW warming in the Weddell Sea has not been determined unambiguously. Indeed, because of a sparsity of long-term measurements from the boundary currents of this region, the WSDW warming signal has not yet been conclusively demonstrated here. However, a warming signal in the WSBW has been shown [*Fahrbach et al.*, 2004], and the processes that lead to formation of these water masses are notably similar. Further work is needed to monitor the evolving WSDW properties directly in the Weddell Sea and understand better the causes of the warming signal exported to lower latitudes.

2.4.3. Freshening of Waters in the Indian and Pacific Sectors

[63] In addition to changes in temperature, changes in ocean salinity are of particular importance in the Southern Ocean. This is a consequence of the equation of state for seawater: at low temperatures, density is dominated by salinity; hence stratification, mixed layer depth, and geostrophic flows all depend very strongly on the changing freshwater budget. Furthermore, the large adjoining sea ice and glacial ice fields mean that very significant changes to the freshwater inputs are possible. That large changes are happening is clearly illustrated by *Boyer et al.* [2005], who used a compilation of ocean salinity measurements from various sources and noted large decreases south of 70°S (Figure 15).

[64] The causes and regional dependence of this freshening have been elucidated in other works. In the Ross Sea, a remarkable freshening was first detailed by *Jacobs et al.* [2002], who proposed a link with melting of glacial ice, based on oxygen isotope measurements. *Jacobs* [2006] demonstrated that the "upstream" region of the Amundsen shelf and upper slope showed a freshening between 1994 and 2000; this is the region in which *Shepherd et al.* [2004] postulated that a warming ocean is eroding the West Antarctic ice sheet, possibly explaining the anomalous freshwater injection. Other notable freshenings include the region between 140°E and 150°E on the George V continental shelf and rise [*Jacobs*, 2004] and at depth in the

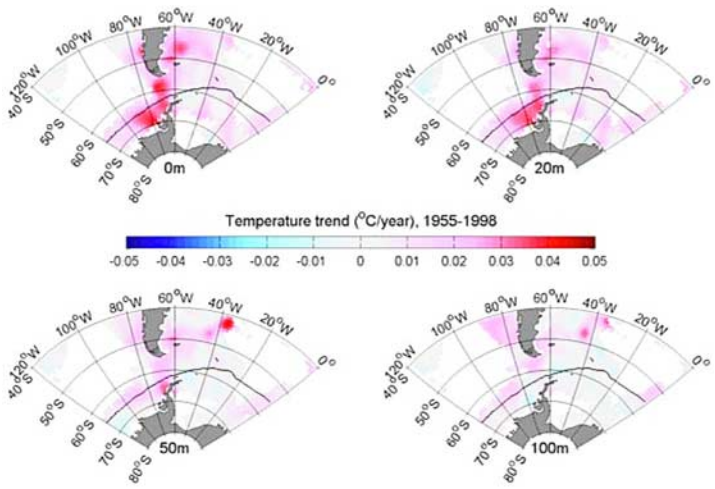

Figure 16. Trends in ocean summer temperature during 1955–1998, for four different depth levels (surface, 20 m, 50 m, and 100 m). Grid cells with no data are left white. Note the significant warming trend observed close to the western Antarctic Peninsula: this is strongly surface intensified, decaying virtually to zero at 100 m depth. From *Meredith and King* [2005, Figure 2].

Australian-Antarctic Basin [*Whitworth*, 2002]. Recently, *Aoki et al.* [2005] observed significant changes in the Adelie Land Bottom Water between the mid-1990s and 2002–2003, based on repeat summer hydrographic observations along 140°E. The putative explanation given was a continuing freshening of source waters supplying bottom water to the Australian-Antarctic Basin. This was expanded upon by *Rintoul* [2007], who observed a freshening in the AABW from both the Indian and Pacific Ocean sector sources. The cause of the decline in shelf water salinity is not yet clear, although an increase in glacial ice melt, increased precipitation, and reduced sea ice production have been identified as possible contributors [*Gordon*, 1998; *Jacobs*, 2004; *Rintoul*, 2007].

2.4.4. Rapid Ocean Warming at the Western Antarctic Peninsula

[65] Atmospherically, the region in the Southern Hemisphere that has been changing most rapidly in recent decades is that to the west of the Antarctic Peninsula [e.g., *Turner et al.*, 2005a]. While the role of the ocean in this warming has been widely speculated upon, it has been difficult to investigate this role in practice because of the strong seasonal bias of data collection from the ocean (the data are almost entirely collected during the austral summer). In essence, waters within the influence of the upper ocean mixed layer show shorter-period variability than those beneath, meaning that higher-frequency sampling is needed. Recently, *Meredith and King* [2005] used a large compilation of in situ hydrographic profiles collected between the 1950s and 1990s to tackle this issue. They found an extremely strong surface-intensified warming (of greater than 1°C) based on austral summer measurements (Figure 16) and a coincident strong summertime salinification. They argued that these were both caused by the reduction in sea ice production in this sector since the 1950s, with the salinification being due to the strong seasonal bias in sampling (wintertime measurements, if available, would likely have shown a freshening). It was noted that the trends observed were positive feedbacks, acting to sustain and enhance the atmospheric warming and induce further reductions in sea ice formation. The profound consequences for the ocean ecosystem in this sector were also noted [cf. *Atkinson et al.*, 2004; *Peck et al.*, 2004].

2.5. Changes in Southern Ocean Circulation

[66] Notwithstanding the magnitude of the challenge in observing and explaining the changes in ocean properties, the aspect of Southern Ocean change that has remained most elusive to date is the variability and change in the circulation. This is a serious problem, as Southern Ocean circulation change is thought to have played a pivotal role in driving past global climatic transitions [*Watson and Naveira Garabato*, 2006; *Toggweiler et al.*, 2006] and stands out as a key element of the oceanic response to recent and projected atmospheric trends in model simulations of climate change [*Saenko et al.*, 2005; *Hallberg and Gnanadesikan*, 2006]. In both past and future global climate evolution, variations in the strength and character of the Southern Ocean's meridional overturning circulation take a central stage. These are driven by changes in the intensity of wind and buoyancy forcing and constitute an important teleconnection between the regional atmosphere and cryosphere and the global deep ocean.

[67] The response of the ocean circulation to the increasing circumpolar westerly winds caused by the increasing SAM has received significant recent attention [e.g., *Hall and Visbeck*, 2002; *Oke and England*, 2004]. For example, it has been argued that the increasing SAM may have led to a latitudinal shift and increase in transport of the ACC [*Fyfe and Saenko*, 2006], with potential consequences for water mass properties in the Southern Ocean. While there is good observational evidence that the ACC transport depends strongly on the SAM on time scales from days and weeks

[*Aoki*, 2002; *Hughes et al.*, 2003] to years [*Meredith et al.*, 2004], it has also been argued that the trend in winds is more likely to result in a trend in circumpolar eddy activity rather than one in ACC transport [*Meredith and Hogg*, 2006]. This is currently the subject of significant ongoing research effort.

[68] In present views of the three-dimensional ocean circulation, the rate of upwelling of Circumpolar Deep Water in the Southern Ocean is enhanced in persistent high SAM index states and reduced in low ones [e.g., *Toggweiler et al.*, 2006; *Hallberg and Gnanadesikan*, 2006]. There is some modeling evidence for this correspondence carrying over to the rate of formation and northward export of Subantarctic Mode Water at the northern edge of the ACC and Antarctic Intermediate Water at the Polar Front [*Rintoul and England*, 2002; *Oke and England*, 2004; *Hallberg and Gnanadesikan*, 2006]. Furthermore, it has been suggested that SAM-related changes in the wind stress curl north and south of the ACC drive variability in the strength of the Southern Hemisphere subtropical gyre [*Cai et al.*, 2005] and the subpolar gyre [*Fahrbach et al.*, 2004]. In the Weddell gyre, these changes may lead to the episodic formation of large open ocean polynyas and associated onset of deep convection, a mode of Southern Ocean ventilation that is rare in the modern ocean but that may have been prevalent in colder global climatic states [*Gordon et al.*, 2007]. Variability in the strength of the Weddell gyre has also been put forward as a driver of complex changes in the export of Antarctic Bottom Water [*Meredith et al.*, 2001]. Without more data, the extent to which any of these circulation changes are occurring remains unclear, as does their relationship to the observed decadal-scale variability in ocean properties.

2.6. Changes in Sea Ice

[69] Sediment and ice-rafted debris from deep sea cores are interpreted to suggest that during the last glacial maximum a sea ice cover persisted for 6–7 months of the year as far north as 56.4°S at longitude 145.3°E [*Armand and Leventer*, 2003]. This is 6.3° farther north than the mean contemporary maximum ice edge location determined from satellite data at the same longitude [*Worby and Comiso*, 2004]. The maximum northerly ice extent at this location during the satellite era (since 1972) was 59.8°S.

[70] Since the days of the earliest explorers, vessels have kept written logs of their encounters with sea ice around Antarctica. However, these logs are sparse and usually only reported the location of the ice edge. Cook, the first to circumnavigate Antarctica in 1777, frequently reported the presence of sea ice as he tried to push south toward the continent, as did Bellingshausen during his exploration in 1831. The data from these log books provided the first in situ observations of Antarctic sea ice used for climate research, when *Parkinson* [1990] compared them with the location of the 1973–1976 ice edge derived from satellite data. Some significant differences between the data sets were observed, but Parkinson found no compelling evidence for the present-day ice edge being much different to that of 200 years ago.

[71] From the 1920s to 1930s, the UK Discovery Committee undertook a series of cruises to the Southern Ocean using the ships *Discovery, Discovery II*, and *William Scoresby.* The purpose was to investigate oceanographic properties and plankton distribution in relation to whale conservation, with the results published by the UK government in a long series of *Discovery Reports*. During this period, direct observations of sea ice extent were made when the ship encountered the pack ice, and from these occasional observations, *Mackintosh and Herdman* [1940] compiled a circumpolar map of the monthly variation of the average position of the ice edge. *Mackintosh* [1972] later updated these analyses with additional observations. Using whale catch data as a proxy, rather than direct observations of the ice edge location, *de la Mare* [1997] reported that there had been a 25% decline in summer (January) Antarctic sea ice between the mid-1950s and early 1970s. This result is in agreement with an analysis of MSA concentration in a coastal ice core from Law Dome, Antarctica [*Curran et al.*, 2003], which shows a strong correlation with ice extent in the region 80–140°E. The explanation for this relationship relies on the fact that MSA is produced from dimethyl sulphide, which, in turn, is produced from sea ice algae. The hypothesis is that in years of greater sea ice extent there will be more sea ice algae and therefore a greater concentration of MSA in precipitation along the coast of Antarctica. However, the strong correlation reported off East Antarctica has not been replicated in other cores, predominantly around the Antarctic Peninsula region, and further investigations are underway.

[72] *Worby and Comiso* [2004] also showed that modern ship-based observations of ice edge position are well correlated with satellite-based data for the ice growth (March–October) period but subject to a consistent offset of between 1 and 2° of latitude during the summer months. This is due to the fact that satellite passive microwave instruments are not able to detect the diffuse, saturated ice edge conditions typical in summer. *Ackley et al.* [2003] also showed that when the offset (between satellite estimates and ship estimates of edge location) is applied, there is good agreement between the range of modern (1979 onward) satellite-based ice edge positions and the ship-based ice edges observed in the 1920s and 1930s for circumpolar mean latitude extent; for example, the 25% decline reported by *de la Mare* [1997] was not confirmed.

[73] Evidence of a regional decrease in ice extent is apparent in the Weddell Sea/Antarctic Peninsula area. *Thompson and Solomon* [2002] attribute some of this change to an air temperature-driven ice retreat effect, all within the period 1969–1998, caused by a shift in the SAM, albeit primarily a change in summer, verified by instrumental temperature records and the changes in ice shelves in that region. In the modern era, continued decreases in this region are balanced by an increase in extent elsewhere, primarily in the Ross Sea [*Gloersen et al.*, 1992], leading to little observed change in the modern era for circumpolar mean

extent. While *Zwally et al.* [2002a, 2002b] suggest no statistically significant change in sea ice extent over recent decades, *Parkinson* [2004] shows a trend in the reduction of the length of ice season for the Antarctic Peninsula region of between 1 and 6 days for the period 1979–2002. This is consistent with the observed temperature increases reported in this region [e.g., *Vaughan et al.*, 2003]. The same study shows an increase in the length of the ice season around much of the rest of Antarctica, with the exception of some regions of the outer pack ice zone around East Antarctica. *Turner et al.* [2003] note exceptional changes in sea ice in the Bellingshausen Sea.

[74] In terms of sea ice thickness, there is still no single measurement technique that provides global, daily coverage, in the same way that passive microwave data provide ice concentration and extent. Consequently, our knowledge of Antarctic sea ice thickness is limited to a compilation of point measurements using a range of techniques from in situ sampling and ship-based observations, to data from upward looking sonars, and to electromagnetic soundings from aircraft-based and ship-based instruments. The SCAR Antarctic Sea Ice Processes and Climate project has been instrumental in compiling the available field data from ships operating in the Antarctic sea ice zone since 1980, and has released a compilation, from 89 voyages, of data that show regional and seasonal variability in the thickness distribution of the Antarctic sea ice zone. However, monitoring of Antarctic sea ice thickness is currently not possible, and important changes in the thickness of Antarctic sea ice may currently be going unnoticed. Ice core researchers are in the process of developing proxies for sea ice, a critical component in the climate system, through studies of sulfur compounds such as $SO_4^=$ and MSA [*Welch et al.*, 1993; *Curran et al.*, 2003; *Dixon et al.*, 2005]. ENSO–sea ice connections are investigated over the last ~500 years utilizing South Pole ice core MSA concentrations as a proxy revealing that in general, increased sea ice extent is associated with a higher frequency of El Niño events, while decreased sea ice corresponds to lower El Niño frequency [*Meyerson et al.*, 2002].

2.7. Mass Balance Changes in the Antarctic Ice Sheet, Antarctic Peninsula, Sub-Antarctic Islands, Southern South America, and New Zealand

[75] The mass balance, or mass budget, of a glacier or ice sheet is defined as the difference between (1) mass input (by net snow/ice accumulation at the surface caused by precipitation, drifting snow, and solid deposition from water vapor, or by subsurface accumulation caused by superimposed ice, where the base of the ice goes from wet to frozen, or by basal accretion in the case of ice shelves) and (2) mass loss (by basal melting, surface sublimation and melting, and ice calving). Balance estimates can be prepared in a number of different ways and with varying levels of complexity and confidence. For small glaciers, field measurement of net annual balance at surface stakes is a straightforward and reliable method [*Paterson*, 1994]. However, for larger glaciers, including ice sheets and ice shelves, widely distributed field measurements are often logistically impractical, and satellite remote sensing methods assume a greater importance [*ISMASS Committee*, 2004].

[76] There are, broadly, three different approaches to obtaining ice sheet mass balance estimates, a topic extensively reviewed by *Eisen et al.* [2008]. Mass flux methods are appropriate for catchment-scale studies [e.g., *Whillans and Bindschadler*, 1988]. The method entails computing the difference between accumulation rate in a catchment and the depth-averaged ice flux through a gate, often defined as the grounding line. Catchment accumulation rates are usually based on interpolation between isolated firn core measurements, best accomplished using ground-penetrating radar profiling [*Richardson and Holmlund*, 1999; *Frezzotti et al.*, 2004; *Spikes et al.*, 2004], or satellite microwave radiometry [*Vaughan et al.*, 1999; *Giovinetto and Zwally*, 2000; *Arthern et al.*, 2006], or regional atmospheric climate models [*Van den Broeke et al.*, 2006; *Van de Berg et al.*, 2006]. Outgoing flux is obtained from the product of ice velocity and ice thickness. Satellite remote sensing methods such as feature tracking [*Scambos et al.*, 1992] and radar interferometry [*Goldstein et al.*, 1993; *Bamber et al.*, 2000] provide the most efficient means of mapping surface velocity fields. Flux calculations are useful but subject to large uncertainties in the interpolation of accumulation rates and in the parameterization of depth-averaged velocity. A variant of the flux method is the submergence velocity technique [*Hamilton et al.*, 2005], in which point measurements of accumulation rate from cores and vertical velocity from GPS surveys are compared to yield the rate of thickness change. While the results of this method are precise, its shortcoming is that the results apply only to the locations where the measurements are made.

[77] Geodetic methods of assessing mass balance are increasingly used. One such technique entails repeat measurements of surface elevations by airborne altimeters [e.g., *Spikes et al.*, 2003] and spaceborne altimeters [e.g., *Zwally et al.*, 2005]. Radar altimeters on board Seasat, Geosat, ERS-1, and ERS-2 have been measuring ice sheet surface elevations since 1978. The relatively long record of radar altimeter observations allows interannual variability in surface elevations to be identified and, to some extent, removed from resulting mass balance estimates [*Li and Davis*, 2006; *Wingham et al.*, 2006]. In recent years, similar measurements have been made by a laser altimeter on board ICESat [*Zwally et al.*, 2002a]. Laser altimetry has the advantage of surveying smaller spatial footprints (~60 m or less) than radar altimetry (~20 km). Altimetric methods, regardless of instrument, yield volume changes with time. The conversion of these estimates to rates of mass change is complicated by poorly quantified processes, such as crustal isostatic adjustment and the variable rate of firn compaction, and, in the case of radar sensors, uncertainty in the depth from which the radar signal is being returned. The role of firn compaction has been analyzed by means of a physically based model [*Li et al.*, 2007] and also by means of regional atmospheric climate models [*Van den Broeke et al.*, 2006]. Spaceborne gravimetry is a relatively new technique that uses tandem satellites to measure spatial and temporal

TABLE 1. Summary of Mass Balance Estimates for the Antarctic Ice Sheet[a]

Mass Balance (Gt a^{-1})	Period	Reference	Observations	Method
+45 ± 7	1992–2003	*Davis et al.* [2005]	~70% of EA and WA; AP not considered	satellite radar altimetry
−24 ± 25	1995–2000	*Rignot and Thomas* [2002]	average for EA and WA; AP not considered	mass flux method
−30 ± 12	1992–2001	*Zwally et al.* [2005]	~80% of Antarctic coverage, including AP; rest of the ice sheet interpolated	satellite radar altimetry
+27 ± 29	1992–2003	*Wingham et al.* [2006]	~70% of Antarctic coverage, including the AP	satellite radar altimetry
−139 ± 73	2002–2005	*Velicogna and Wahr* [2006]	includes AP	GRACE satellite gravity data
−196 ± 92	2006	*Rignot et al.* [2008]	~85% of Antarctica's coastline, including the AP	satellite radar interferometry and regional climate modeling

[a]EA, East Antarctica; WA, West Antarctica; AP, Antarctic Peninsula.

variations in the Earth's gravity field. These changes can be inverted to an ice mass change, on the assumption that the gravity change is the result of a change in mass on the Earth's surface. Several mass balance estimates have been derived using this technique [e.g., *Chen et al.*, 2006; *Ramillien et al.*, 2006; *Velicogna and Wahr*, 2006], but the results are sensitive to processing strategies and the treatment of sources of error.

[78] For the grounded interior ice sheet of Antarctica, losses by surface melting, sublimation, and basal melting are negligible, as are mass inputs by superimposed ice and deposition from water vapor. However, this is not the case for the Antarctic Peninsula and Southern Ocean glaciers, where all components can potentially contribute to the mass balance. The following is a brief summary of the most recent mass balance calculations for Antarctic and Southern Ocean glaciers.

2.7.1. Recent Mass Balance for the Antarctic Ice Sheet

[79] Taking a conservative approach, it appears that there is still no consensus as to the value of Antarctica's mass balance or even its sign (Table 1). However, recent results of *Rignot et al.* [2008] show a near-zero mass balance for East Antarctica of −4 ± 61 Gt/a, which suggests that negative mass balances in coastal areas are larger than any mass increase in the interior. The lack of consensus is largely a function of the differing methods of mass balance assessment described in section 2.7. While the continent's mass balance as a whole remains uncertain, some progress has been made in estimating the mass balance for individual components of the ice sheet. The East Antarctic Ice Sheet appears to be acting as a net sink for global sea level [*Davis et al.*, 2005; *Zwally et al.*, 2005; *Wingham et al.*, 2006], mostly because of ice sheet growth due to a hypothesized modest increase in snowfall. If this is the case, the increase in snowfall must be a very recent event and limited to the time period of the altimeter surveys (1992–2003) because other studies [e.g., *Van de Berg et al.*, 2005; *Van den Broeke et al.*, 2006; *Monaghan et al.*, 2006a, 2006b; *Rignot et al.*, 2008] do not show any substantial trend in snow accumulation over the last few decades. Only two regions of East Antarctica (the Totten and Cook glaciers) have significantly negative mass balance conditions [*Rignot*, 2006; *Shepherd and Wingham*, 2007]. These are regions of enhanced ice flow, and the negative balances might be an ice dynamics response to the removal of buttressing ice shelves at some point in the past [*Rignot and Thomas*, 2002; *Zwally et al.*, 2005].

[80] The smaller West Antarctic ice sheet is generally understood to be losing mass, primarily as a result of an ice dynamics perturbation of glaciers draining into the Amundsen Sea [*Shepherd et al.*, 2002; *Thomas et al.*, 2004; *Holt et al.*, 2006; *Vaughan et al.*, 2006]. Recent data show an ice sheet loss increase of 59% between 1996 and 2006, which amounts to 132 ± 60 Gt/a in 2006 [*Rignot et al.*, 2008]. The two largest glaciers in this basin, Pine Island and Thwaites glaciers, have retreated, accelerated, and thinned since the 1990s [*Shepherd et al.*, 2002]. An inflow of warmer ocean waters is hypothesized as the trigger for the observed changes in ice dynamics [*Payne et al.*, 2004], pointing to a delicate relationship between Antarctic glacier grounding lines and ocean conditions. The negative balance of the Amundsen Sea basin is offset to some extent by steady state conditions in the ice sheet interior [e.g., *Hamilton et al.*, 2005; *Zwally et al.*, 2005], growth in the Siple Coast catchment [*Joughin and Tulaczyk*, 2002] largely due to the shutdown of the Kamb Ice Stream, and positive mass balance of the ice stream catchments that drain into the Filchner-Ronne ice shelf [*Joughin and Bamber*, 2005].

[81] One of the largest sources of uncertainty in the determination of Antarctic mass balance, and its potential role in sea level, is the surface mass balance [*Vaughan et al.*, 1999]. Snowfall is the dominant surface balance term at regional and larger scales, accounting for about 90% of the surface mass balance [*Bromwich*, 1988]. Comprehensive studies of precipitation characteristics over Antarctica are given by *Bromwich* [1988], *Turner et al.* [1999], *Genthon and Krinner* [2001], and *Bromwich et al.* [2004b].

[82] Because of the paucity of spatially and temporally coherent snowfall observations, satellites and numerical atmospheric models are most often used to assess the variability of the Antarctic surface mass balance at continental scales. Results from studies employing these techniques indicate that agreement has yet to be reached as to the long-term distribution of annual surface mass balance over Antarctica. Current estimates range from +119 to +197 mm/a [*Van de Berg et al.*, 2005; *Ohmura et al.*, 1996]. Such discrepancies lead to great uncertainty in estimates of the contribution of the Antarctic ice sheets to sea level rise [e.g., *Vaughan*, 2005]. For example, *Van den Broeke et al.* [2006] show that differences in snowfall accumulation estimates in the basins where the Pine Island and Thwaites glaciers are

accelerating and thinning [*Thomas et al.*, 2004] cause the calculations of sea level contributions from these regions to vary by a factor of two. Clearly, it is imperative to drive down uncertainty in future work.

[83] Numerical models provide a methodology for assessing the temporal variability of Antarctic surface mass balance. A series of the latest studies employing global and regional atmospheric model records to assess changes in Antarctic surface mass balance indicate that averaged over the continent, no statistically significant change has occurred since ~1980, although there are regions of both positive and negative trends [*Van de Berg et al.*, 2005; *Monaghan et al.*, 2006a; *Van den Broeke et al.*, 2006]. More recently, a 50-year record of Antarctic snowfall, constructed by synthesizing atmospheric model output with snow accumulation observations primarily from ITASE ice cores, showed that there has been no statistically significant change in snowfall over an even longer period extending back to the International Geophysical Year of 1957–1958 [*Monaghan et al.*, 2006b]. An additional finding of that study was that the interannual and interdecadal variability of Antarctic snow accumulation is so large that it may be another decade before short-term trends in total ice sheet mass balance from satellite altimetry and gravity measurements can be distinguished from the noise [e.g., *Wingham et al.*, 2006].

[84] Synoptic weather patterns control the distribution of snowfall at larger scales, but at subregional scales, wind-driven redistribution is often of first-order importance [*Gow and Rowland*, 1965; *Whillans*, 1975; *Frezzotti et al.*, 2002; *Ekaykin et al.*, 2002]. ITASE research reveals high variability in surface mass balance such that single cores, stakes, and snow pits do not always represent the geographical and environmental characteristics of a local region [*Richardson and Holmlund*, 1999; *Frezzotti et al.*, 2004; *Hamilton*, 2004; *Spikes et al.*, 2004]. For example, *Frezzotti et al.* [2004] show that spatial surface mass balance variability at subkilometer scales (as is typically represented in ice cores) overwhelms temporal variability at the century scale for a low-accumulation site in East Antarctica. Emerging data collected by ITASE and associated deep ice core projects (e.g., EPICA) reveal systematic biases in long-term estimates of surface mass balance compared to previous compilations. The biases are presumably related to the small-scale spatial variability [*Oerter et al.*, 1999; *Frezzotti et al.*, 2004; *Magand et al.*, 2004; *Rotschky et al.*, 2004]. The extensive use, along ITASE traverses, of new techniques like geolocated ground penetrating radar profiling (GPR) integrated with ice core data provides detailed information on surface mass balance [*Richardson and Holmlund*, 1999; *Urbini et al.*, 2001; *Arcone et al.*, 2004; *Rotschky et al.*, 2004]. At some sites, stake farm and ice core accumulation rates differ significantly, but isochronal layers in firn, detected with GPR, correlate well with ice core chronologies [*Frezzotti et al.*, 2004]. Several GPR layers within the upper 100 m of the surface were surveyed over continuous traverses of ~5000 km and can be used as historical benchmarks to study past accumulation rates [*Spikes et al.*, 2004].

2.7.2. Recent Mass Balance for the Antarctic Peninsula

[85] The ice shelves of the Antarctic Peninsula, some of which are several thousand years old, are not just retreating but are also undergoing rapid collapse in response to regional warming [*Vaughan et al.*, 2003]. As an example, the massive collapse in 2002 of Larsen B ice shelf is unprecedented during the last 10,000 years [*Domack et al.*, 2005]. This collapse was preconditioned by structural weakening related to retreat sometime in the 20 years preceding the event [*Glasser and Scambos*, 2008]. Because they are already floating, the contribution of ice shelves to sea level rise is negligible. However, the removal of buttressing ice shelves has resulted in significant flow acceleration of several inland glaciers [*De Angelis and Skvarca*, 2003; *Scambos et al.*, 2004; *Rignot et al.*, 2004], with the potential to contribute to a rise in sea level. Ice shelves located farther south along the peninsula may collapse in the near future if warming continues. This may be the case for Larsen C, where significant thinning has already been reported [*Shepherd et al.*, 2003]. Widespread glacier recession is reported for 87% of the Antarctic Peninsula marine glacier fronts on the basis of analysis of the behavior of 244 glaciers over the past 61 years [*Cook et al.*, 2005].

[86] Associated ice thinning has been reported at low altitudes in glaciers of the Antarctic Peninsula [*Morris and Mulvaney*, 1995; *Smith et al.*, 1998]. At higher-elevation sites there is some evidence for accumulation increase. This is the case for Dyer Plateau located at an altitude of 2000 m at 70.7°S, where an accumulation increase of 17% has been found in the period 1790–1990 on the basis of ice core data from a depth of 235 m and ice flow modeling [*Raymond et al.*, 1996]. A doubling in snow accumulation during the period 1855–2006 has been reported at a 136 m deep ice core site located at an altitude of 1130 m in the southwestern Antarctic Peninsula at 73.6°S [*Thomas et al.*, 2008], which has been linked to a positive shift of the Southern Annular Mode. Recent satellite interferometry data for more than 300 glaciers on the west coast of the Antarctic Peninsula show that summer flow velocities increased significantly by 12% from 1992 to 2005 [*Pritchard and Vaughan*, 2007], which is explained as a dynamic response to frontal thinning. The thinning and associated flow increase has led to an accelerated mass loss of 140% in the period 1996–2006, with a value of 60 ± 46 Gt/a in 2006 according to *Rignot et al.* [2008]. The enhanced melting near the coast has resulted in an increased risk for glacier travel and snow runway operations [*Rivera et al.*, 2005].

[87] Ice shelves on the western side of the Antarctic Peninsula, such as the Wordie and Wilkins, have also been retreating in response to the rise of air temperatures described in section 2.1, with signs of incipient collapse recently reported for Wilkins. In contrast, an increase in northerly winds in this region has resulted in a rise in the number of precipitation events, leading to greater snowfall and therefore greater accumulation [*Turner et al.*, 2005b].

2.7.3. Recent Mass Balance for New Zealand, Southern South America, and Southern Ocean Island Glaciers

[88] Mountain glaciers and ice caps distributed around the Southern Ocean show a strong wasting during the last century, with increasing rates during the last few decades coincident with hemispheric warming.

[89] New Zealand has 3155 glaciers exceeding 0.01 km^2 with a total area of 1159 km^2 [*Chinn*, 1989], mostly concentrated in the Southern Alps and also on Mount Ruapehu, North Island. Most of these glaciers have been receding during the past century, together with an overall rise in the equilibrium line altitudes [*Chinn*, 1995]. For example, the Tasman Glacier has been thinning at a rate of 1–2 m/a during the past century [*Kirkbride and Warren*, 1999]. A few maritime glaciers such as the Franz Josef Glacier advanced during the period from the early 1980s until 2000 because of increased local and regional precipitation [*Chinn et al.*, 2005], although in the last few years the trend has reversed, presumably because of decreased precipitation in combination with enhanced melting in the ablation areas.

[90] The region in South America located between 40 and 56°S, including Patagonia and Tierra del Fuego, accounts for ~70% of all Andean glaciers, representing more than 20,000 km^2 of ice shared between Chile and Argentina. The main areas in this region are the Northern Patagonian Ice Field with an area of 3953 km^2 [*Rivera et al.*, 2007], the Southern Patagonian Ice Field (SPI) with an area of 13,362 km^2 [*De Angelis et al.*, 2007], and Cordillera Darwin with an estimated area of 2000 km^2 [*Casassa*, 1995]. These ice fields consist mainly of temperate ice and contain the largest glaciers in the Southern Hemisphere outside of Antarctica. They are potentially valuable sources of present and past environmental information from the midlatitudes, providing a link between the southern tropical and equatorial regions and Antarctica. The ice fields and glaciers in Patagonia and Tierra del Fuego show a generalized and accelerated retreat and thinning in response to regional warming [*Casassa et al.*, 2007]. There are a few cases of advancing glaciers such as Moreno Glacier [*Skvarca and Naruse*, 1997] and Pío XI Glacier on the SPI [*Rivera et al.*, 1997] and some south facing glaciers of Cordillera Darwin that flow to the Beagle Channel [*Holmlund and Fuenzalida*, 1995], probably because of an increase in local precipitation and/or ice dynamic effects. There is recent evidence of a small but significant thickening in the accumulation area of Gran Campo Nevado at 53°S [*Möller et al.*, 2007], which is probably driven by increased precipitation due to enhanced westerly circulation. Retreat during the past century reached maximum values of 15 km for O'Higgins Glacier in the SPI [*Casassa et al.*, 1997], with maximum retreat rates of 787 m/a and an area decrease of 2.75 km^2/a for Marinelli Glacier in Cordillera Darwin [*Porter and Santana*, 2003] and a record thinning of 30 m/a detected locally in the SPI [*Rignot et al.*, 2003]. Although southern South American glaciers store a total equivalent sea level rise of only a few centimeters [*Rivera et al.*, 2002], which represents much less than 10% of the total volume of mountain glaciers of the world, they are presently contributing more than 10% of the total global sea level rise from mountain glaciers [*Kaser et al.*, 2006]. Although human population in the area is limited, southern South American glaciers are important in terms of water resources in the region, including generation of hydroelectric energy. There have also been several reported cases of glacial lake outburst floods originating from enhanced melting and ice avalanches and also mudflows related to volcanic eruptions, although no casualties have been reported.

[91] Elsewhere in the Southern Hemisphere, south of 49°S and apart from Antarctica, there are several ice masses located on Southern Ocean islands. A precise glacier inventory is still lacking, but an estimate including sub-Antarctic islands geographically closer to the Antarctic Peninsula shows a total glacier area of about 7100 km^2 [*Haeberli et al.*, 1989]. These islands include South Georgia, Kerguelen, Heard, Bouvet, Macquarie, and the South Sandwich Islands. All glaciers show a general trend for recession [e.g., *Colhoun and Goede*, 1974], although studies are very limited, with mass balance data only available for 1958 for two glaciers in South Georgia. For example, in 1947 the glaciers of Heard Island covered 288 km^2 or 79% of the island. By 1988 this had decreased by 11% to 257 km^2, with about half of this change thought to have occurred during the 1980s [*Allison and Keage*, 1986; *Truffer et al.*, 2001]. Further and increased retreat occurred during the 1990s. These changes are producing profound changes on the landscapes of these islands, which are home to unique ecosystems. Glacier retreat and related landscape and climate change on South Georgia has been accompanied by a population explosion of rats that were introduced to the island 200 years ago by whaling vessels. The rats are rapidly exterminating colonies of local seabirds.

2.8. Summary of Main Climate Events of the Last 50 Years

[92] A summary of the main climatic events of the last 50 years developed from section 2 appears in Figure 17 to set the context for discussing the future of the Antarctic and Southern Ocean climate system in section 3 of this review.

3. NEXT 100 YEARS

[93] The main tools available for predicting how the climate of the Earth will evolve in the future are coupled atmosphere-ocean climate models. Many of the current generation of climate models struggle to represent key aspects of polar climate such as sea ice and near-surface temperature [*Carril et al.*, 2005]. However, the success of many of the models included in the 2007 IPCC Fourth Assessment Report (AR4) at qualitatively reproducing the observed near-surface warming over the Antarctic Peninsula over the last 50 years indicates an improvement of the representation of regional change compared to the previous IPCC Third Assessment Report models [*Lynch et al.*, 2006]. In some cases, there is a large probability that natural

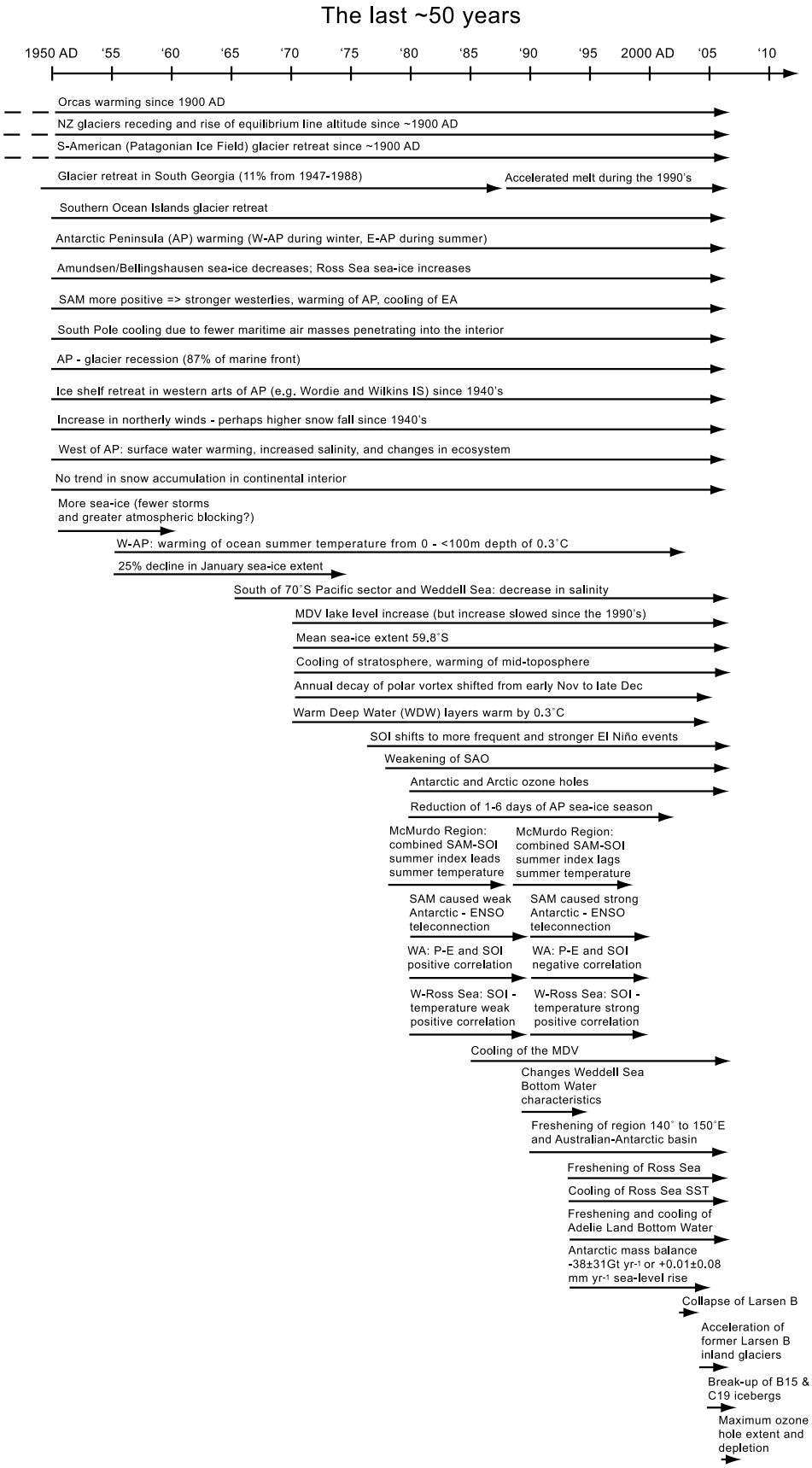

Figure 17. Main climatic events of the last 50 years: the Antarctic context. AP, Antarctic Peninsula; EA, East Antarctica; WA, West Antarctica; MDV, McMurdo Dry Valleys; SOI, Southern Oscillation Index; SAM, Southern Annular Mode; SAO Semiannual Oscillation; P-E, precipitation minus evaporation; SST, sea surface temperature; ENSO, El Niño–Southern Oscillation.

variability contributes to differences between the models and observations. For instance, large variability of sea ice extent means that the negative trend of sea ice extent simulated by the AR4 models is not significantly different from the slight increase suggested from observations over the last 20 years [*Arzel et al.*, 2006].

[94] Future projections from climate models contain two key sources of uncertainty: uncertainty associated with factors that influence climate change and uncertainty associated with model error. The range of scenarios produced for the IPCC is designed to span the likely range of expected future outcomes, and since the probability of a given scenario being realized cannot practically be quantified, all scenarios are considered equally likely.

[95] A popular way to assess modeling uncertainty is to conduct ensemble experiments. This can be particularly valuable for regional climate projections, where the predicted uncertainty is often larger than the global average. Two types of modeling uncertainty that can be assessed using ensemble experiments are process uncertainty and structural modeling uncertainty [*Murphy et al.*, 2004]. Process uncertainty can be assessed by running perturbed physics ensembles, which are multiple versions of a given model with parameters that control important physical processes perturbed across their range of observational uncertainty. Structural modeling uncertainty takes into account factors such as the use of different parameterizations and model resolution and is most commonly done by composing an ensemble of output from different models from a variety of modeling centers. For seasonal forecasting the multimodel ensemble method is suggested to be more robust than a single model ensemble [*Hagedorn et al.*, 2005]. Although climate model projections cannot be verified as rigorously as seasonal predictions, the significant structural modeling uncertainty that is known to exist (e.g., cloud and radiances parameterizations over Antarctica [see *Hines et al.*, 2004]) motivates the use of multimodel ensembles such as the output provided as part of the IPCC effort.

[96] The *IPCC* [2007] estimated that over the period 1990–2100, globally averaged surface temperatures would increase by 1.8° to 6.4°C. This is for 35 greenhouse gas emission scenarios and a number of climate models. Weighted averages of projections of Antarctic climate change over the 21st century were derived by *Bracegirdle et al.* [2008] from 19 of the 24 models that were submitted for the Intergovernmental Panel on Climate Change AR4 and that used the A1B scenario, which predicts an approximate doubling of CO_2 in the atmosphere over the next century and for which predictions for 2100 are about the middle of the range of the various scenarios in terms of effect on temperature. The weighted average of the model runs predicts that the annual mean atmospheric surface temperatures in the Antarctic sea ice zone would increase by 0.24° ± 0.10°C/decade [*Bracegirdle et al.*, 2008]. In contrast to the current near-surface temperature trends over the Antarctic continent, which show a strong warming over the Antarctic Peninsula and little significant change elsewhere, the projected pattern of temperature change shows strong warming over the high interior of 0.34° ± 0.10°C/decade, indicating that the present warming will spread beyond the peninsula. The simulated strong warming over the continent may cause weakening of the katabatic winds from the interior, especially in summer.

[97] On the basis of the reduction in emission of CFCs, the WMO predicts that Antarctic ozone values will return to pre-1980 levels around 2060–2075, which is about 10–25 years later than previously thought [*WMO*, 2006]. However, other factors influencing ozone production and destruction are difficult to forecast. Because stratospheric ozone is influenced by temperature and wind, stratospheric cooling induced by global warming can extend the time period over which polar stratospheric clouds form and therefore could lead to increased winter ozone depletion. In contrast, cooling of the upper stratosphere, above the zone of formation of polar stratospheric clouds, promotes decreased photochemical ozone destruction and hence leads to higher ozone values [*Fahey et al.*, 2007]. In addition, warmer surface temperatures lead to higher natural halogen emissions from the Earth's surface and hence accelerated destruction of ozone in the stratosphere. This could be further amplified through increased stratospheric water vapor, observed over the last 2 decades, serving as nuclei for polar stratospheric clouds [*Fahey et al.*, 2007]. Given that many recent studies [*Shindell and Schmidt*, 2004; *Arblaster and Meehl*, 2006] suggest that stratospheric ozone depletion has the strongest impact on the recent SAM trends, future Antarctic temperature trends, which are dependent upon the SAM [*Schneider et al.*, 2006], may depend strongly on future stratospheric ozone variability and/or recovery.

[98] With the higher tropospheric temperatures, there is expected to be an increase in snowfall over the Antarctic, with an increase of 25–50% being experienced in many areas. The weighted averages of the IPCC models [*Bracegirdle et al.*, 2008] suggest an increase of 25 ± 11% in annual accumulated snowfall over the grounded Antarctic ice sheet. However, there will not be a significant increase of melting, since near-surface temperatures will still mainly remain below freezing. This could lead to an increase of the frozen water mass stored in grounded ice sheets and contribute negatively to sea level rise [*Gregory and Huybrechts*, 2007]. However, other factors such as changes in ice dynamics should also be taken into account for a complete assessment of the mass balance.

[99] Most model predictions suggest that with increasing levels of greenhouse gases, the SAM may be in its positive phase for a greater length of time [*Miller et al.*, 2006]. This will result in lower atmospheric pressures over the Antarctic and higher pressures at midlatitudes. With the greater pressure gradient, there will be a further increase in the westerly winds over the Southern Ocean, and the center of the continent will be more isolated from maritime air masses. Without that isolation, the predicted warming of the interior might be expected to be higher. Changes of the SAM over the next 100 years are not projected to be as strong as they have been in the last 20 years, possibly

because of the less rapid recovery of stratospheric ozone [*Shindell and Schmidt*, 2004]. The response of the ocean to these changes is not clear but could include a strengthening of the ACC, increased eddy activity and poleward eddy heat flux, and possibly a shift in ACC location. Also possible is an intensification of overturning in the Southern Ocean, with potentially global consequences.

[100] The weighted average model analysis [*Bracegirdle et al.*, 2008] suggests that there will be an overall reduction of sea ice area of 33 ± 9% by 2100 around much of the Antarctic, accompanied by a decrease in sea ice concentrations [e.g., see *Arzel et al.*, 2006]. There will also be an increase of the amplitude of the seasonal cycle of sea ice extent. Radar and laser satellite altimetry are currently yielding promising results for measuring ice or snow surface elevation over Antarctic sea ice, which, through conversion algorithms, are used to estimate sea ice thickness. While the techniques are promising, a number of serious difficulties remain, including the variation in physical properties of the sea ice and snow cover within the footprint of the instruments, assumptions related to snow and ice density, and correction for the geoid.

[101] Any assessment of future change using the output from climate models must include a consideration of the ability of the models to represent the processes leading to the change of interest. One example of this is the Antarctic Peninsula warming, which has strong contributions from different mechanisms in different seasons [*Turner et al.*, 2005a; *Marshall*, 2007]. The summer warming to the east of the peninsula has been attributed to SAM changes, which are well represented in most climate models. The winter warming to the west of the peninsula has also been linked to SAM changes and to other processes as well, such as the retreat of sea ice (decreasing albedo) and changes to the frequency and intensity of El Niño events; replicating these processes accurately in models at the regional level represents a significant modeling challenge, and attributions of change should be treated with caution.

[102] Forecasting change in mass balance in response to climate change is needed as the basis for forecasting changes in sea level. The former is difficult because models of ice sheet decay do not yet take changes in ice dynamics adequately into consideration. The 2007 report of the Intergovernmental Panel on Climate Change forecast that sea level will rise by less than 1 m by 2100 in response to thermal expansion of the ocean and the melting of glaciers at midlatitudes and in polar regions [*IPCC*, 2007] (and see http://www.ipcc.ch/). That forecast is conservative because it does not take ice dynamics into consideration. The possibility that sea level may rise by up to 5 m by 2100, because of melting of parts of the Greenland and West Antarctic ice sheets, much as it did when temperatures last rose by 2°–3°C 125,000 years ago [*Jansen et al.*, 2007] has been raised [*Hansen*, 2007].

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