Scientific Committee on Antarctic Research

ANTARCTIC CLIMATE CHANGE AND **THE ENVIRONMENT**

A contribution to the International Polar Year 2007-2008

Edited by:

John Turner, *British Antarctic Survey, UK*
Robert Bindschadler, *National Aeronautics and Space Administration, USA*
Pete Convey, *British Antarctic Survey, UK*
Guido di Prisco, *Institute of Protein Biochemistry, Italy*
Eberhard Fahrbach, *Alfred Wegener Institute, Germany*
Julian Gutt, *Alfred Wegener Institute, Germany*
Dominic Hodgson, *British Antarctic Survey, UK*
Paul Mayewski, *Climate Change Institute, University of Maine, USA*
Colin Summerhayes, *Scientific Committee on Antarctic Research*

Antarctic Climate Change and the Environment

This volume provides a comprehensive, up-to-date account of how the physical and biological environment of the Antarctic continent and Southern Ocean has changed from Deep Time until the present day. It also considers how the Antarctic environment may change over the next century in a world where greenhouse gas concentrations are much higher than occurred over the last few centuries. The Antarctic is a highly coupled system with non-linear interactions between the atmosphere, ocean, ice and biota, along with complex links to the rest of the Earth system. In preparing this volume our approach has been highly cross-disciplinary, with the goal of reflecting the importance of the continent in global issues, such as sea level rise, the separation of natural climate variability from anthropogenic influences, food stocks, biodiversity and carbon uptake by the ocean. One hundred experts in Antarctic science have contributed and drafts of the manuscript were reviewed by over 200 scientists. We hope that it will be of value to all scientists with an interest in the Antarctic continent and the Southern Ocean, policy makers and those concerned with the deployment of observing systems and the development of climate models.

JOHN TURNER is a research scientist at the British Antarctic Survey in Cambridge, UK where he leads a project investigating recent Antarctic climate change and how it may change over the next century. He has had a long involvement with SCAR and was the Chief Officer of the Physical Sciences Standing Scientific Group from 2002 to 2006 and chaired the steering committee of the SCAR programme on Antarctica and the Global Climate System from 2005 to 2008. He is the co-author of 'Antarctic Meteorology and Climatology' and 'Polar Lows: Mesoscale Weather Systems in the Polar Regions', both of which were published by Cambridge University Press. He was awarded the International Journal of Climatology Prize of the Royal Meteorological Society in 2005.

ROBERT BINDSCHADLER is Chief Scientist of NASA's Hydrospheric and Biospheric Sciences Laboratory, a Senior Fellow of the Goddard Space Flight Center, a Fellow of the American Geophysical Union and a past President of the International Glaciological Society. He maintains an active interest in glaciers and ice sheets and has led 15 Antarctic field expeditions to study dynamics of the West Antarctic ice sheet. During his 29 years at NASA, he has developed numerous unique applications of remote sensing data for glaciological research including measuring ice velocity and elevation using both visible and radar imagery, monitoring melt of the ice sheet by microwave emissions, and detecting changes in ice-sheet volume by repeat space-borne radar altimetry. He has testified before Congress, briefed the U.S. Vice President, published over 140 scientific papers, including numerous review articles and is often quoted commenting on glaciological impacts of the climate on the world's ice sheets and glaciers.

PETER CONVEY is a research scientist at the British Antarctic Survey in Cambridge, UK where he is the senior terrestrial ecologist in a wide-ranging programme investigating ecosystem structure and function in the Antarctic, and how this responds to environmental variability and change. He has wide research interests ranging from genomics and ecophysiology to historical biogeography. He is Co-Chair of the current SCAR Science Research Programme 'Evolution and Biodiversity in Antarctica' (2005-2013), and was a Steering Committee member of its predecessor programme 'Regional Sensitivity to Climate Change in Antarctica' (2000-2005). He is a co-editor of 'Trends in Antarctic Terrestrial and Limnetic Ecosystems: Antarctica as a Global Indicator', published by Springer.

GUIDO DI PRISCO is a Director of Research of the Italian National Research Council (CNR). His expertise lies in Biochemistry and Molecular Biology, and his particular interests in molecular adaptation and evolution in polar oceans and the impact of climate change. He has undertaken 13 expeditions at Mario Zucchelli Station in the Ross Sea, and 6 at Palmer Station on the Antarctic Peninsula, and participated in 3 research cruises in the Antarctic/sub-Antarctic. A member of SCAR's Standing Scientific Group on Life Sciences, he is Co-Chair of SCAR's research programme on Evolution and Biodiversity in the Antarctic, and helped to steer its predecessor biological programmes. Guido has been Guest Editor of 5 books in his subject area, and of a number of Special Issues of high impact journals in his field. He is author or co-author of over 280 peer-reviewed articles.

EBERHARD FAHRBACH is a research scientist at the Alfred-Wegener-Institut für Polar-und Meeresforschung and head of the Observational Oceanography section. He has the position of scientific coordinator of RV "Polarstern". He is working on water masses and circulation in Polar oceans. He has participated in 18 cruises to the Antarctic and the Arctic most of them as chief scientist. He was and is a member of a variety of national and international steering groups and committees. In 2007 he was awarded the Georg Wüst Prize by the Deutsche Gesellschaft für Meeresforschung.

JULIAN GUTT is a research scientist at the Alfred Wegener Institute for Polar and Marine Research in Bremerhaven, Germany, and has been involved in polar research for 25 years. His interests focus on ecosystem responses to climate-induced disturbance. In 2006/07 he was the principal investigator of an expedition to the area where the Larsen A and B ice shelves had recently disintegrated, to examine the way in which these areas were being recolonised by new organisms. He is a member of the steering committee of the "Census of Antarctic Marine Life". Together with John Turner, he chairs SCAR's Action Group on "Prediction of Changes in the Physical and Biological Environments of the Antarctic". Besides many scientific articles he has written several chapters for popular books on marine biodiversity and environmental change.

DOMINIC HODGSON is a research scientist at the British Antarctic Survey in Cambridge, UK. He is a Quaternary Scientist with a particular interest in high latitude environmental change. His current research involves using lake sediments, as well as polar marine sediment and ice cores, to investigate how different parts of the Earth System have interacted to produce the large climate changes that occurred naturally in the past. Recent career highlights include developing new relative sea level curves for Antarctica, reconstructing the Holocene history of Antarctic Peninsula ice shelves, documenting climate history from lakes that have survived through glacial cycles, and working on a recently emerged subglacial lake. Dr Hodgson has led research in many regions of Antarctica including east Antarctica, the Antarctic Peninsula, Maritime Antarctic and Subantarctic Islands.

PAUL ANDREW MAYEWSKI is Director and Professor of the Climate Change Institute at the University of Maine. He has led more than 50 research expeditions throughout polar and high mountain regions including leadership of the 25 institution Greenland Ice Sheet Project Two and the 21 nation International Trans Antarctic Scientific Expedition. He has authored more than 300 peer reviewed articles and appears regularly in public media venues discussing climate change. In 2006 he was awarded the SCAR Medal for Excellence in Antarctic Research.

COLIN SUMMERHAYES is a geologist with an extensive career in oceanography, whose most recent scientific research has been on how climate change affected ocean currents and oceanic productivity during the past 100,000 years. A former Director of the UK's Institute of Oceanographic Sciences Deacon Laboratory, now part of the National Oceanography Centre, Southampton, UK, he spent 7 years as Director of the Global Ocean Observing System Project for UNESCO's Intergovernmental Oceanographic Commission in Paris before joining SCAR as Executive Director in 2004. He is a co-author of 'Oceans 2020, Science, Trends and the Challenge of Sustainability, published by Island Press; of 'Oceanography: an Illustrated Guide, published by Manson Publishing; and of 'Upwelling in the Oceans' published by Wiley.

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Preface

To understand how planet Earth works we study it increasingly as a system – a collection of interdependent parts or spheres – the lithosphere, the hydrosphere, the cryosphere, the biosphere and the atmosphere. Understanding how these spheres are connected and how they interact improves our ability to forecast how one or a combination of them may change in response to external forcings caused for example by the advent of volcanic eruptions, solar variability or human activities. One of the remotest parts of the Earth system is Antarctica, a continent larger than either Australia or Europe. We will not be able to fully understand how the Earth system works without comprehensive knowledge of the physical, biological, chemical and geological processes taking place within and above Antarctica and its surrounding Southern Ocean. That is a huge challenge given that these processes take place among some of the remotest and harshest environments anywhere on the Earth's surface.

Currents and waves in the global ocean and the atmosphere ensure that Antarctica is affected by what happens elsewhere on the planet. Equally, ocean and atmospheric processes ensure that what happens in Antarctica may affect the rest of the world. It is the world's freezer.

Much has been achieved in acquiring knowledge of Antarctica's physical, biological, chemical and geological processes, especially since a network of permanent scientific stations was established for the first time on the continent during the International Geophysical Year of 1957-58. Many more results will emerge from the recently completed International Polar Year of 2007-2008. Nevertheless, the practicalities and expense involved in getting scientists to these remote and harsh places means that this region will remain under-sampled for years to come, constraining what can be achieved in the way of both understanding and forecasts.

What we set out to do in this volume is to review present understanding of the physical and chemical climate system of the Antarctic region, the way it varies through time, and the profound influence of that variation on life on land and in the ocean around the continent. We then use this as the basis for predicting, albeit crudely within the limits of the dearth of information compared with the other continents, what may happen in the future as greenhouse gases build up in the atmosphere and as the ozone hole starts to diminish. This volume should be taken as a companion to the Arctic Climate Impact Assessment published in 2005.

The work has been carried out under the editorial control of representatives of three of SCAR's five major scientific research programmes: Antarctica in the Global Climate System (AGCS), Antarctic Climate Evolution (ACE), and Evolution and Biodiversity in the Antarctic (EBA). The process began in 2005 when the SCAR Executive Committee meeting in Sofia, Bulgaria (July 11-13, 2005) agreed that an Antarctic Climate Impact Assessment should be produced for the guidance of policy makers in the Antarctic Treaty System and to inform the public. The plan for the assessment was fleshed out at the first SCAR Cross-Linkages Workshop, in Amsterdam (November 15-17, 2005) (http://www.scar.org/researchgroups/crosslinkages/Amsterdam_Meeting_Report.pdf). Phase I focused on the physics of the climate system. The plans were presented to policy makers at the Antarctic Treaty Consultative Meeting in Edinburgh in 2006 (http://www.scar.org/treaty/atcmxxix/atcm29_ip089.pdf). Initial results were presented to policy makers at the Antarctic Treaty Consultative Meeting in New Delhi in 2007 (http://www.scar.org/treaty/atcmxxx/Atcm30_ip005_e.pdf) and published in the journal Reviews in Geophysics in January 2009 (Mayewski et al., 2009). In the meantime work had begun on Phase II, incorporating biology and chemistry, and preliminary results were presented to policy makers at the Antarctic Treaty Consultative Meeting in Kiev in 2008 (http://www.scar.org/treaty/atcmxxxi/ATCM31_IP62_ACCE.pdf), with final results being presented to the Antarctic Treaty Consultative Meeting in Baltimore in 2009 (http://www.scar.org/treaty/atcmxxxii/Atcm32_ip005_e.pdf). A summary paper on the science is currently being prepared for the journal Antarctic Science.

Inputs were obtained through a process of open consultation with the wider community in which scientists affiliated with SCAR or known to be active in climate and environmental sciences in Antarctica or the Southern Ocean were asked for text on key topics. A first draft was then circulated to the wider community for comment in June 2008. The editorial board then modified the text for a second round of open consultation in March 2009. Parties to the Antarctic Treaty, representatives of the Commission for the Conservation of Antarctic Marine Living Resources (CAMLR), and representatives of the Council of Managers of National Antarctic Programmes (COMNAP) were also asked for input.

The book is available as 500 hard copies, and as individual downloadable chapters on the SCAR web site, so as to encourage its widespread use as a research and teaching resource. It is SCAR's intention to update the report at regular intervals and to provide updates of advice on the topic to the Antarctic Treaty Consultative Meetings.

This volume is a contribution to the International Polar Year 2007-2008. It is also a contribution to the goals of the World Climate Research Programme (WCRP), and in particular to its Climate and Cryosphere programme (CliC), of which SCAR is a co-sponsor. In addition it is intended that it be made available to attendees at the meeting of the UN Framework Convention on Climate Change in Copenhagen in December 2009, and subsequently to the Intergovernmental Panel on Climate Change.

There is a historical context to the development of the ACCE volume, which can be considered as an eventual outcome of the work of the SCAR Group of Specialists on Antarctic Climate Research formed in 1980 to plan the Antarctic contribution to the WCRP (then about to be formed). The first product of the Group *was "Antarctic Climate Research: Proposals for the Implementation of a Programme of Antarctic Research contributing to the World Climate Research Programme". I Allison (ed), 1983. Cambridge, SCAR, 65 pp.* Many of the recommendations of that report continue to be carried out, and their pursuance over the years has enabled us to arrive at the ACCE review.

Further plans for climate-related research were developed in the late 1980s and early 1990s by SCAR's Steering Committee for the newly formed International Geosphere-Biosphere Program (IGBP). They were published as: (i) *The Role of Antarctica in Global Change. Prepared by the SCAR Steering Committee for the IGBP: 28 pp., 1989. ISBN 0 930 35718 3*, and *The Role of the Antarctic in Global Change: An International Plan for a Regional Research Programme, Prepared by the SCAR Steering Committee for the IGBP: 54 pp., 1992. ISBN 0 948277 15 7.* Following the recommendations of those reports SCAR created programmes such as Antarctic Sea Ice Processes and Climate (ASPeCt), Ice Sheet Mass Balance and Sea Level (ISMASS) and the International Trans-Antarctic Scientific Expedition (ITASE) to explore aspects of the physical and chemical elements of the climate system in the Antarctic. In parallel, SCAR developed programmes such as ANTOSTRAT to look at the geological history of climate change and the development of the ice sheet in the distant past, and SCAR biologists began working on EVOLANTA to examine the evolution of organisms through time in response to external forcings. With SCAR's reorganisation in 2004, the physical climate activities became absorbed into AGCS, the biological activities became absorbed into EBA, and the geological ones into ACE. AGCS, ACE and EBA all collaborated on ACCE.

The ACCE editorial board was responsible for overall editing, with some editors operating as lead authors on certain chapters. The editorial board takes full responsibility for errors. Editors are listed alphabetically following the lead editor, Dr. John Turner.

The editors would like to thank the 100 scientists from various fields who provided substantial textual input and who are listed alphabetically as authors of the appropriate chapters after the chapter editor. Many others provide editorial comments and corrections of matters of fact. We also thank Ms. Gill Alexander from the British Antarctic Survey for collating all the references and formatting the material for publication. Mr. Jamie Oliver from BAS prepared the front cover.

EXECUTIVE SUMMARY

Introduction

1. The Antarctic climate system varies on time scales from orbital, to millennial to sub-annual and is closely coupled to other parts of the global climate system. The Antarctic Climate Change and the Environment review discusses these variations from the perspective of the geological record and the recent historical period from which we have instrumental data (the last 50 years), discusses their consequences for the biosphere, and shows how the latest numerical models project changes into the future, taking into account human interference in the form of the release of greenhouse gases and chlorofluorcarbons into the atmosphere. This report highlights the profound impact that the ozone hole has had on the Antarctic environment over the last 30 years, shielding the continent from much of the effect of global warming. However, this situation will not last. Over the next century we expect ozone concentrations above the Antarctic to recover, but if greenhouse gas concentrations increase at the present rate then temperatures across the continent will increase by several degrees and there will be about one third less sea ice.

The Geological Dimension (Deep Time)

2. Studying the history of Antarctica's climate and environment provides the context for understanding present day climate and environmental changes. It allows researchers to determine the processes that led to the development of our present interglacial period and to define the ranges of natural climate and environmental variability on timescales from decades to millennia that have prevailed over the past million years. Knowing this natural variability enables researchers to identify when present day changes exceed the natural state. The palaeorecords show that change is normal and the unexpected can happen.
3. Concentrations of the greenhouse gas CO_2 in the atmosphere have ranged from roughly 3000 ppm (parts per million) in the Early Cretaceous 130 million years ago (Ma) to about 1000 ppm in the Late Cretaceous (at 70 Ma) and Early Cenozoic (at 45 Ma), leading to global temperatures 6° or 7°C warmer than present. These high CO_2 levels were products of the Earth's biogeochemical cycles. During these times there was little or no ice on land.
4. The first continental-scale ice sheets formed on Antarctica around 34 Ma, most likely in response to a decline in atmospheric CO_2 levels caused by a combination of reduced CO_2 out-gassing from mid-ocean ridges and volcanoes, and increased carbon burial. This decline resulted in a fall in global temperatures to around 4° C higher than today. At a maximum these early ice sheets reached the edge of the Antarctic continent, but were most likely warmer and thinner than today's. Further sharp cooling took place at around 14 Ma, probably accelerated by the growing physical and thermal isolation of Antarctica as other continents drifted away from it, and as the Antarctic Circumpolar Current (ACC) developed, rather than due to any change in atmospheric CO_2 levels. At that time the ice sheet thickened to more or less its modern configuration. During the Pliocene (5-3 Ma), mean global temperatures were 2-3° C above pre-industrial values, CO_2 values may have reached 400 ppm, and sea levels were 15-25 m above today's.
5. The earliest cold-climate marine fauna is thought to date from the latest Eocene-Oligocene (c. 35 Ma). The establishment of the Polar Front separating warm water in the north from cold water in the south created a barrier for migration of shallow and open water marine

organisms between the Antarctic and lower latitudes. This promoted adaptive evolution to cold temperature and extreme seasonality to develop in isolation, and led to the current Antarctic marine biota, which is second only to coral reefs in terms of species diversity and biomass. In contrast to marine faunas elsewhere the Antarctic fish fauna is dominated by a single highly endemic taxonomic group - the suborder Notothenioidei. The evolution of antifreeze proteins and the loss of haemoglobin in some members of this order is a particularly advanced adaptation to the environment. The development of sea-ice made the success of krill possible and, consequently, shaped all trophic levels of the open ocean ecosystem. Invertebrates inhabiting the deep-sea experienced considerable exchange with northerly adjacent areas due to sharing similar environmental conditions.

6. In an analogous fashion, circumpolar atmospheric circulation patterns isolated terrestrial habitats from potential sources of colonists at lower latitudes. In some contrast with the marine environment, the combination of continental scale ice sheet formation and advance, and extreme environmental conditions, led to large-scale (but incomplete) extinction of pre-existing biota, and to evolutionary radiation amongst the remaining survivors. Fossil evidence shows the change from species associated with the arid sub-tropical climates of Gondwana to cool temperate rainforest and then cold tundra when the Antarctic was isolated by the opening of the Drake Passage and the separation of the Tasman Rise. Most of these species are now extinct on the continent but recent molecular, phylogenetic and fossil evidence suggests that some species groups have adapted to the environmental changes including chironomids, mites, copepods, springtails, nematodes and cyanobacteria, many of which have endemic representatives on the continent.

The Last Million Years

7. The long periods of cold of the Pleistocene glaciation (post 1.8 Ma) were subject to cycles of warming and cooling with frequencies of 20,000, 41,000 and 100,000 years in response to variations in the Earth's orbit around the Sun. These led periodically to the development of short warm interglacial periods like that of the last 10,000 years. Over the past 400,000 years interglacial periods have recurred at intervals of around 100,000 years.

8. Ice core data from glacial cycles over the last 800,000 years show that CO_2 and mean temperature values ranged globally from 180 ppm and 10° C in glacial periods to 280 ppm and 15° C in interglacial periods. Temperature differences between glacial and interglacial periods in Antarctica averaged around 9°C. Ice sheets expanded in glacial periods, with sea level dropping by 120 m on average. Ice cores from both Antarctica and Greenland show that over the past 400,000 years interglacial temperatures were between 2-5ºC higher and sea levels were 4-6 m higher than they are today.

9. Diatom data from sediment cores show that at the Last Glacial Maximum (LGM), about 21,000 years before present, Antarctic sea ice was double its current extent in both winter and summer. Related sea surface temperature calculations show that both the Polar Front and the subAntarctic Front shifted to the north during the LGM by between 2° and 10° in latitude from their present locations.

10. The expansion and contraction of the Antarctic ice sheets undoubtedly led to the local extinction of biological communities on the Antarctic continent during glacial periods. Subsequent interglacial re-colonisation and the resulting present-day biodiversity is a result of whether the species survived the glacial maxima in refugia, then recolonised

deglaciated areas, or arrived through post-glacial dispersal from lower latitude lands that remained ice free, or are present through a combination of both mechanisms.

11. Continuous evolutionary development, especially during glacials in isolated refugia, is assumed to be a major driving force explaining the relatively high biodiversity of benthic (bottom-dwelling) marine organisms. Expansions and contractions of the sea ice will also have had an impact on marine mammal and seabird distributions. These various expansions and contractions continued into the Holocene.

The Holocene

12. The transition from the LGM to the present interglacial period (the Holocene, beginning around 12,000 years ago) was the last major global climate change event. Geological evidence from land in the Antarctic shows that there were two marked warm periods in the Holocene, one between 11,500 - 9000 years ago, and one between 4,000 - 2,000 years ago. Some marine records show evidence of a climate optimum between about 7,000 - 3,000 years ago. These warm periods were likely to have raised temperatures by no more than around 0.5°C.

13. The ice core record shows dramatic changes in atmospheric circulation in the Antarctic, first at 6000 years ago with strengthening of the southern hemisphere westerlies, followed at 5400-5200 years ago with abrupt weakening of the southern hemisphere westerlies, then around 1200 years ago with intensification of the westerlies and the Amundsen Sea Low Pressure cell.

14. Links between the climates of the northern and southern hemispheres exist, but through most of the Holocene and in the prior ice age northern hemisphere climate events lagged southern hemisphere ones by several hundred years. In contrast, in recent decades the northern hemisphere signal of rising temperature since about 1850 AD has paralleled that of the southern hemisphere. Temperature change in the two hemispheres (at least as far as West Antarctica is concerned) now appears to be synchronous - a significant departure from former times, which suggests a new and different forcing, most likely related to anthropogenic activity in the form of enhanced greenhouse gases. There is no evidence in Antarctica for an equivalent to the northern hemisphere Medieval Warm Period, and there is only weak circumstantial evidence in a few places for a cool event crudely equivalent in time to the northern hemisphere's Little Ice Age.

Changes During the Instrumental Period

15. The instrumental period began in the Antarctic with the International Geophysical Year, about 50 years ago. Here we discuss key changes in the atmosphere, ice and ocean system.

The large scale circulation of the atmosphere

16. The major mode of variability in the atmospheric circulation of the high southern latitudes is the Southern Hemisphere Annular Mode (SAM), a circumpolar pattern of atmospheric mass displacement in which intensity and location of the gradient of air pressure between mid-latitudes (high pressure) and the Antarctic coast (low pressure) changes in a non-periodic way over a wide range of time scales. Over the past 50 years the SAM became more positive, as pressure dropped around the coast of the Antarctic and increased at mid-

latitudes. Since the late 1970s this change increased westerly winds over the Southern Ocean by 15-20%. The combination of the stronger westerly winds around the continent, with the off-pole displacement of Antarctica, has led to a deepening of the Amundsen Sea Low, with consequent effects on temperature and sea ice in the coastal region of West Antarctica.

17. The SAM changed because of the increase in greenhouse gases and the development of the Antarctic ozone hole, the loss of stratospheric ozone having by far the greatest influence. The ozone hole occurs in the Austral spring. At that time of year the loss of stratospheric ozone cools the Antarctic stratosphere, so increasing the strength of the polar vortex – a large high altitude cyclonic circulation that forms in winter in the middle and upper troposphere and stratosphere over the Southern Ocean around Antarctica. During the summer and autumn the effects of the ozone hole propagate down through the atmosphere, increasing the atmospheric circulation around Antarctica at lower levels. As a result, the greatest change in the SAM, which is indicative of surface conditions, is in the autumn.

18. Changes in the SAM from 1958-1997 led to a decrease in the annual and seasonal numbers of cyclones south of 40°S. There are now fewer but more intense cyclones in the Antarctic coastal zone between 60-70°S, except in the Amundsen-Bellingshausen Sea region.

19. In recent decades there have been more frequent and more intense El Niño events. During some of them a signal of the El Niño cycle can be seen in the Antarctic. There is no evidence that this trend has affected long term climate trends in the Antarctic.

Atmospheric temperatures

20. Surface temperature trends show significant warming across the Antarctic Peninsula and to a lesser extent West Antarctica since the early 1950s, with little change across the rest of the continent. The largest warming trends occur on the western and northern parts of the Antarctic Peninsula. There the Faraday/Vernadsky Station has experienced the largest statistically significant (<5% level) trend of +0.53° C per decade for the period 1951 - 2006. The 100-year record from Orcadas on Laurie Island, South Orkney Islands, shows a warming of +0.20°C per decade. The western Peninsula warming has been largest during the winter, with winter temperatures at Faraday/Vernadsky increasing by +1.03°C per decade from 1950 - 2006. There is a high correlation during the winter between sea ice extent and surface temperatures, suggesting more sea ice during the 1950s - 1960s and reduction since then. This warming may reflect natural variability.

21. Temperatures on the eastern side of the Antarctic Peninsula have risen most during the summer and autumn (at +0.41°C per decade from1946 - 2006 at Esperanza), linked to the strengthening of the westerlies that took place as the SAM shifted into its positive phase, primarily as a result of the ozone hole. Stronger westerly winds bring warm, maritime air masses across the Peninsula to the low-lying ice shelves on the eastern side.

22. Based on data from satellites and automated weather stations, West Antarctica has warmed by about 0.1°C/decade, especially in winter and spring. Ice core data from the Siple Dome suggest that this warming began around 1800. There have been few statistically significant changes in surface temperature over the instrumental period elsewhere in Antarctica.

23. On the plateau, Amundsen-Scott Station at the South Pole shows a statistically significant cooling in recent decades, interpreted as due to fewer maritime air masses penetrating into the interior of the continent.

24. Temperatures reconstructed from ice cores show large inter-annual to decadal variability, with the dominant pattern being anti-phase anomalies between the continent and the peninsula, which is the classic signature of the SAM. The reconstruction suggests that Antarctic temperatures increased on average by about 0.2°C since the late nineteenth century.

25. Antarctic radiosonde temperature profiles show that the troposphere has warmed at 5 km above sea level, and that the stratosphere above it has cooled over the last 30 years. This pattern would be expected from increasing greenhouse gases. The tropospheric warming in winter is the largest on Earth at this level. It may, in part, be a result of the insulating effect of greater amounts of polar stratospheric cloud during the winter. These clouds form in response to stratospheric cooling related also to the ozone hole.

Snowfall

26. On average, about 6 mm global sea level equivalent falls as snow on Antarctica each year, but with no statistically significant increase since 1957. Snowfall trends vary from region to region. Snowfall has increased on the western side of the Peninsula, where it has been linked to decreases in Adélie penguin populations, which prefer snow-free nesting habitat.

The Antarctic ozone hole

27. Stratospheric ozone levels began to decline in the 1970s, following widespread releases of CFCs and halons into the atmosphere that destroyed virtually all ozone between heights of 14 and 22 km over Antarctica. Owing to the success of the Montreal Protocol, the amounts of ozone-depleting substances in the stratosphere are now decreasing by about 1% per year. As a result the size and depth of the ozone hole have stabilised; neither are yet reducing.

Terrestrial biology

28. The clearest example of Antarctic terrestrial organisms responding to climate change is given by the two native flowering plants (*Deschampsia antarctica* and *Colobanthus quitensis)* in the maritime Antarctic, which have increased in abundance at some sites. Warming encourages the growth and spreading of established plants and increased establishment of seedlings. Changes in temperature and precipitation have increased biological production in lakes, mainly due to decreases in the duration and extent of lake ice cover. Some lakes have become more saline due to drier conditions.

29. Alien microbes, fungi, plants and animals introduced through human activity occur on most of the sub-Antarctic islands and some parts of the continent. They have impacted the structure and functioning of native ecosystems and their biota. On Marion Island and Gough Island rates of establishment through anthropogenic introduction outweigh those from natural colonization processes by two orders of magnitude or more.

The terrestrial cryosphere

30. Ice shelves in the Antarctic Peninsula have changed rapidly in recent decades. Warming has caused retreat of ice shelves on both sides of the Peninsula. Loss of ice on the eastern side results from warm air being brought over the Peninsula by the stronger westerlies forced by changes in the SAM, driven ultimately by the development of the ozone hole. Ice-shelf retreat results from increased fracturing via melt-water infilling of pre-existing crevasses, and the penetration of warm ocean masses beneath ice shelves. Removal of ice shelves has led to the speeding up of glacier flow from inland.

31. Some formerly snow- and ice-covered islands are now increasingly snow-free during the summer. Glaciers on Heard Island reduced by 11% since the 1940s, and several coastal lagoons have formed there. On South Georgia, 28 of 36 surveyed glaciers are retreating, 2 are advancing, and 6 are stable. On Signy Island ice cover has reduced by around 40%.

32. Of the 244 marine glaciers that drain the ice sheet and associated islands of the Antarctic Peninsula, 212 (87%) have shown overall retreat since 1953. The other 32 glaciers have shown small advances.

33. The Amundsen Sea sector is the most rapidly changing region of the Antarctic ice sheet. The grounding line at Pine Island has retreated, and the Pine Island Glacier is now moving at speeds 60% higher than in the 1970s. The Thwaites Glacier and four other glaciers in this sector show accelerated thinning. Smith Glacier has increased flow speed 83% since 1992. The Pine Island and adjacent glacier systems are currently more than 40% out of balance, discharging 280 ± 9 Gt per year of ice, while they receive only 177 ± 25 Gt per year of new snowfall. The current rate of mass loss from the Amundsen Sea embayment ranges from 50 to 137 Gt per year, equivalent to the current rate of mass loss from the entire Greenland ice sheet, and making a significant contribution to sea level rise.

34. The changes result from warming of the sea beneath the ice shelves connected to the glaciers. The stronger winds associated with the more positive SAM drive warm Circumpolar Deep Water up against the western Peninsula coast and the Amundsen Sea coast.

35. Changes are less dramatic across most of the East Antarctic ice sheet, with the most significant changes close to the coast. The ice sheet shows interior thickening at modest rates and a mixture of modest thickening and strong thinning among the fringing ice shelves. Increasing coastal melt is suggested by recent passive microwave data.

Sea level changes

36. Data from tide gauges and satellite altimeters suggest that in the 1990s-2000s global sea level rose at a rate of 3 mm per year or more, which is higher than expected from IPCC projections, though the rate has since slowed to 2.5 mm/yr. There is concern about possibly large contributions resulting from the dynamic instability of ice sheets during the 21st century. Around 2005, the Antarctic Peninsula was estimated as contributing to global sea-level rise at a rate of 0.16 ± 0.06 mm per year

The Southern Ocean

37. In surface waters change is difficult to detect because an intensive seasonal cycle can induce large errors when there are only a few samples. However, around South Georgia observations exist since 1925 that are frequent enough to resolve the annual cycle and reveal a significant warming averaging 2.3°C over 81 years in the upper 150 m, and being about twice as strong in winter than in summer.
38. The waters of the ACC have warmed more rapidly than the global ocean as a whole, increasing by 0.06°C/decade at depths between 300 – 1000 m over the 1960s to 2000s, and by 0.09°C/decade since the 1980s. The warming is more intense on the southern side of the ACC than north of it, and a maximum increase of 0.17ºC/decade is found in Upper Circumpolar Deep Water at depths of 150-500m on the poleward side of the Polar Front. The changes are consistent with a southward shift of the ACC in response to a southward shift of the westerly winds driven by enhanced greenhouse forcing. North of the ACC a significant freshening of -0.01 salinity units per decade is observed since the 1980s. There is no evidence for an increase in ACC transport. Recent studies suggest that an increase in wind forcing causes an increase in the meridional transport of heat and salt by eddies rather than a change in zonal transport by the current.
39. Changes are evident in the character of the Ross and Weddell Seas, but neither show major long-term trends apart from slight freshening. During the 1970s a persistent gap in the sea ice, the Weddell Polynya, occurred for several winters. The polynya had a significant impact on deep-water properties, being a place where the ocean lost a great deal of heat to the atmosphere. Deep water masses in the Weddell Sea show significant decadal and regional change making it difficult to detect long-term trends. Central Weddell Sea bottom water is warming and becoming saltier over decadal time scales, while bottom water in the Western Weddell Sea and the Australian sector (including the Ross Sea and Adélie Land) is cooling and becoming fresher. The changes are of interest because Antarctic Bottom Water originates in these places and changes there will spread into the world ocean.

Biogeochemistry

40. The Southern Ocean ventilates the global oceans and regulates the climate system by taking up and storing heat, freshwater, O_2 and atmospheric CO_2. From 1991 - 2007 the concentration of CO2 in the ocean increased south of 20°S in the Southern Indian Ocean. At latitudes pole-ward of 40°S, CO2 in the ocean increased faster than it did in the atmosphere, suggesting that the ocean became less effective as a sink for atmospheric CO_2. The stronger westerly winds lead to surface oceanic water being mixed with deeper water rich in CO2, which saturates the carbon reservoir of the surface water, thus limiting its ability to absorb carbon dioxide from the atmosphere These changes seem to be linked to the increase in wind strength driven by the more positive SAM. An increase in the ocean's CO_2 content makes the ocean more acidic.

Sea ice

41. For the first half of the Twentieth Century, ship observations suggest that the extent of sea ice was greater than seen in recent decades, although the validity of such observations is questioned. The sea ice extent data derived from satellite measurements from 1979 - 2006 show a positive trend of around 1% per decade. The trend is positive in all sectors except

the Bellingshausen Sea, where sea ice extent has been significantly reduced. The greatest increase, at around 4.5% per decade, is in the Ross Sea. The reduction of ice extent in the Bellingshausen Sea and its increase in the Ross Sea are both a result of the deepening of the Amundsen Sea Low, which is linked to the ozone hole.

Permafrost

42. There is little information on permafrost in the Antarctic. On Signy Island the active layer (the layer experiencing seasonal freeze and thaw) increased in depth by 30 cm from 1963 - 1990, when Signy Island was warming, then decreased by the same amount from 1990 – 2001, when Signy Island cooled. In McMurdo Sound, the permafrost temperature at -360 cm has remained stable, despite a slight decrease in air temperature of 0.1°C per year.

Marine biology

43. The Southern Ocean ecosystem was significantly disturbed by whaling during the early part of the Twentieth Century, and by sealing before that. About 300,000 blue whales were killed within the span of a few decades, equivalent to more than 30 million tonnes of biomass. Most were killed on their feeding grounds in the southwest Atlantic in an area of some 2 million km^2, which translates to a density of one blue whale per 6 km^2. Following near-extinction of some whale populations, the krill stock was expected to increase due to release from grazing pressure. It did not. While predation by seals and birds increased, the total bird and seal biomass remains only a fraction of that of the former whale population. Benthic habitats have been regionally devastated by bottom fishing; fish stocks and invertebrates are expected to recover only very slowly.

44. For the past 50 years the Antarctic marine ecosystem has been affected by climate change especially on the western side of the Peninsula, with its warming water and declining sea-ice. There we find a decline in krill stocks, in some areas a decrease in others an increase in phytoplankton and a southward shift in the population of gelatinous salps. The decline in phytoplankton may reflect a decrease in iron input from the continental margin that is in turn related to a reduction in the formation of sea ice in this region and hence to climate change. All sea-ice related components of the pelagic Antarctic ecosystem experienced consequences of the regional decrease of sea ice west of the Antarctic Peninsula, whilst for the bottom-inhabiting fauna so far only hints but not evidence for impact of climate change exists.

The Next 100 Years

45. Determining how the environment of the Antarctic will evolve over the next century presents challenges yet has implications for science and for policymakers. Climate evolution can only be predicted with some degree of confidence by using coupled atmosphere-ocean-ice models. The model outputs improve on simple projections of current trends, as they take a large number of parameters into consideration. They are the only means we have of providing synoptic views of future environmental behaviour, albeit crudely and at coarse resolution. The models do not quite accurately simulate the observed changes that have taken place over the last few decades, so there is still some uncertainty about their forward projections, particularly at regional scales. The models used in the IPCC Fourth

Assessment gave a wide range of predictions for some aspects of the Antarctic climate system, such as sea ice extent, which is sensitive to changes in atmospheric and oceanic conditions. The models can be weighted according to their skill in simulating modern conditions.

46. Numerically based biological models cannot yet approach the relative sophistication of models of the physics of the climate system, while physical models do not approach the level of spatial scale or resolution required for application to biological systems, providing yet a further limitation on what can be said about biotic and ecosystem responses.

47. The degree to which the Earth's climate will change over the next century is dependent on the success of efforts to reduce greenhouse gas (GHG) emissions. The ACCE report focuses on outputs from IPCC models that assumed a doubling of CO_2 and other gases by 2100. These outputs may be too conservative, given that some indicators are already changing faster than predicted in IPCC projections.

Atmospheric circulation

48. The recovery of the ozone hole combined with a continued increase in greenhouse gas emissions should continue strengthening the positive phase of the SAM, but with a trend that is less rapid than seen in the last two decades. We can thus expect to see further increases of surface winds over the Southern Ocean in the summer and autumn. This will lead to continuance of the pole-ward shift in the Southern Ocean storm track.

Temperature

49. The models project significant surface warming over Antarctica to 2100 AD, by 0.34°C/decade over land and grounded ice sheets, within a range from 0.14 to 0.5°C/decade. Over land, the largest increase is projected for the high-altitude interior of East Antarctica. Despite this change, the surface temperature by the year 2100 will remain well below freezing over most of Antarctica and will not contribute to melting inland.

50. The largest atmospheric warming projected by the models is over the sea ice zone in winter (0.51 ± 0.26°C/decade off East Antarctica), because of the retreat of the sea-ice edge and the consequent exposure of the ocean.

51. While we are confident in the overall projection of warming, we have low confidence in the regional detail, because of the large differences in regional outcomes between models.

52. The annual mean warming rate in the troposphere at 5 km above sea level is projected to be 0.28°C/decade - slightly smaller than the forecast surface warming.

53. As yet we cannot forecast either the magnitude or frequency of changes to extreme conditions over Antarctica – something that biologists need to assess potential impacts. The extreme temperature range between the coldest and warmest temperature of a given year is projected to decrease around the coasts and to show little change over the interior.

54. The warming of 3°C over the next century is faster than the fastest rate of rise seen in Antarctic ice cores (4°C per 1000 years), but it is comparable to or slower than the rates of temperature rise typical of Dansgaard-Oeschger events during glacial times in Greenland, of the Bolling-Allerød warming in Greenland 14,700 years ago, and of the warming in

Greenland at the end of the Younger Dryas around 11,700 years ago. Thus however unlikely such rapidity may appear, it is feasible in terms of what we know about the natural system.

Precipitation

55. Current numerical models generally underestimate precipitation for the 20^{th} century. This is due to problems in parameterizing key processes that drive precipitation (e.g. because polar cloud microphysics is poorly understood), and because the smooth coastal escarpment in a coarse resolution model causes cyclones to precipitate less than they do in reality. Warmer air temperatures and associated higher atmospheric moisture in most models cause net precipitation to increase in the future. Most climate models simulate a precipitation increase over Antarctica in the coming century that is larger in winter than in summer. Model outputs suggest that the snowfall over the continent may increase by 20% compared to current values. With the expected southward movement of the mid-latitude storm track we can expect greater precipitation and accumulation in the Antarctic coastal region. The form that precipitation takes is of biological significance (liquid water is immediately available to biota), with the balance between rain and snow expected to change, especially along the Antarctic Peninsula.

The ozone hole

56. By the middle of the 21^{st} century we expect that springtime concentrations of stratospheric ozone will have significantly recovered, but not to 1980 values. That is because increasing greenhouses gases will have accumulated in and warmed the troposphere, so further cooling the stratosphere. Destruction of ozone is more likely to continue in a colder stratosphere.

Tropospheric chemistry

57. Various trace gases, such as dimethyl sulphide (DMS) generated by plankton, are released from the oceans around Antarctica. Any projected loss of sea ice will enhance these emissions, which have seasonal cycles closely linked to the extent of sea ice and to the Sun. In a warmer world, with reduced sea ice extent, emissions of gases such as DMS would likely increase, with a wintertime minimum and an extended summer maximum. DMS is a source of cloud condensation nuclei (CCN) via its oxidation to sulphate. Increasing the number of CCN may increase cloudiness and albedo, thus influencing the Earth's climate.

Terrestrial biology

58. Increased temperatures may promote growth and reproduction, but also cause drought and associated effects. Changes to water availability have a greater effect than temperature on vegetation and faunal dynamics. Future regional patterns of water availability are unclear, but climate models predict an increase in precipitation in coastal regions. An increase in the frequency and intensity of freeze–thaw events could readily exceed the tolerance limits of many arthropods. With increases in temperature, many terrestrial species may exhibit faster metabolic rates, shorter life cycles and local expansion of populations. Even

subtle changes in temperature, precipitation and wind speed will probably alter the catchment of lakes, and of the time, depth and extent of their surface ice cover, water volume and chemistry, with resulting effects on lake ecosystems. Warming also increases the likelihood of invasion by more competitive alien species carried by water and air currents, humans and other animals.

The terrestrial cryosphere

59. Existing ice-sheet models do not properly reproduce the observed behaviour of ice sheets, casting doubt on their predictive value. The models fail to take into account mechanical degradation (e.g. water causing cracks to propagate in summer), changing lubrication of the base of the ice by an evolving subglacial hydrological regime, or the influence of variable sub-ice-shelf melting on the flow of outlet glaciers and ice streams. Predictions of the future state of ice sheets are based on a combination of inference from past behaviour, extension of current behaviour, and interpretation of proxy data and analogues from the geological record.

60. The most likely regions of future change are those changing today. Warmer waters will continue to well up onto the continental shelf in the Amundsen Sea, eroding the underside of the ice sheets and glaciers. It has been suggested that there is a 30% probability that loss of ice from the West Antarctic ice sheet could cause sea level to rise at a rate of 2 mm per year, and a 5% probability it could cause rates of 1 cm per year. In addition there is a concern that the ice in the Amundsen Sea Embayment could be entering a phase of collapse that could lead to de-glaciation of parts of the West Antarctic Ice Sheet. Ultimately, this sector could contribute 1.5 meters to global sea level, so a contribution from this sector alone of some tens of centimetres by 2100 cannot be discounted. These estimates are based upon the assumption that the ice sheets will respond linearly to warming and that sea level contributions will be confined to the West Antarctic ice sheet. Evidence for past abrupt changes in climate and inclusion of all marine-based regions could raise these estimates significantly.

61. On the Antarctic Peninsula, most of the effects leading to loss of ice are confined to the northern part, which contains a few centimetres of global sea-level rise. Increased warming will lead to a southerly progression of ice shelf disintegrations along both coasts. These may be preceded by an increase in surface melt-water lakes, and/or progressive retreat of the calving front. Prediction of the timing of ice shelf disintegration is not yet possible. Removal of ice shelves will cause glaciers to speed up. The total volume of ice on the Peninsula is 95,200 km^3, equivalent to 242 mm of sea-level or roughly half that of all glaciers and ice caps outside of Greenland and Antarctica. Increased warming may lead to the Peninsula making a substantial contribution to global sea level.

Sea level

62. The IPCC's Fourth Assessment Report projected a range of global sea-level increase from 18 to 59 cm between 1980-1999 and 2090-2099. This did not include a contribution from the dynamically driven changes in flow for portions of either the Greenland or Antarctic ice sheets. Recent modelling suggests that by 2100 global sea level may rise by up to 1.4m rather than the IPCC's suggested 59 cm. Sea level will not rise uniformly. The spatial

pattern of sea-level rise projections shows a minimum in sea-level rise in the Southern Ocean and a maximum in the Arctic Ocean.

Biogeochemistry

63. Model projections suggest that the Southern Ocean will be an increased sink of atmospheric CO_2. The magnitude of the uptake will depend on how the ocean responds to increases in ocean warming and stratification, which can drive both increases in CO_2 uptake through biological and export changes, and decreases through solubility and density changes.

The ocean circulation and water masses

64. Ocean models are a component of General Circulation Models (GCMs) and are deficient in having a typical grid spacing of around 100 km in the horizontal, which is larger than the typical ocean eddy, and so limits the ability of the models to simulate ocean behaviour - an important constraint given the key role of ocean eddies in north-south heat transport in the Southern Ocean. The models generally predict an intensification of the ACC in response to the forecast southward shift and intensification of the westerly winds over the Southern Ocean. The increase in ACC transport projected for 2100 reaches a few Sverdrups (1 Sv = one million cubic metres/second) in the Drake Passage. The enhanced winds induce a small southward displacement of the core of the ACC (less than 1° in latitude on average).
65. The observed mid-depth and surface layer warming of the Southern Ocean is projected to continue, reaching nearly all depths. Close to the surface, the warming is expected to be weaker than in other regions. Ocean ventilation could be enhanced because of enhanced divergence of surface waters induced by the increase in the wind stress and associated upwelling. Model calculations suggest that bottom waters warm by 0.25°C by 2100. That will decrease the density and hence the ventilation of Antarctic Bottom Water.

Sea ice

66. The models suggest that the annual average total sea-ice area will decrease by 2.6×10^6 km^2, or 33%. Most of the retreat is expected to be in winter and spring, so decreasing the amplitude of the seasonal cycle of sea ice area. The current generation of climate model are not able to provide a precise regional picture of the changes to be expected.

Permafrost

67. It is likely that there will be a reduction in permafrost area, accompanied by subsidence of ground surface and associated mass movements. Change is most likely in the northern Antarctic Peninsula and the South Shetland and South Orkney Islands and coastal areas in East Antarctica. The forecast changes imply risks to infrastructure.

Marine biology

68. Most evidence for how benthic organisms may cope with temperature rise is experimental. Being typically 'stenothermal' (able to live within a limited range of temperature) is a key trait of Antarctic marine animals. If they are truly so limited they would be highly sensitive to significant warming. Experiments show most species have upper lethal temperatures below 10°C. Some can survive just a 5°C change. However, a rise of this magnitude in the Southern Ocean is extremely unlikely by 2100. That being said, the behaviour of organisms can be affected at lower temperatures long before lethal levels are reached; whether populations or species will survive future temperature rises may be dictated by their ability to carry out critical activities e.g. feeding, swimming, reproduction.

69. Model projections suggest that bottom water temperatures on the continental shelf by 2100 AD are likely to be warmer by between 0.5 and 0.75°C, except in the Weddell Sea where the warming is likely to be less. The lack of projected warming of surface and bottom waters by more than about 0.75°C suggests that the effects of warming on the marine biota may be less than has been feared from laboratory experiments, at least over this timescale. Projected warming greater than 1.5°C is restricted to the surface waters near the core of the ACC.

70. Several Antarctic taxa are distributed across sites or depths with a much greater temperature range than 'typical Antarctic conditions'. South Georgia has populations of many typical Antarctic species, despite experiencing maximum summer temperatures 3°C warmer than the Antarctic Peninsula. Thus there may be a conflict between experimental and ecological evaluations of vulnerability. The ecological context may be crucial.

71. If the sea-ice cover continues to decrease, marine ice algae will begin to disappear due to loss of habitat, which may cause a cascade through higher trophic levels in the food web. Given a complete removal of sea ice we might expect extinction of those species that presently depend on it for survival, including some fish, penguins, seals and whales. Climate models suggest that complete removal is unlikely within the next 100 years, and indeed it does not seem to have been removed completely during previous interglacials. With the decline in sea ice there should be more algal blooms supplying food to benthic organisms on the shelf. A resulting increase in phytodetritus on the shelf may cause a decline in suspension feeders adapted to limited food supplies, and to their associated fauna.

72. When ice shelves collapse, the changes from a unique ice-shelf-covered ecosystem to a typical Antarctic shelf ecosystem, with high primary production during a short summer, are likely to be among the largest ecosystem changes on the planet.

73. If surface ocean pH levels become more acid by 0.2 to 0.3 units by 2100 it seems likely that there will be some thinning of the aragonite skeletons of the pteropods that are an important part of the plankton at the base of the food chain; also benthic calcifiers such as corals are potentially threatened. The Southern Ocean is at higher risk from this than other oceans because it has low saturation levels of $CaCO_3$.

74. Given the slow rates of growth and high degree of endemism in Antarctic species, continued ocean warming and expanded tourism and scientific activity may lead to the wider establishment of non-indigenous species by 2100, and consequent reduction or extinction of some locally endemic species. Invasion by new species will likely remain restricted to

isolated areas where invaders can survive at their physiological limits. As yet it is unclear if the finding of a very few 'non-indigenous' macroalgae and invertebrate animals represents rare occurrences at their natural southern distribution limits, or the first stages of a marine biogeographical shift induced by warming.

75. Acute agents of disturbance to the marine biota include: 1) increased ice-loading and coastal concentrations of large icebergs from ice shelf collapse, resulting in more ice scour; 2) increased coastal sedimentation associated with ice melt, smothering benthos and hindering feeding; 3) freshening of surface waters leading to stratification of the water column; and 4) thermal events like those associated with El Niño events.

76. Chronic impacts of climate change include: 1) ice shelf disintegration, exposing new habitats; 2) long-term decreases in ice scour by icebergs, leading to decreased local but increased regional biodiversity; 3) the physiological effect of direct warming, leading to reduced performance of critical activities and thus geographic and bathymetric migration; 4) benthic responses to changes in the pelagic system, especially in the food web; 5) increased acidification, leading to skeletal synthesis and maintenance problems; and 6) slight deoxygenation of surface waters, ultimately leading to more serious deoxygenation in deeper layers. The absence of wide latitudinal and environmental gradients around the Antarctic continent minimises the advantage of migration for survival.

77. Species such as fur seals are likely to respond most to changes in extreme climate events, for instance caused by changes in the El Niño - Southern Oscillation. Emperor penguins and other ice-dependent species depend on the sea-ice habitat to complete their life cycle. A significant decline in sea ice is likely to affect their populations, and may lead to true Antarctic species being displaced by immigrating subAntarctic species.

78. It seems likely that only a few species will become extinct by 2100, either because they proved unable to cope ecologically and physiologically with such an increase, or because they were restricted in their occurrence to an area with an above average temperature increase. Studies of biodiversity, coupled to sound data handling and dissemination, will bring a better understanding of how life has evolved in the marine environment, and to what extent it can potentially respond to change. Bridges between different disciplines and international programmes will provide a legacy of knowledge for future generations in the form of a comprehensive information system.

Concluding remarks

79. The climate of the high latitude areas is more variable than that of tropical or mid-latitude regions and has experienced a huge range of conditions over the last few million years. The snapshot we have of the climate during the instrumental period is tiny in the long history of the continent, and separation of natural climate variability from anthropogenic influences is difficult. However, the effects of increased greenhouse gases and decreases in stratospheric ozone are already evident. The effects of the expected increase in greenhouses gases over the next century, if they continue to rise at the current rate, will be remarkable because of their speed. Removal of the cooling effect of the ozone hole as it diminishes in extent will exacerbate the problem. We can make reasonably broad estimates of how quantities such as temperature, precipitation and sea ice extent might change, and consider the possible impact on marine and terrestrial biota. We cannot yet say with confidence how the large ice sheets of Antarctic will respond, but observed

recent rapid changes give cause for concern – especially for the stability of parts of West Antarctica.

80. Marine biologists need more information from ecophysiological field-studies and laboratory experiments about the sensitivity of ecological key species, along with more information on the geography of the hot- and cold-spots of Antarctic biodiversity and their ecosystem functioning, and identification of the main biological and physical driving forces. These challenges should provide the basis for further development of spatially explicit numerical simulations of the state-of-the-art of the Antarctic ecosystem, and extrapolations from this –with the support of, or in combination with results from physical models - into the future, using various climate scenarios.

Chapter 1

The Antarctic Environment in the Global System

Chapter Editor:

Colin Summerhayes

Contributing Authors:

David Ainley, Peter Barrett, Robert Bindschadler, Andrew Clarke, Pete Convey, Eberhard Fahrbach, Julian Gutt, Dominic Hodgson, Mike Meredith, Alison Murray, Hans-Otto Pörtner, Guido di Prisco, Sigrid Schiel, Kevin Speer, Colin Summerhayes, John Turner, Cinzia Verde and Anne Willems

1.1 The Physical Setting

This first chapter sets out the nature of the physical-chemical environment of the Antarctic region and briefly describes key features of life there today.

The Scientific Committee on Antarctic Research (SCAR) considers the Antarctic region to include the continent, its offshore islands, and the surrounding Southern Ocean (Figure 1.1) including the Antarctic Circumpolar Current, the northern boundary of which is the Sub-Antarctic Front (SAF) (Figure 1.2), and we will do the same here. Southern Ocean islands that lie north of the SAF and yet fall into SCAR's area of interest include Ile Amsterdam, Ile St. Paul, Macquarie Island and Gough Island.

Antarctica is renowned as being the highest, driest, windiest and coldest continent, boasting the lowest recorded temperature on Earth, -89.2°C, at Russia's Vostok Station (Figure 1.1), which is located on the Polar Plateau (the white area of East Antarctica in Figure 1.2). The continent covers an area of 14×10^6 km^2, which is about 10% of the land surface of the Earth. That area includes the ice sheet, the floating ice shelves (offshore continuations of the continental ice sheet) and the areas of fast ice (sea-ice that has become frozen to ice shelves or to the land and does not drift with wind and currents). Most of the continent apart from the northern part of the Antarctic Peninsula lies south of the Antarctic Circle (at latitude 66° 33' 39"S), beyond which there are 24 hours of continuous daylight at the austral summer solstice in December, and 24 hours continuous darkness at the austral winter solstice in June. The sun rises at the South Pole on 22 September, the austral vernal equinox, and sets on 23 March, the austral autumn equinox.

The land surface rises rapidly away from the coast (Figure 1.2), and the continent has the highest mean elevation of any continent on Earth, at around 2,200 m. Much of East Antarctica comprises a high, domed Polar Plateau, with a crest at 4,093 m at Dome A (80° 22'S, 77° 32'E). The South Pole is much lower, at 2,835 m, reflecting the fact that although it lies on the Plateau it is not the geographic centre of the continent, which lies instead at the Pole of Inaccessibility at 85° 50'S, 65° 47'E. West Antarctica, with a mean elevation of 850 m, is much lower than East Antarctica, though it includes the highest Antarctic mountain - Mt Vinson in the Ellsworth Mountains, at 4,892 m (78° 35'S, 85° 25'W). Rocky outcrops like Mt Vinson rise above the surrounding ice sheet in several places, notably in the Transantarctic Mountains, which separate East and West Antarctica along a line from the Ross Sea to the Weddell Sea (Figure 1.1) and rise to a maximum height of 4,528 m. Exposed rock and soil total only about 46,000 km^2 of continental Antarctica (approximately 0.33% of the land area) (Fox and Cooper, 1994).

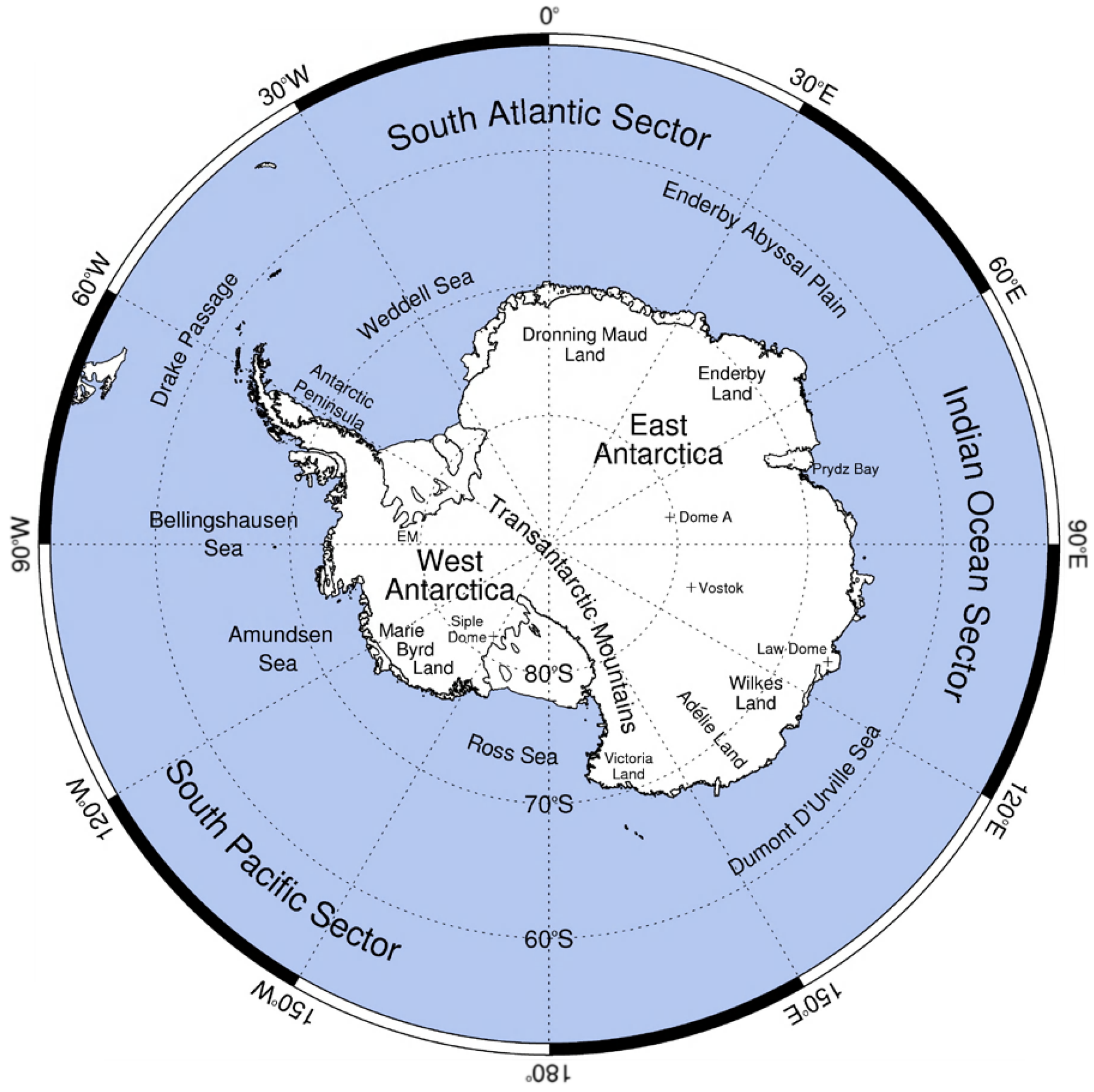


Figure 1.1a A map of Antarctica showing selected topographic features and locations.

The surrounding seabed (Figure 1.2) comprises a number of features, including the continental shelf, the deeper parts of which are typically grooved by furrows carved by glaciers during the ice ages, when sea level was lower and the ice margin extended out towards the edge of the continental shelf. The shallower parts of the shelf are still being furrowed by modern icebergs, whose keels may reach depths of several hundred metres. The continental slope (red and yellow in Figure 1.2) is cut by submarine canyons, and merges with the adjacent deep sea floor (>3,500 m below sea level, and blue in Figure 1.2), which includes abyssal plains built by sediment deposited from episodic turbidity currents that flowed down the canyons. Other features of note are abyssal hills mantled with pelagic clays, and mountainous ridges of various kinds, like the Scotia Arc (red and yellow in Figure 1.2), which connects the tip of the Antarctic Peninsula to South America); and a surrounding deep-sea mountain range consisting of the southern parts of the global Mid-Ocean Ridge System (pale green with yellow crests in Figure 1.2). The crest of the Mid-Ocean Ridge System outlines the borders of the Antarctic Plate, one of the Earth's great tectonic plates. These various topographic features form key elements of the habitats of marine organisms, and constrain ocean circulation.

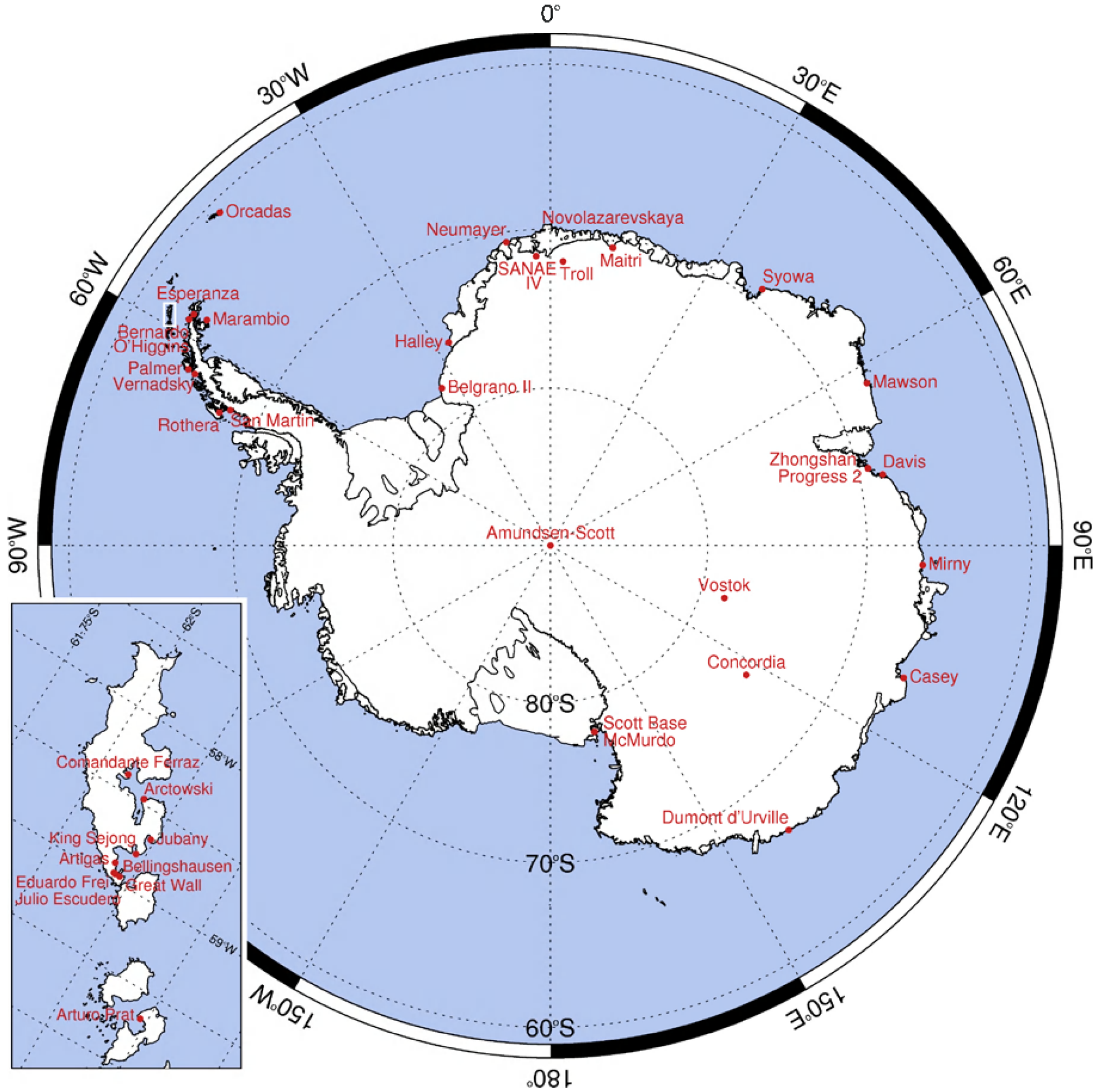


Figure 1.1b A map of Antarctica showing selected stations.

Many benthic (bottom-dwelling) organisms live on the Antarctic continental shelf, which comprises almost 15% of the global continental shelf area (Clarke and Johnston, 2003) – in total around 4.6×10^6 km^2. The shelf is unusually deep, reaching 800 m in places, as a side effect of the continent being depressed by the weight of its massive ice sheet. More than 95% of the shelf lies at depths beyond the reach of the scouring effects of sea ice or wave action, is below the reach of sunlight (i.e. outside the euphotic zone), and is inaccessible to scuba divers. Floating ice shelves (Figure 1.3) cover about one third of the continental shelf; the rest is covered by sea ice for around half the year. Both the sea and the seabed below the ice shelves are among the least known habitats on Earth.

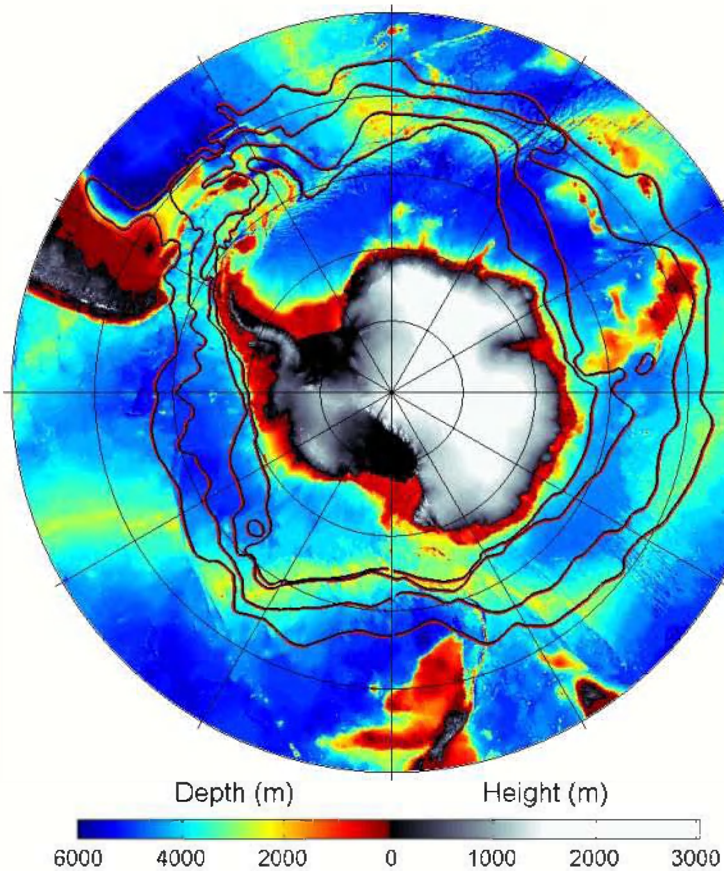


Figure 1.2 The land topography and the bathymetry of the seabed around Antarctica (in metres). Four major oceanic fronts are shown (not labelled), which are (north-to-south): the Sub-Antarctic Front (SAF), the Polar Front (PF), the Southern ACC Front (SACCF) and the Southern Boundary (SB). The Antarctic Circumpolar Current runs between the SAF and the SB. Figure produced by M.P. Meredith (BAS; pers. comm.) using frontal locations adapted from Orsi et al., 1995) and bathymetry from the General Bathymetric Chart of the Oceans (GEBCO) Centennial digital topography data (www.gebco.net).

1.2 The Antarctic Cryosphere

The continent is dominated by the Antarctic Ice Sheet (Figures 1.2 and 1.3), a vast contiguous mass of glacial ice that covers the Antarctic continent and surrounding seas. It is the single largest solid object on the surface of the planet, containing around 30×10^6 km^3 of ice or 70% of the Earth's freshwater, and covering around 99.6 % of what we generally consider to be the Antarctic continent (Fox and Cooper, 1994). The ice sheet is made up of three distinct morphological zones, consisting of East Antarctica (covering an area of 10.35×10^6 km^2), West Antarctica (1.97×10^6 km^2) and the Antarctic Peninsula (0.52×10^6 km^2) (Figure 1.1).

East and West Antarctica are separated by the Transantarctic Mountains, which extend from Victoria Land to the Ronne Ice Shelf. The Antarctic Peninsula is the only part of the continent that extends a significant way northwards from the main ice sheet, reaching latitude 63°S. It is a narrow mountainous region with an average width of 70 km and a mean height of 1,500 m. This north-south trending mountain barrier has a major influence on the west-east trending oceanic and atmospheric circulations of the high southern latitudes.

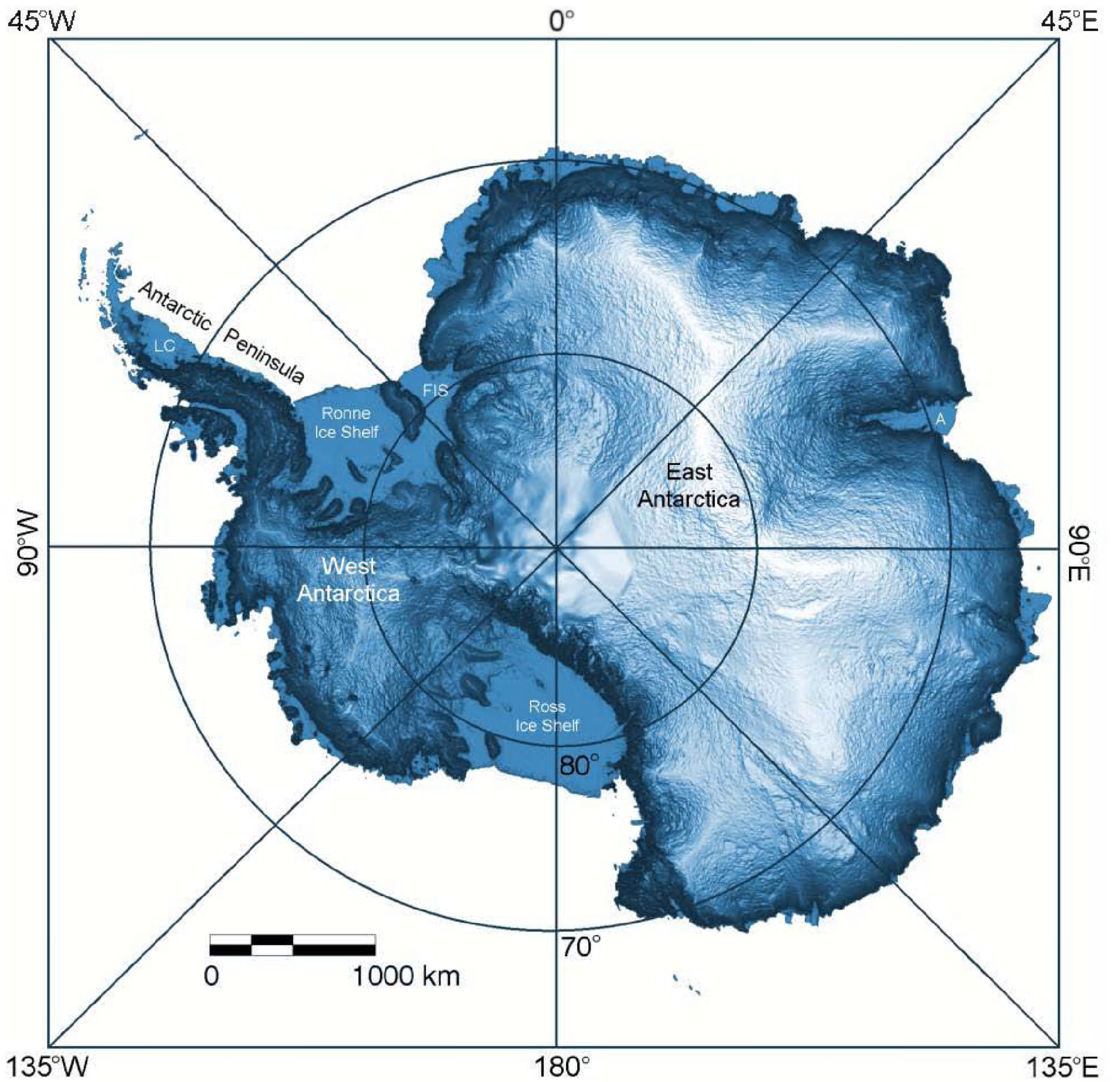


Figure 1.3 Surface elevation illuminated from directly overhead shows the general shape of the continent as well as the smaller scale roughness. Topographic divides between major catchments are bright (white) sinuous ridges. Fringing ice shelves are extremely flat (shown as pale grey matt). Smaller scale roughness is often associated with subglacial relief. The smoother surface surrounding the South Pole is the result of sparser and less accurate elevation data south of 86°S. (from Bamber et al., 2008).

The East Antarctic ice sheet (EAIS) comprises by far the largest part of the ice sheet. Lying primarily in the Eastern Hemisphere, it is bounded clockwise by the Ronne Ice Shelf near 30ºW, the Transantarctic Mountains, and the coast of Victoria Land, which lies along the 165ºE meridian on the western side of the Ross Sea) (Figures 1.1, 1.2, 1.3). The EAIS contains the coldest ice and is frozen to the bed over much of its area, which restricts its rate of flow. However, the bed in many places is at the pressure melting point as a consequence of the thick blanket of ice and heat flow from the Earth beneath, giving rise to some 145 subglacial lakes (Siegert et al., 2005).

The EAIS lies on a landmass predominantly above sea level with a few broad basins depressed below sea level largely due to the weight of the ice sheet (Figure 1.4).

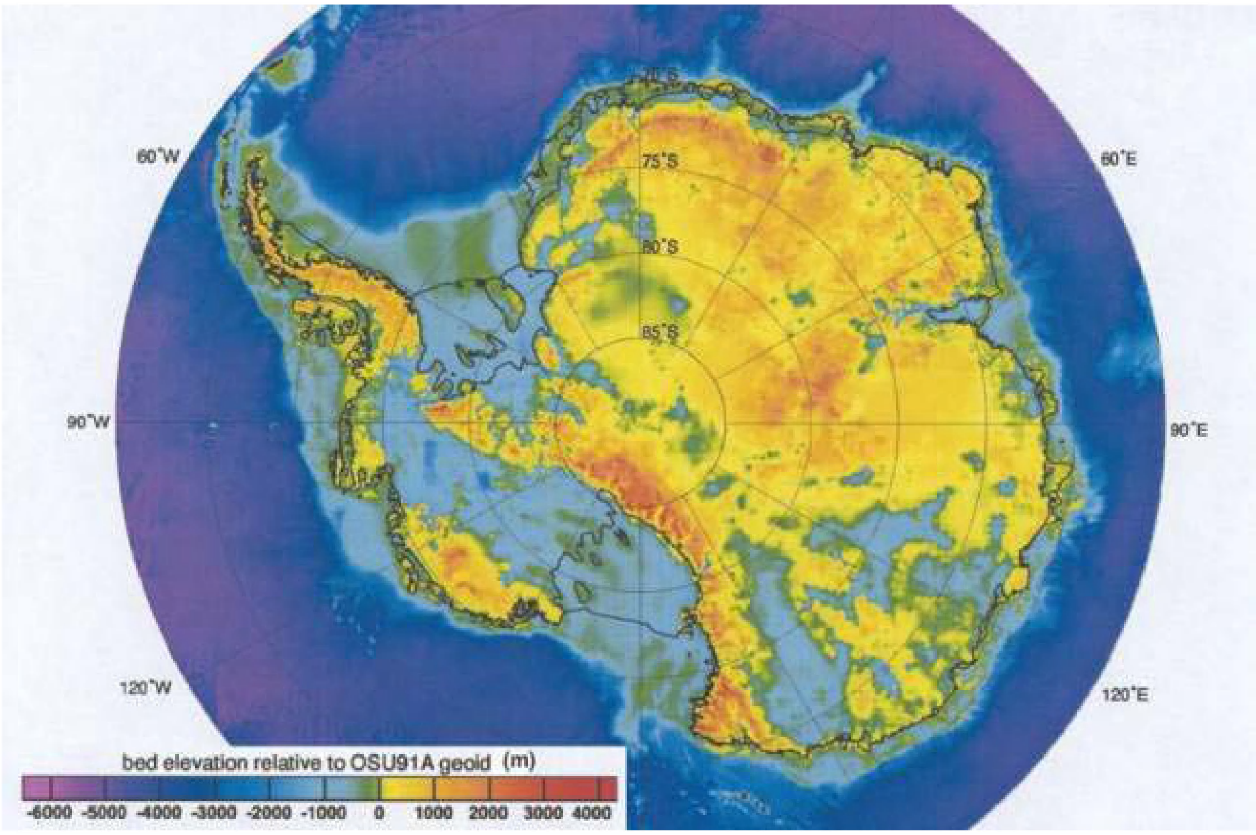


Figure 1.4 Bed elevation illustrating major regions (in blue and green) below sea level, the major subglacial continent underneath East Antarctica, and the mountain ranges separating East and West Antarctica and along the Antarctic Peninsula. Note the topography shown here has not been corrected for isostatic rebound. Taken from Lythe, Vaughan and BEDMAP Consortium, 2001.

The West Antarctic ice sheet (WAIS) comprises the ice that lies mostly in the Western Hemisphere and occupies the region from the Transantarctic Mountains counter clockwise through Marie Byrd Land to 150^{o}W (but not including the Antarctic Peninsula) (Figure 1.1). In contrast to the EAIS, most of the WAIS rests on a bed that is substantially below sea level (Figure 1.4), and would remain so even if the ice sheet were removed. For this reason the

WAIS is described as a "marine ice sheet" and is considered inherently unstable, being vulnerable to collapse if it lost its fringing ice shelves (e.g. see Bamber et al., 2009). In many places, the bed beneath the WAIS is over one thousand metres below sea level – far deeper than most of the world's continental shelves. A few of its buried archipelagos (Ellsworth Mountains, Executive Committee and Flood Ranges, Whitmore Mountains and the Ellsworth Land coast – not shown here) are sufficiently high to rise above the ice sheet. The WAIS is generally warmer, both at the surface and at the bed, than its East Antarctic neighbour, with basal ice close to the melting point in most areas.

The ice sheet that covers the Antarctic Peninsula is quite different from either the EAIS or the WAIS. It consists of much smaller and much thinner ice caps that cover the central mountainous spine and some of the larger outlying islands. These ice caps drain into the sea through relatively narrow, but steep and fast-moving, alpine-type glaciers. In contrast to the WAIS and the EAIS, which lose mass primarily through iceberg calving and melt from the base of ice shelves, the Antarctic Peninsula experiences much higher summer temperatures making runoff from surface melt a significant component in the budget of its ice sheet.

The ice sheets are nourished at their surface by deposition of snow and frost, which remains frozen because of the year-round cold, and accumulates year-on-year. As the surface snows are buried by new snowfall, they are compressed and eventually transformed into solid ice, a process that captures a chemical record of past climates and environments. From this record in East Antarctica it is possible to reconstruct Antarctica's contribution to planetary climate change over the past 800,000 years.

The ice sheet is up to 4,776 m thick in Terre Adélie (at around 140ºE in East Antarctica), averaging 1,829 m (Figure 1.5). In places the deepest ice may be more than one million years old; older ice is likely to have been so squeezed out by compression from above that it is in practical terms undateable as well as being very thin. Remnants of much older glaciers survive in parts of the McMurdo Dry Valleys, with ice at one location dated from volcanic ash at over 8 million years (Sugden et al., 1995). However, that age is in dispute (Hindmarsh et al., 1998).

All ice flows out from the central ridges and domes to the edge of the continent, converging as ice streams and outlet glaciers that move at speeds of up to 500 m per year (Figure 1.6). The entire Antarctic ice sheet transports ice from the interior to the coast at a rate of around 2000 billion tonnes per year. Once the ice streams reach the edge of the continent they either calve into icebergs or float on the ocean as ice shelves. The ice shelves constitute 11% of the total area of the Antarctic, with the two largest being the Ronne-Filchner Ice Shelf in the Weddell Sea and the Ross Ice Shelf in the Ross Sea, which have areas of 0.53×10^6 km^2 and 0.54×10^6 km^2 respectively (Figure 1.6). The ice shelves are several hundreds of metres thick and the ocean areas under them are among the most isolated and unvarying environments on Earth. Nevertheless, ocean currents carry water masses into the cavities beneath the ice shelves. There the water interacts with the undersides of the ice shelves either melting the ice or, at times, adding to it by refreezing. The cooled ocean water becomes fresher with the addition of meltwater and emerges from beneath the ice shelves in this modified condition (Nicholls et al, 2009).

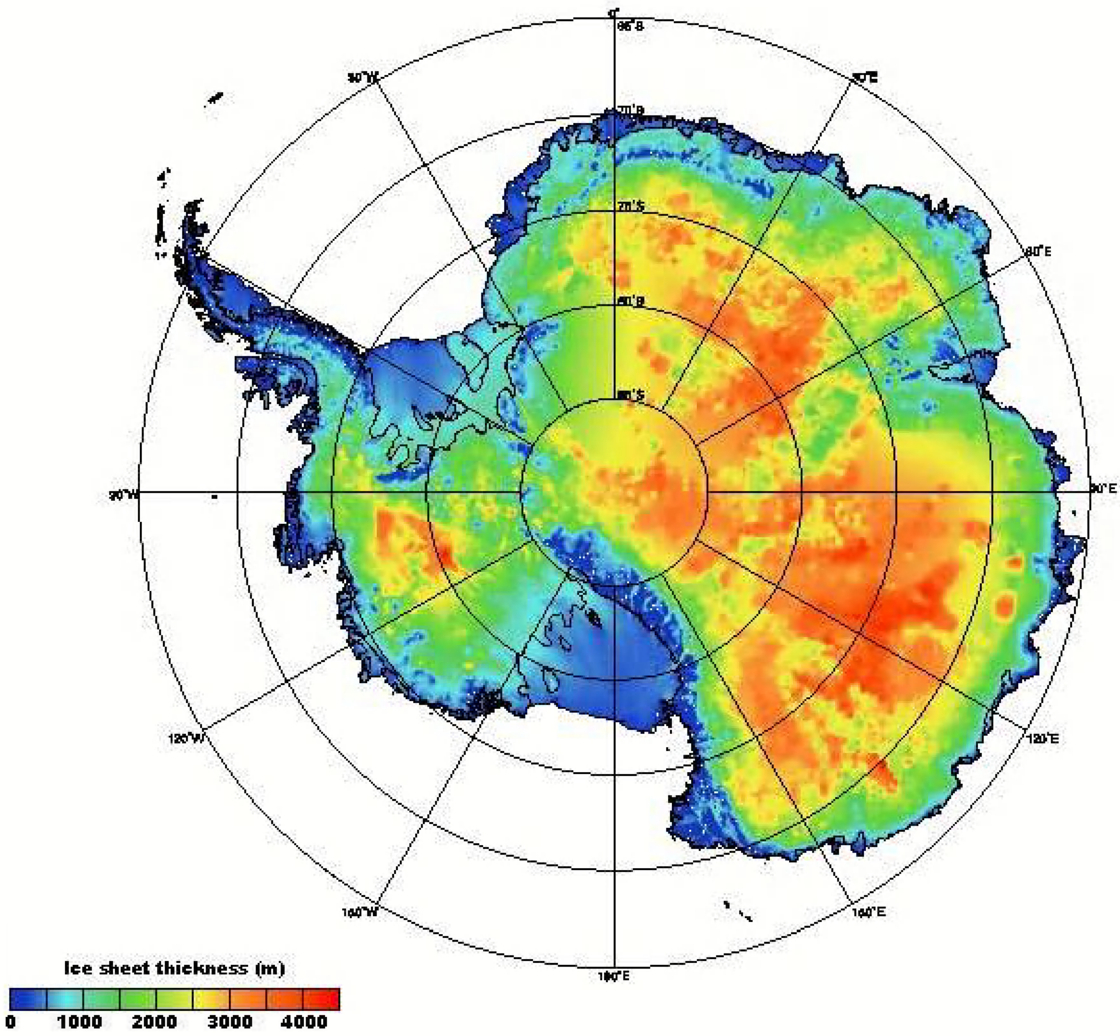


Figure 1.5. Ice thickness (difference between surface and bed elevations) in metres, showing the much thicker East Antarctic ice sheet, gradual thinning from the interior to the coast and the many deep outlet glaciers. Seawards of the continental outline, the dark blue areas are the permanent ice shelves that represent floating extensions of the continental ice sheet. From Lythe, Vaughan and BEDMAP Consortium, 2001.

The continent is surrounded for most of the year by a zone of frozen seawater 1 or 2 m thick (Figure 1.7). By late austral winter, this sea ice covers an area of 20×10^6 km^2, more than the area of the continent itself. At this time of year the northern edge of the sea ice is close to 60°S around most of the continent, and near 55°S north of the Weddell Sea. Unlike the Arctic, most of the Antarctic sea ice melts during the austral summer, so that by autumn it only covers an area of about 3×10^6 km^2. Most Antarctic sea ice is therefore thin first year ice, with the largest area of multi-year ice being over the western Weddell Sea.

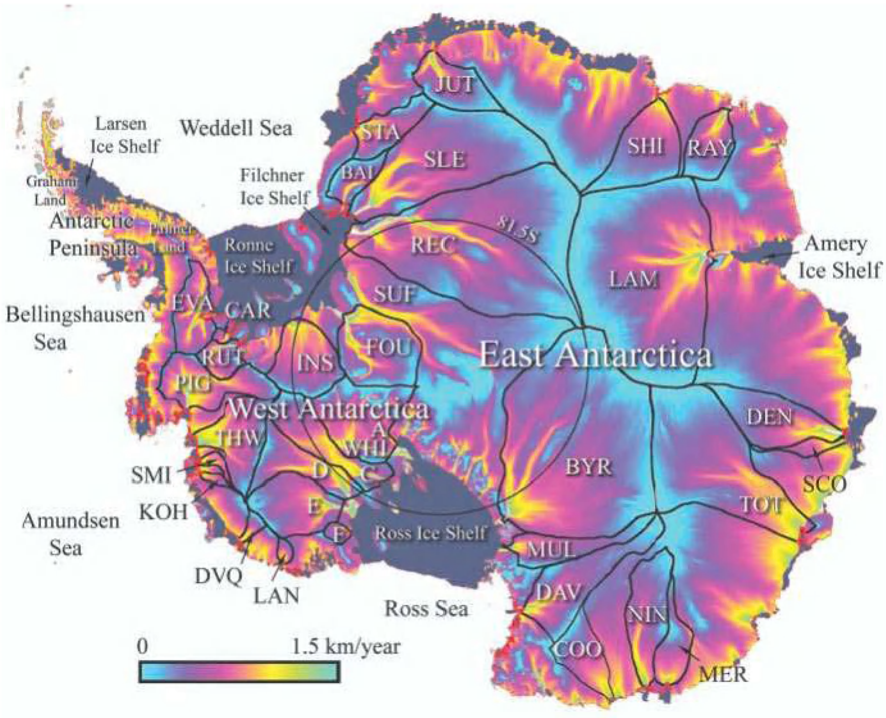


Figure 1.6 Balance velocity, calculated at any point as the velocity (averaged over the thickness) required to balance upstream accumulation, to illustrate the spatial pattern of ice flow that is required to maintain the ice sheet shape in the present climate (Rignot and Thomas, 2002).

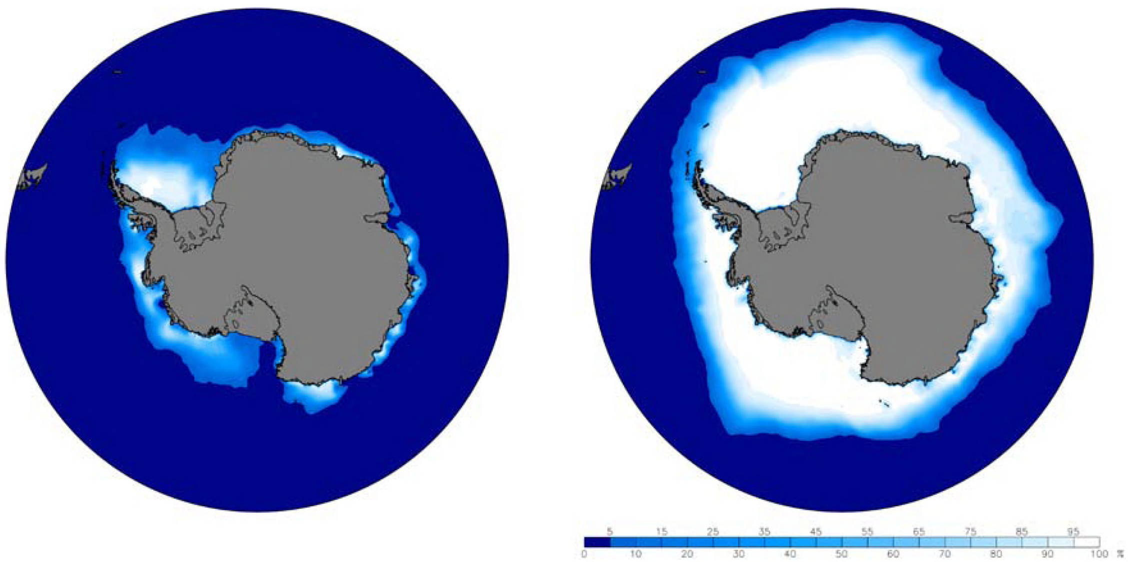


Figure 1.7. Mean 1979 – 2007 sea ice extent. Left, February and right, September. From NSIDC (http://nsidc.org/cgi-bin/bist/bist.pl?config=seaice_index.)

From time to time substantial areas of ice-free water can form like lakes within the sea ice. These are polynyas, which may occur where relatively warm water rises to the surface or where winds drive sea ice away from the coast (Renfrew et al, 2002). The latter kind of polynyas are 'factories' for the continuous formation of sea ice, which deposits salt in the water making it dense enough to sink and contribute to the formation of new deep water masses (Markus et al, 1998).

An important feature of any polar environment is permafrost, defined as any earth material that remains below 0ºC for two or more consecutive years. The uppermost layer, which seasonally can experience temperatures above 0°C is called the 'active layer'. Permafrost is much less extensive in the Antarctic compared to the Arctic, because of the very large area covered by the major ice sheets. Permafrost underlies most of the ice-free surfaces, some parts of the ice sheet and the glaciated areas, and even some spots of the adjacent sea floor. It is widespread in continental Antarctica, across parts of the Antarctic Peninsula and also in the islands of maritime Antarctica. Here, permafrost is a significant feature of the environment and it is important in the support of terrestrial ecosystems. Permafrost is also found in the McMurdo Dry Valleys and along the narrow coastal zone of East Antarctica. The McMurdo Dry Valleys have lakes with open water in the summer, and the Wright Valley has the Onyx River, which flows for about 2 months each year. Melting snow patches on rock form small runnels at elevations up to around 2,000 m there. The active layer is typically 30-50 cm thick. Permafrost ranges from 240 to 900 m thick in continental Antarctica, and 15 to 180 m thick in maritime Antarctica, and occurs only sporadically in the South Shetland and South Orkney Islands. The temperature of permafrost is –12 to –24°C in continental Antarctica but is likely warmer in maritime Antarctica.

1.3 The Role of the Antarctic in the Global Climate System

The global climate system is driven by solar radiation, most of which, at any one time, arrives at low latitudes. Over the year as a whole the Equator receives about five times as much radiation as the poles, creating a large Equator-to-pole temperature difference. The atmospheric and oceanic circulations respond to this large horizontal temperature gradient by transporting heat polewards (Trenberth and Caron, 2001). In fact the climate system can be regarded as an engine, with the low latitude areas being the heat source and the polar regions the heat sink. Although the dynamics of atmosphere-ocean interaction are well known, the complexities of the heat exchange engine and the interactions of ice shelves and land ice with the ocean and atmosphere make predictions about climate change a challenge.

The combination of tropical heating, poleward moving air and the Coriolis Force imposed by the Earth's rotation, leads to the development of the Hadley Cell, an atmospheric circulation in which air rises at the equator, creating the tropical belt of low pressure, and descends in the subtropics, forming the subtropical high pressure belt. At higher latitudes (60-65°S) the air ascends again, creating a low-pressure zone. The pressure gradient at the Earth's surface between the high pressure in the subtropics and the low pressure at 60-65°S forces air to move eastwards under the influence of the Earth's rotation, creating the mid-latitude westerlies that help to drive surface waters east in the ACC. The air ascending at 60-65°S moves poleward at upper levels and sinks again over the poles, forming a high-pressure system over the Antarctic continent. The pressure gradient from the low pressure at 60-65°S to the high pressure over the continent gives rise to easterly winds along the Antarctic coast, driving the westward-directed Antarctic Coastal Current over the continental shelf; that shelf current tends to be focused along the fronts of ice shelves (Deacon, 1937), Heywood et al, 2004, Ismael Núñez-Riboni and Fahrbach, 2009). The main westward flow is that associated with the Antarctic Slope Front, over the continental slope. The north-to-south distribution of surface pressure around Antarctica is subject to remarkable variability in the intensity of the meridional pressure gradient and its zonal location. Due to the circumpolar character of this variation it is called the Southern Annular Mode (of variability) (SAM) (See Trenberth and Jones (2007) for a review). The SAM can also be seen as a measure of the intensity of the westerly winds that propel the ACC. The SAM is the southern hemisphere equivalent of the Arctic Oscillation (also known as the Northern Annular Mode) or the related North Atlantic Oscillation, which is measured from the pressure difference between the Azores and Iceland. Variations in the SAM (which can be thought of as variations in a North - South pressure gradient) drive variability in the Southern Ocean's winds and currents: the steeper the gradient, the stronger the winds. In addition to the SAM, there are other significant modes of variability with meridional or zonal patterns.

Both the atmosphere and ocean play major roles in the poleward transfer of heat (Trenberth and Caron, 2001), with the atmosphere being responsible for 60% of the heat transport, and the ocean the remaining 40%. In the atmosphere, heat is transported by both low-pressure systems (depressions) and the mean flow. Clockwise-circulating depressions carry warm air poleward on their eastern sides and cold air towards lower latitudes on their western flanks. The atmosphere is able to respond relatively quickly to changes in the high or low latitude heating rates, with storm tracks and the mean flow changing on scales from days to years. The process by which air arrives at the poles also imports pollutants from industrialised areas, though quantities are tiny compared with the Arctic, not least because most industrialised areas are in the north, and the mean flow is zonal around Antarctica rather than more meridional - as in the Arctic.

The oceans carry heat and salt south towards the pole in upper ocean surface currents moving down the western sides of the Atlantic, Pacific and Indian Ocean basins (e.g. Schmitz, 1995, Lumpkin and Speer, 2007, Figure 1.8). In addition, heat and salt move south through the Atlantic Ocean in the sub-surface in North Atlantic Deep Water, which rises to the surface near the Antarctic coast (Figures 1.8 and 1.9), as a consequence of upwelling driven by the divergence of surface water forced north by the westerly winds of the Southern Ocean (e.g. Rintoul et al, 2001). In addition to their transport by the large-scale currents, heat and salt are transported by mesoscale eddies (with diameters of tens to hundreds of kilometers). In the Southern Ocean, where the meridional currents are normally small, meridional transport by eddies is significant (e.g. de Soeke and Levine, 1981; Bryden, 1979; Hughes and Ash, 2001; Rintoul et al, 2001; Hogg et al., 2008).

Eastward circulation is focused in the ACC, which lies broadly between the Southern Boundary Front (SB) and the Sub-Antarctic Front (SAF) (Rintoul, 2001, compare Figures 1.2 and 1.9). The core of the ACC lies roughly beneath the core of the predominant westerly winds. The ACC is an outstanding feature in the global ocean's circulation. Stretching over a length of around 20,000 km, it is the only current to completely encircle the globe. It transports around 140×10^6 m^3 of water per second (140 Sverdrups), making it the world's largest ocean current. And it links the three main ocean basins (Atlantic, Pacific and Indian) into one global system by transporting heat and salt from one ocean to another.

The westerly winds of the Southern Ocean act on the surface waters, which are forced north under the influence of the Coriolis Force of the Earth's rotation. Deep water from below wells up to replace these surface waters, as is evident from Figure 1.9. This ascent (e.g. Schmitz, 1995; Lumpkin and Speer, 2007) from the deep in the ACC forms the upward part of the global overturning circulation. The northward moving surface water is cold and dense and sinks at the Polar Front to form Antarctic Intermediate Water (Wüst, 1935), which sinks northward at intermediate depths to permeate the word's oceans especially in the Southern Hemisphere (McCartney, 1982). Some of the surface water reaches the SAF, where it sinks to form Sub-Antarctic Mode Water (Hanawa and Talley, 2001). Due to the excess of precipitation over evaporation at these latitudes, the surface water gets lighter as is moves north (gain of buoyancy in Figure 1.9), which explains how the Mode Water (lighter) comes to overlie the Intermediate Water (heavier), despite their having the same source. South of the divergence zone, where upwelling takes place, the surface waters reaching coastal seas are cooled by contact with ice, and pick up salt rejected when surface waters freeze to form sea ice, so losing buoyancy (Figure 1.19) and becoming dense enough to sink. The sinking waters cascade down the continental shelf (Baines and Condie, 1998; Foldvik et al. 2004) and slope to form Antarctic Bottom Water, which spreads around Antarctica (Orsi et al, 1999) and aerates most of the global deep ocean floor (e.g. Wüst, 1935; Hogg, 2001). These sinking cold waters provide a fairly uniform cold environment for bottom dwelling (benthic) organisms.

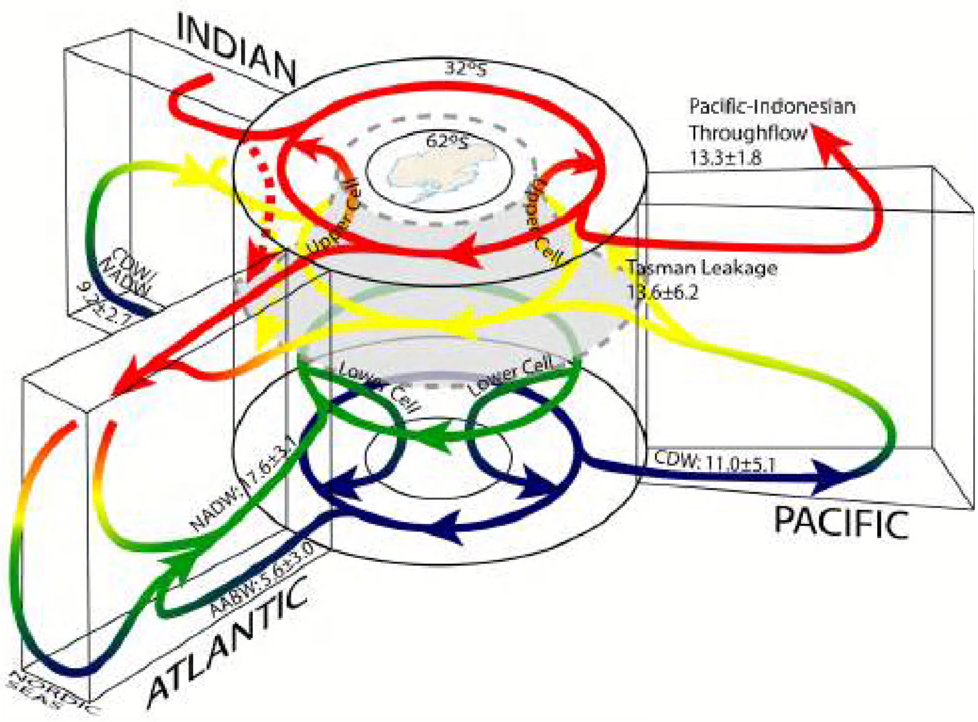


Figure 1.8 Model of the global ocean circulation, emphasising the central role played by the Southern Ocean. From Lumpkin and Speer (2007). NADW = North Atlantic Deep Water; CDW = Circumpolar Deep Water; AABW = Antarctic Bottom Water. Units are in Sverdrups (1 Sv = $10^6 \times m^3$ of water per second). The two primary overturning cells are the Upper Cell (red and yellow), and the Lower Cell (blue, green, yellow). The bottom water of the Lower Cell (blue) wells up and joins with the southward flowing deep water (green or yellow), which connects with the upper cell (yellow and red). This demonstrates the global link between Southern Ocean convection and bottom water formation and convective processes in the Northern Hemisphere.

Jets along the SAF and the PF typically carry a large fraction of the transport of the ACC (Figures 1.9 and 1.10; and Cunningham et al., 2003), but other fronts can have comparable transports across individual hydrographic sections. In fact, the ACC is not a single front but a complex system of fronts (e.g. Sokolov and Rintoul, 2002), several of which are thought to be of circumpolar extent (Orsi, et al., 1995). This complex system approach provides a new paradigm for considering the response of the ACC to climate change.

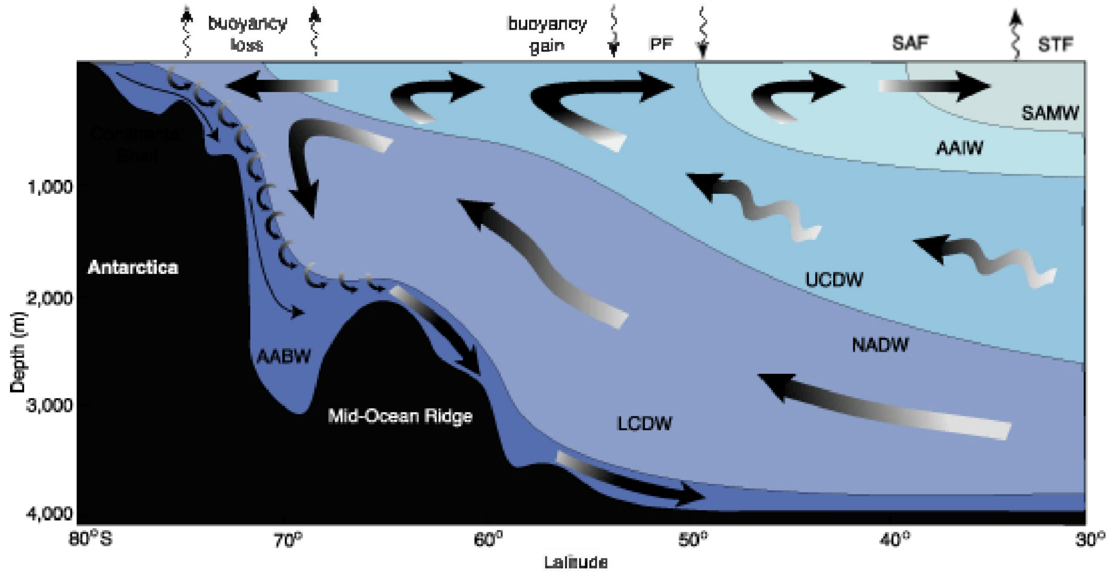


Figure 1.9 South (left) to north (right) section through the overturning circulation in the Southern Ocean. South-flowing products of deep convection in the North Atlantic are converted into upper-layer mode and intermediate waters and deeper bottom waters and returned northward. Marked are the positions of the main fronts (PF – Polar Front; SAF – Sub-Antarctic Front; and STF – Subtropical Front), and water masses (AABW – Antarctic Bottom Water; LCDW and UCDW, Lower and Upper Circumpolar Deep Waters; NADW – North Atlantic Deep Water; AAIW – Antarctic Intermediate Water and SAMW – Sub-Antarctic Mode Water) (from Speer et al., 2000). Note that as well as water moving north to south or vice versa, it is also generally moving eastward (i.e. towards the observer in the case of this cross section), except along the coast where coastal currents move water westward (away from the observer).

Superimposed on the circumpolar circulation system are regional clockwise gyres, principally the Weddell Gyre and Ross Gyre (Figure 1.10), whose southern boundaries are the west-moving Slope Current, and whose outer limits reach to the east-moving ACC. These gyres constitute the detailed pathway for transport to, from, and along the continental margin. They are visible in traditional hydrographic surveys as dome-shaped structures surrounded by downward sloping masses of equal density that extend towards the coast in the south and towards the ACC in the north. They are also clearly revealed in Southern Ocean numerical model outputs. Estimates of the extent of transport in the gyres are few. The Weddell Gyre carries 30 +/- 10 Sverdrups (Sv) in the Weddell Sea (Fahrbach et al., 1994) and 56 +/- 8 Sv across the Greenwich Meridian (Klatt et al, 2005). The Ross Gyre transports 40 Sv across longitude 150°W, and the Australian-Antarctic Gyre 76 +/- 26 Sv across longitude 110°E (McCartney and Donohue, 2007).

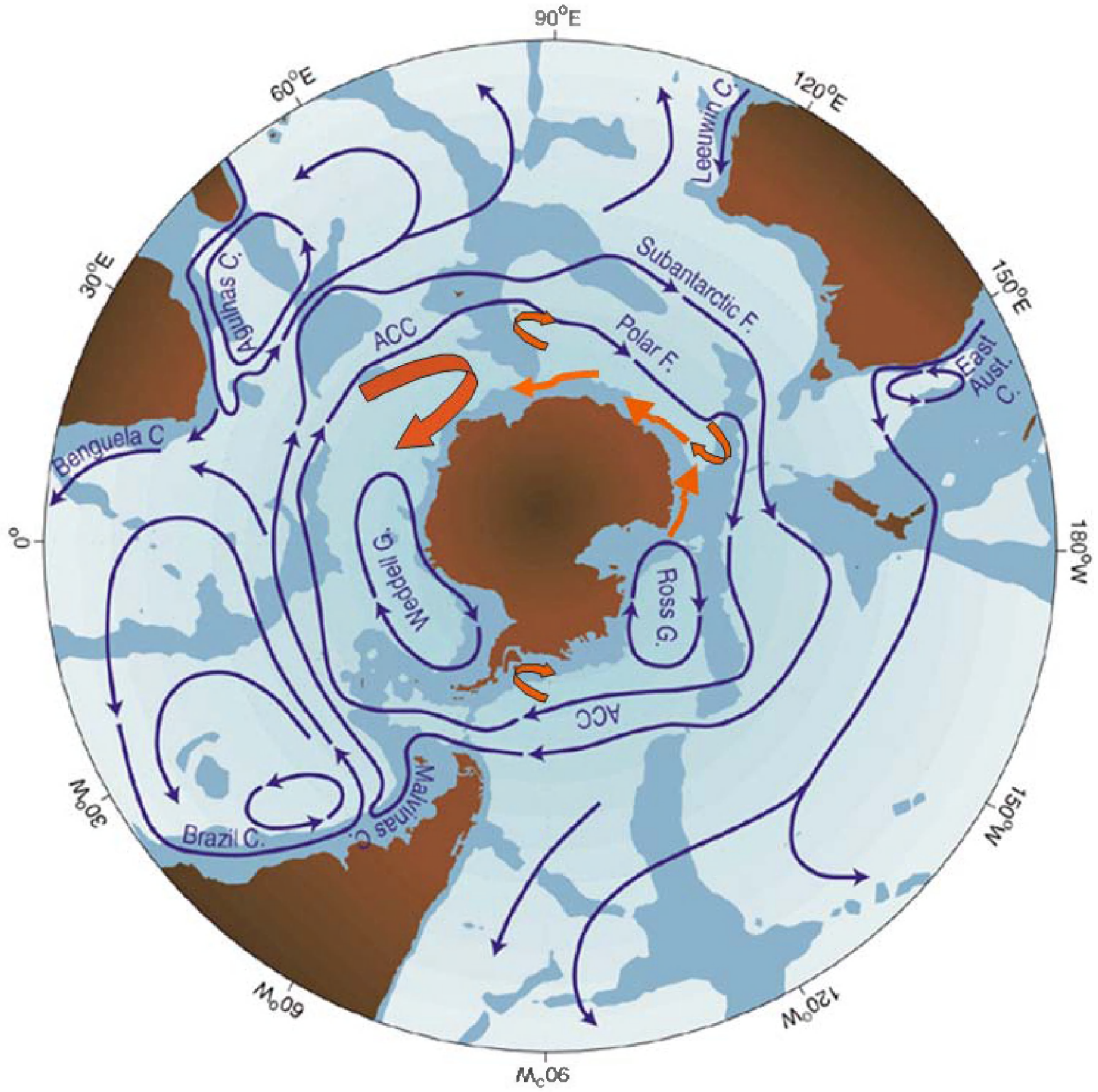


Figure 1.10 Schematic map of major currents south of 20ºS (F = Front; C = Current; G = Gyre) (Rintoul et al, 2001); showing (i) the Polar Front and Sub-Antarctic Front, which are the major fronts of the Antarctic Circumpolar Current; (ii) Other regional currents; (iii) the Weddell and Ross Sea Gyres; and (iv) depths shallower than 3,500m shaded (all from Rintoul et al, 2001). In orange are shown (a) the cyclonic circulation west of the Kerguelen Plateau, (b) the Australian-Antarctic Gyre (south of Australia), (c) the slope current, and the (d) cyclonic circulation in the Bellingshausen Sea, as suggested by recent modelling studies (Wang and Meredith, 2008), and observations – e.g. eastern Weddell Gyre - Prydz Bay Gyre (Smith et al, 1984), westward flow through Princess Elizabeth Trough (Heywood et al, 1999), and circulation east of Kerguelen Plateau (McCartney and Donohue, 2007).

The distribution of land and sea in the two polar regions is responsible for the very different atmospheric and oceanic circulations observed in each. Antarctica is a continent surrounded by ocean, while the Arctic is an ocean surrounded by land. The Antarctic continent is a sub-circular dome lying over the pole and surrounded by a broad swath of deep ocean, apart from the slight constriction of Drake Passage between the Antarctic Peninsula and South America (Fig. 1.10). As a result, the mean atmospheric flow and surface ocean currents are zonal in nature (parallel to latitude). The ACC, which is the major oceanographic feature of the Southern Ocean, flows unrestricted around the continent (Figures 1.2 and 1.8), isolating the cold high latitude land and sea areas from warm temperate mid- and low-latitude surface waters. This has only been the case since about 30 million years ago when the Drake Passage opened up, as discussed in greater detail in Chapter 3. Before then the ocean currents had more of a meridional component (parallel to longitude), which allowed greater poleward penetration of warm waters from temperate latitudes. In contrast, as mentioned earlier, surface ocean circulation in the Arctic tends to be meridional, especially the northward flow of warm water at the surface in the North Atlantic Current.

The Southern Ocean plays a key role in the global carbon cycle. The upwelling deep water south of the Polar Front (Figures 1.8 and 1.9) brings to the surface dissolved nutrients and carbon dioxide (CO_2), and releases this gas to the atmosphere. In contrast, the Intermediate Water and Mode Water masses sinking north of the Polar Front (Figure 1.9) take up CO_2 from the atmosphere. These complementary processes make the Southern Ocean both a source and a sink for atmospheric CO_2 (see Figure 1.11; from Sabine et al., 2004).

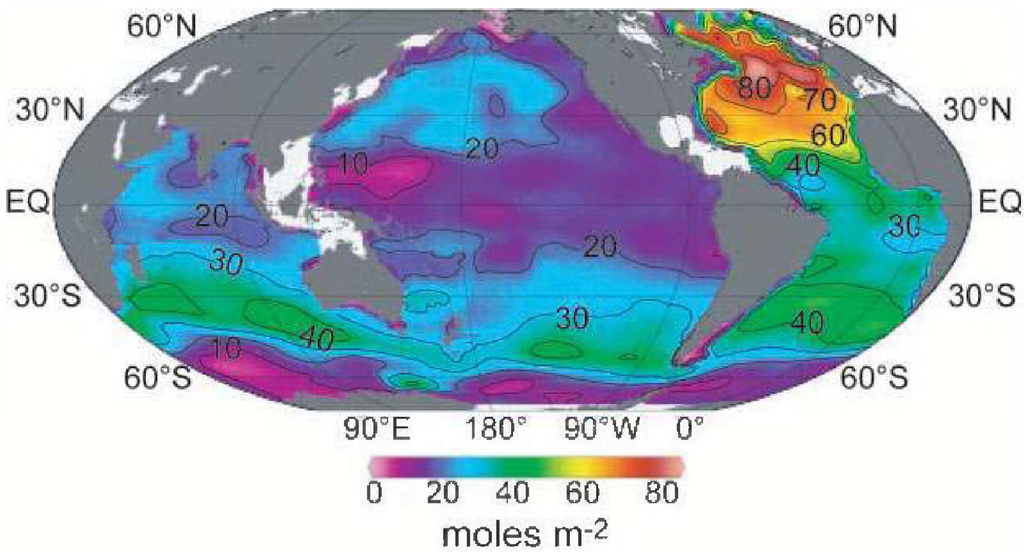


Figure 1.11 Column inventories of anthropogenic CO_2 in the ocean. Dissolved old CO_2 is lost to the atmosphere south of the Polar Front, where NADW wells up to the surface close to the coast (purplish colours); most of this old CO_2 is not anthropogenic. The cold sinking Mode and Intermediate Waters north of the polar front (green colours) have extracted considerable anthropogenic CO_2 from the atmosphere (from Sabine et al., 2004).

Because of its upwelling nutrients, the Southern Ocean is the world's most biologically productive ocean. Nonetheless, its productivity is limited by the low availability of micronutrients such as iron, except around the islands that are scattered through the ACC. As a result the Southern Ocean is classified as High Nutrient Low Chlorophyll (HNLC). Through photosynthesis, the growth of phytoplankton extracts CO_2 from the atmosphere and pumps it to the seabed or into subsurface waters through the sinking of decaying organic matter. Without this process, and without the solution of carbon dioxide in cold dense sinking water near the coast, the build up of carbon dioxide in the atmosphere would be much faster.

CO_2 is exchanged between the ocean and the atmosphere primarily through air-sea fluxes, which are highly variable in space and time (Mahadevan, et al., 2004; Volk and Hoffert, 1985) balancing to within 2% (net) when integrated globally (Watson and Orr, 2003).

Surface ocean CO_2 levels (pCO_2) and their related atmospheric CO_2 levels can be described as being controlled by the combination of physical and biological processes that move CO_2 from the upper ocean into the deep ocean. These processes can be separated into the "physical pump" and the "biological pump" (Figure 1.12). The biological pump is the most effective at removing carbon from the system through burial in sediments. Both pumps are effective at transferring CO_2 to deep water.

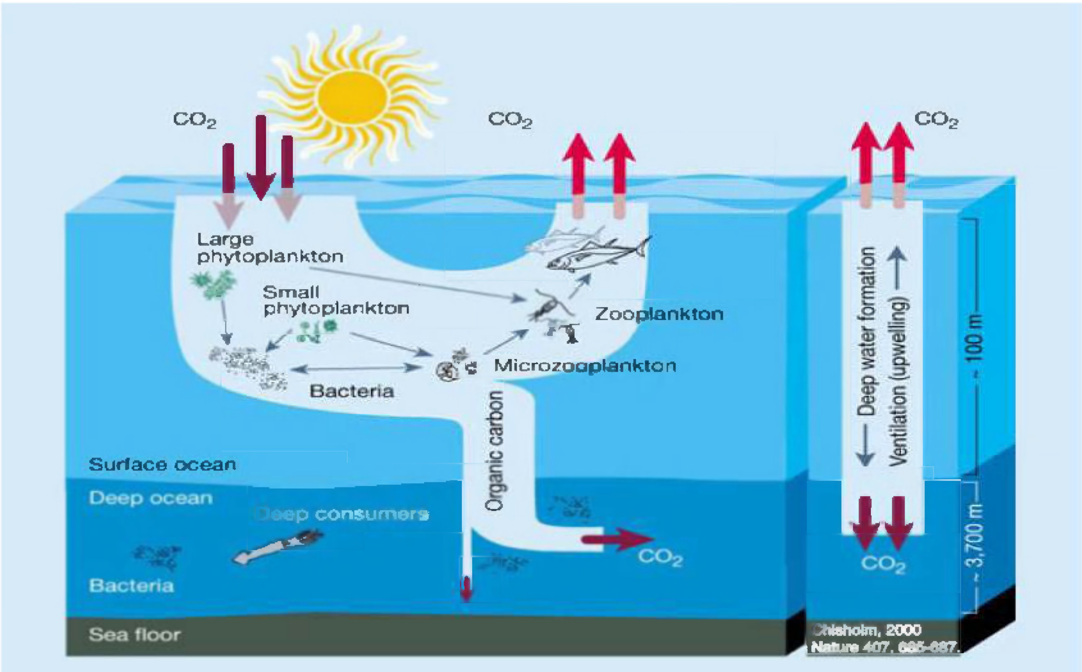


Figure 1.12 A simple schematic representation of the biological pump (left) and physical pump (right) (Chisholm, 2000). Reprinted by permission from Macmillan Publishers Ltd: Nature doi: 10.1038/35037696, copyright 2000.

The biological pump refers to the biological cycling of ocean carbon into the ocean interior. It is a complex process operating over timescales from hours to months, and is dependent on the physical processes of ocean mixing and transport (Anderson and Totterdell, 2004). The first stage is fixing dissolved inorganic carbon (DIC) into dissolved organic carbon (DOC) by photosynthesis in the euphotic zone; the amount of carbon fixed is crudely proportional to the nutrients available. Most of this DOC is then processed, consumed and recycled by the marine ecosystem within the euphotic zone. Some of the organic material in the euphotic zone is exported, sinking under gravity to be either remineralised by the benthos (converted back into DIC) or buried in the seafloor as sediment. As a rule the biological pump always acts to reduce surface seawater pCO_2 levels.

The physical pump describes the role of ocean dynamic and thermodynamic processes in changing the distribution of dissolved carbon between the upper ocean and the deep ocean.

Dynamical processes that can impact the distribution of CO_2 in the upper ocean include changes in the rate or volume of upwelling or downwelling waters, and buoyancy or wind driven mixing. The thermodynamic response of the carbonate system is such that at a fixed atmospheric pCO_2, the upper ocean (in contact with the air) can hold more dissolved inorganic carbon at cooler ocean temperatures than at warmer temperatures.

The carbon in the ocean is a mixture from both natural and anthropogenic sources. The anthropogenic fraction comes from human-induced emissions of CO_2 into the atmosphere that have continued at an increasing rate since the start of the industrial revolution and in 2007 were approaching 10 PgC/yr (where 1 petagram (Pg) = 10^{15} grams CO2 equivalent, which is equal to 10^9 metric tons C) (Canadell, et al., 2007). It is estimated that 30% of total anthropogenic emissions annually are taken up from the atmosphere and sequestered by the ocean (Sabine et al., 2004). Regardless of source (natural or anthropogenic) the carbon in the ocean follows the carbon cycle described above.

1.4 Observations for Studies of Environmental Change in the Antarctic

As we shall see in more detail in Chapters 2 and 4, lack of observations has long been a problem for scientists working in the Antarctic. The early expeditions were mainly carried out during the brief Antarctic summer, and there is still a bias towards summer observations in some fields – for instance many of the research vessels investigating oceanographic conditions in the Southern Ocean are not equipped to work in the ice-strewn seas of the austral winter, and 24 hour darkness restricts work on land at that time. A major advance was the establishment of research stations that operated year-round. Many stations were constructed around the time of the International Geophysical Year in 1957/58. Most of them still operate today, though not all are year-round. Those with year-round capability monitor many aspects of the Antarctic environment, such as meteorology, solar-terrestrial interactions in the Earth's outer atmosphere (known as 'geospace'), coastal sea ice conditions, and sea level. These observations have produced instrumental records that are in many cases 50 years long, providing important data sets for global change studies.

Most research stations are located in the Antarctic coastal region, with only Amundsen-Scott station at the South Pole and Vostok and Concordia on the polar plateau providing *in-situ* observations from the interior. The advent of satellite remote sensing, beginning with the launch of the meteorological satellite TIROS-1 in April 1960, was therefore crucial for obtaining data from the data-sparse interior of the continent. Initially only visible and infrared imagery were available, but increasingly advanced instruments have been developed that have provided a rich source of data for studies in meteorology, glaciology, geology and observing the surface of the ocean.

To compensate for the difficulties of humans collecting data everywhere on the continent and year-around, there has been a significant rise in the use of automated recording instruments in remote parts of the Antarctic. Automatic weather stations have been deployed at many isolated locations since the early 1980s, to provide data for weather forecasting and climate change studies. Such instruments have sometimes been coupled with equipment to provide upper atmosphere and geospace observations, which is very operationally efficient. Biological studies have also increased, including environmental monitoring at small, biologically relevant scales, although such datasets are typically of short duration. There have been few attempts to link environmental observations across different scales of measurement or between scientific disciplines. Ocean regions have long been a data void for sub-surface oceanographic observations, in particular during winter, but new autonomous systems like Argo floats now complement the subsurface moored systems that provided data over the last two decades in the form of quasi-continuous time series at selected locations. Floating systems are now providing frequent measurements from the subsurface across wide areas, and have the potential to revolutionise our understanding of ocean conditions.

While satellites and autonomous systems can provide valuable data, in some fields the primary forms of data are still collected using *in-situ* techniques. Ice and sediment cores are the main means of obtaining high-resolution paleoclimate information for the Antarctic. Short cores can be fairly easily collected and provide high horizontal and vertical resolution coverage through the last few hundred years. Deeper ice cores, like those collected at Vostok and Dome C, have provided new insights into the glacial cycles of much of the last million years. Even earlier periods can be investigated with sediment cores from ocean areas, although the resolution there is coarser. Over-snow traverses such as those during IGY and more recently the 21-nation International Trans Antarctic Scientific Expedition (ITASE) serve effectively as polar research vessels. These traverses have now covered many tens of thousands of kilometres. They offer the ground-based opportunities of traditional style traverse travel coupled with the modern technology of GPS navigation, crevasse detecting radar, remote sensing, satellite communications and multi-disciplinary research. By operating as a ground-based transport system, over-snow traverses offer scientists the opportunity to experience the dynamic environment under study.

Naturally, most biological studies involve *in-situ* measurement and observation, although here again there is some application of satellite or autonomous data, and further potential in this field such as in the monitoring of vegetation extent and the identification of vertebrate colony locations.

In Chapter 2 we describe the various kinds of measurements being made, and the advances that they have led to.

1.5 The Climate of the Antarctic and its Variability

In the summer, when the sun is above the horizon for long periods, the Antarctic receives more solar radiation than the tropics, but the highly reflective ice- and snow-covered surfaces reflect much of this radiation back to space, aided by a relatively cloud-free atmosphere that contains little water vapour. This reflection is one of the important feedback mechanisms found in the ice-covered polar regions, because it enhances cooling. Where snow melts, exposing large patches of bare (dark) ground, or where sea ice melts exposing ice free (dark) ocean, solar radiation will be absorbed rather than reflected, and the environment will warm.

The Polar Plateau of East Antarctica experiences very low temperatures because of its high elevation, the lack of cloud and water vapour in the atmosphere, and the isolation of the region from the relatively warm maritime air masses found over the Southern Ocean. The very cold temperatures in the interior of Antarctica year-round, coupled with its isolation from warm, moist air masses, mean that precipitation there is very low, with only about 5 cm water equivalent falling per year (King and Turner, 1997). That makes much of Antarctica a desert and the driest continent on Earth. The low temperatures mean that there is very little evaporation and sublimation, so that although the amount of precipitation is small, it builds up year by year to form the ice sheet. Many blizzards tend to be fallen snow resuspended, rather than new snow.

Temperatures are much less extreme in the Antarctic coastal region than on the plateau. In general, at most of the coastal stations the monthly mean summer temperatures never rise above freezing point, although daily temperatures may show excursions above freezing in places during summer. There are exceptions. The highest temperatures on the continent are found on the western side of the Antarctic Peninsula where there is a prevailing northwesterly wind; there temperatures can rise to several degrees above freezing during the summer, and monthly means are positive for 2-4 months of the year. Temperatures also tend to be above freezing at times in places like the Schirmacher Oasis near the coast in Dronning Maud Land, where there are ice-free lakes in the austral summer.

Because of the lack of incoming solar radiation, the Antarctic stratosphere in winter is extremely cold. A strong temperature gradient develops between the continent and mid-latitudes (Figure 1.13), isolating a pool of very cold air above Antarctica. Very strong winds develop along this thermal gradient. They are stronger than the equivalent winds found in the Arctic, because the Equator-to-pole temperature difference is larger in the south. The pool of cold air and its strong surrounding winds together form the polar vortex. It plays an important part in determining the atmospheric circulation of the high southern latitudes, as well as in the formation of the ozone hole, where the polar vortex acts as a 'containment vessel' allowing chlorofluorocarbon compounds (CFCs) to build up during the winter.

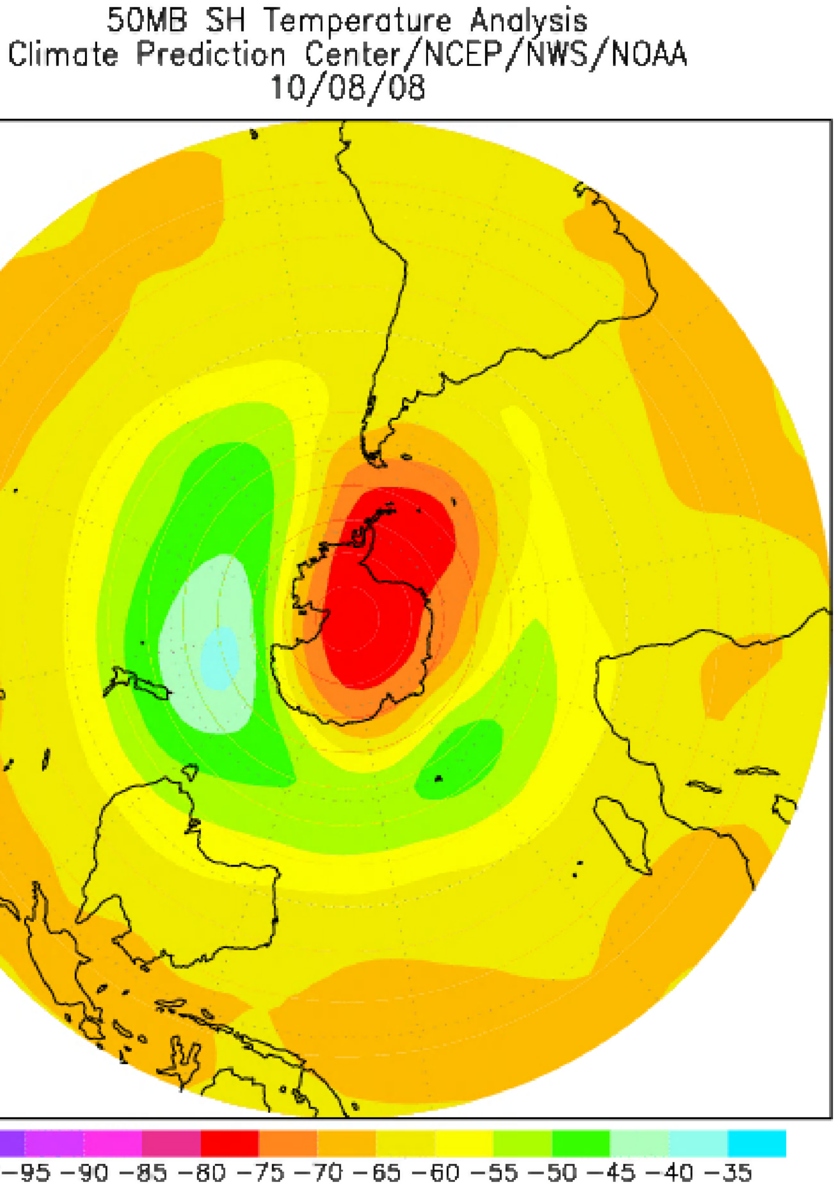


Figure 1.13 The polar vortex above Antarctica (indicated by the red colours) in mid-winter (August) as seen via the temperatures in degrees C on the 50 hPa pressure surface (roughly 20 km elevation above mean sea level). Generated from the NCEP reanalysis using their online, interactive chart drawing system at: www.cdc.noaa.gov/data/reanalysis/reanalysis.shtml.

The Antarctic climate system varies on time scales from the sub-annual to the millennial and is closely coupled to other parts of the global climate system. On the longest time scales it fluctuates on Milankovitch frequencies (20 ka, 41 ka, 100 ka) in response to variations in the Earth's orbit around the sun that cause regular variations in the Earth's climate on these recurrent time scales. We discuss the variability of the Antarctic climate system on these longer timescales in Chapter 3.

Proxy data from ice cores show that since the Last Glacial Maximum (LGM) at about 21 ka before present (BP) there have been a number of climatic fluctuations across the continent. One of the most marked was the Mid Holocene warm period, which is present in various records from Antarctica. Ice cores also reveal shifts in the intensity of the atmospheric circulation, with the westerlies weakening at 5,200-5,400 years ago and strengthening around 1,200 years ago.

Because long time-series of observations did not begin until the IGY (1957-58), the instrumental period in Antarctic is only about 50 years long (measurements started before then but were not continuous). Since proxy data show oscillations on longer time scales than 50 years, it is accepted that the instrumental period only provides a snapshot of change in the Antarctic. Nevertheless, it shows the complexity of change and a mix of natural climate variability and anthropogenic influence, as discussed in Chapters 2 and 4.

Reliable weather charts for high southern latitudes have only been available since the late 1970s. They reveal a great deal about the patterns in the atmospheric circulation of high southern latitudes. The most pronounced climate variability over this period is evident in changes in the SAM. Over the last few decades there has been a marked drop in pressure around the Antarctic coast and an increase in mid-latitudes. The increase in the pressure gradient across the Southern Ocean has strengthened the surface winds, which in turn has affected ocean currents and the distribution of sea ice. Studies of coupled ocean-atmosphere general circulation models show that the strength of the SAM should increase as the Earth warms, confirming what is known from observations.

But the strength of the SAM is also modulated by the ozone hole over Antarctica. Stratospheric ozone is an important constituent of the upper atmosphere above the Antarctic, where ozone levels began to decline in the late 1970s following widespread releases of CFCs and halons into the atmosphere. We now know that the presence of CFCs in the Antarctic stratosphere results in a complex chemical reaction during the spring that destroys virtually all ozone at altitudes between 14 and 22 km, especially within the polar vortex where temperatures are coldest. The depletion in ozone maintains the cooling within the polar vortex, which accentuates the winds around the vortex. This strengthening of the winds at high altitude in spring then propagates downwards in the atmosphere through time, strengthening surface winds during the summer and autumn. Thus one effect of the ozone hole has been to accentuate the SAM signal, further strengthening the westerly winds around the Antarctic during the summer and autumn.

In the strengthened SAM, with its stronger winds, we see two examples of changes from outside the region having a profound impact on the Antarctic environment. One impact comes from the CFCs responsible for the 'ozone hole', and the other comes from greenhouse gas emissions responsible for global warming. Both CFCs and greenhouse gases have mainly been released in the Northern Hemisphere during the industrial era. Both have had, and continue to have, a profound effect on the radiation balance of the Antarctic atmosphere.

Although the Antarctic is far removed from where most of the solar energy enters the system at tropical latitudes, it is still influenced by variability in tropical conditions, and signals of low latitude climate variability can be identified in Antarctica and the Southern Ocean (Turner, 2004). In addition, there is increasing evidence that signals can also be transmitted in the opposite direction from high to low latitudes.

Statistically significant long-range linkages between high and low latitudes go by the name of teleconnections. They can take place via the atmosphere and/or the ocean, although the timescales are usually rather different. The most rapid teleconnections generally occur via the atmosphere, with storm track changes occurring on the scale of days or weeks. The El Niño-Southern Oscillation (ENSO) is one of the largest climatic cycles on Earth, functioning on the scale of years to decades. It has its origins in the tropical Pacific Ocean, but its effects can be felt across the world. As discussed in a number of places in this volume, ENSO signals can be identified in the physical and biological environment of the Antarctic, although some of the links are not robust and there can be large differences in the extra-tropical response to near-identical events in the tropics.

The circulation of the upper layers of the ocean can change over months to years, but the deep ocean and the global thermohaline circulation (THC) require decades to centuries to respond. At the other extreme, fast wave propagation in the ocean takes place on timescales of just a few days.

About 30 years of reliable atmospheric reanalysis outputs are available for the high southern latitudes, so it has been possible to establish the nature of the broad-scale teleconnections typical of the Southern Hemisphere. There is evidence of decadal timescale variability in some of these linkages, but with such a short data set it is not possible at present to gain insight into how the teleconnections may vary on longer timescales. High resolution ice core records collected from areas where the rate of precipitation of snow is high shed some light on teleconnections over the century timescale, and where the rate of accumulation is high enough they can even give seasonal data, which is important since some teleconnections are only present in individual seasons.

Links between the climates of the northern and southern hemispheres can be found, but they vary with time as discussed in Chapter 4. Through most of the Holocene (roughly the past 12,000 years) northern hemisphere events have lagged behind southern hemisphere ones by several hundred years. That has changed in recent decades, and the northern hemisphere signal of rising temperature since about 1850 AD has paralleled the southern hemisphere one. Temperature change in the two hemispheres now appears to be synchronous - a radical departure from former times, which suggests a new and different forcing, most likely related to anthropogenic activity in the form of enhanced greenhouse gases. Comparing data on winds and temperatures from northern and southern hemisphere ice cores confirms that the wind/temperature fields of today differ from those of the recent past indicating that the modern day atmosphere is not an analogue for that of the so-called Medieval Warm Period.

Changes associated with the changing climate are now plainly evident in the Antarctic, and will be discussed in detail in Chapters 3 and 4. They range from the glaringly obvious, like the collapse of the Larsen B ice shelf in February-March 2002, and the warming of various parts of the Antarctic Peninsula over the past 50 years (King, 1994), to the rather more subtle, like the 0.2°C warming of the Southern Ocean, which - though a small amount - represents a major transfer of heat when summed over a vast area.

1.6 Biota of the Antarctic

In the Earth System, the poles fulfil a very special role. Their slowly changing physico-chemical features have engineered life processes so that organisms surviving the ensuing severe selection can prosper in such extreme habitats. It is unreasonable to investigate life in Earth's extreme environments without also addressing the impacts of current climate changes on organisms whose adaptations to climatic conditions have slowly evolved over geological time to reach a required equilibrium. This equilibrium is delicate, and bound to be upset by rapid changes such as those that are occurring at the present time.

Evolution is the major unifying principle of biology, and evidence of the evolutionary process pervades all levels of biological organisation from molecules to ecosystems. Although the influence of evolution extends to the fauna and flora of the most isolated continent, Antarctica is not usually included among notable evolutionary sites such as the Galapagos, Hawaii, Australia, Madagascar, the East African Great Lakes and Lake Baikal (as discussed by Eastman, 2000). In fact, however, the evolution of the marine fish group Nototheniidae on the Antarctic continental shelves ranks equally to the species radiations at these other sites (see below). The exploration of remote habitats such as Antarctica has proved to be far more valuable than merely documenting the existence of unusual faunas and floras. Discoveries made at the Galapagos stimulated Charles Darwin to begin erecting the framework for evolutionary thought. Darwin's voyage in HMS *Beagle* (1831-36), and subsequent research by numerous scientists, has made the Galapagos the premier evolutionary site in the world. At about the same time (1839-43), James Clark Ross with HMS *Erebus* and *Terror*, and other explorers (Wilkes and Dumont d'Urville) explored the high latitudes of Antarctic coastal waters. During these voyages many distinctive species were discovered, including an endemic tribe of four seals, six species of penguins and about a dozen species of notothenioid fish, which more recent research has shown to be unique. A comprehensive picture of the fauna did not emerge and Antarctica remained for decades unappreciated as a continental island with an endemic marine fauna. Antarctica has now joined the Galapagos as a destination of tourist ships and fishing vessels, but unlike the Galapagos, whose unique fauna is mostly terrestrial, it still does not have the recognition it deserves as a centre of evolution.

In that context it is perhaps not surprising that the unique fish fauna of the Antarctic's continental shelf, and especially that of the Ross Sea, the largest of Antarctica's shelves (see Eastman, 2005, and Eastman and McCune, 2000) is somewhat under-appreciated. Like the fresh-water fishes of Lake Baikal or the birds of the Galapagos and Hawaii, most Ross Sea fishes are derived from a common ancestor, and are now recognized within a single group, the notothenioids. Unlike elsewhere in the world, the Eocene fauna of the Antarctic continental shelf, particularly the Ross Sea, has been completely replaced by a modern fauna dominated by this single group. In isolation and in the absence of competition they have come to dominate the Ross Sea's species diversity, abundance and biomass at levels of 77%, 92% and 91%, respectively. This is unique in the marine realm. They are also one of the few examples of a marine "species flock". Some of the most recent molecular phylogenies suggest that the neutrally buoyant water column group, including toothfish (*Dissostichus mawsoni*) and silverfish (*Pleuragramma antarcticum*), was the first clade (group) to diversify into this habitat about 15-20 million years ago.

The Antarctic and its biota command increasing attention in a world attuned to changes in global climate, loss of biological diversity and depletion of marine fisheries. The links between long and short-term global climate change and evolution are among the least understood natural events in the history of the Earth. Excellent examples are available for study in the Antarctic, but will be lost if areas of the Southern Ocean are not protected in the same way that terrestrial systems have been protected in a network of Specially Protected Areas. For instance, the nototheniid fauna of the northern insular shelves, e.g. Scotia and Kerguelen, has been devastated by industrial fishing (Kock, 1992, 2007). Understanding the impact of past, current and predicted environmental change on biodiversity and the consequences for Antarctic-ecosystem adaptation and function is a primary goal. The critical examination of Antarctic ecosystems undergoing change provides a major contribution to the understanding of evolutionary processes of relevance to life on Earth. How well are Antarctic organisms able to cope with daily, seasonal and longer-term environmental changes? Will climate change result in relaxation of selection pressure on genomes, or tighter constraints and ultimately extinction of species and populations? The Antarctic holds great potential for evolutionary studies, through which evolutionary biology can play an important role in understanding the biological response to climate change within the whole Earth system.

There is evidence that climate change and modifications of the Earth system occur at faster rates than elsewhere in the polar regions. The uniquely adapted fauna of these regions is vulnerable to shifts in climate. Therefore, it is urgent to establish the state of these communities, and in particular their diversity, as well as to protect them from direct human impacts (e.g. fishing), if we are to understand the impacts of climate change. These impacts will have many types of consequences, among which the loss of biodiversity is of highest concern. To reliably assess the extent of future changes in polar ecosystems, a mass of data should be aggregated to identify undisputable evidence of change in ecosystem structure/functioning/services. Sound science plans, efficient data management and an unprecedented collaborative research effort in the International Polar Year framework and beyond will provide scientists, environmental managers and decision-makers with a solid benchmark against which future changes can reliably be assessed. Studies of the biodiversity of polar marine organisms, coupled to sound data handling and dissemination, within a network of protected areas, will bring a better understanding of how life has evolved in the marine environment of the poles, and to what extent it may potentially respond to change. Bridges between different research disciplines and international programmes will not only bring a crucial support to polar science, but will provide a legacy of knowledge for future generations in the form of a sustainable information system.

All organisms and the communities to which they belong are shaped by both ecological and evolutionary factors. On longer time scales, marine, lake, and terrestrial assemblages reflect the influence of evolutionary events, invasions, extinctions, tectonics and climatic change. Evolutionary processes in the biosphere have developed efficiently under specific Antarctic conditions for the past 30 million years, and it is these processes that are most closely linked to future developments, since it is adaptation that governs the survival or extinction of species and communities. On shorter time scales organisms and communities are shaped by ecological factors such as dispersal, predation, competition, habitat, disturbance, food supply, and pure chance.

1.6.1 Terrestrial

Looked at from the perspective of the terrestrial and freshwater biologists and biogeographers, the various bits of land are subdivided for the purposes of this volume into three zones - the ‘continental Antarctic’ (comprising most of the continent), the ‘maritime Antarctic’ (comprising the Antarctic Peninsula and associated islands and archipelagos as well as the South Shetland, South Orkney, and South Sandwich Islands, and Bouvetøya), and the ‘sub-Antarctic’ (those islands that lie in or around the Antarctic Polar Frontal Zone (PFZ).

Levels of terrestrial macro-biodiversity in the Antarctic are strikingly lower than those of the Arctic, although the sub-Antarctic hosts greater diversity than the maritime and continental regions. This is the case even relative to the superficially environmentally extreme and isolated High Arctic Svalbard and Franz Josef archipelagos at around 80°N. In comparison with about 900 species of vascular (higher or flowering) plants in the Arctic, there are only two on the Antarctic continent and up to 40 on any single sub-Antarctic island. Likewise, the Antarctic and sub-Antarctic have no native land mammals, against 48 species in the Arctic. The continuous southwards continental connection of much of the Arctic is an important factor underlying these differences. Despite the apparent ease of access to much of the Arctic, few established alien vascular plants or invertebrates are known from locations such as Svalbard (Rønning, 1996; Coulson 2007), in comparison with the 200 species introduced to the sub-Antarctic by human activity over only the last two centuries or so (Frenot et al., 2005, 2008). It may be the case that species comparable to the many sub-

Antarctic 'aliens', being cosmopolitan northern hemisphere and boreal 'weeds', have had greater opportunity to reach polar latitudes by natural means in the north than the south.

Antarctic and sub-Antarctic floras and faunas are strongly disharmonic, with representatives of many major taxonomic and functional groups familiar from lower latitudes being absent. Sub-Antarctic plant communities do not include woody plants, and are dominated by herbs, graminoids and cushion plants; flowering plants (phanerogams) are barely represented (two species) in the maritime and not at all in the continental Antarctic. Sub-Antarctic floras have developed some particularly unusual elements – 'megaherbs are a striking element of the flora of many islands, being an important structuring force within habitats, and a major contributor of biomass (Meurk et al., 1994a,b; Mitchell et al., 1999; Fell, 2002; Shaw, 2005; Convey et al., 2006). These plants present an unusual combination of morphological and life history characteristics (Convey et al., 2006), and their dominance on sub-Antarctic islands is thought to have been encouraged by a combination of the absence of natural vertebrate herbivores (Meurk et al., 1994a; Mitchell et al., 1999), and possessing adaptive benefits relating to the harvesting and focussing of low light levels and aerosol nutrients (Wardle, 1991; Meurk et al., 1994b). The recent anthropogenic introduction of vertebrate herbivores to most sub-Antarctic islands has led to considerable and negative impacts on megaherb-based communities (Frenot et al., 2005; Shaw et al., 2005; Convey et al., 2006b).

The tables below provide summary information on terrestrial biodiversity in the Antarctic.

Zone	Flowering plants	Ferns and club-mosses	Mosses	Liverworts	Lichens	Macro-fungi
sub-Antarctic	60	16	250	85	250	70
maritime Antarctic	2	0	100	25	250	30
continental Antarctic	0	0	25	1	150	0

Table 1.1 Biodiversity of plant taxa in the three Antarctic biogeographical zones. Note that figures presented are approximate, as it is likely that (i) new species records will be obtained through more directed sampling, (ii) a significant number of unrecognized synonymies are likely to exist and (iii) taxonomic knowledge of some Antarctic groups is incomplete.

Group	Sub-Antarctic	Maritime Antarctic	Continental Antarctic and continental shelf
Protozoa *	83		33
Rotifera *	> 59	> 50	13
Tardigrada	> 34	26	19
Nematoda *	> 22	28	14
Platyhelminthes	4	2	0
Gastrotricha	5	2	0
Annelida (Oligochaeta)	23	3	0
Mollusca	3/4	0	0
Crustacea (terrestrial)	4	0	0
Crustacea (non-marine)	44	10	14
Insecta (total)	210	35	49
(Mallophaga	61	25	34)
(Diptera	44	2	0)
(Coleoptera	40	0	0)
Collembola	> 30	10	10
Arachnida (total)	167	36	29
(Araneida	20	0	0)
(Acarina *	140	36	29)
Myriapoda	3	0	0

Table 1.2 Biodiversity of native terrestrial invertebrates in the three Antarctic biogeographical zones. Data obtained from Block, in Laws (1984), Pugh (1993), Pugh and Scott (2002), Pugh et al. (2002), Convey and McInnes (2005), Dartnall (2005), Dartnall et al. (2005), Greenslade (2006), Maslen and Convey (2006). ND - number of representatives of group unknown; * - large changes likely with future research due to current lack of sampling coverage, expertise and/or synonymy.

Representing the animal kingdom, across the Antarctic and sub-Antarctic there are no native land mammals, reptiles or amphibians and very few non-marine birds. Instead, terrestrial faunas are dominated by arthropods, including various insects, arachnids, the microarthropod groups of mites and springtails, enchytraeids, earthworms, tardigrades, nematodes, beetles, flies and moths, with smaller representation of some other insect groups (Gressitt, 1970; Convey, 2007a). Although levels of species diversity are low relative to temperate communities, population densities are often comparable, with tens to hundreds of thousands of individuals per square metre. Few of these invertebrates are thought to be true herbivores, and the decomposition cycle is thought to dominate most terrestrial ecosystems, even in the sub-Antarctic, with the exception of some beetles and moths, although detailed autecological studies are typically lacking (Hogg et al., 2006). However, despite the preponderance of detritivores, decay processes are slow. Carnivores are also present (spiders, beetles on the sub-Antarctic islands, along with predatory microarthropods and other microscopic groups throughout), but predation levels are generally thought to be insignificant (Convey 1996).

Although microbial biodiversity is dominant in most Antarctic terrestrial and freshwater systems, these communities are generally considered to be relatively simple, with a limited trophic structure (Hogg et al., 2006). Furthermore, only a relatively small proportion, 0.33% (Fox and Cooper, 1994) of the continent's surface area is ice-free and available for terrestrial biota.

Terrestrial microbial diversity has been explored, although not extensively enough. Data are available particularly from the McMurdo Dry Valleys (Priscu et al., 1998; Gordon et al., 2000; Shravage et al., 2007; Babalola et al., 2009) and endolithic communities (de la Torre et al., 2003; de los Rios et al., 2007), the Pridz Bay area (Smith M. et al., 2000; Taton et al., 2006) and also from the Antarctic Peninsula (Hughes and Lawley, 2003). Terrestrial dark crusts are found throughout Antarctica and are commonly dominated by cyanobacteria (Broady, 1996; Mataloni and Tell, 2002; Adams et al., 2006). Freshwater systems have been sampled in studies of benthic habitats in continental Antarctic lakes (Bowman et al., 2000; Brambilla et al., 2001; Sabbe et al., 2004; Taton et al., 2003, 2006; Van Trappen et al., 2002, 2004, 2005). These studies have revealed a considerable amount of new biodiversity, concerning various eubacterial phyla including cyanobacteria. The diversity and function of the microbial lake communities have been reviewed by Ellis-Evans (1996). Molecular genetic tools allow specialist habitats such as hot mineral soils (Soo et al., 2009), cryoconites (Christner et al., 2003) and droppings of the Adélie penguins (Banks et al., 2009) to be studied in detail. Application of these tools to a diverse range of samples from across Antarctica should reveal more diversity and provide insights into distribution patterns, the forces that drive them, the presence and extent of endemism, and the impact of global change.

1.6.2 Marine

With the opening of the Drake Passage, separation of Antarctica from South America was complete. This event allowed the inception of the ACC and the establishment of the Polar Front, which acts as an efficient barrier for migration in both directions. Antarctica is an island continent, with the dominant fauna inhabiting the sub-zero water rather than the ice-covered landmass. Unlike other large marine ecosystems, the waters of the continental shelf around Antarctica resemble a closed basin, isolated from other shelf areas in the Southern Hemisphere by distance, current patterns and sub-zero water temperatures. As these isolating conditions developed over the past 30-40 million years, the marine fauna became adapted to a new shelf habitat and their ranges became highly circumscribed. Rates of endemism reach 97% in the case of some marine groups. The attention of evolutionary biologists is drawn to these isolated habitats because of their unusual faunas, which make the waters of the Antarctic shelf comparable to classic evolutionary sites such as the Galapagos.

Over the past 30-40 million years, the physico-chemical features of the Antarctic marine environment experienced a slow and discontinuous transition from the warm-water system of the early Tertiary (15°C) to the cold-water system of today (-1.87°C). As it separated from the other southern continents during the fragmentation of Gondwanaland and moved into its present polar location, Antarctica played a key role in altering ocean circulation and forcing the global climate toward cooling and glaciation. During that time there has been a nearly complete replacement on the Antarctic shelf of the diverse, cosmopolitan temperate fish fauna from the late Eocene by the highly endemic, cold adapted modern fauna. The Antarctic continental shelf offers several striking examples of faunal change and radiation, with fish and some invertebrate groups among the best studied.

1.6.2.1 Benthos and demersal fish

The Antarctic continental shelf is very deep in comparison with other continents, reaching 800 m in places, depressed by the weight on the continent of the massive ice sheet, and with troughs reaching to 1,000 m. More than 95% of the shelf is at depths outside the reaches of sea ice keels scouring the seafloor, wave action, scuba divers and sunlight or the photosynthetically active radiation (PAR) that illuminates the euphotic zone. Some 33% of the Antarctica's continental shelf is covered by floating ice shelves, the largely inaccessible areas below them belonging to the least known habitats on Earth.

The benthic marine fauna of Antarctica is comparatively well known thanks to historical surveys, modern international scientific initiatives such as European Polarstern Study (EPOS), Ecology of the Antarctic Sea Ice Zone (EASIZ), Evolution in the Antarctic (EVOLANTA), Census of Antarctic Marine Life (CAML), Latitudinal Gradient Project (LGP), Food for Benthos on the Antarctic Continental Shelf (FOODBANCS), and individual national projects, although at the regional scale many large gaps in survey data remain (e.g. Griffiths et al., 2009). This fauna currently comprises over 4,000 described species (White 1984; Arntz et al., 1997; Clarke and Johnston, 2003), although it has been estimated that the total macrofauna of the continental shelf may exceed 17,000 species (Gutt et al. 2004). Recent faunistic expeditions, especially those sampling the relatively unknown Antarctic deep-sea (Brandt et al., 2007), reveal tens to hundreds of putative new but so far undescribed species.

The modern Antarctic shelf-inhabiting benthic fauna is very much an epifauna of sessile filter and particle feeders associated with poorly sorted glacial substrates. In most systematic groups the fauna is highly endemic, eurybathic and possibly stenothermal. Figure 1.14 shows, contrary to an old paradigm, that species richness is quite high; the percentage of the Antarctic species richness as a proportion of global diversity is not much below Antarctica's percentage of the world's continental shelf. Two groups, the Antarctic polychaetes and pycnogonids, exceed this value and consequently, have an above global average species richness.

At its most abundant, Antarctica's benthic fauna is typified by dense, stratified communities of sponges, anemones, ascidians, gorgonians, hydroids, corals, bryozoans, cirripedes, crinoids and dedrochirote holothurians that often form a three-dimensional biogenic architecture below the zone of anchor ice formation and scour by the keels of sea ice (Figure 1.15; Arntz et al., 1994, 1997; Gutt, 2000). Associated with these sessile forms is a wandering fauna of ophiuroids, asteroids, echinoids, pycnogonids, isopods, amphipods, shrimps, nemerteans and gastropods. Aside from the dominant sessile suspension feeders we find assemblages in which the fauna benefits from deposited phytodetritus; these organisms include the filter feeding infauna, such as bivalves, or the vagrant deposit feeders, such as elasipode holothurians or ophiuroids (Figure 1.16, Gutt, 2007). Benthic communities cover a full range from an extremely high biomass of several kg wet-weight per m^2 to extremely low biomass, abundances and metabolic processes below ice shelves (Azam et al., 1979). Regional and local patchiness can be high due to differences in environmental conditions, especially food supply by currents, biological interactions, disturbance and maybe even incomplete recolonization after the past glaciation. Despite peculiarities in the fauna west of the Antarctic Peninsula, it seems that at coarse spatial resolution there can be said to be a circumpolar fauna typical of the Antarctic.

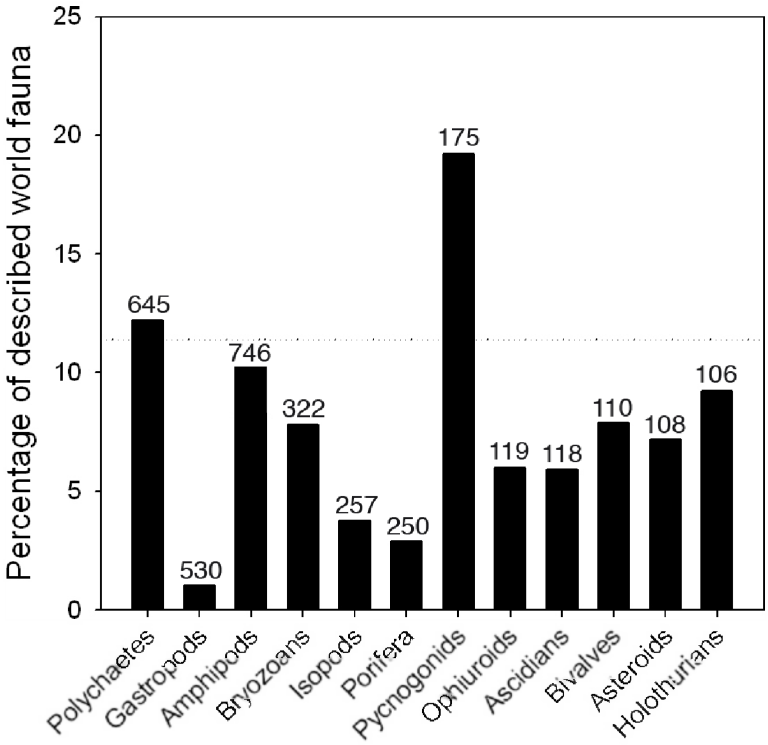


Figure 1.14 Percentage of Antarctic macrobenthic invertebrate species of the described world fauna. Systematic groups are arranged from left to right according to their species richness in the Antarctic indicated by the numbers above bars; the horizontal dotted line shows percentage of the world's continental shelf (data modified from Aronson et al., 2007 with additional information on amphipods from De Broyer et al., 2007).

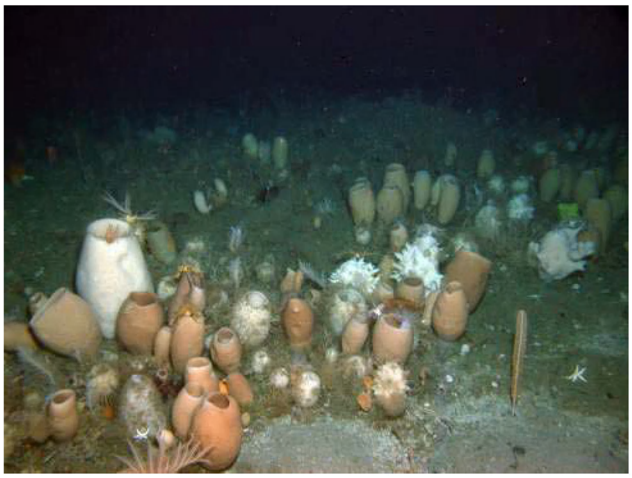


Figure 1.15 Rich megabenthic community on the shelf of the southeastern Weddell Sea at about 250 m depth, consisting mainly of suspension feeders such as glass sponges of the genus *Rossella* and *Scolymatra* with associated sea-fans and feather-stars. © J. Gutt, AWI

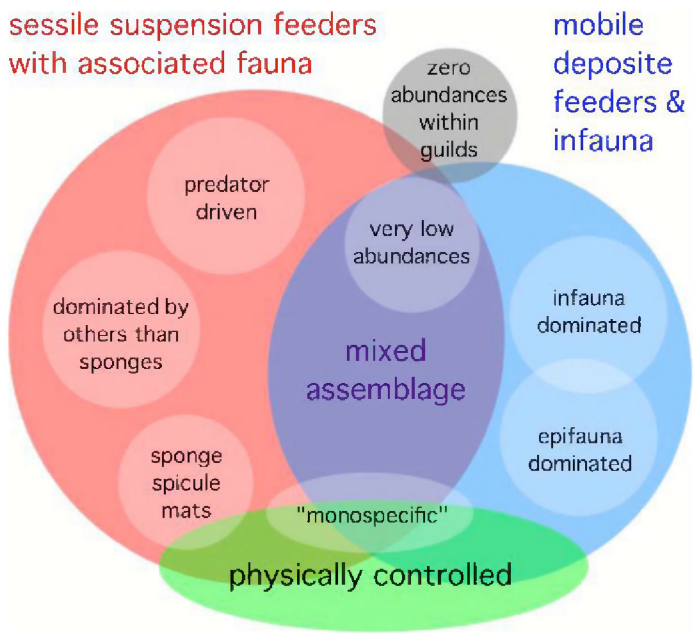


Figure 1.16. Classification of macro-benthic communities on the continental Antarctic shelf (changed after Gutt, 2007).

One of the most important features of the Antarctic benthic fauna is the lack of the durophagous (skeleton-breaking) predation that is a common characteristic of shallow waters elsewhere. Crabs, lobsters and sharks are essentially absent, and there is only a very limited diversity of ray-finned fish (teleosts) and skates (Aronson and Blake, 2001). The benthic fish fauna of Antarctica is no less remarkable in its taxonomic balance. With the exception of a small number of rays known from deeper waters around South Georgia, there are no selachians in the Southern Ocean. Furthermore, many teleost groups are almost completely absent and the fauna is dominated by striking radiations in two groups: the notothenioids, principally on the continental shelf, and the liparids in the deeper waters of the continental slope (Eastman, 1993, Eastman and Clarke, 1998).

The perciform suborder Notothenioidei, mostly confined within Antarctic and sub-Antarctic waters, is the dominant component of the Southern Ocean fauna. Indirect indications suggest that notothenioids appeared in the early Tertiary (Eastman, 1993) and began to diversify on the Antarctic continental shelf in the middle Tertiary, gradually adapting to the progressive cooling.

Densities and biomass of the Antarctic meiobenthos (45 µm to 1 mm) are often higher than in other parts of the world (Ingels et al., 2006; Gutzmann et al., 2004; Sebastian et al., 2007). In shelf and slope sediments, nematode communities are generally very diverse at genus level (Vanhove et al., 1999), while the abyssal diversity is comparable with other oceans (Sebastian et al., 2007). In contrast with many macrobenthic taxa (larger than 1 mm), no endemic nematode genera have been identified yet. While there may be no endemic nematode genera, there are endemic nematode species. Information at species level was until recently limited to some scarce taxonomical descriptions. In recent years high local and regional species richness has been revealed in a number of selected dominant nematode genera. Several species found in the Antarctic do show a wide geographical distribution, but most species investigated were new to science (Vermeeren et al., 2004; Fonseca et al., 2006; Ingels et al., 2006; De Mesel et al., 2007).

The present number of macroalgae described from Antarctica is 120 species (Wiencke and Clayton, 2002). Due to the remoteness of the continent and due to incomplete collecting this number is certainly an underestimate. Species numbers are highest in red algae (75) followed by brown and green algae (26 and 17 species, respectively), whilst golden-brown algae are only represented by one species. The proportion of species being endemic to the Antarctic region is highest among the brown and golden-brown algae (Heterokontophyta) (44%), and lowest among the green algae (18%).

1.6.2.2 Microorganisms

Studies of marine microbial diversity of the Southern Ocean are increasing (Murray and Grzymski, 2007). Approaches have both targeted cultivable heterotrophic bacteria (e.g. Michaud et al. 2004) and used molecular biological approaches. In the genome of the marine psichrophile *Pseudoalteromonas haloplanktis* TAC125, three genes encoding monomeric hemoglobins and one encoding a flavo-hemoglobin have been discovered (Giordano et al., 2007; Verde et al., In Press). This multiplicity is very rare; their physiological role is under investigation, and may be linked to cold adaptation. From a global ocean perspective, the microbial diversity is moderate and depends on the timing and location of sampling. Broadly, the phylogenetic composition of the microorganisms living in the open water and in sea ice is similar to that found in Arctic Ocean studies and is spread among the different classes of the Proteobacteria (in which Alphaproteobacteria and Gamaproteobacteria are most dominant), Bacteriodetes, and the Archaeal domain. Both Crenarchaeota and Euryarchaeota have been detected in numerous studies (DeLong et al., 1994, Murray et al., 1998, Church et al., 2003, Bano et al., 2004). Both activities and abundance (Pearce et al., 2008) as well as diversity vary significantly seasonally (Murray et al., 1998, Murray and Grzymski, 2007) in coastal waters off the Ingrid Christensen Coast (Prydz Bay, East Antarctica) and Antarctic Peninsula respectively. Summertime samples are dominated by Alphaproteobacteria in the Scotia Sea (Topping et al. 2006), whereas in regions influenced by sea ice melt in the Ross Sea Gammaproteobacteria and Bacteriodetes dominated (Gentile et al., 2006). The extent of diversity in the polar winter is less well known. New directions of research aim to describe microbial diversity associated with Antarctic marine invertebrates (Webster et al., 2004, Webster and Bourne, 2007, Riesenfeld et al., 2008), and the genome sequences of both cultivated isolates and environmental genome fragments (Grzymski et al., 2006) in oceanic and sea ice communities.

1.6.2.3 The pelagic system

The open ocean ecosystem of the Southern Ocean is well known for its latitudinal zonation. Three different large-scale subsystems coincide with the distribution of water masses and the ice cover (Hempel, 1985; Nichol and Alison, 1997; Arrigo, 1998; Brierly and Thomas, 2002; Thomas and Dieckmann, 2002, Arrigo and Thomas, 2004).

The northern zone occupies the ice-free part of the oceanic West Wind Drift, which is relatively poor in biomass and production. Primary production seems to be limited not by nutrients but by the trace element iron. Despite its High Nutrient – Low Chlorophyll (HNLC) character, there is abundant zooplankton biomass, especially near the Antarctic Convergence and decreasing towards lower latitudes within the West Wind Drift. Zooplankton is dominated by copepods, gelatinous salps, small euphausiids - relatives of the Antarctic krill (*Euphausia superba*), chaetognaths and amphipods, with copepods constituting more than 60% in most regions and seasons (e.g. Voronina, 1966; Hopkins, 1971). Krill is mostly absent except for the area around South Georgia. Typical representatives of the apex predators are fur seals as well as king and macaroni penguins.

The intermediate zone is the seasonal pack-ice zone, which is typically ice-covered in winter/spring and ice-free in summer/autumn. This zone includes most of the East Wind

Drift, the northern branch of the Weddell Gyre, and the Antarctic Peninsula. Large parts of the East Wind Drift are relatively shallow, and this region is the most productive in the Southern Ocean. The Antarctic krill is the dominant species and its summer distribution coincides with that of the sea ice cover in winter (Hempel, 1985; Nichol et al., 2008). The main feature is the coincidence of largest krill concentrations with mixing of water masses of different origins or with sharp changes in bottom topography. Thus, the highest concentrations of krill are found at the shelf slope break in the Bransfield Strait, and in the vicinity of fronts and eddies like the zones of the Antarctic Divergence and the Weddell-Scotia Confluence. Krill can occur in high densities of up to 30,000 individuals per m^3 close to the ice edge (e.g. Brierley et al., 2002), where they feed on a great variety of organisms ranging from phytoplankton (particularly diatoms) to larger zooplankton such as copepods (Smetacek and Nicol, 2005). However, recent studies observed krill also in high numbers at depths of up to 3,500 m, probably feeding on abundant phytodetritus (Clarke and Tyler, 2008). In winter, krill have been observed underneath the sea ice exploiting the sea ice algal standing stock (Ross et al., 1996; Daly, 2004). Krill itself provides an important food source for higher trophic levels, including notothenioid fish, seabirds, chinstrap and rockhopper penguins, elephant- fur- and crabeater-seals and the large baleen whales, and hence plays a pivotal role within the Antarctic food web.

The southern zone of the Antarctic open ocean ecosystem is the permanent pack-ice zone and comprises the cold ice shelf water along the continental shelf, particularly in the shallow parts of the inner Weddell and Ross Seas (Hempel, 1985). Phytoplankton production is limited to a short but intensive season, but there is a much larger season of ice algae production. The zooplankton abundance and biomass is low. Small neritic euphausiid species, *Euphausia crystallorphias,* and juveniles of the pelagic silverfish *Pleuragramma antarcticum* are typical species of this zone (Hempel, 1985; Boysen-Ennen and Piatkowski, 1988). As is typical for polar systems, the copepod community is characterised by the dominance of only a few species. In general, copepods are a very diverse taxon within the Antarctic zooplankton, accounting for a total of more than 100 species (Schnack-Schiel et al., 1998, 2008).

1.6.2.4 Higher Predators

The permanent pack-ice zone represents the habitat for a highly confined community of seabirds, the most unvarying of any seabird assemblage in the Southern Hemisphere (Ribic and Ainley, 1988). It is composed of Adélie and Emperor penguins, Snow and Antarctic petrels, with the addition during the summer of South Polar skua and Wilson's storm-petrel. Included as well, and unique to the Antarctic pack-ice zone, are four species of seals: Crabeater, Weddell, Leopard and Ross; and, among cetaceans, the Antarctic minke whale, the ecotype-C of the killer whale (Erickson et al., 1971, Laws, 1977a,b; LeDuc et al., 2008) and a number of rare toothed whales. These species are in most cases immensely abundant, within their respective groups. A large number of sub-Antarctic species of birds, pinnipeds and cetaceans occur in northern ice-free waters of the Southern Ocean, and move south in the summer as the pack ice recedes.

1.6.2.5 Species-specific adaptation to low temperature

Most Antarctic marine species are highly stenothermal, with the vast majority having experimental upper lethal temperatures between 5°C and 10°C (Somero and DeVries, 1967; Peck and Conway, 2000). The most stenothermal can only survive in a temperature window between -2°C and +4°C (Peck 1989, Pörtner et al., 1999a). The physiological processes setting the temperature tolerance limits, at least in marine ectotherms, are associated with reductions in whole animal aerobic scope (Pörtner et al., 1998, 1999b; Peck et al., 2002). Recently, Pörtner (2002) has elucidated the physiological basis of temperature limits at different levels, and shown a hierarchy of tolerance from the molecular to whole animal. This showed that the tightest limits were set at the whole animal level, with progressively wider tolerance at each step down the physiological hierarchy, and he argued that, in general, adding organismal complexity reduces thermal tolerance. Thus the physiological processes evident in response to varying temperature, at least in acute to medium-term experiments, are a progressive reduction in aerobic scope to a point where it is lost completely and tissues transfer to anaerobic metabolism, the critical physiological limit of Pörtner et al. (1998), and this may have a basis in mitochondrial function (Pörtner et al., 2007). Beyond this point survival is dictated by organismal tolerance to anaerobiosis.

In longer-term studies several Antarctic fish species have been shown to be able to acclimate to 4°C, but not above (Gonzalez-Cabrera et al., 1995; Lowe and Davison, 2005; Seebacher et al., 2005; Podrabsky and Somero, 2006; Jin and DeVries, 2006). Invertebrates, however, appear less able to acclimate to elevated temperatures, as attempts to acclimate animals to temperatures above 2°C failed for the scallop *Adamussium colbecki* (Bailey et al., 2005). In long-term temperature elevation trials the brachiopod *Liothyrella uva* survived at 3.0°C but failed at 4.5°C (Peck, 1989) and the bivalve *Limopsis marionensis* failed at 4°C (Pörtner et al., 1999a). Attempts to acclimate the clam *Laternula elliptica* (S. Morley, pers. comm.) and the brittle star *Ophioniotus victoriae* (M. Clark pers. comm.) to 3°C have also failed.

The important criteria for population or species survival in a given area is not, however, dictated directly by its physiological tolerance limits, but by the ecophysiological constraints on ability to perform critical biological functions such as feeding, locomotion and reproduction, and how changes in these characters affect ecological balances. Recent investigations of activity in a range of Antarctic marine herbivores have indicated a surprising sensitivity to temperature and a progressive decline in capability consistent with declining aerobic scope (Peck et al., 2004). The large infaunal bivalve mollusc *L. elliptica* has an experimental upper lethal temperature of 9°C and transfers to anaerobic metabolism at around 6°C (Peck et al., 2002). However, it ceases to rebury after removal from sediment at 5°C, and 50% of the population lose this ability when temperatures reach 2.5°C (Peck et al., 2007). Likewise the limpet *N. concinna* has an upper lethal temperature of 9.5°C (Peck 1989), but 50% of the population loses the ability to right themselves when turned over at around 2°C, and the scallop *A. colbecki* dies at 5-6°C, but loses the ability to swim between 1°C and 2°C. These are all major activities that involve extensive muscular activity. The most eurythermal Antarctic marine benthic species identified to date is the starfish *Odontaster validus*, that survives in raised temperature experiments to 15°C, is capable of performing activity (righting itself when turned over) to 9.5°C, and continues to feed normally and complete a full digestive cycle (Specific Dynamic Action of feeding, SDA, Peck (1998)) to 6°C (Peck et al., In Press). However, in interpreting these studies there remains a considerable gap between the longest experimental periods achievable (with the slowest rates of imposed temperature change), and the rates of change experienced over either evolutionary or contemporary climate change timescales. Thus there are a number of marine invertebrates present on both the Antarctic Peninsula and sub-Antarctic South Georgia, even though the thermal regime of South Georgia is above the experimental limits determined for some of these species living around the Peninsula. Clearly, experimental approaches cannot yet replicate the ability to adapt to climate achievable through evolutionary processes.

Chapter 2

Observations, Data Accuracy and Tools

Chapter Editor:

John Turner

Authors:

Todd Arbetter, Rob Arthern, Carlo Barbante, Roberto Bargagli, Dana Bergstrom, Nancy Bertler, Robert Bindschadler, James Bockheim, Claude Boutron, David Bromwich, Steve Chown, Andrew Clarke, Josifino Comiso, Pete Convey, Alison Cook, Guido di Prisco, Eberhard Fahrbach, Jim Fastook, Andrew Fleming, Mauro Gugliemin, Julian Gutt, Hartmut Hellmer, Françoise Hennion, Karen Heywood, David Holland, Ad Huiskes, Adrian Jenkins, Anna Jones, Anthony Lenton, Dominic Hodgson, Sungmin Hong, Stan Jacobs, Katrin Linse, Gareth Marshall, Paul Mayewski, Mike Meredith, Nicolas Metzl, Andrew Monaghan, Alberto Naveira-Garabato, Steve Rintoul, Howard Roscoe, Jonathan Shanklin, Ted Scambos, Mike Sparrow, Kevin Speer, Mark Stevens, Colin Summerhayes, John Turner, Kees van der Veen, David Vaughan, Cinzia Verde, Zhaomin Wang, David Webb, Philip Woodworth, Tony Worby, Roger Worland, Takashi Yamanouchi

2.1 Observations, data accuracy and tools

2.1.1 Introduction

In this chapter we consider the various types of data and models that are available to investigate climatic and environmental change in the Antarctic. We examine the availability and accuracy of *in-situ* physical and biological data, including the data that can be extracted from ice cores, along with the information that can be obtained from satellite systems that have become increasingly important over the last couple of decades.

Mathematical models have been utilized increasingly in Antarctic research. Initially their applications were in the more physical areas of climate, ice sheet, sea ice and ocean modelling. However, they are now starting to be employed in biology – a trend that seems set to continue in the future.

2.1.2 Meteorological and ozone observing in the Antarctic

2.1.2.1 Meteorological observations

The International Geophysical Year (IGY) provided a significant impetus towards setting up continuously operating stations in the Antarctic, and over forty were established, of which over a dozen are still operating today, alongside more recently established stations. This was the peak of manned observation in Antarctica, and since then the number of staffed stations has declined, though this is offset by an increasing number of automatic stations. A further boost to the observing network has taken place in the International Polar Year of 2007 – 2008.

Most manned stations are built on ice-free ground at coastal sites, primarily so that stores can easily be transported ashore and to simplify construction. The weather recorded at these sites is not a true representation of the continent as a whole, as the coastal areas are much milder than the interior due to the moderating influence of the sea and, at the more local scale, the surrounding rock. Automatic stations are much more widely spread across the continent and give a broader picture of the meteorology.

At most manned stations meteorological observations are made regularly throughout the "day" according to WMO standards, but there is increasing reliance on automatic systems during the "night". Surface temperature, humidity, solar radiation, atmospheric pressure, wind speed and direction are largely measured by automated instruments, but an observer is needed to estimate visibility and the amount, type and height of clouds, although automatic instruments are being introduced to deal with these parameters. The observer also needs to monitor the weather: rain, snow, fog, gales etc., as well as noting more unusual phenomena: diamond dust, halos, mirages and the *aurora australis*. Traditional weather observing on the polar plateau brings additional problems, with the combination of very low temperature and high altitude.

The observations are expressed in a numeric code and sent via geostationary satellites to meteorological centres, largely in the Northern Hemisphere, where they join thousands of other observations from all over the world. They are processed by super-computers and used to forecast the weather.

Automatic weather stations (AWS; Figures 2.1 and 2.2) generally measure a reduced range of parameters compared to the staffed stations, usually just atmospheric pressure, temperature and wind, although some may measure humidity, along with snow depth, which can also be used to estimate when the instruments may become buried. Where there is significant snow accumulation these stations require annual maintenance visits, however others may not be revisited after deployment.

Stations near the Antarctic coast are quite cloudy because of the frequent passage of depressions and the influence of the sea. The further a station is inland, the less cloudy it becomes. Signy has an average cloud cover of 86%, Halley 66% and the South Pole an average of 41%. Visual observation of cloud height is difficult at stations on ice shelves or the polar plateau, where the high albedo reduces contrast and there are no references to estimate height. Cloud lidars improve these measurements, and can also monitor precipitation falling from clouds.

The Antarctic atmosphere is very clear, as there are few sources of pollution. On a clear day it is possible to see mountains well over 100 km away. In these conditions, estimating distances can be very deceptive. Objects may appear to be close by, when in fact it would take many hours of travel to reach them. Automatic instruments, which use infra-red scintillation and scattering to measure visibility are becoming more common, although some have difficulty in discriminating variation in visibility above 20 km horizontal distance.

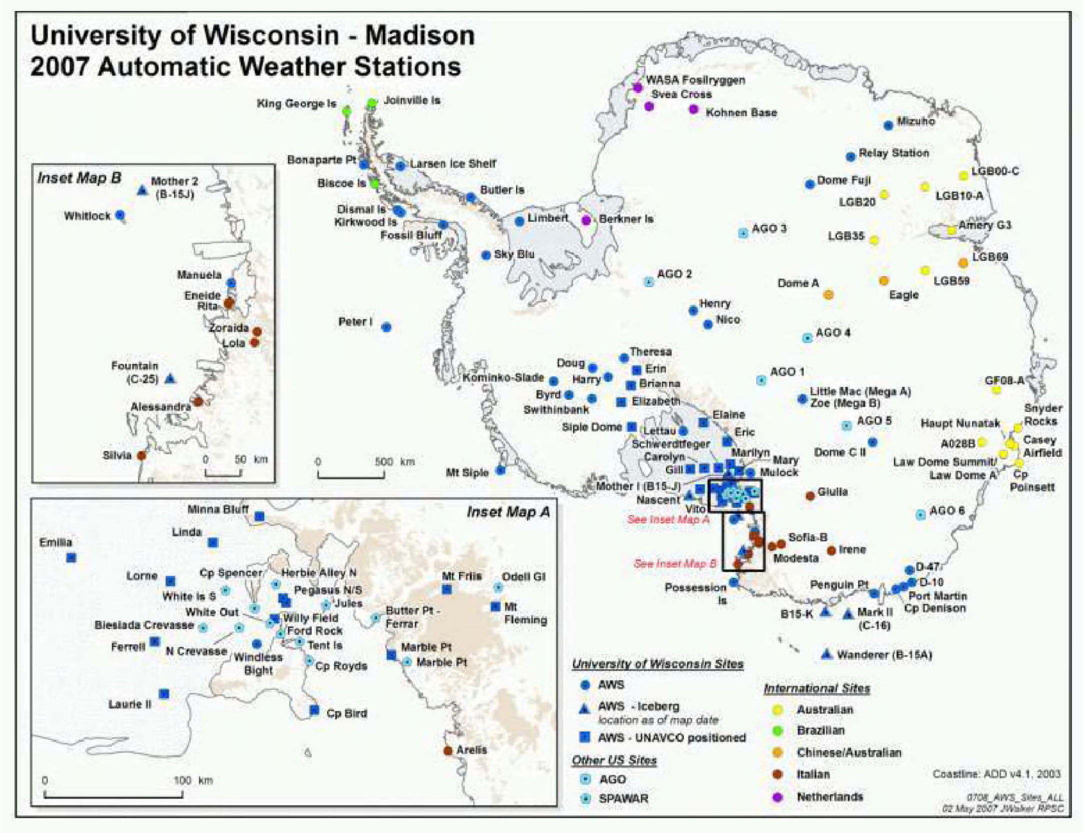


Figure 2.1 Location of automatic weather station (AWS) sites (Source: University of Wisconsin – Madison)

Figure 2.2 Servicing the AWS at Butler Island (Courtesy Jon Shanklin)

At approximately a dozen stations balloons are launched once or twice a day, each carrying a package of meteorological instruments known as a radiosonde. The instrument package signals back to base the temperature, humidity and pressure up to an altitude of over 20 km, with wind speed and direction being found by tracking the package with GPS sensors. One particular problem affects some balloons during winter: the combination of low ambient temperature and lack of heating from solar radiation makes the balloon fabric brittle and they burst early. The traditional remedy is to briefly dip the balloon in a mixture of oil and avtur immediately prior to launch, and to allow excess fluid to drain off. This plasticizes the fabric and gives much improved performance.

Special balloon ascents are sometimes made to help study the lower part of the atmosphere called the troposphere, where weather systems are active. These include flights to investigate very stable conditions in the lowest layer, which mainly occur during the winter, and other flights to study, for example, depressions forming off shore. Such studies are augmented by atmospheric profiles measured using captive packages carried aloft by kites or blimps, or by sodars (sonic radars). Further studies are made using instrumented aircraft, for example to study the composition of clouds in situ.

2.1.2.2 Ozone

Ozone is a compound of oxygen containing three atoms instead of the two in the oxygen we breathe. The largest concentrations of ozone in the atmosphere are found above the well-mixed lower atmosphere (the troposphere). The sunlight absorbed by these large concentrations of ozone heats the atmosphere locally, so the temperature increases with height, which leads to very little vertical wind in this ozone layer – hence its other name, the stratosphere. Above Europe, the ozone layer is typically from 13 to 50 km altitude. In Antarctica it is a few kilometres lower in altitude because of the reduced convective mixing of the troposphere in the reduced sunlight there, and because the colder temperatures result in higher densities that cause the same pressures to be at lower altitudes.

Ozone is created by the photochemical destruction of oxygen, which liberates a free oxygen atom to combine with an oxygen molecule:

$O_2 + h\nu \rightarrow 2O$	where $h\nu$ is a photon of radiation of wavelength < 240 nm
$O_2 + O + M \rightarrow O_3 + M$	where a third molecule (usually nitrogen or oxygen), M, is required to sequester or remove excess energy

The total amount of ozone in the atmosphere is routinely measured by a Dobson spectrometer, which measures the ratio of intensities of sunlight at two ultraviolet wavelengths between 305 and 340 nm, one of which is absorbed strongly by ozone, the other weakly. Wavelengths are selected by the use of prisms and a series of slits. A well-calibrated instrument can measure the amount of ozone to within a few percent. A typical amount of ozone in the atmosphere is around 300 milli-atmosphere-centimetres, equal to 300 Dobson Units (DU). If this was pure ozone at sea level, it would form a layer in the atmosphere just 3 mm thick.

Vertical profiles of ozone are measured by electrochemical sondes on balloons. Air is pumped through a cell containing potassium iodide solution, which generates a current proportional to ozone concentration. Ozone concentration is usually low in the troposphere, but increases in the stratosphere to a maximum between 15 and 25 km, above which ozone concentration decreases.

Figure 2.3 The Dobson ozone spectrophotometer at Halley, during an intercomparison of instruments prior to the announcement of the ozone hole in 1985 (courtesy British Antarctic Survey).

In common with most Antarctic data sets, that of total ozone in Antarctica starts in 1956 when stations began operation in the run up to the IGY of 1957 to 1958. Later some stations began measuring the vertical profiles of ozone by sondes. In the 1970s ozone measurements from polar orbiting satellites began. The long record of total ozone from Halley (Figure 2.3) and its location within the winter polar vortex made it the best site for discovering what is now known as the Antarctic ozone hole (Farman et al., 1985).

The ozone hole was discovered via ground-based observations from Antarctica, and most manned stations continue with long-term measurements of the ozone column from the ground. Ground based sensors include the traditional Dobson ozone spectrophotometer, the Brewer spectrometer and the SAOZ (Système d'Analyse par Observation Zénithales) spectrometer, or variants of these. All measure the differential absorption of sunlight as it passes through the ozone layer. At a few stations ozone sondes are flown on balloons; these sondes give precise profiles of the ozone in the atmosphere.

2.1.2.3 Observational problems

Although many of the meteorological observations made in Antarctica are done in exactly the same way throughout the world, some additional problems are encountered. Temperature measurements are often made using a platinum resistance or a traditional thermometer in a Stephenson screen, or variations of it. The screen shields the thermometer from direct solar radiation and from precipitation falling on the thermometer. In low wind speeds in summer, the radiation reflected from the high albedo surface can give anomalously high readings, whilst under clear skies in winter the reverse can occur. Many stations use aspirated screens, where air is sucked over the thermometer bulb at a constant flow rate, to provide more consistent data. A further problem occurs below -38°C, when mercury freezes, and so is either doped with thalium for use down to -61°C, or coloured ethanol is used as a fluid. In blizzard conditions a screen can fill with drifted snow, giving a uniform temperature environment unless it is quickly cleared. Some protection can be afforded by the use of snow boards, which temporarily block the louvres whilst the blizzard is in progress.

Measuring the precipitation itself can be difficult. The snow is generally dry and any that falls into a standard rain gauge just as easily blows out again. Equally, precipitation that has fallen elsewhere or at a previous time can be blown around by the wind and into the gauge. Specially designed snow gauges provide a partial solution, and another is to measure the depth of freshly fallen snow, and assume that in the long term there is a balance between transported and falling snow. Electronic precipitation detectors using scintillation in an infrared beam are now being deployed in Antarctica, and combination of the outputs of two detectors at different heights may provide the necessary discrimination between precipitation and transport.

Wind measurements were traditionally made using large, heavy cup anemometers. These required a significant wind speed before they started turning and did not respond well to gusts. They were replaced by anemometers with light-weight plastic cups, which performed better in these circumstances, but which frequently became coated with rime deposited from fog near the coastal regions, and hence under recorded wind speeds. Propeller type combined vanes and anemometers suffer less from rime, but can fail mechanically. The modern replacement is the sonic anemometer, which can measure both wind-speed and direction, can be heated to dispel any riming, and has no moving parts apart from the ultrasonic source. Even these suffer problems, particularly in conditions of heavy blowing snow, when the pinging of snow grains on the sensor saturates the detector, so they do not have widespread use over the continent as of this date.

Sunshine amounts were traditionally measured using the Campbell-Stokes recorder, a sphere of glass that focuses the sun's rays onto a fixed card, where they burn a hole in bright sunshine. This suffers from a universal problem of over-recording in patchy cloud conditions, and also experiences another problem in polar latitudes, where there can be 24 hours of daylight. In high southern (or northern) latitudes two recorders have to be mounted back to back to measure sunshine throughout the long day. Modern electronic recorders get around both problems and allow continuous recording, and are often more effective at recording sunshine at low solar elevations.

Early measurements of humidity were made using dry and "wet" bulb psychrometry. For much of Antarctica the "wet" bulb is an ice bulb, and a thin ice layer needs to be maintained for the technique to work reliably. This is not practical in an automatic system, and most sensors now use a capacitive technique to measure humidity. These sensors are generally reliable, although suffer a decline in accuracy when exposed to frequent freeze/thaw cycles. These conditions are common at many coastal sites, so that annual replacement of sensors often becomes necessary.

2.1.2.4 Data archiving

The Antarctic *in-situ* meteorological observations have been brought together within a SCAR project called REference Antarctic Data for Environmental Research (READER). The primary sources of data are the Antarctic research stations and automatic weather stations, and the data have, where possible, been obtained from the operators who run the stations. The observations have been quality controlled and are available on the web. For more details see http://www.antarctica.ac.uk/met/READER/. For periods prior to the instrumental era of

Antarctic meteorological stations, ice core records that have been calibrated to observed data can provide highly valuable proxies. Data for Antarctic ice cores has been assmbled together within a SCAR project called Ice REference Antarctic Data for Environmental Research (Ice READER) (Figure 2.4) available on the web. For more details see http://www.icereader.org/icereader/.

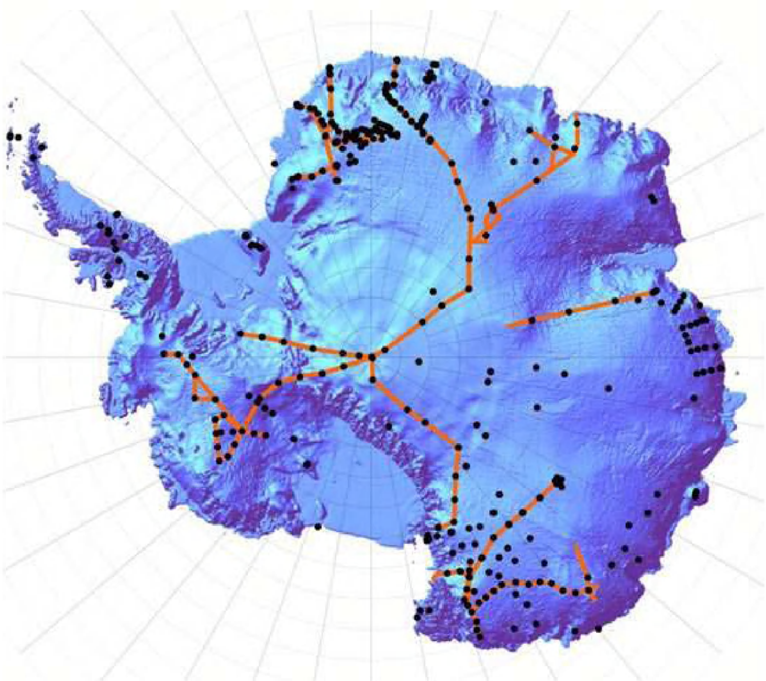


Figure 2.4. Location of Antarctic ice cores (black dots) and ground penetrating radar (GPR) routes (red lines) from Ice READER.

2.1.2.5 Meteorological analyses

Some surface synoptic weather charts were prepared in the early years of the Twentieth Century, but these did not have a great deal of accuracy in the Antarctic, as there were very few observations over ocean areas. From the 1970s, an increasing number of observations became available from the polar orbiting weather satellites, which allowed the production of increasingly reliable atmospheric analyses at high southern latitudes. Of particular importance are the atmospheric temperature sounders that have been flown on the polar orbiting satellites since the mid 1970s. These provide profiles of temperature and humidity from the surface into the stratosphere and are similar in nature to the radiosonde ascents, although with the broad coverage provided by a satellite system.

Over the last few years the historical archives of *in-situ* and satellite observations have been re-processed using current data assimilation schemes to produce the so called 're-analysis' data sets, which provide a particularly valuable source of information for investigation of climate variability over approximately the last three decades. However, prior to satellite sounder data becoming available in 1974, the re-analysis fields for the Southern Hemisphere are poor and cannot be used to investigate atmospheric circulation change.

2.1.2.6 Satellite Sounding of the Atmosphere

Satellite sounding instruments measure radiation at infrared or microwave wavelengths that has been emitted by the atmosphere itself, and thus provide information on the temperature and composition (e.g. humidity, ozone amount) of the atmosphere over a range of altitudes. Such instruments have improved our knowledge of the climate of the polar regions, and hence the forecasts made by Numerical Weather Prediction (NWP) models. Today, regions such as the Southern Ocean, previously a data void, are filled by data with a spatial resolution of tens of kilometres. The wavelengths of these types of instrument are selected according to the absorption and emission properties of carbon dioxide, nitrous oxide, oxygen, ozone and water vapour. As the distribution of these gases is relatively constant and their radiative properties understood, variations in the radiance received by the satellite (the brightness temperature) can be related via mathematical inversion methods to the temperature of that part of the atmosphere from which most of the signal emanates. By using a number of wavelengths that have different absorption characteristics a complete profile of temperature through the atmospheric column can be derived (see Figure 2.5).

The principal satellite sounding instruments are the TIROS-N operational vertical sounder (TOVS) sensor systems, which have flown on polar-orbiting satellites since 1978, and their successor Advanced TOVS (ATOVS), which began operating in 1998. TOVS comprises three radiometer arrays: (i) the high-resolution infrared radiation sounder version 2 (HIRS/2), the microwave sounding unit (MSU), and the stratospheric sounding unit (SSU), while in ATOVS the Advanced Microwave Sounding Unit-A (AMSU-A) and the Advanced Microwave Sounding Unit-B (AMSU-B) replace the MSU and the SSU, and HIRS/3 replaces the HIRS/2. Significant improvement in forecast accuracy has come from using the AMSU data.

Unfortunately, comparison against radiosondes launched from Arctic sea ice demonstrated marked deficiencies in the original TOVS retrieval algorithms over sea ice during the winter (Francis, 1994). Near-surface temperature inversions were missed, primarily because of surface temperature retrievals over sea ice being 5-15°C too high. This was traced to a tendency for the algorithm to over-estimate total column water vapour (Francis, 1994). Similar problems have been identified in the ECMWF 40 year reanalysis (ERA-40) prior to 1996 (Bromwich et al., 2007). In the Antarctic there are the additional problems that the high Antarctic Plateau causes for an un-tuned retrieval algorithm (Lachlan-Cope, 1992). However, a comparison of integrated water vapour from AMSU-B and limb-sounding observations by a GPS receiver on board the Low Earth Orbiter (LEO) CHAMP satellite (obtained using radio occultation techniques) showed good agreement between the two independent methods, indicating that modern satellite instruments can obtain accurate data on moisture over the plateau, which could be usefully assimilated into NWP forecasts (Johnsen et al., 2004).

2.1.2.7 Atmospheric data derived from ice cores

Temperature proxies

The Antarctic presents a complex picture of temperature change in recent decades (Turner et al., 2005a). While the Antarctic Peninsula and some coastal portions of the Antarctic Ice Sheets have experienced pronounced warming over the last 50 years, most continental stations have shown little or no significant surface warming over this same period. Despite this, significant warming trends have been identified in the mid-troposphere in winter (Turner et al., 2006). (See also Steig et al, 2009)

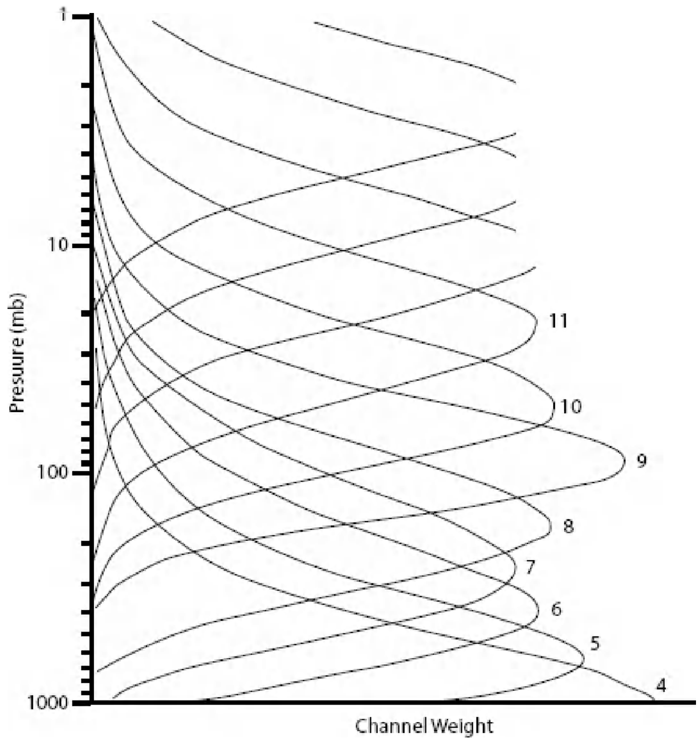


Figure 2.5 The weighting functions for channels 4-14 of the AMSU-A instrument.

High-resolution (sub-annual) proxy temperature data are derived from stable water isotope measurements on ice cores, and records of such data presently range from several centuries up to around 1,000 years (e.g. Graf et al., 2002; Morgan and van Ommen, 1997; Steig et al., 2005). The largest-scale proxy reconstruction presently available in Antarctica (Schneider et al., 2006) extends the record of temperature changes over the continent (excluding the Peninsula) to 200 years, using ice cores from 5 sites. The proxy data show a long-term increase in temperature of around 0.2°C since the late nineteenth century. This change is in phase with the Southern Hemisphere mean until around 1975, when this primarily interior Antarctic temperature record diverges, showing a very slight cooling trend. The proxy reconstruction and observations reveal significant decadal-scale variability in temperatures associated with variability in the Southern Annular Mode (SAM). This suggests that this surface cooling may, at least in part, reflect recent changes in the SAM (Thompson and Solomon, 2002), a connection also noted by Turner et al. (2005a).

Atmospheric circulation proxies using ice cores

Advances in the sampling and analysis of chemical ion concentrations and stable isotope composition of snow and firn have enabled the collection of multivariate data sets from firn and ice cores, which have increasing application as proxy climate data. Glaciochemical time series in conjunction with accumulation rate time series have been used to unravel the likely moisture sources and atmospheric circulation history for many global locations in the last decade. For a review of glaciochemistry and an introduction to its early use as climate proxy data see Legrand and Mayewski (1997).

Coupled to the advancements in geochemical analyses is the availability of global climate reanalysis data, and instrumental station climate data that have enabled the calibration of ice core proxy climate data, particularly pressure. The two main climate reanalysis data sets are: the US National Centers for Environmental Prediction (NCEP) - National Center for Atmospheric Research (NCAR) climate reanalysis data set that is referred to as the NNR data set (Kalnay et al., 1996; Kistler et al., 1985) available from the US National Oceanic & Atmospheric Administration (NOAA); and ERA-40 climate reanalysis data available from the ECMWF. These data sets contain gridded 6 hourly, daily and monthly data for the full range of climate parameters, from the surface through the atmosphere. The monthly data are of most relevance to the calibration of annual to sub-annually resolved ice core data. Sea-salt aerosols, calcium, nitrate and MS aerosols are the most widely used proxies for atmospheric circulation in the Antarctic region. Examples of the use of these proxies to reconstruct atmospheric circulation and precipitation are given below.

Ice core nitrate as a proxy for atmospheric ridging and the SAM. The Queen Maud and Wilkes Land regions of East Antarctica lie entirely within the strong katabatic wind zone where erosional surface winds from east-southeast to south-southeast drain cold air from the ice sheet's interior down-slope to the coast. Snow precipitation accompanied by an east to east-southeast wind occurs during synoptic-scale maritime cyclonic incursions over the katabatic slope, at least to the 2,000 m elevation. Cullather et al. (1998) established that the strength of the ridging over Wilkes Land influenced the circumpolar storm tracks, resulting in cyclones being steered into Wilkes Land. Murphy and Simmonds (1993) and Bromwich et al. (1993) have established that tropospheric ridging has a significant influence on the surface windfield. Katabatic winds were found to intensify in response to stronger than average surface temperature inversions produced by an enhanced high surface pressure and at the 500 hPa level over the East Antarctic interior. These high pressure anomalies over East Antarctica are associated with a blocking anticyclone to the south east of New Zealand (170 to 180°E), and approaching low pressure systems in the circumpolar trough at 100-120°E. Strong wind events are the product of enhanced katabatic wind flow down slope together with strong geostrophic wind flow across the Wilkes Land slopes (Murphy and Simmonds, 1993). Research by Goodwin et al. (2003) has demonstrated that the nitrate time series from Wilkes Land sites, located in convergent windfields, is a proxy for the strength of the surface windfield. They established from the annual sea salt and nitrate concentration time series that anomalously high accumulation rate and nitrate concentrations at these sites are the product of the accumulation of blown snow, plus increased surface wind speed and/or wind pumping efficiency, rather than synoptic precipitation. Goodwin et al. (2003) have shown that the annual mean nitrate concentration in eastern Wilkes Land snow has a strong statistical relationship with winter (June, July and August) meridional mean sea level pressure (MSLP) indices (Macquarie Island - Scott Base, and Kerguelen Island - Casey MSLP) that describe the oscillations in the SAM (Thompson and Solomon, 2002). The decreasing trend in annual mean nitrate concentration in snow after 1964 suggests a reduced incidence of mid-latitude atmospheric ridging into Wilkes Land during winter (Figure 2.6). This is in agreement with the observed deepening of the circumpolar trough since the mid 1960s (Allan and Haylock, 1993) and the trend towards the high index phase of the SAM (Thompson and Solomon, 2002).

Ice core sea-salt as a proxy for mid-latitude and circum-Antarctic sea-level pressure and the SAM. Goodwin et al. (2004) have derived a 700 year proxy record (at monthly resolution) for winter (May, June, July (MJJ)) MSLP variability over the Southern Ocean, by analysing sea-salt (sodium) aerosol concentrations in the DSS ice core from Law Dome in East Antarctica. The relationship between modern patterns of mid-latitude and sub-Antarctic atmospheric circulation and variations in sodium (Na) delivery to Law Dome ice was identified by analysing co-variations between Na concentrations, MSLP and wind field data. The observed relationship was then used to hindcast a proxy record of early winter MSLP anomalies and the SAM. The proxy mid-latitude MSLP time series is most correlated to atmospheric variability in the region southeast of Australia and south of the Tasman Sea (Macquarie Island and Campbell Island). This is a function of the Rossby Wave number 3 pattern, the periodic atmospheric blocking in this region, and interaction with the SAM. The MSLP patterns (MSLP anomalies that are associated with high and low sea salt aerosol deposition in the DSS (Dome Summit South, Law Dome) core) are shown in Figure 2.7 (Goodwin et al., 2004). The full 700 year record of early winter (MJJ) sea salt sodium and MSLP variability is shown in Figure 2.8 (Goodwin et al., 2004).

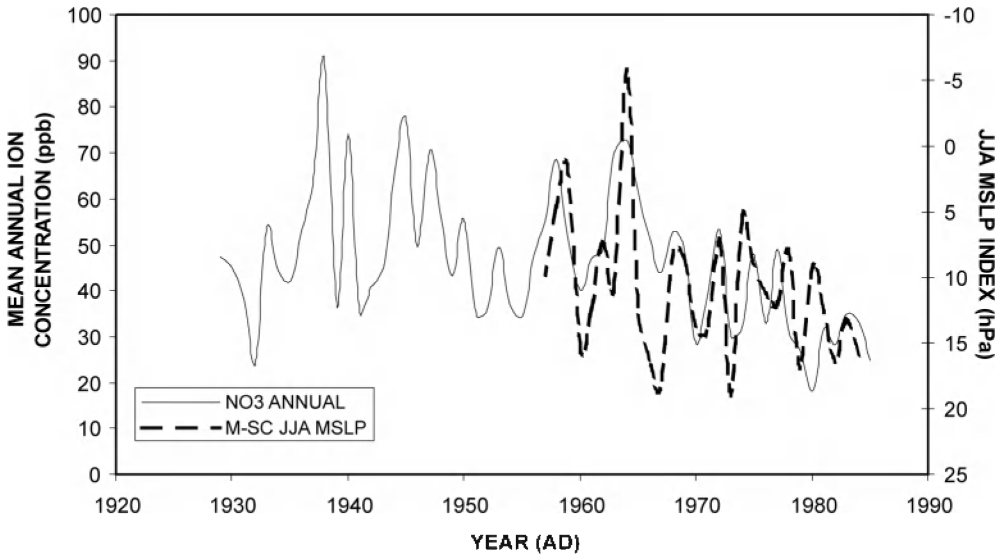


Figure 2.6 Mean annual nitrate concentration at GD09 (eastern Wilkes Land, East Antarctica) plotted against the average mean sea level pressure (MSLP) gradient between Macquarie Island and Scott Base for June, July and August (JJA). The MSLP index represents the difference in MSLP across the circumpolar trough to the east of Wilkes Land, where ridging is prevalent. Low values of the MSLP index are indicative of ridging (meridional conditions) across the circumpolar trough in the 130 to 160° E longitudes, whilst high values represent strong zonal conditions.

Ice core sea-salt as a potential proxy for mid-latitude winter rainfall variability. Recent work by Goodwin (in prep.) has focused on the application of the proxy mid-latitude MSLP and SAM index time series, to investigate methods for hindcasting southern Australian rainfall over the past millennium. Proxy May - July MSLP and SAM data have been cross-correlated against May - July total rainfall data for stations located in southwest and southern Western Australia. Preliminary results, indicate that the highest correlation between the data was calculated for stations close to the coastal escarpment in southwest Western Australia, indicating the strong relationship between winter rainfall, the phase of the SAM and the circumpolar longwave circulation pattern, particularly the meridional location of the Rossby wave number 3 troughs and ridges in the Indo-Australian region. Interdecadal winter rainfall variability across coastal Southern Australia appears to be strongly associated with the time-varying behaviour of the longwave pattern and the SAM. The work in progress indicates that southwest Western Australia experienced periods of higher mean winter rainfall, with high interdecadal variability during 1300 to 1600 AD, followed by lower mean but less variable winter rainfall from 1600 to 1900 AD, which is similar to the past 50 years (Goodwin, in prep.). These preliminary results, illustrate the potential for using high-resolution (monthly) ice core glaciochemical data to reconstruct and predict atmospheric circulation and rainfall distribution patterns across Southern Australia at interannual to interdecadal resolution. Since there is a strong correlation between mid-latitude MSLP, South Indian Ocean sea surface temperatures (SST) and rainfall over southwest Western Australia (Smith I. et al., 2000) ice core sites in Queen Mary Land, East Antarctica appear to have the best potential for reconstructing MSLP and rainfall variability over southwest Western Australia for the past few hundred years. This long record would be of enormous economic benefit to all water users in Western Australia.

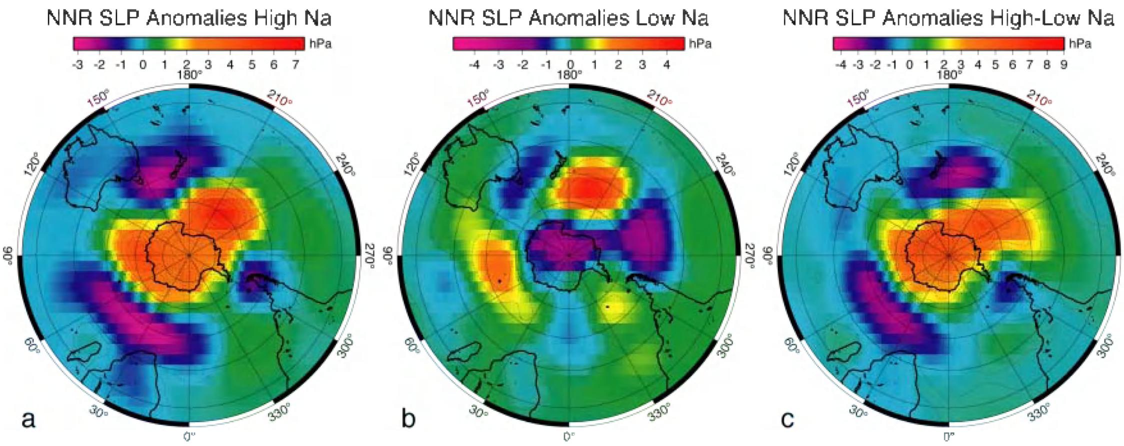


Figure 2.7 The spatial pattern of 'early winter' May - July NNR MSLP anomalies calculated for: a) the 6 highest (1990, 1977, 1992, 1970, 1986 and 1975) Na concentration years at DSS between 1970 to 1995; b) the 6 lowest (1982, 1981, 1995, 1985, 1993 and 1971) Na concentration years; and, c) from the difference between the 6 highest and 6 lowest Na concentration years. The difference plot accentuates the circulation changes between SAM phases in the mid-latitude trough in the southwest Indian and Pacific Ocean sectors, together with the enhanced ridging in the circumpolar trough centred on 110° E during high Na concentration years at DSS.

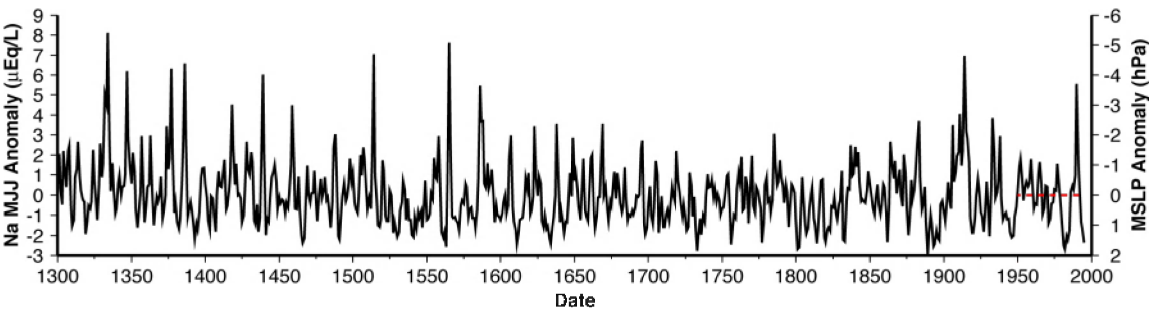


Figure 2.8 Sodium May – July time-series from 1300-1995 in µEq/L are shown as anomalies from the 1950-1995 mean (dashed line). Data are low filter. Right (inverted) axis shows equivalent MSLP anomalies for the southwest Pacific region (Campbell Is. and Macquarie Is.) using the observed calibration slope of -0.65 hPa/µEq/L derived from the instrumental period (see text). Negative (positive), states of the SAM correspond to negative (positive) MSLP anomalies at mid latitudes.

Although full-scale reconstructions of past climate over Antarctica have yet to be finalized, SCAR's International Trans-Antarctic Scientific Expedition (ITASE) (Mayewski et al., 2005a) has, as noted above, pioneered calibration tools and reconstruction of climate indices and evidence for climate forcing using single-sites through to multiple arrays of sites. Initial syntheses of combined ITASE and deep ice core records demonstrate that inclusion of instrumentally calibrated ITASE ice core records allows previously unavailable reconstruction of past regional to continental scale variability in atmospheric circulation and temperature (Mayewski et al., 2005a). Emerging results demonstrate the utilization of ITASE records in testing meteorological reanalysis products (e.g. Genthon et al., 2005). Connections have also been made between ITASE climate proxies and global scale climate indices such as ENSO (Meyerson et al., 2002; Bertler et al., 2004) and to major atmospheric circulation features over the Southern Hemisphere - such as the Amundsen Sea Low, East Antarctic High, Southern Hemisphere westerlies and SAM (Kreutz et al., 2000a; Souney et al., 2002a; Goodwin et al., 2004; Proposito et al., 2002; Shulmeister et al., 2004; Kaspari et al., 2005; Yan et al., 2006). ITASE research is also focused on understanding the factors that control natural climate variability over Antarctica and the Southern Ocean, through, for example, documentation of the impact of solar forcing via UV induced changes in stratospheric ozone concentration on the strength of the zonal westerlies at the edge of the polar vortex (Mayewski et al., 2005b).

ITASE proxy circulation reconstructions provide an estimate for the state of modern Antarctic climate. Evidence is emerging for inland penetration of marine tropospheric air masses over the past few decades in summer into portions of coastal West Antarctica near the Amundsen Sea (Dixon et al., 2006) and the Antarctic Peninsula. However, ice core proxy reconstructions for the Amundsen Sea Low suggest that this system is still within the range of variability established over the last 1,200 years (Mayewski and Maasch, 2006), while proxy reconstruction of the Southern Hemisphere westerlies reveals intensification in the 1980s (Dixon et al., in review) consistent with the impact of human-induced changes in stratospheric ozone on the strength of the westerlies around the edge of the polar vortex (Thompson and Solomon, 2002).

2.1.2.8 Snow Accumulation

Field campaigns traversing many thousands of kilometres have revealed the broad pattern of snow accumulation across the featureless plateau of the ice sheet (the input term in the ice-budget). This can be measured by sampling snow from pits or ice cores then counting seasonal cycles of isotopes, dust, and other chemical impurities (e.g. Mayewski et al., 2005a; Frezzotti et al., 2005; Monaghan et al., 2006a). If the density is also measured, the mass of snow deposited within each year can be found. Repeated observations of snow stakes provide similar information (Kameda et al., 2008; Takahashi and Kameda, 2007). One of the most reliable methods is to detect the layers contaminated by radioactive fallout from atomic bomb tests performed in the 1950s and 1960s (Kamiyama et al., 1989). By measuring the mass of snow that has accumulated above this layer, an unambiguous measurement of snow accumulation is provided.

There are now several thousand observations of snow accumulation across the ice sheet, and these have been compiled into maps of the mean annual snow accumulation rate (Vaughan et al., 1999; Takahashi et al., 1994). Satellites and atmospheric models provide some information about snow accumulation away from traverse routes, and this can help to interpolate the field observations (Giovinetto and Zwally, 2000; Arthern et al., 2006; van de Berg et al., 2006). The published maps are broadly similar in character, showing high coastal accumulation rates and a cold, precipitation-starved interior. The largest differences between estimates are found in the coastal regions where poor weather has hitherto limited the amount of data collected by field parties.

Measurements of snow accumulation have been supplemented by field-based radar. The radar reveals layering within the ice that has been deposited over time. Comparisons with independently dated ice cores have confirmed that bright reflecting layers accurately trace layers of snow that were deposited synchronously, referred to as isochrones. By dating a particular isochrone using an ice core, then following how its depth varies from place to place, the spatial gradients of snow accumulation can be determined (e.g. Richardson-Naslund, 2004; Spikes et al., 2004). Ice penetrating radars are already flown on aircraft, and in the future will be flown on satellites, such as Cryosat-2 (Hawley et al., 2006). These will provide more complete coverage of the stratigraphy of the ice sheet. For a review of accumulation measurements using such methods see Eisen et al. (2008).

When multiple layers are traced, a complete spatio-temporal picture of how snow has accumulated along the traverse route can be established. Surveys of this kind have revealed that the accumulation upon the ice sheet is not uniform, but varies erratically: drifting depletes some locations of snow and provides others with a surfeit. The effect can be especially severe in low accumulation regions where drifting produces dune-like structures (Fahnestock et al., 2000; Frezzotti et al., 2002). The effect of drift makes it difficult to analyse trends in Antarctic snowfall over recent decades. Nevertheless, combining many core records and forming decadal averages gives a measure of how snowfall has varied in Antarctica.

Most of Antarctica is too cold for surface melt to occur, but there is some melting around the periphery. This can be observed from space because the signature of wet snow is quite different from dry snow in the microwave spectrum (Ridley, 1993; Zwally and Fiegles, 1994). This method has been used to study how melt rates vary from year to year. Between 1980 and 2006, melt seasons over the ice shelves of the Antarctic Peninsula lengthened (Torinesi et al., 2003). In the latter decade of this interval (1996 to 2006) melt seasons shortened in the Peninsula, Ronne-Filchner and Amundsen Sea sectors of Antarctica, but lengthened over much of East Antarctica and the Ross Ice Shelf (Torinesi et al., 2003). The pattern of melting generally follows trends in air temperatures (Liu et al., 2006). Snow surface temperature can be measured from space using images from the Advanced Very High Resolution Radiometer (AVHRR), which is sensitive to infra-red radiation emitted by the surface (Comiso, 2000). This temperature information has proved useful in glaciological applications, including modelling snow compaction (Li et al., 2007) and detecting glazed surface (Fujii et al., 1987), which develops on steeper slope of the undulating ice sheet suraface (Furukawa et al., 1996).

Although microwave radar altimeters are designed to measure surface elevation, the backscattered power can also provide information on snow properties (Legresy and Remy, 1997; Arthern et al., 2001). Passive microwave radiometers carried by satellites and scatterometers provide complimentary information on the snow structure and how it changes (Surdyk and Fily, 1994; Sherjal and Fily, 1995; Long and Drinkwater, 1999; Rotschky et al., 2006). It has proved difficult to derive physical and climatic quantities related to snow structure directly from microwave observations, but these satellites can provide spatial information that extends the usefulness of field observations. Microwave observations from the Advanced Microwave Scanning Radiometer (AMSR-E) instrument have been combined with temperatures measured by AVHRR to assist the interpolation of sparsely-spaced observations of snow accumulation rate (Arthern et al., 2006).

The scale of the Antarctic ice sheet, and the difficulties of working there, mean that remote sensing methods using satellites are the only practical way to get an overview of how the entire ice sheet is changing. The distinction between satellite remote sensing and fieldwork is becoming more blurred as field operations are targeted towards important changes identified by satellite and as fieldwork provides calibration and validation measurements for specific satellite campaigns. Although the time series of available satellite observations grows longer each year, it is still a short record compared to observations from the surface. Field-based observations using radar, seismic and ice cores can resolve extra dimensions in depth, and hence time, through the layered structure of the ice sheet. They can also give invaluable insights into what has caused changes that are observed using satellites.

2.1.3 In-situ ocean observations

The Southern Ocean plays a critical role in driving, modifying, and regulating global change. We need to be able to address the magnitude, variation, causes, and consequences of such change by monitoring the state of the ocean. However, conducting research in the Southern Ocean can be extremely challenging. Sea ice, high winds, rough seas, poor visibility, sub-zero temperatures, 24 hour winter darkness, and the icing of ships' superstructures do not make for an easy environment in which to work. To obtain the required data requires a mixture of tried and tested techniques as well as the use of novel technology (e.g. Smith and Asper, 2007).

2.1.3.1 Ship-based measurements

Taking measurements from a ship, whether by estimating currents from ship drift, taking samples of water to examine its properties or by catching and examining animals from the sea has been carried out to some extent since humans first took to the oceans. Even now taking measurements or samples from a ship is still the mainstay of *in situ* oceanographic observations.

Figure 2.9 Water sampler-CTD package being lowered into the water (photograph courtesy of Ovide Ifremer)

Many different sampling systems are used from ships, but by far the most common is the use of a package consisting of a water bottle array with a Conductivity-Temperature-Depth (CTD) probe that is lowered into the ocean when the ship stops on station (Figure 2.9). The CTD package most commonly includes sensors for conductivity (from which salinity can be derived) and temperature plus a pressure sensor. Other sensors, including those used to measure dissolved oxygen and photosynthetically active radiation as well as fluorometers, transmissometers and Acoustic Doppler Current Profilers (ADCPs) (to measure water velocity) are also routinely included. Water samples from the bottle array are taken to calibrate the conductivity sensors as well as to measure oxygen and its isotopes, nutrients, chlorofluorocarbons (CFCs), dissolved inorganic carbon and alkalinity, trace metals and a host of other parameters important to climate, biogeochemical and ecological studies.

When steaming between CTD stations, most research ships - and many so called 'ships of opportunity' - are equipped with a variety of underway samplers, for example sensors on the ships' masts to record marine meteorological parameters such as wind speed and humidity, as well as other equipment taking measurements of the ocean such as thermosalinographs (that measure the temperature and salinity of the water a couple of metres below the surface) or hull-mounted ADCPs and echosounders.

2.1.3.2 Towed and tethered instruments

Instruments that can either be towed behind or tethered to ships (remotely operated vehicles) have provided substantial insights into biological and physical features and processes and are becoming increasingly sophisticated. SeaSoars (towed instruments that undulate in the upper 250 m or so of the ocean, and are equipped with temperature, salinity, fluorescence, and other sensors) have been used extensively in the Antarctic. For example, Hales and Takahashi (2007) were able to determine the three-dimensional structure of hydrographic, chemical and biological variables in the southern Ross Sea. Remotely operated Vehicles (ROVs) can be equipped with sensors and samplers for water mass properties. Other towed systems such as the Continuous Plankton Recorder are also used routinely in the Southern Ocean (e.g. Hunt and Hosie, 2003).

2.1.3.3 Autonomous vehicles

The types of autonomous vehicle vary widely, depending in part on the nature of propulsion and the sensor payload. AUVs (Autonomous Underwater Vehicles) use batteries or fuel cells as the primary power source for propulsion whereas gliders use variable buoyancy and have smaller power requirements, however paid by smaller speeds which is critical in strong currents.

As an example, the AUV AutoSub-2 (Figure 2.10) has conducted high resolution surveys of krill abundance in the marginal ice zone of the Antarctic and found that krill were concentrated in a narrow band within roughly 1 km of the ice edge, in densities exceeding those in the open water and elsewhere in the marginal ice zone (Brierley et al., 2002). Such small-scale distributions have profound implications for the foraging ecology and energy transfer within open-water food webs, and could not have been determined using ship-based surveys.

Both AUVs and gliders are capable of multidisciplinary sampling. As a result, there is increasing demand for better sensor development and more complex sensor configurations. A particular challenge is the use of such devices under ice where navagiation by returning to the surface is not possible. However, Autosub III recently spent significant time under the Pine Island Glacier, on traverses of up to 60 km and 30 hours under not just sea ice but several hundred metres of the floating glacier tongue.

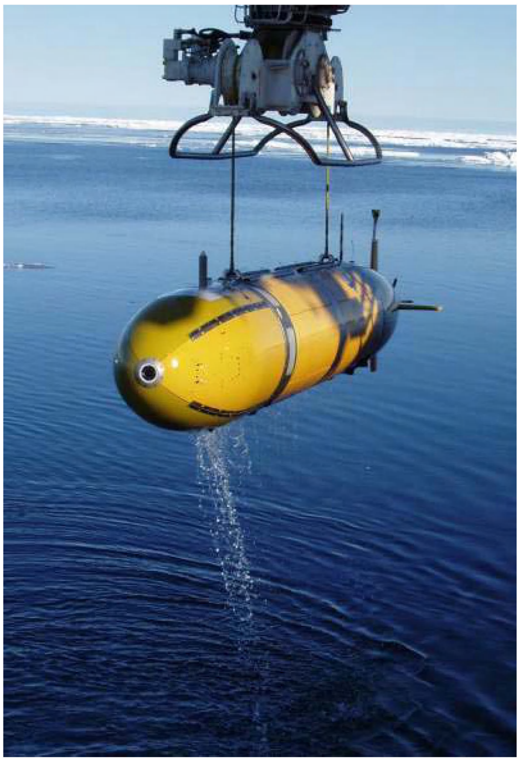


Figure 2.10 Autosub-2 (source Gwyn Griffith)

2.1.3.4 Time series sites

Time series sites (e.g. http://www.oceansites.org/) consist of instrumentation measuring a range of parameters at a particular location repeatedly over time. They may consist of returning to a particular location to take repeat measurements or of instruments moored semi-permanently. Fixed instrumented moorings allow data collection of a suite of parameters, e.g. current or ice velocity, temperature and salinity for extended periods at a single location. Moorings are usually anchored to the seafloor with instruments extending upward by means of flotation, but may also be mounted within the ice, with a free hanging weight suspended below. As an example, the Weddell Sea Convection Control (WECCON) experiment has been investigating large-scale processes and long-term variations of convection in the Weddell Sea since 1996 (e.g. Fahrbach et al., 2004). Moored instruments are particular useful in key locations to monitor outflow plumes of water masses from the Weddell Sea or determine transports through straits, such as the Drake Passage.

Tide gauges are essential for observing changes in sea level as well as for estimating volume transport variability, for example in the Antarctic Circumpolar Current (ACC) (e.g. Meredith et al., 2004). The Antarctic tide gauge network includes gauges at coastal stations on the Antarctic continent, and also many gauges at Antarctic and Subantarctic Islands. However there are large gaps in some key areas, such as the Amundsen Sea.

A combination of bottom pressure recorders and Inverted Echosounders (PIES) which measure the traveltime of sound signals from the sea bed to the surface and back allow us to derive fluctuations of heat and mass transports through the change of bottom pressure and the sound velocity profile in the water column. The data can be related to the vertical distribution of temperature and salinity by use of the gravest empirical mode.

2.1.3.5 Floats and drifters

Data obtained from surface drifters and subsurface floats, particularly the Argo array (http://www.argo.ucsd.edu/), have made a huge impact on our understanding of the world's oceans and hold tremendous promise for acquiring data to address critical questions for the Southern Ocean. Because of the problem of ice damage to floats and drifters, novel technologies and techniques have been developed to cope with these conditions, e.g. in the case of Argo, experimenting with subsurface tracking using Sound Fixing and Ranging (SOFAR) technology, storage of the profiles, and the adaptation of software to detect ice as the float attempts to surface to transmit its data to satellite (e.g. Klatt et al., 2007). The installation of arrays of sound sources to be used by floats and gliders are a logistical and financial challenge. The application of ice-tethered platforms consisting of autonomous surface platforms on the ice with profiling instruments under the ice which are now common in the Arctic is restricted in the Antarctic due to the almost completely disappearing sea ice in summer.

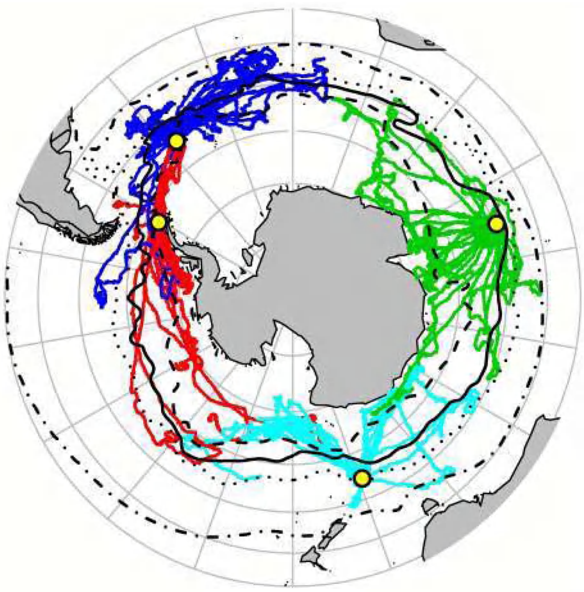


Figure 2.11 Tracks of elephant seals tagged with instruments to record temperature and salinity profiles in the Southern Ocean (source Biuw et al., 2007).

2.1.3.6 Sensors on animals

The use of innovative techniques, such as the deployment of small sensors on wide-ranging marine animals (Figure 2.11) has provided a wealth of data on their movements and behaviour as well as providing near real-time monitoring of ocean properties for long-term weather and climate analyses and forecasting (e.g. Biuw et al., 2007).

2.1.3.7 Oceanographic data resources

In order to facilitate international exchange of scientific data a number of oceanographic data centres have been set up both as access points and repositories for such data. As well as National Data Centres there exist a number of international Data Assembly Centres that are dedicated to the collation, archival and delivery of data and products (see e.g.

http://www.clivar.org/data/dacs.php). As an example, the CLIVAR (Climate Predictability and Variability Project) and Carbon Hydrographic Data Office is a repository and distribution centre for CTD and Hydrographic data sets (http://cchdo.ucsd.edu/). Other organizations both act to facilitate data exchange as well as providing access to the data itself, for example the World Data Center for Oceanography (http://www.nodc.noaa.gov/General/NODC-dataexch/NODC-wdca.html), the IOC's International Oceanographic Data and Information Exchange (IODE) (http://www.iode.org/) and, covering the Southern Ocean region only, the Joint Committee for Antarctic Data Management (JCADM; http://www.jcadm.scar.org/).

A number of other resources exist for oceanographic data, for example the online Southern Ocean Atlas (http://woceatlas.tamu.edu/) includes a range of data products for the region south of 30°S (e.g. Figure 2.12). Static atlas products are available for browsing and downloading, but a suite of fully interactive tools are also provided where users can construct atlas illustrations using their own choice of parameters.

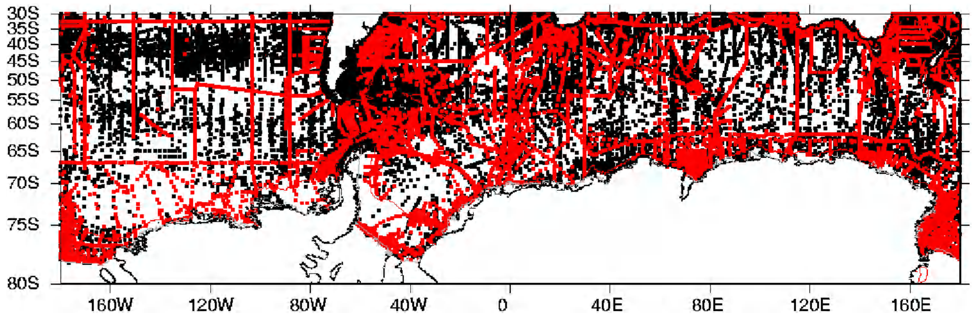


Figure 2.12 Example figure produced using the Southern Ocean Atlas (http://woceatlas.tamu.edu/). Dots show all data from 1900 to 2007 inclusive, including bottle data. Red dots are CTD data only.

Southern Ocean READER is a portal for links to temperature, salinity and ocean current data from the Southern Ocean (http://www.antarctica.ac.uk/met/SCAR_ssg_ps/OceanREADER/).

The Joint World Meteorological Organisation Intergovernmental Oceanographic Commission Technical Commission for Oceanography and Marine Meteorology *in situ* Observing Platform Support Centre (JCOMMOPS) provides coordination at the international level for oceanographic and marine observations from drifting buoys, moored buoys in the high seas, ships of opportunity and sub-surface profiling floats (http://www.jcommops.org; see Figure 2.13).

2.1.3.8 Observational problems

There is a suite of methods available to study the Southern Ocean. However, despite the importance of the region to global change, it is still one of the most data-sparse regions on the planet.

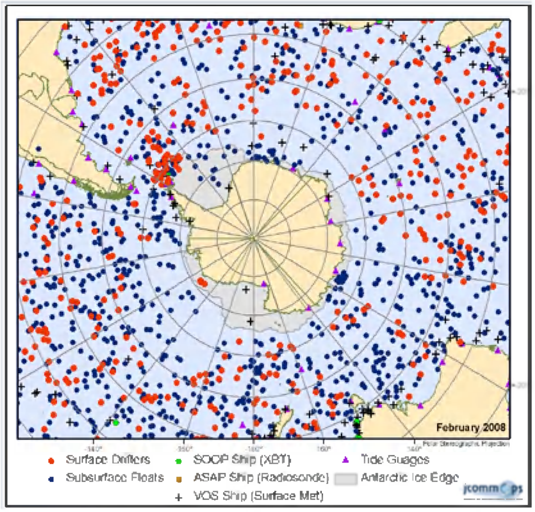


Figure 2.13a Observations available in the Southern Ocean region from JCOMMOPS (as of February 2008).

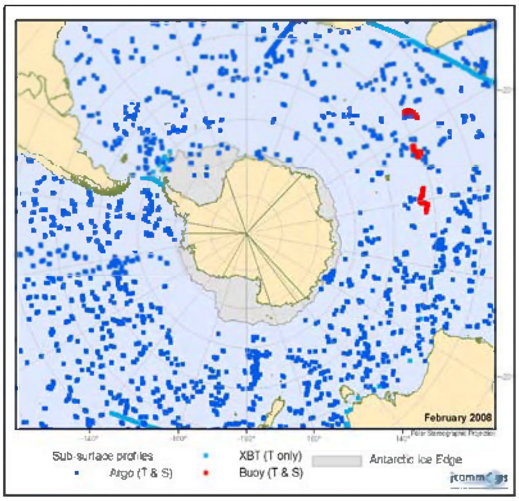


Figure 2.13b Sub-surface profiles available in the Southern Ocean region from JCOMMOPS (as of February 2008).

Currently we are not able to routinely monitor the characteristics of the ocean in the seasonally and permanently ice-covered region, which covers an area the size of Antarctica itself during the southern winter months. This is despite the efforts of extending Argo to the sea ice zone, the use of ice-tethered profilers and the inclusion of sensors on marine mammals that forage under the ice. Argo is currently also limited to the upper 2,000 m. In order to properly understand the processes that contribute to global change (e.g. understanding the overturning circulation) our monitoring efforts need to be extended to the deep ocean.

Measurements in the ice shelf environment, especially in the ice cavity regions beneath the ice shelves, are particularly difficult. Routine sustained monitoring is virtually unknown, but is required to understand how the ocean/ice shelf interaction will change as the climate alters, and what the impacts are for deep and bottom water formation and the global overturning in the ocean.

Because of the unique problems encountered at high latitudes, development of new sensors and methodologies is key, for example the addition of biogeochemistry sensors to Argo floats, or technology to study the long-term impact of seasonal ice cover on pelagic and benthic communities.

It is imperative to sample the polar oceans routinely and cost-effectively with an appropriate level of coverage to capture the main oceanographic and marine meteorological processes taking place that contribute to global change.

During the International Polar Year 2007-2008 - and beyond – one of the key aims is to monitor the Southern Ocean in a sustained manner (e.g. Summerhayes et al., 2007). This is already underway with the development of a Southern Ocean Observing System (e.g. Sparrow, 2007) that will incorporate the whole range of observations available to observe the marine physics and surface atmosphere, biogeochemistry and carbon, cryosphere and sea ice of the Southern Ocean.

2.1.3.9 Satellite Observations of the Ocean

Satellites underpin a vast amount of modern oceanography, and their utility is nowhere greater than in data-poor regions of the Southern Ocean. They include, for example: - altimeters, scatterometers, infra-red and microwave sensors for sea-surface temperature and ice extent and visible-wavelength radiometers for ocean colour. Ensuring their continuity is of paramount importance.

Sea Surface Temperature

Sea surface temperature is an important geophysical parameter, providing the boundary condition used in the estimation of heat flux at the air-sea interface. On the global scale this is important for climate modelling, study of the Earth's heat balance, and obtaining insight into atmospheric and oceanic circulation patterns and anomalies (such as El Niño). On the mesoscale, SST can be used to study ocean structure, such as eddies, fronts and upwellings and to assess biological productivity. In the remote Southern Ocean where *in-situ* measurements are particularly sparse, satellite observations of SST are key inputs to studies of physical and biological aspects of oceanography.

Satellite-derived SST is one of the longest duration continuous remote sensing datasets. Observations began in November 1981 using data collected by the AVHRR instruments on the NOAA polar orbiting operational meteorologicall satellites. This series is still operational and plans are in place to continue availability of AVHRR data from the European MetOp and U.S. National Polar-orbiting Operational Environmental Satellite System (NPOESS) series.

The direct broadcast capability of instruments such as the AVHRR adds value for real-time support to scientific cruises in the Southern Ocean.

Improved and consistent processing of AVHRR data has resulted in the Pathfinder dataset (http://podaac.jpl.nasa.gov/PRODUCTS/p216.html), which provides global datasets from 1985 to current at 4 km pixel spacing with improved ice and land masks. This has provided improved SST data sets of the Southern Ocean, which is a very cloudy region where detection of cloud and ice has been a problem in the past.

Other processing efforts include the Global High-Resolution Sea Surface Temperature (GHRSST) project (http://www.ghrsst-pp.org/), which aims to provide a new generation of global multi-sensor high-resolution (<10 km) SST products to the operational oceanographic, meteorological, climate and general scientific community.

Satellite SST data are also available from a range of other satellites, including the European Space Agency Along Track Scanning Radiometer (ATSR) instruments and more recently the MODIS instruments on the NASA Terra and Aqua satellites (Figure 2.14).

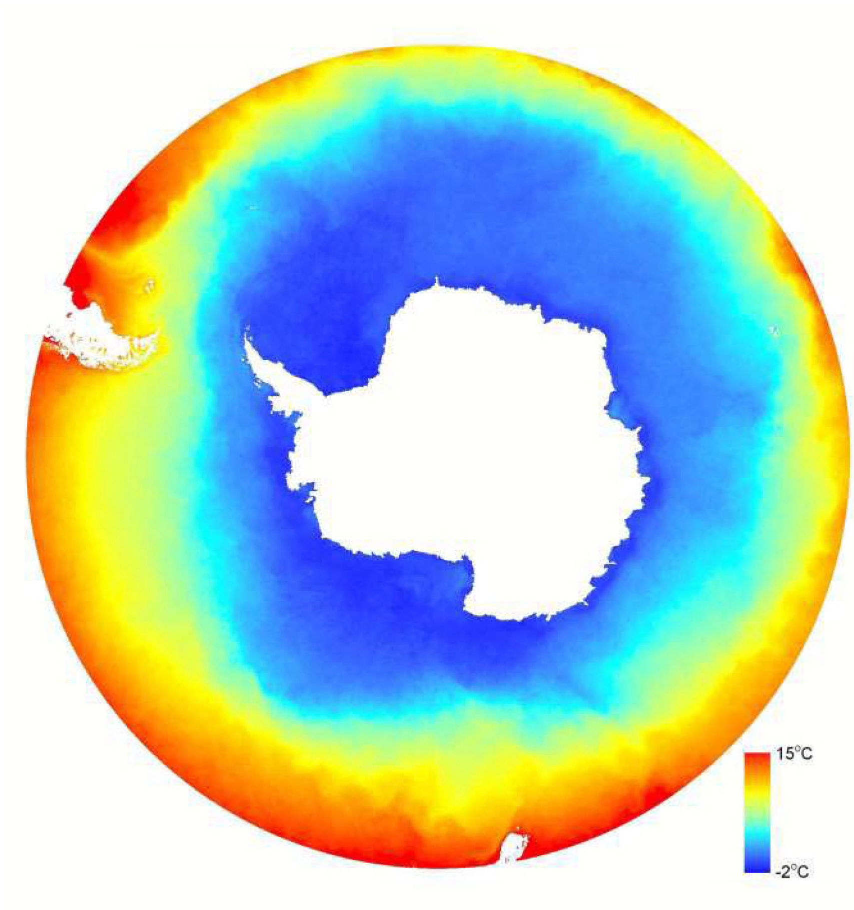


Figure 2.14 Southern Ocean averaged SST for January from Aqua MODIS sensor

Of particular application to the study of climate change are the ATSR instruments onboard the European Space Agency ERS and Envisat satellites. Operating since July 1991, ATSR is now available as consistently processed dataset with a target SST accuracy of 0.3°C (Corlett et al., 2006). This series of accurate space-based observations of SST is to be extended as part of the European Space Agency Sentinel series, onboard Sentinel-3 and currently planned for launch in 2012.

In addition to SST data derived from thermal infrared measurements, observations are also made using passive microwave measurements. Accuracy and resolution is poorer for SST derived from passive microwave measurements. The advantage gained with passive microwave is that radiation at these longer wavelengths is largely unaffected by clouds, and generally easier to correct for atmospheric effects. Instruments that have been used include the Scanning Multichannel Microwave Radiometer (SMMR) carried on Nimbus-7 and Seasat satellites, the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) and data from the AMSR instrument on the NASA EOS Aqua satellite.

Altimetry

Satellite altimeters provide data that are helpful for understanding both the ocean circulation (Gille et al., 2000; Hughes et al., 2001) and the geophysical characteristics of the sea floor (Sandwell and Smith, 1997) (Figure 2.15). Satellite altimeters provide long-term observations of mesoscale circulation patterns in the Southern Ocean and the variability of features such as the ACC. Satellite altimetry also provides information that is of use in mapping sea surface wind speeds and significant wave heights.

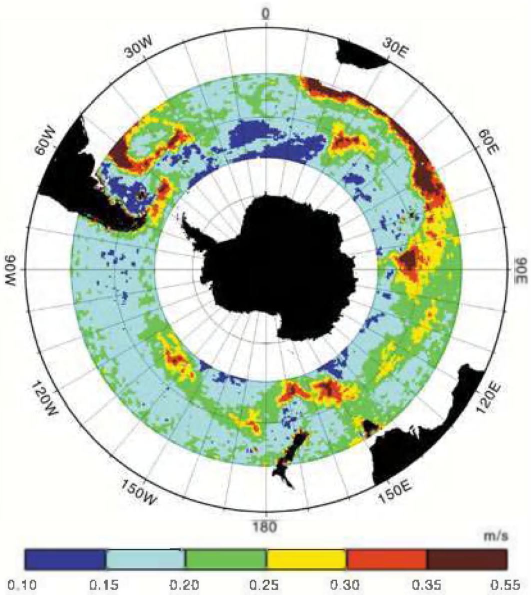


Figure 2.15 Satellite altimetry showing high Southern Ocean eddy kinetic energy in the core of the ACC (Gille and Sandwell, 2001)

In the polar oceans, sea ice and its associated snow cover is a major regulator of the heat, mass and momentum between the atmosphere and the ocean. Although ice extent and concentration are routinely measured from space, sea ice and snow thickness, particularly in the Antarctic, are not well measured and are highly uncertain. Methods for estimating sea ice thickness in the Arctic using satellite altimetry may have application for measuring Antarctic sea ice thickness, despite the difficulty of determining the smaller freeboard measurements of dominant first year ice in the Antarctic (Giles et al., 2006).

Radar altimeters are non-imaging radar sensors, which use the ranging capability of radar to measure the surface topographic profile parallel to the satellite track. They provide precise measurements of a satellite's height above the ocean and, if appropriately designed, over land/ice surfaces, by measuring the time interval between the transmission and reception of very short electromagnetic pulses.

To date, most spaceborne radar altimeters have been wide-beam (pulse-limited) systems operating from low Earth orbits. Radar altimetry data have been collected from a variety of instruments including Seasat (1978), Geosat (1985-1989), ERS-1 (1991-1996), Topex-Poseidon (since 1992), ERS-2 (since 1995), Jason-1 (since 2001) and Envisat (since 2002). The future availability of satellite altimetry observations is assured with new missions such as Jason-2 and the ESA Sentinel satellites. ESA's CryoSat-2 mission will also provide new data from the SIRAL instrument specifically designed to deliver higher resolution measurements for ice sheet and sea ice observations.

In addition to radar altimetry, the first satellite laser altimeter was launched in 2003 onboard the Ice, Cloud, and land Elevation Satellite (ICESat) satellite. The Geoscience Laser Altimeter System (GLAS) is the sole instrument on the ICESat (Figure 2.16). The main objective of the ICESat mission is to measure ice sheet elevations and changes in elevation through time. Secondary objectives include sea ice thickness, measurement of cloud and aerosol height profiles, land elevation and vegetation cover (Schutz et al., 2005). NASA's ICESat-2 mission will continue the legacy of ICESat from around 2014.

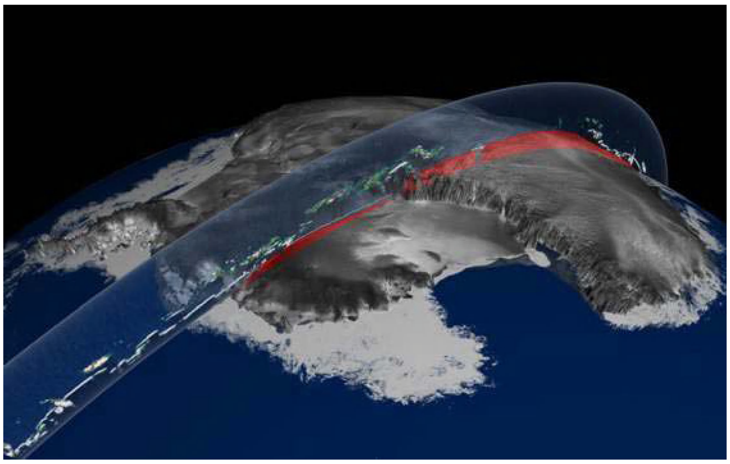


Figure 2.16 Schematic of a single ICESat repeat track indicating surface altimetry measurements (red) and cloud observations (green/white/pink). Source: NASA/GSFC.

Ocean colour

Satellite ocean colour sensors provide data on the concentration of chlorophyll in the ocean surface waters. Variations in ocean chlorophyll concentration contribute information on the biology of the Southern Ocean and are a key input for modeling of the South Ocean ecosystem. In a region characterised by high-nutrient, low-chlorophyll status, ocean colour satellite imagery has greatly enhanced our understanding of the true spatial and temporal extent of phytoplankton blooms (Figure 2.17) and has revealed that chlorophyll-*a* (chl-*a*) biomass can be particularly high in regions such as the Scotia and Weddell Seas (E.J. Murphy et al., 2006). Combined with other observations, chlorophyll measurements help to provide an understanding of how ocean productivity changes with weather, oceanographic variations such as the El Niño/South Oscillation, and other fluctuations in ocean temperature – work that could provide hints as to how future climatic change could affect ocean productivity (Behrenfeld et al., 2006).

Parts of the Southern Ocean ecosystem have also been highly perturbed as a result of harvesting over the last two centuries, and significant ecological changes have also occurred in response to rapid regional warming during the second half of the twentieth century. This combination of historical perturbation and rapid regional change suggests that the Scotia Sea ecosystem is likely to show significant change over the next two to three decades, which may result in major ecological shifts. Satellite measurements of chlorophyll will help obtain a comprehensive understanding of the evolution of the Antarctic marine ecosystem.

The accuracy of chlorophyll measurements varies from region to region globally. In the Southern Ocean it has been observed (Holm-Hansen et al., 2004) that satellite data underestimate chlorophyll values recorded *in situ* at high chlorophyll concentrations, and slightly over-estimate the shipboard data at lower chlorophyll concentrations.

Data suitable for measurement of chlorophyll concentration began in 1979 with the launch of the Coastal Zone Colour Scanner (CZCS) onboard the Nimbus-7 satellite. Intermittent collection of data continued until 1986. However, it was not until 1997 that global ocean colour observations resumed with the launch of the SeaWiFS (Sea-Viewing

Wide Field-of-View Sensor) instrument onboard the Orbview-2 satellite. SeaWiFS has now collected data for more than a decade, but continuity of ocean colour data continues concurrently with the MODIS instruments on the Terra and Aqua satellites and the MERIS instrument on the Envisat satellite. Future plans for the continuity of satellite chlorophyll measurements include instruments on the ESA Sentinel-2 satellite and the NOAA Polar-orbiting Operational Environmental Satellite System (NPOESS) series.

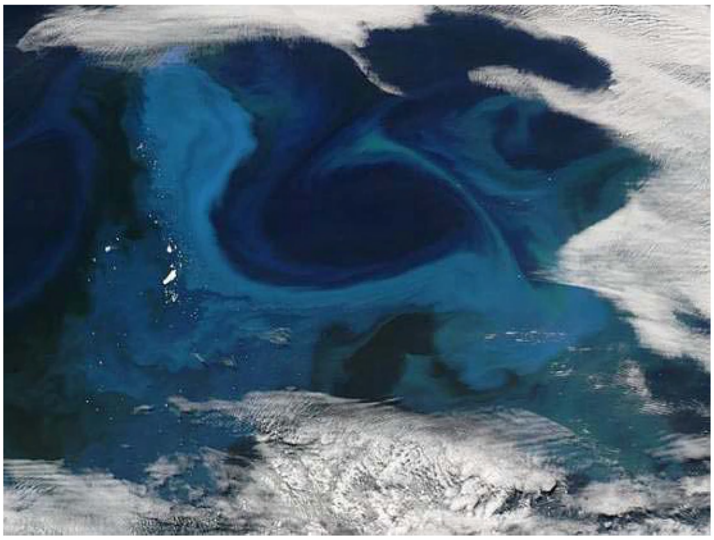


Figure 2.17 Phytoplankton bloom off South Georgia in the Southern Ocean acquired by the Aqua MODIS instrument

Scatterometer data

Radar scatterometer instruments have been designed principally to capture the near-surface wind field over the oceans. By assuming that the energy transmitted back to the radar from the ocean is dependent only upon that component of sea surface roughness that is a product of the frictional interaction of wind on the surface, a model may be used to relate this roughness, through the radar backscatter coefficient, to wind speed and direction. This is possible because the roughness is anisotropic, with crests and troughs generally orthogonal to the wind direction.

The primary use of scatterometer data in southern high latitudes is as data to be assimilated into numerical weather prediction (NWP) models. For example, Andrews and Bell (1998) showed a marked reduction in rms errors of UK Met Office forecasts over the Southern Ocean that assimilated scatterometer winds. However, the spatial resolution of scatterometer data — 25 km for the instrument on the ERS-1 satellite — means that it can also be used for case studies. It has been utilised to study both mesoscale and synoptic-scale weather systems around Antarctica (Marshall and Turner, 1997a, 1999) and coastal katabatic winds off the Ross Ice Shelf (Marshall and Turner, 1997b).

Figure 2.18, taken from Marshall and Turner (1999) demonstrates the type of information that scatterometer data can provide about weather systems over the Southern Ocean. Feature A is a mesocyclone and it can be observed in the scatterometer data: wind speeds are greater on the equatorward side and weaker on the poleward side of the centre than the background flow, and the wind field shows troughing in an equatorward direction. Feature B is a trough that is not apparent in the cloud imagery: the scatterometer data indicate the marked wind shear associated with this trough, with the wind direction changing by 135° within the resolution of the data. Finally, feature C is a dissipating synoptic-scale system located close to the Antarctic Peninsula. The scatterometer winds reveal a weak but closed surface circulation.

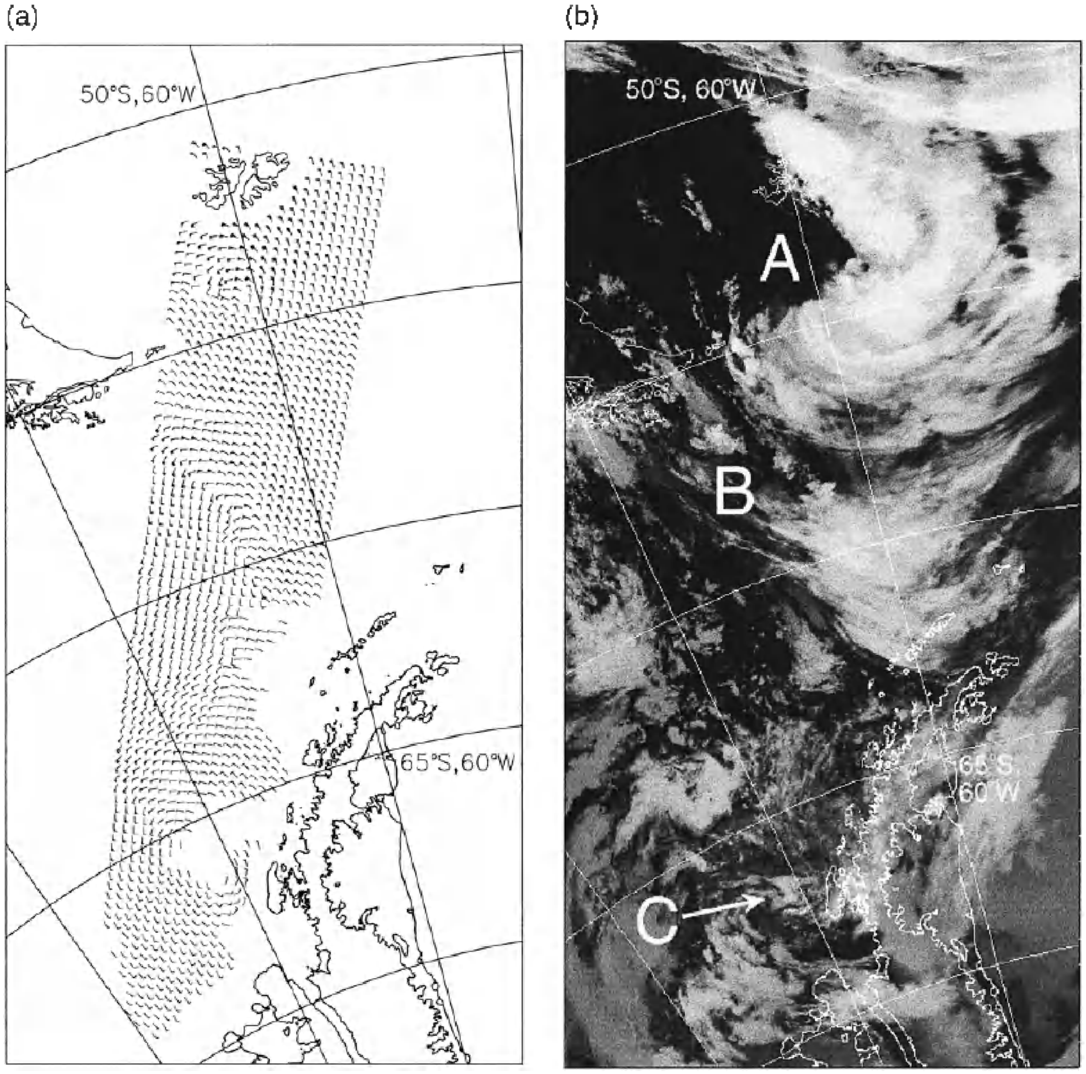


Figure 2.18 Near-coincident data over the Southern Ocean on 15 Jan 1995. (a) ERS-1 scatterometer swath acquired between 13:09 and 13:15 GMT; (b) Thermal infrared AVHRR imagery obtained at 13:11 UTC. Wind feathers point in the direction the wind is blowing; half a barb represents 2.5 m/sec, and a full barb 5.0 m/sec. Labelled features are described in the text. (Marshall and Turner, 1999; courtesy of the American Meteorological Society).

Scatterometer data can also provide useful information from ice-covered surfaces (Long et al., 2001): the return signal is dependent on the roughness and the dielectric properties (a measure of the ability of a medium to resist the formation of an electric field within it) of the surface and near-surface. Despite the relatively poor spatial resolution, useful information on the thermodynamic state, distribution and dynamics of sea ice at a regional scale can be tracked easily because of the frequent repeat coverage at polar latitudes (1-2 days). Similarly, over terrestrial ice sheets large-scale patterns of seasonal melt can be observed.

Synthetic Aperture Radar

Imaging Synthetic Aperture Radar (SAR) systems provide imagery of the ocean and land surfaces by recording the reflected signals of emitted microwave radiation (Figure 2.19). The active nature of these imaging instruments means data can be obtained throughout the year, day or night and regardless of cloud and weather conditions. Whilst initially difficult to interpret, they are useful in resolving features of sea ice, land ice and snow, and the oceans.

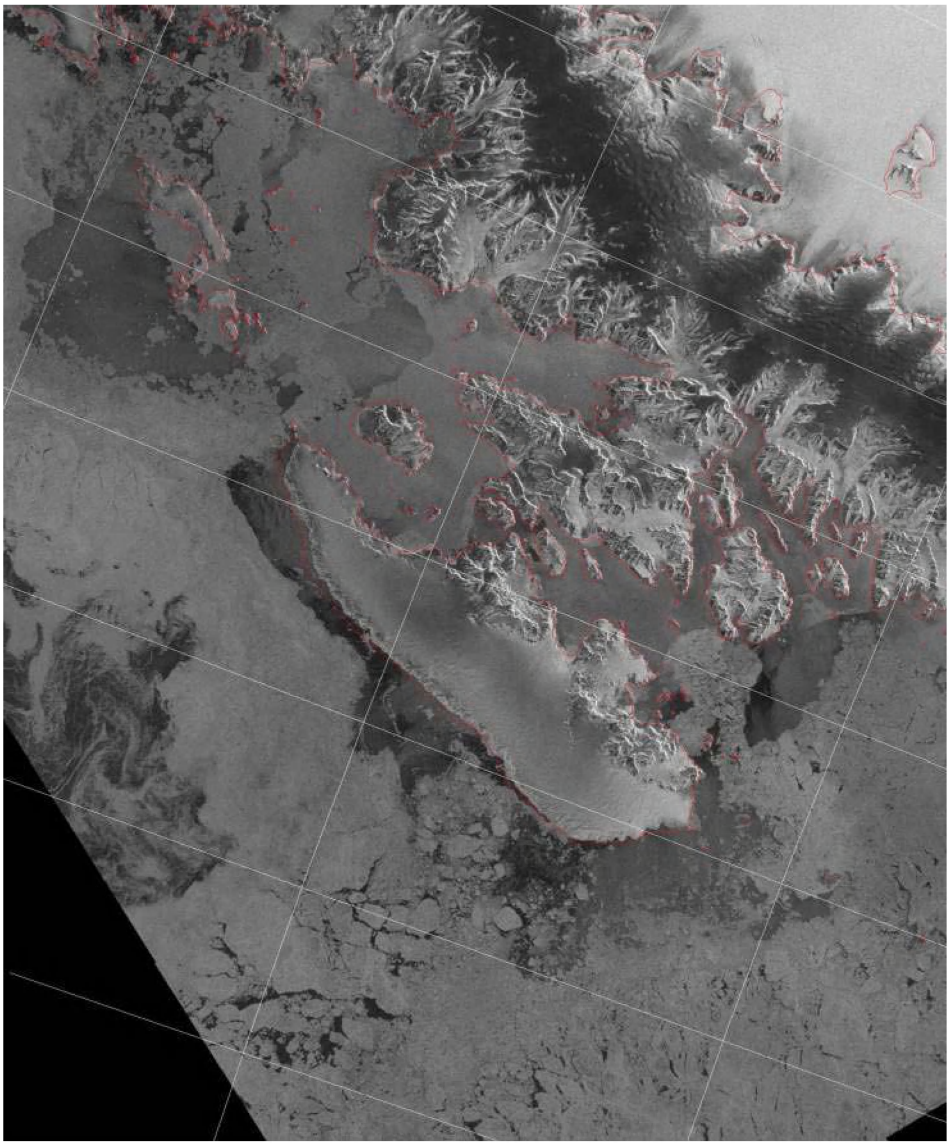


Figure 2.19 Envisat ASAR image showing sea ice around Adelaide Island, West Antarctic Peninsula

SAR imagery also has wide application to oceanographic studies. Imagery provides information about the interactions at the sea ice edge and the resolution of currents, frontal boundaries and internal waves.

Available SAR imaging satellites began in 1991 with the launch of the European ERS-1 satellite. Since then, ERS-2, Envisat and Radarsat satellites have provided continuity of C-band SAR imagery. In 1997, Radarsat even made an unprecedented in-orbit manoeuvre to allow the entire Antarctic continent to be imaged including areas south of 80° S, which are not normally viewable. This resulted in the RAMP dataset (http://bprc.osu.edu/rsl/radarsat/radarsat.html, which has subsequently been updated in 2000 to provide the MAMM (Modified Antarctic Mapping Mission) dataset, which includes repeat images of glaciers in order to determine velocity. Recently, other SAR systems including different wavelength radars have been launched, including Radarsat-2, the Japanese ALOS, TerraSAR-X and COSMO-SkyMed systems. In the future there are plans for continuity of the European SAR systems from the ESA Sentinel series, a Radarsat constellation, a planned addition to the TerraSAR system called Tandem-X that will deliver detailed elevation data and even discussion of a P-band radar system that may allow imaging of internal ice sheet structure.

2.1.4 Sea ice observations

2.1.4.1 The pre-satellite era

Since the days of the earliest explorers, ships' logs have recorded encounters with sea ice. Captain James Cook frequently reported the presence of sea ice as he tried to push south toward the continent, as did Captain Fabian von Bellingshausen during his exploration in 1831. Mackintosh and Herdman (1940) compiled a circumpolar map of the monthly variation of the average sea ice edge, based on data from ships' logs during the 1920s and 1930s. These were later updated and republished by Mackintosh (1972).

Whaling vessels also made many valuable observations of the postion of the sea ice edge. The region close to the ice edge is rich in food and many whales congregate there, attracting the whaling fleets. Most of these observations are for the summer season. De la Mare (1997) examined the whaling records, which provide the location of every whale caught since 1931. He suggested that there had been a big change in the location of the whaling vessels between the 1940s and the 1970s, with the summer sea ice edge having moved southward by 2.8° of latitude between the mid 1950s and the early 1970s. He inferred this as meaning that there had been a decrease of 25% in the area covered by sea ice. However, there is a great deal of debate over whether or not the locations of whale catches can be translated into ice edge estimates that are comparable to those made from satellite observations. It is particularly unfortunate that there is no overlap between the period covered by the whale catch data and the modern satellite observations of the ice edge. At the moment the de la Mare results are questioned in many quarters and cannot be regarded as proof of a major decrease in sea ice extent between the 1940s and 1970s.

Coastal stations have also made valuable sea ice observations, albeit from a very limited number of locations. Nevertheless, direct observation of the ice edge from land is difficult because the ice edge extends well into the Southern Ocean for much of the year, out of sight from most coastal observatories. Some island stations, such as Signy in the South Orkney Islands, have provided information on sea ice variability over many years, and revealed details of some important modes of climate variability, such as the Antarctic Circumpolar Wave (Nowlin and Klinck, 1986).

2.1.4.2 Satellite observations of sea ice extent and concentration

From the 1960s it was possible to obtain a broad-scale view of the distribution of sea ice from visible and infra-red satellite imagery. However, the imagery was only of value in cloud-free or partly cloudy conditions, which was a major handicap as the Antarctic sea ice zone is characterised by extensive low cloud cover. With the introduction of reliable satellite passive microwave observations in the early 1970s (Gloersen et al., 1992), the extent (the area bounded by the ice edge, which is often taken as 15% ice concentration) and area (the integrated area of ice within the ice edge) of Antarctic sea ice became confidently measureable.

The US Nimbus-5 Electrically Scanning Microwave Radiometer (ESMR) was launched in December 1972 and allowed the first all-weather mapping of Antarctic sea ice. This instrument only had one channel at 19 GHz, but the large contrast in the emissivity of sea ice and ice-free ocean enabled the development of an ice concentration algorithm, allowing the production of sea ice concentration maps from 1973 to 1976. The ESMR data provided the first observational data on the growth and decay patterns of sea ice for the entire Antarctic region.

Despite hopes for further developments in sea ice monitoring with the launch in 1975 of Nimbus-6/ESMR-2, with its dual polarized 37 GHz radiometer the instrument failed to perform well and no useful data were obtained. Further useful passive microwave data were obtained with the launch of the Scanning Multichannel Microwave Radiometer (SMMR), first on board the SeaSat satellite in July 1978, and then on Nimbus-7 in October 1978. The SMMR was a multifrequency system covering five frequencies from 6 to 37 GHz, all of them dual polarized (horizontal and vertical). The sensor was also conically scanning (i.e., incidence angle constant), and with its multifrequency capability ice concentrations were derived at a much better accuracy than with ESMR data. SMMR lasted for about 9 years, before it failed in August 1987, the DMSP/Special Scanning Microwave Imager (SSM/I) was already in operation and provided overlap data from mid-July to mid-August 1987. The SSM/I sensor has only 7 channels from 19 to 89 GHz, among which is the same set used for generating ice concentration maps from SMMR. The sensor is also conically scanning with similar resolutions, but has a wider swath; it has provided continuous data up to the present. The overlap allowed for comparison of the performance of the two radiometers and a confirmation that data from both sensors provided approximately the same results. In May 2002, the EOS/Advanced Microwave Scanning Radiometer (AMSR-E) was launched, and with 14 channels from 6 to 89 GHz, and much higher resolution, it has provided the baseline for sea ice studies.

The ESMR data set was very valuable and was unique when it first came out, but there were problems using it together with the other sets of data for time series studies. First, since it is a one-channel instrument, the ice concentration data are not as accurate because variations in temperature and emissivity of the ice cover could not be taken into account. Second, it is a horizontally scanning radiometer going from nadir to around 50^{o} with varying resolution and with different incident angles. Third, there were many missing bits in the data stream, causing the elimination of a large fraction of the data and big data gaps in the time series. And fourth, there was no overlap of ESMR and SMMR data to enable assessment of differences of ice edge locations and concentrations derived from the two sensors. For uniformity, and accuracy in the trend analysis in Section 4.7, we use data from the two sets of similar sensors (i.e., SMMR and SSM/I) to evaluate the variability and trends in the ice cover over the last 28 years. We also discuss, how we can utilize ESMR data as well as some ship observations during the pre-satellite era to improve our understanding of the long-term trend.

SARs flown on spacecraft can provide high resolution data on sea ice and reveal mesoscale ice motion and deformation, the development of leads and polynyas, ice type discrimination, sea ice roughness data and iceberg detection. In addition to the scientific applications of these images, ship operators in polar seas have benefited from recent advances in the near real time processing of SAR data, allowing them to be delivered quickly enough to assist ship navigation in sea ice.

2.1.4.3 Observations of sea ice thickness

While the early explorers made many observations of sea ice location and type, their logs do not contain information on sea ice thickness. Only in recent decades have vessels become more ice capable and spent more time south of the ice edge in support of logistic and scientific activities. Consequently the sea ice logs from these ships have become more comprehensive and often include an estimate of sea ice thickness, or ice type from which thickness can be inferred.

In 1997, SCAR established the Antarctic Sea Ice Processes and Climate (ASPeCt) programme. One of the programe's first objectives was to collate the many disparate sea ice logs kept from icebreakers operating in the Antarctic sea ice zone. This effort focused primarily on the Australian, German, US and Russian national Antarctic programmes, which were known to have dozens of data sets containing information on the concentration, thickness and snow cover characteristics of the Antarctic sea ice zone. The data constituted a compilation of 23,391 individual ship-based observations collected from 81 voyages to Antarctica over the period 1981 – 2005, plus 1,663 aircraft-based observations. The ship-based observations are typically recorded hourly and include the ship's position, total ice concentration and an estimate of the areal coverage, thickness, floe size, topography, and snow cover characteristics of the three dominant ice thickness categories within a radius of approximately 1 km around the ship (Worby and Allison, 1999). Not all observations contain this level of information, but at a minimum the partial ice concentrations and thicknesses (or ice types) were necessary for inclusion in the data set. The data are publicly available via the ASPeCt website (http://www.aspect.aq) or from the Australian Antarctic Data Centre (http://data.aad.gov.au/).

Time series of sea ice thickness can be measured by moored upward looking sonars. They consist of echosounders moored about 150 m below the ocean surface which record by travel time changes the variation of the ice draft. Mooring motions and changes of the sound velocities require extended corrections. Significant efforts occurred in the Weddell Sea to quantify the annual cycle of sea ice thickness and horizontal transports in the gyre circulation.

A new method for sea ice thickness measurements is by means of non-destructive electromagnetic inductive (EM) sounding which can be performed by moving an EM instrument above the snow surface either airborne or terrestrially.

It is hoped that in the future, processing of satellite altimeter data will allow the recovery of sea ice freeboard and therefore ice thickness. This may be easier in the Arctic since there sea ice is generally thicker and there is more multi-year ice present. In the Antarctic, the predominance of first year ice may make this a great challenge.

2.1.4.4 Sea ice extent proxies

ITASE and other Antarctic ice coring programmes are in the process of developing proxies for sea ice, a critical component in the climate system, through studies of sulfur compounds such as sulfate and MSA (methane sulphonic acid) (Welch et al., 1993; Curran et al., 2003; Dixon et al., 2006, Abram et al., 2007a). ENSO-sea ice connections are noted utilizing ice core MSA and sulfate series over the Ross Sea embayment region (Meyerson et al., 2002). The sea ice proxies rest on the premise that biogenic sulfur emissions from within the sea ice zone, when integrated regionally and through time, may be related to the total sea ice extent. This approach has been supported by observed correlations between MSA and sea ice (Curran et al., 2003; Foster et al., 2006, Abram et al., 2007a), however it is likely to be dependent upon the regional sea ice regime. For example, the positive relationship observed by Curran et al. is in the East Antarctic, where there is little residual summer sea ice and very little multi-year ice. Results from East Antarctica show evidence of a 20% decline in mean sea ice extent since the mid-20th century, consistent with results reconstructed from whaling records (de la Mare, 1997) although the validity of such observations is questioned (Ackley et al., 2003). However this continuous sea ice proxy illustrates large decadal-scale variations superimposed on the trend.

Curran et al. (2003) have demonstrated that MSA variability in the coastal DSS ice core can be used as a proxy for regional sea ice extent. Accordingly they reconstructed the sea ice extent between 80° to 140° E back to 1840 AD, and showed the consistent decline in sea ice extent since the 1950s (decreasing MSA in the DSS ice). However, regional sea ice extent is a function of ocean temperature and circulation and atmospheric windfield divergence and convergence patterns. Hence, changes in the longwave atmospheric pattern may cause contrasting regional temperature and windfield patterns. Preliminary analysis of MSA concentrations in the GF12 core (Queen Mary Land, East Antarctica) indicates that MSA concentrations increased from 1950 to 1986 when the core was retrieved. These analyses may indicate a growth in regional sea ice to the west of 90° E. This is consistent with increased southerly wind outflow from the Lambert Basin and cool SST anomalies in this region. Hence, ice core sites located at different longitudes have the potential to provide historical data on regional sea ice changes around Antarctica. Figure 2.20 shows the June – August MSLP pattern associated with June - August MSA variations recorded in the GF12 ice core in Queen Mary Land, near 100° E.

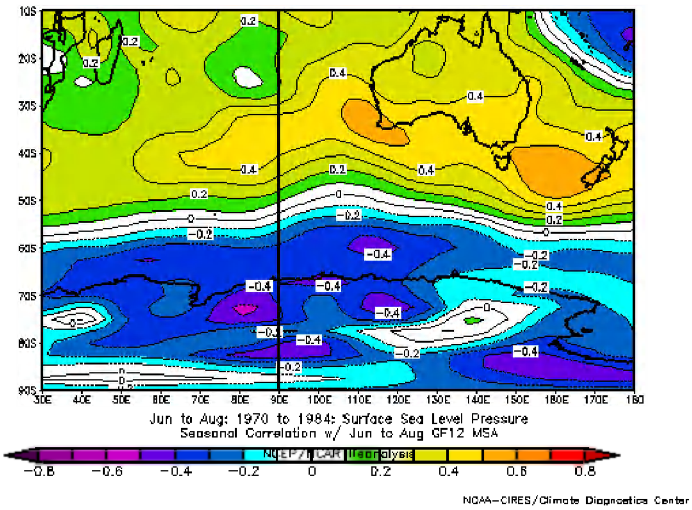


Figure 2.20 The cross correlation pattern with R values for June - August MSLP anomalies for the Indian Ocean and Southern Ocean region and MSA variability at GF12, (97° E) in Queen Mary Land. This shows the strong relationship between MSA and the SAM.

Sea-salt levels have been proposed as an additional proxy of sea ice extent (Wolff et al., 2003) and applied to the long EPICA ice core record (Wolff et al., 2006). The basis of this proxy is the observation from near-coastal sites that ‘frost flowers’ are a significant source of sea-salt species. These ‘flowers’ form on new ice under calm conditions, producing fragile crystalline structures that are vulnerable to destruction and transport by wind. Salt species from frost flowers provide a distinguishing signature resulting from fractionation of sulfur content relative to other species. At coastal sites, where winter snow can be isolated, this signature is evident, however it appears to diminish in relative importance at sites further inland (M. Curran, personal communication) and its utility as a proxy at sites in the interior has not yet been clearly demonstrated.

2.1.5 Observations of the ice sheet and permafrost

2.1.5.1 Introduction

Since the earliest exploration of Antarctica scientists have visited the ice plateau in search of new discoveries. Despite the logistic and physical difficulties imposed by the environment and weather, field-based observations have been made throughout Antarctica, and have proved invaluable in explaining the dynamic character of this body of ice. An era of international fieldwork, sparked by the IGY of 1957-58, has continued to the recent International Polar Year of 2007-08. In the intervening decades, glaciological fieldwork adopted new technologies and revealed behaviour that could not have been observed by any other means. Some measurements can now be made over wide areas from satellites or aircraft, but direct *in-situ* observations are still needed to provide calibration and validation for these surveys. Moreover, fieldwork provides detailed information about the processes that operate to change the Antarctic environment. The combination of fieldwork, aircraft, and satellite measurements has revealed changes in Antarctica on many scales and given a better understanding of their causes.

2.1.5.2 Surface elevation

Oversnow traverses provided the first indications of the surface shape of the ice sheet. The existence of broad, flat ice shelves bounded by tall mountain ranges that retained a high, frigid and harsh polar plateau played into the strategies used by explorers to reach the South Pole. However, such sparse sampling could never hope to attain sufficient coverage of the ice sheet to detect its shape in detail. In the mid-1970s, random sampling of the surface from radar altimeters onboard balloons released to circulate above the continent at constant atmospheric pressure greatly increased the mapping of the ice sheet's surface elevation (Levanon et al., 1977). Radar altimeters use the ranging capability of radar to measure the surface topographic profile beneath the sensor by measuring the time interval between the transmission and reception of very short electromagnetic pulses. These data showed the ice sheet surface topography contained a number of separate domes in East Antarctica and that this ice sheet was much higher than the West Antarctic ice sheet (Figure 1.2). With this information, the drainage basins of individual outlet glaciers could begin to be defined (Figure 1.3).

Radar altimeters onboard satellites established systematically repeated surface elevation observations of the earth. To date, most spaceborne radar altimeters have been wide-beam (pulse-limited) systems operating from low Earth orbits. Radar altimetry data of Antarctica has been collected since 1978 from a variety of instruments including Seasat (1978), Geosat (1985-1990), ERS-1 (1991-1996) and ERS-2 (since 1995), and Envisat (since 2002). Seasat reached only to 72° S whereas subsequent missions have increased coverage to 81.5° S. The measurement accuracy of such systems is sub-metre across the smoother ice sheet interior, but degrades to a few metres as the surface slope increases, especially nearer the coast and in mountainous regions. The future availability of satellite altimetry observations is assured with new missions such as ESA's Sentinel satellite series. ESA's CryoSat mission will also provide new data from the SIRAL instrument specifically designed for ice sheet and sea ice observations.

In addition to radar altimetry, the first satellite laser altimeter was launched in 2003 onboard the ICESat satellite. The Geoscience Laser Altimeter System (GLAS) is the sole instrument on the Ice, Cloud, and land Elevation Satellite (ICESat) (Figure 2.16). A laser altimeter has the advantage of ranging directly to the surface without penetrating the upper snow layers, but requires cloud-free conditions. Microwave radar is unaffected by clouds, but penetrates several metres into the snow (Legresy and Remy, 1997; Arthern et al., 2001). Alteration of the snowpack by weather and climate could change the depth of penetration, with a risk that this is mistaken for a real change in the surface elevation. Recent studies have corrected for this effect using an empirical, location-dependent correlation between penetration depth and backscattered power (Wingham et al., 1998; Zwally et al., 2005), although some residual error may be unavoidable (Alley et al., 2007). Both radar and laser altimetry can be affected by changes in the density of the upper layers of the snowpack, so these must be estimated separately or corrected for (Arthern and Wingham, 1998; Zwally et al., 2005).

The primary objective of the ICESat mission is to measure ice sheet elevations and changes in elevation through time, with a precision of a few centimetres, surpassing that achieved with radar altimetry. Secondary objectives include measurement of cloud and aerosol height profiles, land elevation and vegetation cover, and sea ice thickness (Schultz et al., 2005). The southern limit of ICESat data is 86° S. The increase in surface topographic detail in these data is illustrated in Figure 1.3 by noting the smoother apparent surface in the region closest to the South Pole; a visual artifact caused by the absence of precise altimetric data in this region.

Most modern airborne geophysical measurement systems operated in Antarctica include a laser altimeter instrument. These have provided independent validation of ICESat data. Their primary function is to focus on regional studies and refine the spatial pattern of elevation change, as indication of ice-sheet or outlet glacier thickness changes.

Satellite altimeter systems have the drawback that their data usually do not extend away from the sub-satellite ground track. ICESat's GLAS system can be pointed off-nadir, but these occasions are rare. Photoclinometry, or shape from shading interpolated between precisely measured elevation profiles by using optical imagery (Landsat, ASTER and MODIS). These systems collect images of the surface using reflected sunlight in a variety of spectral bands. Because most of Antarctica is snow covered and the albedo of snow at any given wavelength is relatively constant, the brightness variations of ice-sheet imagery are predominantly related to the surface slope in the direction of the solar illumination (Bindschadler and Vornberger, 1998). This relationship allows the image to be used to interpolate surface elevations between satellite altimetry profiles and provide more complete and accurate elevation fields.

A more traditional use of optical imagery to produce elevations is stereo photogrammetry. This can be problematic over ice sheets where distinct features visible in modest-resolution imagery can be uncommon, but with the recent addition of higher spatial-resolution sensors such as ASTER and SPOT-5, the technique is now more often successful. A dedicated set of SPOT stereo images has been acquired around the Antarctica perimeter and elevation data are being produced as users request them.

2.1.5.3 Bed elevation and ice thickness

As little as fifty years ago, relatively few traverses had been made across the interior of Antarctica, and the thickness of the ice was uncertain over much of the continent. Ice thickness can be determined by the seismic method of detonating explosives and recording the delay before arrival of echoes from the base. This can reveal details of the rock or sediment beneath the ice, as well as the ice thickness (Blankenship et al., 1986). A similar technique using radar, rather than sound waves can be performed from the ground (e.g. Conway et al., 1999; Catania et al., 2006), or from aircraft (Robin et al., 1970; Mae and Yoshida, 1987; Siegert et al., 2005a).

Airborne radar (see section 2.1.2.8), sometimes flown by airplanes based at remote field camps, is now the principle method of determining ice thickness (Figure 2.21). Present technology allows the surface of the underlying rock and sediment to be measured to a vertical precision of tens of metres. Interpolating the rugged bed topography between surveyed tracks remains difficult, and even now there are large regions of Antarctica where few measurements have been recorded. Nevertheless, compiling data from many survey flights can give a good estimate of the volume of ice stored in Antarctica. The most recent compilation of these data is BEDMAP (Lythe et al., 2000) and BEDMAP-2 is underway. Figure 1.5 is a visual illustration of the Antarctic bed elevation field that has resulted from this data compilation. From these data, the total volume of the Antarctic ice sheet is calculated as equivalent to 57 m of sea level. Of this, 52 m is locked away in the thick ice of East Antarctica, and around 5 m is stored in the more dynamic ice sheet covering West Antarctica (Lythe et al., 2000).

Figure 2.21 Twin Otter aircraft fitted with ice penetrating antennas mounted under the wings. Aircraft fitted with geophysical instruments have collected a wealth of ice thickness data throughout Antarctica.

Several areas of Antarctica have recently benefited from focused airborne-geophysical surveys. Coastal Dronning Maud Land (Steinhage et al., 1999), Thwaites Glacier (Holt et al., 2006), Pine Island Glacier (Vaughan et al., 2006) have been surveyed in detail. These and similar surveys reveal the amounts of ice present. They also characterise deep troughs in the underlying bed that make some regions more vulnerable than others to rapid change. Furthermore they provide geometric boundary conditions needed for modeling the evolution of the ice sheet.

2.1.5.4 Velocity

In-situ observations of ice-flow can be made by repeatedly surveying the position of marker poles carried along by the ice (e.g. Naruse, 1979). Nowadays this same direct approach makes use of precise locations derived from GPS. Modern dual-frequency GPS receivers and post-processing of data can give positional accuracy of centimetres or better (King, 2004). The movement of AWS over time along the ice sheets, which have GPS on-board, can also provide invaluable information about ice sheet velocity. Ferrell AWS, located on the Ross Ice Shelf, is estimated to have moved approximately 0.65 - 0.75 km/year over a period from 1979 to 2003. Ice motion is almost everywhere greater than a metre per year, so surface velocities are measured to within one percent in a single year. For faster flowing ice, the relative accuracy is even higher. Point velocities measured in this way provide constraints to calibrate remote sensing methods such as feature tracking or satellite radar interferometry (discussed below).

Features seen from space can serve the same role as surveyed marker poles or GPS antennas. Sequential imagery, once coregistered, offer the opportunity to measure the surface motion of features identified on multiple images. Image coregistration is simplified when there is exposed rock visible. When not, long-wavelength surface forms can be used, because these are surface expressions of fixed basal undulations and, therefore, also fixed in space. Image processing techniques are then employed to first filter out the higher wavelength features and then cross-correlate static long-wavelength features in multiple images. Once coregistered, the same cross-correlation techniques are employed to determine surface feature motion (Scambos et al., 1992). Most often applied to sequential Landsat imagery with a 30 m pixel resolution, the errors in measured displacements are 2-3 pixels with roughly equal parts random error associated with tracking individual features, and systematic error associated with image coregistration.

The optical imagery record began in earnest in 1972 with the launch of ERTS-1, which became the first of the Landsat series. Earlier and recently declassified satellite photographs have extended a sparse data set back to 1960, soon after the first year that satellites were used for military and national security purposes. Spatial resolution, temporal coverage and radiometric sensitivity have steadily improved to the present, when sub-metre pixel resolution, near daily coverage and extraordinary surface detail can be obtained. However, not all of these are available from any single sensor, and the use of the data will help determine what sensor is most suitable.

Much information about land ice and snow is obtained from SAR imagery, in particular from SAR Interferometry (InSAR) techniques. This method, which uses phase differences in returned radar signals, is able to measure centimetric changes in surface displacement over very wide areas (Figure 2.23). These new techniques have provided information to monitor glacier discharge (Pritchard and Vaughan, 2007), map grounding lines (Gray et al., 2002; Yamanokuchi et al., 2005), contribute to mass balance studies (Rignot and Thomas, 2002), determine glacier and ice sheet motion (Fischer et al., 2003) and monitor propagation of ice shelf rifts (Fricker et al., 2002). In contrast, an approach taken in matching the amplitudes of SAR images can be used in correlating scenes over a long time period. Nakamura et al. (2007) applied the techniques to the Shirase glacier, one of the fastest-moving ice streams in the Antarctica, and obtained the seasonal variation of velocity near the grounding line, that is the highest in summer and lowest in winter.

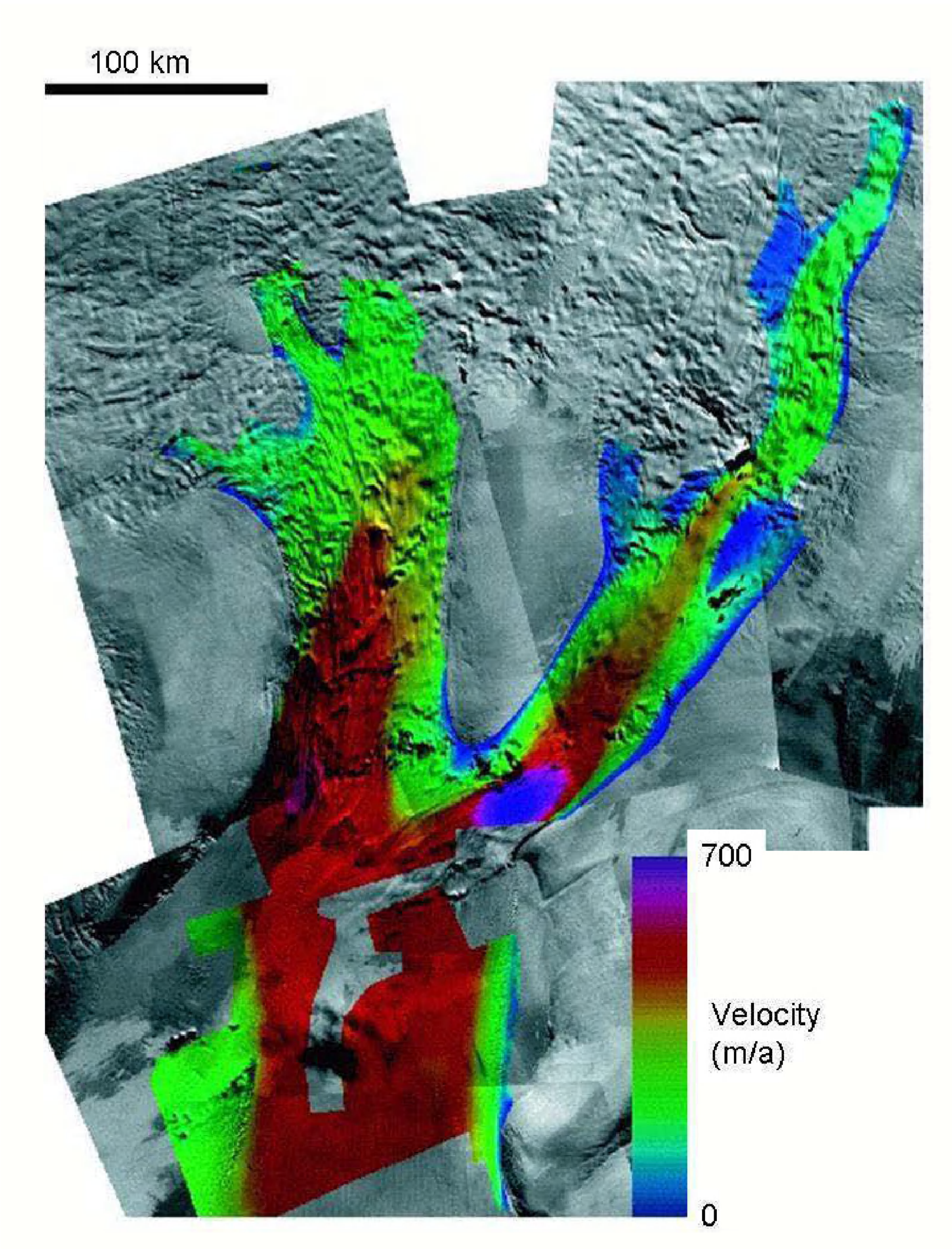


Figure 2.22 Speed of Bindschadler (right) and MacAyeal (left) Ice Streams, West Antarctica. Flow direction is generally from top to bottom of figure with speed indicated by color superimposed on a mosaic of 14 Landsat images enhanced to highlight surface features. Faster ice occurs in the central portion of each ice stream but the spatial complexity indicates a spatial variation in the lateral, longitudinal and basal forces that determine internal stresses, strains and speeds of the ice. Velocity in metres/year.

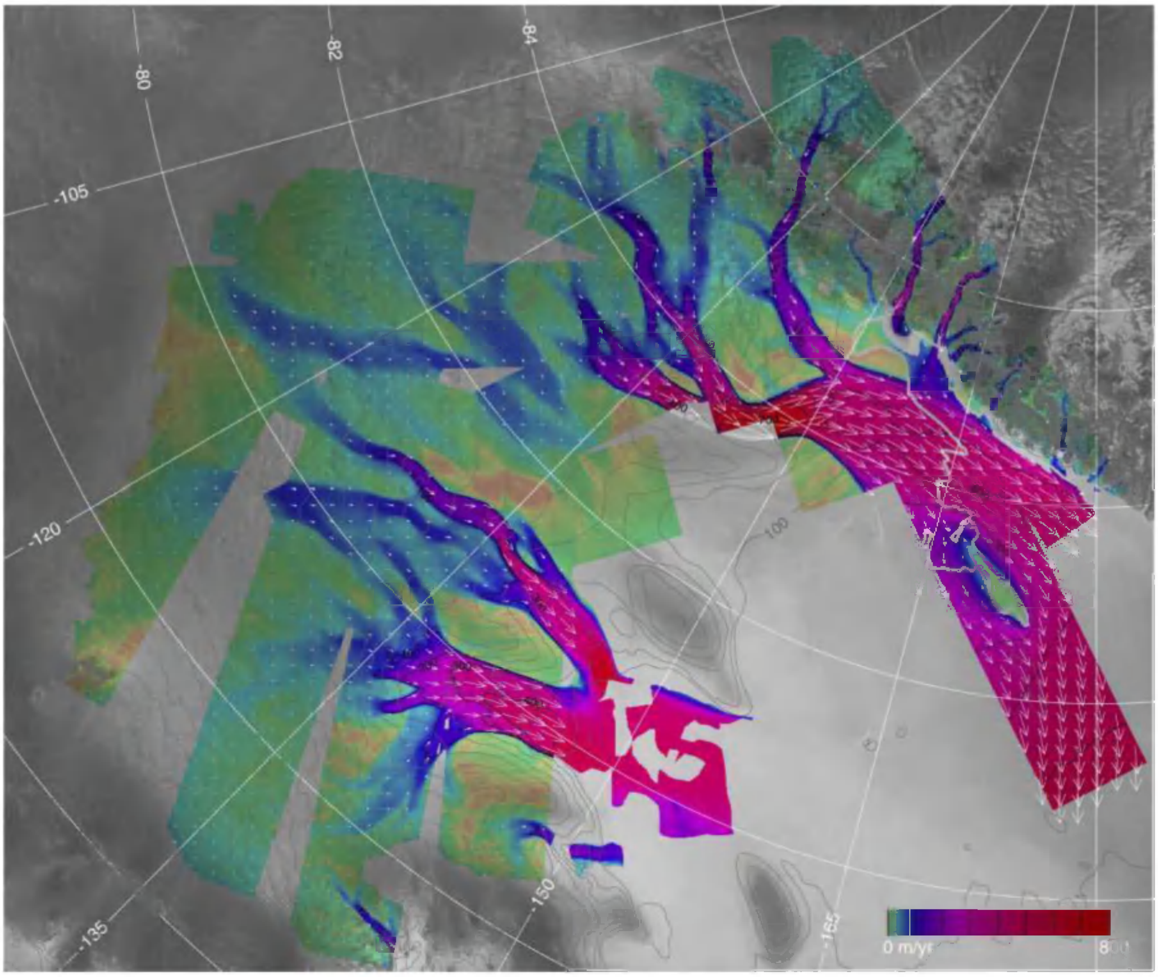


Figure 2.23. Surface velocity (in metres/year) of the Siple and Gould Coast regions of West Antarctica. Flow direction, indicated by the arrows, is generally left to right, and speed, indicated by colour, increases from left to right. The underlying layer is a portion of the Radarsat mosaic with the Transantarctic Mountains in the upper right and the Ross Ice Shelf on the right. The network of tributaries begins to emerge from the surrounding ice at speeds of about 100 m per year and is guided by unseen valleys in the underlying bed. From top to bottom, the major ice streams (red) are the Whillans (with the green Crary Ice Rise in its mouth), Bindschadler and MacAyeal Ice Streams.

Figure 2.24 is the most complete compilation of surface speed yet produced (Jezek, 2008).

Repeated measurements of horizontal velocity using the GPS technique have shown that the speed of ice streams can change on very short timescales. The speed of ice tens or hundreds of kilometres inland can vary from hour to hour, responding to forcing by the tides (Bindschadler et al., 2003). Other areas respond differently to tides. Flow speed on the Rutford Ice Stream follows a 14 day cycle (Gudmundsson, 2006). Analysing the response of ice streams to known tidal forcing has opened up a new way of studying the friction that impedes the motion of ice.

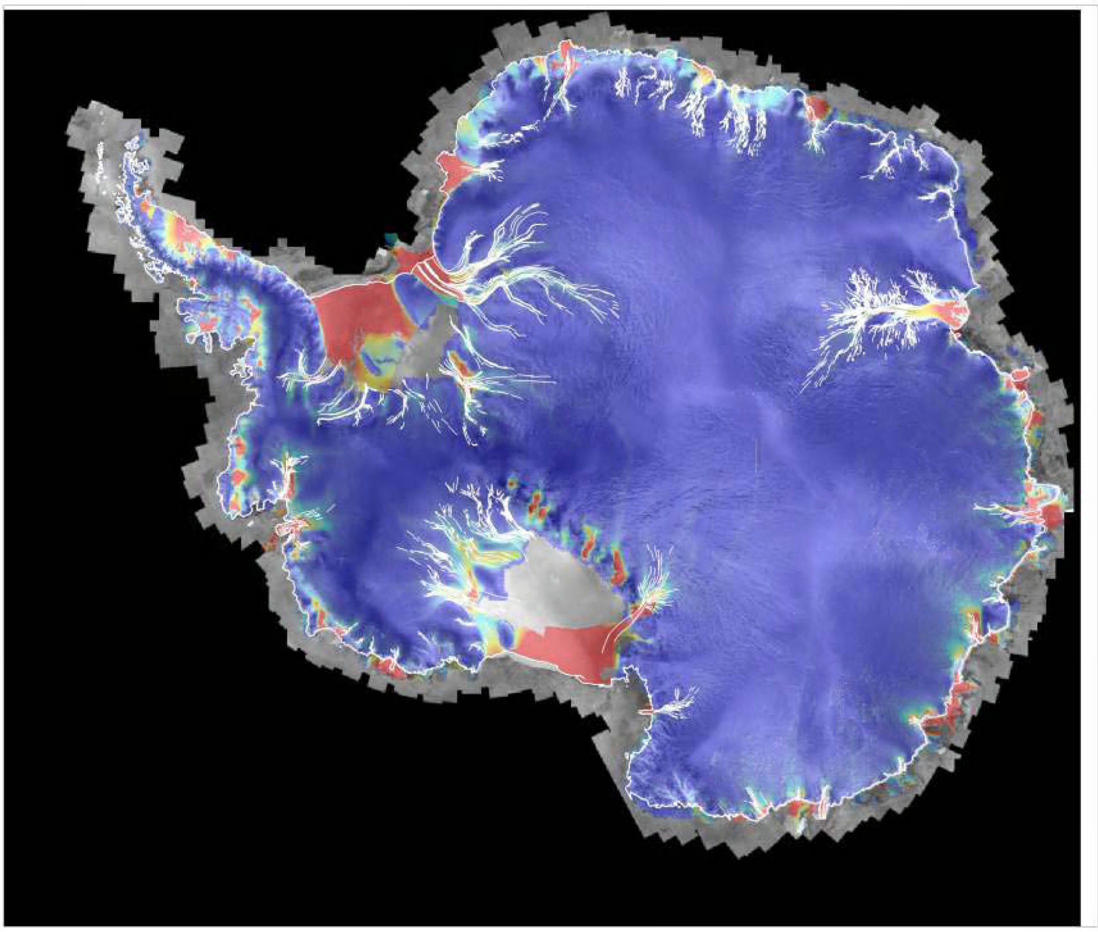


Figure 2.24 Composite of surface speed. North of 80° S the velocities are derived from the Modified Antarctic Mapping Mission (MAMM) whose data were collected by Radarsat-1 in Autumn 2000. South of 80° S the velocities are calculated balance velocities. Overlain on these two data sets are flow lineations (in white) from the SAR imagery indicating the location of faster ice motion and shear. Speed is represented by a color log10 scale (0 in deep blue to 1,000 m/yr in red) to illustrate the wide range and patterns of surface flow. The underlying layer is a gray-scale version of the Radarsat mosaic where bright and dark areas correspond to regions of high and low radar backscatter. From Jezek, 2008.

A meaningful diagnostic of the ice sheet state is to calculate the "balance velocity" and compare it with the measured surface velocity. The balance velocity is calculated from the ice thickness, accumulation pattern and flow direction (defined as the direction of maximum surface gradient). It is neither a measured, nor an actual velocity, but corresponds to the depth-averaged velocity at any point that is required to transport all the mass accumulated upstream of that point. If the actual velocity equals the balance velocity everywhere, then the shape of the ice sheet remains unchanged. Thus, balance velocity becomes a convenient way of comparing the geometric shape and meteorological input to the ice speed. Figure 1.6 presents the continental pattern of balance velocity. It compares favourably to the surface speed (Fig 2.24) indicating that a majority of the ice sheet is near equilibrium. Areas of change, including areas changing very rapidly, do occur and will be discussed in Chapter 4.

2.1.5.5 Mass balance

The overall condition of an ice sheet is usually described as its “mass balance”. It is a fundamental expression of the ice sheet’s health. Mass balance is the sum of all mass inputs to the ice sheet, such as accumulation, minus all mass removals, such as melting. It expresses the rate of growth or shrinkage of the ice sheet and relates directly to the amount of mass transferred to the oceans to affect sea level. Ice leaving the grounded region of the ice sheet affects sea level immediately, even if it remains attached to the ice sheet in the form of a floating ice shelf. Icebergs calved from floating ice shelves and sub-ice shelf melting remove mass from the ice sheet, but do not alter sea level. Sometimes ice shelf mass changes are included in mass balance calculations, because they reflect changes in ice sheet volume, but care must be taken not to equate these mass balance quantities to Antarctica’s contribution to sea level without removing the ice shelf component.

There are currently three independent methods of measuring changes in the mass of the grounded portion of the ice sheet: satellite altimetry, ice-budget, and gravity surveys. All are subject to errors from different sources: principally, snow compaction and electromagnetic penetration for altimetry; thickness errors and accumulation errors for the ice-budget; unmodelled postglacial rebound and atmospheric effects for gravity surveys. The problem is over-determined (three methods for measuring one number) and each of the methods has errors that are (more or less) independent of the others. This means that a better assessment of the state of balance of the ice sheet should be possible by combining all the observations. There are problems doing this, because all the measurements are recorded at different locations, and over different time intervals. A promising approach is to use data assimilation methods similar to those used by numerical weather prediction centres to produce daily weather forecasts (Arthern and Hindmarsh, 2006). This should allow a better analysis of the present-day changes, and opens up the possibility of forecasting the evolution of the ice sheet over the coming decades. Even before such formal methods are applied, comparison of the three approaches provides a valuable consistency check.

The concept of mass balance can also be applied to any portion of the ice sheet. To measure the amount of ice transported, the thickness and the profile of velocity through the column are needed. If the base is slippery, or the ice is afloat, an assumption that ice speed is constant at all depths throughout the column is appropriate. Then the ice flux is simply the product of surface velocity and thickness. For ice thickness of the order of kilometres, the depth error from radar surveys is of the order of one percent, and dominates over the velocity error. If shearing is present in the column it must be corrected for, and this can introduce errors larger than one percent in the flux. The accuracy that can be achieved for the flux leaving any particular drainage basin is typically of the order of ten percent of the turnover (e.g. Fricker et al., 2000; Rignot and Thomas, 2002).

Since measurements began, the overall mass balance of the ice sheet (i.e., the rate of change of the total mass of ice held within it) has been slightly negative, but with large uncertainties. This belies the enormous regional variability of mass changes that express different dynamic characteristics and responses to different climate forcings. Some signals, such as thinning of the Amundsen Sea sector in West Antarctica, are common to all three approaches, and this agreement provides a strong confirmation that this part of Antarctica is contributing to sea level rise. Meanwhile East Antarctica seems close to balance, or thickening slightly, according to these independent assessments.

Point measurements of elevation change, achieved by repeat GPS positioning, can also contribute to mass balance studies (Smith et al., 1998; Hamilton, 2005; Wendt et al., 2009). The vertical component of position derived by GPS is less accurate than the horizontal. The vertical motion of the marker poles must be corrected for along-slope advection, gradients of snow accumulation, and snow compaction. Placing the markers at the bottom of boreholes drilled through the upper layers of the ice sheet, where most of the density changes occur, can lessen the impact of variable snow compaction on these measurements (Hamilton, 2005). The submergence velocity of markers, compared with the long-term rate of snow accumulation (from ice cores), provides a local estimate of the state of balance of the ice sheet.

2.1.5.6 Ice shelves

Ice shelves, the floating fringe of the ice sheet, are often overlooked because of the absence of a direct impact on sea level, however they have a very important indirect effect. Ice shelves impart forces on grounded glaciers and ice streams, slowing the delivery of ice to the ocean (Thomas, 1973). This means that collapse or melting of a floating ice shelf can trigger a subsequent rise in sea level, as glaciers accelerate when this force is removed. This force is determined not only by the size, lateral extent and temperature of the ice shelf, but also its shape (Walker and Holland, 2007). Drilling allows temperature profiles through the ice to be recorded, and samples of the ice to be recovered for mechanical tests of strength and deformation strength (e.g. Rist et al., 2002). Local mass balance assessments are important, as melt rates at the base of ice shelves can reach several metres a year. Drilling provides access to the oceanographic environment beneath the ice shelf, so that the processes that control melting from the base can be investigated (Nichols and Jenkins, 1993; Craven et al., 2004). Phase-sensitive radar can be used to measure thinning rates (Corr et al., 2002). This can provide a direct estimate of the basal melt-rate with an accuracy of a few centimetres per year, revealing whether the ice shelf is in a steady state (Jenkins et al., 2006). GPS observations can record the flow velocities of the floating ice, and how it is changing (King et al., 2007). These *in-situ* observations provide an important test of melt rates, flow rates and thinning rates observed using satellite.

The very definition of where the ice shelves begin needs quantification, and its sensitivity to local ice thickness changes lends it to be an important diagnostic observable. The transition from grounded ice to floating occurs at the "grounding line". It can be mapped with optical imagery by noting the change in surface roughness: floating ice is smoother due to the absence of basal friction, which supports smaller spatial scale surface undulations on grounded ice. Vertical flexing of the grounding line also can be observed either by interferometric SAR analysis or repeat laser altimetry (Vaughan, 1995; Padman and Fricker, 2005).

The sensitivity of the grounding line to thickness changes results from a shallow basal slopes. Relatively small changes of a few metres in thickness are effectively amplified by many orders of magnitude, depending on the bed slope, to horizontal displacements of hundreds of metres, or even kilometres. In many locations, the grounding line is extremely complex, and patches of ephemerally grounded ice must be included in the observations.

2.1.5.7 Subglacial hydrology

Ice flow rates in the faster moving ice streams and glaciers of Antarctica are orders of magnitude greater than can be explained by ice deformed by gravitational forces. The additional speed is the result of basal sliding and directly related to lubrication at the ice-bed interface. The degree of lubrication is controlled by the presence of water and sediment (glacial till) underneath the ice. A supply of water from melting at the base of the ice sheet can pressurise subglacial water to the point that the ice is close to flotation. This lessens the load on the underlying sediment and allows it to deform easily, lubricating the sliding. Drag from the edge can be comparable to drag from beneath, even when the width of an ice stream greatly exceeds its thickness. The energy that is needed for melting is provided partly by geothermal heat, and partly by frictional heating (Raymond et al., 2001). Frictional heating is greater for faster sliding, but is also modulated by the degree of lubrication, so a feedback loop links melting to lubrication, speed, friction, and further melting. Depending upon the environment, this loop may reinforce itself, so that ice streams accelerate or decelerate rapidly. In other circumstances variations in velocity are muted, especially if other physical controls, such as trough geometry, limit the margin migration (Raymond et al., 2001). Field measurements have shown that ice streams can flow steadily, have episodes of rapid flow, or shut down completely, depending on details of supply, storage and transport of water and sediment (Retzlaff and Bentley, 1993; Stokes et al., 2007).

In addition to the direct observations of surface velocity (discussed earlier), seismic methods have illuminated some of the processes that control ice stream flow. Although the base of ice streams tend to be well lubricated, the importance of small areas of concentrated friction has been identified by seismicity characteristic of their stick-slip motion (Anandakrishnan and Alley, 1994). These 'sticky spots' provide significant retardation to the flow of ice and have a number of possible causes, including reduction of basal water pressure by freezing, channel formation, or redirection of subglacial water flow (recently reviewed by Stokes et al., 2007). Analysis of waveforms reflected from the bed during active seismic sounding can reveal information about whether the sediments are mobile and fluidized, or whether they are lodged and strong enough to support large shear stress retarding the flow of ice (Smith, 2007).

Understandably, there are not many direct observations from beneath the ice streams, but in several places access to the bed has been achieved by hot-water drilling. Video cameras lowered down the borehole (Carsey et al., 2002) have revealed clear ice at the bottom of Kamb ice stream, supporting the interpretation that basal freezing may have contributed to its shutdown around 140 years ago (Vogel et al., 2005). Underneath the ice stream, the camera revealed a water layer 1.6 metres in depth at one site, but just centimetres or less at another. The water flux and linkages between such water cavities are important in determining the basal water pressure and this, in turn, affects the amount of lubrication (Christoffersen and Tulaczyk, 2003; Vogel et al., 2005). The route taken by subglacial water flow is sensitive to surface and bed topography, and changes in surface slope that alter this routing may affect the water pressures, and hence the flow of ice (Alley et al., 1994).

The discovery of thicker reservoirs of subglacial water has led to more direct inference from repeat satellite observations that large volumes of water move between subglacial lakes (Gray et al., 2005; Wingham et al., 2006; Fricker et al., 2007) and that these larger volumes of water transfer change ice flow rates. These build on a series of discoveries that flow rates change on a wealth of time scales from minutes to millennia. A directed programme of fieldwork has illuminated many causes of this fast and changeable flow (e.g. Alley and Bindschadler, 2001; Bindschadler, 2006). Much of this dynamism can ultimately be traced to the slipperiness of basal sediment pressurised by meltwater.

2.1.5.8 Mass balance data from ice cores

Ice core records offer a valuable tool to assess natural variability and recent trends in snow accumulation. Climate model predictions lead to a general expectation of an increase in Antarctic precipitation with projected warming. Such an increase acts to offset sea level increases arising from mass loss. Recent results, based on records from ice cores, snow pits and stakes suggest an absence of a warming signal in precipitation over the last 50 years (Monaghan et al., 2006a), with no significant changes in overall precipitation. The data do show, however, a large temporal and spatial variability that underscores the need for mass balance studies to include long-term records from a number of sites.

ITASE research reveals high variability in surface mass balance, such that single cores, stakes, and snowpits do not always represent the geographical and environmental characteristics of a local region (Richardson and Holmlund, 1999; Frezzotti et al., 2004; Spikes et al., 2004; Nishio et al., 2002). For example, Frezzotti et al. (2004) show that spatial surface mass balance variability at sub-kilometre 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 reveals 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 GPR profiling integrated with core data, provides detailed information on surface mass balance (Richardson and Holmlund, 1999; Urbini et al., 2001; Arcone et al., 2005; 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 have been surveyed over continuous traverses of 5,000 km and can be used as historical benchmarks to study past accumulation rates (Spikes et al., 2004).

2.1.5.9 Observations of Antarctic Permafrost

The permafrost thermal regime is monitored by recording the temperature at different depths within boreholes. Traditionally, active layer measurements are performed by annual probing of the maximum thickness of seasonal thaw within a 100 × 100 m grid with a span of 10 m in each of the 121 grid points marked on the field, according to the CALM protocol (Nelson et al., 1998).

To monitor the depth of the 0°C isotherm, the temperature of the active layer is recorded at different depths, at least during the summer season.

However, in Antarctica the probing method has been tested for some years (Guglielmin, 2006) with poor results because of the coarse grain size of the main part of the terrain.

For these reasons, a practical and basic method to monitor the active layer thickness has been proposed for Antarctica. It consists in measuring at each grid point the ground temperature at different depths (2, 5, 10, 20 and, where possible, 30 cm) with a needle thermistor at one time during the period of maximum thawing. The active layer thickness is calculated by interpolating the two deeper temperature measurements (Guglielmin, 2006).

Active layer thickness depends primarily on ground surface temperature (GST) and the thermal properties of the ground, especially its ice/water content. Not all sites are sensitive to climate change, because heat convection, lateral heat advection and the thermal properties of the ground produce a “thermal offset” (Romanovsky and Osterkamp, 1995). This is the difference between mean annual GST and mean annual permafrost temperature, which can change with time, making difficult the modelling of the relationships between climate and active layer thickness.

Both the permafrost thermal regime and the active layer thickness are mainly related to air temperature (Guglielmin, 2004) and snow cover (Guglielmin, 2004, Zhang and Stammes, 1998), although the incoming radiation can be important especially on bare ground surfaces.

Vegetation cover can significantly influence the ground thermal regime by changing the snow thickness and permanence, the wind flow near the surface, and, therefore, the sensible heat and latent heat fluxes and, consequently, the net energy balance of the surface (Oke, 1987; Cannone et al., 2006; Guglielmin et al., 2007).

Permafrost and active layer monitoring network development in the last 50 years

Since the 1960s on many occasions, especially in maritime Antarctica, ground temperature has been measured for different purposes, through different protocols, in general for short periods (up to 2-3 years). In many cases only the active layer was investigated (Chambers, 1967; Kejna, M. and Laska, 1999; Matsuoka et al., 1990; McKay et al., 1998; Nichols and Ball, 1964; Robertson and Macdonald, 1962; Sawagaki, 1995; Thomson et al., 1971; Walton, 1977; Wójcik, 1989).

Permafrost and active layer monitoring sites are located both in continental Antarctica (Victoria Land and Queen Maud Land) and in maritime Antarctica (Table 2.1). Five deep boreholes (up to 282 m) drilled during the 1970s during the Dry Valley Drilling Project (DVDP) may also be available to measure permafrost temperature (Decker, 1974).

Despite the relatively small extent of ice-free areas, the distribution of the monitoring sites is insufficient to characterise these areas, which may show large spatial variability in the active layer (see Figure 2.25 left). In continental Antarctica active layer monitoring is generally more difficult than in Arctic areas, because spatial variability is very large due to the heterogeneity of the coarse sediments as well as the high surface roughness.

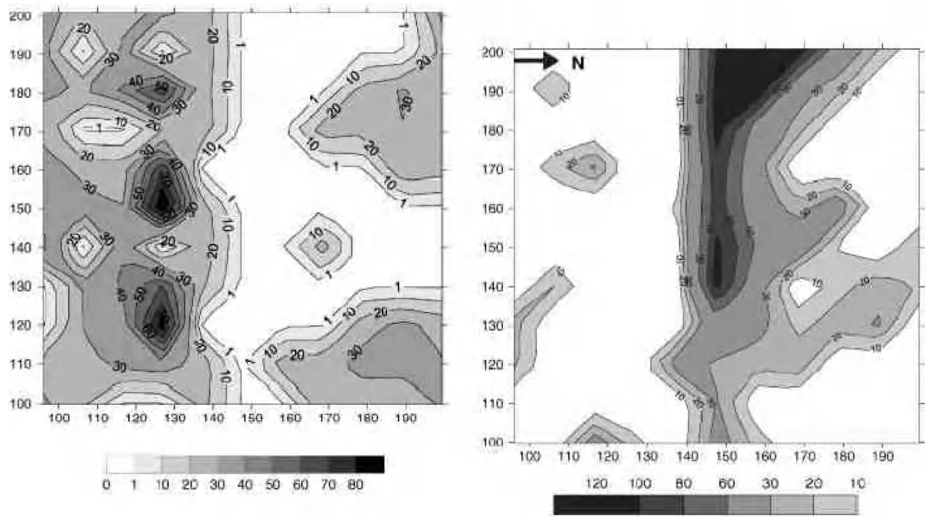


Figure 2.25 Example of active layer spatial variability. The Boulder Clay CALM grid (100 × 100 m) is located on a flat area close to the Mario Zucchelli Station. Top - active layer thickness (as maximum depth of 0° C isotherm, in cm) on 7 January 2002; lower - snow thickness recorded on the same date. Note the influence of snow accumulation on the active layer thickness distribution. (from Guglielmin, 2006).

The monitoring grid (100 × 100 m) located at Boulder Clay, a flat area close to the Mario Zucchelli station in Northern Victoria Land (Guglielmin, 2006), provides a striking example of the large areal variations in the active layer (Figure 2.25 left), which is mainly driven by different snow accumulation (Figure 2.25 right) influenced by surface roughness. There are only two other grids (Simpson Crags and Signy Island) but in both cases monitoring could not be repeated every year.

Future sites used in the framework of the IPY project “ANTPAS” will improve the network.

Site	*coordinates*	*Elevation (m)*	*Depth (m)*	*Monitoring type*	*P.I.*
		Southern Victoria Land			
Scott Base	77°51'S 166°46'E	38	1.2	P, C (1999)	1
Beacon Valley	77°51'S 160°36'E	1273	19	P,R	4
Mt. Fleming	77°33'S 160°17'E	1697	0.75	P,C (2002)	1
Bull Pass	77°31'S 161°52'E	152	1.2	P,C (1999)	1
Wrigth Valley	77°31'S 161°51'E	150	29.7	P,C (2007)	2
Minna Bluff	78°30'S 166°45'E	38	0.84	P,C (2003)	1
Marble Point	77°25'S 163°41'E	50	1.2	P,C (1999)	1
Marble Point	77°24'S 161°51'E	90	30,2	P,C (2007)	2
Victoria Valley	77°20'S 161°37'E	412	1.1	P,C (1999)	4
Victoria Valley	77° 20'S 161°37' E	380	11	P,R (2003)	3
Granite Harbour	77°00'S 162°31'E	4,6	1,2	P,C (2003)	1
		Northern Victoria Land			
Adelie Cove	74°47'S 163°58'E	35	6,1	P,R (2003)	3
Boulder Clay	74°45'S 164°01'E	205	3,6	P,C (1996)	3
Oasi	74°42'S 164°08'E	84	15,5; 30,3	P,C (2001;2008)	3
Mt. Keinath	74°33'S 163°59'E	1100	1	P,C (1998-2004)	3
Simpson Crags	74°26'S 162°53'E	830	7,8	P,C (1998-2002)	3
		Maritime Antarctica			
Livingston Island 1	62°39'S 60°21'W	35	2.4	AL,C (2000)	5
Livingston Island 2	62°39'S 60°21'W	275	1.1	P,C (2000)	5
King George Island	58°17'W 62°13'S	17	3; 6	P,R (1989)	6
James Ross Island	63°54'S 57°40'W	25	8,3	P,R(2000)	7
James Ross Island	63°54'S 57°39'W	10	2,3	P,R(1999)	8
Marambio Island	64°14'S 56°37'W	200	8	P,R (1999)	8
Signy Island	60°44'S45°36'W	90	2,5	P,C (2005)	9
		Queen Maud Land			
Svea	74°34'S 11°13'W	1286	1,2	P,C (2003)	10
Wasa	73°02'S 13°26'W	450	1,2	P,C (2003)	10
Fossilbryggen	73°24'S 13°02'W	550	1,2	P,C (2003)	10

Table 2.1 – locations of the sites where active layer and permafrost (P) or only active layer (AL) temperature are recorded. The temperature measurements are carried out continously all year round for more than 2 years (C) or not (R). The number indicated between parentheses is the first year of measurement. The column P.I. indicates the Principal Investigator or the data source: 1) www.wcc.nrcs.usda.gov; 2) Guglielmin M. and Balks M; 3) Guglielmin M.; 4) Sletten R. ;5) Ramos M; 6) Chen X); 7) Guglielmin M., Strelin J., Sone T., Mori J.; 8) Sone et al., 2001; 9) Guglielmin M. and Worland R.; 10) Boelhouwers J.

The thickness of permafrost in Antarctica varies with region (Bockheim and Hall, 2002). In the McMurdo Dry Valleys of interior Antarctica (77° S, 161-166° E), it ranges from 240 to 970 m (Decker and Bucher, 1977). In North Victoria Land (74° S, 164° E), permafrost varies from 400 to 900 m in thickness (Guglielmin, 2006). Along the northeastern Antarctic Peninsula at Seymour and James Ross Island, it ranges from 15 to 180 m in thickness depending on elevation above sea level (Borzotta and Trombotto, 2003). Permafrost is sporadic in the South Shetland Islands and monitoring programmes show that it is generally continuous above c. 100-150 m asl and discontinuous lower down. Since the islands are very mountainous, most of the South Shetlands show the presence of permafrost (Ramos and Vieira, 2003). Permafrost temperatures, normally measured at a depth of 50 m, range from -14 to -24° C in continental Antarctica (Decker and Bucher, 1977). The temperature of permafrost at a depth of 10 m in NVL ranges between -12 to -17° C (Guglielmin, 2006).

The moisture content of permafrost in Antarctica is reflected by permafrost form. Whereas the gravimetric moisture content of ice-bonded permafrost averages 40%, the moisture content of dry-frozen permafrost may be <3% (Campbell and Claridge, 2006). A minimum of 6-7% moisture is required for ice bonding. There is considerable small-scale variation in moisture content of permafrost in Antarctica (Campbell and Claridge, 2006).

Although the age of Antarctic permafrost is not known, it is likely that it developed after the final breakup of Gondwana and the initiation of glaciers at the Eocene-Oligocene boundary, ca. 40 million years ago (Gilichinsky et al., 2007). Buried glacial ice in upper Beacon Valley (77.83° S, 159.50° E) may be 8 million years in age (Marchant et al., 2002).

Active-layer dynamics

The active layer refers to the layer of seasonal thawing. In Antarctica seasonal thawing is at a maximum in early February. The active layer varies between 5 and 80 cm in the MDVs (Guglielmin et al., 2003). The Circumpolar Active Layer Monitoring – Southern Hemisphere (CALM-S) Project is monitoring active layer dynamics at 16 sites in Antarctica, including 12 sites in the MDV (Figure 2.26), 2 in North Victoria Land, and 2 in the South Shetland Islands.

Variations in active layer thickness bear a strong relationship to fluctuations in air temperature, particularly during the summer months. For example, at Marble Point (77.4° S, 163.8° E), the seasonal thaw or active layer thickness varied from 30 cm in 2001 to 60 cm in 2002 (Figure 2.27); 2002 had unusually high summer temperatures and extensive flooding in the MDV. The gravimetric moisture content of the active layer in southern Victoria Land typically ranges between 1 and 10% (Campbell et al., 1997).

Ongoing ground temperature monitoring

Current ground temperature monitoring in Antarctica is done under the auspices of the Global Terrestrial Network for Permafrost (GTN-P). Ground temperature is being monitored in 11 boreholes in Antarctica to depths ranging from 2.4 to over 30 m, including five boreholes in the MDV, four in North Victoria Land, one in the South Shetland Islands, and one in the South Orkney Islands. Data loggers at these sites are monitoring temperature within the active layer and the permafrost. Electrical Resistivity Tomography (ERT) and Refraction Seismic Tomography (RST), and other electrical techniques are being used to detect and characterize structures containing frozen materials in Antarctica (Borzatta and Trombotto, 2003; Hauck et al., 2007).

At the 16 CALM-S sites, soil temperature is measured at approximately 10 cm increments in the upper 1 m and at lower frequencies to a depth of 7.8 m (Guglielmin et al., 2003).

PERMAMODEL, a project involving the University of Alcalá and the University of Lisbon, is studying permafrost dynamics on Livingston and Deception Islands in the South

Shetland Group (Ramos and Vieira, 2003). The project entails (i) long-term monitoring of permafrost and active layer temperatures, (ii) identification of factors controlling ground temperatures, (iii) inverse modeling of climate signals from ground temperature data, and (iv) spatial modeling of permafrost distribution.

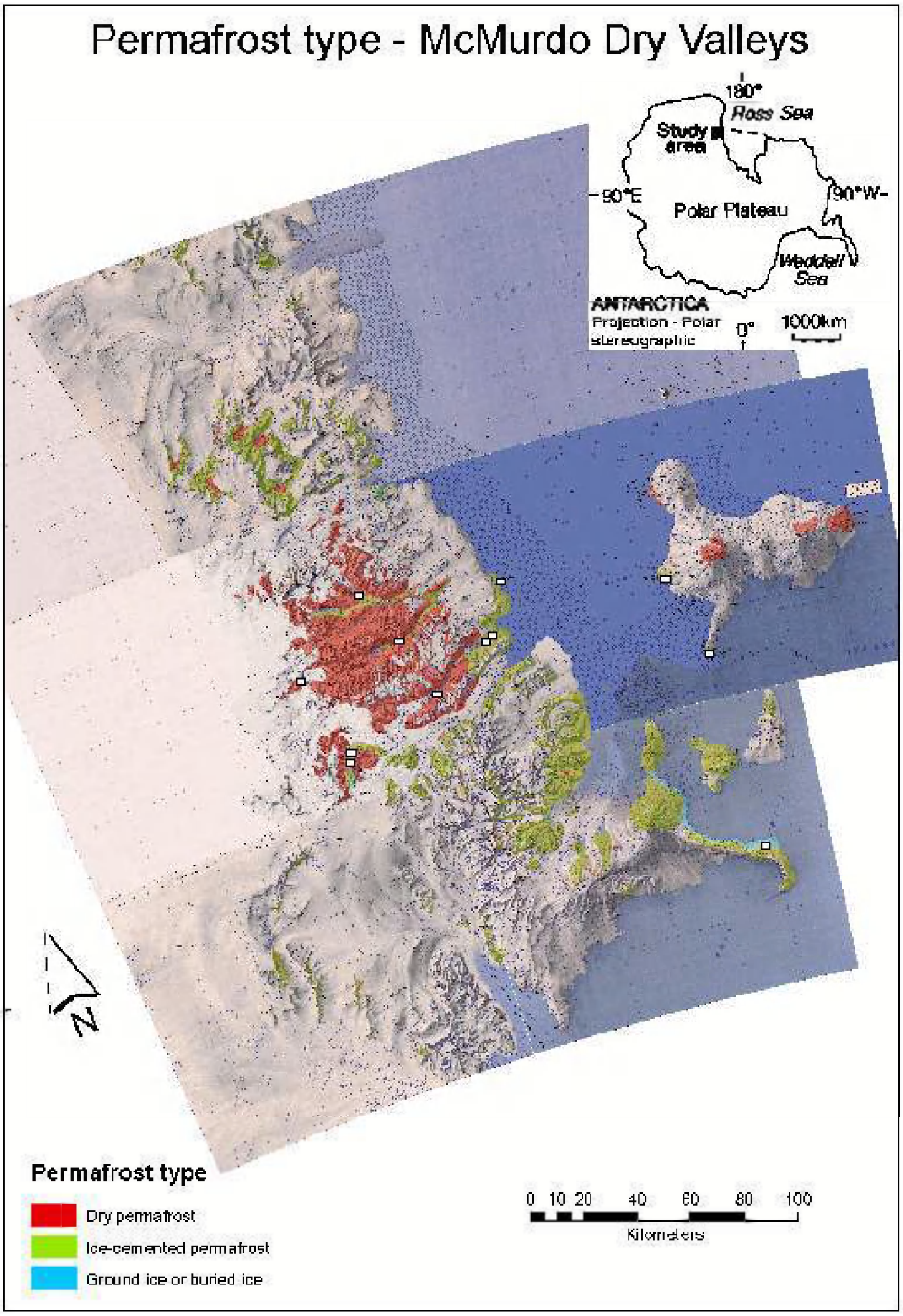


Figure 2.26 Reconnaissance map (1:2 million scale) showing permafrost distribution by form in the McMurdo Dry Valleys (Bockheim et al., 2007). Squares show Circumpolar Active-Layer Monitoring – South (CALM-S) sites.

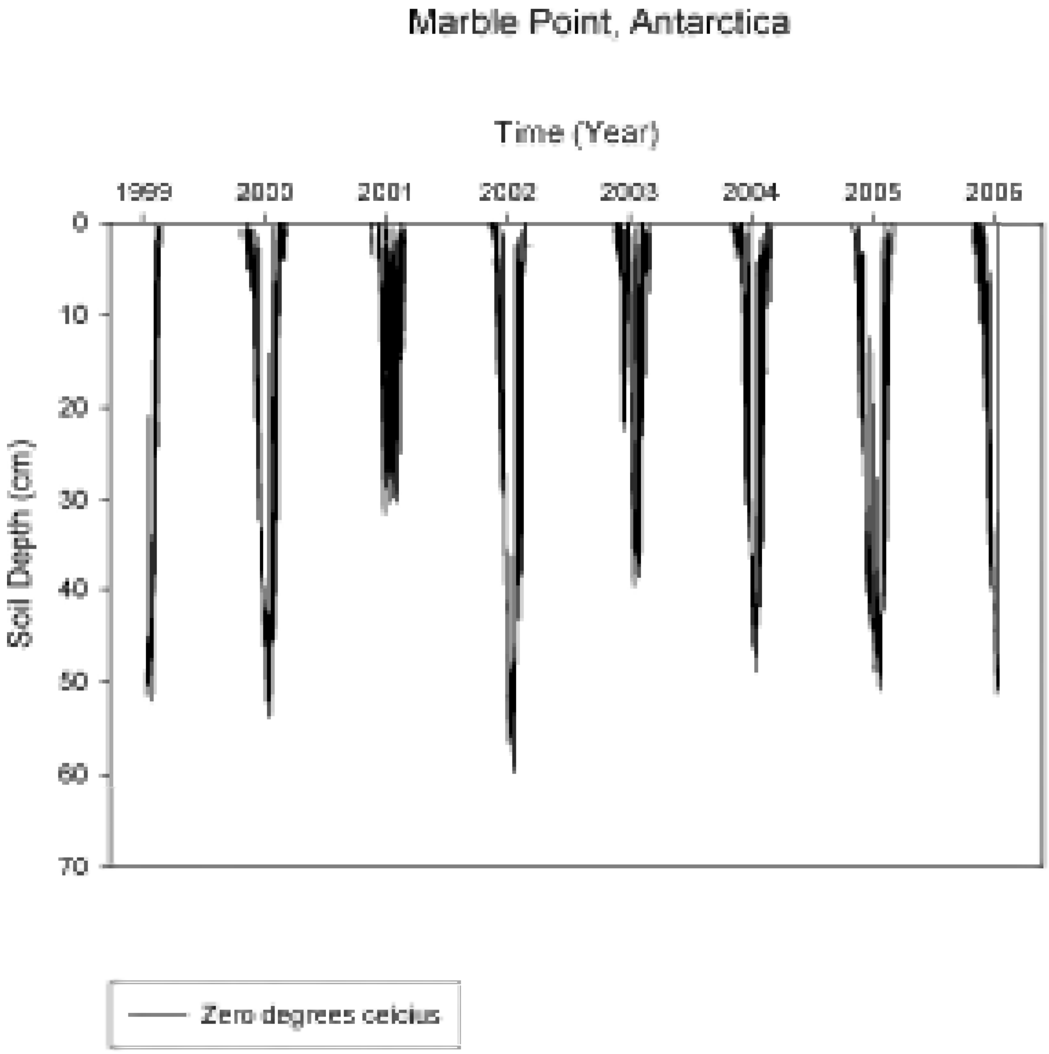


Figure 2.27 Variation in active layer thickness at Marble Point (77.4° S, 163.8° E), McMurdo Dry Valleys, Antarctica during the period 1999 to 2006 (Seybold, unpublished).

2.1.6 Sea Level

2.1.6.1 Monitoring sea level

As Antarctica could play a potentially large role in 21st century sea level change, it is disappointing that we have such poor knowledge of 20th century and present-day rates of change of sea level around the continent itself. Of course, this situation is due primarily to the great difficulty of acquiring extended time series of sea level measurements in environmentally hostile areas to the same standard as is possible elsewhere. Most sea level measurements during the 19th and 20th centuries were made with float and stilling well gauges, technology which presented operational problems in Antarctica. In some locations, there were also major issues to do with datum control (e.g. the establishment of adequate local benchmark networks and maintenance of tide gauge calibration with respect to those marks in local surveys conducted during brief annual visits).

While many short tide gauge measurements have been made around Antarctica, primarily for the determination of tidal parameters (e.g. IHB, 2002), there are few records that satisfy the quality criteria required by the Permanent Service for Mean Sea Level (Woodworth and Player, 2003) and that are long enough to be of interest for long-term change studies. The outstanding record is that from Vernadsky station (formerly Faraday station) in the Argentine Islands on the western side of the Antarctic Peninsula (Figure 2.28). It is now operated by the National Antarctic Scientific Center of Ukraine and contains a conventional float and stilling well gauge maintained in collaboration with the Proudman Oceanographic Laboratory. Hourly sea levels are measured by means of a paper chart recorder, with datum control provided by daily comparisons of tide gauge and tide pole observations. The sea level record from this venerable gauge commenced in 1958, the equipment having been installed at the then British Antarctic Survey Faraday base during the International Geophysical Year, thereby providing the longest sea level time series in Antarctica. The gauge received a major upgrade in the early 1990s when a pressure sensor gauge was added, and a new pressure sensor gauge with satellite transmission capability was installed in 2007.

The PSMSL data catalogue (www.pol.ac.uk/psmsl) provides one list of sea level data available from Antarctica. Notable records can be found from the Japanese Syowa base from the mid-1970s and from the three Australian bases of Mawson, Davis and Casey from the early 1990s. Other long term records are known to exist that are as yet not included in international data banks. A particularly interesting one, as it is far from the other long records mentioned above, is from the New Zealand Scott base and has been acquired since 2001 with the use of a bubbler pressure gauge attached to the reverse osmosis water pipe for the base, around which there is a permanent gap in the sea ice. France has made major efforts to instrument Dumont D'Urville. This station has been operated since 1997 with gaps and various upgrades, and is currently being updated to real time transmission as part of the Indian Ocean Tsunami Warning System. Details of this and other gauges operated by various nations in Antarctica are often included in the national reports of the Global Sea Level Observing System (GLOSS: IOC, 1997; Woodworth et al., 2003) (see www.gloss-sealevel.org). In addition, a list of Antarctic stations with tide gauges is maintained by the Scientific Committee on Antarctic Research (SCAR) (www.geoscience.scar.org/geodesy/perm_ob/tide/tide.htm). However, an important point to make about Antarctic data is that very little of it is downloadable and as readily analysable as data from elsewhere. In particular, much data have been obtained with pressure sensors, which are subject to drifts and biases. Any analyst must consider carefully the possible data problems and the impacts on the application to which they are put.

2.1.6.2 Sea Level Data for Ocean Circulation Studies

One application of sea level data is in ocean circulation studies, for which analysts usually require sub-surface pressure (SSP) rather than sea level itself. SSP can be obtained at a tide gauge site either with the use of a shallow-water pressure sensor, or by adding local air pressures to the data from a gauge (e.g. float or acoustic) that records true sea level. An alternative to coastal equipment in such studies is provided by bottom pressure recorders (BPRs), which have been employed in the Drake Passage and at other Antarctic locations at various times since the International Southern Ocean Studies (ISOS) programme (Whitworth and Peterson, 1985), and more recently since the World Ocean Circulation Experiment of the 1990s (Spencer and Vassie, 1997; Woodworth et al., 2002). Many of these deployments have been by UK groups and most records are readily available for analysis via www.pol.ac.uk/ntslf/acclaimdata/bprs.

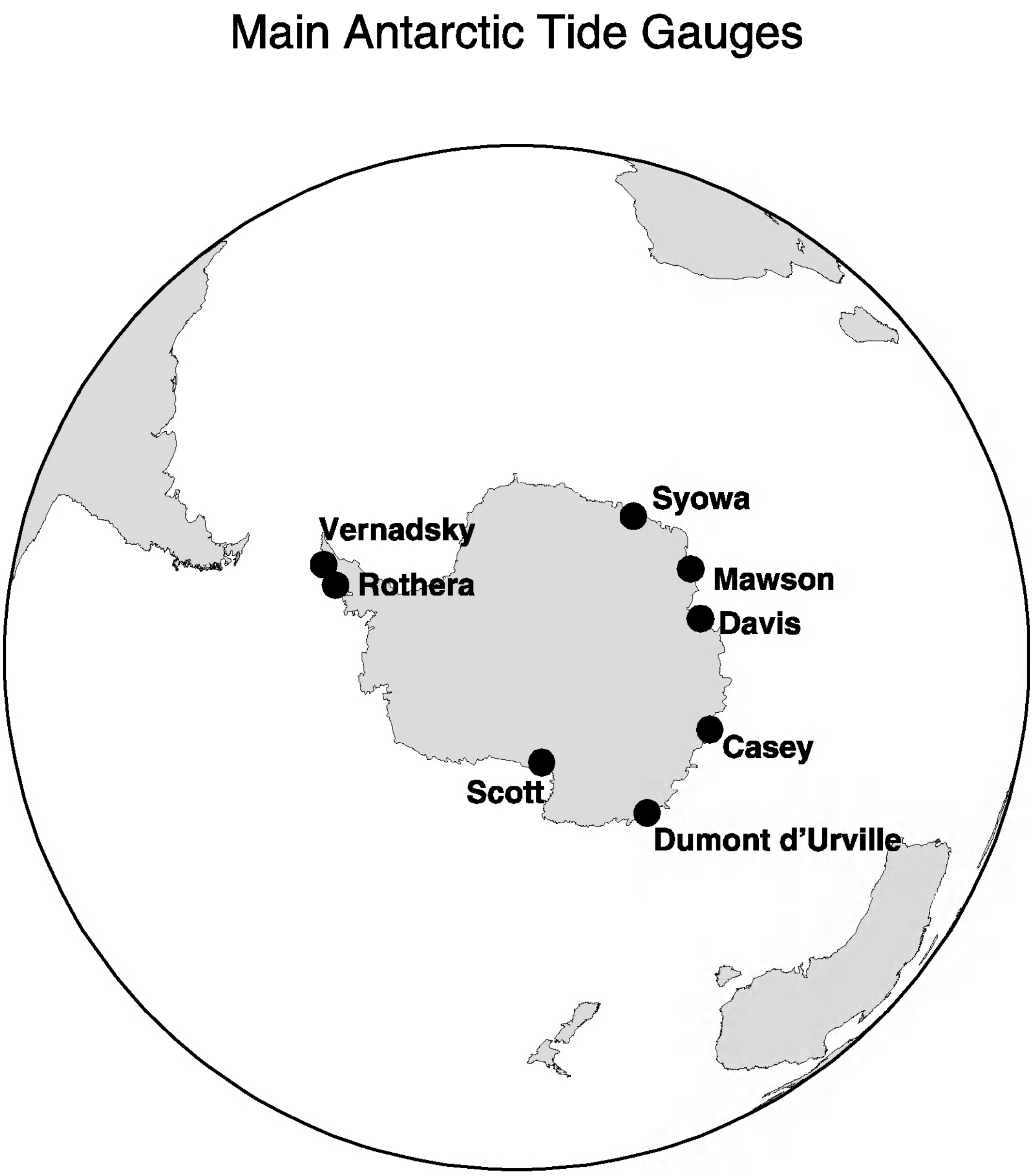


Figure 2.28 Main Antarctic tide gauges described in the text.

Antarctic sea level data have great importance in understanding the variability in the Antarctic Circumpolar Current (ACC), and thereby the role of the ACC in the global climate and, ultimately, in sea level change itself. A series of papers (Woodworth et al., 1996; Hughes et al., 1999; Aoki, 2002; Hughes et al., 2003; Meredith et al., 2004) have demonstrated that SSP fluctuates similarly around the entire Antarctic continent and that the SSP fluctuations can be related to changes in the circumpolar ocean transport around Antarctica. SSP data can be obtained either from measurements by BPRs deployed to the south of the main ACC axis or from coastal gauges as described above. The relationship between SSP, ACC transport and the SAM applies at least on intra-seasonal timescales (i.e. periods of more than a month and less than a year but excluding the quasi-regular seasonal cycle). It also applies on inter-annual timescales, despite the presence of baroclinic variability in the ocean at these longer periods (Meredith et al., 2004). However, the relationship at longer (decadal) timescales remains to be tested. The importance of sea level data in monitoring the circumpolar transport around Antarctica became more apparent with the realisation that many of the other techniques commonly employed are subject to critical aliasing, resulting in unrealistically high estimates of variability (Meredith and Hughes, 2005).

Major efforts have been made recently to provide sea level data from Antarctica in real-time, resulting in more rapid determination of ACC transport than has been possible to date (Woodworth et al., 2006). This development is also part of a general effort by GLOSS to have as many gauges as possible in the global network delivering data in real-time data, thereby enabling faults to be identified and corrected faster than would otherwise be the case. Rothera real-time data became available in 2007, while data from Vernadsky and King Edward Point, South Georgia in the South Atlantic will follow. The latter will largely replace an older installation at Signy, South Orkney Islands. All such UK data will be obtainable via www.pol.ac.uk/ntslf/acclaimdata. Data from Syowa are available in real-time from www1.kaiho.mlit.go.jp/KANKYO/KAIYO/jare/tide/tide_index.html, while data from Australian stations are available in 'fast' rather than 'real time' mode (i.e. with a short delay of typically 1-2 months).

2.1.7 Marine Biology

2.1.7.1 From historical surveys to high-tech instrumentation

Historical expeditions in Antarctic offshore waters dating back to the second half of the 19th century can provide data relevant to studies of the effects of climate on the ecosystem, and can provide a baseline for long-term observations, as has been demonstrated in studies of krill and salp populations (Atkinson et al., 2004). Early taxonomic surveys and corresponding species descriptions provide much of the basis for advanced studies on ecosystem functioning, especially where historical taxonomic and biogeographic data are included. Such data are generally accessible through databases managed by by SCAR-MarBIN, SOMBASE (Figure 2.29), FISHBASE, OBIS or, for a more multidisciplinary approach, by PANGAEA.

Modern taxonomic and biogeographic surveys include programmes such as CAML (Census of Antarctic Marine Life) and ICEFISH (International Collaborative Expedition to collect and study Fish Indigenous to Sub-Antarctic Habitats), both linked to the SCAR/IPY biology program EBA (Evolution and Biodiversity in the Antarctic – The Response of Life to Change). Long-term observations are recognised by an increasing number of scientists as an essential tool not only to achieve sustainable yields of natural resources such as krill and fish and to study the impact of over-exploitation e.g. by CCAMLR surveys (Convention on the Conservation of Antarctic Marine Living Resources), but also to monitor the impact of climate change at any level of ecosystem organisation.

Satellite sensors, combined with 'ground truth' data from *in situ* surveys, are contributing to a better understanding of pelagic systems by providing large scale and long-term data on biological bulk parameters such as chlorophyll, and on ecologically relevant physical parameters, such as ice cover, sea surface temperature, and iceberg production. ADCP's and plankton-video-recorders provide further information on the pelagic habitat to supplement results from traditional net hauls and CTD stations. Moored sediment traps and current meters provide data on the pelagic environment and on biological coupling between the surface and the seabed. High resolution side scanning sonars, benthic landers, cameras and ROVs provide further information. These methodologies provide the basis for classifying the benthic system and for carrying out controlled sampling, for instance of sediments and organisms at a cold seep in the former Larsen B ice shelf area, for observing small scale patterns of recolonisation after ice berg scouring, and for carrying out fine-scale experiments on the effects of increased supplies of planktonic phytodetritus as a source of food for the sea-floor.

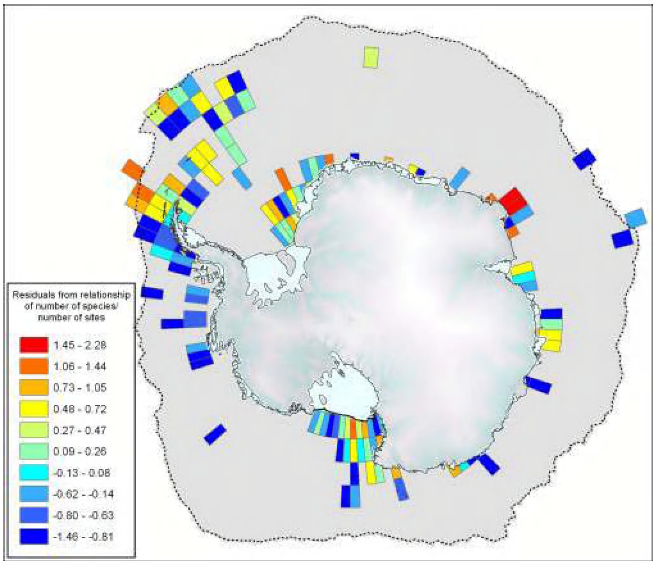


Figure 2.29 Large-scale circum-Antarctic depiction and analysis of semi-standardized biodiversity-datasets is possible for the first time using databases that contain information from various sources, historical and modern. In this example, from the SOMBASE database (Griffiths et al., 2003), species richness of bivalves were plotted against sampling effort (number of sampling sites) and are shown as the residual values from this regression. Red/orange indicate species richness above the mean, green/blue values below the mean; grey areas: no representative data. This example shows the high spatial heterogeneity obviously shaped by ecological conditions and being of high relevance for predictions of climate change impact and natural conservation strategies.

Continuous recording of underwater acoustics, e.g. from the year-round operating PALAOA observatory beneath the Ekström Ice Shelf, provides information on the activities of whales and seals that can be used to study aspects of the ecology of the top predators and of nature conservation (Boebel et al., 2008).

In the future, the routine use of AUVs and gliders will provide a picture at even higher spatial resolution of ocean behaviour and ecology, in addition to permitting collection of data from formerly inaccessible areas under the sea ice or beneath ice shelves. Instrumented AUVs could also contribute to comprehensive mapping of sea-floor habitats as the basis for characterising seabed topography and sediments for classifying benthic communities surveying krill and other planktonic concentrations and, finally, establishing ecosystem function. Fine scale experiments and observations can be continued through the use of instrumented ROVs and the invention of independently operating crawlers triggered by events. Automatically operating underwater observatories can be used to measure *in situ* a variety of ecologically relevant parameters for detecting biological response to climate change and for large scale ecological mapping.

Intensive inshore biological research supported by land stations began in the 1960s. Even at that early stage, sampling was accompanied by simple but effective *in situ* experiments carried out by scuba divers. Later, complicated laboratory experiments were made possible by new opportunities to maintain living animals in aquaria with running sea-water, not only on bases on the Antarctic continent, but also on research vessels carrying fish and invertebrates to home institutes. Such physiological studies contribute especially to a better understanding of the presence and future of marine biota if they are combined with field obervations and field surveys. Modern station-based and scuba-diving-based ecosystem field research focuses on determining the ecological effects of retreating and melting glaciers and ice scouring at the organism, species and community levels, for instance at Rothera, McMurdo, and Jubany Stations.

2.1.7.2 Molecular biology, genetics, genomics

The role of molecular analyses is exponentially growing. A new powerful aid to evolutionary biology and ecology has been recently provided by *genomic technologies*, whose fast development has brought biological sciences to the threshold of a revolution. The availability of sequences greatly enhances our understanding of the structure, function, content, variation and evolution of individual genes. Most importantly, it allows sequencing of whole genomes. Sequencing genomes of model organisms is a great challenge for biological sciences. In the last decade, scientists have developed methods to align and compare sequenced genomes, but the analysis of a given sequence provides much information on the genome structure, yet only to a lesser extent on the function. Comparative genomics is a useful tool for functional and evolutionary annotation of genomes. In principle, comparison of genomic sequences may allow for identification of evolutionary selection (*negative or positive*) the functional sequences have been subjected to over time. Positively selected genome regions are the most important ones for evolution, because most changes are adaptive and often induce biological differences in organisms.

Genome mining is clearly high on the agenda and it also provides unparalleled tools for studying natural selection in action, and investigating the link between organisms and environment (*environmental genomics*). DNA sequence data on Antarctic bacteria are increasingly available. Genomics will provide invaluable information on the molecular basis of adaptation to the extreme marine environment at the poles, and on the response to climate changes. DNA microarrays will lead to high throughput analysis of gene expression (functional genomics) in marine polar organisms. Structural genomics, i.e. sequencing genomic libraries (bacterial artificial chromosome (BAC) libraries) and whole-genome sequencing, would complement the information from transcriptomics, to understand how the extreme polar environment has shaped genome evolution. Genome sequencing will also provide the means for a population genomic (rather than genetic) approach to the study of gene flow, population structure and dynamics of marine polar organisms. Large sequencing projects generally exceed the capacity of single research groups and even national programmes, and therefore cooperative international actions are highly desirable. An additional goal will be to make marine biologists, ecologists, physiologists, and other polar scientists familiar with genomics. The overall objective is to review and coordinate current knowledge of gene/protein evolution, adaptations and systematics (*proteomics*) in polar marine organisms.

Genomics is high-profile science, impacting on all areas of biology, so it is not surprising that it is playing an increasing role in polar studies (Clark et al., 2004). Future research in genomics will not only depend on the new technologies, but also on the integration with ecology, physiology and biochemistry.

Nowadays molecular biology is essential for studying evolution, and important insights come from nucleic acid and protein sequences, as well as from large-scale chromosome change (e.g. Stam et al., 1997, 1998; di Prisco, 1998, 2000; Bargelloni et al., 2000a; Pisano et al., 1998, 2003; Held, 2000; Pawlowski et al., 2002; Papetti et al., 2007; Wilson et al., 2007a; Mahon et al., 2008; Smith et al., 2008; Strugnell et al., 2008). Relevant examples of molecular phylogeny, e.g. hemoglobin, antifreeze glycoprotein, myoglobin (Cheng, 1998; Bargelloni et al., 2000b; Near, 2004; Verde et al., 2006a,b; Giordano et al., 2007), have expanded our knowledge on adaptive evolution, mostly related to the fish realm.

Cold adaptation is an important part of a refined physiological equilibrium, which remains unmodified in the absence of disturbances due to climate change. When such disturbances do occur, evaluating the response of cold-adapted organisms will yield indications which can be extrapolated to organisms of lower latitudes.

Knowledge on the capacity of the polar marine fauna to respond to on-going climate changes is beginning to emerge, e.g. information is available on fish cold adaptation. Over the last twenty years, important advances have been made in understanding the molecular mechanisms involved in evolutionary adaptation to temperature change. Further progress is likely to occur as scientists exploit new molecular approaches to environmental changes in the polar regions. Increasing knowledge of protein structure and function is progressively helping to solve the puzzle of cold adaptation. Although in some proteins a coherent picture of the molecular changes involved in evolutionary adaptation to temperature is emerging, much remains unknown as yet. Relatively few cold-adapted enzymes have been examined, and many important cellular processes have hardly been studied at all. Nowadays, the possibility of sequencing whole genomes may provide the amount of data needed to allow rejection or acceptance of some of the classical hypotheses currently invoked to explain protein thermal adaptation.

The use of molecular techniques in diversity, phylogenetic and population research has opened new ways to analyse patterns in biogeography and evolution. DNA markers, for example mitochondrial genes (e.g. encoding cytochrome oxidase subunit I) 12S and 16S rDNA, or nuclear genes 18S and 28S rDNA, rhodopsin or pax-6, can be used to identify species and to analyse macroevolution (phylogenetics – studying genome mutations over time – and phylogeography) and microevolution (population genetics – studying relative frequency of genome variation and how they change over time). In the past decade several molecular studies on Antarctic marine and terrestrial fauna have identified the existence of cryptic (hidden) species (e.g. Held and Wägele, 2005; Linse et al., 2007; Raupach and Wägele, 2006; Brökeland and Raupach, 2008). Cryptic species are morphologically very similar but vary in their nucleic acid sequences. Since its introduction, DNA barcoding has been used to identify species (e.g. Webb et al., 2006a; Rock et al., 2008) and to reveal new species (e.g. Lörz et al., 2007; Smith et al., 2008), mostly using the COI gene.

A robust knowledge of species distributions is important as it gives insights into Antarctica's faunal and floral past across time scales from recent ecological scale disturbance (e.g ice scour in marine habitats) through glacial cycles to the fragmentation of Gondwanaland (Convey et al., 2008, in review). Recent phylogeographic studies have provided an insight as to the origins and maintenance of current Antarctic diversity by examining how biota have responded to past and current climate and other environmental fluctuations (e.g. Raupach et al., 2004; Stevens and Hogg, 2003; Stevens et al., 2006; Allegrucci et al., 2006; Van Vuuren et al., 2007; Wilson et al., 2007a). Calibrating phylogenies with fossil records reduces the error in molecular clock estimates and allows a tighter comparison between inferences from molecular data and the historical climate and geological records (e.g. Page and Linse, 2002; Strugnell et al., 2008).

2.1.8 Terrestrial biology

For the purposes of this volume, the Antarctic terrestrial and freshwater biome includes the main continental landmass (the 'continental Antarctic' to biologists), the Antarctic Peninsula and associated islands and archipelagos (South Shetland, South Orkney, South Sandwich Islands, Bouvetøya) (the 'maritime Antarctic'), and the sub-Antarctic islands which lie on or about the Antarctic Polar Frontal Zone (PFZ) (Figure 2.30). These geographic regions are also meaningful biogeographical regions (see Smith, 1984; Chown and Convey, 2006, 2007; Huiskes et al., 2006; Convey, 2007b)

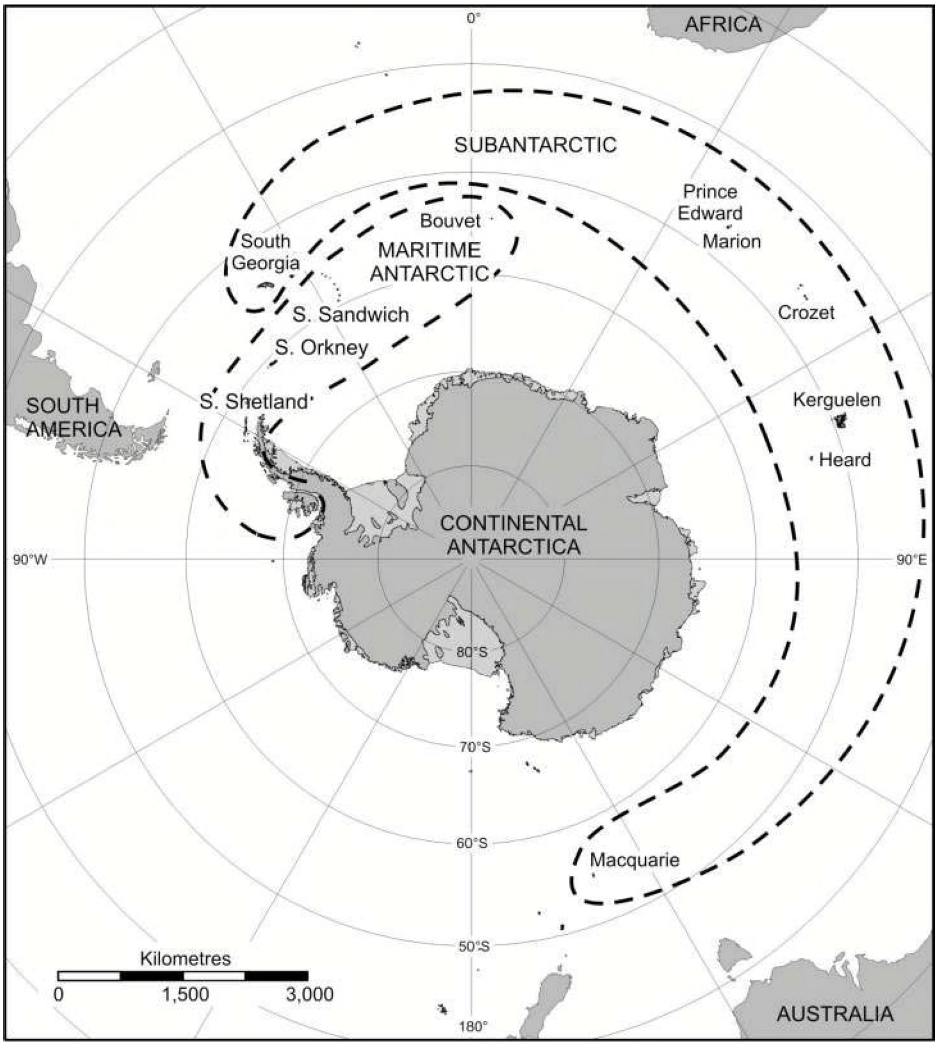


Figure 2.30 Commonly used terrestrial biogeographic regions within the Antarctic.

Over both short and long timescales, there are three major potential colonisation mechanisms likely to have played a role in shaping contemporary Antarctic biodiversity and biogeography, these being simple transport in the air column, incidental transport on other biota and debris, and transport on/in the ocean. Although both oceanic and atmospheric circulation patterns have acted to isolate Antarctica from lower latitudes, at least since the initiation of the Antarctic Circumpolar Current, this barrier is certainly not a hermetic seal, and terrestrial environments of Antarctica and the sub-Antarctic have experienced a fairly constant if low level rate of invasion from temperate or closer regions over evolutionary time, as well acting as a source and exporting biota northwards (Barnes et al., 2006).

Terrestrial and freshwater biological knowledge is unevenly distributed both across these regions and across the different biological groups present (Adams et al., 2006; Chown and Convey, 2007; Peat et al., 2007). Historically, biological research effort has focused on areas easily accessible from research stations, with an understandable but unfortunate tendency to select those areas with obvious biological development. In practice, this means that the majority of biological research (relating to both biodiversity survey and the study of biological adaptation and function) has taken place at a limited number of locations around the continent, and focused on a limited number of organisms. Prime amongst these locations are Signy Island (South Orkney Islands), various locations on the South Shetland Islands, the western coast of the Antarctic Peninsula (Anvers Island, the Argentine Islands, Marguerite Bay), locations along the Victoria Land coastline, and the Victoria Land Dry Valleys. Less accessible and more 'barren' areas have historically not received priority for logistic support or science funding, meaning that very little information is available from most inland regions, and many coastal areas and islands remote from research stations. Furthermore, in terms of important but subtle aspects of biodiversity research, in particular the functional role of organisms within an ecosystem, and the provision of ecosystem services, autecological studies of most species in most groups are non-existent (Convey, 1996a; Hogg et al., 2006). This means that functional interpretations of the biology of Antarctic terrestrial biota are often based on (untested) generalisations from the literature on a small number of species that have been targeted, and that on related species and genera from lower latitude ecosystems.

While biodiversity records are obviously available from a much wider range of locations across the continent than the focus areas mentioned, these are often the result of single field campaigns (sometimes directly involving an appropriate specialist, more often opportunistic collections subsequently passed to specialists). It is rare, even where specialists are engaged in field studies, for organised and replicated surveys to be completed. This is owing to multiple reasons - the practical logistics of supporting remote fieldwork, the very patchy distribution and small physical scale of habitats, the typically aggregated distributions of many of the biota involved, and the potential environmental impact and damage caused by sampling sensitive and fragile habitats. Many of the records that do exist are of limited taxonomic usefulness, while even where identifications to species level are available, they often represent the work of a single taxonomist and, especially for the smaller groups of soil invertebrates (e.g. nematodes, tardigrades), have not been re-assessed for veracity since the original collections, in some cases made by the early exploring expeditions (Adams et al., 2006). In many cases, the type material originally described no longer exists or is too degraded to be useful. The contemporary shortage of specialist taxonomic expertise is a problem recognised globally (references from Sands et al., 2008), but is particularly acute with reference to the faunal, floral and microbial groups that constitute Antarctic terrestrial and freshwater ecosystems.

A new tool recently available to Antarctic biologists is that of molecular taxonomy (e.g. Sands et al., 2008). This relies on the use of DNA or RNA sequence substitutions that build up over evolutionary time in order to calculate what are in effect likelihood trees (phylogenetic trees) expressing the evolutionary relationships between different organisms. There are many assumptions inherent in this approach, in particular relating to the rate of substitution over time, how to integrate molecular and classical taxonomic studies, and how to independently 'ground truth' the dating of divergence events, but its utility is now generally accepted. Even without attempting to interpret relationships, the use of selected DNA sequences as a molecular 'barcode' identifying specific taxa is also becoming widely accepted, and can be seen as an alternative means of assessing biodiversity in the absence of either appropriate taxonomic expertise, or of distinguishing morphological characters.

The limitations of contemporary survey data available as a baseline against which to compare and monitor future trends are amply illustrated by the bryophyte flora, one the best-known and researched groups of Antarctic biota. Peat et al. (2007), based on a comprehensive dataset of confirmed herbarium and literature records, provide a visual illustration and quantification of the level of diversity knowledge of bryophytes across the continent by the simple and coarse means of dividing the continental area in one degree latitude/longitude boxes, identifying all boxes that include at least one ice-free area of > 100 m^2, and then identifying how many of these have at least one herbarium or verified literature record of a plant's occurrence. On this basis, almost exactly 50% of boxes identified have no plant records (although it is then not possible to separate those that have simply not been visited from those that have had any form of visit or survey). Database compilations of diversity for other major groups of Antarctic terrestrial biota are less spatially representative even than that of the bryophytes, but are now becoming available (references in Pugh and Convey, 2008; Convey et al., 2008), starting to generate a baseline against which future changes can be compared.

2.1.9 Models

2.1.9.1 Introduction

The operation of the atmosphere or the ocean or the ice sheet, and the interactions between them, whether in the Antarctic or elsewhere, are of such enormous scale and complexity that it is impossible for a given observer to perceive these operations and interactions synoptically, or to envisage how they may change in the future, not to mention the fact that many of the operations are in themselves non-linear and may be chaotic and so are not amenable to simplistic prediction. Atmospheric and ocean physicists and glaciologists have therefore long known that it is necessary to simplify these processes to the point where they can be represented in a numerical model that can be run on a powerful computer. The complexities involved are such that simulating the behaviour of a coupled atmosphere-ocean-ice system requires the use of the most powerful supercomputers available. As we shall see, below, these models have their limitations. Nevertheless, every attempt is made to keep the models 'honest' by ensuring that to the extent possible, they can simulate today's environment (in other words the acid test of a model to be used for indicating what may happen next is to see how well it can reproduce what is happening now). In some fields, conceptual models are valuable for understanding certain aspects of the environment and are still used when analysing some types of data from the Antarctic. But numerical models run on computers are the mainstay of many modern atmospheric, oceanic, sea ice and increasingly, biological studies. In this section we examine some of the major classes of models used in Antarctic studies.

2.1.9.2 Coupled Atmosphere-Ocean Models

Predictions of future climate changes can only be made by using coupled global climate models (GCMs). These are also primary tools with which to simulate the past and present-day climates. Simulation of present day climates by these models provides a way of rigorously testing their ability to reproduce reality. Such reality checks are often crude, given that most of the commonly used GCMs do not operate at a finer scale than one-degree squares.

There are two main classes of coupled climate models: models of reduced complexity (see Claussen et al., 2002) and three-dimensional coupled atmospheric and oceanic general circulation models. In the following text, coupled atmosphere-ocean models means three-dimensional coupled atmospheric and oceanic general circulation models. These models are continually evolving so as to more closely reflect reality, through improvements to resolution on the one hand, and to increases in the number of potentially interacting properties of the climate system used in each model on the other hand.

In the early 1990s, there were only a few coupled global climate models; by 2006, as noted in the fourth assessment report of the Intergovernmental Panel on Climate Change (IPCC, 2007), around 24 such models were available to join the model inter-comparison project. Here we give a brief description of a coupled atmosphere-ocean model, and discuss their strengths and weaknesses, before presenting some results relevant to the Antarctic from the models used in the IPCC AR4.

A coupled atmosphere-ocean model or climate model has generally four major components: the atmosphere, the ocean, the land surface and the ice (cryosphere). More components are now being added to coupled models, such as biogeochemistry, in order to simulate the carbon cycle, and atmospheric chemistry.

The atmospheric component of such a model consists of a dynamical core that uses the equations of momentum and thermodynamics expressed in a set of partial differential equations, along with many physical processes, to observe change through time away from a set of initial conditions. The horizontal components of these partial differential equations are often solved using spectral methods on a supercomputer or network of PCs. Typically the horizontal resolution is 2^{o} to 3^{o}. In the vertical direction, variables are held on surfaces of constant pressure or height, or on surfaces that follow the orography. Typically, there are 20 to on order of 100 levels in the vertical direction. The water cycle and cloud-radiation interactions play very important roles in the climate system (e.g. clouds can both reflect incoming solar radiation, and absorb outgoing infrared radiation from the Earth's surface), but they are two of the largest areas of uncertainty in the current generation of climate models. Processes operating at the sub-grid scale, such as the formation of cumulus clouds, have to be parameterized using the quantities that the model can resolve. Since the processes of cloud formation are not well understood, there are many uncertainties in these parameterizations – nevertheless their adequacy can be tested by their ability to simulate known modern conditions.

The oceanic component has many similarities to the atmospheric element, apart from the fact that the ocean is seen as a basin with side boundaries. To compensate for the existence of these side boundaries, the equations are solved using finite difference methods in the horizontal direction, with a typical resolution of 1-2^{o}. In the vertical direction, most models use height as a coordinate, although some models use quasi-isopycnic coordinates (equal density) (for example, in the Miami Isopycnal Coordinate Ocean Model). Usually 20-40 levels are used in the current generation of models. Because sea water is much denser than air, ocean currents are much slower than winds, and a much higher spatial resolution is needed to resolve the processes happening in the ocean on the same time scale as the processes operating in the atmosphere. The details of the physical processes that need to be represented in the atmospheric and oceanic components determines the computing resources per grid square for each domain. It is still a matter of debate to what extent variations of small scale turbulence, sea ice properties, brine rejection etc have to be represented in large scale ocean models.The typical spatial scale of mesoscale eddies in the ocean is around 10 km, which is too fine to be resolved by the ocean components of GCMs. Eddy processes in the ocean, like clouds in the atmosphere, therefore have to be parameterised. However, the differences between results with models of different resolution suggests that the present parametrisations are far from being perfect. Eddy processes are important, because they are actively involved in the buoyancy and momentum transport across the ACC and in the boundary layers.

The land-surface component of GCMs is important in the global energy and water balance. Land surface temperature and soil moisture content are two basic variables in the energy and water exchanges between the atmosphere and the land surface. Some key properties including the roughness and albedo (reflectivity) are usually prescribed from the observed datasets. Some prescribed key properties can now be interactively determined in coupled models. For example, a dynamic global vegetation model can produce active vegetation types that can be used to determine the likely surface roughness and albedo. Land surface processes can also strongly affect the oceanic heat and freshwater fluxes.

Sea ice and snow over land are two important elements that are explicitly modeled in coupled GCMs. Glaciers and ice shelves are usually not resolved; instead, large-scale ice sheets (including ice shelves) are prescribed as land-surface topography with a high albedo. In addition to a high surface albedo, sea ice has very important insulation effects that inhibit the momentum, heat and freshwater exchanges between the ocean and the atmosphere. Sea ice formation or melting has strong effects on the freshwater flux. Early coupled climate models had very simple sea ice models, but recently more coupled GCMs employ sea ice models with more complex thermodynamics and sea ice dynamics.

Initialization

Initial conditions are needed for the ocean component because it is a slowly varying component. Observed present-day ocean temperature and salinity are usually used to initialize the ocean component. Another slowly varying component, the continental ice sheet is prescribed in terms of the present-day condition.

Flux adjustment

After the initial conditions are provided, the model is coupled together by allowing the exchanges of momentum, heat and freshwater between the atmosphere and ocean. In the early days of the development of coupled models, it was necessary to adjust the momentum, heat and freshwater fluxes to avoid large drifts of the model steady state away from the present-day climate. The use of flux adjustments was due to the existence of relatively large errors, mainly in the atmosphere and ocean components, and the fact that errors in one component could lead to further errors in the other components. Today, most coupled climate models have improved to the point where they no longer need to use flux adjustments.

Although significant progress has been made, large errors in the coupled models still exist, especially in southern high latitudes. For example, in the UK Hadley Centre climate model (HadCM3), the drifts of atmospheric and oceanic heat transport are largest in the southern high latitudes (see Figure 16 of Gordon et al., 2000). It is difficult to represent small-scale topographic features over the Antarctic and in the Southern Ocean, yet these play key roles in steering currents and ice streams. Shelf processes are believed to be very important in the Antarctic deep/bottom water formation and need to be represented correctly in models. Melting and freezing at the base of ice shelves are completely unrepresented in the current generation of coupled climate models; the role played by sea ice in the southern high latitude water cycle is untreated, because of the lack of observed sea ice volume and sea ice drift velocity. As a result the various models have large uncertainties, and the simulated fluxes of freshwater into the ocean are highly varied. Currently no model can realistically resolve the Antarctic ice sheet and ice shelves and their effects on the ocean circulation.

Ocean circulation simulation is still variable across the range of models. ACC transport is significantly biased in most IPCC AR4 models (Russell et al., 2006a). The simulated strength of the largest sub-polar gyre - the Weddell Gyre, has a very different value in different models (Figure 2.31). The changes of the Weddell Gyre strength during the 20th century in 18 AR4 models are also shown in this figure. The mean precipitation over the Antarctic and its change in 18 AR4 models is shown in Figure 2.32. Sea ice extents simulated by AR4 models are also very different (see Parkinson et al., 2006).

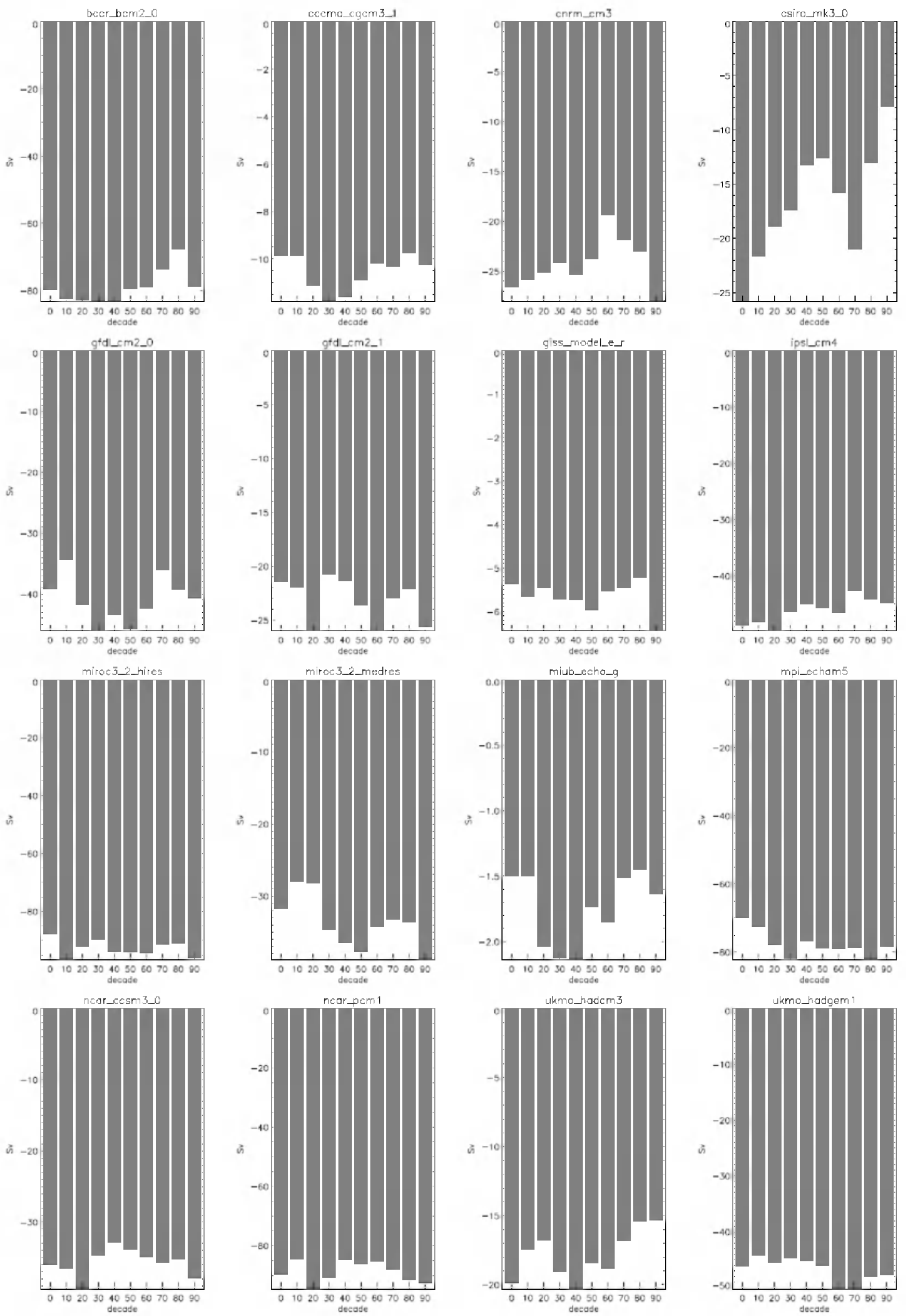


Figure 2.31 Simulated decadal means of Weddell Gyre strengths by 16 IPCC AR4 coupled climate models during the 20th century (IPCC, 2007). The Weddell Gyre strength is defined as the absolute transport of the southern limb across the Prime Meridian, whose observed value is 56 ± 8 Sv (Klatt et al., 2005).

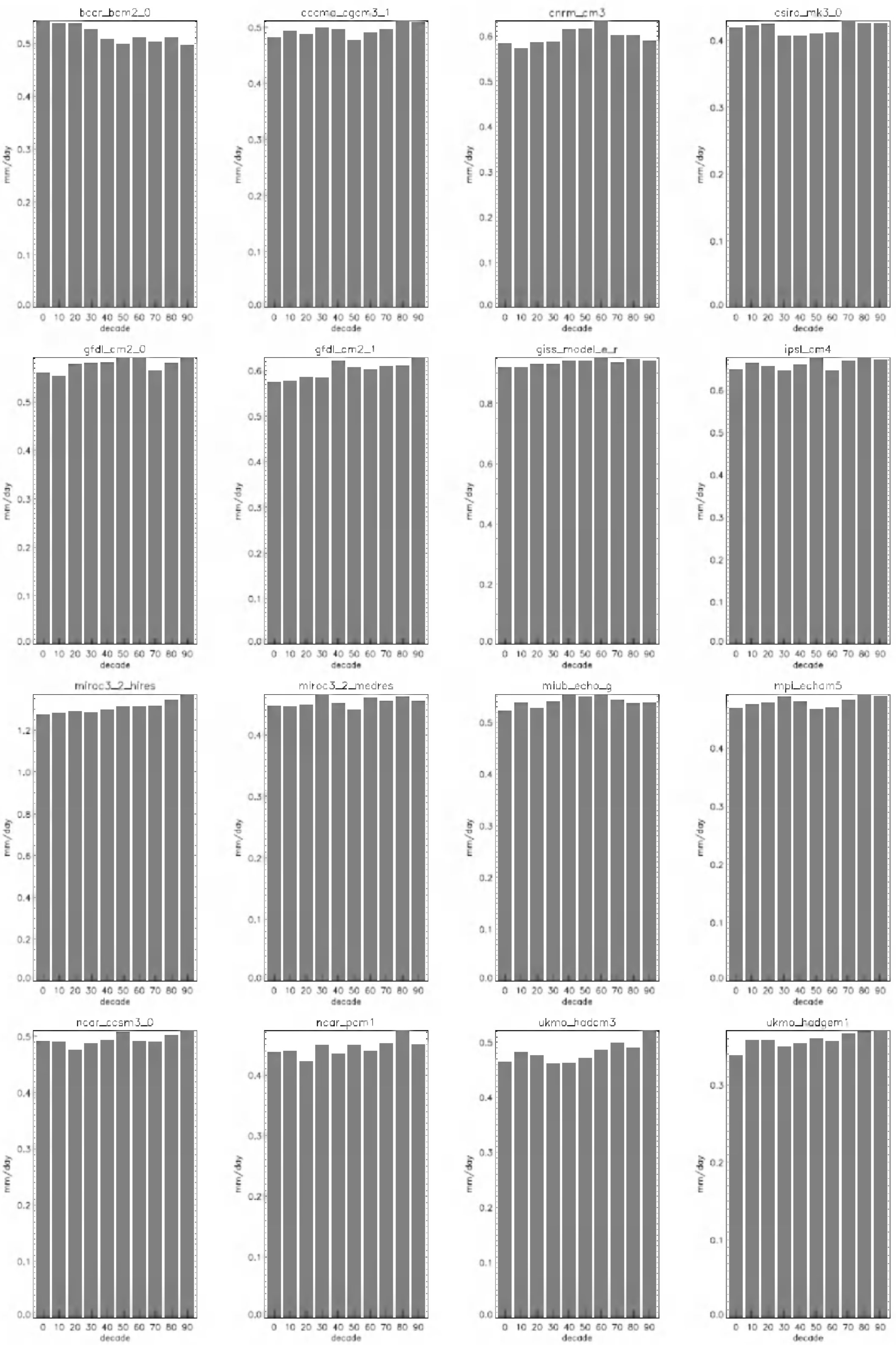


Figure 2.32 Simulated decadal means of precipitation averaged over the Antarctic continent by 16 IPCC AR4 coupled climate models during the 20th century.

A correct simulation of the Antarctic climate needs correct representation of the atmosphere, ocean and cryosphere components. The interactions between these components are very complicated and difficult to represent in models. Over the last 50 years changes in surface temperature across the Antarctic have been dominated by the winter season warming on the western side of the Antarctic Peninsula. The ensemble mean of the IPCC AR4 models when run through this period with natural and anthropogenic forcing correctly identified this area as having some of the largest warming in the Antarctic. However, the magnitude of the warming was only about a quarter of that observed in reality and there were very large differences between the models. As we discuss below, part of the problem is that the resolution of the GCMs is very coarse, preventing us from resolving features such as the Antarctic Peninsula, where some key climate changes are known to be occurring.

In addition there is still a dearth of appropriate observations against which to test models. The IPY 2007-2008 provided an opportunity for climate modellers to further validate and improve their models, and provided additional insights to improve understanding of the complex climate system in the southern high latitudes. Before we are able to point out the future directions of the model improvements, we have to have enough observational data to assess the model performance. In the meantime, more model intercomparison work needs to be carried out to find out the model ensemble means and the model differences.

2.1.9.3 Regional Climate models

GCMs have increased in resolution and complexity since the first models were developed in the middle of the 20th Century. However, until recently, little attention has been devoted to modeling of the polar regions. This is partly due to the large amount of computer time needed by the complex atmosphere and ocean models, but also due to a lack of observations and knowledge of the cryospheric components, including sea ice and ice shelves as well as snow, glaciers and permafrost. Compared to the effort devoted to development of parameterizations for mid-latitude processes, the cryosphere is under-represented. Nevertheless, there has been and continues to be improvement in the cryospheric components of GCMs. Even so, because GCMs need by definition to be global, they lack the grid resolution to sufficiently represent or parameterize processes that occur on a subgrid scale – like the scale of the Antarctic Peninsula. While the spatial resolution of GCMs will increase, their capabilities will continue to be limited by computational constraints such as processor speed and disk storage space.

Because of this, there is a niche for Regional Climate Models (RCMs), which can be either atmosphere-only models or coupled atmosphere-ocean models, and which can use the GCMs to supply boundary conditions.

Atmosphere-only regional climate models and stretched-grid global models

There are many aspects of the atmospheric physics and thermodynamics that are very specific to the Antarctic region, including for instance strong and persistent surface inversions, katabatic winds, 'clear-sky' precipitation, etc, which pose special challenges to the models. Some, such as clear sky precipitation, still remain a challenge to represent in models. But other aspects, such as the katabatic wind system, are being handled better now that high resolution regional models are being run.

Such models have been applied to regions of the Arctic and Antarctic with some success. For example the model Polar MM5 (mesoscale model 5), based on the Penn State model MM5 has been used to examine a number of problems in Antarctic meteorology, including the katabatic wind system (Bromwich et al., 2001).

The group at the Institute for Marine and Atmospheric Research, University of Utrecht, has also used a regional model to examine many aspects of the Antarctic climate. Their RACMO model is based on the German ECHAM4 model and has been used to look at the impact of the SAM on the atmospheric circulation of the Antarctic (van Lipzig et al., 2006).

Such models obviously have a lateral boundary, and boundary conditions are usually obtained from a global run of a coarse resolution model. In addition, it is also necessary to specify ocean forcing, such as sea ice extent/concentration and sea surface temperatures.

Stretched-grid general circulation models avoid the lateral boundary conditions issue in limited-area models. They are global but the grid may be horizontally stretched to refine resolution over a region of particular interest. For Antarctica, this approach was pioneered at the Glaciology Laboratory in Grenoble, France (Krinner et al., 1997). The finest (~60 km over Antarctica) climate change predictions to date have been obtained with this technique (Krinner et al., 2008; Genthon et al., 2008).

Coupled regional climate models

Implementing a limited area, coupled atmosphere-ocean climate model is much more difficult because of the need to obtain both atmospheric and oceanic data at the lateral boundary and to maintain stability in the atmosphere/ocean fluxes. However, the value of such a system is that important features of the Antarctic climate, such as the under ice shelf cavity, which is not present in the coarse resolution global models, can be included.

Such coupled models are starting to be developed, but there are many challenges in obtaining good coupling between the elements. They will become increasingly important in the future.

2.1.9.3 Ocean models of the Southern Ocean

Introduction

All numerical models involve a compromise between cost and physical realism. This is especially true for ocean models because the computational cost is always high, and can be hundreds of times that of a comparable atmospheric model.

This underlying conflict arises because the scale of ocean features is small compared with the size of the ocean and because the computer effort required is proportional to the cube of the horizontal resolution. However, the time step of integration may be longer for the ocean models because velocities are smaller (stability condition depends on speed). Thus, although ocean models need to resolve smaller scales, they save on time-step compared to atmosphere models.

Horizontal scales in the ocean are usually determined by the Rossby radius - a measure of how far the lowest internal wave mode can travel before being affected by the Earth's rotation. In the sub-tropics where ocean models were first developed, the Rossby radius is typically 25 km[1]. In polar regions, where the ocean is less stratified, this can drop to 8 km or less. The weak stratification also means that the influence of bottom topography is much stronger in polar regions, so when steep topography is involved a fine horizontal resolution is again required.

The lack of sufficient synoptic data for initialising and validating ocean models is also an issue because it means that any estimate of skill has to be based largely on qualitative judgements. Even so, because we have a good theoretical understanding of the problems involved, long model runs usually show up gross errors. We also have satellites, which provide good data on the surface fields, and the Argo programme is starting to provide a good data set for the subsurface and near surface layers. Detailed studies of density currents, the sea ice field and other key regions provide further local checks on model performance.

1 In the atmosphere the Rossby radius is nearer 250 km.

The major model choices

Most ocean models represent the spatial structure of the ocean by storing the model variables on a regular horizontal and vertical grid. A few (Iskandarani et al., 2003) split the ocean into finite elements within which the spatial variation is represented by linear or higher order functions. The latter approach is widely used by engineers modelling steady state structures, but it has not been widely adopted by oceanographers, possibly because of cost. If the finite elements are not on a regular grid there are also problems with spurious reflection and refraction.

Arakawa (1966) investigates the five standard ways that velocity and other model variables can be arranged on a regular horizontal grid. Of these he shows that two, the Arakawa B and C grids, are most accurate at representing the large-scale circulation of the ocean and atmosphere. The B-grid is slightly better at representing geostrophic flows within the ocean, i.e. flows in which the main balance is between the horizontal pressure gradient and the Coriolis force. For this reason the B-grid is used by many large-scale ocean models such as MOM (Griffies at al., 2004) and OCCAM (Coward and de Cuevas, 2005; Webb and de Cuevas, 2007).

The C-grid is slightly better at representing gravity waves and so is usually used for models near the coast where tides are important or where turbulent effects are large, so that flows are not geostrophic. It is also used for some large-scale models (Penduff et al., 2007), where the improved representation of gravity waves makes it easier to add sea ice. On the B-grid, gravity waves on the 'black' and 'white' sub-grids (to use a chess analogy) are only weakly coupled. In regions where internal waves are active this can produce an apparently noisy temperature field in the surface layers, which, if left uncorrected, affects the sea ice field.

In the vertical dimension, three standard schemes are used. Many, following the original Bryan and Cox model, use fixed horizontal layers (Bryan and Cox, 1968; Semtner, 1974; Cox, 1984). Together with the horizontal grid this generates a 3-D array of grid-boxes, each model variable nominally placed at the centre of each box defining the mean value of the variable within the box. Modern versions usually extend the vertical grid to include a free upper surface, allowing tides and other waves, and a variable thickness bottom box in each column, for better representation of topography.

The one major problem of the horizontal level scheme arises because most density surfaces within the ocean are gently sloping and there is little mixing between water masses of different density. The level scheme only allows fluxes through the horizontal and vertical faces of each box and this produces spurious numerical mixing between water masses of different density. To overcome this, isopycnal models have been developed (Bleck et al., 1992; Hallberg, 1997; Chassignet et al., 2007) in which the layers correspond to constant potential density surfaces within the ocean. The scheme works well in removing numerical mixing. It can also handle overflows well. There are problems in handling mixed layers (Gnanadesikan et al., 2007), in handling the non-linear effect of temperature on compressibility (so really a unique constant potential density surface does not exist) and in handling regions where weak stratification means that only a few model layers are present. However the gain from reduced numerical mixing is often more important.

A third scheme, used most often near coastlines, is to split the water column into a fixed number of layers (Haidvogel et al., 1991; Song and Haidvogel, 1994; Blumberg and Mellor, 1987; Mellor, 2003). This then gives extra vertical resolution in the shallow water. In deep water it suffers from the same numerical mixing problem as fixed layer models (Willebrand et al., 2001). The sloping surfaces also introduce small errors in calculating the horizontal pressure gradient (Shchepetkin and McWilliams, 2003). This does not matter in turbulent shallow water but in the deep ocean it can generate spurious currents.

Once the grid is chosen, grid variables are used to represent the standard momentum and tracer equations describing the ocean (Griffies, 2006). They are written so that momentum, heat and salinity are conserved, but as they miss sub-gridscale processes, they can only provide approximate solutions to the full set of differential equations.

Most of the errors discussed can be reduced by using higher order numerical schemes or by using finer horizontal or vertical resolution. There is even a suspicion that we are starting to see a convergence in the results obtained from the different fine resolution models, so that in the end computational efficiency may be the only criteria left at this level of model choice.

Sub-grid scale and process models

Within the large scale model framework there are usually a number of process models representing the surface mixed layer, the bottom boundary layer, sea ice, floating ice shelves, small scale mixing and other processes – especially those that act at a scale smaller than the model resolution. Usually such process models can be adapted for use in all of the different large-scale models and, because a large variety of such process models have been developed, there is usually a selection available for each run of a large scale model (Griffies et al., 2004). The skill of the final model may depend critically on the choices made at this stage.

Sea ice and the ice shelves

Sea ice in the Southern Ocean differs from that of the Arctic in two key ways. First, away from the coastline, the ice usually lasts for only a single year, so complex multi-year ice models are not essential. Secondly, the primarily westerly winds mean that the flow is generally divergent, so modelling of ice rheology is not so essential. As a result Southern Ocean models that use simple ice models (i.e. Semtner, 1976; Hibler, 1979), give good results away from the coastlines. Near the coast, especially in parts of the Weddell and Ross Seas, more complicated models are required. In such areas the Hunke and Dukowicz (1997) elastic–viscous–plastic scheme is often chosen.

In the past the ocean under floating ice shelves has often been ignored in ocean models. Recently the situation has changed, and successful models of the flows under such ice shelves have been developed. These are discussed later. As confidence in these models develops they are likely to be more widely adopted.

Overflows and bottom boundary layer

The cooling and ice formation processes that occur on the continental shelf around Antarctica result in an important source of very dense water which eventually sinks down the continental slope to form some of the densest waters of the deep ocean. This 'overflow' occurs in a very thin layer, typically 30 m thick, which has proved almost impossible to represent correctly unless the model itself has a finer vertical resolution (Legg et al., 2006).

Initially it was thought that analytic sub-grid scale models could be used (Baringer and Price, 1997), but with realistic flows and topography these were found to be unstable. The lack of a realistic sub-grid model has its greatest effect on horizontal or z-layer models but schemes such as those of Döscher and Beckmann (2000) can be used to reduce the error.

Isopycnal models can be more successful as long as one of the density layers corresponds to the thin descending plume (Willebrand et al., 2001). Density is not fixed but depends non-linearly on both pressure and water properties, so the assumed existence of constant potential density layers produces small errors. Also the large range of densities found in the ocean, and the necessity of modelling small density differences in regions of weak stratification, means that the number of density layers in the model needs to be large.

Sigma coordinates have the advantage that extra resolution can be provided near the bottom. The bottom layer also follows the descending plume, its thickness increasing with depth. In principal, constant thickness lower layers can also be added to the z-layer models, the normal preference for global ocean models.

For long-term climate integrations, the realistic representation of overflows remains a major problem that still needs to be solved. Its importance arises because it affects the replenishment of bottom waters and thus the large-scale vertical structure of the ocean. However for the study of short-term and near surface processes, the effect of any such error is usually small.

Upwelling, subduction and the mixed layer

Offshore in the Southern Ocean, key processes include the upwelling of dense water in the south, the transport northwards of this water in the surface Ekman layer, and the mixing and sinking of intermediate waters in the north. If the wind stress is correct, then momentum conservation ensures that the total transport in the surface Ekman layer is also correct. The velocity of the Ekman layer, which affects sea ice, depends on near surface mixed layer processes, and these vary from model to model. Thus for any research involving sea ice, a good mixed-layer model is essential.

Available mixed layer models include those of Pacanowski and Philander (1981), Mellor and Yamanda (1982), Price et al. (1986), Large et al. (1994) and Gaspar et al. (1990), and extend from simple bulk models to detailed turbulent closure models. Their effectiveness has not been seriously tested with the range of conditions (ice, stratification and surface forcing) found in the Southern Ocean, so at present all should be used with caution.

Both upwelling and subduction involve advection of water along sloping density layers. This is handled best by isopycnal models (Willebrand et al., 2001). However subduction also involves interaction with the mixed layer and capping at the end of winter and here the isopycnal layer models have problems (Large and Nurser, 1998).

Mixing – tides, topography, currents

The representation of mixing in the ocean is a huge subject, which can only be briefly discussed here. Near the surface the effect of wind and breaking surface waves is included in the mixed layer models. On continental shelves the extra effect of bottom turbulence due to the currents may fully mix the water column, and as tides produce their own currents, their influence also needs to be included.

Away from the boundaries, vertical mixing occurs primarily due to breaking internal waves. The energy for these waves may come from the wind acting via the surface mixed layer, from the propagation of internal tides and from the interaction of currents with bottom topography. However the processes are still only poorly understood. Most mixing models represent such effects by simple Laplacian diffusion, possibly with larger values near topography. Recent research indicates that in the Southern Ocean mixing is largest in areas of strong bottom currents so there is a case for increasing the values in these regions as well. However while numerical mixing remains a problem it is likely that, except in isopycnal models, the effective vertical mixing will be too large.

Horizontal mixing is also important in the ocean, especially in frontal regions where gradients are large. In low-resolution models the main sub-gridscale process that needs to included is baroclinic instability. The Gent and McWilliams scheme (Gent and McWilliams, 1990; Gent et al. 1995; Griffies, 1998) has been used to represent such processes, but there are concerns about how realistic it is, especially near the ocean surface or bottom topography. As a result, if frontal regions are important then it is best to use a model with a resolution of less than the Rossby radius.

Finally, because the Southern Ocean is only weakly stratified, bottom topography effectively steers the currents throughout the whole water column. It is therefore essential that topography is accurately represented. If smoothing is carried out, as it is usually done in sigma coordinate models to reduce errors in the pressure term, then the errors produced by the smoothing need to be addressed.

Available models of the Southern Ocean

For large-scale studies of the Southern Ocean it is probably best to start with the fine resolution global models for which model data is readily available. These are OCCAM and the Parallel Ocean Model POP (Maltrud and McClean, 2005; Collins et al., 2006) with a resolution of 0.1 degrees or less. POP is available in both the original and the NCAR community versions.

OCCAM and POP are related to the original Bryan-Cox-Semtner code. If you want to run or develop your own version of this code, the best supported version is MOM (Griffies et al., 2004, 2005) but both POP and OCCAM have made code available, for example, for developing biological models (Popova et al., 2006). The NEMO (Nucleus for European Modelling of the Ocean), ORCA025 and ORCA12 (Madec et al., 1998) models are primitive equation models adapted for regional and global ocean circulation problems. With a resolution of ¼° or 1/12° (Madec, 2008) NEMO is intended to be a flexible tool for studying the ocean and its interactions with the others components of the Earth's climate system (atmosphere, sea-ice, biogeochemical tracers etc) over a wide range of space and time scales.
Other global studies which have been carried out at lower resolution include the HYCOM (Chassignet et al., 2006, 2007) and POM (Mellor, 2003) models. HYCOM is an isopycnal model adapted to use level coordinates in the near surface layer. POM is a widely used sigma coordinate model. At low resolution we are also starting to see more operational models. These combine one of the regular models with data from satellites and other sources to provide a more accurate view of the ocean and its circulation (Chassignet and Verron, 2006).

Regional Models

A major problem that arises when developing regional models is how best to deal with the open boundary. Flow through the boundary usually dominates the large-scale circulation within the region under study, so any errors seriously affect the results. In such cases the best solution is to specify the boundary conditions using data taken from one of the global models. A less satisfactory solution is to specify the boundary conditions using climatology.

At the largest scale there have been three major models that cover just the Southern Ocean. The earliest, FRAM, was a rigid-lid level model without sea ice. It relaxed to climatology at the surface and at the open boundary (FRAM Group, 1991).
The later BRIOS model is a sigma co-ordinate model based on Haidvogel's SPEM code (Haidvogel et al., 1991), which uses a Hibler type ice model. The Southern Ocean versions use 24 layers in the vertical and have a horizontal resolution of 1.5 degrees or less, with finer resolution in areas such as the Weddell Sea (Beckmann et al., 1999). It is important because it is the first of the large-scale models to include the ocean under the ice-shelves.

Another important large-scale model is the isopycnal model which Hallberg and Gnanadesikan (2006) used to investigate the effect of horizontal resolution in the Southern Ocean. The model uses 20 density layers but has no sea ice. Of the three model types this is probably the best suited to studies of the transport of mid-depth water masses through the ocean.

The BRIOS Model

In the past, most large-scale models ignored the regions of ocean under the ice shelves. This was a serious omission but it arose because no suitable computer codes had been developed for handling the revised upper boundary condition.

Such codes have now been developed, one of the first to be widely used being part of the BRIOS model, discussed above. It was originally used to study flows in the Weddell Sea region (Beckmann et al., 1999; Timmermann et. al., 2002) but has also been used elsewhere around Antarctica (Assmann et al., 2003). The model extends the sigma coordinate scheme under the ice shelves with the 'ocean surface' coordinate following the bottom contour of the ice shelf. As a result, vertical resolution is good under the ice shelf. The fact that the model can be run in circumpolar mode also means that problems with the open boundary condition have little effect on the shelf circulation. As with other sigma coordinate models the main problems are due to numerical mixing and the necessity to smooth topography to reduce pressure gradient errors.

Isopycnal model

An alternative approach is that of Holland (Holland et al., 2003; Jenkins et al., 2004), who has modified the MICOM isopycnal model to include ice shelves. As with other pure isopycnal schemes, vertical resolution is obtained by making a judicious choice of model density levels for the area under study. The model typically uses ten density layers, but the weak stratification of some regions of the ice cavity means that only a few layers are involved, so the effective vertical resolution can be very coarse. However the advantage of the method is that it does not suffer from numerical mixing in the same way as the sigma coordinate model, and there is no pressure gradient error. The model is thus a useful independent check on the circulation.

The model of Dinniman et al. (2007)

A second sigma coordinate model for use under the ice shelves has been developed by Dinniman et al. (2007), based on the ROMS model (Shchepetlin and McWilliams, 2005). Both this and the BRIOS model use 24 layers in the vertical with a concentration of layers towards the top and bottom, so the main differences are at the process model level. Thus where the BRIOS model uses a Hibler ice model, Dinniman et al. preferred to impose an ice climatology based on satellite observations. They did this because during the period of study (2001-2003) the sea ice was affected by large ice islands and behaved in a complex way, which was unlikely to be reproduced by a standard ice model. Such parallel developments need be encouraged because of the insights they give into the strengths and weaknesses of different approaches.

Concluding comments

In future there are likely to be two areas where specialised models of the Southern Ocean need development. The first is in the study of the biology of the Southern Ocean, and especially the communities that develop under the immense areas of sea ice. The second is in the study of land ice and its response to climate change. Here the processes occurring under the ice shelves may have a significant impact.

For the biological studies, the main weaknesses of the physical models is likely to be in the representation of the surface mixed layer and the detailed properties of the surface ice field. It is easy to suggest possible improvements to the process models, but what are lacking are sets of good year-round data from the Southern Ocean that can be used to test them. Data from the Argo floats is helping to fill the gaps but there is still very little data from the large areas of open ocean covered by sea ice.

For the flows under the ice shelves data are becoming available from boreholes and by other means. Here a model intercomparison experiment, along the lines of the DYNAMO project (Willebrand et al., 2001), would be useful. This should compare the results of a sigma coordinate model with isopycnal and z-layer models of comparable vertical and horizontal resolution, and be designed initially to investigate the size and effect of the error terms. Once these are quantified then people would have a lot more confidence in using the models to predict future changes.

2.1.9.4 Under Ice Shelf Models

It is important to have realistic models of the ocean flow under ice shelves because of the crucial role that the shelves play in the climate system of the Antarctic, and their sensitivity to changes in water masses. At present global and regional climate models do not include sub-ice shelf cavities; these must be included in the future and the current generation of ocean/shelf models provides a step in this direction.

At the moment high resolution ocean models are run across limited areas of the Southern Ocean with atmospheric forcing being provided by the reanalysis data sets or global or regional atmospheric models. These allow the investigation of the changes of water masses under the ice shelves and the interaction with the broader scale ocean environment (e.g. Hellmer, 2004).

2.1.9.5 Ice sheet models

The first numerical modeling involving the entire Antarctic Ice Sheet was the diagnostic study of Budd et al. (1971) aimed at deriving physical characteristics of the ice sheet, in particular the temperature distribution (obtained from a moving-column model) and balance velocities. In the early 1980s, depth-averaged time-dependent models were first applied to simulate evolution of the Antarctic Ice Sheet in the studies of Budd and Smith (1982) and Oerlemans (1982a, b). Subsequently, models have been modified to include calculation of englacial ice temperature, inclusion of ice-shelf flow, and deformation of subglacial sediments. A recent overview of the status of Antarctic models is provided by Huybrechts (2004). For the present discussion, suffice to note that for land-based flow the so-called shallow ice approximation (SIA) is adopted with the local driving stress balanced by drag at the glacier base (Nye, 1957; Hutter, 1983). In that case, the dominant strain rate is vertical shear and an analytical expression for the depth profile of the horizontal velocity can be readily derived (Van der Veen, 1999a, section 5.1). Essentially, this incorporates the same simplified ice dynamics as did the pioneering work of Mahaffy (1976). Thermodynamics are included following Jenssen (1977) with conservation of energy considered at discrete depth layers extending from the ice surface to the bed or including several layers into the bed underneath. Seddik et al. (2008) proposed an application of a continuum-mechanical model for the flow of anisotropic polar ice to the EDML core, Antarctica.

Realistic model simulations of the behaviour of the Antarctica Ice Sheet over time require adequate surface and lower boundary conditions. More than 2000 reports of SMB determinations are available over Antarctica (e.g. Vaughan et al., 1999), and many of these are on the plateau. However, the Antarctic is very big and more measurements are still needed. While some airborne radar sounding has been conducted, resulting maps of basal topography (Drewry, 1983; Lythe et al., 2000) reflect at best the large-scale topography. Localized channels, if existing, are not captured in these compilations but may exert important controls on ice drainage, especially through outlet glaciers. Discussion here focuses on the ice-dynamical component of Antarctic ice-sheet models, and, specifically, on their ability to simulate contemporaneous rapid changes. A more expanded discussion of the shortcomings of existing ice sheet models is provided by Van der Veen and ISMASS (2007).

Over the past two decades or so, evidence for active ice sheets – both in the past and present-day – has mounted and the traditional view of ice masses responding sluggishly to external forcings has been replaced by the understanding that large ice sheets can undergo rapid change. In West Antarctica, ice streams draining into the Ross and Ronne-Filchner ice shelves have been identified and some of these have been the subject of extensive field campaigns aimed at better understanding the controls on ice streams (c.f. Alley and Bindschadler, 2001, for a collection of papers summarizing earlier findings). These ice streams appear to be capable of rapid changes including margin migration and complete shut-down. Based on mapping of grounding-line positions of Pine Island Glacier using satellite radar interferometry, Rignot (1998a) inferred a retreat of 1.2 ± 0.3 km/yr between 1992 and 1996, corresponding to a thinning rate of 3.5 ± 0.9 m/yr for this glacier. Satellite radar altimetry over the period 1992 to 1999 confirmed this inferred thinning rate and also showed thinning to extend far into the interior (Shepherd et al., 2001). Comparison of the satellite altimetry data with airborne laser altimeter surveys showed that thinning rates near the coast during 2002-2003 were significantly larger than those observed during the 1990s and the Amundsen Sea sector of the West Antarctic Ice Sheet appears to be out of balance by as much as 60% (Thomas et al., 2004a). In the Antarctic Peninsula, several of the peripheral ice shelves have disintegrated or retreated, with a total area in excess of 14,000 square kilometers lost over the past two decades. Vaughan et al. (2003) linked ice shelf collapse to southward migration of the -9ºC isotherm, presumed to correspond to the thermal limit of ice shelf viability. While Vaughan (1993) reported no significant acceleration of input glaciers following the breakup of the Wordie Ice Shelf, elsewhere in the Peninsula ice shelf break-up has led to flow acceleration of grounded glaciers (e.g. Rott et al., 2002; De Angelis and Skvarca, 2003; Rignot et al., 2004a). These observations suggest that the West Antarctic Ice Sheet may be on the verge of contributing to future sea level rise, and have reinvigorated the long-standing debate about the stability of this marine-based ice sheet and to what extent buttressing ice shelves control drainage from the interior.

The current generation of whole ice-sheet models cannot reproduce the observed rapid changes, which led James Hansen to conclude that "ice sheet models cannot be used with confidence for assessing expected sea level change until they demonstrate realistic forcing yielding realistic rates of ice sheet demise" (Hansen, 2005, p. 273). From an ice-flow perspective, the most important yet least understood processes to be included in Antarctic models are dynamics of ice streams and the transition to ice shelf spreading, grounding-line migration and stability, and the interaction between ice shelf break-up and discharge from grounded glaciers formerly draining into these shelves. New analytical treatments of the ice stream to ice shelf transition (Schoof, 2007) suggest the way forward, but at present, the best global models have been unable to include the recently observed activity of ice sheets (IPCC, 2007).

Ice streams, embedded in slow-moving ice, are the primary drainage conduits evacuating ice from the interior to peripheral ice shelves and from thereon to the oceans. In West Antarctica, these ice streams rest on a layer of weak and possibly deforming till offering little resistance to ice flow, implying that flow resistance is concentrated at the lateral shear margins and transferred to excess basal drag under the adjacent interstream ridges (Van der Veen et al., 2007). The exact nature of this transfer of stress is not well known but may ultimately determine inward or outward migration (Raymond, 1996; Raymond et al., 2001; Jacobson and Raymond, 1998). Moreover, Van der Veen et al. (2007) estimate that meltwater production under the shear margin adjacent to Whillans Ice Stream is comparable to that under the ice stream itself, suggesting that these margins could be an important source of water for maintaining basal lubrication under the ice stream. These processes are not included in existing continental-scale models whose spatial resolution is usually insufficient to capture an entire ice stream, let alone the much narrower shear margins. Instead, ice streams are usually simulated through enhanced basal sliding where the basal ice reaches the pressure-melting point. Interestingly, a model study on the dynamics of the Siple Coast ice streams based on this concept generated a cyclicity with stagnant and active ice streams, caused by competition between several preferred ice-flow pathways in the area (Payne, 1998). However, this result may be more fortuitous than reflecting the real physical processes, as noted by the author of that study.

Ice streams represent the transition from interior flow to ice shelf spreading. At the inland boundary of the West Antarctic ice streams, smaller tributaries form within well-defined troughs (Joughin et al., 1999) and over sedimentary basins (Anandakrishnan et al., 1998; Bell et al., 1998) that coalesce into major ice streams. As the streams flow outward and enter the ice shelf, the flow regime becomes more akin to ice shelf spreading. Conceptually, ice streams may be viewed as the transition from internal flow dominated by vertical shear, to ice shelf spreading controlled by longitudinal stress gradients and lateral drag. MacAyeal (1989) developed the so-called “shelfy-stream” model to incorporate this transition into a numerical model simulating large-scale flow over a viscous sediment. This approach, or one similar to it, has yet to be incorporated into whole ice sheet models.

Ever since the pioneering work of Mercer (1968, 1978) and Weertman (1974), glaciologists have speculated that removal of peripheral ice shelves and floating ice tongues will result in increased discharge of interior ice. Over the last two decades, ice shelves in the Antarctic Peninsula have disintegrated, probably in response to a local warming trend that caused the thermal limit of ice shelf viability to migrate progressively southward (Vaughan and Doake, 1996; Vaughan et al., 2003). Velocity measurements on grounded glaciers formerly draining into these ice shelves show that a speed-up followed collapse of the ice shelves (De Angelis and Skvarca, 2003). While many in the glaciological community have interpreted these observations as evidence for the instability models proposed by Mercer and Weertman, and many others since, a more careful analysis of the sequence of events is needed to establish unambiguously to what extent forcings at the calving front propagate upstream and influence discharge from the interior. For example, it could be that increased speeds on the grounded portions resulted from the same surface melt event(s) that led to the collapse of the floating part, rather than reflecting the glacier response to loss of ice shelf buttressing. If, indeed, discharge from the interior is affected by ice-marginal processes, an important question is whether ice shelf collapse necessarily leads to irreversible glacier retreat or whether interior flow will adjust to the perturbation by reaching a new equilibrium. Modeling experiments on the response of Pine Island Glacier to perturbations at the grounding line indicate that a new equilibrium is reached after ~150 years following an imposed instantaneous change on the ice plain (Payne et al., 2004).

At present, understanding of the effects of ice shelf weakening or break up on discharge from the interior is insufficient to incorporate into numerical models. To gain better understanding, targeted data collection and hypothesis testing is needed for identifying processes responsible for rapid glacier changes. Fortuitously, perhaps, the southward migration of ice shelf collapse and consequent flow adjustment of grounded ice in the Antarctic Peninsula offers the opportunity to study a range of glacier settings. The importance of observing a range of grounding-line behaviours is evidenced by the study of Vieli and Payne (2005), who compared various model formulations applied to the study of marine ice sheets. They found that predicted grounding-line migration is dominantly controlled by the way grounding line motion is treated in the numerical model (e.g. fixed grid versus a moving grid), and how the governing equations are discretisized. Physics incorporated into the numerical models, such as longitudinal momentum coupling between the ice shelf and the grounded ice sheet, appeared to be of secondary importance only. The implication of this model comparison is that there is an urgent need to develop better models whose predictions are not dictated by numerical specifics. As concluded by Vieli and Payne (2005), "further model development also requires a better observational history of grounding line migration (in terms of both the timing and spatial extent) and also indicates the need of a test data set for the modeling community."

Rapid ice-sheet changes originate in, and spread from, restricted regions of fast flow such as ice streams and outlet glaciers. Existing models are based on the shallow-ice approximation (SIA) and do not include longitudinal stresses and the buttressing effects of ice shelves that may restrain ice-stream flow in these key regions. The SIA is appropriate only for slow-moving inland ice where resistance to glacier motion is entirely concentrated at the glacier bed. On the other end of the modeling spectrum is the Morland-MacAyeal (MMF) formulation for ice shelf spreading in which basal drag is set to zero and ice flow is assumed to be depth-independent. It seems likely that flow of fast-moving ice streams and outlet glaciers falls somewhere in between these two model regimes. Thus, the best way to model the dynamically important ice streams is to solve the full stress equations without a priori simplifying assumptions. Doing so on a sufficiently fine grid to resolve the ice streams is too computationally intensive for the current generation of computers; a possible solution would be to follow the climate modelers' lead and develop variable-resolution models, through use of nested mesoscale models embedded in coarse-grid models, or variable-element size models with adaptive regridding if needed.

Traditional models of ice sheets employ a fixed horizontal resolution over the whole domain and so either fail to resolve these features adequately (if they employ a relatively coarse ~20 km resolution), or would require unfeasible computing resources (if they employ a more reasonable ~1 km resolution). A solution to this dilemma is the application of a nested grid in which the whole domain is modeled at a coarse resolution, while areas of supposed importance such as ice streams are modeled at a series of finer resolutions with the remainder of the slow-flowing interior omitted. A wide range of variable resolution techniques are available ranging from simple ones in which the areas to be modeled at finer resolution are predetermined and held fixed, to ones in which the solution algorithm itself determines which areas are to modeled at the finer resolution. Similarly, the way information passes between the various grids can vary in sophistication from the coarse grid providing boundary conditions to the finer grid, to full multi-grid techniques in which information flows both ways in an iterative fashion. This type of approach is well developed in ocean and atmospheric modeling, and a number of software libraries exist to facilitate the use of nested grids.

To place any confidence in model predictions, it is first necessary to demonstrate that the models can successfully reproduce past glacier variations. Consequently, it is imperative that data sets be developed against which the skill of the numerical models can be tested. In this respect, it is important to separate data used for model calibration (i.e. parameter adjustment) from those used to evaluate the model performance. A more extensive discussion of the more philosophical underpinnings of model evaluation can be found in Van der Veen (1999b), while Van der Veen and Payne (2003) discuss a more pragmatic approach.

An Example Model

Recent results from the University of Maine Ice Sheet Model (UMISM) are illustrative of current capabilities of the latest generation of ice sheet dynamic models. This example focuses on simulating the ice kinematics of the Amundsen Sea Embayment region (Fastook and Sargent, 2004). Recent airborne geophysical surveys of the region measured new details of the ice thickness and bed elevations (Holt et al., 2006; Vaughan et al., 2006). These data are now available as a 5 km gridded data set in a format convenient for modeling. Figure 2.33 shows the new dataset (right) contrasted with the older BEDMAP dataset (left) for comparison. Gross details are found to be similar, but details within the trunks of the Pine Island Glacier (PIG) and the Thwaites Glacier (TW) allow models to simulate observed flow more accurately (Lang et al., 2004; Rignot et al., 2004b).

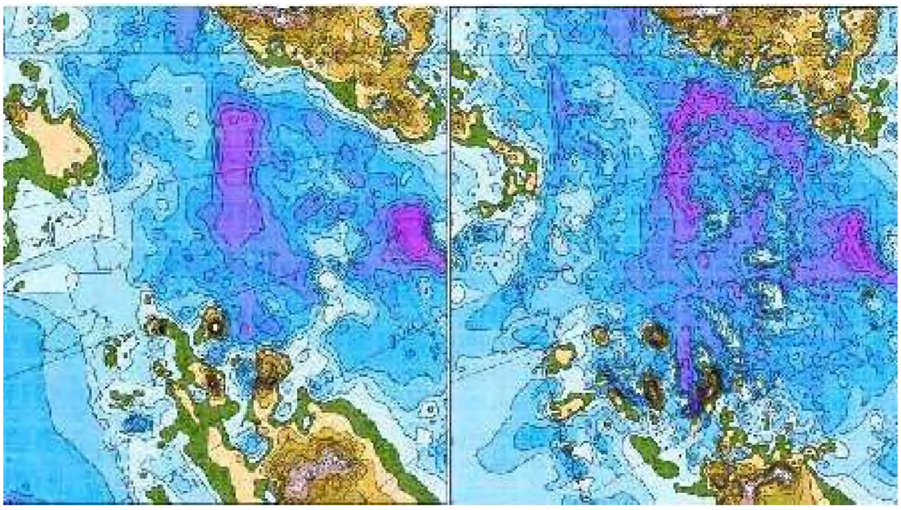


Figure 2.33 Bed elevations of the Amundsen Sea Embayment area: (left) from circa 2001 (BEDMAP) data compilation (Lythe et al., 2001) and (right) from newer circa 2005 airborne geophysical surveys (Holt et al., 2006; Vaughan et al., 2006). Blue is below sea level, with deeper blue signifying greater depths. The Amundsen Sea is in the lower left corner. Comparison shows that the recent surveys provided a more detailed representation of the subglacial topography.

The nest of UMISM embedded models begins with a 40 km grid of the entire ice sheet with 16,263 nodes. Embedded in this is a 10 km grid (9000 nodes) that includes the entire Amundsen Sea Embayment measurement area. Nested inside this medium-resolution grid are two 5 km grids encompassing Thwaites and Pine Island Glaciers (6402 and 10,920 nodes, respectively). This procedure produces high-resolution results with very reasonable runtimes. The model reaches the modern position by being run for 40,000 years, from an initial configuration to the Last Glacial Maximum and then from the LGM to the present. The 40,000 year cycle ensures that the final configuration includes time-dependent effects such as transient internal temperatures and isostatic adjustment. It is not sufficient to run the model in steady state for the present configuration and expect the model snapshot to correspond with reality. Model advance and retreat is controlled by a "thinning-at-the-grounding-line" parameter because the internal calculation of grounding line dynamics remains an extremely challenging issue and one that has only recently been solved analytically (as discussed earlier). The grounding line position is determined in this model through spatially non-uniform specification of this parameter, so retreat is guaranteed to halt at the correct position.

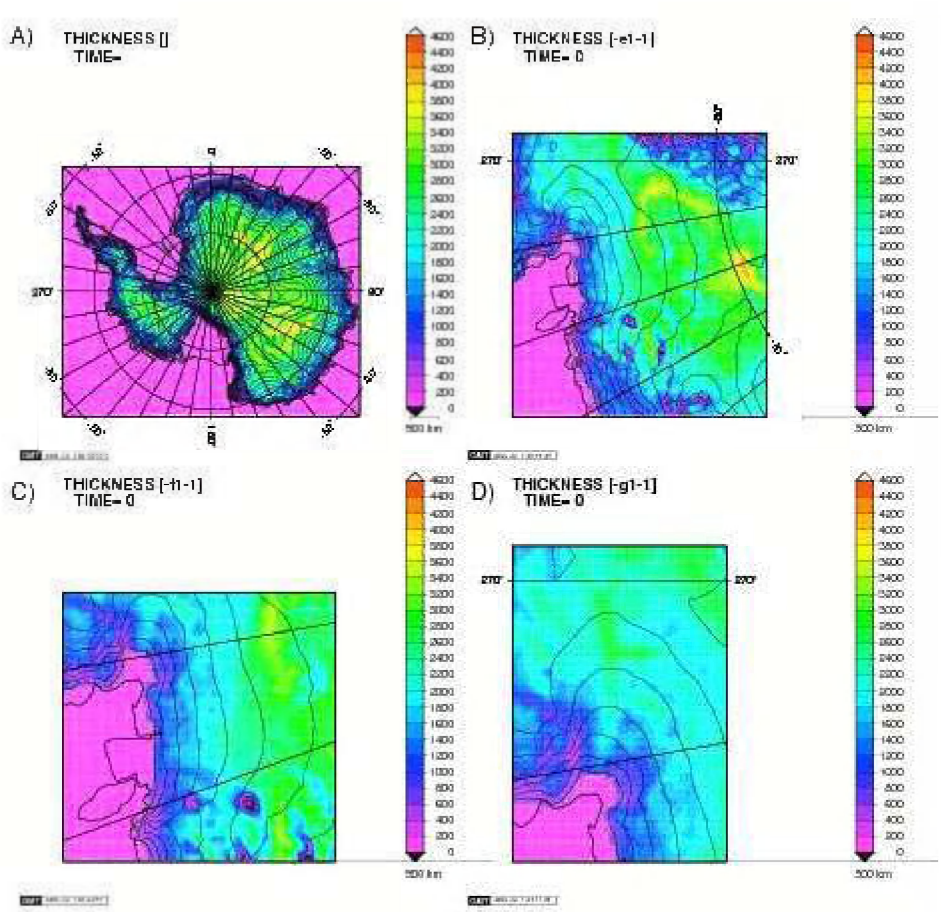


Figure 2.34 Nested grids, with ice thickness fields, for the UMISM. (A), the full ice sheet at 40 km resolution; (B), a medium resolution, 10 km grid encompassing the whole airborne geophysical dataset (shown in Figure 2.33); full 5 km resolution grids for (C) Pine Island Glacier and (D) Thwaites Glacier. Color elevation scale is shown at right.

Basal conditions are critical and in this case, water is produced, which leads to basal lubrication and higher ice flow rates, when the temperature reaches the pressure melting point. This happens first in the deeper channels and is one reason the dendritic pattern of modeled velocities agrees well with the observed field, however, such a mechanism does not guarantee the excellent agreement between modeled and observed velocity magnitudes (Figure 2.35).

The viscosity of ice is a strong function of temperature, so it is important to include the thermal evolution in predictive models of the ice flow. Retreat of grounding lines to colder, stiffer ice inland could act to stabilise the margin. Measurement of englacial (within-ice) temperatures requires boreholes to be drilled into the ice. This has been possible at a number of sites using hot water drilling (Engelhardt, 2005), but there is still a shortage of information for testing models that couple the ice flow and temperature. Because the presence or absence of water plays such a large role in the dynamics of the ice sheet, the temperature within the ice, and the flux of geothermal heat can have a critical role in determining the rates of ice discharge.

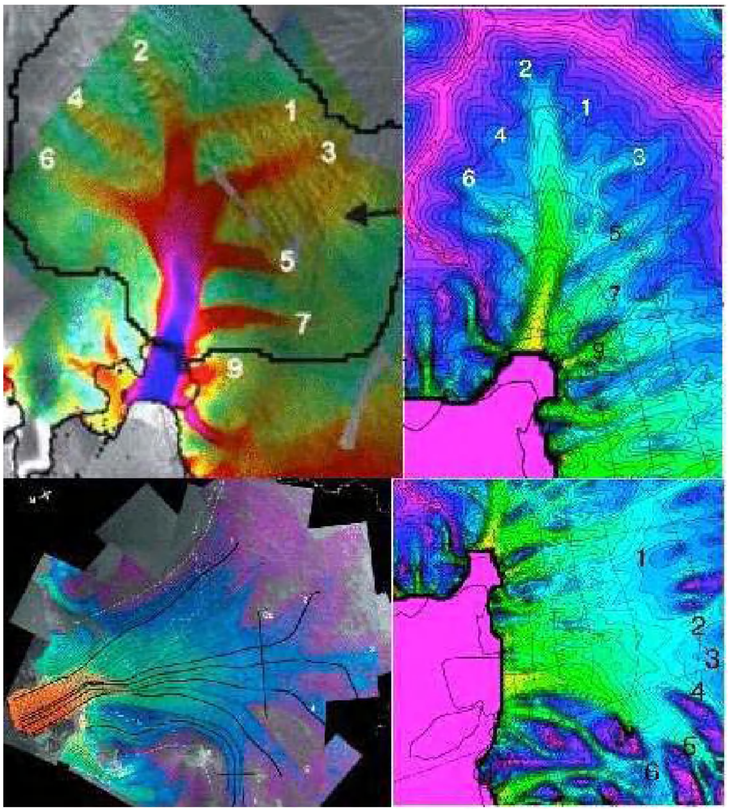


Figure 2.35 Comparison of model and observed surface velocities for Pine Island Glacier (upper right and left (Rignot et al., 2004b), respectively) and for Thwaites Glacier (lower right and left (Lang et al., 2004), respectively).

2.1.9.6 Biological Modelling and Bioregionalization

In the Antarctic environmental conditions, particularly on land but also in the sea can show high variability at both, spatial and temporal scales. This is especially relevant in the context of climate change, because the area of intensive atmospheric and sea water surface warming west of the Antarctic Peninsula coincides largely with the off-shore area that is central to large krill swarms, their population dynamics and consequently, higher trophic levels of the open ocean ecosystem. Further, inshore benthic communities range regionally from those with the world record biomass values within filter feeding assemblages (Gutt, 2000) to others with possibly the lowest community metabolic rates in the extremely depauperate species poor benthic systems that thrive below ice shelves (Azam et al., 1979). Temporal variations can be very important, e.g. in krill and penguin populations, benthic recruitment, and primary production, and are either supported or reduced by the occurrence of sea ice (Beaman and Harris, 2005; Gutt et al., 2007).

As a consequence of these constraints, advanced ecosystem research, independently of whether it refers to singe species or to communities, to ecosystem functioning or complex structures, demands spatially and temporarily explicit models. The approach of comparing various ecological layers (i.e. parameters) to understand the functioning of the ecosystem, including its spatial variability, is known as "bioregionalization" (Figure 2.36), and has the aim of identifying ecological "hot-" and "cold-spots". This concept can also be used to answer specific ecological questions - e.g. to understand the habitat of an Antarctic midwater fish and its trophic position in the pelagic ecosystem, with the objective of predicting fish recruitment (Loots et al., 2007).

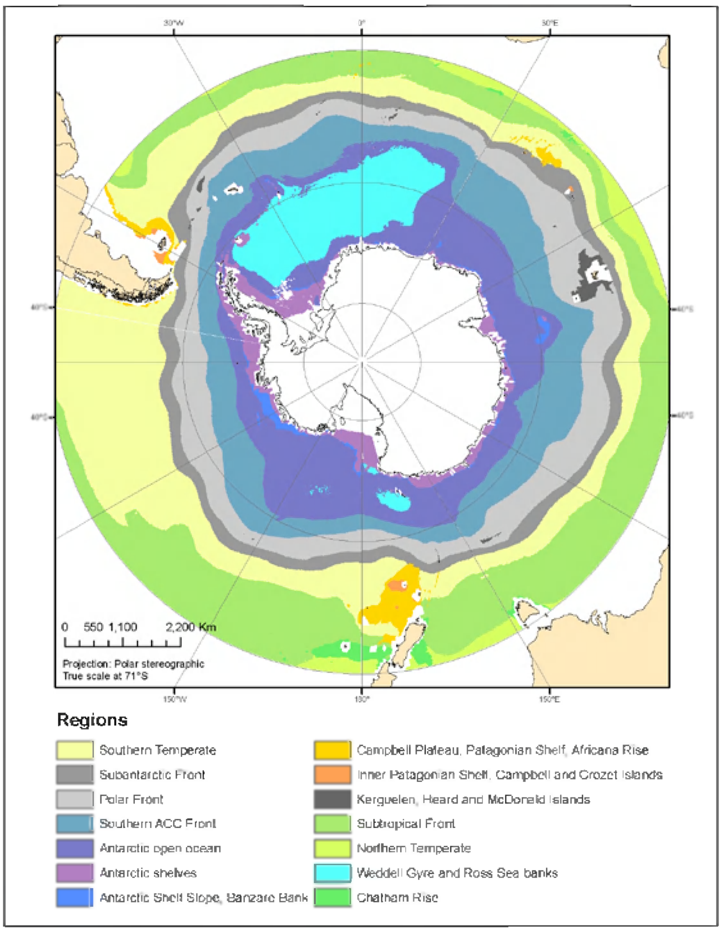


Figure 2.36 Primary regionalisation of the Southern Ocean based on depth, sea surface temperature, silicate, and nitrate concentrations (white areas represent cells with missing data that were not classified in these analyses). From Grant et al. 2006.

In cases where a complete coverage of important ecological parameters and organisms, even within a limited area is not possible (e.g. for bottom communities), a series of proxies such as water depth, bathymetry, water masses and currents as well as ecological key species can be used to estimate benthic structures and processes. Later, the potential for circumpolar application can be tested. For a general understanding of selected processes even individual based biodiversity-models can elucidate the relationship between, for example, climate-induced iceberg disturbance in combination with biological characteristics, such as dispersal, reproduction, mortality, and longevity of model organisms (Johst et al., 2006; Potthoff et al., 2006). Such models can cover periods of faunistic succession which exceed by orders of magnitudes the life span of ordinary scientific projects, and simulate expected climate change scenarios, e.g. the response of the benthos to the absence of iceberg scouring or to extremely high magnitudes of disturbance (Figure 2.37).

In addition spatially explicit biological models can contribute to solve methodological problems such as autocorrelation and can help to maximize the efficiency of sampling designs.

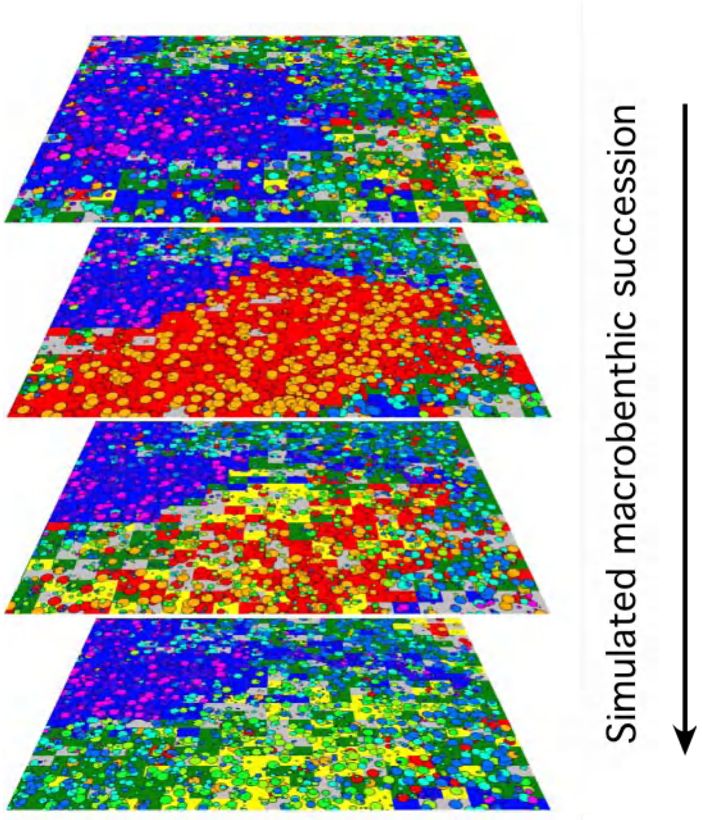


Figure 2.37 Climate change is potentially altering iceberg scouring frequency due to enhanced deglaciation in Antarctica. Thus, spatially explicit models can be a useful tool predicting biodiversity changes under these different environmental conditions. In addition, time spans clearly exceeding periods of research programmess can be easily simulated by different model runs. The results contribute to a better understanding of ecosystem functioning and allow predictions of future developments of the Antarctic biosphere. This individual-based simulation of a benthic succession with a total of 12 benthic model species (colours) demonstrates that iceberg scouring, indicated by the patch in the centre, increases macrobenthic biodiversity due to the coexistence and succession of model species (red/orange: pioneer species, green/yellow: intermediate colonizers, blue/purple: climax species). (model design: M. Potthoff).

With advanced studies of ecosystem research uni-dimensional parameters can either be relatively easily sampled using satellite data (e.g. in the case of ice cover and sea surface temperature), or demand high effort field studies (e.g. in the case of krill or fish stocks). Most multidimensional biological approaches will never be able to cover the entire area of the Southern Ocean and a complete range of organisms. To study parameters, which explain biodiversity, ecosystem functioning (such as reproduction, dispersal, feeding, or physiological tolerance), and climate sensitivity ecological key specied can be selected.

In the next step of advanced bioregionalization research, some of the approaches described above could be merged into one model with the option of working at different levels of resolution, spatially, temporarily, physically and biologically. Finally, such knowledge can provide the best possible basis to identify “Vulnerable Marine Ecosystems” (VME), options for Marine Protected Areas (MPA) and produce numerical predictions for the future of the Antarctic ecosystem. This has recently been identified by SCAR as an important challenge, resulting in the formation of an action group on “Prediction of Changes in the Physical and Biological Environment of the Antarctic”.

Based on such knowledge the UN Environmental Programme, World Conservation Monitoring Centre (UNERP-WCMC) developed in cooperation with the Conference on Biological Diversity (CBD), other international institutions and specialists criteria to identify "Ecologically and biologically significant areas" (EBASs) to propose these as "Marine Protected Areas" (MPA's). In parallel the Convention on the Conservation of Antarctic Marine Living Ressources (CCAMLR) has started to declare "Vulnerable Marine Ecosystems" (VME) to protect them from anthropogenical impact, mainly ground fishing.

2.1.9.7 Sea ice biological and physical modeling

Most of our knowledge on sea ice microalgal growth and production was derived from observational data and some experimental work. Recently, new experimental data have revealed the complex and heterogeneous character of the sea ice ecosystem (Thomas and Dieckmann, 2002), and the new information is improving the parameterization of biological sea ice models . The dependence of microalgal growth on light, nutrient and temperature is mechanistic, which means all processes involved are in accordance with purely physical and chemical relationships. However it has often been modeled using empirical relationships, where microalgal growth is parameterized as a function of the most limiting resource (Liebig's law). The difference between empirical and mechanistic models resides in the use of self-adapting variables by the latter to predict future states of microalgal biomass, allowing a dynamic adaptation of the biological processes to transient environmental changes. Such a dynamical approach depends on the physiological status of the cells. For Antarctic sea ice, where microalgal cells are subject to strong seasonal and daily variations in light, temperature and nutrients, only a mechanistic model can well represent the physiological responses of microalgal cells to environmental changes. The sea ice ecosystem differs considerably from water column conditions, and is characterized by extremely low temperatures, high brine salinities and strong downward attenuation of solar radiation. The sea ice seasonal cycle has two main implications for the spatial distribution of its microalgal communities: first, the ice growth season provides a large space surface area available for colonization in e.g. the brine channels of newly formed sea ice, followed by a drastic ice retreat during the transition to summer, dispersing the ice communities into the water column; and second, the role of ice drift during the growth season in the lateral transport of biomass through the seasonal ice zone, which significantly contributes to the horizontal distribution of food resources for other organisms associated with sea ice (e.g. the Antarctic krill and upward organisms further up in the trophic chain). The link between the establishment of microalgal communities and sea ice formation itself begins during direct interactions of ice crystals formed in seawater with individual organisms in the water column. This process is not selective and the mechanism by which microalgal cells, organic matter and other organisms are incorporated into developing sea ice is described in detail by Garrison et al. (1989) and Weissenberger and Grossmann (1998). Garrison (1991) demonstrated that although the large number of species that inhabit the underlying water column and sea ice brine channels (bacteria, microalgae, protists, small metazoans and some crustaceans), the extreme environmental conditions which the organisms are exposed to are very selective, and most of the sea ice biomass is dominated by small diatoms which are responsible for almost all sea ice primary production. Gleitz et al. (1998) showed that for more than 100 different diatoms species already found in sea ice habitats, less than 20 contribute significantly to the total biomass commonly found in the ice-pack while Lizotte (2001) inferred that Fragilartiopsis cylindrus and F. curta are the most dominant microalgae in sea ice. This low diversity may be related to the physiological capacity of these diatoms to maintain relatively high growth rates under extreme conditions of low light and temperature when compared to water column species. This apparent dominance of few species is also found between heterotrophic protists that inhabit sea ice.

Garrison and Buck (1989), working with pack-ice microbial communities in the Weddell Sea found that ciliates contributed 70% of the total protozoa biomass, followed by heterotrophic flagellates and other small protists, which play a key role in the cycling of material (e.g. excretion of nitrogen) and in controlling microalgal growth (Garrison, 1991).

The first attempt to model the Antarctic sea ice primary production was made by Arrigo et al. (1993), where the prognostic variables (C, N and Si) were modeled using static physiological parameterizations as a function of light, temperature and salinity, using a nutrient limitation scheme based on Monod growth kinetics and the Liebig's law. The thermterm static comes from the absence of variables in the model that describe the state of the cells, as well as the use of Redfield ratios to couple the microalgal growth function with the uptake of N and Si to carbon. The Carbon-equivalent (Ceq) for the nutrients N and Si used by Arrigo and co-workers were C:N:Si of 107:17:40 molar-ratio, which was consistent with some sea ice observations (see details in Arrigo et al., 1995, 1997). The shortcoming arising from such an approach is that microalgal growth limitation by light, temperature or salinity, affects only the rate of carbon incorporation in biomass, without any co-limitation and the system is basically controlled by the balance of nutrients and biomass in a fixed ratio (given by the molar ratios). When the environment is nutrient depleted, than growth rate relies only on the Monod growth kinetics. Thus, their model results show that the maximum biomass produced by microalgae depends only on the nutrient availability, and the rate of remineralization. The exact concentration is only related to the molar ratios between carbon, nitrogen and silicon. Although Redfield Ratios are commonly in use by many modelers (Arrigo et al., 1995; Fritsen et al., 1998), there are many cases where elemental composition does not match with these ratios and the use of Redfield is questionable.

There are many approaches to creating a biological model for sea ice microalgal growth and production and modern studies shown that a quasi-realistic model must be based on self-adapting mechanisms to relate the physiological status of sea ice microalgae to their primary production, including grazing by sea ice protozoa. Sea ice microalgae in models must grow over two essential co-limiting nutrients, dissolved inorganic nitrogen [N] and silicate [Si], and biomass is represented in terms of carbon content (P C), chlorophyll-a (PChl), particulate organic nitrogen (PN) and biogenic silica (PSi), allowing variable cellular N:C and Si:C quotas to decouple primary production and inorganic nutrient availability. Light- and temperature dependence of microalgal growth must be included in the model, treated by a set of ordinary differential equations, which describes the balance between light and the chlorophyll-a:carbon ratio, to simulate the photo acclimation mechanism. There is a lack of data about heterotrophic protists feeding rates on microalgae accumulating carbon and nitrogen biomass (ZC and ZN, respectively) and excreting the N-excess to the medium. However, remineralization of silicon is neglected in most experiments, as well as the uptake of other elements, like phosphate, iron, aluminum and vitamins.

Developing a sea ice coupled biological-physical model, to investigate the influence of transient changes in environmental conditions on sea ice biological communities, must include physiological self-adaptive schemes to simulate microalgal photoacclimation and uptake of dissolved nutrients in response to transient changes of light, temperature and nutrient supply. Grazing may be simulated by the incorporation of heterotrophic protists in the model with specific organic carbon and nitrogen pools, although there is almost no field data about these points. The excretion of nitrogen excess by the sea ice protozoa in some laboratory experiments proved to be of importance in sea ice nutrient dynamics.

Also, to simulate sea ice biology, the model must be coupled with a one-dimensional thermodynamic sea ice model including a numerical scheme to resolve the heat conduction in sea ice, as well as the incorporation and vertical redistribution of biological material due to brine flux. Belem (2002) showed that these physical processes play a key role in the vertical structuring of sea ice biological communities and their associated primary production. The inclusion of brine fluxes in the physical model shows that vertical differentiation in the salinity profiles, commonly observed in ice cores collected in the Antarctic pack ice, results from variable ice growth rates and consequently salt partitioning, which cause distinct bulk salinity profiles. These differences associated with the vertical position of the impermeable layer play a key role in the formation of banded layers where high accumulation of biological material is commonly observed.

However, there are a few key questions to be assessed in such coupled biological-physical model. One of these is a Lagrangian approach, where the time-dependent position of simulated floes extracted from ice velocity fields can be used to compute model forcing parameters (e.g. air temperature, oceanic heat flux, solar radiation). These processes are essential to determine the thermodynamic history of floes and to obtain a good agreement between model results and field observations. During the drift, synoptical changes in environmental conditions lead to a differentiation of vertical sea ice characteristics that most models do not take into account. With exception of the model cited, all other sea ice models have an Eulerian approach.

One of the most important environmental parameters governing the sea ice primary production is the incoming solar radiation. The model includes the simulation of the spectral solar radiation attenuated by the atmosphere and clouds and a bio-optical sea ice model, determining the vertical distribution of light within sea ice. The solar radiation is also of significant importance for the sea ice thermodynamics, governing the temperature profile within sea ice during the summer months. Simulations with the spectral bio-optical sea ice model in many works showed that part of the marginal pack ice during the winter receives enough light to support biological production. This fact is of extreme importance for overwintering organisms associated with the sea ice.

Thermodynamic processes controlling ice formation are fundamental in governing the vertical distribution of biological communities and the impermeable layer formed due to vertical temperature gradients and brine segregation acts as a barrier to the flux of biogenic material entrapped in the sea ice. Observed vertical variability in chlorophyll profiles from collected ice cores in the Weddell Sea are caused mainly by differential incorporation rates of biological material during the ice growth. Figure 2.28 represents the seasonal cycle of the Antarctic pack ice with emphasis on the physical processes controlling the incorporation and redistribution of biological material within sea ice.

During autumn, dynamic accumulation of microorganisms in the water column due to frazil ice scavenging results in high incorporation rates of biological material within newly accreted ice layers. During winter, the thermodynamic growth of the pack ice controls the vertical distribution of ice assemblages, mainly due to the effect of the brine flux. With the onset of the melting season in spring and summer and increasing solar radiation over sea ice, the biological activity reaches its maximum followed by a rapid decline in the primary production rates between February and March. The sea ice retreat constrains the primary productivity to persistent ice fields in the summer, characterized by nutrient depleted conditions and heavy snow cover, indicating a significant degree of limitation for microalgal production.

Belem (2002) demonstrated in a coupled biological-physical sea ice model that the total sea ice productivity in the Weddell Sea shows an annual carbon production of approximately 11 Tg C. During the winter, the sea ice productivity ranges from 0.16 to 0.6 Tg C per month, contributing to 17% of the total annual production. These results are lower than previous estimates for the Weddell Sea pack ice computed only for infiltration communities associated with flooded snow and snow-ice formation. However, the coupled biological-physical sea ice models described before do not consider the formation of snow-ice and, therefore, infiltration communities are not taken into account. Snow flooding is restricted to the later spring/early summer in some areas of the Weddell Sea and cannot be generalized to the whole seasonal ice zone. Alternatively, most of the ice cores collected in the Weddell Sea indicate that infiltration communities occur in only a small fraction of the pack ice and model results revealed a dominance of interior and bottom ice assemblages.

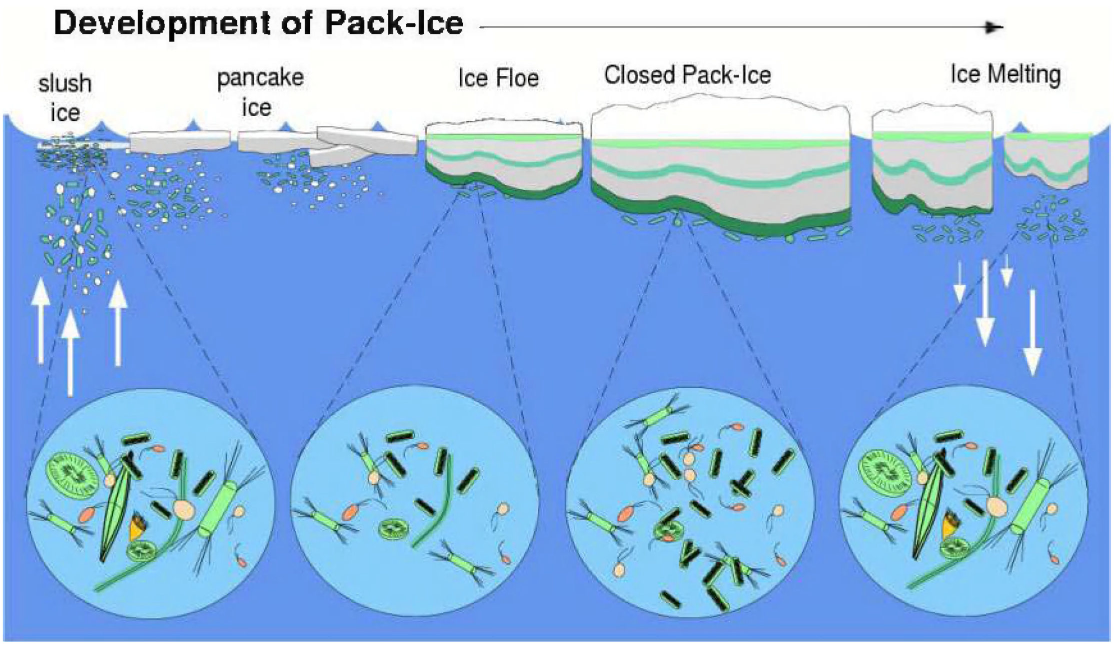


Figure 2.28 Schematic representation of the seasonal cycle of the Antarctic pack ice, with emphasis on important processes of incorporation of biological material and the physical evolution of sea ice until the melting season. Adapted from Belem (2002).

The ecological distinction between these two assemblages is relevant to understanding the time course of sea ice colonization. Garrison and Buck (1991) found that most of the sea ice infiltration assemblages are contained within a porous layer of hard ice near the freeboard level (i.e. the space between the ice surface and the seawater level) and apparently developed from internal assemblages. They also showed that interior assemblages are apparently established at the time of ice formation and thus overwinter survival in the ice is an important consideration. In contrast, an infiltration assemblage, which is inoculated when flooding occurs, would be heavily influenced by the composition of the plankton at the time of flooding. The results of coupled sea ice models suggest that the surface assemblages are related to the impermeable characteristic of topmost ice layers. However, the low temperatures associated with this vertical level must limit primary production.

Therefore, any estimates of the sea ice primary production obtained by models should be considered conservative, since the simulated pack ice consisted only of first-year ice. Many other ice habitats like multi-year ice, rafted ice floes, refrozen leads, infiltration layers, which are potential sites for the settlement of biological sea ice assemblages were not considered. The contribution of these communities to the total sea ice primary production needs to be further investigated. However, sea ice coupled physical and biological models are good tools to understand biological processes and its physical driven processes.

2.2 Future developments and research needs

Lack of data still remains a significant problem for researchers in many areas of Antarctic science. The Antarctic continent is large and there are logistical difficulties in getting to many areas. Autonomous systems have been deployed increasingly in recent decades and this trend is certain to continue in the future.

Satellites have become of great importance over the last few decades and future missions will greatly aid the science as new instruments are flown. Several satellite missions are scheduled for launch in coming years, many of which will directly benefit studies of the Antarctic. While not exhaustive because of changing plans, some of these are briefly highlighted below.

SMOS - ESA's Soil Moisture and Ocean Salinity (SMOS) mission has been designed to observe soil moisture over the Earth's landmasses and salinity over the oceans. It is due to be launched in 2009. Ocean salinity observations will be important to studies and models of the circulation of the Southern Ocean and the role of sea ice. These observations are also hoping to contribute to improved characterisation of ice and snow covered surfaces and studies of accumulation rates in the Antarctic.

Cryosat-2 - After the loss of the first CryoSat in October 2005 due to a launch failure, the decision was made to build and launch the CryoSat-2 satellite. The mission's objectives remain the same as before – to measure ice thickness on both land and sea very precisely so as to provide conclusive proof as to whether there is a trend towards diminishing polar ice cover, furthering our understanding of the relationship between ice and global climate. CryoSat-2 is due for launch in 2009.

Landsat continuity mission - The Landsat series of satellites have provided detailed visible satellite imagery of the Antarctic over the past 30 years. The current satellite, Landsat-7 has developed problems with the ETM+ instrument and plans are now in place to launch a mission to provide continuity of Landsat data in 2011. Plans are also being put in place to implement various options to bridge any Landsat data gap.

GOCE - Launched in March 2009, ESA's Gravity field and steady-state Ocean Circulation Explorer (GOCE) is the latest satellite designed to provide an accurate and detailed global model of the Earth's gravity field and geoid. These data will contribute to the quantitative determination, in combination with satellite altimetry, of ocean currents and ocean circulation. These data will also provide estimates of the thickness of the polar ice sheets through the combination of bedrock topography derived from gradiometry and ice sheet surface topography from altimetry.

However, there is still a strong case for high quality observations from the staffed stations. This is the situation with sea level data, for example. There are good climate change and oceanographic arguments for Antarctica to be equipped with a well-maintained and long-term network of about a dozen high quality sea level stations distributed around the continent. These fundamental stations would automatically be components of the Global Sea Level Observing System, and would be supplemented by secondary stations where possible (e.g. on the Antarctic Peninsula), as redundancy of data provision will always be an important consideration, and further complemented by targeted bottom pressure recorders deployments for ocean circulation studies. Such evident minimum requirements are compatible with many other earlier statements of need within, for example, WOCE, Climate Variability and Predictability (CLIVAR) and Global Climate Observing System (GCOS) studies.

For the ocean it is clear that both the scientific research and operational forecasting communities would benefit from implementation of a Southern Ocean Observing System (e.g. Sparrow, 2007) to provide regular and routine observations of the ocean and surface meteorology in situ, to complement the ocean data obtained by satellites (which relate solely to the ocean surface). These data are essential for understanding ocean processes and as inputs to models for forecasting future change.

On land and in the sea there is a clear and urgent need for considerable improvement in biological survey data, without which both the baseline description of ecosystems and the identification of any biological responses or changes influenced by the physical environment are severely hampered. Coupled with this there is an urgent need for the establishment of robust and targetted biological monitoring programmes against which to identify change. Highlighting this point, despite much-used citation of clear biological responses to the currently rapid physical environmental changes along the Antarctic Peninsula (see Chapter 4), in reality there is currently only a single terrestrial location in the entire Antarctic where biological survey and resurvey data are available extending back to near to the start of the current warming phase (see Fowbert and Smith, 1994; Parnikoza et al., in review). This is at the Argentine Islands, coincidentally but positively a station with one of the longest, most detailed and best understood climate records within the Antarctic.

At the level of modelling and prediction, there is a clear need not only in the Antarctic but globally to develop future climate and oceanographic models that are both based on and predict at scales of resolution that are relevant to biological communities and processes. Coupled with this is the need to develop a more robust framework linking the description of climatic variables and trends at macro and micro scales. For instance, while macroclimatic trends of warming have been identified beyond dispute at many stations along the western Antarctic Peninsula and Scotia arc, in reality there are currently no data available (through the lack of long term monitoring studies driven by the normal culture of short-term research programme national funding) demonstrating either the existence of any form of trend in biologically-relevant microclimate variables, or whether any such trend mirrors that in microclimate (see Convey et al., 2003 for discussion of this lack, and an example of an alternative biological proxy). This is a fundamental gap in knowledge.

Chapter 3

Antarctic climate and environment history in the pre-instrumental period

Chapter Editor

Dominic A. Hodgson

Authors:

Nerile Abram, John Anderson, Luca Bargelloni, Peter Barrett, Michael J. Bentley, Nancy A.N. Bertler, Steven Chown, Andrew Clarke, Peter Convey, Alistair Crame, Xavier Crosta, Mark Curran, Guido di Prisco, Jane E. Francis, Ian Goodwin, Julian Gutt, Dominic A. Hodgson, Guillaume Massé, Valérie Masson-Delmotte, Paul A. Mayewski, Robert Mulvaney, Lloyd Peck, Hans-Otto Pörtner, Regine Röthlisberger, Mark I. Stevens, Colin P. Summerhayes, Tas van Ommen, Cinzia Verde, Elie Verleyen, Wim Vyverman, Christian Wiencke, Lorenzo Zane.

3.1 Introduction

This chapter reviews Antarctic climate and environmental history from deep time through to the early part of the instrumental period. First, the large scale changes that have shaped the climate and environments of Antarctica and its surrounding ocean are described; from its early origins as the centrepiece of the Gondwana supercontinent, though its subsequent breakup and isolation by the circumpolar Southern Ocean c. 180 Ma (million years) ago. This is followed by a description of the formation of the first ice sheets on the continent c. 34 Ma, and later, the development of the 100 ka (thousand years) glacial cycles that have characterised the last 0.9 Ma. The climate at the end of the last major glaciation, retreat of the ice sheet and its effects on global sea level, and the changing distribution of sea ice and its effect on climate are then described.

Our present interglacial, the Holocene is described in detail from a combination of ice cores, marine sediments, lake sediments and terrestrial records with a focus on both regional and continent-wide climate and environmental variability. This provides a background, against which the climate changes in the instrumental period (Chapter 4) and the predicted climate changes of the future (Chapter 5) can be compared. In order to facilitate such comparison, towards the end of the chapter we examine the evolutionary and biogeographical factors that have shaped the contemporary biota of the continent, including some of the physiological adaptations in the marine realm that have led to the distinctive marine fauna seen today. This section also highlights some of the reorganisations in species distributions that have occurred during the relatively minor natural climate changes of the Holocene that are being used to inform our understanding of the climate and biological changes that might be anticipated in the near future (Chapter 5).

The modern climate over Antarctica and the Southern Ocean results from the interactions within the ice sheet – ocean – sea ice – atmosphere system. Knowledge of how this system responds to past and present climate forcing and the phasing of climate events on regional to hemispheric scales is essential for understanding the behaviour of the Earth's past and future climate. Studying the history of Antarctic climate and environment is also important as it provides the context for understanding present day climate and environmental changes. Together, climate and environmental studies allow researchers to determine the processes that led to the development of our present interglacial state, and to define the ranges of natural climate and environmental variability on timescales from decades to millennia that have prevailed over millions of years in the past. Once this natural variability is known we can more confidently determine when changes have exceeded the natural state. We can also study, for example, past warm periods to identify and understand the processes that might have caused them.

In Antarctica, past climates and environments can be reconstructed from a range of palaeoenvironmental records. Each of these provides different, yet often complementary, information on the patterns and mechanisms of change and they span a range of different timescales. Instrumental data from satellites, ground-based instruments and oceanographic surveys provide detailed information about past climates for much of the last five decades. However, for records that pre-date this 'instrumental period' it is necessary to use the stratigraphic records contained in lake sediments, marine sediments and ice cores, and for the longest time periods, geological records on land and in deep ocean sediments.

Geological records are used to study the distant past, or 'deep time'. A particular value of studying climate in the distant past is that we need to go back more than 30 million years to find atmospheric CO_2 concentrations at more than twice pre-industrial levels, which we are projected to encounter by the end of this century. Geologists have termed this the "greenhouse" world, and both the biota and geothermometers indicate it was indeed much warmer, in contrast to the "icehouse" world that was initiated with the first ice sheets on Antarctica 34 million years ago. In the pages that follow we will briefly outline the main features of this world and chart the changes that have taken place as CO_2 levels declined and persistent ice sheets grew in the Antarctic.

Geological records of past climate on land in Antarctica are generally accessible only in mountains around the margin or in the Transantarctic Mountains due to the thick continental ice cover (Figure 3.1). These records either largely predate Gondwana breakup, which began around 180 million years ago, or in the case of sediments deposited since glaciation began they are mostly patchy and difficult to date (Barrett, 1996). However a few places, like the Antarctic Peninsula, have well-exposed post-breakup sedimentary strata that predate the first ice sheets. James Ross Island and Seymour Island in particular, near the tip of the Antarctic Peninsula, have yielded a great variety of plant and animal fossils from Late Cretaceous and early Cenozoic times (100-34 Ma ago) (Francis et al., 2008a). In addition, deep drilling into marine sediments on the continental margin of Antarctica (Figure 3.1) has provided records of Antarctica's climate during the tens of millions of years that it has been ice-covered (Barrett, 2009).

Deep drilling for direct records of Antarctic climate history has taken place in only a few locations around the margin (e.g. ship-based drilling in the Antarctic Peninsula, Prydz Bay and Ross Sea, and drilling from sea- and shelf-ice platforms in McMurdo Sound). In contrast, the collection of relatively short piston cores and vibracores of sediments from the seas around Antarctica has been widespread. These cores, typically less than 10 m long, but on occasion up to 40 m, provide typically late Quaternary sequences that record changes in, for example, the duration or extent of sea ice, the proximity and stability of glaciers and ice shelves, changing oceanic temperature, surface water productivity, and clues to the behaviour of ocean currents (Anderson, 1999). In addition, collections of deep-sea drill cores from the wider world ocean provide information from oxygen isotopes in plankton shells about the water temperature, and information from carbon isotopes about the water flux. This can be linked to production of cold deep bottom water from Antarctica and changes in the global signature of ice volume, which we can use to interpret the role of Antarctica in the global climate system over time (Kennett, 1982; Miller et al., 2008; Zachos et al., 2008).

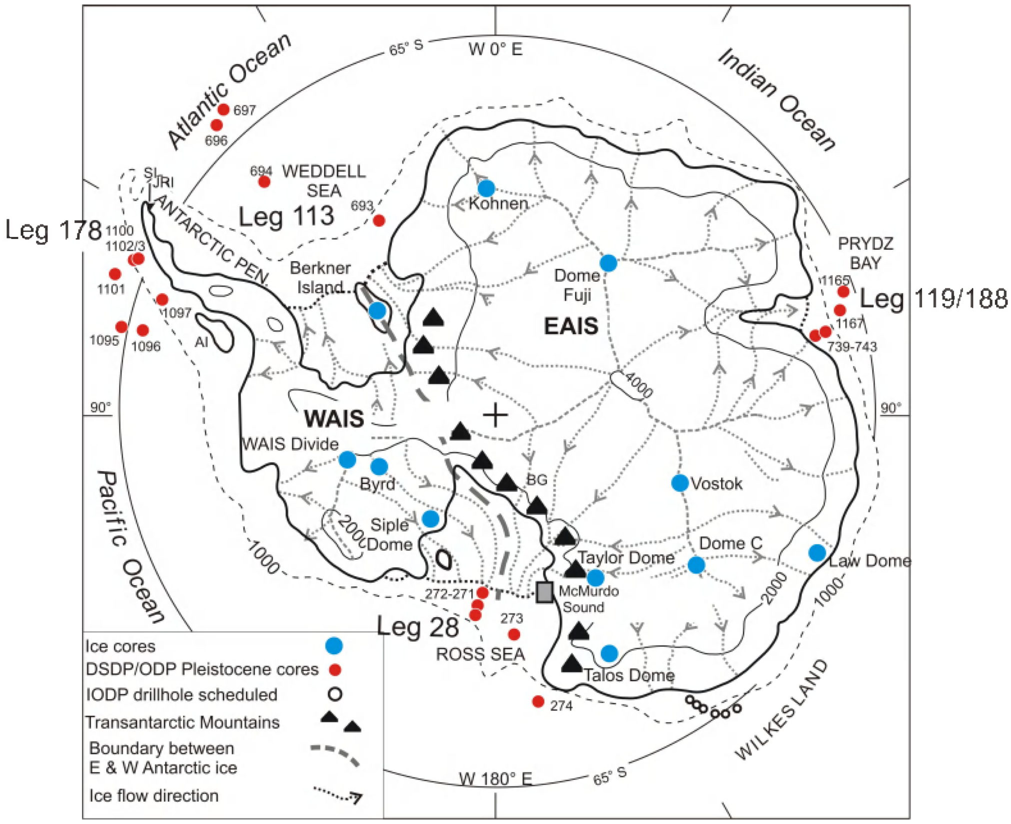


Figure 3.1 Map of Antarctica showing main geographic features (EAIS – East Antarctic Ice Sheet; WAIS – West Antarctic Ice Sheet) and locations of selected deep ice and sediment cores. Note: The McMurdo Sound area (box) includes 23 sediment cores, 14 on land and 9 from floating ice and ranging in length from ~ 50 to 1,285 m (see Naish et al., 2008a; Barrett, 2009 for details). Further records of Antarctic geological history are found at Seymour Island (SI), James Ross Island (JRI), Alexander Island (AI) and the Beardmore Glacier region (BG).

Ice cores provide a continuous high-resolution record of environmental change on timescales of up to c. 800 ka. Here water molecules, soluble and insoluble chemistry, particles, gases and clathrates trapped in the ice provide valuable archives of change in past atmospheric composition, temperature, precipitation, major trace chemistry and other environmental parameters such as changes in atmospheric circulation, sea ice extent, and volcanic activity. They offer the possibility of improving knowledge of key climate forcings (solar, volcanic) and of key climate feedbacks (greenhouse gases, aerosols). The ice core data are also crucial in benchmarking climate and ice sheet models against observations. The longest ice core records date back over 800 ka (Jouzel et al., 2007). However, from marine sediments, we know that just before this time the pattern of climate variability was different, with glacial cycles of only 40 ka length, compared with the 100 ka cycles today. If we are to understand the natural variability that has led us to our current climate, it is critical to understand what caused the length of the cycles to change. It is therefore a key challenge to identify the locations where ice older than 1 Ma can be found (IPICS, 2008) in order to test the hypothesis that the change from 40 ka to 100 ka cycles was caused by a lowering of atmospheric carbon dioxide concentrations. This is because the longer ice cores enable assessment of the natural relationship between greenhouse gases and climate, allowing us to deduce the underlying rules governing climate including the processes that control exchange of carbon dioxide between reservoirs. Many ice core studies also focus on the last 40 ka which includes the transition from the last glacial into the present warm interglacial and a sequence of abrupt swings in climate which are particularly apparent in Greenland ice cores and other climate archives, and to a lesser extent in Antarctic ice cores. This glacial-interglacial transition is the best documented global response to very large-scale changes in climate boundary conditions, and the ice cores enable researchers to reconstruct the precise sequence of events (including forcings and responses) through the last glacial-interglacial transition and related changes in greenhouse gas concentrations, accumulation and ice sheet extent, ocean surface conditions (e.g. sea ice and marine biological productivity) and changing thermohaline circulation patterns (IPICS, 2008).

Glacial geomorphological records such as moraines and trimlines, dated using cosmogenic isotopes from exposed rocks, also provide information on past ice sheet extent and thickness, and the timing of glaciation and deglaciation extending back from tens of thousands to millions of years (e.g. Summerfield et al., 1999; Mackintosh et al., 2007; Jamieson and Sugden, 2008). On land, the drilling of sediments that have accumulated in lakes in the ice-free regions of Antarctica provides geological records of climate that typically span just the present Holocene interglacial (last 11.7 ka), the most recent period of relatively stable global climate, and a time that has seen the rise of human civilizations. These sediments contain detailed and high resolution records of changes in temperature-related variables such as lake ice cover, moisture balance, biological productivity, ecology and species composition, together with records of deglaciation, isostasy and relative sea level change (Hodgson et al., 2004a).

Using geological records, the next section describes the geological setting of Antarctica and its features during "greenhouse" times, the changes as dynamic ice sheets grew and pulsated from orbital forcing, and the shift to a colder more persistent East Antarctic Ice Sheet 14 million years ago and finally the stabilisation of the West Antarctic Ice Sheet in the last million years.

3.2 Deep Time

Geologists have long held to the principle that "the present is the key to the past". Equally, the past may hold clues about the future that can be teased out from the geological record, particularly for environmental scenarios in past high CO_2 greenhouse worlds for which there are no current modern analogues, such as forests in polar regions. Palaeorecords show that profound changes in earth's past climate have followed well-defined patterns and shifts on time scales of thousands to millions of years. Hence studies of the past can be useful in providing insights into how the system responds and what can be expected to occur in response to future change.

3.2.1 The Greenhouse world: from Gondwana breakup to 34 million years

Some 200 million years ago (Ma) Antarctica was the centrepiece of the Gondwana super-continent, which began to break up around 180 Ma in the Jurassic Period of the Mesozoic Era. The Gondwanan fragments separated by sea-floor spreading largely between around 100 to 65 Ma, during the Cretaceous Period, although by this time Antarctica had already moved into a position over the South Pole (Figure 3.2). A rich record of plant and animal fossils from Late Cretaceous and early Tertiary times has been found on Seymour Island and Alexander Island in the Antarctic Peninsula. This demonstrates that, despite its polar position, Antarctica had a warm temperate climate that allowed the growth of lush forests (Figure 3.3). These were inhabited by dinosaurs in the Cretaceous and mammals during the early Tertiary (Francis and Poole, 2002).

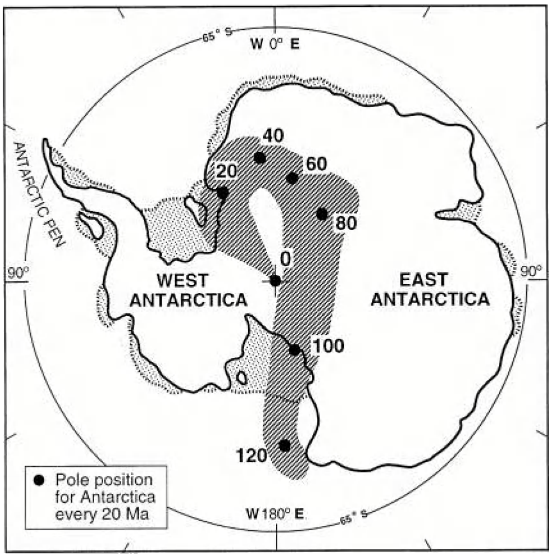


Figure 3.2 Apparent polar wander path for East Antarctica over the last 120 Ma (modified from DiVenere et al., 1994). The shaded area represents the modelled error envelope.

Analysis of changing biodiversity in the forests, and of the climate record preserved in the foliage, indicates that climate cooling signalled the decline of Antarctic warmth during the Middle Eocene, about 45 Ma, when heat-loving plants were lost from Antarctica and replaced by types such as species of *Nothofagus* trees that could tolerate cold climates (Francis et al., 2008a).

Over time, South America, Africa, India, Australia and New Zealand moved away from Antarctica, opening the South Atlantic, Indian and Southern Oceans (Figure 3.4). As the proto-Pacific Plate was subducted beneath the Antarctic Plate an active volcanic arc grew, the eroded remnants of which now form the Antarctica Peninsula. The western margin of Antarctica was thus characterised by volcanism and lateral movement of fragments of the formerly continuous Gondwana terrain (for example see McCarron and Larter, 1998). This western arc broke apart first at around 85 Ma south of New Zealand, but much later between South America and the Antarctic Peninsula to create the Drake Passage.

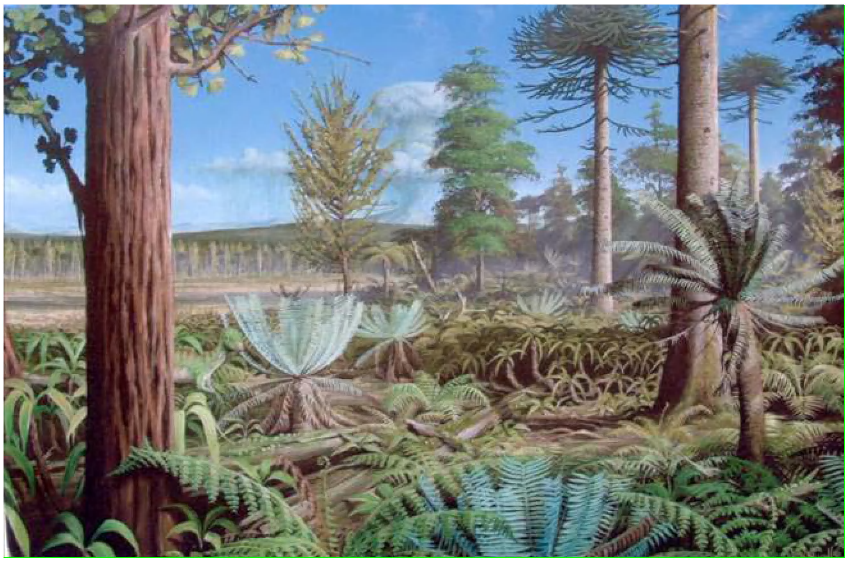


Figure 3.3 Reconstruction of mid-Cretaceous forests of Alexander Island, Antarctica. Based on the work of J. Howe, J.E. Francis, and numerous British Antarctic Survey geologists. Painted by Robert Nicholls (Francis et al., 2008a).

Over the last 40 million years, the Antarctic shelf experienced a series of tectonic, climatic and oceanographic events that lead to isolation from other oceans, establishment of colder conditions and complete replacement and impoverishment of the fish fauna. The deepening of the separations from other continental masses and the removal of the last barriers to circumpolar flow, allowed the establishment of the Antarctic Circumpolar Current (ACC) and, at its northern border, of the Antarctic Polar Front (APF), a roughly circular oceanic system running between 50°S and 60°S and extending to 2,000 m in depth.

Two key events allowed the establishment of the ACC: (1) the opening of the Tasman Seaway between Antarctica and Australia, which according to tectonics and marine geology occurred approximately 35.5-30 Ma (Kennett, 1977; Lawver and Gahagan, 2003; Stickley et al., 2004; Wei, 2004); (2) the opening of the Drake Passage between southern South America and the Antarctic Peninsula. The two openings allowed the development of the ACC. The timing of the opening of the Passage is controversial, with estimates ranging between 40 and 17 Ma (Barker and Burrell, 1977; Livermore et al., 2004, Scher and Martin, 2006). Such a relatively wide time window is due to contrasting views on the palaeo-elevation of key parts of the Scotia Arc and the Shackleton Fracture Zone (reviewed in Barker et al., 2007). Recently, using neodymium isotopes to detect the presence of Pacific seawater in the Atlantic sector Scher and Martin (2006) proposed that the opening of the Drake passage occurred as early as 41 Ma, and was thus completed before the establishment of the Tasman Seaway.

The timing of the final onset of the ACC is also uncertain (Barker et al., 2007). Estimates from sediment records range from late Eocene (40 Ma) to latest Miocene (8-10 Ma) (Wei and Wise, 1992; Scher and Martin, 2006). Concerns have been raised that some of these estimates could rather be climate-related and reflect a change in either temperature or productivity independent of ocean circulation (Barker et al., 2007). The older view that the ACC developed around the Eocene-Oligocene boundary leading to the first continental ice sheets on Antarctica by reducing heat transport into the region (Kennett, 1977) has been questioned by DeConto and Pollard (2003). Their modelling study shows the first ice sheets could have formed from a drop in atmospheric CO_2 levels from 3 to 2 times pre-industrial levels and that the opening of Drake Passage could provide only 20% of the cooling. The formation of the circum-polar deep water current may have subsequently helped the intensification of glaciation (Francis et al., 2008b).

Despite the uncertainty on both the timing of final onset of the ACC and its importance as a driver of climatic change, palaeoclimatic reconstructions leave little doubt about the fact that oceanographic and climatic processes responsible for current day glacial conditions in the Antarctic started in the same period of time. In fact, switching of climate conditions from "greenhouse" to "icehouse" in the Antarctic started in the late Eocene, 42 Ma, with an initial phase of strong but short-lived glaciations that match with the most recently published time estimates of formation of the Drake passage (Scher and Martin, 2006). A further drastic change occurred at the Eocene-Oligocene boundary ~34 Ma, (Tripati et al., 2005; Barker et al., 2007), leading to ice-sheet coverage similar in extent to the present ice sheet and contemporaneous with the opening of the Tasman Seaway.

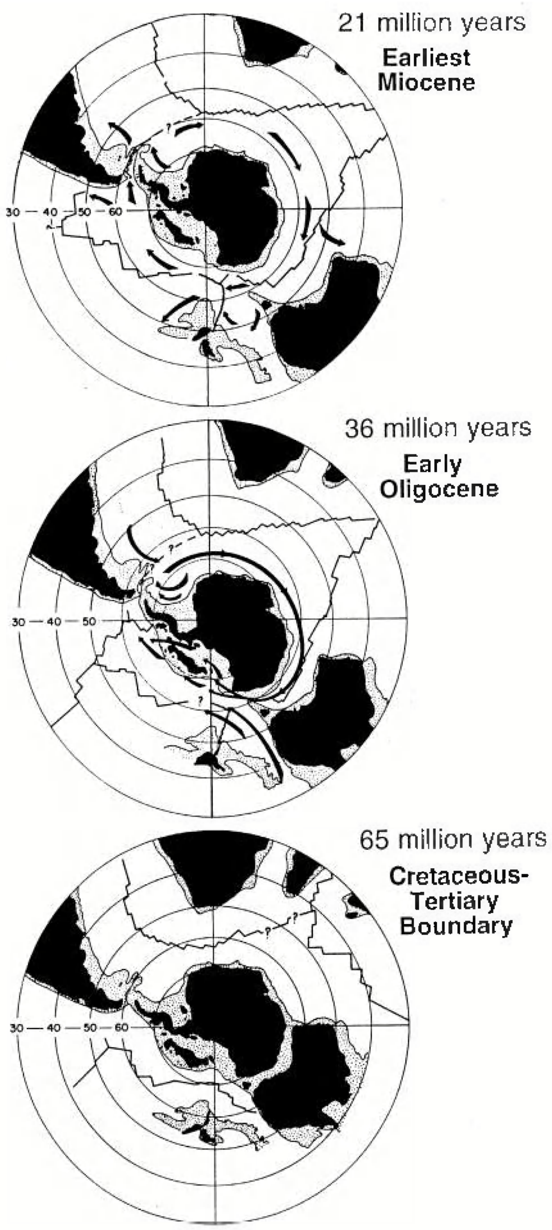


Figure 3.4 Three maps showing the progressive separation of Antarctica and the other Southern Hemisphere continents, leading to the opening of "ocean gateways" that allowed the development of the Antarctic Circum-polar Current sometime after 34 Ma. Modified from Kennett (1978).

The palaeontological record suggests these climatic events also influenced the Southern Ocean and its fish fauna. The best studied Antarctic fish fossils have been discovered in La Meseta Formation on Seymour Island (at the tip of the Antarctic Peninsula). The deposit from the late Eocene showed a fauna still "cool and temperate in character" that possibly lived in waters such as those found today around Tasmania, New Zealand and southern South America (Eastman, 2005). Accordingly, the earliest cold-climate marine Antarctic faunas are thought to date back to the latest Eocene-Oligocene (35 Ma).

With the ACC, the Southern Ocean became the cold isolated habitat that we know today. However, ice sheet coverage oscillated throughout the Cenozoic and probably produced repeated shifts of species distribution in Antarctic coastal waters, allowing allopatric speciation and diversification of Antarctic taxa (Rogers, 2007). In addition, the present APF is now suspected to be "leaky" and to allow transport of plankton north in mesoscale eddies with cold cores (Clarke et al., 2005).

Atmospheric CO_2 levels ranged from ~3,000 ppm in the Early Cretaceous (at 130 Ma) to around 1,000 ppm in the Late Cretaceous (at 70 Ma) and Early Cenozoic (at 45 Ma) (e.g. Pagani et al., 2005; Royer, 2006), leading to global temperatures at least 6 or 7°C warmer than present at times. These high CO_2 levels were probably the consequence of volcanic outgassing. Temperatures peaked at ~85 Ma during the mid-Late Cretaceous when sub-tropical climates prevailed over the pole (Francis et al., 2008a). Even in this Cretaceous greenhouse world from oxygen isotope and backstripping data some researchers have inferred sea level changes in the order of tens of metres and intermittent polar ice sheets (Miller et al., 2008), although there is as yet no geological evidence (Thorn et al., 2007). Temperatures peaked again at 50 Ma (Figure 3.5) and subsequently declined, as did atmospheric CO_2 levels (Pearson and Palmer, 2000; Pagani et al., 2005; Royer, 2006). Superimposed on the high CO_2 world of the Early Cenozoic, deep-sea sediments provide evidence of the catastrophic release of more than 2,000 gigatonnes of carbon into the atmosphere from methane hydrates around 55 Ma ago, at the Paleocene-Eocene boundary, raising global temperatures by a further ~4-5°C, although they recovered after about 100,000 years (Figure 3.5) (Zachos et al., 2003; Zachos et al., 2005).

3.2.2 Into the Icehouse world: the last 34 million years

The first continental-scale ice sheets formed on Antarctica close to the Eocene-Oligocene boundary around 34 Ma), and are physically recorded in strata from Prydz Bay, McMurdo Sound and Seymour Island (Francis et al., 2008b). Deep-sea isotope data suggest they were similar in size to that of today (Miller et al., 2008). According to DeConto and Pollard (2003) the development of these first Antarctic ice sheets was triggered in an interval when the Earth's orbit of the Sun favoured cool summers as atmospheric CO_2 levels declined below a critical threshold (~2.8 times pre-industrial). This decline has been documented (Pagani et al., 2005; Siegert et al., 2008) and has been ascribed to reduced CO_2 outgassing from ocean ridges, volcanoes and metamorphic belts and increased carbon burial (Pearson and Palmer, 2000), dropping global temperatures from at least 6 to around 4°C higher than today (Figures. 3.5 and 3.6). The fall in CO_2 levels at this time is also reflected in a 1-km drop in the calcium compensation depth in the tropical Pacific Ocean (Coxall et al., 2005).

The pulsating style of Antarctic glaciation in Oligocene-early Miocene times is best recorded from 1,500 m of nearshore marine sediments drilled off Cape Roberts in the southwest Ross Sea (Figures 3.1 and 3.7) and in the middle to late Oligocene Polonez Cove Formation, exposed on south-eastern King George Island, South Shetland Islands (Troedson and Smellie, 2002). At Cape Roberts, sediments resulting from 55 glacial-interglacial cycles have accumulated close to sea level on the subsiding margin of the Victoria Land Basin, spanning the period from 33 to 17 Ma (Barrett, 2007). The site was close to the edge of the continental ice sheet and recorded its cyclic expansion and contraction on Milankovitch frequencies (41 ka and 100 ka, Naish et al., 2001; Naish et al., 2008b; Huybrechts, 2009; Naish et al., 2009; Pollard and DeConto, 2009). The changes in sediment type that characterize the cycles - from glacial deposits through nearshore sand and offshore mud to sand again - indicate sea level changes on a scale of tens of metres (Dunbar et al., 2008). Despite episodes of extensive ice cover palynological studies indicate a coastal vegetation ranging from low woodland (*Nothofagus*) to tundra persisting through the Oligocene with a slight cooling in early Miocene times (Prebble et al., 2006).

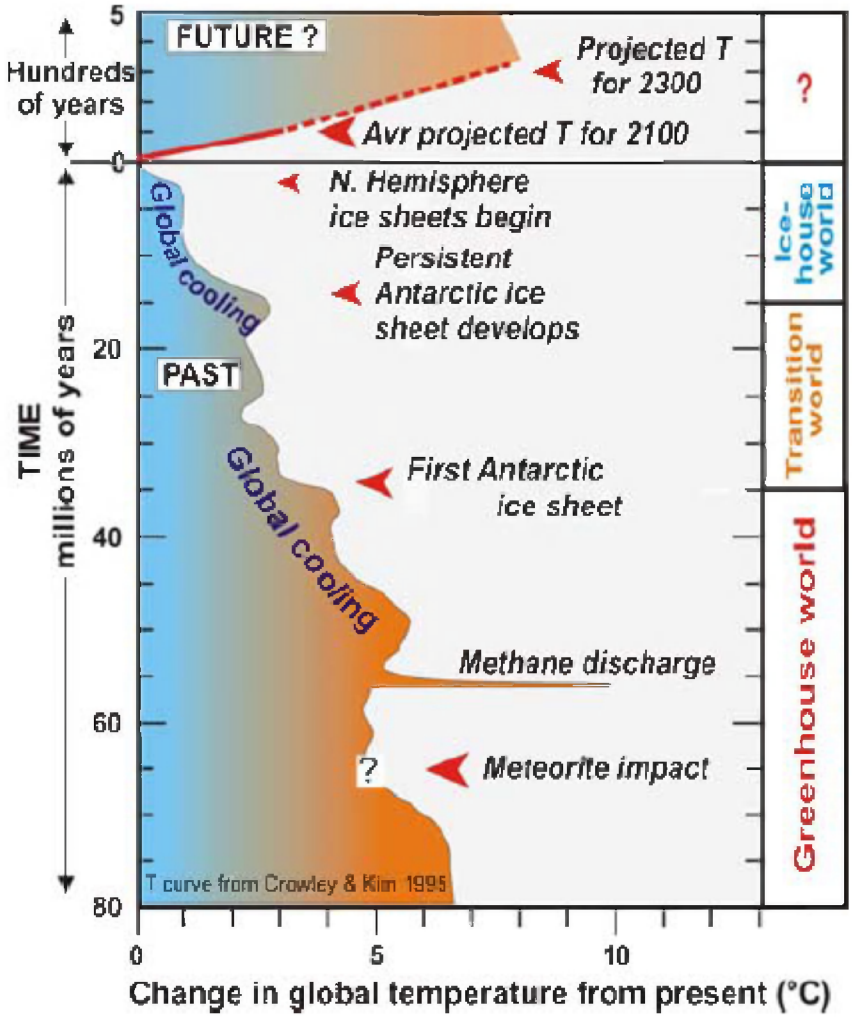


Figure 3.5 Change in average global temperature over the last 80 Ma. The temperature curve of Crowley and Kim (1995) is modified to show the effect of the methane discharge at 55 Ma (Zachos et al., 2003; Zachos et al., 2005). The future rise in temperature expected from energy use projections shows the Earth warming back into the 'greenhouse world'(from Mayewski et al., 2009).

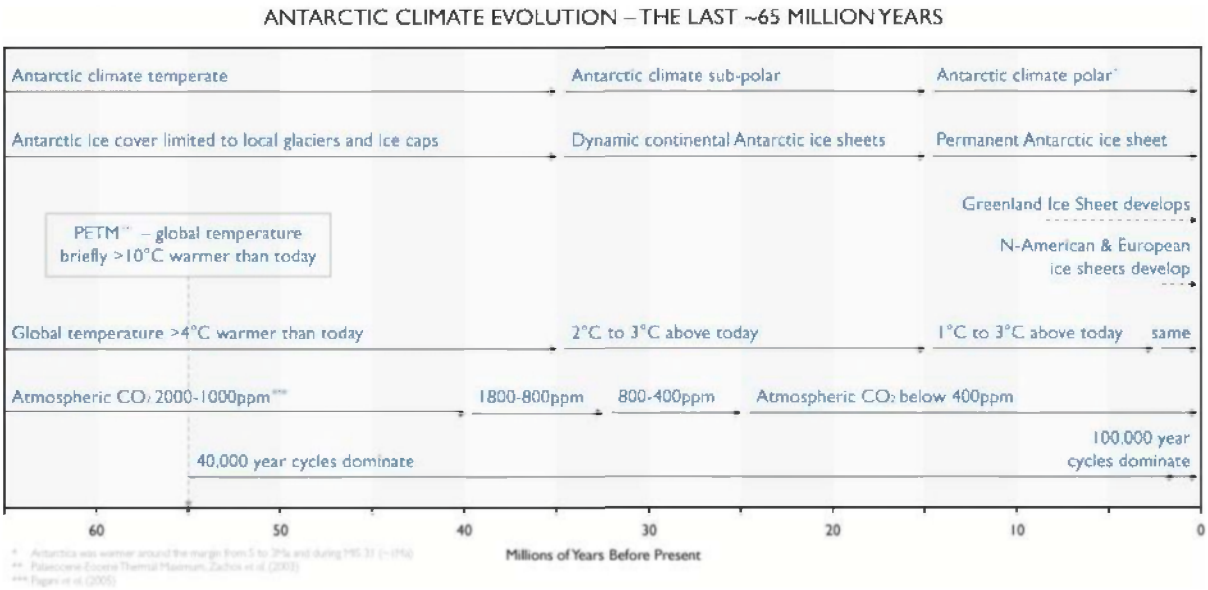


Figure 3.6 Main climatic events of the last 65 million years in the Antarctic context (adapted from Mayewski et al., 2009).

Recent studies of an ancient glacial landscape in the Olympus Range on the inland edge of the McMurdo Dry Valleys have revealed warm-based glacial deposits overlain by ridges of cold-based gravelly debris, each bearing volcanic ash beds ~14 and ~13.6 Ma respectively. The landscape has changed little since that time, the lack of alteration being ascribed to a persistent frozen state from that time on, recording a sharp Middle Miocene cooling (Lewis et al., 2007). Ash-bearing proglacial lake beds, also dated at ~14 Ma, include an ostracod fauna, *Nothofagus* pollen and beds of moss, and are regarded as possibly the last vestiges of this fauna and flora in the region (Ashworth et al., 2007; Lewis et al.; 2008). A well preserved flora of *in situ Nothofagus* dwarf shrubs, mosses and cushion plants, along with the remains of beetles, molluscs, fish and parts of flies, from the Sirius Group, Oliver Bluffs in the Dominion Range, Transantarctic Mountains also provide good evidence for tundra conditions in this region only 300 miles from the South Pole. The fossils are preserved in a glacio-fluvial-lacustrine-palaeosol layer that represents a warmer interval that prompted glacial retreat between colder intervals during which glaciers were present at that site (Francis and Hill, 1996; Ashworth and Cantrill, 2004). Unfortunately an undisputed age for these deposits is not available.

The sharp cooling in the Middle Miocene has long been known from deep-sea isotopic studies (Shackleton and Kennett, 1975), and was probably caused by the growing thermal isolation of Antarctica and related intensification of the Antarctic Circumpolar Current described above, accompanied by a drop in atmospheric CO_2 (Shevenell et al., 1996). This thickened the ice sheet to more or less its modern configuration, which is thought to have persisted through the early Pliocene warming from 5 Ma to 3 Ma (Kennett and Hodell, 1993; Barrett, 1996; McKay et al., 2008).

During the Pliocene, mean global temperatures were 2-3°C above pre-industrial values (Figure 3.6) with CO_2 values less than 400 ppm and sea levels 15-25m above modern levels (Raymo et al., 1996; Jansen et al., 2007, and references therein). The Antarctic margin also records Pliocene temperatures several degrees warmer than today in diatom-bearing coastal sediments (Harwood et al., 2000) and offshore cores (Whitehead et al., 2005).

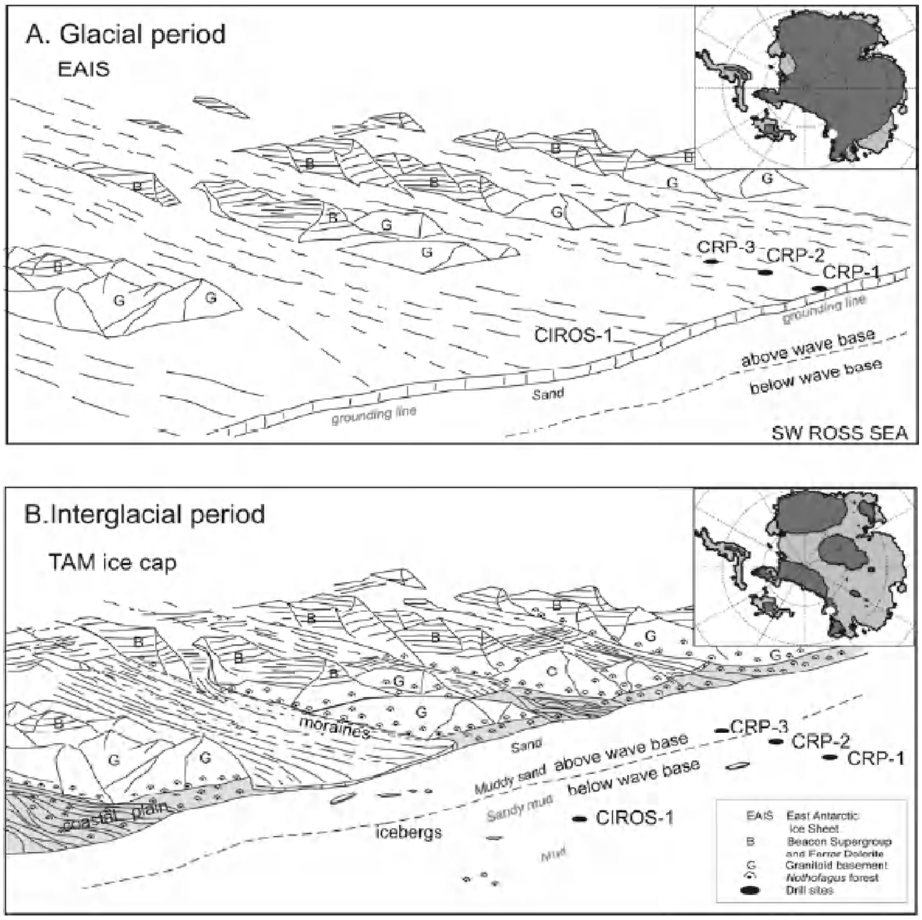


Figure 3.7 View of the Victoria Land coast off Cape Roberts during Oligocene and early Miocene times (from Barrett, 2007). (A) Glacial period, with an expanded inland ice sheet feeding thin temperate piedmont glaciers depositing sediment to a shallow shelf to be reworked by waves and currents nearshore with mud settling out offshore. (B) Interglacial period, with higher sea level and a much reduced ice sheet. Sediment carried to the coast largely by proglacial rivers. Low woodland beech forest at lower elevations. Insets for A and B show examples of the modelled extent of an ice sheet that might have existed during glacial and interglacial periods in early Oligocene times, representing 21 × 10^6 km^3 of ice (50 m of sea level equivalent) in A, and 10 × 10^6 km^3 of ice (24 m of sea level equivalent) in B (DeConto et al., 2007). The location of Cape Roberts is shown as a white filled circle. TAM = Transantarctic Mountains, EAIS = East Antarctic Ice Sheet.

A particularly instructive record comes from the ANDRILL McMurdo Ice Shelf site, where over 1,200 m of strata dating back to 13 Ma were cored from a deep-water basin south of Ross Island (Naish et al., 2007; 2008a,b, 2009). The record comprises many cycles of sedimentation, alternating between deposition beneath grounded ice (diamictite) and ice shelf ice (mudstone) in the last million years. However, from 1 to ~5 Ma depositional environments ranging from grounded ice (diamictite) to open water (diatomite) (Figure 3.8). The diatomite beds show the drill site to have been essentially ice-free at this time, i.e. no McMurdo Ice Shelf and hence no Ross Ice Shelf. The lack of buttressing from the loss of the Ross Ice Shelf in turn implies a much reduced West Antarctic Ice Sheet (Mercer, 1978; Dupont and Alley, 2005a). However geomorphological evidence, and the antiquity of high level surfaces in the Transantarctic Mountains (Sugden et al., 1993), along with modelling

Pliocene ice sheets (Hill et al., In Press), supports the persistence of an ice sheet in the East Antarctic interior through this period. A new ice sheet model by Pollard and DeConto (2009) confirms the persistence of East Antarctic ice and the disappearance of the West Antarctic ice sheet during warm Pliocene times and even as recently as the MIS 31 interglacial stage a little over a million years ago (Figure 3.8).

Global cooling from around 3 Ma onwards (Ravelo et al., 2004) led to the first big ice sheets on North America and NW Europe around 2.6 Ma (Shackleton et al., 1984), enhancing the Earth's climate response to orbital forcing with a 40,000 year cyclicity, and taking us to the Earth's present intense "ice house" state. For the last million years (Figure 3.8) this has alternated between (i) longer (100,000 years) glacial cycles, when much of the Northern Hemisphere was ice-covered, global average temperature was around 5°C colder, and sea level was approximately 120 m lower than today, and (ii) much shorter warm interglacial cycles like that of the last ~10,000 years, with sea levels near or slightly above those of the present.

3.3 The Last Million Years

This section describes the Antarctic climate and environmental history of the last 1 Ma, including the development of the current pattern of glacial-interglacial cycles. This is followed by a description of the transition from the height of the last ice age – the Last Glacial Maximum (LGM, c. 21 ka BP), to our present interglacial state, a period known as Termination 1. It includes an account of the deglaciation of the continental shelf, coastal margin and continental interior and the impact of this loss of ice on global sea level. Finally we review the influence of changing sea ice distributions on climate.

3.3.1 Glacial interglacial cycles: the ice core record

Ice cores spanning up to the past 800 ka provide some of the best records of Antarctic climate history and reveal not only the high degree of responsiveness of the ice sheet to changes in orbitally induced insolation patterns, but also a close association between atmospheric greenhouse gases and temperature (Figures 3.9 and 3.10). Unlike today, the CO_2 signal follows the temperature signal (Caillon et al., 2003; Ahn et al., 2004; Loulergue et al., 2007; Lüthi et al., 2008), signifying that changes in the concentrations of atmospheric CO_2, as well as other greenhouse trace-gases, at the scale of the glacial-interglacial cycles were initiated by climatic mechanisms acting on their different reservoirs. Nevertheless, once CO_2 began rising it would provide a positive feedback on temperature (Raynaud et al., 2003; Jansen et al., 2007). Köhler and Fischer (2006) used various paleo-climatic records from the EPICA Dome C Antarctic ice core and from oceanic sediment cores covering the last 740 kyr to force a carbon cycle box model to provide constraints on the factors causing variation in the CO_2 signal and its relation to temperature including: Southern Ocean vertical mixing, ocean temperature, iron limitation of the marine biota in the Southern Ocean, carbonate compensation, North Atlantic deep water formation, gas exchange due to a decreasing sea ice cover, sea level rise and rising terrestrial carbon storage. Other models have identified the importance of on and off states of the southern overturning circulation, a result of a positive feedback that involves the mid-latitude westerly winds, the mean temperature of the atmosphere, and the overturning of southern deep water. Cold glacial climates seem to have equatorward shifted westerlies, which allow more respired CO_2 to accumulate in the deep ocean. Warm climates like the present have poleward shifted westerlies that flush respired CO_2 out of the deep ocean (Toggweiler et al., 2006).

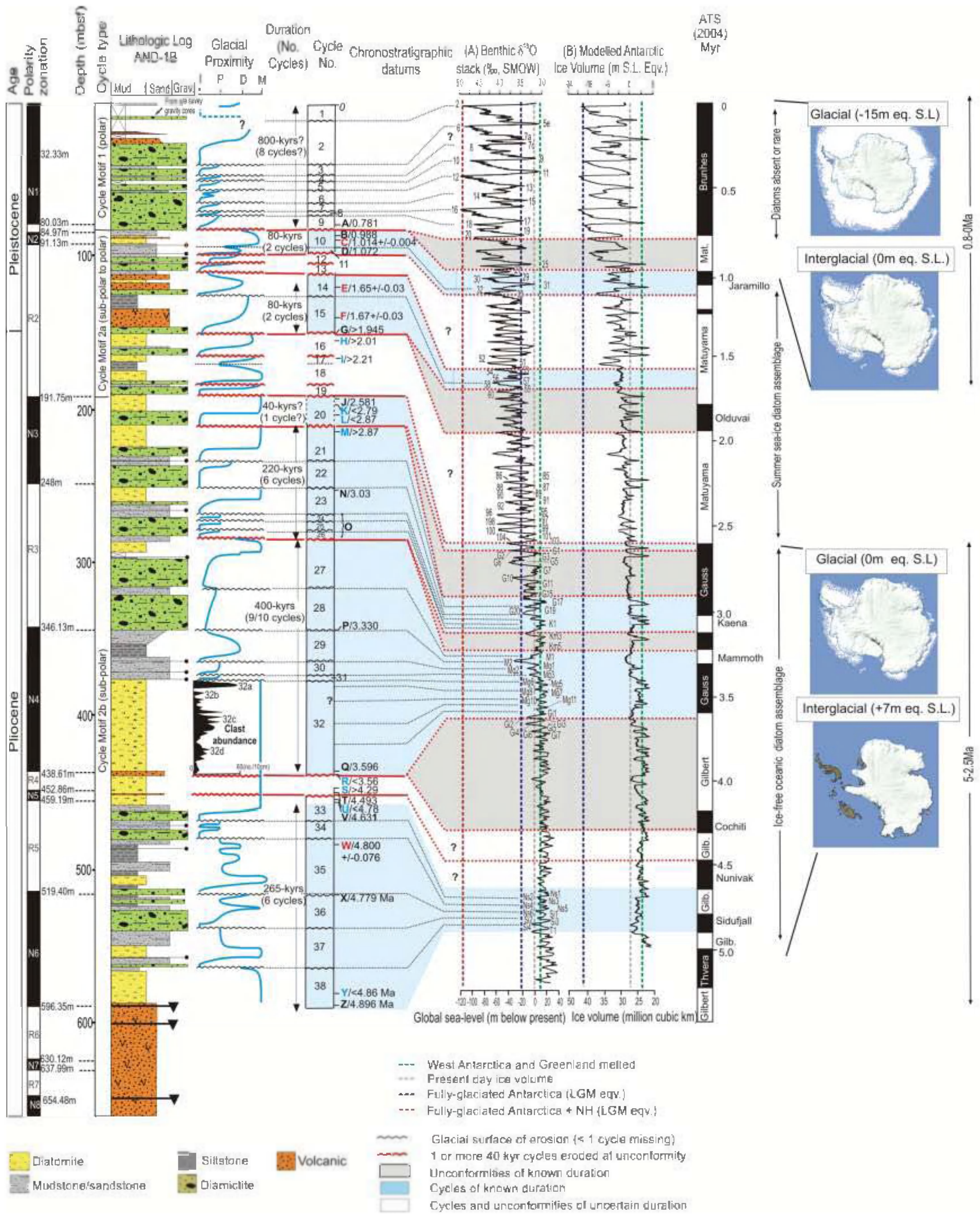


Figure 3.8 Lithological column and inferred glacial proximity curve for the upper 600 m of the AND-1B drillcore recovered by the ANDRILL McMurdo Ice Shelf Project (Naish et al., 2009), compared with the continuous deep-sea oxygen-isotope record (Lisiecki and Raymo, 2005) and the modelled contribution of Antarctic Ice Sheet to sea level change for the last 5 million years (Pollard and DeConto, 2009). The "glacial proximity" column shows glacial surfaces of erosion, which mark boundaries of orbital-scale, glacimarine sedimentary cycles, and the prospect of cycles missing on account of erosion. The glacial proximity curve tracks the cycles themselves from inferred depositional environment - I = ice contact, P = proximal glacimarine, D = distal glacimarine, M = open marine. Pairs of images of the Antarctic Ice Sheet on the right show hypothesized glacial and interglacial states for the last million years and the period between 1 and ~5 Ma from Pollard & DeConto (2009). Figure adapted by Tim Naish from Naish et al. (2009).

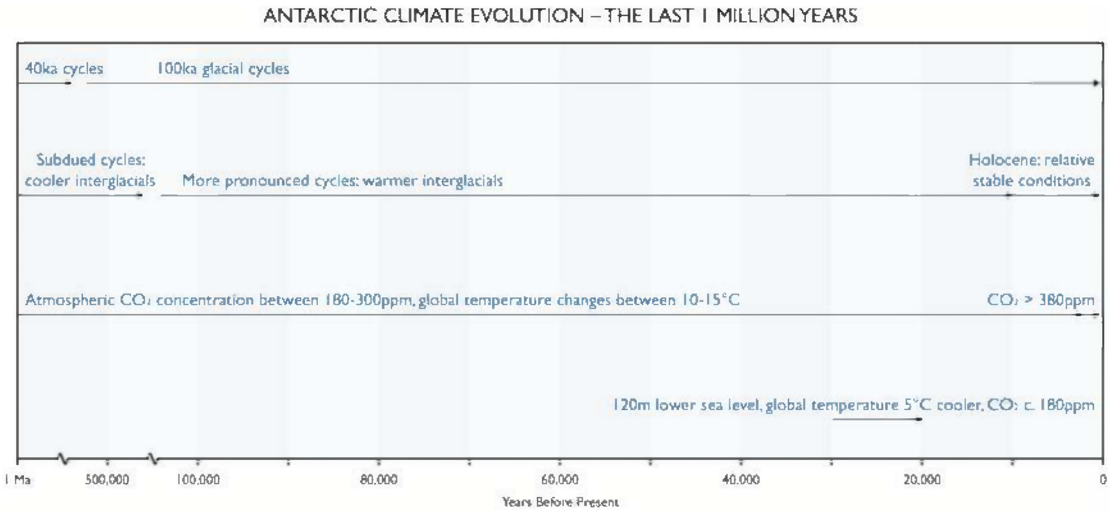


Figure 3.9 Main climatic events of the last 1 Million years in the Antarctic context

The Antarctic ice core data show that the Earth's climate has oscillated through eight glacial cycles over the last 800 ka, with CO_2 and mean temperature values ranging from 180 ppm and 10°C in glacials, to 300 ppm and 15°C in interglacials (Jouzel et al., 2007). The pattern has changed through time, with a fundamental reorganisation of the climate system at 900-600 ka from a world that for the preceding 33 million years had been dominated by 41 ka oscillations in polar ice volume, to a 100 ka climatic beat. It is critical to understand what caused the length of these cycles to change; a challenge that will likely be met by finding Antarctic locations where ice older than 1 Ma is present (IPICS, 2008). As stated previously, the effect on global sea level was profound, with sea level dropping by 120 m on average during glacial periods (e.g. Figure 6.8 in Jansen et al. (2007)).

In order to separate the roles of orbital and greenhouse gas forcings on global climate, and to test Milankovitch theory of glacial cycles as the driver for the 100-kyr cycles, an accurate ice-core chronology to within ~2 kyr (~10 % of a precession cycle) is required. Such a chronology for 80-400 kyr BP have been realized by orbital tuning of the record of oxygen-to-nitrogen concentration ratio (O_2/N_2) in trapped air with the local summertime insolation (Kawamura et al., 2007; Suwa and Bender, 2008) (Figure 3.10, A-C). O_2/N_2 in the ice cores is depleted from the original atmospheric ratio because of physical fractionation during air trapping at ~100 m depth (Severinghaus and Battle, 2006), and its magnitude is presumably controlled by physical properties of snow, which in turn is modified by local summer insolation when the layer was originally at the surface (Bender, 2002). Although exact mechanisms are not yet well known, empirical evidence indicates that the O_2/N_2 variation is nearly phase-locked to the local summer solstice insolation, with negligible climatic influences (Kawamura et al., 2007).

The Dome Fuji and Vostok temperature records on the new chronology closely follow boreal summer insolation (Figure 3.10, D). Thus, the previous arguments of early Antarctic warming following the southern summer insolation to trigger northern deglaciation are not supported. Further, the onsets of the last four Antarctic terminations are found to lag behind the minima of 65°N June solstice insolation by 2 -7 kyr. For the last three glacial inceptions, Antarctica cooled in phase with the decrease in northern summer insolation and before significant decreases seen in CO_2 and sea level (Figure 3.10, D-F) curves. These timings are consistent with the Milankovitch theory that high northern latitude summer insolation drives the 100-kyr glacial cycles by changing summertime temperature and thus long-term glacial mass balance (Raymo, 1997; Denton et al., 2006), with large amplification by albedo and

CO_2 feedbacks. This view is supported by the analysis of marine sediment data (Bintanja and Van De Wal, 2008) that shows the decrease in ice volume did not lag behind the increase of air temperature at the 100-kyr terminations after ~700 kyr BP.

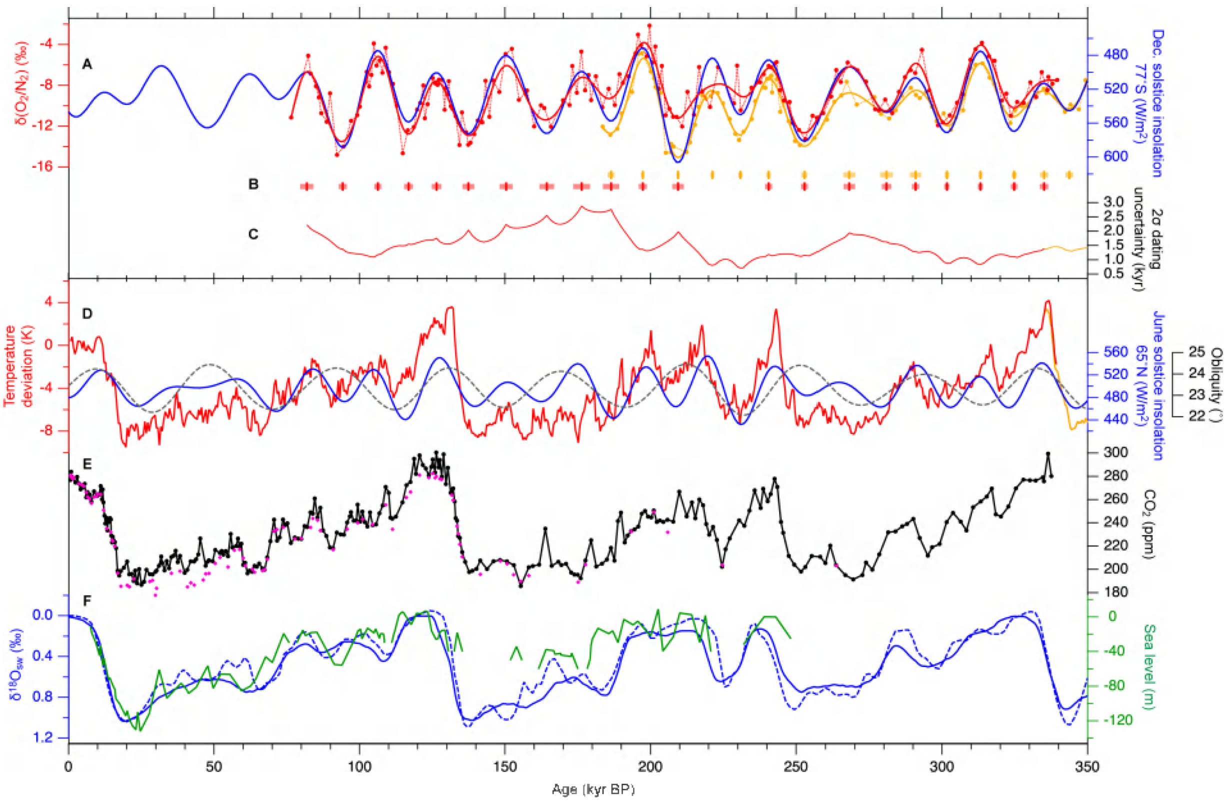


Figure 3.10 Accurately dated Antarctic ice-core records and the comparison with sea level proxies (Kawamura et al., 2007). A) Orbital tuning of O_2/N_2 in the Dome Fuji core (red) and Vostok core (orange) with local summer insolation (SI; blue). B) Tuning tie points with 2σ error bars due to the noise of the O_2/N_2 data. C) 2σ dating uncertainty by combining those of the tie points and interpolation procedure. D) Dome Fuji temperature (red), Vostok temperature (orange), Northern Hemisphere summer insolation (SI; blue) and obliquity (gray). E) CO_2 concentration from the Dome Fuji core by wet extraction (black) and dry extraction (purple). Note that the wet values are too high in LGM and dry values are too low in MIS 3 due to artifacts during extractions. F) Sea level reconstructions using radiometrically dated corals blue; (Thompson and Goldstein, 2005), and orbitally tuned $\delta^{18}O$ of seawater based on marine sediment records through ice sheet modelling blue solid; (Bintanja et al., 2005) and regression analyses blue dashed; (Waelbroeck et al., 2002). Note they are plotted on their own chronologies.

An alternative view is that Antarctic temperature is independent of northern climate and is controlled by the duration of Antarctic summer, which shows nearly identical pattern to the northern summer insolation (intensity) (Huybers and Denton, 2008). However, for the last termination, the onset of sea level rise (at 19 kyr BP) was earlier than that of the Antarctic warming (and CO_2 rise), and there may be causal relationship between them through the bipolar seesaw mechanism (Clark et al., 2004). In addition, tight bipolar coupling (see below) seems to operate for orbitally driven warming events (A4 and A7 events in Blunier and Brook (2001); see also EPICA (2006)). Global climate modeling would be required to quantitatively estimate the contributions of local and distant orbital forcings to the Antarctic temperature (and CO_2) changes.

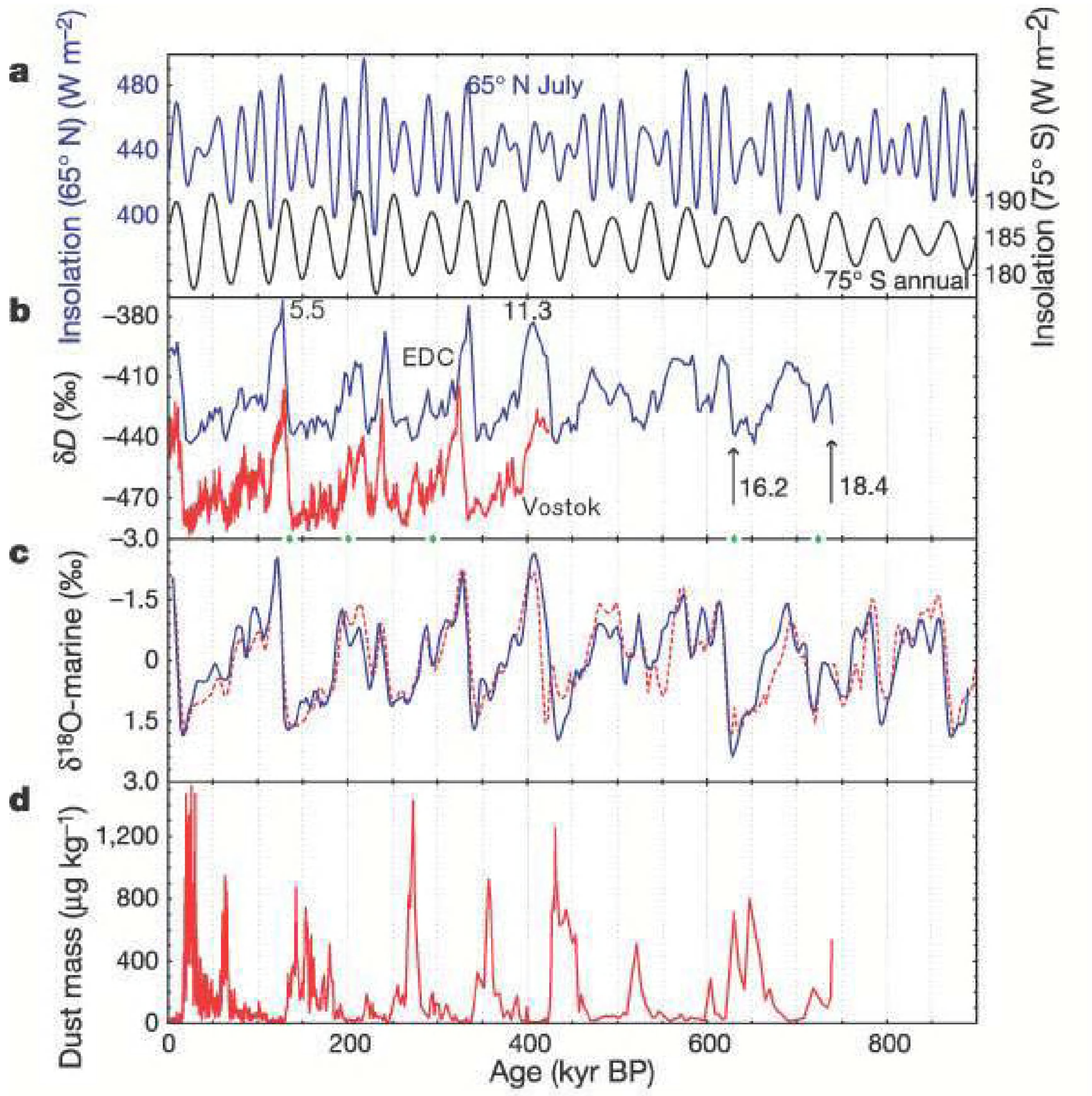


Figure 3.11 (a) Insolation records. Upper blue curve (left axis), mid-July insolation at 65°N; lower black curve (right axis), annual mean insolation at 75°S, the latitude of the Dome C ice core. (b) δD from EPICA Dome C (3,000-yr averages). Vostok ice core δD (red) is shown for comparison and some Marine Isotope Stage (MIS) numbers are indicated; the locations of the control windows (below 800-m depth) used to make the timescale are shown as diamonds on the x axis. (c) Marine oxygen isotope record. The solid blue line is the tuned low-latitude stack of drill sites MD900963 and ODP677; to indicate the uncertainties in the marine records we also show (dashed red line) another record, which is a stack of seven sites for the last 400 ka but consisting only of ODP drill site 677 for the earlier period. Both records have been normalized to their long-term average. (d) Dust from the EPICA Dome C ice core (EPICA, 2004).

Summer insolation intensity is dominated by the precession signal, whose amplitude is modulated by eccentricity. However, the obliquity signal becomes dominant if longer periods (e.g., half year average) or summer energy (time-integrated insolation over the days whose daily insolation exceed a certain threshold) (Huybers, 2006) with a low threshold is taken into consideration. The last four terminations indeed started when obliquity was relatively large (Huybers and Wunsch, 2005; Suwa and Bender, 2008). With regard to the relative importance of precession and obliquity, termination III is interesting because the precession (as in northern summer insolation) and obliquity were out of phase just before the termination. Here, the pattern of temperature evolution is similar to northern summer insolation. A longer ice core record with the O_2/N_2 chronology is needed for investigating more terminations before this problem can be resolved. With the second Dome Fuji core; (~720 kyr BP Motoyama, 2007) and future deep cores (IPICS, 2008), the accurate chronology will be extended to permit rigorous testing of the phase stability of terminations with respect to both precession and obliquity (Huybers and Wunsch, 2005).

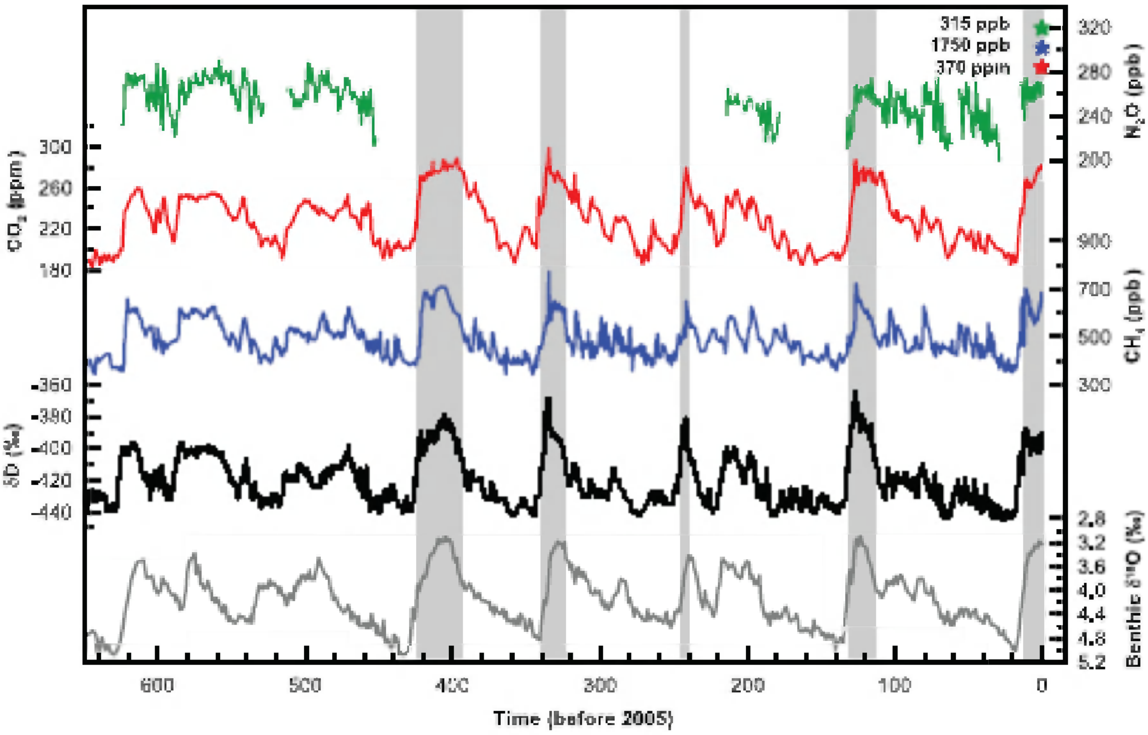


Figure 3.12 Variations in deuterium (δD; black), a proxy for local temperature, and the atmospheric concentrations of the greenhouse gases CO_2 (red), CH_4 (blue), and nitrous oxide (N_2O; green) derived from air trapped within ice cores from Antarctica and from recent atmospheric measurements (Petit et al., 1999; Indermühle et al., 2000; EPICA, 2004; Siegenthaler et al., 2005a; Siegenthaler et al., 2005b; Spahni et al., 2005). This points to a strong coupling of the climate and the carbon cycle. The shading indicates the last interglacial warm periods. Interglacial periods also existed prior to 450 ka, but these were apparently colder than the typical interglacials of the latest Quaternary. The length of the current interglacial is not unusual in the context of the last 650 ka. The stack of 57 globally distributed benthic $\delta^{18}O$ marine records (dark grey), a proxy for global ice volume fluctuations (Lisiecki and Raymo, 2005), is displayed for comparison with the ice core data. Downward trends in the benthic $\delta^{18}O$ curve reflect increasing ice volumes on land. Note that the shaded vertical bars are based on the ice core age model (EPICA, 2004), and that the marine record is plotted on its original time scale based on tuning to the orbital parameters (Lisiecki and Raymo, 2005). The stars and labels indicate atmospheric concentrations at year 2000 (From IPCC, 2007).

Ice cores from both Antarctica and Greenland show that temperatures were between 2-5°C higher than today in recent past interglacials (Jansen et al., 2007). This is matched by biological evidence in the lake sediment record in Antarctica (Hodgson et al., 2006a). At the same time, global sea levels were 4-6 m higher than today's (Jansen et al., 2007). During the Pliocene, when temperatures were about the same amount above today's values, sea levels were 15-25 m higher than today's (Jansen et al., 2007). The 10-20 m difference in sea levels between the Pliocene and the last interglacial, despite the fact that both shared more or less the same global temperature, may reflect the longer time available in the Pliocene to melt significant parts of the Antarctic ice sheet (the longest past interglacial lasted 30 ka – see Jansen et al. (2007)). Given that modern CO_2 levels of around 380 ppm are higher than at any time in the last 850 ka (EPICA, 2004) (Figure 3.13), and most likely in the last 25 million years (Royer, 2006; Jansen et al., 2007), the data from these past warm periods suggest that if the currently rising levels of CO_2 drive temperatures to levels last seen in past interglacials or in the Pliocene, then high sea levels may be expected.

Understanding the dynamics of the Earth's climate system requires knowledge of the phasing of climate events on regional to hemispheric scales. Methane emitted from wetlands is rapidly mixed throughout the atmosphere and can be used to correlate ice cores from Antarctica and Greenland. Using this technique it can be shown that climatic events of millennial to multi-centennial duration in the north and south polar regions are related (Figure 3.13, and EPICA (2006)), but exhibit a see-saw relationship. Antarctic warm events correlate with but precede those in Greenland, and vice versa (Figure 3.13). Warming in the Antarctic begins when Greenland is at its coldest during the so-called Dansgaard/Oeschger (D/O) events (Figure 3.13). One theory exploring this see-saw relationship states that D/O cold events, and their associated influx of meltwater, reduce the strength of the North Atlantic Deep Water current (NADW), weakening the Northern Hemisphere circulation and therefore resulting in an increased transfer of heat polewards in the Southern Hemisphere (Maslin et al., 2001). This warmer water results in melting of Antarctic ice, thereby reducing density stratification and the strength of the Antarctic Bottom Water current (AABW). This allows the NADW to return to its previous strength, driving Northern Hemisphere melting and another D-O cold event. Eventually, the accumulation of melting reaches a threshold, whereby it raises sea level enough to undercut the Laurentide ice sheet - causing armadas of icebergs to discharge into the North Atlantic in a Heinrich event and resetting the cycle. The signals differ from one hemisphere to the other, warming being gradual in the Antarctic, but abrupt in Greenland (Figure 3.13). The relationship is thought to reflect the nature of the connection between the two hemispheres via the ocean's meridional overturning circulation (MOC), with the lag reflecting the slow transfer of warmth from south to north. Heat retention in the Southern Ocean, and ocean circulation around Antarctica and globally may also reflect changes in the Antarctic ice sheet and sea ice extent (Stocker and Wright, 1991; Knorr and Lohmann, 2003). In any case it would appear that the paired and phase lagged events are linked responses to single climatic forcing events. The fact that today we see no phase lag in temperature between the two hemispheres is evidence for a new and different kind of forcing, as discussed in more detail later.

While the changes from glacial to interglacial states (Figure 3.13) are to some extent predictable, relying as they do on the Earth's orbital behaviour, other kinds of change are not. These include the abrupt Dansgaard-Oeschger (D/O) cooling events in Greenland and their Antarctic counterparts, which occurred within the last glaciation (Figure 3.13). Equally, millennial to centennial scale variability observed within the past 12 ka in ice cores (see later) is not yet understood to the point of providing a firm basis for predictions of future change. The origins of these various cycles and events most likely relate to changes in global ocean circulation, atmospheric circulation, albedo dynamics, and solar variability. Their persistence through time, strong during glacials and weaker during interglacials (like the Holocene, see later), further suggests that such fluctuations can be expected in the future. Ice core data from the last glacial period in Greenland show that change at that time could proceed rapidly - with several increases of more than 10°C within a decade to possibly one to two years (Figure 3.13); comparable changes in Antarctica were much slower.

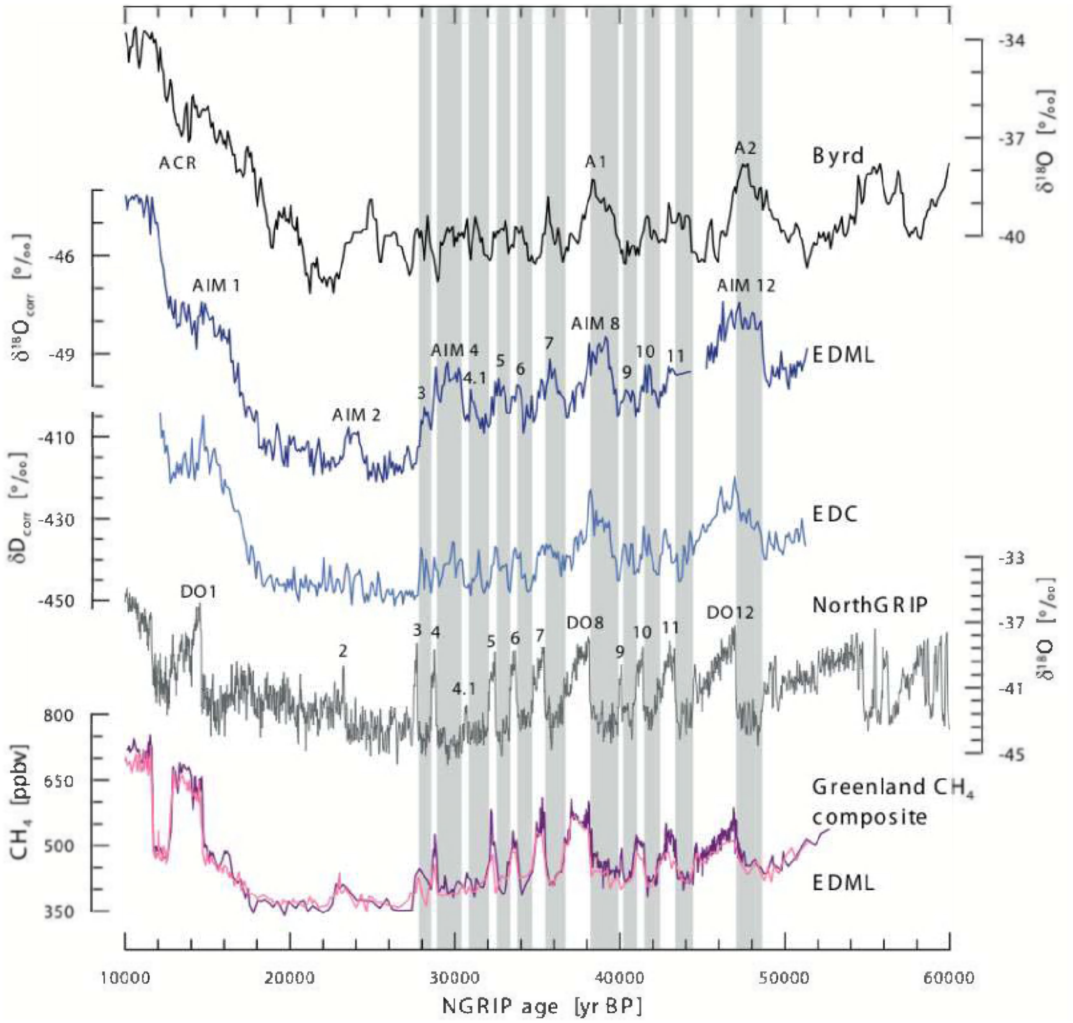


Figure 3.13 Methane (CH_4) synchronization of the ice core records of $\delta^{18}O$ as a proxy for temperature reveals one-to-one association of Antarctic warming (AIM) events with corresponding Greenland cold (stadial) events (D/O) covering the period 10-60 ka ago. EDML = EPICA core from Dronning Maud Land Antarctica, Byrd = core from West Antarctica; EDC = core from East Antarctica; NGRIP = core from North Greenland. Gray bars refer to Greenland stadial periods. Figure modified from EPICA (2006) by H. Fischer (from Mayewski et al., 2009).

3.3.2 The transition to Holocene interglacial conditions: the ice core record

The transition from the LGM (beginning about 21 ka BP) to the present interglacial period (Termination I) was the last major global climate change event. The Northern Hemisphere experienced dramatic changes during the deglaciation starting with an abrupt warming (the Bølling-Allerød) at 14.7 ka B2K (i.e. before AD 2000; an ice core dating convention introduced by Rasmussen (2006) as an alternative to the conventional Before Present (BP) where 'Present' is 1950), followed by a return to colder conditions (the Younger Dryas episode) and a final rapid warming leading to the Holocene at 11.7 ka B2K (Rasmussen et al., 2006). In contrast, long climate records extending back to 800 ka BP, recently obtained from deep Antarctic ice cores, have revealed the exceptional character of this last termination:

neither the Antarctic temperature proxies nor the methane records capture such abrupt changes during the earlier terminations (Spahni et al., 2005; Jouzel et al., 2007).

Because it is the most recent transition between a glacial and an interglacial period, Termination I can be examined with high resolution data and a robust age control (EPICA, 2006; Rasmussen et al., 2006). Ice core records can be synchronised either regionally using aerosol tracers (e.g. dust particles including calcium from continental aerosols, or sulphate from volcanism) (Severi et al., 2007) or globally using well-mixed atmospheric gas records (such as CH_4 or $\delta^{18}O$ of O_2) (Blunier et al., 1997; Morgan et al., 2002; Blunier et al.. 2007). The accuracy of the Greenland ice core layer-counted age scale GICC05 is between 100 and 200 years over the last transition (Rasmussen et al., 2006). The error on the transfer of this age scale to Antarctic ice cores using CH_4 synchronisation is estimated to be at most 250 years for the Younger Dryas period (the abrupt cooling interrupting the Northern Hemisphere deglaciation). Since the pioneer works of the 1960s, Greenland and Antarctic ice core records have been used to determine the magnitude of climatic and environmental changes and the precise sequence of events during the last major climate transition.

Termination I is now documented at a high resolution in more than 10 Antarctic ice cores located in both East and West Antarctica (Figure 3.14). The beginning of the Holocene at ~11.7 ka BP (Rasmussen et al., 2006) can be found at various depths depending mainly on the accumulation rate and the ice flow at different locations: ranging from ~270 m at Vostok to ~1120 m at Law Dome; note that it is found even deeper in Greenland (~1490 m at NorthGRIP, ~1620 m at GRIP). The full length of the transition varies between ~20 m at Taylor Dome or Law Dome, 100 m at Siple Dome, ~135 m at Vostok, ~155 m at EPICA Dome C, ~290 m at Byrd and up to 300 m at EPICA Dronning Maud Land. Climate records of Termination I from different Antarctic ice cores can therefore offer varying temporal resolutions. The records are affected by a strong thinning at coastal locations and may be partly affected by changes in ice sheet elevation and ice origin from upstream areas at inland locations, especially those not located on domes. The quality of the ice core samples may be affected when the transition is located in the brittle zone – cores taken from the range approximately 400 to 1,000 m depth, often described as brittle, can be easily damaged by drilling and handling.

Figure 3.15 shows selected examples of Antarctic ice core records of Termination I. These records offer the potential to compare local (Antarctic site temperature / accumulation), regional (sea salt and marine biogenic sulphur concentrations), hemispheric (concentration of different Antarctic dust fractions) and global (greenhouse gas concentrations in the atmosphere) climate and environmental parameters which are described in more detail below. The stable isotopic composition of ice (δD or $\delta^{18}O$) is classically used to quantify past Antarctic surface air temperature changes. The strong linear relationship observed between snowfall isotopic composition and temperature (Masson-Delmotte et al., 2007) is caused by the progressive cooling and distillation of air masses along their trajectories from oceanic moisture sources to inland Antarctica. General atmospheric circulation models have shown that this relationship remains stable in central Antarctica at the glacial-interglacial scale, leading to uncertainties of 20-30% on reconstructed temperatures (Jouzel et al., 2003). Temperature estimates based on stable isotope records may be biased by deposition effects related to the seasonality of snowfall. Climate models suggest that this effect remains limited inland in Antarctica (Werner et al., 2001). However, past changes in evaporation conditions may affect Antarctic snowfall isotopic composition. The deuterium excess parameter ($d=\delta D-8\delta^{18}O$), mainly affected by a kinetic fractionation effect, has been used to quantify past changes in moisture source and site temperatures more precisely (Stenni et al., 2001; Vimeux et al., 2002). The moisture source temperature is an integrated quantity that may correspond to either changes in sea surface temperature in the dominant evaporation areas or to geographical changes of the main moisture origin with time.

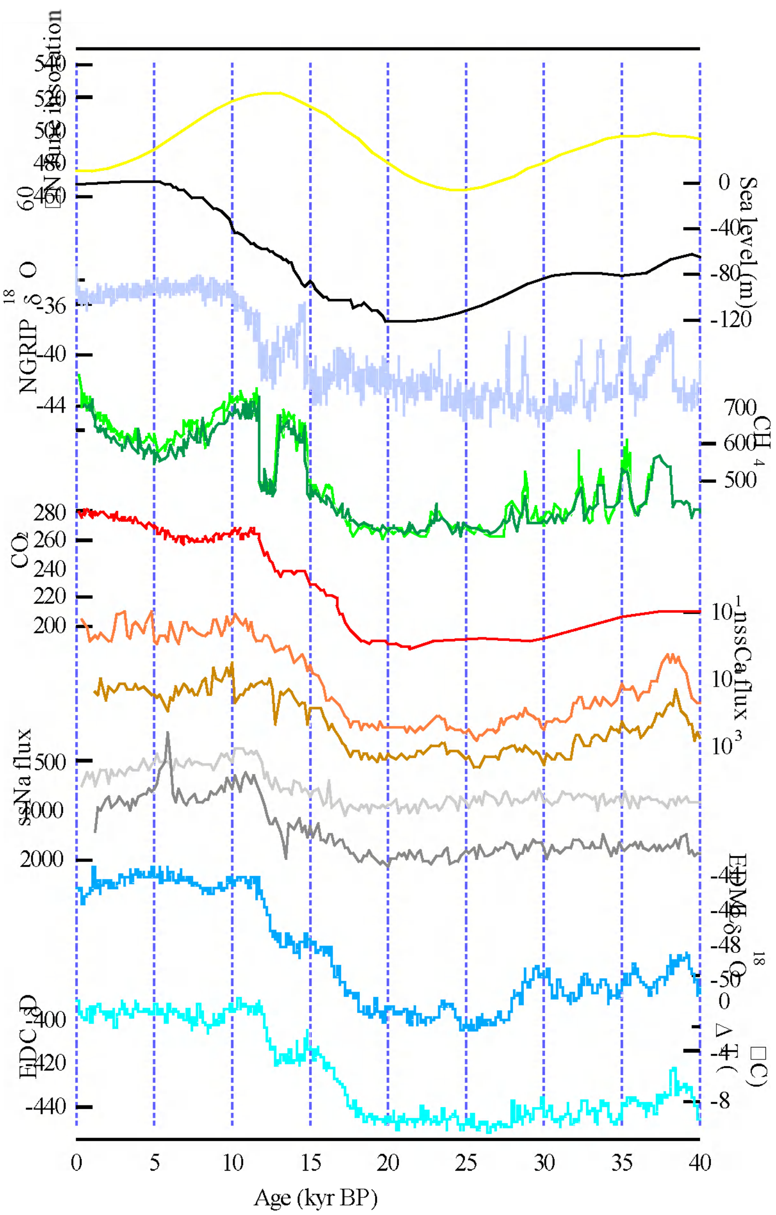


Figure 3.14 Climate and environmental records of Termination I. From top to bottom: yellow, 60°N June insolation (W m^{-2}) (Berger and Loutre, 1991); black, global sea level change (m) (Lambeck and Chappell, 2001; Waelbroeck et al., 2002); blue-grey, Greenland NorthGRIP ice core $\delta^{18}O$ (‰), a proxy of Greenland temperature change (NorthGRIP-community-members, 2004); composite Greenland (light green) and EPICA Dome C (EDC) (dark green) atmospheric methane concentration records (ppbv) (EPICA, 2006) (Loulergue et al., submitted); red, Antarctic EDC and Vostok stacked atmospheric CO_2 concentration record (ppm) (Monnin et al., 2001; Petit et al., 1997); non sea-salt calcium flux, a proxy for continental aerosols, from EPICA Dronning Maud Land (EDML) (brown) and EDC (orange) ice cores (µg/m²yr) (Wolff et al., 2006; Fischer et al., 2007a); sea-salt sodium flux, a proxy for marine aerosols, from EDML (light grey) and EDC (dark grey) (µg/m²yr) (Wolff et al., 2006; Fischer et al., 2007a); blue, EDML $\delta^{18}O$ (‰), a proxy of EDML temperature (EPICA, 2006); and finally, in light blue, EDC δD (‰), a proxy of EDC temperature (Jouzel et al., 2007).

Most Antarctic ice cores exhibit comparable magnitudes of glacial-interglacial water stable isotope changes, with any major departure likely to be due to changes in ice sheet elevation. The magnitude of glacial-interglacial temperature change in central Antarctica has been estimated to be 9°C with an uncertainty of 2°C (Stenni et al., 2001). In central Antarctica, the coherence between the various temperature records based on stable isotopes is very high (within 1°C) (Watanabe et al., 2003). Based on stable isotopes, Antarctic temperatures appear very constant during the period from 19 to 23 ka BP without a clear "glacial maximum" (i.e. temperature minimum). A first Antarctic warming phase associated with Termination 1 is identified as starting at about 19 ka BP in central Antarctica. The warming is interrupted by a 1.5°C cooling from 14 to 12.5 ka BP, the "Antarctic Cold Reversal" (Jouzel et al., 1995). The second warming phase into the early Holocene is the fastest temperature rise detected in the 800 ka EPICA Dome C record, with a pacing of ~4°C per 1,000 years (Masson-Delmotte et al., 2006). Intergovernmental Panel on Climate Change (IPCC) Assessment Report 4 coupled ocean-atmosphere-sea ice climate models run under LGM and present-day conditions tend to underestimate the range of glacial-interglacial temperature changes in central Antarctica (Masson-Delmotte et al., 2006): when considering an unchanged ice sheet elevation, the median simulated glacial-interglacial central Antarctic surface temperature change is 4.5°C. Future climate change in Antarctica, in response to anthropogenic CO_2 emissions, is expected to reach 2°C in 100 years, which is much faster than the fastest change of the last transition (Masson-Delmotte et al., 2006). Other IPCC models described later in this volume predict temperature changes of up to 4-6°C

Over the last transition in Antarctica, warming seems to start first, followed after a few hundred years by an increase in atmospheric CO_2 concentrations (Monnin et al., 2001). The rise in CH_4 also appears to lag the initial rise in Antarctic temperature, but while the significant rapid excursions of CH_4 apparent in the Greenland record (which are related to the abrupt warming into the Bølling-Allerød, the cooling to the Younger Dryas and the warming to the pre-Boreal) are mirrored in the Antarctic CH_4 record, these rapid changes seem to have no direct climate parallel in Antarctic temperature. The abrupt Bølling-Allerød warming recorded in the Northern Hemisphere takes place at the beginning of the Antarctic Cold Reversal, but the precise cross-dating of these events remains disputed (Morgan et al., 2002). Finally, the end of the Younger Dryas occurs when Antarctic temperatures have already reached their early Holocene optimum. The north-south sequence of events is affected by changes in orbital forcing, oceanic and atmospheric circulations, including see-saw effects related to changes in the thermohaline circulation, and biogeochemical cycle feedbacks. The precise role of Northern Hemisphere or Southern Hemisphere insolation changes on the termination onset remains uncertain (Schulz and Zeebe, 2006). Similarly, and despite the strong correlation between CO_2 variations and Antarctic temperature, the mechanisms involved in glacial-interglacial changes in greenhouse gas concentrations and their lags with Antarctic temperature are still a challenge for modellers (Köhler et al., 2005).

Across the transition, the concentration of the majority of aerosol species falls from high glacial levels to lower levels in the Holocene. However, the concentration measured in ice is a combination of changes in both the atmospheric aerosol loading and the accumulation rate. While the use of concentration or flux (expressed in $\mu g\ m^{-2}\ year^{-1}$) is still debated for high accumulation sites, it is generally agreed that for the low accumulation central Antarctic sites, where dry deposition dominates wet deposition, the ice core flux measurement gives a better indication of the initial atmospheric aerosol concentration (Fischer et al., 2007b). At Dome C, the flux of sea-salt sodium and non-sea-salt calcium decrease by factors of 2 and 30, while in Dronning Maud Land, they decrease by factors of 2.6 and 10. The coherence of the sea-salt records from both sites may indicate the waning of the sea ice cover in the Indian and Atlantic Ocean sectors, with the greater magnitude of changes in Dronning Maud Land attributed to a greater retreat in sea ice cover and an additional role of summer sea-ice extent in the Weddell Sea sector. The non-sea-salt Ca is the fraction that is considered representative of mineral dust of continental origin, and the likely source region for both Dronning Maud Land and Dome C is Patagonia (Smith et al., 2003). The higher concentration of dust in Dronning Maud Land is indicative of a greater input of aeolian dust to the Weddell Sea region (Fischer et al., 2007a), presumably as a consequence of the closer proximity to the source region. Models have indicated that there is surprisingly little change in transport strength and atmospheric residence time of atmospherically entrained dust. Change in residence time makes no sense in this case between South America and Dome C during glacial and interglacial periods, so the large reduction in flux during the transition seen at both sites, which began at about 18 ka BP, is thought to be primarily resulting from the changes in the aridity and extent of the dust source region (Wolff et al., 2006; Fischer et al., 2007a). An alternative hypothesis is that changes in the dust flux may be influenced by changes in the position of the polar front and strength as demonstrated by calibration of non sea-salt Ca and westerlies (Yan et al., 2005).

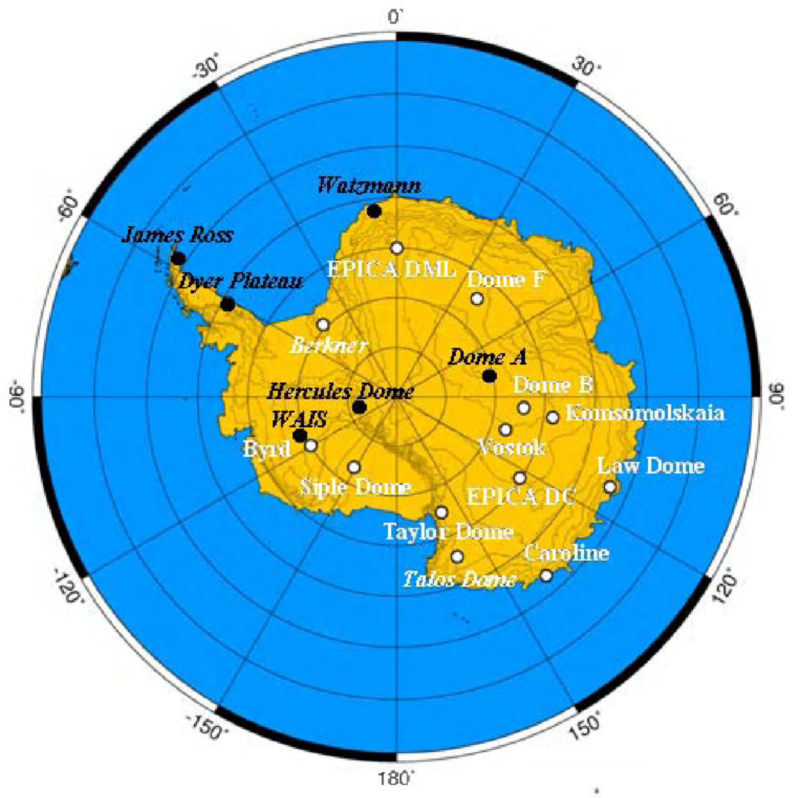


Figure 3.15 Map of Antarctica (created using http://www.aquarius.ifm-geomar.de/omc/) showing some deep ice core sites where climate records have been obtained back to the LGM and beyond: Byrd (Hammer et al., 1994; Blunier et al., 1997), Caroline (Yao et al., 1990), Vostok (Petit et al., 1999), Komsolmolskaia (Nikolaiev et al., 1988), Dome B (Jouzel et al., 1995), Law Dome (Morgan et al., 2002), Taylor Dome (Steig et al., 1998), Dome C with a first deep drilling (Lorius et al., 1979) and the EPICA deep ice core (EPICA, 2004), Dome F (Watanabe et al., 2003), Siple Dome (Brook et al., 2005) and EPICA Dronning Maud Land (EPICA, 2006) . The surface elevation is represented as grey contours (100, 200, 500, and each 1,000 m). Locations of existing deep ice cores going back to the LGM are displayed in white. Names in italics indicate recent ice cores spanning the last termination but not yet published. Further ice cores have recently been drilled at the Detroit Plateau on the northern Antarctic Peninsula and at Titan Dome near the South Pole. Future deep ice core drilling projects are displayed in black.

Although the broad features of the last termination are now well documented in Antarctica, improvements can still be made in the detailed sequence of events with better resolution both temporally (for instance, in the phasing of changes in Antarctic climate, environmental parameters and atmospheric composition) and spatially, particularly in near coastal regions. This calls for further high resolution records of the last deglaciation and a comparison of records from ice drilling sites reflecting different oceanic basins. In addition, for a better understanding of climate dynamics, new ice cores are also needed to improve the knowledge of Antarctic ice sheet dynamics over the past deglaciation. Recently drilled records at Berkner Island and Talos Dome spanning the last deglaciation and beyond, and future new deep ice cores (Figure 3.15) will help to progress the understanding of the climate and biogeochemical cycle responses to climate forcings, the climate variability around Antarctica and its source areas, and the reaction of the Antarctic ice sheet to major climate changes of the past.

3.3.3 Deglaciation of the continental shelf, coastal margin and continental interior

One key to understanding the response of glacial systems to the evolving Antarctic climate during Termination 1 is knowledge of how the ice sheet retreated from the continental shelf, not least because that retreat heralded the arrival of current interglacial conditions. The Antarctic cryosphere consists of three main components, which together contain enough ice to raise sea level 57 metres (Intergovernmental Panel on Climate Change, 2007, Chapter 4). The largest is the East Antarctic Ice Sheet (EAIS), which presently is mostly land-based and is therefore considered to be the most stable. The second largest is the West Antarctic Ice Sheet (WAIS), which is largely a marine-based ice sheet and, as such, is generally considered to be unstable because it is potentially subject to direct melting and grounding line instability by a warming ocean. The third component is the Antarctic Peninsula, which today is characterized by ice caps, outlet glaciers and valley glaciers, but during the LGM was covered by a small ice sheet that extended to the edge of the continental shelf. While the Antarctic Peninsula Ice Sheet (APIS) has, according to some models, been a relatively minor (1-2 m) contributor to post-LGM sea level rise, it has been the most sensitive of the Antarctic ice sheets to climate change and sea level rise (Domack et al., 2003a).

The configuration of the Antarctic Ice Sheet during the LGM (c. 21 ka BP) was modelled in the CLIMAP reconstruction by Stuiver et al. (1981) and later revised by Denton et al. (1991). The grounding line was placed near the continental shelf edge, based on an ice surface profile reconstructed from the Ross Sea (Stuiver et al., 1981). Subsequently, numerous studies have addressed the questions of how large the ice sheets were during the LGM, and when they began their retreat.

There is unambiguous evidence that ice sheets grounded on the continental shelf all around Antarctica during the Late Pleistocene. Sediments from piston cores from all the continental shelves studied to date consist of glacially deposited tills overlain by glacial-marine sediments (Anderson, 1999). Seismic records from these areas show subglacial facies resting on regional glacial erosion surfaces, along with geomorphic features typical of grounded ice sheets. Swath bathymetry surveys of the continental shelf yield spectacular subglacial geomorphic features that have been collectively referred to as the "Death Mask of the Ice Sheet" (Wellner et al., 2006). More detail follows below, based on recent reviews by Anderson (1999); Anderson et al. (2002); and Bentley (1999).

3.3.3.1 East Antarctic Ice Sheet expansion and retreat

To date, results from studies in East Antarctica have yielded mixed results with regard to the size of the ice sheet during the LGM (21 ka BP) and its subsequent retreat history. Several onshore coastal locales apparently experienced only limited or no ice cover during the LGM, including the Bunger Hills (Gore et al., 2001) Larsemann Hills (Burgess et al., 1994; Hodgson et al., 2001), the Lützow-Holm Bay area (Igarashi et al., 1995). Colhoun (1991) summarizes other lines of evidence for a thinner (< 300 m) EAIS in coastal areas than had been previously hypothesized. These results are supported by marine geological data from the eastern Weddell Sea, where glaciomarine sediments that directly overly tills have yielded radiocarbon ages older than 20 Ka, indicating that the ice sheet retreated from the continental shelf prior to the LGM (Anderson and Andrews, 1999; Anderson et al., 2002), and the George V Coast (Presti et al., 2005).

Studies from other parts of the East Antarctic sector show ice sheets grounding on the continental shelf during the LGM, followed by retreat of the ice sheet from the shelf. Off Wilkes Land, continental shelf swath bathymetry data show lineations that extend across the shelf, and grounding zone wedges, that record the retreat of the ice sheet from the shelf (McMullen et al., 2006). Radiocarbon dates of glacial-marine sediment that directly overlies till indicate that the transition from subglacial to glaciomarine sedimentation on this shelf occurred prior to ~ 9 ka BP, and that the retreat of the grounding line to its present coastal position was complete by ~ 2 ka BP (Domack et al., 1989; Domack et al., 1991). These results are consistent with those from the Windmill Islands, west of the George V Coast between 110°E and 111°E, where de-glaciation occurred between 8 and 5 ka BP (Goodwin, 1993).

In Prydz Bay, O'Brien (1994), O'Brien and Harris (1996), O'Brien and Leitchenkov (1997), and Domack et al. (1998) used bottom profiler data, seismic data and sediment cores to reconstruct the Late Pleistocene ice sheet configuration. The reconstruction of Domack et al. (1998) shows that during the LGM the ice sheet was grounded on the shelf, except in the deeper portions of troughs. Radiocarbon dates indicate that the ice sheet retreated from Prydz Bay sometime around 11.5 ka BP. A mid-Holocene re-advance of the Lambert/Amery system occurred between 7 and 4 ka BP (Verleyen et al., 2005).

In the western Ross Sea region, the most widespread high-elevation moraine unit is the Ross Sea Drift; it is perched 240 to 610 m above present sea level (Stuiver et al., 1981). These moraines merge with the inland ice plateau near modern glacier heads, indicating that the inland EAIS stood at about the same elevation as today (Denton et al., 1991). Radiocarbon dates on Late Pleistocene deposits confirm that the LGM occurred here between 21.2 and 17 ka BP (Stuiver et al., 1981), which is consistent with the date we use here for the LGM (21 ka BP) and the initial warming at 19 ka BP. Swath bathymetry records show glacial lineations that extend to the outer continental shelf (Shipp et al., 1999). These observations confirm earlier results from sedimentological and petrographic studies of sediment cores that indicated the EAIS was grounded on the continental shelf. The ice sheet begun its retreat from the shelf shortly after the LGM, and retreated from the outer shelf at a fairly constant rate, with the grounding line being located somewhere near the present ice shelf front by about 7 ka ago (Licht et al., 1996; Domack et al., 1999; Licht et al. 1999). Radiocarbon dates show that the recession of the Holocene grounding line was completed between ~ 7 and 5 ka BP in this region (Denton et al., 1991).

3.3.3.2 West Antarctic Ice Sheet expansion and retreat

Swath bathymetry records from the eastern Ross Sea show glacial lineations that extend to the shelf break and confirm that the WAIS was grounded on the shelf in the past (Mosola and Anderson, 2006). Radiocarbon dates from sediment cores indicate that the ice sheet was grounded there during the LGM. While the resolution of these dates is not sufficient to determine the exact retreat history of the ice sheet from the shelf, the distribution of grounding zone wedges on the shelf indicates that the grounding line retreated in a step-wise fashion.

The extent of the grounded ice sheet on the southern Weddell Sea continental shelf is poorly constrained. Piston cores from the Crary Trough show that an ice sheet was grounded in the trough, but attempts to date the glacial-marine sediments that overly these tills have been hampered by a lack of carbonate material (Bentley and Anderson, 1998). Marine geological studies off Marie Byrd Land, including Pine Island Bay, indicate that the ice sheet extended to the edge of the continental shelf (Evans et al., 2006) and that the ice sheet was in its final phase of retreat from Pine Island Bay by 10 ka cal BP (Lowe and Anderson, 2002).

3.3.3.3 Antarctic Peninsula Ice Sheet Expansion and Retreat

The Antarctic Peninsula Ice Sheet (APIS) advanced to the edge of the continental shelf during the LGM. The expanded ice sheet there was more than double the size of the landmass (Ó Cofaigh et al., 2002; Dowdeswell et al., 2004; Heroy and Anderson, 2005; Ó Cofaigh et al., 2005; Bentley et al., 2006). The retreat of the APIS from the shelf occurred progressively from the outer, middle, and inner continental shelf regions, as well as progressively from the north to the south. Retreat began on the outer shelf of the northern peninsula by ~18,000 cal yr BP and continued southward by ~14 ka cal yr BP on the outer shelf off Marguerite Bay (Heroy and Anderson, 2005). Steps in the data occur at ~14 and possibly 11 ka cal yr BP, coincidental with global melt water pulses MWP 1a and 1b, and show that rapidly rising sea level at those times may have destabilised the marine ice sheet (Heroy and Anderson, 2005).

By ~10 ka cal BP the APIS grounding line reached the inner shelf. From that time on the retreat of the ice sheet was diachronous. Again, this is not unexpected as the highly irregular bedrock relief on the inner shelf was undoubtedly an important factor regulating grounding line retreat (Heroy and Anderson, 2005). Also, as the ice sheet retreated the glacial system evolved into discrete outlet and valley glaciers that responded differently to different forcing mechanisms, a result of their different drainage basin size, elevation, and climate setting.

While the retreat history of the ice sheet was largely in phase with Northern Hemisphere deglaciation, it is not clear if climate warming or other factors caused the ice sheet to retreat. Marine geological data provide clear evidence that the expanded ice sheet was drained by large ice streams, at least during the final stages of the advance. These ice streams would have contributed to thinning of the ice sheet, thereby rendering it more sensitive to global sea level rise. Hence, the retreat of the Antarctic Ice Sheet was probably a response to a combination of climate warming driven by insolation and rising CO_2, changing sea level, and other dynamical processes.

3.3.4 Antarctic deglaciation and its impact on global sea level

As the Antarctic Ice Sheet retreated its meltwaters contributed to an increase in global sea level. This process continues today and is set to accelerate under predicted increases in temperature. However, in 2007 the Intergovernmental Panel on Climate Change noted continued uncertainty about the exact future contribution of the Earth's major ice sheets to sea level (Intergovernmental Panel on Climate Change, 2007). The Antarctic Ice Sheet has least evidence to constrain its past volumes and it is not known if in future net loss of ice mass could occur with dynamical ice discharge, rather than surface melting, dominating the ice sheet mass balance. This is important, as better understanding of how global ice sheets have contributed to sea level across Termination I and into the Holocene is one key to predicting how relatively subtle changes in these ice sheets could influence sea level in the future in a warming world. This is especially relevant as evidence from both Antarctica and

Greenland show that in the last Eemian (MIS5e) interglacial when temperatures were between 2-5°C higher global sea levels were 4-6 m higher than today's (Jansen et al., 2007). When global continental ice sheets grew before the LGM, global sea level was drawn down (Yokoyama et al., 2001). At the end of the last glacial, deglaciation caused sea levels to rise again over a period spanning more than 10 ka. The conventional estimate is that this sea level rise was 120 m (Peltier, 2002), but other evidence suggests that it may have been as much as 130-135 m (Yokoyama et al., 2001). A hundred metres or so was accounted for by the melting of the Northern Hemisphere ice sheets, as indicated by ice sheet models and geological data (Clark et al., 2002). This left the Antarctic Ice Sheet, together with other smaller ice sheets, to supply the rest of the water. At present there are relatively few geological constraints on the extent and thickness of the Antarctic Ice Sheet during and after the LGM and its contribution to global sea level. Most models predict the Antarctic Ice Sheet to have contributed c. 20 m to global sea level rise with estimates ranging from as much as 37 m (Nakada and Lambeck, 1988) to 6-13 m (Bentley, 1999). More recently, an improved three-dimensional thermomechanical model has suggested a contribution of 14-18 m (Huybrechts, 2002). This variability is significant, because when the calculated contributions of the LGM global ice sheets to sea level rise are added together 20% of the water cannot be attributed to any known former ice mass. This is known as the 'missing water' (Andrews, 1992). Another uncertainty concerns the ice masses that contributed to meltwater pulse 1a (MWP 1a), a c. 25 m jump in global sea level dated to c. 14.2 ka BP when peak rates of sea level rise potentially exceeded 50 mm/year, equivalent to the addition of 1.5 to 3 Greenland Ice Sheets to the oceans during a period of one to five centuries (Bard et al., 1996). This rapid meltwater event has been variously attributed via modelling studies to the Antarctic Ice Sheet (Clark et al., 2002; Weaver et al., 2003), or the Laurentide and Fennoscandanavian Ice Sheets (Peltier, 2005). Geological evidence does not rule out a contribution from Antarctica (Verleyen et al., 2005; Bassett et al., 2007). Resolving the debates about the Antarctic contribution to global sea level rise and whether or not the Antarctic contributed to Meltwater Pulse 1a has been hampered by a lack of geological evidence or observational control on Antarctic ice thickness and maximum marine limits during the LGM (Huybrechts, 2002).

One method of providing this observational or geological control is to study the history of relative sea level change. This provides one of the primary data sets for constraining the geometry, volume and melt history of past and present glacial masses. These data can be used to determine the loading and rebound of the continental plates under the increasing weight of a growing ice sheet and the decreasing mass of melting ice sheets, and can be used by modellers to indirectly infer former regional ice sheet thickness. Relative sea level (RSL) is the movement of the shoreline relative to the sea. It is influenced by changes in both global ocean volume (eustatic changes in sea level) and local vertical movements of the land (isostatic changes) that are caused by the rebound of the Earth's crust following the removal of the overlying ice mass. Sometimes, the interaction of these two factors can result in complex curves of RSL change over time, and in other cases they can be influenced by tectonic processes. Typically, these curves are determined for the period since the LGM. Depending on the location of a particular site with respect to former ice masses a number of different shapes of curves can result. For example, in the equatorial regions far from the former ice sheets the RSL curve is dominated by the eustatic component of meltwater returning to the oceans and so shows relative sea level rise. These regions are known as far-field sites. In contrast, at a (near-field) site close to the centre of a former ice sheet mass the RSL change will be dominated by the isostatic component as the crust is unloaded and the land rebounds. If significant ice is removed, and quickly enough, then this rebound is sufficient to outpace the eustatic contribution and so there is a continuous fall in relative sea level like that seen in areas such as Hudson Bay or Sweden today.

Determining RSL change over the Holocene is important because the RSL record provides information on former ice thicknesses in specific regions and on the timing of deglaciation. Crucially, RSL can be determined independently of other glacial geological techniques in two main ways: by using techniques relying on sampling raised marine landforms (beaches, deltas etc), and techniques relying on isolation basins. In the former case, raised marine features are sampled for organic material such as shells, seal skin, penguin remains, or whalebone that can provide an age estimate or constraining date for the age of the beach. For example, whalebone or shells are usually deposited at or just below sea level but over millennia can be reworked into lower beaches and so can only be used as maximum estimates for the age of the beach in which they are found. In the case of isolation basins, cores are taken from lakes near sea level along the coastal margin. Prior to deglaciation, and depending on their altitude, these lakes may have been former marine inlets or basins. As the ice melted and the crust rebounded they became isolated and transformed into freshwater lakes. The sediment in the lakes records this transition, which can be dated to determine when the basin was at sea level. This approach can be used for a staircase of lakes at different altitudes, allowing reconstruction of an RSL curve that is more precise than can be achieved using the raised beach approach.

RSL curves have been developed for many formerly glaciated regions, but those in Antarctica have required the development of some innovative dating techniques. In particular penguin remains and sealskin have been important for constraining the ages of beaches (Baroni and Orombelli, 1994' Hall et al., 2003; Bentley et al., 2005a).

3.3.4.1 Spatial coverage of relative sea level (RSL) curves

To date, almost all Antarctic RSL data come from three areas: the Ross Sea sector, the Antarctic Peninsula, and East Antarctic coastal ice-free oases located between 75-105° E and 30-45° E. There are three main reasons for this restricted distribution. First, much of the Antarctic coast (> 95%) is fronted by glacier ice or ice shelves and so there are few sites where RSL change is recorded in coastal sediments. Second, logistical and cost considerations mean that some locations have not yet been sampled for RSL work. Finally, although there is a need for tight dating of RSL data, some sites with raised marine deposits yield little or no organic material suitable for radiocarbon dating.

In the Ross Sea area, well-dated RSL curves have been developed from the Scott Coast and Victoria Land Coast (Hall et al., 2003) (Figure 3.16). The Ross Sea curves show exponential RSL fall owing to deglaciation of the coast by the retreating West Antarctic Ice Sheet (Conway et al., 1999). The amount of RSL fall (~ 30 m) is low compared to other formerly glaciated regions (e.g. 150 m for the Canadian Arctic). This is because much of the rebound occurred as restrained rebound, when the coast was still covered by thinning ice or an ice shelf and so no beaches were formed, and the Antarctic ice sheet still continues to provide a substantial load on the crust, unlike the near-full deglaciation of the Canadian Arctic.

Zwartz et al. (1998) and Verleyen et al. (2005) determined RSL curves for the last 15 ka for the Vestfold Hills and Larsemann Hills in East Antarctica by dating a series of marine-lacustrine transitions in isolation basins. The curve is typical of sites close to the margin of formerly more expanded ice sheets in that it shows a RSL rise to a maximum, or 'highstand', followed by gradual fall to the present day, athough in the Larsemann Hills there is a decline in the rate of isostatic uplift between c. 7250–6950 and 2847–2509 cal yr BP, due to a mid-Holocene glacier readvance. The RSL highstand in the Larsemann Hills reached approximately +8 m between c. 7570–7270 and 7250–6950 cal yr BP. This shape occurs because the amount of glacio-isostatic rebound at the East Antarctic ice margin in the Early Holocene was sufficiently small that eustatic sea level rise was able to outpace it. When the northern ice sheets had melted and eustatic rise slowed then isostatic rebound outpaced the eustatic signal, leading to a slow RSL fall.

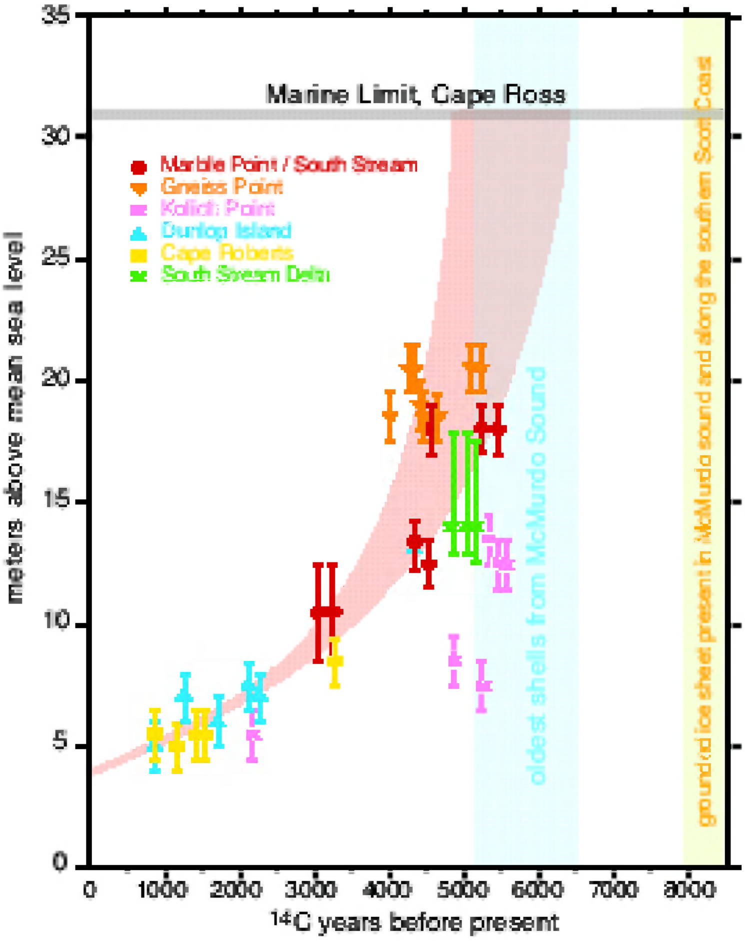


Figure 3.16 Relative sea level curve for the Ross Sea based on radiocarbon dates of elephant seal skin along with shells and penguin remains and guano, in raised beaches on the southern Scott Coast and Terra Nova Bay. This gives valuable information regarding the timing and magnitude of deglaciation (Hall et al., 2003).

Bentley et al. (2005a) developed the first RSL curves for the Antarctic Peninsula using a combination of morphological remains and isolation basin techniques. In the north, a preliminary RSL curve from the South Shetland Islands shows a similar shape to that from the Larsemann Hills (Verleyen et al., 2005). Here, the highstand reaches ~+14.5-16 m at about 4,000 ^{14}C yr BP. In Marguerite Bay (south-central Peninsula) the RSL curve shows exponential RSL fall since deglaciation at ~ 9,000 ^{14}C yrs BP. The curve reflects the removal of a large grounded ice sheet that extended across Marguerite Bay to close to the shelf edge (Ó Cofaigh et al., 2002). Both sites are currently the subject of further studies.

Because they reflect former ice volume changes, RSL curves are central to a number of ongoing important debates. These include determining the size of the Antarctic Ice Sheets at the LGM (Bentley, 1999); testing the hypothesis that Meltwater Pulse 1A was wholly or partly derived from Antarctica (Bassett et al., 2007); and determining the timing and style of deglaciation around the ice sheet margin (Conway et al., 1999). One of the most important applications of existing and new RSL curves is to constrain coupled models of ice sheet volume change and glacial-isostatic adjustment. These models attempt to infer reasonable ice sheet histories by testing which histories can satisfy the constraints provided by far-field and near-field RSL datasets. Initial modelling studies have yielded promising results on former ice sheet volume (Nakada et al., 2000; Ivins and James, 2005) and presence/absence of meltwater pulses (Bassett et al., 2007), but the limited distribution of Antarctic RSL sites means that the models are still not as well constrained as for other continents.

Work on developing new RSL curves, particularly using isolation basin techniques, is ongoing at a number of sites. The long-term aim is to improve both the precision and spatial coverage. There are a small number of regions where there are sites suitable for developing RSL curves. They are generally in logistically challenging locations, but the need to improve the spatial coverage of sites in order to better constrain models provides a strong driver for their development.

3.3.5 Sea ice and climate

Sea ice is a crucial part of the Earth's climate system and its extent can be inferred for up to the past 220 ka from marine sediments (e.g. Crosta et al., 2004) and for shorter periods from some ice cores (Curran et al., 2003; Abram et al., 2007a). Sea ice isolates the polar ocean from the atmosphere and inhibits the exchange of heat and moisture. The formation and melting of sea ice also changes the salinity of the cool surface waters of the polar oceans, so changing their density. This affects the global thermohaline circulation, which, in turn, influences climate around the globe. The high albedo of sea ice also means that it efficiently reflects incoming solar radiation, a process that acts as a positive climate feedback to amplify climate change. Growing sea ice cools the planet; decreasing sea ice warms the planet by exposing more dark and less reflective sea, so enabling more heat to be absorbed than reflected. At the same time decreasing sea ice may also enhance outgassing of CO_2 from the Southern Ocean – see Stephens and Keeling (2000) – which may help to explain the close correlation of sea ice extent with the pattern of CO_2 in ice cores (see Crosta et al., 2004).

3.3.5.1 Sea ice extent

Diatom assemblages from some marine sediment cores can be used to indicate whether or not the sea at the core locations was covered with sea ice in the past (Crosta et al., 2004; Justwan and Koça, 2008). Recently, a novel proxy for sea ice studies was established, the so-called IP25 (Ice Proxy with 25 carbon atoms) produced by diatoms living in the sea ice (Belt et al., 2007; Belt et al., 2008). In the Arctic, sediments containing IP25 have already been dated using radiocarbon methods to at least 9,000 yr (Belt et al., 2007). Given reliable proxies and enough cores it is possible to approximately map the position of the sea ice edge back through time and to determine its summer and winter limits. This has been done for the Southern Ocean, for example at the LGM by Gersonde et al. (2005) see Figure 3.17. From this exercise it can be seen that at the LGM sea ice was double its present extent in winter; LGM sea ice cover was similarly double its present extent in summer due to greater extent off the Weddell Sea and possibly the Ross Sea (Gersonde et al., 2005). This value is however far less than the 5-fold estimate of CLIMAP (1981).

According to Gersonde et al. (2005) the LGM sea ice edge in the Atlantic and Indian sectors reached close to 47°S (Figure 3.17), which is in the modern Polar Frontal Zone and close to the Subantarctic Front that today defines the northern edge of the Antarctic Circumpolar Current. More data are needed to define the sea ice edge in the Pacific sector (Gersonde et al., 2005). Data for the extent of summer sea ice suggest that it was at least as extensive as at present, but with a larger than present summer sea ice extent in the Weddell Sea area. The related sea surface temperature calculations show that the Polar Front in the Atlantic, Indian and Pacific sectors would have shifted to the North during the LGM by around 4°, 5–10°, and 2–3° in latitude, respectively, compared with their present location. In the Atlantic and Indian sector, the Subantarctic Front would have shifted by around 4–5° and 4–10° in latitude, respectively. The Subtropical Front displacement would have been minor, by around 2–3° and 5° in latitude in the Atlantic and Indian sector. The net effect would be to steepen the oceanographic fronts in the Polar Frontal Zone, thereby speeding current flow in the jets along those fronts. A northerly displacement of the wind field is also implied.

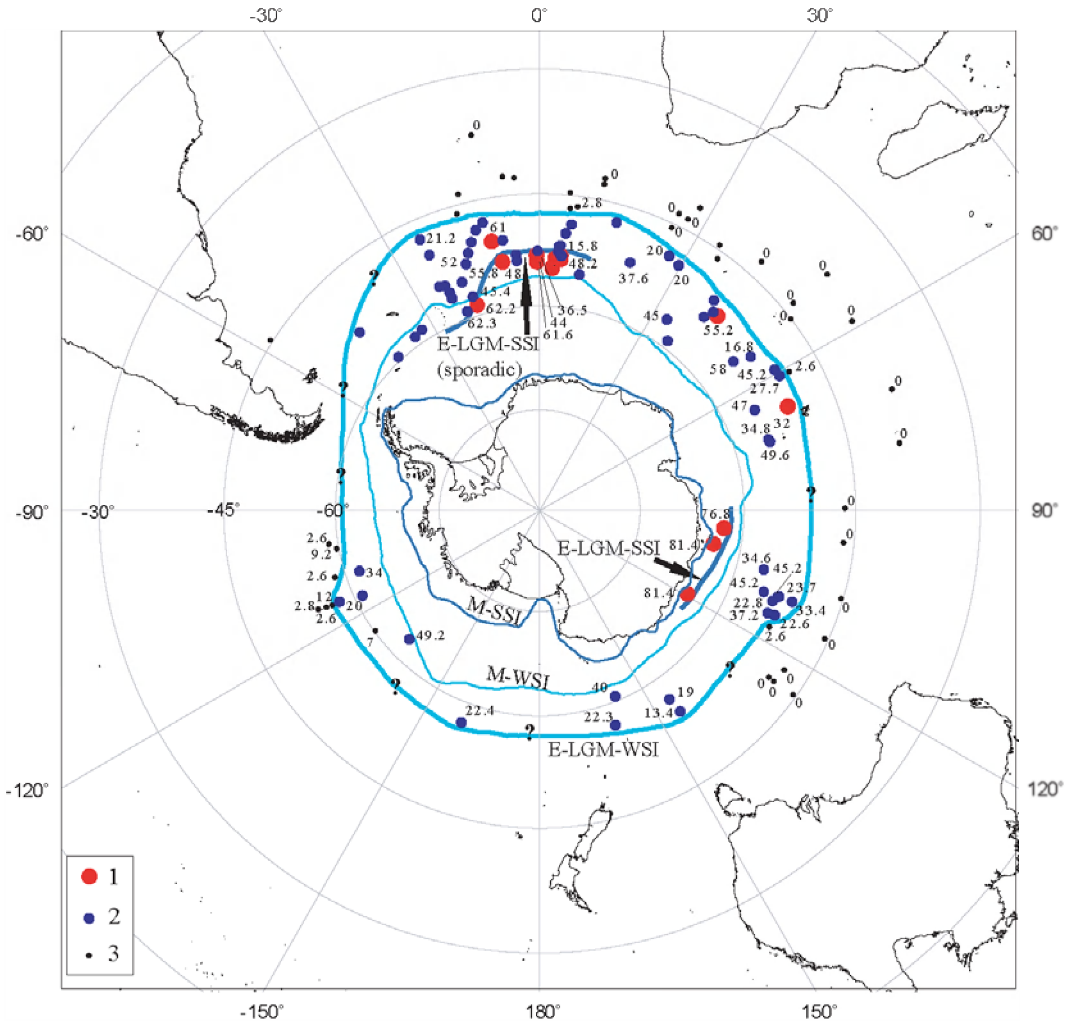


Figure 3.17 Sea ice distribution at the Southern Ocean EPILOG-LGM (E-LGM) time slice. E-LGM-winter sea ice (E-LGM-WSI) indicates maximum extent of winter sea ice (September concentration >15%). Modern winter sea ice (M-WSI) shows extent of >15% September sea ice concentration according to Comiso et al. (2003). Values indicate estimated winter (September) sea ice concentration in percent derived with Modern Analog Techniques and Generalized Additive Models. Signature legend: (1) concomitant occurrence of cold-water indicator diatom *F. obliquecostata* (>1% of diatom assemblage) and summer sea ice (February concentration >0%) interpreted to represent sporadic occurrence of ELGM summer sea ice; (2) presence of WSI (September concentration >15%, diatom WSI indicators >1%); (3) no WSI (September concentration <15%, diatom WSI indicators <1%) (Gersonde et al., 2005)

Fluctuations in the position of the sea ice edge through Late Quaternary and Holocene time can be compared with insolation and sea surface temperature data over about the past 220 ka (Crosta et al., 2004) (Figure 3.18).

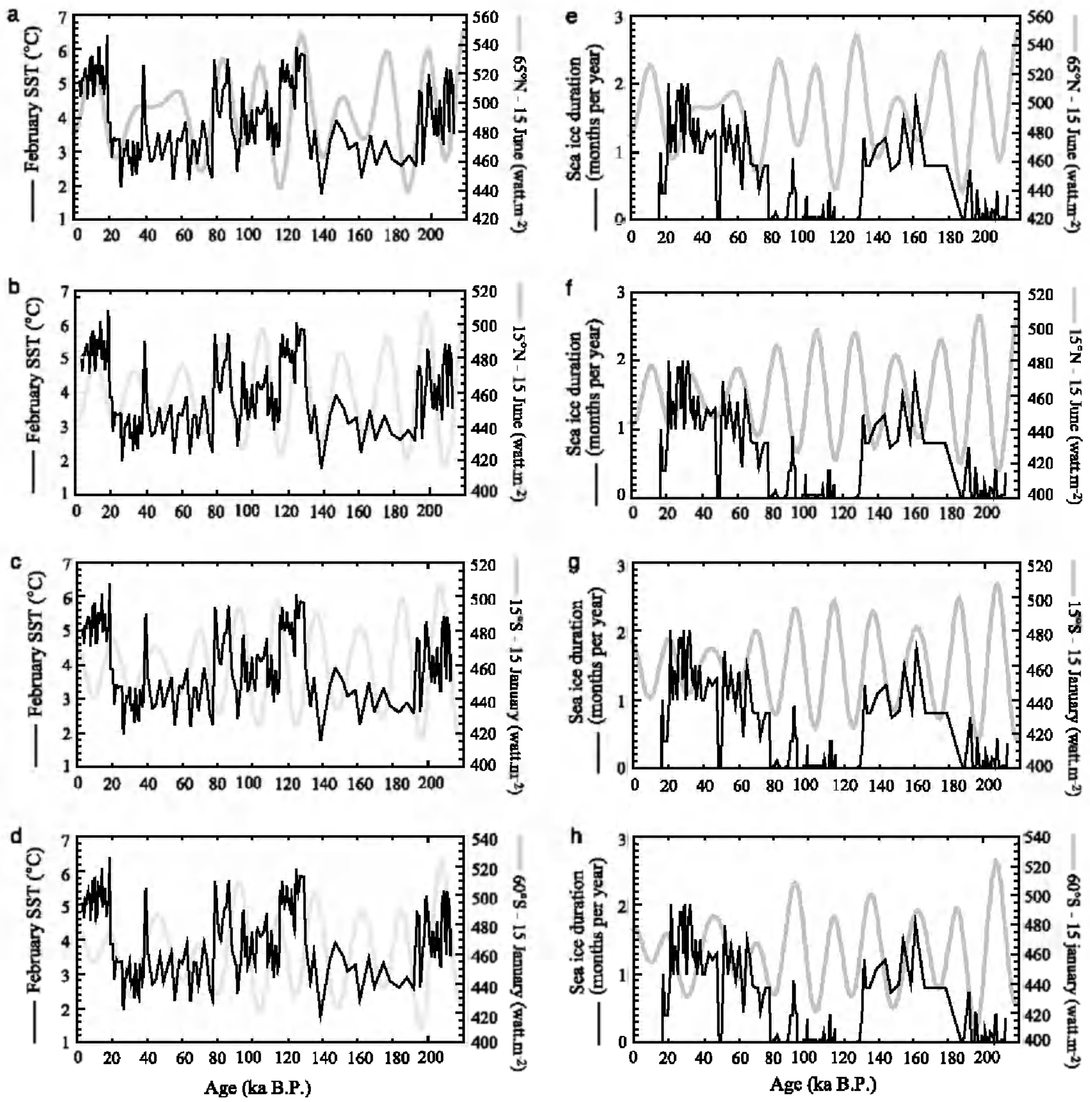


Figure 3.18 Comparison of February SSTs and sea ice duration in core SO136-111 with insolation curves. Parameters estimated by Modern Analog Technique 5201/31 are represented by black lines; insolation curves are represented by grey lines. (a) SSTs versus insolation at 65°N for the 15th of June, (b) SSTs versus insolation at 15°N for the 15th of June, (c) SSTs versus insolation at 15°S for the 15th of January, (d) SSTs versus insolation at 60°S for the 15th of January, (e) sea ice cover versus insolation at 65°N for the 15th of June, (f) sea ice cover versus insolation at 15°N for the 15th of June, (g) sea ice cover versus insolation at 15°S for the 15th of January, (h) sea ice cover versus insolation at 60°S for the 15th of January (Crosta et al., 2004).

Past sea ice extent can also be inferred from assemblages of planktonic organisms or produced biomarkers (such as IP25) that exhibit a relationship to the temperature of the surface waters in which they live. Knowing this relationship at the present day, we can use it to transform assemblage data and IP25 abundances from the recent past into estimates of seawater temperature (Crosta et al., 2004; Massé et al., 2008) and compare these with oxygen isotope records where these are available. In the Southern Ocean, the data indicate that for core sites well offshore, sea ice formation lags temperature decline at the onset of glaciations by about 1,000 years (probably reflecting the time for Southern Ocean temperatures to become cold enough for sea ice to form at these more northerly sites), but that warming and sea ice retreat are simultaneous during glacial terminations (the transitions from glacial to interglacial) (Crosta et al., 2004).

3.4 The Holocene

In this section we review the climate changes that have occurred in the present Holocene interglacial. In many long ice core records this period may superficially appear to be one of relative climate and environmental stability. However, closer inspection of these and other records reveal a complex of changes resulting from the interplay between the ice sheet – ocean – sea ice – atmosphere system to past and present climate forcing. In this section we first examine evidence from the ice core record looking particularly at the phasing of these climate and environmental changes on regional to hemispheric timescales. Second we examine changes in sea ice extent through the Holocene and how these have interacted with changing climate and ocean circulation. Finally we look at the regional patterns of Holocene climate and environmental change experienced in the major regions of Antarctica which are the result of both continental and local forcing mechanisms.

3.4.1 Holocene climate changes: regional to hemispheric perspectives

Over the past ~11.7 ka there have been several abrupt changes in Antarctic climate. These are either related to, or superimposed on, the dynamic response of the ice sheet to past, longer term forcing. As a consequence the current configuration of the Antarctic ice sheet is the result of a multi-millennial scale lagged response to climate forcing. As an example, 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 close to the edge of the current Ross Ice Shelf 7-9 ka ago (Conway et al., 1999). Based on evidence developed from a synthesis of ice core isotopic records this massive retreat was preceded by an early Holocene climatic optimum between 11.5 and 9 ka ago (Masson et al., 2000) which was warmer than today and resulted in the rapid loss of some Antarctic Peninsula ice shelves (Hodgson et al., 2006b). These and other Antarctic data confirm that although the global climate of the Holocene superficially appears to have been relatively stable, there has been sufficient variability during the last ~9 ka to cause major changes to Antarctic ecosystems. This natural variability must be taken into account in understanding modern climate and the potential for future climate change.

A comparison of the behaviour of climate proxies measured in selected ice cores from GISP2 (Greenland) and Siple Dome (West Antarctic) (Figure 3.19) reveals periods of notable change in climate with some coincidence in both Northern and Southern Hemisphere polar regions during the Holocene. From this comparison several general conclusions can be drawn concerning the phasing and the magnitude of changes in atmospheric circulation and temperature between Northern and Southern Hemisphere polar latitudes. These conclusions are relevant to understanding not only the forcing of climate change over the polar regions, but also the implications of change over the polar regions for climate at the global scale.

With respect to changes in atmospheric circulation, Figure 3.19 shows that:

(1) North Atlantic climate (GISP2) displays more frequent and larger shifts in atmospheric circulation than does the Antarctic (Siple Dome). This difference is similar to that seen between Greenland and Antarctica in millennial scale events from glacial age ice core records (EPICA, 2006);
(2) North Atlantic atmospheric circulation (GISP2) generally displays more abrupt onset and decay of multi-centennial scale events than does the Antarctic (Siple Dome);
(3) GISP2 and Siple Dome ice core proxies for Northern and Southern Hemisphere westerlies show considerable similarity in event timing, frequency, and onset style suggestive of some common control; and
(4) the most dramatic changes in atmospheric circulation during the Holocene noted in the Antarctic are (i) the abrupt weakening of the Southern Hemisphere westerlies ~5.2 ka ago, and (ii) intensification of the westerlies and the Amundsen Sea Low starting ~1.2-1 ka ago.

With respect to temperature, Figure 3.19 shows that:

(1) the prominent temperature drop ~8,200 years ago over the North Atlantic, noted in the GISP2 isotope proxy for temperature, is missing in the Siple Dome isotope temperature proxy series, although it is suggested in a composite isotope record covering East Antarctica (Masson et al., 2000);
(2) Siple Dome isotopic temperature reconstructions reveal notable cooling ~6.4-6.2 ka ago followed by relatively milder temperatures over East Antarctica 6-3 ka ago (Masson et al., 2000), lasting until ~1,200 years ago in the Siple Dome area; and
(3) both Siple Dome and GISP2 proxies for temperature show a decline in temperature starting ~1.2-1 ka ago, followed by warming in the last few decades.

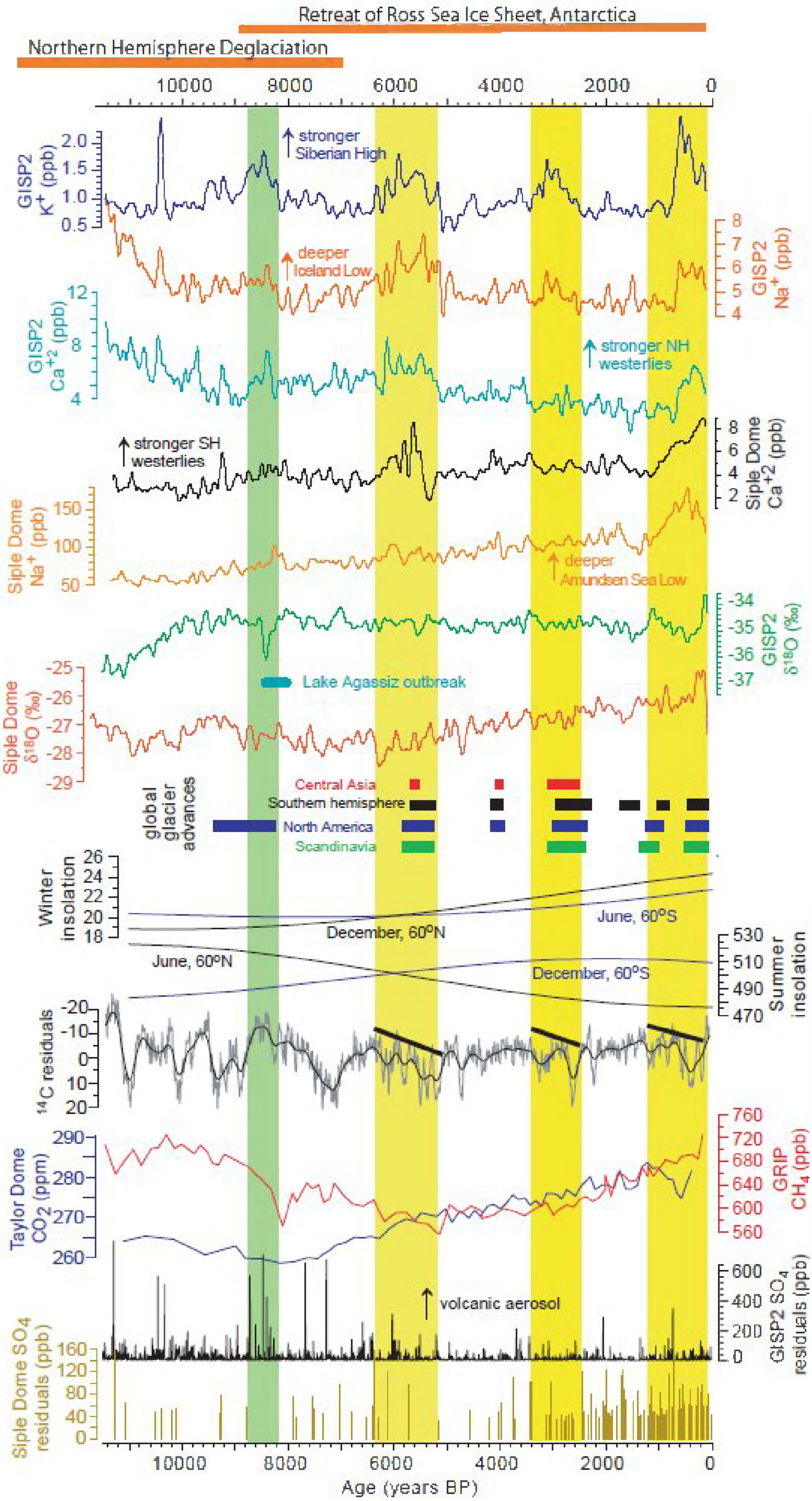
The abrupt climate change event commencing ~1,200-1,000 years ago is the most significant Antarctic climate event of the last ~5,000 years (Mayewski and Maasch, 2006). Its onset is characterized by strengthening of the Amundsen Sea Low and the Southern Hemisphere westerlies, with cooling both at Siple Dome (3.19) and in the East Antarctic composite isotope record (Masson et al., 2000). Consistent with this picture, a comparison of reconstructions of Southern Hemisphere temperature (Mann and Jones, 2003) and ice core proxies for atmospheric circulation covering the last 2,000 years suggests that in general 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).

Several interactive factors provide the forcing for decadal to centennial scale Holocene age climate events. These include variations in insolation, caused by variations in the Earth's orbital properties and in solar output, and in the heating of the atmosphere by greenhouse gases or its cooling by volcanic aerosols. A number of key climate events and their possible forcing are displayed in Figure 3.19. The earliest is intensification of atmospheric circulation in the Northern Hemisphere (stronger Siberian High and westerlies, deeper Icelandic Low) and in the Southern Hemisphere (stronger westerlies and deeper Amundsen Sea Low) ~ 8,200 years ago. This is associated with the cooling of East Antarctica (Masson et al., 2000), and with a drop in CH_4, a long-term decline in CO_2, and an increase in solar output based on the

$\delta^{14}C$ proxy for solar variability. The latter may have led to increased melting with consequent decrease in North Atlantic salinity and decrease in thermohaline circulation in the North Atlantic. Second, is intensification of the Southern Hemisphere westerlies ~6,400-5,200 years ago, associated with cooling ~6,400 years ago. This change is also associated with a crossover in the trend of insolation, a drop in CH_4, a rise in CO_2, a decrease in solar variability, and collapse of the Ross Sea ice sheet (Conway et al., 1999). These changes are coincident with climate change as far north as the Equator (Stager and Mayewski, 1997). Thirdly, is intensification of atmospheric circulation commencing ~1,200-1,000 years ago and lasting to the present, 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 variability, a drop in CO_2, and increased frequency of volcanic source sulfate aerosols over Antarctica. A satisfactory explanation for the forcing of these Holocene Antarctic climate changes remains elusive, though the link to variations in solar output is suggestive.

Evidence of a link to solar radiation comes from a more detailed examination of forcing over the last 2,000 years using the ice cores and other palaeoclimate records. This supports the close association in timing between changes in atmospheric circulation and solar energy output (Maasch et al., 2005). 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 solar variability (Mayewski and Maasch, 2006). 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 AD 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.

Over the last 700 years, evidence for abrupt climate change has been examined using ice core records from East Antarctica (Law Dome) and West Antarctica (Siple Dome). These selected studies reveal that these two regions have operated inversely with respect to temperature and to the strength of atmospheric circulation on multi-decadal to centennial scales (Figure 3.20) (Mayewski et al., 2004a). The exception is a climate change event commencing ~AD 1700 and ending by ~AD 1850, during which circulation and temperature acted synchronously. This cooling period is coincident with an increase in the frequency of penetration of El Niño events assessed using a South Pole ice core proxy (Meyerson et al., 2002) and with an increase in solar 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 Siple Dome ice core record (Mayewski et al., 2005). The close of this event is coincident with a major transition from zonal to mixed flow in the North Pacific (Fisher et al., 2004), possibly suggesting a global scale association between Antarctic and North Pacific climate for this event.



Retreat of Ross Sea 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
GISP2 δ18O (‰)
Lake Agassiz outbreak
Siple Dome δ18O (‰)
global glacier advances
Central Asia
Southern hemisphere
North America
Scandinavia
Winter insolation
June, 60°S
December, 60°N
June, 60°N
December, 60°S
Summer insolation
14C residuals
Taylor Dome CO2 (ppm)
GRIP CH4 (ppb)
volcanic aerosol
GISP2 SO4 residuals (ppb)
Siple Dome SO4 residuals (ppb)
Age (years BP)

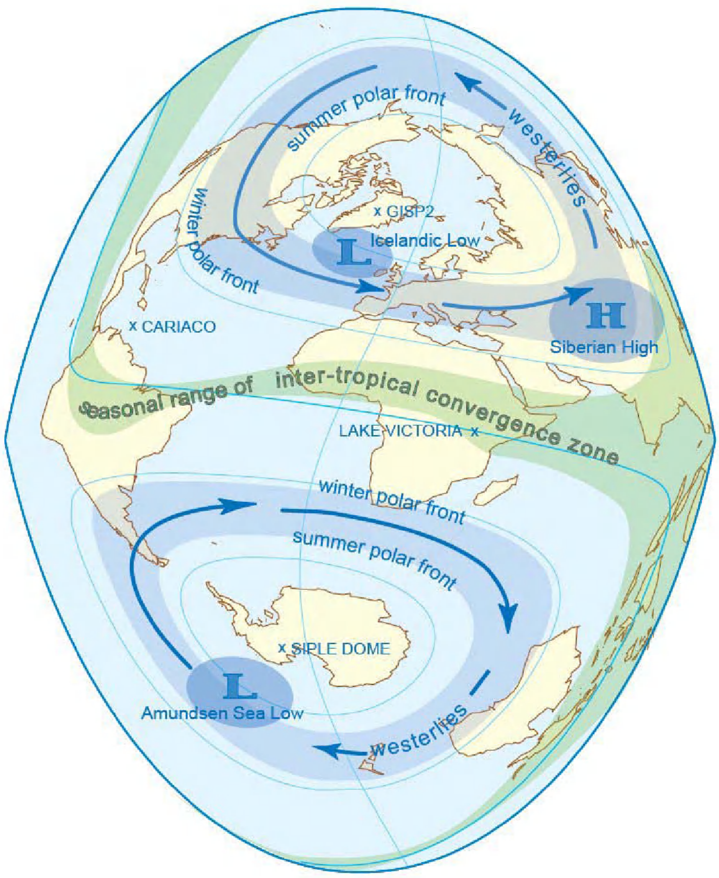


Figure 3.19 Examination of potential controls on, and sequence of, Antarctic Holocene climate change compared with Greenland climate change using 200 year gaussian smoothed data from the following ice cores (top to bottom): GISP2 (Greenland) ice core, K^+ proxy for the Siberian High (Meeker and Mayewski, 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); timing of the Lake Agassiz outbreak that may have initiated Northern Hemisphere cooling at ~8,200 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); 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 (Mayewski et al., 1981) and retreat of Ross Sea Ice Sheet (Conway et al., 1999). Figure modified from Mayewski et al. (2004b, 2005). Green bar denotes the 8,800-8,200 year ago event seen in many globally distributed records associated with a negative $\Delta^{14}C$ residual (Mayewski et al., 2004a). Yellow denotes 6,400-5,200, 3,400-2,400, and since 1,200 year ago events seen in many globally distributed records associated with positive $\Delta^{14}C$ residuals (Mayewski et al., 2004b). Map inset shows location of GISP2, Siple Dome, Icelandic Low, Siberian High, Amundsen Sea Low, Intertropical Convergence Zone, and westerlies in both hemispheres.

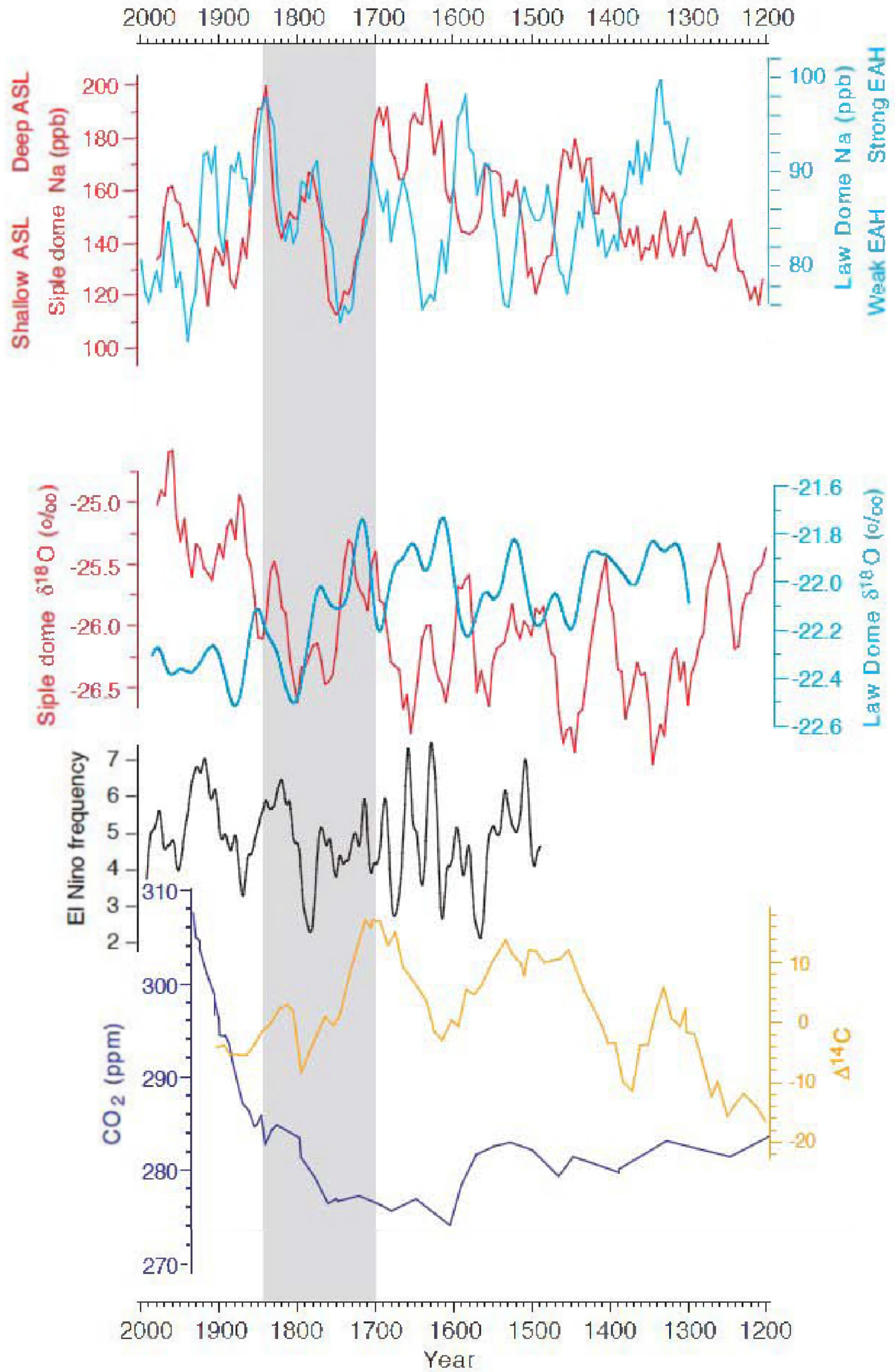


Figure 3.20 25 year running mean of SD (Siple Dome (red)) and DSS (Law Dome (blue) Na^+ (ppb) used as a proxy for the ASL (Amundsen Sea Low) and EAH (East Antarctic High), respectively, with estimated sea level pressure developed from calibration with the instrumental and NCEP reanalysis (based on Kreutz et al., 2000; Souney et al., 2002). Twenty five year running mean SD (red) and DSS (blue) $\delta^{18}O$ (°/oo) 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 (black) based on calibration between the historical El Niño frequency record (Quinn et al., 1987; Quinn and Neal, 1992) and SP MS (methanesulfonate) (Meyerson et al., 2002). Figure from Mayewski et al. (2005). $\delta^{14}C$ series used as an approximation for solar variability (Stuiver and Braziunas, 1993). CO_2 from DSS ice core (Etheridge et al., 1996). Darkened area shows the 1700-1850 AD era climate anomaly discussed in the text.

Recent reductions in ice extent over the Arctic, around the Antarctic Peninsula, and for many mid-low latitude glaciers demonstrate some of the initial impacts of the rise in temperature measured over the last few decades (Arctic Climate Impact Assessment, 2005; Intergovernmental Panel on Climate Change, 2007). Reconstructions of past temperature indicate that this rise is anomalous relative to temperature variability over the last 2,000 years (e.g. Mann and Jones, 2003; Moberg et al., 2005). However, the association between change in temperature and change in atmospheric circulation under natural conditions has not been examined as vigorously. This association is perhaps most critical to investigate in the polar latitudes, where future warming is expected to be the greatest (Intergovernmental Panel on Climate Change, 2007). Synthesis of 50 well-dated, continuous palaeoclimate records covering the current interglacial (the last 11.7 ka, Holocene) reveals the occurrence of at least six periods of naturally forced abrupt climate change, of which several coincide with major disruptions in civilization (Mayewski et al., 2004b). It is within the last 2,000 years that annually resolved dating of stratigraphic records is most accurate and further this time period is characterized by boundary conditions (e.g., ice, ocean, atmosphere) most similar to those of today as well as two analogues for naturally warm (Northern Hemisphere Medieval Warm period) and cool (Little Ice Age) climates. Over the period of the last 2,000 years established, well-dated, proxy records of change in temperature are available (e.g. Mann and Jones, 2003). In addition, for this period, bipolar ice core records provide proxy reconstructions of past changes in regional scale atmospheric circulation, a major component of the climate system that has not received the same detailed attention as that given to past temperature, despite a strong association with temperature over a wide range of timescales (Mayewski et al., 1997; Thompson and Wallace, 2000; Bertler et al., 2004; Masson-Delmotte et al., 2005; Schneider et al., 2006). Comparison between ice core proxies for atmospheric circulation and multiple proxies for temperature reveals associations over the last few decades that are inconsistent with those of the past 2,000 years. Notably, patterns of middle to high latitude atmospheric circulation in both hemispheres are still within the range of variability of the last 6–10 centuries (Mayewski and Maasch, 2006, Figure 3.20), while, as demonstrated by Mann and Jones (2003), Northern Hemisphere temperatures over recent decades are the highest of the last 2,000 years. Further, recent temperature change in the Northern Hemisphere precedes change in middle to high latitude atmospheric circulation unlike the two most notable changes in climate of the past 2,000 years (Little Ice Age and Medieval Warm Period) during which change in atmospheric circulation preceded or coincided with change in temperature (Figure 3.21). In addition, the most prominent change in Southern Hemisphere temperature and atmospheric circulation of the past 2,000 years, and probably of the last 9,000 years, precedes change in temperature and atmospheric circulation in the Northern Hemisphere unlike the most recent change in Northern Hemisphere temperature that leads (Figure 3.21). These findings provide new verification that recent rise in temperature is inconsistent with natural climate forcing and is most likely related to anthropogenic activity in the form of enhanced greenhouse gases. Figure 3.21 also demonstrates that the delayed warming, relative to the Northern Hemisphere, over much of the Southern Hemisphere may be, in addition to other factors (ocean thermal capacity, Antarctic ice sheet and sea ice reflectivity and size) a consequence of underpinning by natural climate variability (Figures 3.21 and 3.22).

Non-sea-salt (nss) Ca records from eight International Trans-Antarctic Science Expedition (ITASE) West Antarctic ice cores, covering the period AD 1400 – 2002, and the Siple Dome deep ice core, including the last 10,000 years, reveal that Southern Hemisphere westerlies intensification since ~1980 is unprecedented for at least the last 5,400 years, supporting the proposed role of human activity in this intensification (Dixon et al., In Review). This study also demonstrates the abrupt termination of the Southern Hemisphere westerlies intensification 5,400 years ago and abrupt terminations to earlier such intensifications of the westerlies. Based on this analogue (Dixon et al., In Review) propose that ozone hole recovery and continued greenhouse gas warming could trigger yet another abrupt weakening of the westerlies, leading to a hastening of Antarctic warming and accelerated changes in Antarctic and Southern Hemisphere climate.

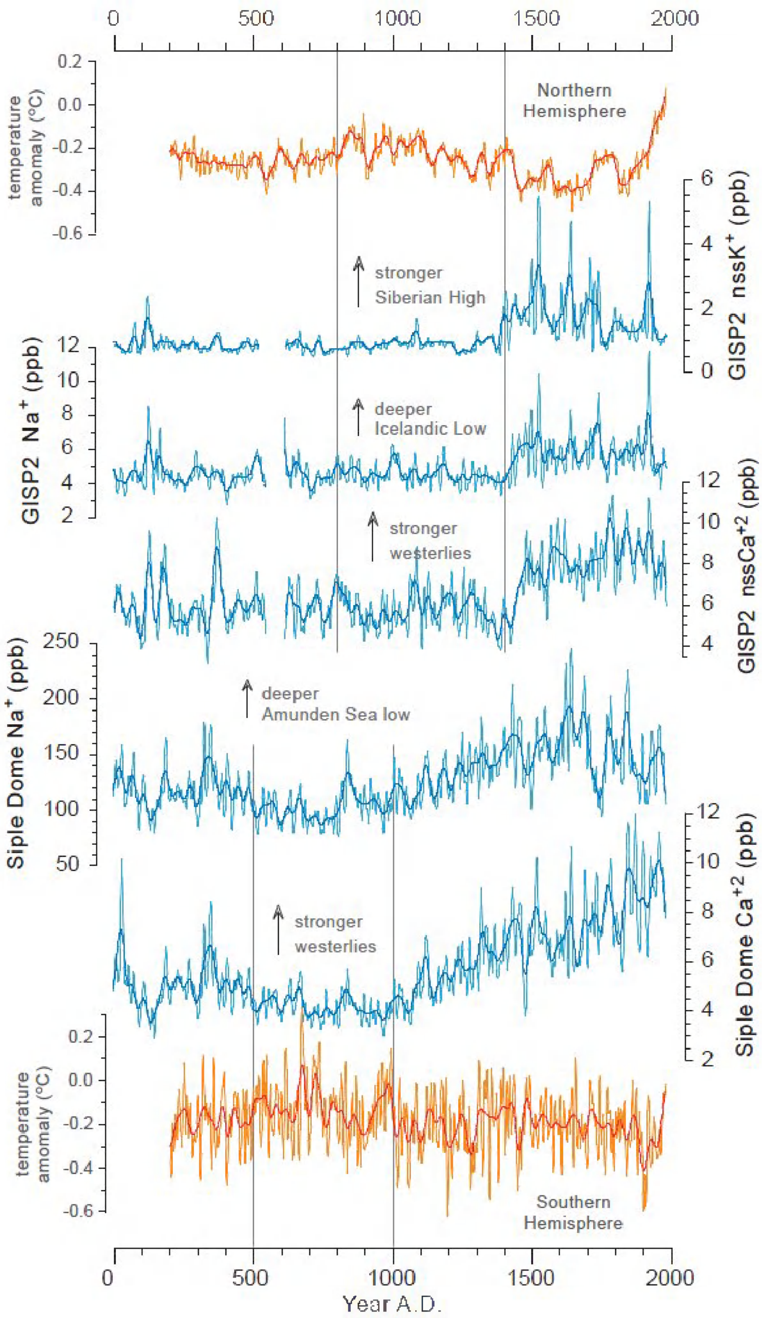


Figure 3.21 Northern and Southern Hemisphere reconstructed temperatures (in red from Mann and Jones, 2003) and ice core reconstructed atmospheric circulation systems (in blue from Mayewski and Maasch, 2006) (Icelandic Low, Siberian High, Northern and Southern Hemisphere westerlies, and Amundsen Sea Low). Data are presented with less than 10-yr signal (light line) extracted to approximate the original annual to multi-annual series and with the less than 30-yr signal (dark line) extracted series to facilitate examination at decadal scales. Vertical lines refer to onset for temperature change (earliest refers to Medieval Warm Period and second to Little Ice Age, the two most recent analogues for naturally warm and cool temperatures, respectively). These ice cores were chosen because they are the highest resolution Antarctic and Greenland ice core data of their kind available. Figure taken from Mayewski and Maasch (2006).

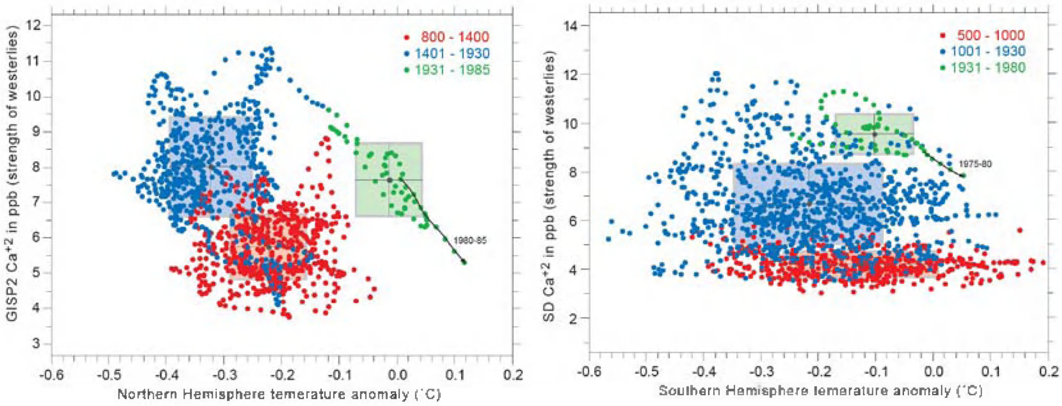


Figure 3.22 (Left side) Phase diagram for Northern Hemisphere temperature versus ice core proxy for Northern Hemisphere westerlies (shown in Figure 3.21, light lines). (Right side) Phase diagram for Southern Hemisphere temperature versus ice core proxy for Southern Hemisphere westerlies (shown in Figure 3.21, light lines). Red dots are data from 800–1400 AD, blue dots 1401–1930 AD, and green dots 1931–1985 AD. The shaded red, blue and green boxes represent the mean ± one standard deviation of the data for each these time periods, respectively. The black arrow labelled 1980–1985 highlights data for the last 5 years of the record. Note that recent decades of westerlies in both hemispheres are within range of variability of the Little Ice Age (blue dots) although westerlies in Northern Hemisphere as of late 1980s are approaching Medieval Warm Period (red dots) conditions. Figures from Mayewski and Maasch (2006).

3.4.2 Changes in sea ice extent through the Holocene

Despite the fundamental importance of sea ice to many aspects of Holocene climate (see section 3.3.5), it is a factor that remains poorly constrained in computer simulations of past and future climate change. One of the reasons for this is the paucity of historical and palaeo records of sea ice extent. Routine satellite measurements of sea ice began only in the 1970s, and the strong interannual variability in these short instrumental records makes it difficult to isolate long-term changes in sea ice cover (Zwally et al., 2002a). Whaling records and early sea ice charts do suggest that Antarctic sea ice cover has undergone a dramatic decline during the 20th century (de la Mare, 1997), although the quality and interpretation of these early historical records is debated. In order to improve the incorporation of sea ice into computer models of future climate change it will be essential to use proxy records to represent how Antarctic sea ice has changed in the past.

Much of our current understanding of Holocene sea ice changes around Antarctica comes from a few marine sediment cores in which the diatom microfossils are used as a proxy for past sea ice coverage over the core site (Hodell et al., 2001), although these have yet to be calibrated with the instrumental record. Instead, qualitative reconstructions of the presence or absence of sea ice at a core site can be made by measuring changes in the abundance of those diatom species that are associated either with the presence of winter sea ice or with the presence of summer sea ice. Quantitative reconstructions of changes in the duration of sea ice cover at particular sites can be made by identifying changes in the relative abundances of a number of diatom species (Crosta et al., 1998a; Crosta et al., 1998b), see Figure 3.23 from Crosta et al. (2007).

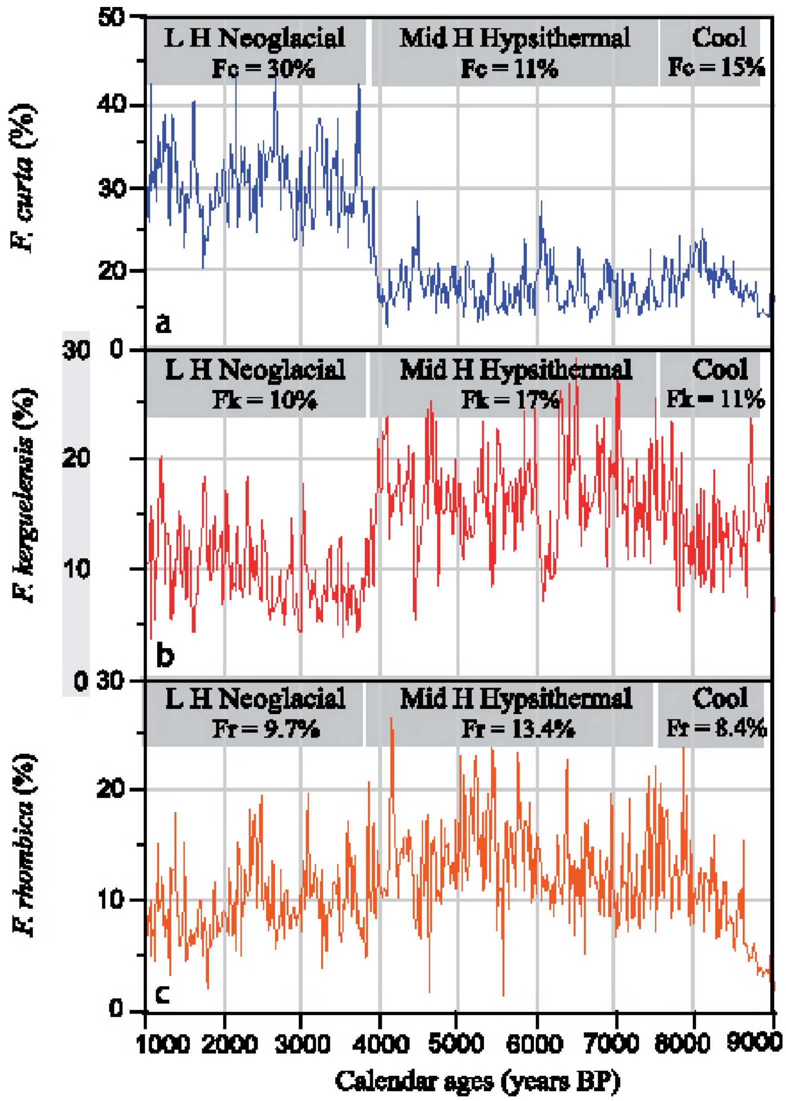


Figure 3.23 Relative abundances of (a) the *Fragilariopsis curta* group and (b) *F. kerguelensis* and (c) *F. rhombica* versus calendar ages in core MD03-2601 off Adélie Land, East Antarctica. Mean abundances of the three species are reported for each climatic period. Fc, F. curta; Fk, *F. kerguelensis*; Fr, *F. rhombica*; LH, late Holocene; Mid H, mid Holocene (Crosta et al., 2007).

Holocene sediment cores from the Southern Ocean, for example off Adélie Land, East Antarctica, generally record reduced sea ice coverage during the early to mid-Holocene (Figure 3.23) (Hodell et al., 2001). This minimum in sea ice cover is broadly coincident with the timing of a Holocene climatic optimum documented in some marine palaeoclimatic records from the Antarctic continent and Southern Ocean (Masson et al., 2000; Domack et al., 2001). Also known as a Hypsithermal, this marine-inferred Holocene climatic optimum seems to have ended at different times around Antarctica. In the Atlantic sector of the Southern Ocean (offshore from Dronning Maud Land and the Weddell Sea) sediment cores show that marine-inferred conditions were warm and ice free at ~50°S until around 6-5 ka BP (Hodell et al., 2001; Nielsen et al., 2004). There is some leeway in the dating, which in

Hodell's core was determined at 5 ka by extrapolation from sedimentation rates, leaving room for uncertainty. This marine-inferred mid Holocene warm period was followed by a marine Neoglacial cool period that lasted until at least ~2 ka BP, when winter sea ice cover extended out to beyond the sediment core sites at ~50°S. This was then followed by a late Holocene warming that saw sea ice retreat southward of 50°S in the Atlantic sector (Nielsen et al., 2004), although the evidence for this warming and sea ice retreat is less obvious in a core site at 53°S (Hodell et al., 2001). At present the timing of this marine-inferred mid-Holocene climate optimum is out of phase with some of the ice core and terrestrial records described later (Section 3.4.3) and requires further study.

In the Southern Ocean sector facing Australia (offshore of Adélie Land in East Antarctica) a diatom record from 66°S suggests that the transition to increased sea ice coverage in this marine-inferred Neoglacial cool period occurred at ~4 ka BP, marked by an abrupt threshold change in the relative abundance of *F. kerguelensis* and *F.curta* which have particular ecological responses at 0.5°C, and a more gradual change in *F. rhombica* (Figure 3.23), the latter giving a signal that is in good agreement with model output (Renssen et al., 2005). This lasted until at least ~1 ka BP (the top of the core) (Figure 3.23) (Crosta et al., 2007). As well as these long-term changes in winter sea ice extent, changes in diatom species also suggest that during the colder Neoglacial period both the autumn formation and spring melting of sea ice occurred later in the season than during the Hypsithermal (Crosta et al., 2008). A sediment core record from the eastern Pacific sector of the Southern Ocean (offshore of the western Antarctic Peninsula) also supports a delayed transition at ~3.6 ka BP to cool marine Neoglacial conditions (Domack et al., 2001).

The drivers of this marine-inferred mid-Holocene cooling in ocean temperature and increase in sea ice around Antarctica after ~6 ka BP are still debated. One hypothesis is that cooling of the Southern Ocean during the mid-Holocene may have been a response to the decrease in summer insolation at high latitudes in the Northern Hemisphere (see the insolation curve in Figure 3.16), which may have led to an increase in the strength of the thermohaline circulation (Hodell et al., 2001; Nielsen et al., 2004). However, a numerical model of the influence of seasonal changes in insolation over Antarctica suggests that the observed regional changes in Holocene temperatures and sea ice extent in the Southern Ocean may instead be consistent with forcing by southern summer insolation (Renssen et al., 2005). This model suggests that the temperature and sea ice trends in the Southern Ocean can be explained by the combination of a 1-2 month lag of ocean temperature to local insolation, and the long memory of the Southern Ocean system - which allows local insolation signals from different seasons to be preserved throughout the year. Sea ice also plays an important role in the modelled results by amplifying the effect of the local insolation signal through ice-albedo and ice-insulation feedbacks. This model simulation also suggests that sea ice cover in the early Holocene was more reduced in the western than in the eastern sector of the Southern Ocean, which could account for the regional differences in the timing/length of the Holocene climatic optimum in palaeoclimate records.

As well as recording the seasonal and inter-annual changes in sea ice during the Holocene, diatoms from sediment cores record decadal to millennial-scale cyclic variations in Antarctic sea ice (Nielsen et al., 2004; Crosta et al., 2008). Based on spectral analysis, it has been proposed that many of these high-frequency oscillations (c. 200 yr) in sea ice are consistent with the periodicities in records of solar activity (e.g. $\Delta^{14}C$). Other periodicities in the Holocene diatom records may be associated with internal climate variability related to the global thermohaline circulation, and possibly act to amplify the influence of high-frequency variations in solar activity (Crosta et al., 2008). Verification of the drivers associated with decadal to millennial-scale cyclical changes in sea ice is hampered by the limited resolution and dating control on most marine sediment cores. However, the possibility of developing highly resolved sea ice reconstructions in the future from varved marine sediments or from ice cores (see below) may help to establish the degree to which external (e.g. solar) and internal (e.g. thermohaline circulation) forcings affect the amount of sea ice around Antarctica.

Efforts are currently underway to establish new proxies to help build reliable reconstructions of Antarctic sea ice through the Holocene, and beyond. Promising proxies for the future may come from chemical markers in marine sediment cores from around coastal Antarctica. The presence of the highly branched isoprenoid biomarker known as IP_{25} in Arctic sea ice and sediments appears to be associated with the *Haslea* spp. sea ice diatom (Belt et al., 2007). Early studies using sediment cores from the Arctic and Antarctic have found that the sea ice reconstructions produced using IP_{25} and similar biomarkers are consistent with historical sea ice reports (Massé et al., 2008) and sea ice estimates based on diatom assemblages (Figure 3.24). These biomarkers can be measured rapidly in extremely small sample sizes and have been readily detected in sediments covering the entire Holocene. Although their long-term behaviour in sediments needs to be established further, it appears that these biomarkers will play a valuable role in the development of proxy sea ice records around Antarctica.

The chemistry of Antarctic ice cores may also provide proxies for reconstructing the Holocene history of Antarctic sea ice. Sea salt in ice cores has traditionally been viewed as a proxy for wind transport strength and storminess (Iizuka et al., 2008). However, the observation of increased sea salt in Antarctic ice cores during winter and during glacial periods, as well as a characteristic sulphate depletion in the sea salt aerosol reaching Antarctica, has recently led to the suggestion that sea salt may instead reflect the formation of brine and frost flowers on top of new sea ice (Rankin et al., 2002; Wolff et al., 2003). Ice core records of sea salt appear to respond to sea ice changes over long timescales (e.g. glacial-interglacial), but more quantitative calibration is required (Wolff et al., 2006; Fischer et al., 2007). While sea salt is a promising proxy from a conceptual point of view, it still needs to be shown whether it is sensitive to changes in sea ice production over short timescales of centuries or less. It also appears that the response of this proxy reaches a threshold level when sea ice is very extensive (i.e. during glacial periods).

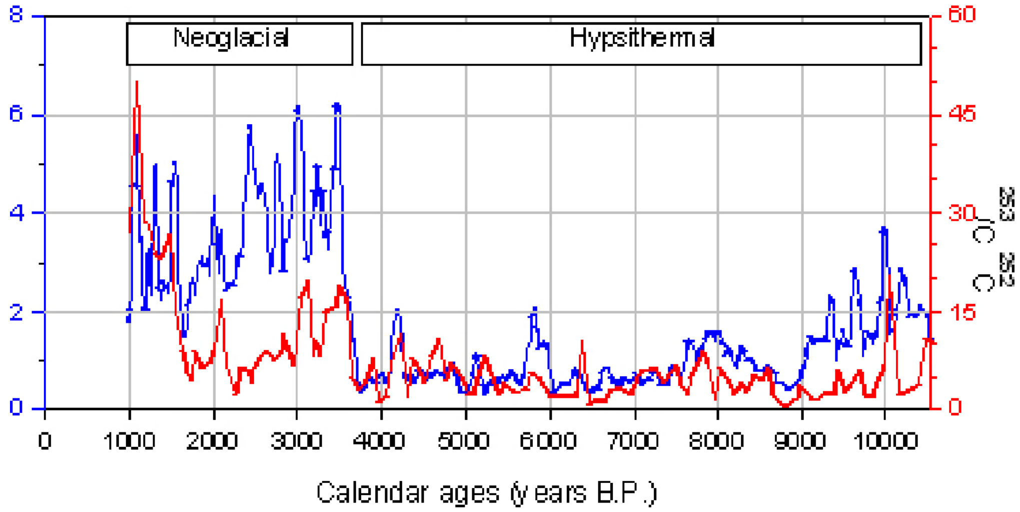


Figure 3.24 Diatom (blue) and highly-branched isoprenoid (red) records from a sediment core retrieved from the Adélie Trough off Wilkes Land, East Antarctica. High values of the *F. curta*/*F. kerguelensis* ratio (left axis) and the $C_{25:2}/C_{25:3}$ ratio (right axis) during the Neoglacial indicate greater sea ice cover over the core location (Massé and Crosta, unpublished data)

Another promising sea ice proxy in Antarctic ice cores is methane sulphonic acid (MSA) (Curran et al., 2003). MSA is ultimately derived from marine phytoplankton that thrive in the marginal sea ice zone. At some ice core sites around coastal Antarctica the amount of MSA deposited in Antarctic snow increases following winters of increased sea ice extent (Welch et al., 1993; Curran et al., 2003; Abram et al.. 2007b). However, at other locations MSA variability appears to be more sensitive to wind direction (Fundel et al., 2006; Abram et al., 2007b). This means that site selection and transport processes are essential considerations in using ice core records as sea ice proxies. Post-depositional changes in MSA records at sites with low accumulation rates and over long time periods also mean that the use of this proxy is essentially limited to high accumulation records of the Holocene.

Reconstructions of Antarctic sea ice extent based on proxy data from ice cores have the advantage of providing a regionally averaged view, rather than being sensitive to sea ice conditions over a single (coring) point in the Southern Ocean. Ice core records also have the potential to provide both long-term records of sea ice changes through the Holocene, and very detailed (i.e. annual resolution) reconstructions of sea ice changes during recent centuries. Early comparisons of high-resolution MSA reconstructions of sea ice extent from sites around the Antarctic continent reveal clear regional differences in the timing and speed of Antarctic sea ice decline (and growth) during the 20th century (Curran et al., 2003; Abram et al., 2007a).

3.4.3 Regional patterns of Holocene climate change in Antarctica

In this section we provide a spatial synthesis of records of climate and environmental changes in East Antarctica (EA), the Antarctic Peninsula (AP), and the Ross Sea region (RS). West Antarctica, which is currently understudied, is also briefly described. We build on previous reviews by Ingólfsson et al. (1998, 2003), Ingólfsson and Hjort (2002), Ingólfsson (2004), Jones et al. (2000) and Hodgson et al. (2004a) and focus on four main periods, namely: (1) the deglaciation history of currently ice-free regions and the Pleistocene-Holocene transition; (2) the period after the early Holocene, (3) the Mid Holocene warm period or Hypsithermal (MHH), and (4) the past 2,000 years with a focus on Neoglacial cooling, the presence of warm periods, the possibility of a Little Ice Age (LIA) like event, and the recent rapid climate changes documented in instrumental and observational records. It should be noted that these climate periods are not always synchronous in different regions of Antarctica which might in part be due to different degrees of chronological control, or to forcing mechanisms operating at different intensities between regions. At present, with the exception of the Antarctic Peninsula (Bentley et al., 2009), the role of various forcing mechanisms in regional climate change is poorly described.

In order to allow comparison between the different records in studies where ^{14}C dates were not calibrated we list the original ^{14}C dates (^{14}C ka BP), together with the upper and lower limits (at 2-std deviations) of the data (cal. ka BP) generated by the radiocarbon calibration method CALIB 5.0.2 (http://calib.qub.ac.uk/calib/). Radiocarbon dates of marine samples were corrected for the reservoir effect by subtracting 1,300 yrs following the Antarctic standard prior to calibration (i.e. the offset from the global marine reservoir was set at 900 years when using the marine calibration curve; Hughen et al. (2004)). For lacustrine ^{14}C ages younger than 11 ka cal yr BP the Southern Hemisphere atmospheric calibration curve was used (McCormac et al., 2004); in all other cases the Northern Hemisphere atmospheric calibration curve (Reimer et al., 2004) was applied. The dates of deglaciation of the current ice-free regions are largely derived from ^{14}C dating of fossils in raised beaches, organic material and fossils in lake sediments, peat deposits and bird colonies; they are thus minimum ages since there is an unknown lag time between deglaciation and colonization of the land by biota (e.g. Gore, 1997; Ingólfsson et al., 2003).

3.4.3.1 East Antarctica (EA)

Deglaciation history and the Pleistocene-Holocene transition: The widespread Antarctic early Holocene optimum between 11.5 and 9 ka BP is observed in all ice cores from coastal and continental sites (Steig et al., 2000; Masson-Delmotte et al., 2004) and coincided with biogenic sedimentation commencing in lakes along the East Antarctic margin and the occupation of ice-free land by biota between c. 13.5 and 10 ka BP (Figure 3.25). The early Holocene climate evolution seems to be consistent in the regions studied so far (i.e., Amery Oasis (70°40'S-68°00'E), the Larsemann Hills (69°20'S-76°50'E), and the Vestfold Hills (68°30'78°00'E)). In the Larsemann Hills some areas escaped glaciation during the LGM, whereas other areas became gradually ice-free between c. 13.5 and 4 ka BP (Hodgson et al., 2001). Diatoms and pigment data point to the establishment of seasonally melting lake ice and snow cover and the development of microbial mats at c. 10.8 ka BP (Hodgson et al., 2005), with evidence for relatively wet conditions between c. 11.5 and 9.5 ka BP in a lake on one of the northern islands (Verleyen et al., 2004b). This is consistent with palaeolimnological evidence from the nearby Vestfold Hills, which were probably more fully glaciated during the LGM, but where lakes became ice-free and diatoms and rotifers inhabited the lakes from c. 11.4 ^{14}C ka BP (c. 13.2-13.4 ka BP); (Roberts and McMinn, 1999a; Cromer et al., 2005). At least parts of Amery Oasis were covered by locally expanded glaciers during the Late Pleistocene (Hambrey et al., 2007), and deglaciation in some areas started around c. 11 ka BP (Fink et al., 2006), whereas biogenic sediments started to accumulate in lakes in other parts of the region at c. 12.5 ka BP (Wagner et al., 2004; Wagner et al., 2007), broadly coincident with deglaciation of the Larsemann and Vestfold Hills. Deglaciation was followed by the establishment of a diatom community in one of the lakes, likely related to increased nutrient inputs and a reduction in ice and snow cover at c. 10.2 ka BP, marking the start of relatively warm conditions (Wagner et al., 2004).

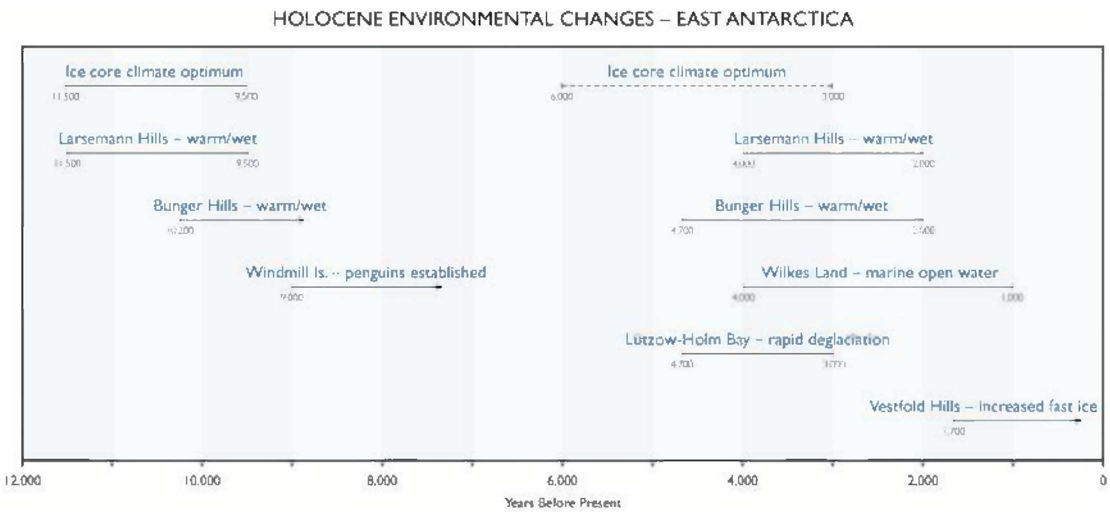


Figure 3.25 Selected Holocene environmental changes – East Antarctica

In Wilkes Land, parts of the Bunger Hills (66°10'S-101°00'E) remained ice-free during the LGM (Gore et al., 2001), whereas the Windmill Islands (66°20'S-110°30'E) were probably glaciated (Goodwin, 1993). Minimum ages for deglaciation in the Windmill Islands are also slightly younger than those from the oases near the Lambert Glacier; post-glacial lake sediments accumulated at c. 10.2 ka BP (Roberts et al., 2004), biogenic sedimentation in the marine bays started around c. 10.5 ka BP (Cremer et al., 2003; Hodgson et al., 2003), and penguins occupied the region from at least c. 9 ka BP (Emslie and Woehler, 2005). Relatively cool summer conditions near the Windmill Islands probably prevailed during the early Holocene, as reflected by the microfossil record in coastal marine sediment cores (Cremer et al., 2003). The Bunger Hills were occupied by snow petrels from at least 10 ka (Verkulich and Hiller, 1994), and organic sediments started to accumulate in the lakes there at the Pleistocene-Holocene boundary in association with extensive and relatively rapid ice melting, which similarly points to an early Holocene warm period at c. 9 +/- 0.5 ka BP (Verkulich et al., 2002), and is followed by a marine optimum (see below, Kulbe et al., 2001). Radiocarbon evidence suggests that large parts of the Southern Bunger Hills were rapidly deglaciated prior to 8 ka BP (Melles et al., 1997).

Although terrestrial climate archives are present in the ice-free regions in Dronning Maud Land (Matsumoto et al., 2006), surprisingly little information is available about the deglaciation and post-glacial climate evolution there. The Untersee Oasis (71°S-13°E) was probably ice-free during the LGM as shown by ^{14}C dating of organic deposits from snow petrels, which indicates an occupation during at least the past 34 ka (Hiller et al., 1988). Some islands in the Lützow-Holm Bay near Syowa Station (69°00'S-39°35'E) are believed to have been ice-free for at least 40 ka and probably longer, as evidenced by AMS ^{14}C dates of individual *in situ* marine fossils from raised beach deposits (Miura et al., 1998).

In summary, parts of some East Antarctic oases escaped glaciation during the LGM, whereas others were probably glaciated and became gradually ice-free at the Pleistocene-Holocene boundary, with some regional differences in the timing of deglaciation and colonization by biota. The early Holocene climate optimum is detected in terrestrial and coastal marine records between c. 11.5-9.5 ka BP, centred on c. 10 ka BP, when most glaciated regions became ice-free and organic deposits started to accumulate, and in ice cores between 11.5 and 9 ka BP (Masson et al., 2000).

After the early Holocene: All eastern Antarctic sites show a weak climate optimum between 6 and 3 ka but in general the period after deglaciation shows complex and less consistent patterns than those observed at the Holocene-Pleistocene boundary. In the oases near the Lambert Glacier, relatively dry conditions occurred on land between c. 9.5 and 7.4 ka BP, and in the Larsemann Hills, lake levels dropped below their present position (Verleyen et al., 2004b). Marine sediments in isolation basins at c. 7.4 and 5.2 ka BP are consistent with a marine climate optimum, which is not clearly evident in the terrestrial sediments. In contrast, warm conditions prevailed on land in the Amery Oasis in the early Holocene (c. 10.2 ka BP), which lasted until c. 6.7 ka BP and with a clear optimum between c. 8.6 and 8.4 ka BP (Cremer et al., 2007), whereas cold conditions prevailed from c. 6.7 ka BP onwards until c. 3.7 ka BP (Wagner et al., 2004). In the Vestfold Hills, isostatic rebound and the emergence of isolation lakes from the sea resulted in a major ecosystem change, which hampers detailed palaeoclimatological inferences from being made for the period after the early Holocene optimum, particularly in lower altitude lakes (Fulford-Smith and Sikes 1996; Roberts and McMinn, 1999b).

In Wilkes Land, open water conditions were inferred from the marine sediments of an isolation basin in the Windmill Island between c. 8 and 4.8 ^{14}C ka BP (c. 9–4.5 ka BP), but the dating uncertainty is large because ^{14}C dates are few and there is a variable reservoir effect throughout the sediment (Roberts et al., 2004). Relatively cool summer conditions were observed in a nearby marine bay with a combination of winter sea ice and seasonal open water conditions between c. 10.5-4 ka BP (Cremer et al., 2003; Hodgson et al., 2003). The peak of this (marine) cooling period was pinpointed at between c. 7 ka and 5 ka BP, when penguin colonies where abandoned on one of the peninsulas (Emslie and Woehler, 2005). In the Bunger Hills cold and dry conditions prevailed between c. 9 and 5.5 ka BP, with a low input of glacial meltwater in the lakes and a permanent lake-ice cover (Verkulich et al., 2002), coincident with the extensive occupation of snow petrels between c. 8 and 6 ka BP (Verkulich and Hiller, 1994) probably as a result of more distal glaciers and snow fields. In contrast, a marine optimum was identified in coastal sediments between c. 9.4 to 7.6 ka BP (Kulbe et al., 2001), followed by cold marine conditions between c. 7.6 and 4.5 ka BP deduced from low organic carbon accumulation rates.

In summary, the period following deglaciation shows complex patterns with a marine climate optimum in some areas apparently out of phase with a terrestrial optimum or coincident with cool and dry conditions on land. Dating uncertainties prevent an in depth correlation between the different anomalies. This might reflect the fact that the organic fraction in marine sediments records spring to autumn conditions (including sea ice blooms), whereas lacustrine biotic assemblages largely reflect summer conditions when the lakes are ice-free and primary production peaks, and in some cases spring (under-ice) blooms (Hodgson and Smol, 2008).

Mid Holocene warm period – Hypsithermal: A Mid Holocene Hypsithermal (MHH) is present in various ice, lake and marine core records from Antarctica (see Hodgson et al., 2004a) for a review) including ice-free oases near the Lambert Glacier (note: the timing of the MHH differs from the marine/sea ice inferred 'hypsithermal' in some of the records discussed in section 3.4.2 - suggesting that sea ice responds to different forcing). In the Larsemann Hills the MHH is dated between c. 4 and 2 ka BP. There, relatively wet conditions occurred on land, and predate the coastal marine optimum observed in isolation basins (Verleyen et al., 2004a; Verleyen et al., 2004b). A short return to dry conditions and low water levels is present in one of the lake records at c. 3.2 ka BP (Verleyen et al., 2004b). The relatively wet MHH is coincident with the restart of biogenic sedimentation in Progress Lake at 3.5 ka BP, after at least 40 ka of permanent lake ice cover (Hodgson et al., 2006a), and the formation of proglacial lakes occupying Stornes, the eastern of the two main peninsulas in the Larsemann Hills between c. 3.8 and 1.4 ^{14}C ka BP (c. 4.4 - 4.1 and 1.3 ka BP, Hodgson et al., 2001). In the Amery Oasis, the relatively warm conditions of the MHH between c. 3.2 and 2.3 ka BP are inferred from abundant organic matter deposition in Lake Terrasovoje (Wagner et al., 2004). In the Vestfold Hills a decline in lake salinity could be inferred between c. 4.2 and 2.2 ka BP, but dates are uncertain (Björck et al., 1996; Roberts and McMinn, 1996, 1999a). This period of low salinity is however broadly consistent with the warm and humid conditions between c. 4.7 and 3 ka BP proposed by Björck et al. (1996) after reinterpretation of previously published results (Pickard et al., 1986). In contrast, Bronge (1992) inferred relatively cold conditions between c. 5 and 3 ka BP and a short but marked cooling event between c. 2.3 and 2 ka BP.

In Wilkes Land, enhanced biological production, probably reflecting more open water conditions and a climate optimum, occurred between c. 4 and 1 ka BP (Kirkup et al., 2002). This coincided with a local marine optimum characterised by open water and stratified conditions caused by enhanced meltwater input (Cremer et al., 2003). In this area the MHH coincided with the readvance of the Law Dome ice margin after c. 4 ka BP, in response to an increase in precipitation (Goodwin, 1996).

In the Bunger Hills a stepwise increase in primary production was reported in the lakes between c. 4.7 and 2 ka BP (Melles et al., 1997). The start of the MHH slightly postdates the start of warm conditions between c. 5.5 and 2 ka BP, interpreted by Verkulich et al. (2002) from the pattern of draining of ice-dammed lakes. A marine optimum occurred in this area between c. 3.5 and 2.5 ka BP, which was preceded by a gradual warming from c. 4.5 ka BP onwards (Kulbe et al., 2001).

In the Lützow-Holm Bay region a rapid isostatic rebound (6 m in c. 1,000 years) occurred between c. 4.7 and 3 ka BP, which was linked to the rapid removal of part of the regional ice mass, most likely as a result of melting caused by the MHH (Okuno et al., 2007).

In summary, there is evidence for a MHH in EA, but dating uncertainties are still high in some areas. Because the MHH acts as one of the historical analogues for the present day warming climate, there is a pressing need for well-dated lake sediment records to study this past climate anomaly and its influence on ecosystem functioning. There is a disappointing lack of long-term high-resolution records from ice-free areas in the Dronning Maud Land region.

Past 2,000 years - Neoglacial cooling, the Little Ice Age and recent climate change: Much attention has been paid to the fluctuations in climate that gave rise in the Northern Hemisphere to the well-documented Medieval Warm Period (900-1300 AD / 1100-700 yr BP), and Little Ice Age (LIA) consisting of cool intervals beginning about 1650, 1770, and 1850 AD, each separated by slight warming intervals / 300-100 yr BP), the earlier part of the latter coinciding with the Maunder Minimum, a period of minimal sunspot activity and low solar output from 1645-1715 AD / 355-285 yr BP). The search in Antarctica for short-term climate signals like these that are apparent in the Northern Hemisphere is an important element in understanding how the Earth's climate system works.

In the Lambert Glacier region there is some evidence of Neoglacial cooling in the Larsemann Hills (Hodgson et al., 2005) leading to dry conditions around 2000 yr BP, 700 yr BP and between c. 300 and 150 yr BP in some of the lakes (Verleyen et al., 2004b). The lake evidence parallels declines in sea bird populations during the past 2,000 years (Liu et al., 2007). In the Amery Oasis, Neoglacial cooling followed the MHH from c. 2.3 ka BP onwards, with a short return to a relatively warmer climate between c. 1.5 and 1 ka BP (Wagner et al., 2004). In the Vestfold Hills an increase in fast ice extent is observed from c. 1.7 ka BP (McMinn, 2000), broadly coincident with the Chelnock Glaciation on land (Adamson and Pickard, 1986). Cold conditions were inferred between c. 1.3 ka BP and 250 yr BP (Bronge, 1992) and low precipitation has tentatively been inferred from c. 1.3-1.5 ka BP, but dating uncertainty is high (Roberts and McMinn, 1999a). Meltwater input into the lakes gradually decreased from 3 ka BP onwards (Fulford-Smith and Sikes, 1996). A palaeohydrological model derived from the reconstructed changes in salinity and water level throughout sediment cores suggested that there was no significant change in evaporation for the last c. 700 years, but that a lower evaporation period is evident between c. 150 - 200 yr BP, suggestive of a mild LIA-like event in the Vestfold Hills (Roberts et al., 2001).

In Wilkes Land, Neoglacial cooling and persistent sea ice cover were observed in the marine bays near the Windmill Islands (Kirkup et al., 2002; Cremer et al., 2003). A slow decrease in lake water salinity was observed on land during the late Holocene, and there is no evidence for a LIA-like event there (Hodgson et al., 2006c). Instead, we see a very rapid salinity rise during the past few decades that has been brought about either by increased evaporation rates or a decrease in precipitation. In the Bunger Hills a rapid growth of the petrel population after c. 2 ka BP was reported by Verkulich and Hiller (1994), and coincides with climate cooling (Melles et al., 1997; Verkulich et al., 2002) and a glacier readvance during recent centuries (Adamson and Colhoun, 1992). After the cooling event at c. 2 ka BP relatively warm conditions, yet colder than during the Hypsithermal, prevailed with an additional cooling trend that started recently (Verkulich et al., 2002).

In the Lützow-Holm Bay region lake sediment cores are starting to provide insights in past climate variability during the late Holocene. An increase in the total organic carbon to total nitrogen ratio was linked to an increase in the aquatic moss vegetation from c. 1.1 ka BP onwards (Matsumoto et al., 2006). However, detailed palaeoclimatic records are still lacking for the region.

In summary, most East Antarctic areas experienced Neoglacial cooling, with some markedly cooler or drier events in places. The apparent differences in the dating of these events from one part of the region to another might be related to dating uncertainties or to the lack of high-resolution records in some areas. With the exception of some ice cores (such as Law Dome where CO_2 mixing ratios were in the range 275-284 ppm with the lower levels during 1550-1800 AD, probably as a result of colder climate (Etheridge et al., 1996)) there is neither convincing evidence for a Little Ice Age event, nor for anything corresponding convincingly, and region-wide, to the Medieval Warm Period of the Northern Hemisphere.

3.4.3.2 Antarctic Peninsula (AP)

Deglaciation history and the Pleistocene-Holocene transition: The pattern and mechanisms of Holocene palaeoenvironmental change in the AP region have recently been reviewed (Bentley et al., 2009). The early Holocene climate optimum detected in ice cores lasted to around 9.2 ka BP (Masson et al., 2000; Masson-Delmotte et al., 2004) occurred at the same time as the continued deglaciation of the AP continental shelf (Bentley, 1999; Ingólfsson et al., 2003; Bentley et al., 2006). Ice sheet retreat around the AP probably began c. 14-13 ka BP (Evans et al., 2005; Heroy and Anderson, 2005), and continued through the Holocene (Figure 3.26). The early Holocene optimum is however absent in the offshore Palmer Deep record, which is characterized by an apparent 'cold' proxy record at that time (11.5 – 9.1 ka BP) (Domack, 2002; Sjunneskog and Taylor, 2002; Taylor and Sjunneskog, 2002). There is little evidence in the terrestrial record in the Peninsula for an early Holocene climate optimum, because most currently ice-free areas were probably still ice-covered before c. 9.5 ka BP (Ingólfsson et al., 1998, 2003).

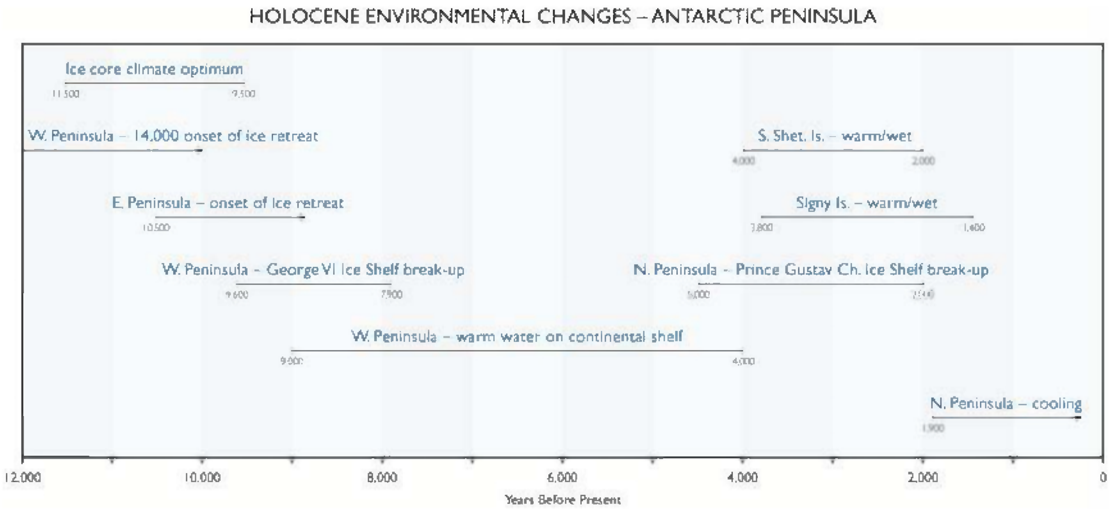


Figure 3.26 Selected Holocene environmental changes – Antarctic Peninsula

After the early Holocene: The period after the early Holocene optimum shows complex patterns in the AP region. Ice shelves on the western side had collapsed, whilst those on the east were still stable (Hodgson et al., 2006b). The onset of marine conditions in epishelf lake sediments on Alexander Island shows that George VI Ice Shelf collapsed at c. 9.6 ka BP, immediately following the early Holocene optimum (Bentley et al., 2005b). At the same time, ocean records from the Palmer Deep, a basin on the continental shelf of the AP, indicate a dramatic increase in the presence of warmer surface waters over the AP's continental shelf (Leventer et al., 2002) so it is likely that the ice shelf was attacked both from above (atmospheric temperatures) and below (warm ocean currents) (Smith J A et al., 2007). The ice shelf reformed from c. 7945 yr BP (Bentley et al., 2005b; Smith et al., 2007; Roberts et al., 2008). In contrast, evidence from the Larsen B Ice Shelf, east of the AP, shows that it was stable throughout the Holocene (from 11.5 ka BP), but has now collapsed (in 2002) due to a combination of long-term (postglacial) thinning and cracking combined with rapid recent warming (Domack et al., 2005). This suggests that there was an intensification of the climate contrast between the two sides of the AP in the early Holocene, with a steepening of the thermal gradients to the north and west (Figure 3.27) (Hodgson et al., 2006b). This is backed up by data on the historical retreat of the Peninsula ice shelves as well as by differences in the timing of deglaciation of middle- to inner-continental shelf sites between the west (~13.3 ka B.P. at Palmer Deep and 15.7 ka BP at Lafond Trough) and the east of the Antarctic Peninsula (~10.6 ka BP at Erebus and Terror Gulf, 10.7 ka B.P. at Greenpeace Trough, and 10.5 ka BP at Larsen B embayment, all using conventional ^{14}C ages) (Domack et al., 2005; Heroy and Anderson, 2005). According to this hypothesis, the earlier deglaciation of the northern and western side may have made the glacial system more susceptible to the advection of warmer ocean currents. This is consistent with the evidence that at least some ice shelves there retreated in periods of early and mid-Holocene atmospheric and ocean warmth, while the thicker ice shelves on the east, such as Larsen B Ice Shelf, remained buffered against these warm periods.

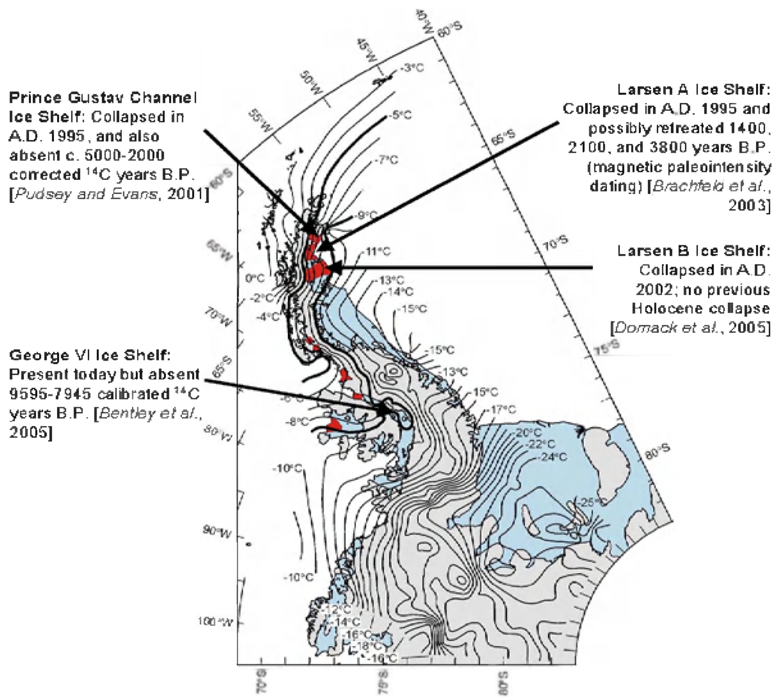


Figure 3.27 Recent retreats of Antarctic Peninsula ice shelves shown together with the –9ºC mean annual isotherm that marks the southern limit of ice shelf stability. Ice shelves that have retreated or collapsed are marked in red and extant ice shelves in blue. Locations of ice shelves studied are marked together with known Holocene retreat events (Hodgson et al., 2006b).

Deglaciation of the currently ice-free regions similarly showed regional differences in time and duration. Sedimentation began from c. 9.5 ka BP onwards in newly exposed lake basins in the north-eastern part of the AP and some islands to the north (Ingólfsson et al., 1998; Jones et al., 2000; Ingólfsson et al., 2003). Coastal areas in Marguerite Bay and parts of the coast on King George Island in the South Shetland Islands, were ice-free immediately after the early Holocene climate optimum (c. 9.5 ka BP) and some lake basins began to accumulate sediments c. 9.5-9.0 ka BP (Mäusbacher et al., 1989; Schmidt et al., 1990; Hjört et al., 1998; Hjört et al., 2003; Bentley et al., 2005a), but other areas in the South Shetland Islands did not become free of ice until much later in the Holocene (Björck et al., 1996; Gibson and Zale, 2006). In general, on the west side of the AP significant glacier thinning and ice margin retreat continued until at least c. 7-8 ka BP (Bentley et al., 2006). The transition from glacial to interglacial conditions was broadly completed by around c. 6 ka BP, when most ice-free areas were colonized by biota (Ingólfsson et al., 2003), but Byers Peninsula on Livingston Island deglaciated as late as c. 5-3 ka BP (Björck et al., 1996).

Evidence for mid Holocene glacier readvances are present on some islands, such as Brabant Island after c. 5.3 ka BP, and northern James Ross Island around c. 4.6 ka BP (Hjört et al., 1997; Ingólfsson et al., 1998). The glacial expansion coincided with cold and arid conditions on land from 5 ka BP onwards as detected in peat and lake sediment cores (Björck et al., 1991a,b; Björck et al. 1996) and cold marine waters with extensive sea ice cover in the bays of King George Island (Yoon et al., 2000) and in Lallemand Fjord beginning at 4.4 ka BP and peaking at 3 ka BP (Taylor et al., 2001).

In summary, there seems to be a regionally different response along the western and eastern coast of the AP, with ice shelf collapse restricted mainly to the west during the early Holocene. In addition, while most East Antarctic oases and nunataks were ice-free at the beginning of the Holocene, different parts of the AP were still ice-covered, and some did not deglaciate until as late as c. 5-3 ka BP. There is some evidence for a mid Holocene glacier readvance, coincident with cold marine water and extensive sea ice cover in the coastal areas.

Mid Holocene warm period – Hypsithermal: It was not until the mid-Holocene that the next period of significant warmth occurred in the AP. This interval is reviewed in detail in Hodgson et al. (2004a). The best-dated records place it between either c. 3.2 to 2.7 ^{14}C ka BP (c. 3.5-3.2 to 2.9-2.7 ka BP) in the AP region (Björck et al., 1991a) or c. 3.3 to 1.2 ^{14}C ka BP (c. 3.6-3.4 to 1.2-0.9 ka BP) just to the north of the AP (Jones et al., 2000; Hodgson and Convey, 2005). This Mid-Holocene Hypsithermal (MHH) is detected as a period of rapid sedimentation, high organic productivity, and increased species diversity in lake sediments ranging from the South Shetland Islands (Schmidt et al., 1990; Björck S. et al., 1996) and James Ross Island (Björck et al., 1996) to Signy Island in the South Orkney Islands (Jones et al., 2000; Hodgson and Convey, 2005). Sites in the northern AP show increased amounts of South American pollen in lake sediments during this period (Björck et al., 1993). It has also been associated with collapse of the Prince Gustav Channel ice shelf in the northern AP between c. 5 and 2 ka BP (Pudsey and Evans, 2001), and fluctuations of the Larsen-A Ice Shelf between c. 4 and 1.4 ka BP (Brachfeld et al., 2003), while the Larsen B Ice Shelf remained stable. This suggests that the steepening of the thermal gradients to the north and west between the two sides of the AP is common to both the early and the mid-Holocene (Figure 3.27).

In summary, whilst there is widespread agreement on the presence of some sort of warm period in the mid-Holocene, the exact timing often varies by hundreds of years, either because the timing varied spatially, or because there are insufficient numbers of dates or dating uncertainty is high, implying that, in the AP region as in EA, there is a need for further well-dated, high resolution sedimentary records.

Past 2,000 years - Neoglacial cooling, the Little Ice Age and recent climate change: The end of the MHH was marked by colder climate conditions. Numerous studies have identified Late Holocene glacier advances but most are poorly dated or even undated. Some of the putative Neoglacial advances may belong to a Little Ice Age (see Ingólfsson et al., 1998 for review). There is good evidence that the Prince Gustav Channel Ice Shelf started to reform after c. 1.9 ka BP, but due to a variable and sometimes large reservoir effect (6,000 years), this date is far from certain (Pudsey and Evans, 2001). Around c. 1.4 ka BP, as the climate began to cool, the Larsen-A Ice Shelf reformed, but here as well, dating uncertainty is high (Brachfeld et al., 2003). Numerous well-dated biological proxy records in lakes and other sites show a temperature-related decline in production at about this time (Björck et al., 1991b; Jones et al., 2000; Hodgson and Convey, 2005). Following the MHH, Midge Lake (Livingston Island) records a gradually deteriorating environment with both warm and cold pulses (Björck et al., 1991a). There was one warm event at c. 2 ka BP, and conditions were generally colder than present between c. 1.5 ka BP and 0.5 ka BP. Lake Åsa (Livingston Island) shows a distinct climate deterioration, with cold, dry conditions starting at c. 2.5 ka BP and continuing until close to the present day (Björck et al., 1993). Penguin populations declined between c. 1.3 to 0.9 ka BP and from c. 2.3 to 1.8 ka BP on Ardley Island (Sun et al., 2000; Liu et al., 2006).

Various outlet glaciers or ice shelves such as Rotch Dome, Livingston Island (Björck et al., 1996), and the Muller Ice Shelf (Domack et al., 1995) are thought to have advanced during a period roughly corresponding to the Northern Hemisphere Little Ice Age. However, the precise timing of those advances is well-constrained at only a few sites, and many of the terrestrial records of glacier advances are as yet undated. There is limited evidence of a LIA from lake proxy evidence. Liu et al. (2005) do however show a decline in penguin populations on Ardley Island, South Shetland Islands between 1790 to 1860 AD.

Instrumental measurements show the spatial pattern and magnitude of the recent rapid regional warming, and in particular the pronounced contrast between west (more warming) and east (less warming) sides of the AP. In proxy records, the warming is seen in increased sediment accumulation rates in some AP lake cores (Appleby et al., 1995), and some high-resolution marine cores (e.g. Domack et al., 2003b). Warming was further detected in a monitoring study of lakes in Signy Island where an increase in air temperature resulted in a significant increase in the amount of ice-free days and 4-fold increase in chlorophyll *a* content, which approximates lake productivity (Quayle et al., 2002, 2004). Few studies have focussed on this period in the proxy records.

In summary, climate conditions probably deteriorated after the Mid Holocene Hypsithermal, coincident with glacier readvance in some regions, yet these are poorly constrained in terms of dating; this is also the case for glacial readvance during the period of the Little Ice Age. Recently rapid climate warming has been observed in various regions of the AP, with glacier fronts retreating (Cook et al., 2005) and the lakes on Signy Island showing a remarkably rapid and magnified response in ecosystem functioning.

3.4.3.3 Ross Sea Region, Victoria Land Coast and Transantarctic Mountains (RS)

Deglaciation history the Pleistocene-Holocene transition: A grounded ice sheet fed from the Ross Embayment filled McMurdo Sound at the LGM and remained at its LGM position until c. 12.7 ^{14}C ka BP (c. 14.7-15.2 ka BP). The following ice recession was slow until c. 10.8 ^{14}C ka BP (c. 12.7-12.9 ka BP) (Denton and Marchant, 2000; Hall and Denton, 2000) (Figure 3.28). The Piedmont Glacier was probably also at its LGM position at c. 10.7 ^{14}C ka BP (c. 12.6-12.8 ka BP) (Hall and Denton, 2000). The grounded ice sheet blocked several valleys in the McMurdo Dry Valleys and fed large proglacial lakes such as Lake Washburn and Lake Wright, which had water over 500 m deep in some valleys (Hall and Denton, 1995) and existed until the early Holocene (Hall et al., 2000; Hendy, 2000; Hall et al., 2001). The large amount of water in these lakes was probably derived from melting glaciers as a result of dry and cold conditions, which led to decreased snowfall and lower albedo values (Hall and

Denton, 2002). Between c. 14 ka and 8 ka BP these large proglacial lakes started to evaporate (Hendy, 2000); the evaporation of Lake Washburn started during the LGM. Lake level lowering was discontinuous, with a series of high and low stands (Wagner et al., 2006). Significant changes evident in the geochemistry of a sediment core from Lake Fryxell between c. 13.5 and 11 ka BP suggest a further desiccation event at the Pleistocene-Holocene transition (Wagner et al., 2006).

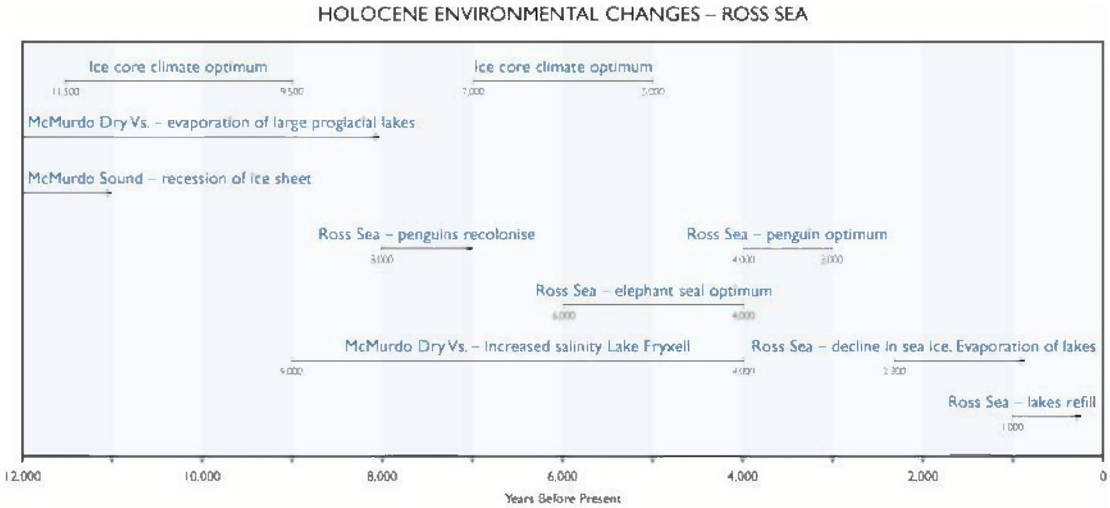


Figure 3.28 Selected Holocene environmental changes – Ross Sea

After the early Holocene: The last remnants of grounded ice in Taylor Valley post-date c. 8.4 ^{14}C ka BP (c. 9-9.4 ka BP), and penguins did not recolonize the RS region until c. 8 ka BP, after an absence of c. 19 Ka (Hall et al., 2006; Emslie et al., 2007). The period from c. 8 ka BP shows some discrepancies. Evidence from relict deltas for an increase in the moisture supply at c. 6 ^{14}C ka BP (c. 6.8 ka BP, Hall et al., 2000)) contrasts with evidence for increased salinity between c. 9 and 4 ka BP in a sediment core from Lake Fryxell (Wagner et al., 2006). This discrepancy could be explained by dating uncertainties or the relict deltas being from smaller local lakes (Wagner et al., 2006).

The presence of hairs from Southern Elephant Seals along with the remains of Adélie Penguins between c. 6 and 4 ^{14}C ka BP, indicates less sea ice than today, but sufficient pack ice for penguins to forage during spring (Emslie et al., 2007). The final retreat of the Ross Ice Sheet and deglaciation is estimated to have occurred around that time; c. 5.4 ^{14}C ka BP (c. 6.3 - 5.9 ka BP, Hall and Denton, 2000), coincident with the start of the recession of parts of the Wilson Piedmont Glacier at c. 5.5 ^{4}C ka BP (c. 6.4-6.27 ka BP) that lasted until at least c. 4.4-3.1 ^{14}C ka BP (c. 5.2-4.8 - 3.5-3.3 ka BP, Hall and Denton, 2000)). This corresponds with a secondary climate optimum detected in the ice cores of the RS sector, between 7 and 5 ka BP (Masson et al., 2000), and at Siple Dome c 5.2 ka (Figure 3.17).

The Mid Holocene: In Lake Fryxell well-developed microbial mats occurred from c. 4 ka BP onwards, which indicates similar environmental conditions and water depths to those found today (Wagner et al., 2006). Between c. 4 and 2.3 ^{14}C ka BP (c. 4.8-4.4 and 2.6-2.3 ka BP) elephant seals are completely absent from the region, whereas the Adélie Penguin population increased between c. 4 and 3 ka BP (Baroni and Orombelli, 1994; Emslie and Woehler, 2005). These data suggest that there was sufficient pack ice, but less than today, implying that conditions were relatively warm (Hall et al., 2006). This so-called ‘penguin optimum’ was followed by a period of high lake levels between c. 3 and 2 ^{14}C ka BP (c. 3.3-2.9 and 2-1.7 ka BP) in water bodies fed by meltwater from alpine glaciers (e.g. Lake Vanda, Hendy, 2000).

The Wilson Piedmont Glacier was less extensive than at present from c. 3.1 ^{14}C ka BP onwards until c. 0.9 ^{14}C ka BP (c. 3.5-3.2 and 1.1-0.8 ka BP; (Hall and Denton, 2000)).

The past 2,000 years – Late Holocene warm period and recent rapid climate change: In contrast to the AP, the warmest conditions in the RS region during the Holocene did not occur during the mid Holocene but rather during the late Holocene, as evidenced by the presence of elephant seal hairs. The warmest period of the past 6,000 years occurred between c. 2.3 and 1.1 ka ^{14}C BP (c. 2.6-2.3 and 1.2-0.9 ka BP) accompanied by the greatest decline in sea ice, as evidenced from an expansion of the elephant seal colonies (Hall et al., 2006), and substantial abandonment of penguin sites (Emslie et al., 2007). This period was followed by a period of enhanced evaporation, which lasted until 1 ka BP, when Lakes Fryxell, Vanda and Bonney evaporated to ice-free hypersaline ponds by c. 1.2-1 ka BP (Lyons et al., 1998; Wagner et al., 2006), and when Lake Wilson, a perennially ice-capped, deep (>100 m) lake further South (80°S) in southern Victoria Land, similarly evaporated to a brine lake (Webster et al., 1996). After 1 ka BP, warmer and wetter conditions led to increasing water levels and primary production in the lakes (Lyons et al., 1998; Wagner et al., 2006). This has been attributed to higher summer temperatures or to an increase in the number of clear, calm and snowless midsummer days (Hendy, 2000). During the last few centuries (< c. 0.2 ^{14}C ka BP, c. 0.4-0.1 ka BP) the Wilson Piedmont Glacier was more extensive than today in some areas until AD 1956 (Hall and Denton, 2002). There is no sign in the McMurdo Dry Valleys of the pattern of glacier advances typical of the Northern Hemisphere Little Ice Age (Hall and Denton, 2002).

Studies in the framework of the US Long Term Ecological Research program have revealed a rapid ecosystem response to local climate cooling in the McMurdo Dry Valleys during recent decades, as evidenced by a decline in lake primary production and declining numbers of soil invertebrates (Doran et al., 2002).

Summary: In general, geological evidence shows that deglaciation of the currently ice-free regions was completed earlier in EA compared with the AP, but all periods experienced a near-synchronous early Holocene climate optimum (11.5-9 ka BP). Marine and terrestrial climate anomalies are apparently out of phase after the early Holocene warm period, and show complex regional patterns but an overall trend of cooling. A warm mid Holocene Hypsithermal is present in many ice, lake and coastal marine records from all three geographic regions, although there are some differences in the exact timing. In EA and the AP (excluding the northernmost islands) the Hypsithermal occurs somewhere between c. 4 and 2 ka BP, whereas at Signy Island it spanned 3.6-3.4 – 0.9 ka BP. Despite this there are a number of marine records that show a marine-inferred climate optimum between about 7-3 ka BP and ice cores in the RS sector that show an optimum around 7-5 ka BP, and the EPICA Dome C ice core, and some others, show a weak optimum between 7.5 and 3 ka BP. The occurrence of a later Holocene climate optimum in the RS is in phase with a marked cooling observed in ice cores from coastal and inland locations (Masson et al., 2000; Masson-Delmotte et al., 2004). These differences in the timing of warm events in different records and regions points to the interplay of a number of mechanisms that we have yet to identify. Thus there is an urgent need for well-dated, high resolution climate records in coastal Antarctica and particularly in the Dronning Maud Land region and specific regions of the AP to fully understand these regional climate anomalies and to determine the significance of the heterogeneous temperature trends being measured in Antarctica today. There is no geological evidence in Antarctica for an equivalent to the Northern Hemisphere Medieval Warm Period, there is only weak circumstantial evidence in a few places for a cool event crudely equivalent in time to the Northern Hemisphere's Little Ice Age but not in phase (Goosse et al., 2004), and it is likely beyond the current signal to noise ratio of ice cores to detect the c. 0.5°C change believed to have occurred at that time.

3.5 Biological responses to climate change

This section reviews the evolutionary and biogeographical factors that have shaped the contemporary terrestrial and marine biota of the Antarctic continent. It includes descriptions of some of the physiological adaptations in the marine realm that have led to the distinctive marine fauna seen today. This section also highlights some of the reorganisations in terrestrial and marine species distributions that have occurred during the relatively minor natural climate changes of the Holocene that are being used to inform our understanding of the climate and biological changes that might be anticipated in the near future (Chapter 5).

The fossil record, stretching back over 500 million years, provides a broad outline of evolutionary history of the continent and its biota. The first signs of temperate biotas, marine and terrestrial, were in the middle Devonian (ca. 375 Ma). During the Cretaceous, the highly seasonal, then warm, high-CO_2 climate and unstable landscape of high-latitude Gondwana may have been the centre of origin for gymnosperm and angiosperm taxa that subsequently spread northward, providing sources for temperate southern hemisphere floras. Environmental change, including climatic change, in Antarctica and its surrounding waters has been dramatic since separation of the other southern hemisphere continents from Antarctica during the Cretaceous and Tertiary. It is possible that some elements of the living marine biotas can be traced back to the Early Cretaceous period (130 Ma). The earliest cold climate marine faunas are thought to be latest Eocene-Oligocene (ca. 35 Ma) in age. Conditions on land fluctuated greatly between cold and warm during the Tertiary, and terrestrial floras changed accordingly. There is local high diversity of invertebrates on the deeper shelf, where environmental conditions are relatively homogenous so that distributional limits of species assemblages cannot be explained only by ecological demands. The "climate diversity pump", a variation of the refugio-survival model, is based on surges of the Antarctic ice cap to the coast and sometimes as far as the shelf break. These episodes are first documented from the Late Eocene (40 Ma) and have continued into the Pleistocene. It is possible that they have acted as a barrier for genetic exchange. During such periods of the separation of populations these continued to evolve and mixed again when the ice retreated but interbreeding was not possible anymore. As a consequence, more species with broadly overlapping ecological niches can be expected after the exposure to large scale disturbances. This might be especially relevant for filter feeders among which competition is not highly effective to reduce the number of species and diversity. Our knowledge of the relevance of the "climate diversity pump" can only improve when glaciological and sedimentological findings are combined with those of ecological surveys using statistics and genetics and a sampling design adapted to this specific question.

3.5.1 The terrestrial environment

With the exception of lake sediment studies (Hodgson et al., 2004a), little terrestrial research has set out to examine changes in biodiversity, distributions and abundance over Holocene and longer timescales. There has been a widely held but untested general assumption that, owing to the much more extensive and thicker ice sheets present on the continent at LGM, many if not all contemporary Antarctic terrestrial biota must be recent colonists. That this is not the case has come to recent prominence (Convey and Stevens, 2007; Convey et al., 2008; Pugh and Convey, 2008), and it is becoming clear from biogeographical and molecular evidence that across the continent and also the sub-Antarctic islands the contemporary biota is a result of vicariance and colonization processes that have taken place on all timescales between pre-LGM and pre-Gondwana-breakup (e.g. Stevens et al., 2006). Nevertheless it is also clear that much of the tiny proportion of Antarctica that is ice-free today has been exposed over only the last few thousand years during post LGM glacial retreat. Through the maritime Antarctic and much of the continental coastline, most exposed ground takes the form of 'islands' of terrestrial habitat of varying size at low altitude and close to the coast, surrounded by either hostile sea or ice (Bergstrom and Chown, 1999). Exceptions to this generalization are provided first by the continental Antarctic 'Dry Valleys' of Victoria Land, providing several thousand square kilometres of ground at least some of which has been continuously exposed since about 12 Ma in the Miocene, and second by inland higher altitude nunataks and mountain ranges, some of which will not have been covered at Pleistocene glacial maxima (Stevens and Hogg, 2003; Stevens et al., 2006).

Taking the maritime Antarctic as an example, it is thus clear (a) that the large majority of areas of currently ice free ground have been exposed post LGM (while longer term refugia are required to explain contemporary biota distributions, their precise locations remain unknown (Hodgson and Convey, 2005; Pugh and Convey, 2008)), and at the same time (b) it is clear that most elements of the regional biota have successfully colonized those areas that have been exposed, and done so rapidly, as their terrestrial communities are in most cases entirely typical of this regional biota. Hodgson and Convey (2005), using terrestrial arthropod abundances obtained from lake sediment cores on maritime Antarctic Signy Island, identified some differences in relative abundances of two common mite species over time (the last 5+ ka) that they proposed to be consistent with climatic changes that would have altered the balance of the different habitats that these species favour. On a longer pre-LGM Pleistocene timescale, Hodgson et al. (2005) have described changes in lake diatom communities in some continental East Antarctic lakes proposed to have survived intact throughout the LGM period. They provide evidence that sub-Antarctic diatom taxa present during the last interglacial period were lost from the community as the LGM approached, leaving only continental taxa, and that the sub-Antarctic taxa have not yet returned to the lakes post LGM.

Liquid water and ice-free refugia during ice ages have meant long availability of habitat for some of the biota (e.g. mites, springtails, chironomids), even extending back to the Gondwana era (Allegrucci et al., 2006; Hodgson et al., 2006a; Convey et al., 2008) (Figure 3.29). Nevertheless, the expansion and contraction of the Antarctic ice sheets has undoubtedly led to the local extinction of biological communities on the Antarctic continent during glacial periods (Hodgson et al., 2006a). Subsequent interglacial re-colonisation and the resulting present-day biodiversity is then a result of whether the species were vicariant (surviving the glacial maxima in refugia, possibly also requiring them to take advantage of diachronous extension and retreat of ice in particular areas, then recolonising deglaciated areas), arrived through post-glacial dispersal from lower latitude islands and continents that remained ice free (Pugh et al., 2002), or are present through a combination of both mechanisms. Evidence can be found to support both vicariance (Marshall and Coetzee, 2000; Stevens and Hogg, 2003; Hodgson et al., 2005; Allegrucci et al., 2006; Hodgson et al., 2006a; Stevens et al., 2006; Stevens and Hogg, 2006) and dispersal (Hodgson et al., 2006a) for a variety of different species, and is based on the level of cosmopolitanism (dispersal model) or endemism (vicariance model) (Gibson and Bayly, 2007), on direct palaeolimnological evidence, or, most recently, on molecular phylogenetic and evolutionary studies (Marshall and Convey, 2004; Allegrucci et al., 2006; Stevens et al., 2006; Stevens and Hogg, 2006).

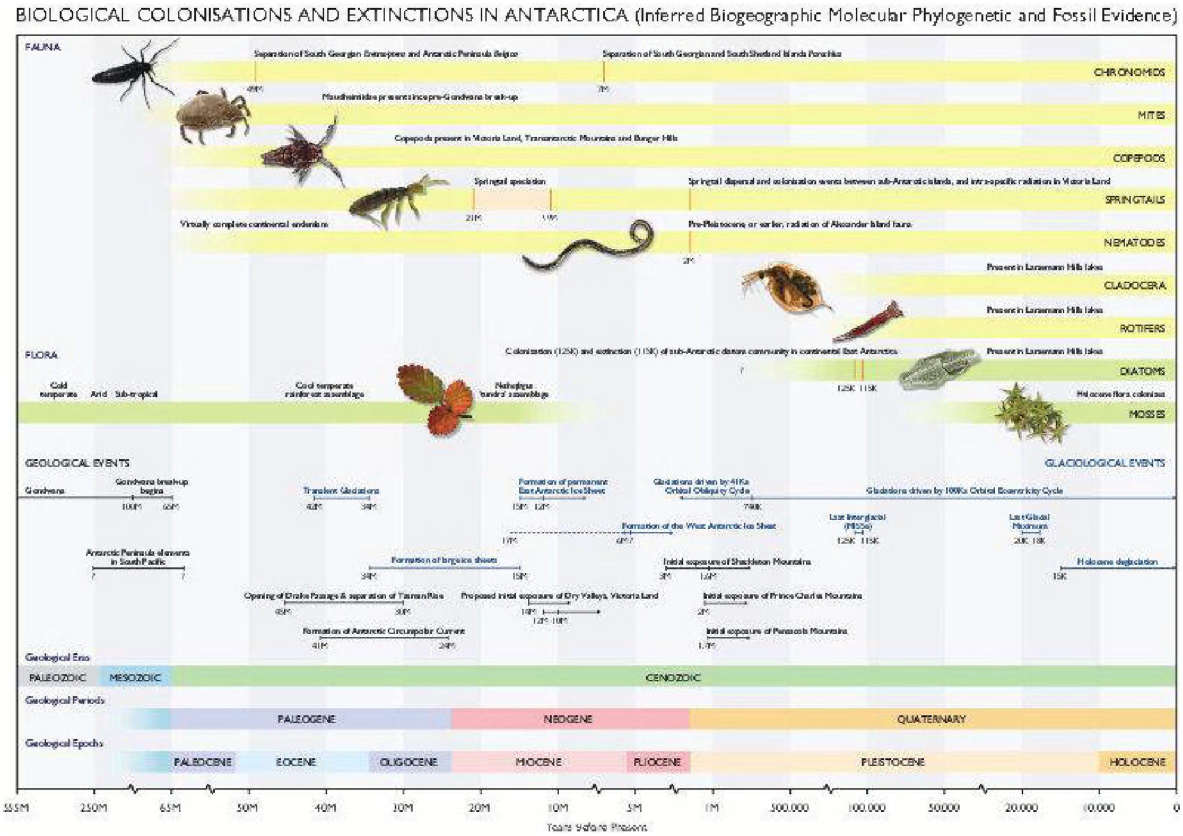


Figure 3.29 Biological colonisations and extinctions in Antarctica since the break-up of Gondwana; based on molecular, phylogenetic, fossil and sub-fossil evidence. The upper panel shows schematic timelines for the survival of different species on the continent. The lower panel shows the major geological and glaciological events in the evolution of Antarctica that will have influenced the flora and fauna. Note that the geological timescale is non-linear and that most micro-organisms are excluded from this schematic diagram

On the oceanic islands, the biotas must have originally arrived via long-distance over-ocean dispersal, with vicariance and terrestrial dispersal playing subsequent roles in shaping the biodiversity across glacial cycles (Marshall and Convey, 2004; Stevens et al., 2006). Species on Southern Ocean islands show conventional island biogeographic relationships, with variance in indigenous species richness explained by factors including area, mean surface air temperature, and age and distance from continental land masses (Marshall and Convey, 2004). For aquatic species, at least for some groups such as the diatoms, diversity is controlled by the 'connectivity' among habitats with the more isolated regions developing greater degrees of endemism (Vyverman et al., 2007; Verleyen et al., 2009).

3.5.1.1 Changing species distributions, abundance and biodiversity

Through the Holocene the changing environmental conditions described in section 3.4.3 have caused marked changes in species distributions particularly for those species that have well defined ecological ranges. This is best recorded in palaeolimnological studies where preserved morphological and geochemical fossils provide a detailed record of changing species compositions in response to changes in lake water chemistry and other environmental variables at many sites around the continent.

The historical record of changing terrestrial species distributions is more sparse. Much of our knowledge is based on changes that have been recently been observed in the Antarctic Peninsula region where increasing temperatures in the last 50 years have resulted in the local expansion of population ranges of a number of plant and animal species, and the establishment (albeit with human assistance) of new species that appear not to have survived on the continent before (Frenot et al., 2005; Barnes et al., 2006).

The growth and life cycle patterns of many invertebrates and plants are fundamentally dependent on regional temperature regimes and their linkage with patterns of water availability. Distinct patterns in sexual reproduction are evident across the Antarctic flora and are most likely a function of temperature variation. In addition, phenology (the study of periodic plant and animal life cycle events and how these are influenced by seasonal and interannual variations in climate) of flowering plants is cued to seasonality in the light regime. In regions supporting angiosperms, wind is assumed to play a major role in the pollination ecology of grasses and sedges, resulting in cross-pollination. The lack of pollinators in the native fauna, combined with high reproductive outputs in non-wind pollinated species implies a high reliance on self-pollination.

The Antarctic biota shows high development of ecophysiological adaptations relating to cold and desiccation tolerance, and displays an array of traits to facilitate survival of these conditions. While patterns in absolute low temperatures are clearly influential in determining survival, perhaps more influential are the patterns of sub-lethal environmental stresses experienced and the freeze-thaw regime, with repeated freeze-thaw events being more damaging than a sustained freeze event. How these patterns change in the future will be an issue of major importance to ecosystems.

In contrast to many Antarctic marine organisms, the terrestrial biota often has a wide environmental tolerance. It includes some of the most robust life forms on Earth, the cyanobacteria, which can survive extremes of low temperature, water availability, light and high UV radiation (Hodgson et al., 2004b). These are particularly abundant in extreme habitats, such as parts of the Transantarctic Mountains, where they have no or few competitors. Other groups, however, do have well defined ranges within which they can survive.

It is already well known that Antarctic terrestrial biota possess very effective stress tolerance strategies, in addition to considerable response flexibility. The exceptionally wide degree of environmental variability experienced in many Antarctic terrestrial habitats, on a range of timescales between hours and years, means that predicted levels of change in environmental variables (particularly temperature and water availability) are often small relative to the range already experienced. Given the absence of colonisation by more effective competitors, predicted and observed levels of climate change may be expected to generate positive responses from resident biota of the maritime and continental Antarctic. The picture is likely to be far more complex on the different subantarctic islands, and many already host (different) alien invasive taxa, some of which already have considerable impacts on native biota (Frenot et al., 2005).

3.5.2 The marine environment

Although the fossil record of the Antarctic marine faunas is far from complete, there is growing evidence to suggest that a distinctive, temperate, shallow benthos has characterised the southernmost high latitudes since at least the late Palaeozoic Era (Crame, 1994). Such fauna, which were not necessarily continuous through time, show two consistent features: they are invariably less diverse than their lower latitude counterparts (Crame, 2001), and they are characterised by bipolar elements (Shi and Grunt, 2000). By the Late Cretaceous the Antarctic shallow marine fauna formed part of a distinctive Weddellian Province that could be traced around the southern Gondwana margins from Patagonia, through West Antarctica, to New Zealand and south-eastern Australia (Zinsmeister, 1982). There is no doubt that it was affected by the mass extinction event at the end of the Cretaceous, but precisely to what extent is still uncertain. The late Paleocene-early Eocene fossil record of Antarctica is incomplete, but by the Middle Eocene it is apparent that benthic molluscan diversity had more than recovered to its pre-mass extinction levels. The prolific middle-late Eocene La Meseta Formation of Seymour Island has now yielded in excess of 170 molluscan taxa, with more than half of these being bivalves (Stilwell and Zinsmeister, 1992). Thereafter shallow-marine benthic molluscan diversity was substantially reduced to its present day levels. Whether this was a gradual process or perhaps punctuated by further mass extinctions is unknown. Bivalves appear to have been particularly badly hit in the intervening 35 Ma and it is possible that this happened for a variety of reasons.

The broad history of continental glaciation in Antarctica is reasonably well established (Barrett, 2008). However, the details that are important to an understanding of the evolution of the Antarctic marine fauna are still lacking. The important question in terms of the evolutionary history of the Southern Ocean shallow water marine fauna is to what degree fluctuations in sea level, and the extent of the continental ice-sheet, have driven changes in the depth and area of habitat on the continental shelves around Antarctica (Clarke and Crame, 1989). Although there is clear geophysical evidence for extensions of the ice-sheet having reduced considerably the area of continental shelf at least once in the past, we cannot yet say how often this has occurred in the past or how widespread these extensions might have been (see Convey et al., submitted.).

What we can assume from these constraints for the evolution of Antarctic species is that natural climate change has always played an important and multi-fold role. The genesis of the circumpolar current together with the cooling of the Southern Ocean isolated most populations from those living farther north, creating a state in which Antarctica is sometimes considered an "evolutionary incubator". This isolation was associated with an increase in invertebrate diversity, but other factors were also involved. First of all, some animal groups became extinct during the cooling and left ecological niches unoccupied. Others survived with very few representatives that provided the seeds to radiate into many new species. Other groups were pre-adapted, e.g. isopods and amphipods, by their breeding behaviour, but they continued to evolve and adapted to the new Antarctic habitats and food sources. The glaciation of the continent prevented terrestrial fluvial river run-off and, consequently, supported in many places the growth and development of filter feeders. Despite the availability of some information about the food preferences and life cycles of selected species, e.g. amphipods, our knowledge of the occupation of specific ecological niches or, alternatively, broadly overlapping environmental demands, especially among filter feeders, is still very poor. The slow growth of filter feeders, encouraged among other factors by the low temperature, can explain why rather few species might have been historically outcompeted (Gutt, 2006). As a consequence, modern filter feeders are rich in species, provide microniches for a species associated mobile fauna and large benthic predators are comparatively scarce. This image of the modern Antarctic shelf benthos was interpreted as the result of a mainly convergent development of Palaeozoic and modern communities (Gili et al., 2006) and reminds regionally due to its 3-D structure that of a modern coral reefs. During glacial maxima, most species that survived on the shelf or upper slope, especially those being dependant on phytodetritus, had to adapt to a low food supply (Bonn et al., 1998). During summers of interglacial periods these animals might experience a clear food surplus. The moderate number of benthic species compared to global numbers could be explained by this switching back and forth from conditions with low food supply, creating high diversity, to conditions with a high food supply, supporting high competition pressure.

Another effect of the cooling of the Antarctic continent and surrounding waters is a similarity of environmental conditions and, consequently, of the fauna, between habitats on the Antarctic shelf, at its continental slope and in the adjacent deep-sea. Independently of whether deep-sea animals colonized the Antarctic shallow waters or vice versa, this long-term dispersal shows that Antarctic benthic life is not as isolated from northerly adjacent areas as is the open ocean ecosystem by the Antarctic Convergence.

3.5.2.1 Adaptation of algae

Macroalgae from the Antarctic show an adaptation to considerably lower temperatures compared with Arctic species (Wiencke et al., 1994). This difference is the result of the different times of exposure of these groups to low temperatures, 14 million years in the Antarctic compared with 3 million years in the Arctic. The northern distribution limit of endemic Antarctic species is mainly set by the temperature requirements for macroscopic growth (Wiencke and tom Dieck, 1989). In the ecologically important brown macroalgae (Desmarestiales), the northern distribution limit is determined by the temperature requirements for growth of the macroscopic stages (sporophytes). These species are only found in areas south of the Antarctic convergence with temperatures < 5°C. During the ice ages the coastline of the Antarctic continent was probably mostly inhospitable for macroalgae, and suitable refuges were probably provided by some of the sub-Antarctic islands and/or southernmost South America. At the LGM the northern boundary of the Antarctic region just touched South America. However, habitats as far south as the South Shetland Islands may have served as refuges (Bischoff-Bäsmann and Wiencke, 1996). Species with relatively high temperature demands were able to extend their northern distribution limit to lower latitudes during the ice ages. Species or species pairs with a bipolar distribution such as *Desmarestia viridis/confervoides* (Peters and Breeman, 1992), *Acrosiphonia arcta* (Bischoff and Wiencke, 1995b) and *Urospora penicilliformis* (Bischoff and Wiencke, 1995a) probably drifted across the equator during the Pleistocene decline in water temperatures in the tropics (Wiencke et al., 1994). The most resistant developmental stages of these species tolerate 25-27°C, temperatures slightly above the minimum equatorial sea surface temperatures during the last glaciation (23-25°C). Ice recrystallization inhibition proteins (IRIPs) and freeze tolerance have recently been discovered in the cryophilic Antarctic hair grass by John et al. (2009) and ice-binding face of a plant antifreeze protein e.g. by Middleton et al. (2009) and Janech et al. (2006).

3.5.2.2 The evolution of Antarctic fish

Much of our knowledge of the effects of the environment on vertebrate physiology and evolution has come from fish, which share many basic physiological mechanisms with humans. The close physical and physiological interaction with the aquatic environment makes them sensitive sentinels of environmental challenge and offers important advantages for defining the organism-environment interface and the mechanisms of temperature adaptation. Fish have developed cellular and molecular mechanisms of cold adaptation, and these are fully representative of the suite of strategies adopted by organisms under strong evolutionary pressure. In the course of evolution, Antarctic fish have developed specialised adaptations, some of which characterise these organisms as unique. In strong contrast to the continental shelf faunas elsewhere, the fish fauna of the Southern Ocean around Antarctica is also unique in being overwhelmingly dominated by the single, highly endemic group of Notothenioidei. From many viewpoints Notothenioidei are the best characterised group of fish in the world. The dominance by a single taxonomic group of fishes provides a simplified natural laboratory for exploring the wealth of their physiological, biochemical and ecological adaptations, and of their evolution. Understanding the patterns of adaptation can tell us much about the process of evolution, and on the impact of climate change on a highly specialised group of fish. Notothenioids are the best example of adaptive radiation by a vertebrate group in the sea; they provide a truly unique opportunity to study the evolution of a single group of fishes in a known thermal and tectonic context, and also to study reactions of highly specialised organisms to anthropogenic changes. The amount of information available on cold adapted polar fish will provide invaluable clues on the development, impact and consequences of climate challenges, with powerful implications for the future of the Earth.

A suborder of Perciformes, Notothenioidei comprise 8 families with 44 genera and 129 species (Eastman, 2005). Notothenioids represent 35% of all species in the Southern ocean and 76% of all species in the shelf waters of Antarctica (90-95% of the fish biomass). The majority of the species are endemic to the Antarctic waters, where they have successfully diversified into several ecological niches. As for other examples of adaptive radiation, this was made possible by the isolation of Antarctic coastal waters and its separation from other continents by large and deep water masses, with no shallow water connections (undersea mountain ridges or plateaus). The presence of the Antarctic Polar Front, which acts as an oceanographic barrier (albeit perhaps "leaky" (Clarke et al., 2005)), further reduces the exchanges between Antarctic and sub-Antarctic waters. As described earlier, this isolation was established after the opening of the Drake Passage (Scher and Martin, 2006) and the Tasman gateway (Stickley et al., 2004).

The subzero water temperatures and the progressive extension of ice sheets likely determined the extinction of most of the temperate fish fauna, leaving space to those species that evolved some adaptation to the freezing conditions. Antarctic notothenioids are stenothermal, and their ability to cope with the ongoing increases in environmental temperatures might be reduced, due to losses in the level of temperature-mediated gene expression, including the absence of a heat-shock response (Hofmann et al., 2005; Somero, 2005) described above. Thus, the question to what extent Antarctic fish may adapt to environmental change becomes a very important issue.

Three mechanisms will be mentioned below in some detail: the first two of which are development of antifreezes and hematological features. Polar fish are the only vertebrates endowed with these two specialisations. Both have required costly and complex anatomical, ecological, physiological and biochemical adjustments and compensations, and both have tight links with the temperature of the environment. Hence warming, albeit small, may have a significant impact. The third mechanism is ecological adaptation to the new habitat.

Another physiologically relevant peculiarity of Antarctic fish is the high content of mitochondria in slow muscle fibres, towards the upper end of the range reported for teleosts with similar lifestyles, and up to 50% higher in Channichthyidae (Johnston, 2003). The phyletically basal bovichtids, pseudaphritids and eleginopids do not possess antifreeze glycoprotein gene sequences in their genomes (Cheng et al., 2003).

(1) Antifreeze compounds in polar fish: Most invertebrate species have body fluids whose osmotic pressure is the same as seawater so that they do not freeze unless the water around them freezes. Freezing is generally unlikely for organisms inhabiting depths more than a few metres below the intertidal. Fish, on the other hand, have body fluids less concentrated than seawater, usually around 300–600 mOsm per litre, and must invest the costs associated with production of antifreeze glycoproteins (AFGPs) year-round to avoid freezing (Devries, 1971; 1982; Clarke, 1998; Cheng and Chen, 1999). The importance of this attribute is underlined by the large number of copies of antifreeze genes in the genome of Antarctic notothenioids (Wang et al., 1995; Cheng and Detrich III, 2007), and the fact that the trait has evolved on several separate occasions (Chen et al. 1997a,b; Cheng and Chen, 1999) and meets the criteria for a "key innovation" (Eastman, 2005). Biochemical and cellular adaptations showing strong temperature compensation of functional capability have been identified in polar marine species, and these include changes in enzyme isoforms or activation energies (Clarke, 1998; Vetter and Buchholz, 1998), adaptation of rate of microtubule assembly (Detrich et al., 1989), and mitochondrial proliferation in red muscles of polar fish (Johnston et al., 1998). These adaptations all allow specific intracellular processes to proceed at rates similar to those in temperate species. However, these are rarely, if ever translated into whole-animal compensation, because metabolic activity and growth rates are predominantly slower in polar marine environments (Peck et al., 2002). That some cellular functions are cold-compensated and proceed at rates similar to those from other latitudes supports the argument that slow growth, development and metabolic rates are dictated by resource considerations (Clarke, 1991a,b), and that these characteristics all reduce ATP demand in the face of seasonally or ecologically reduced resource availability (Clarke, 1998).

At the evolutionary level, antifreeze compounds enabled the ancestral notothenioid to fill the many ecological niches made available by ice-driven extinctions. Indeed, estimates obtained using "molecular clock approaches" indicate that the main notothenioid diversification started 23-15 Ma (Bargelloni et al., 1994; Near, 2004), in parallel with the establishment of permanent sea ice.

(2) The molecular evolution of hemoglobins in polar fish: Specialised hematological features are among the most striking adaptations developed by the Antarctic ichthyofauna during evolution. Antarctic waters are cold and oxygen-rich. Oxygen binding is generally favoured at low temperature. The metabolic demand of fish for oxygen is relatively low, the solubility of oxygen in the plasma is high, but the energetic cost associated with circulation of a highly corpuscular blood fluid is also large (Wells, 1990; di Prisco et al., 1991; Eastman, 1993). Notothenioids have evolved reduced hematocrits, hemoglobin concentration/multiplicity and oxygen affinity. The growing knowledge of the phylogenetic relationships among the notothenioid families (Eastman, 1993; Bargelloni et al., 1994; Bargelloni et al., 2000a,b; Eastman, 2000; Near et al., 2003; Dettaï and Lecointre, 2004; Near, 2004; Near et al., 2004; Eastman, 2005; di Prisco et al., 2007, Near and Cheng, 2008; Negrisolo et al., 2008; Verde et al., 2008b,c; Giordano et al., 2009a,b) is producing compelling evidence to answer some questions about thermal adaptation. Certainly the identification of the sister group of the suborder will be necessary to understand how these species have achieved adaptations to temperature and how they have been affected by climate change in the late Eocene (38-35 Ma is the time of widespread continental glaciation and sharp drops in Southern Ocean surface temperatures (di Prisco et al., 1991; Eastman, 1993; Clarke, 1998). Bovichtidae, Pseudaphritidae, Eleginopidae, Nototheniidae, Harpagiferidae, Artedidraconidae, Bathydraconidae and Channichthyidae (icefishes) are the families of the suborder, thought to have arisen in Antarctica through adaptive radiation from the single ancestral stock. Seven families have hemoglobin (Hb)-containing erythrocytes in the blood, whereas Channichthyidae (the crown group) are devoid of Hb (Ruud, 1954; Cocca et al., 1995; Zhao et al., 1998). Bovichtidae (only one out of ten species is Antarctic) and monotypic Pseudaphritidae and Eleginopidae presumably diverged during the Eocene and became established in waters around areas that now correspond to New Zealand, Australia and high-latitude South America.

The globin-gene status in notothenioids, leading to the unique vertebrate specialisation in the hemoglobin-less family Channichthyidae (Figure 3.30), has been studied in detail (Cocca et al., 1995; Zhao et al., 1998; di Prisco, 2000; di Prisco et al., 2002; Hudson and Coyne, 2002; Near et al., 2006). Why have icefishes alone taken such a radical course leaving the other Antarctic families with only partial reductions in hemoglobin? Does hemoglobin remain absolutely vital for adequate oxygen transport in the other Antarctic notothenioids, or is it a vestigial relict which may be redundant under stress-free conditions? Are the hematological features of the modern families a result of life-style adaptation to extreme conditions? What is the sensitivity of this specialised oxygen-transport system to warming?

Figure 3.30 Icefishes: *Chionodraco hamatus*. Photo: G. di Prisco

The different phylogenetic histories of polar fish depend on the differences in the respective habitats. As a result of isolation, the genotype of Notothenioidei diverged with respect to other fish groups in a way interpreted as a giant species flock (Eastman and McCune, 2005). The Arctic ichthyofauna is thriving in a much more complex ocean system than the Antarctic one. Antarctic waters are dominated by a single taxonomic group; Arctic waters are instead characterised by high diversity. This is reflected in phylogeny, as shown for instance by the amino-acid sequences of globins (Eastman, 1997; Verde et al., 2002, 2006a,b,c; Giordano et al., 2007; Verde et al., 2007; Dettaï et al., 2008; Verde et al., 2008a,b,c; Verde et al. 2009).

(3) Heat shock proteins: Protein synthesis at low temperature may be difficult and hence costly, which may restrict the amount of useful protein available for growth (Fraser et al., 2002). This factor would again be a limitation of resource, or energy available for growth and development, rather than a direct limitation of rate by reduced temperature. Hofmann et al. (2000) have shown that there is no classic heat shock response in the Antarctic fish *Trematomus bernacchii*, although the position is more complex than this in other species. Some Antarctic marine invertebrates express heat shock genes when temperatures are elevated experimentally to over 10°C, although this is not something they ever experienced in nature (Clark et al., 2008a), whereas other invertebrates appear to be similar to the fish data reported by Hofman et al. (2000) in having no heat shock proteins (HSP) response to elevated temperature (Clark et al., 2008b). Surprisingly, in the context of laboratory studies, where there was no induction of response in the limpet *Nacella concinna* until 15°C, wild intertidal populations do show a heat shock response when emersed, even though foot temperatures do not exceed 3°C (Clark et al., 2008a). It is not known whether the lack of a heat shock response in some Antarctic species is due to the deletion or dysfunction of genes, instability of messenger RNAs, the absence of a functional heat shock factor, or some other character. The loss of this response is, however, a large energetic cost saving, and could only be successful in an environment with very stable temperatures over long evolutionary periods. Such cost-saving adaptations as the loss of a heat shock response contrast markedly with adaptations developed in the much more variable terrestrial environment.

(4) Ecological adaptation of notothenioid fish: Isolation and extinctions, together with the evolution of AFGPs and reduction of reliance on hemoglobin provided the opportunity

for notothenoids to radiate, but alone cannot explain the great number of species. Additional adaptations or modifications led to notothenioid radiation. The ancestral notothenioid was a benthic fish, yet acquisition of neutral buoyancy allowed the repeated colonisation of pelagic habitats (pelagic, epipelagic, cryopelagic). Even for truly benthic species, diversification occurred through partition of depth range, from very shallow waters (0-30 m) to great depths (2,950 m, *Bathydraco scotiae*). Variation in body size (from the few centimetres of *Pleuragramma antarcticum* to close to two metres of the large pelagic predator *Dissostichus mawsoni*) and diverse feeding habits further contributed to diversification among notothenioid species. Besides niche partitioning, habitat fragmentation provided the means for species divergence, as often observed in adaptive radiations. Recent studies have demonstrated reduced gene flow between populations of notothenioid fish distributed around the continent, even in the presence of homogenising circumpolar currents (Patarnello et al., 2003; Zane et al., 2006). Speciation was also promoted by the presence of sub-Antarctic islands and archipelagos (e.g. South Georgia, Kerguelen Islands).
A clear example of how all the above described factors shaped the evolution of notothenioids is provided by Chen et al. (1998). These authors mapped ecological and geographic distributions of the species belonging to a notothenioid family, Channichthyidae, onto a molecular phylogenetic tree. It emerged that speciation was always associated either with a shift in ecological habits (feeding behaviour, depth range) or with disjunct geographic distribution. The evolution of notothenioids represents an extraordinary example of adaptive radiation in the marine environment, with all the canonical characteristics of this mode of speciation (isolation, mass extinctions, key adaptations, ecological shifts and habitat fragmentation).

3.5.2.3 Changes in the marine ecosystem in the Quaternary

Life on the Antarctic shelf in the Quaternary has altered in response to varying degrees of cooling, ice shelf dynamics, isolation and changing oceanography. The major impacts of climate change on glacial timescales in the marine environment have been the glacial-interglacial expansion and contraction of the Antarctic ice sheet across the continental shelf, the consequent loss and recovery of benthic marine habitats, and the interglacial fluctuations in summer and winter maximum sea ice extent. Direct evidence for this can be seen in the marine geological record from the near shore continental shelf off the Windmill Islands (66°S, 110°E), in EA, where the expanding ice sheet eliminated habitat, while the shrinking ice sheet led to recolonisation and succession (Hodgson et al., 2003). Maximum expansion of the Antarctic ice sheet to the continental shelf edge at the LGM was diachronous, about 90% of the continental shelf was covered by grounded ice shelves (Harris and O'Brien, 1996), though perhaps not all simultaneously. This ground was therefore unavailable to benthic species. Even so, at any given time refugia were most likely available. This is consistent with molecular evidence that a distinct Antarctic marine biota has survived on the continental shelf, or at the shelf break through multiple glacial cycles. Species richness increased significantly in this phase when populations became isolated during glacial maxima, as antifreeze glycoproteins permitted continued evolution, and, at the end of the glacial period when the populations mixed again, they were unable to interbreed. Such processes are generally termed vicariance events and are described specifically for the Antarctic as a climate-diversity pump (Clarke and Crame, 1989). Relatively high species numbers might also have been supported by low extinction rates, being the result of slow ecological processes reducing the likelihood of competitive displacement (Gutt, 2006). Wherever the benthos lived during glacial periods, its physical and biological environment differed significantly from today's, especially with regard to food supply (see Bonn et al., 1998). Only during interglacials were the sea-floor and the overlying water column suitable habitats for

rich benthic and pelagic communities (Gutt, 2007; Clarke et al., 2004). Within the present Holocene interglacial, physical conditions have been comparatively stable.
Nevertheless, even on Holocene timescales, marine mammal and seabird distributions have been affected by warm periods and expansions and contractions of the sea ice (Hall et al., 2006) ice shelves and possibly epidemic viruses (Nelson et al., 2008). Changes in the Holocene distribution of marine birds can be tracked through the changing distributions of their nesting sites (Emslie and Woehler, 2005; Emslie et al., 2007). Peaks in penguin populations correlate with a period of open water associated with a warm period in the RS between 4 and 3 ka corr.^{14}C BP (Baroni and Orombelli, 1994). Extensive occupation by elephant seals shows the warmest period occurred there between c. 2.3 and 1.1 ka ^{14}C BP (c. 2.6-2.3 and 1.2-0.9 ka BP), correlated with a significant decline in sea ice (Hall et al., 2006). In the Bunger Hills, isotopic concordance between a marine sediment core and fossil stomach oil of snow petrels (mumiyo), and a significant correlation between mumiyo δD and δ^{13}C, suggest that past δ^{13}C variation in plankton was transferred through diet to higher trophic levels and ultimately recorded in the stomach oil of snow petrels (Hiller et al., 1988; Verkulich and Hiller, 1994). Divergence in signals during cold periods may indicate a shift in foraging by the petrels from ^{13}C-enriched neritic prey to normally ^{13}C-depleted pelagic prey, except for those pelagic prey encountered at the productive pack-ice edge during cooler periods, a shift forced by presumed greater sea ice concentration during those times (Ainley et al., 2006). For ^{13}C, both mumiyo and marine sediment were enriched during the warmer ocean conditions experienced during the mid-Holocene (ca. 7.5 to 5.5 cal ka BP) in the Bunger Hills (Ainley et al., 2006). In general most long-term data for high-latitude Antarctic seabirds (Adélie and emperor penguins and snow petrels) indicate that winter sea ice has a profound influence. However, some effects are inconsistent between species and areas, and other effects trend in opposite directions at different stages of breeding and life cycles.

3.6 Concluding remarks

This chapter has summarised our current understanding of Antarctic climate and environment history prior to the instrumental record, and the biological effects of climate and environmental change. Although the broad features of the climate and environmental history of Antarctica are being progressively documented, improvements can still be made in understanding the detailed sequence of events with better resolution both temporally (for instance, in the phasing of changes in Antarctic climate, environmental parameters and atmospheric composition) and spatially, particularly in near coastal regions. This calls for enhanced geographical coverage of high resolution records and a better understanding of climate dynamics.

Research priorities for the ice core community in Antarctica include collecting a 1.5 million year record of climate and greenhouse gases from Antarctica (spanning the time period where Earth's climate shifted from 40,000 year to 100,000 year cycles); a 40,000 year bipolar network of records of climate forcing (to improve understanding of the detailed sequence of events across Termination 1, particularly at coastal sites representative of different oceanic basins), and a network of ice core climate and climate forcing records for the last two millennia (IPICS, 2008). These will help to progress understanding of the climate and biogeochemical cycle responses to climate forcings, the climate variability around Antarctica and its source areas. Similarly, and despite the strong correlation between CO_2 variations and Antarctic temperature, identifying the mechanisms involved in glacial-interglacial changes in greenhouse gas concentrations and their lags with Antarctic temperature is still a challenge for modellers (Köhler et al., 2005). In addition, for a better understanding of climate dynamics, new ice cores are also needed to improve the knowledge of Antarctic ice sheet dynamics over the past deglaciation. In particular, results from new ice cores from Detriot Plateau and James Ross Island in the Antarctic Peninsula region are eagerly anticipated as these are located in the fastest warming region of Antarctica.

For marine and terrestrial records the priority is to improve the geographical coverage and spatial resolution of the data so that the underlying links with climate forcing mechanisms emerge. A particular challenge is to constrain the timing of major climate events so that leads and lags may be identified. For example, records of climate optima in the marine and terrestrial environment in the mid to late Holocene are often out of phase and, with the exception of the Antarctic Peninsula (Bentley et al., 2009), little attention has been paid to identifying the mechanisms driving the different palaeoclimate patterns in various archives at a regional scale. Another example is the occurrence of a later Holocene climate optimum in the Ross Sea which is in phase with a marked cooling observed in ice cores from coastal and inland locations (Masson et al., 2000; Masson-Delmotte et al., 2004). These differences in the timing of warm events in different records, and in different regions, point to a number of mechanisms that we have yet to identify. Priority areas include coastal Antarctica, the Dronning Maud Land region, West Antarctica and specific regions of the Antarctic Peninsula. Here, modellers can also be of significant help in carrying out data model comparisons and providing sound theoretical frameworks within which to identify and answer the major questions

In terms of glacial history further work on the response of the ice sheet to climate change is required, particularly geological records of past ice sheet extent, including submarine surveys, and records of past relative sea level change. New biological evidence of floras that have survived through glacial cycles also challenges existing ice sheet models where most of the available habitats are covered at the LGM.

For past sea ice studies more marine sediment core data are needed to reconstruct past sea ice extent particularly in the Pacific sector of the Southern Ocean. Further detailed comparisons are required of marine sediment core-based and ice core-based sea ice reconstructions on a region by region basis. Sea ice is also a factor that remains poorly constrained in computer simulations of past and future climate change primarily because of this paucity of historical and palaeo records of sea ice extent.

The study of the response of biological communities in Antarctica and the Southern Ocean to historical and contemporary climate change, beyond the instrumental record, is still in its infancy, and nearly absent at smaller (microorganism to molecular) scales, yet this potentially offers many analogues with which to better understand the climate changes predicted to occur in the next 100 years. For example, on the Antarctic Peninsula, which is the fastest warming region of Antarctica, there is a pressing need for high resolution stratigraphic records that include evidence of changing species distributions. This region is likely to have been amongst the most responsive to climate changes of the past and already increasing temperatures in the last 50 years have resulted in the local expansion of population ranges of a number of plant and animal species, and the establishment (albeit with human assistance) of new species that appear not to have survived on the continent before. Similarly, climate impacts on biodiversity and production on land and in the ocean are still poorly constrained despite their wider economic consequences.

Chapter 4

The Instrumental Period

Chapter Editor:

John Turner

Authors:

Byron Adams, Rob Arthern, Angus Atkinson, Carlo Barbante, Roberto Bargagli, Dana Bergstrom, Nancy Bertler, Robert Bindschadler, James Bockheim, Claude Boutron, David Bromwich, Steve Chown, Josifino Comiso, Pete Convey, Alison Cook, Guido di Prisco, Eberhard Fahrbach, Jim Fastook, Jaume Forcarda, Josep-Maria Gili, Mauro Gugliemin, Julian Gutt, Hartmut Hellmer, Françoise Hennion, Karen Heywood, Dominic Hodgson, David Holland, Sungmin Hong, Ad Huiskes, Enrique Isla, Stan Jacobs, Anna Jones, Andrew Lenton, Gareth Marshall, Paul Mayewski, Mike Meredith, Nicolas Metzl, Andrew Monaghan, Alberto Naveira-Garabato, Kevin Newsham, Covadonga Orejas, Lloyd Peck, Hans-Otto Pörtner, Steve Rintoul, Sharon Robinson, Howard Roscoe, Sergio Rossi, Ted Scambos, Jon Shanklin, Victor Smetacek, Kevin Speer, Mark Stevens, Colin Summerhayes, Phil Trathan, John Turner, Kees van der Veen, David Vaughan, Cinzia Verde, David Webb, Christian Wiencke, Philip Woodworth, Tony Worby, Roger Worland, Takashi Yamanouchi,

4.1 Introduction

The instrumental period began with the first voyages to the Southern Ocean during the Seventeenth and Eighteenth centuries when scientists such as Edmund Halley made observations of quantities such as geomagnetism. During the early voyages information was collected on the meteorological conditions across the Southern Ocean, ocean conditions, the sea ice extent and the terrestrial and marine biology.

The continent itself was discovered in 1820, although the collection of data was sporadic through the remainder of the Nineteenth Century and it was not possible to venture into the inhospitable interior of Antarctica.

At the start of the Twentieth Century stations were first operated year-round and this really began the period of organised scientific investigation in the Antarctic. Most of these stations were not operated for long periods, which is a handicap when trying to investigate climate change over the last century.

The International Geophysical Year (IGY) in 1957/58 saw the establishment of many research stations across the continent and this period marks the beginning of many of the environmental monitoring programmes. Thankfully many of the stations are still in operation today so that we now have some 50 year records of many meteorological parameters.

The ocean areas around the Antarctic have been investigated far less than the continent itself. Here we are reliant on ship observations that have mostly been made during the summer months. Satellite observations can help in monitoring the surface of the ocean, but not the layers below. Even here quantities such as sea ice extent have only been monitored since the late 1970s, when microwave technology could be flown on satellites.

4.2 Changes of Atmospheric Circulation

4.2.1 Modes of variability

4.2.1.1 Variability

Within the atmospheric circulation of the high southern latitudes several so-called modes of variability can be discriminated. These circulation patterns appear more frequently than would be expected in a random sample, and can 'describe' a significant proportion of the total circulation variability.

4.2.1.2 The Southern Annular Mode

As pointed out in Chapter 1, the Southern Annular Mode (SAM) is the principal mode of the Southern Hemisphere extra-tropical atmospheric circulation. It can be described either as a 'flip-flop' of atmospheric mass between mid- and high-latitudes, such that there are synchronous pressure (or geopotential height) anomalies of opposite sign in these two regions (e.g. Rogers and van Loon, 1982), or as a north-south shift in the mid-latitude jet resulting from both latitudinal vacillations in the jet and fluctuations in jet strength (Fyfe and Lorenz, 2005). When pressures are higher (lower) than average over the Southern Hemisphere mid-latitudes (Antarctica) the SAM is said to be in its positive phase (see Figure 4.1) and *vice versa*.

The spatial pattern of the SAM varies negligibly with height in the atmosphere and is revealed as the leading mode of variability in many atmospheric fields (see Thompson and Wallace (2000) and references therein). Model experiments demonstrate that the structure and variability of the SAM results from the internal dynamics of the atmosphere (e.g. Limpasuvan and Hartmann, 2000). Synoptic-scale (a horizontal scale of greater than 1,000 km) weather systems interact with the zonal mean flow to sustain latitudinal displacements of the mid-latitude westerlies. The SAM contributes a significant proportion of Southern Hemisphere climate variability (typically ~35%) from high-frequency to very low-frequency timescales, with this variability displaying a greater tendency to low frequency variability.

Gridded reanalysis datasets have been used to derive time series of the SAM (e.g. Renwick, 2004). Due to the poor quality of current reanalyses products at high southern latitudes prior to the assimilation of satellite sounder data in the late 1970s (Hines et al., 2000), long-term SAM time-series cannot be derived from them. Based on a definition by Gong and Wang (1999), Marshall (2003) produced a SAM index based on 12 appropriately located station observations in the extra-tropics and coastal Antarctica. This index itself is limited to the period following the IGY. Attempts have been made to reconstruct annual to century-scale records based on the SAM and associated atmospheric circulation features, such as surface pressure over Antarctica and the Southern Hemisphere westerlies. Souney et al. (2002a) used variability in Na concentrations from the Law Dome ice core, while Jones and Widmann (2003) employed tree-ring width chronologies: both these studies highlight the decadal variability in their derived SAM time series. Yan et al. (2005) reconstructed past behaviour of the Southern Hemisphere westerlies at the edge of the polar vortex using the variability in Ca concentrations from West Antarctic ice cores.

The SAM has shown significant positive trends over the past few decades, particularly during austral autumn and summer (e.g. Thompson et al., 2000; Marshall, 2003). The positive trend was especially pronounced from the mid-1960s until the end of the Twentieth Century, since when there has been a similar frequency of seasons with positive and negative SAM values (Figure 4.2). The positive trend in the SAM has resulted in a strengthening of the mean circumpolar westerlies of ~15% (Marshall, 2002), and contributed to the spatial variability in Antarctic temperature change (e.g. Thompson and Solomon, 2002; Kwok and Comiso, 2002; Schneider et al., 2004; Marshall, 2007), specifically a warming in the northern Peninsula region and a cooling over much of the rest of the continent. The SAM also impacts the spatial patterns of variability in precipitation across Antarctica (e.g. Genthon et al., 2003).

The imprint of SAM variability on the Southern Ocean system is observed as a coherent sea level response around Antarctica (Aoki, 2002; Hughes et al., 2003), and by its regulation of Antarctic Circumpolar Current (ACC) flow through the Drake Passage (Meredith et al., 2004). Modelling studies show that a positive phase of the SAM is associated with northward (southward) Ekman drift in the Southern Ocean (at 30°S) leading to upwelling (downwelling) near the Antarctic continent (~45°S) (Hall and Visbeck, 2002; Lefebvre et al., 2004). These changes in oceanic circulation impact directly on the thermohaline circulation and may explain recent patterns of observed oceanic temperature change in the Southern Ocean described by Gille (2002). Although the SAM is essentially zonal, a wave-number 3 pattern (three troughs and three ridges around the hemisphere) is superimposed, with a marked low pressure anomaly west of the Peninsula when the SAM is positive, leading to increased northerly flow and reduced sea ice in the region (Liu et al., 2004). Raphael (2003) reported that diminished summer sea ice may in turn feed back into a more positive SAM. However, modelling work by Marshall and Connolley (2006) showed that increased sea surface temperatures (SSTs) at high southern latitudes will warm the atmosphere, and, through thermodynamical processes, cause the atmospheric centre-of-mass to rise and geopotential height to increase thus producing a more negative SAM.

4.2.1.3 The Pacific-South American (PSA) pattern

The El Niño–Southern Oscillation (ENSO) phenomenon is the largest climatic cycle on Earth on decadal and sub-decadal time scales, and can influence the weather and climate well beyond the low-latitude Pacific Ocean where it is most marked. ENSO is a coupled atmosphere-ocean phenomenon that involves a major reversal of the atmospheric and oceanic flows across the tropical Pacific Ocean. During the La Niña phase there is intense storm activity close to Indonesia and strong westward moving atmospheric and ocean flow across the Pacific near the Equator. During the El Niño phase the storm activity moves close to the date-line, with deep convection giving upper atmosphere divergence. This results in a quasi-stationary Rossby Wave (atmospheric long wave) pattern becoming established in both hemispheres, so providing teleconnection patterns between high and low latitudes.

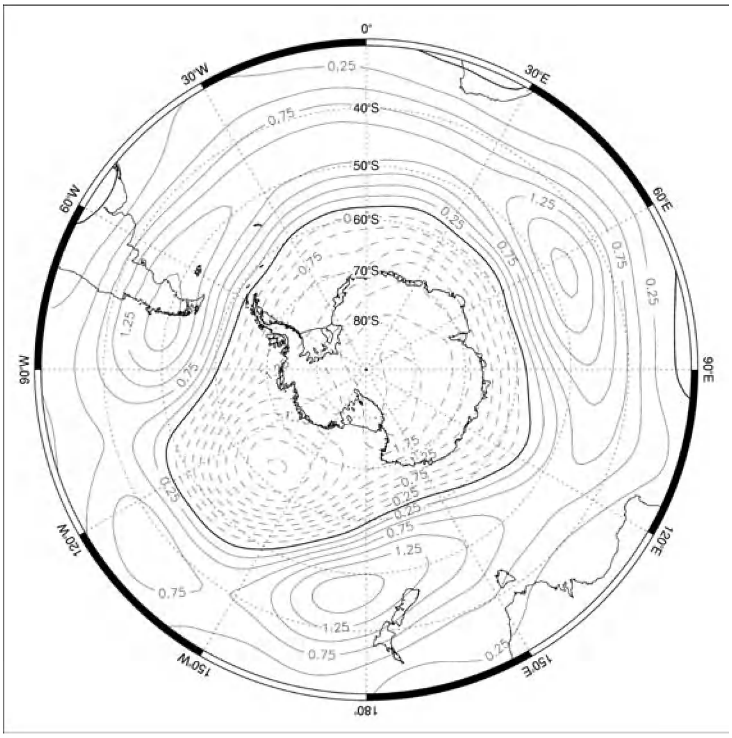


Figure 4.1 The austral summer SAM derived from gridded 500 hPa geopotential height monthly anomaly data (the 500 hPa surface is approximately at an elevation of 5 km above mean sea level) for 1989-2008. Here the SAM is in its positive phase with negative anomalies over the Antarctic and positive anomalies over the Southern Ocean.

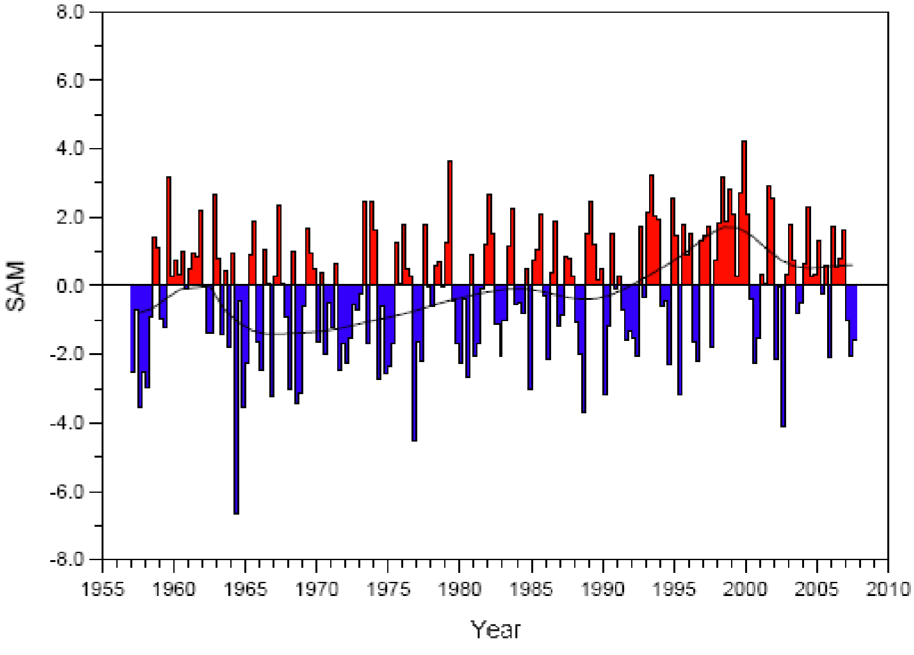


Figure 4.2 Seasonal values of the SAM index calculated from station data (Marshall, 2003). The smooth black curve shows decadal variations. This is an updated version of Figure 3.32 from IPCC (2007).

In the Southern Hemisphere, the Rossby wave train is known as the Pacific-South American Association (PSA) and it links the central tropical Pacific and the Amundsen-Bellingshausen Sea (ABS) (Figure 4.3). The most robust signals of ENSO are found across the South Pacific during winter and at this time of year, the wave train can be identified as anomalously high mean sea level pressure (MSLP) and upper level heights (heights of constant pressure surfaces) across the ABS, with anomalously low values to the east of New Zealand (Figure 4.4) (Karoly, 1989). Such circulation anomalies give generally colder temperatures across the Antarctic Peninsula, with more extensive sea ice, and warmer air arriving from the north across West Antarctica and the Ross Ice Shelf (Figure 4.4).

During the La Niña phase of the cycle these anomaly patterns are broadly reversed with more cyclonic activity over the ABS and warmer winters experienced over the Antarctic Peninsula (Turner, 2004) (Figure 4.4).

The PSA pattern is thought to be primarily related to ENSO. Nonetheless, it is prominent at many timescales even in the absence of a strong ENSO signal. Thus, the PSA may also be an internal mode of climate variability. Note that Mo and Higgins (1998) state that the PSA actually comprises two separate modes, each associated with enhanced convection in different parts of the Pacific and suppressed convection elsewhere.

An alternative mechanism for the PSA teleconnection was proposed by Liu et al. (2002). They suggest that the increased convection associated with El Niño events alters the mean meridional atmospheric circulation through longitudinal changes to the Hadley circulation and subsequent alterations of the subtropical jet position and strength. Yuan (2004) suggests that the two mechanisms operate in phase and are comparable in magnitude.

While there are general patterns of high-latitude climate anomalies that co-vary with ENSO, correlations suggest that teleconnections are small. A modelling study by Lachlan-Cope and Connolley (2006) shows that this is because (i) Rossby wave dynamics are not well-correlated to 'standard' definitions of ENSO, because the relationship between upper level divergence and SSTs via deep convection is complex, and (ii) natural variation in the zonal flow of Southern Hemisphere high-latitudes can swamp the ENSO signal. Moreover, Fogt and Bromwich (2006) demonstrated that a weaker high-latitude teleconnection in spring during the 1980s compared to the following decade was due to the out-of-phase relationship between the PSA and SAM at this time: subsequently, the tropical and extra-tropical modes of climate variability had an in-phase relationship.

Ice core records have provided insight into the tropical-Antarctic teleconnections in the pre-instrumental period. Schneider and Steig (2008) found positive temperature anomalies in West Antarctica over the period 1936-45, which they attributed to the major 1939-42 El Niño event.

Although apparent throughout the year, the PSA demonstrates the strongest teleconnections to the Antarctic region in austral spring and summer. During these seasons an El Niño (or a La Niña) event is associated with significantly more (or less) blocking events in the southeast Pacific (Renwick and Revell, 1999). This pressure anomaly in the ABS is primarily responsible for the Antarctic Dipole (ADP) in the interannual variance structure in the sea ice edge and SST fields of the Southern Ocean, which are characterised by an out-of-phase relationship between anomalies in the central/eastern Pacific (ABS) and the Atlantic (Weddell Sea) sectors (Yuan and Martinson, 2001: see Figure 4.4). In addition, the ADP sea ice anomalies are reinforced by ENSO-related storm track variability, which influences the Ferrel cell (the vertical circulation cell located between the Hadley Cell of the tropics and the Polar Cell of the high latitude areas) by changing the meridional heat flux divergence/convergence and shifting the latent heat release zone, which in turn modulates the mean meridional heat flux that impacts sea ice extent. Correlations with ENSO indices imply that up to 34% of the variance in sea ice edge is linearly related to ENSO (Yuan and

Martinson, 2000), while Gloersen (1995) showed that different regions of sea ice respond to ENSO at different periodicities. The strongest sea ice correlations associated with ENSO occur at 120-132° W (in the ABS) lagging the tropical temperature anomaly by 6 months; so the austral spring/summer ENSO signal is observed in the subsequent ice growth period in autumn/winter (Yuan and Martinson, 2000). Note that the temporal quasi-periodic nature of both ENSO and the ice anomalies prevents the identification of direction of causality. Kwok and Comiso (2002) state that recent trends in sea ice extent in both the Bellingshausen Sea and Ross Sea are related directly to ENSO variability.

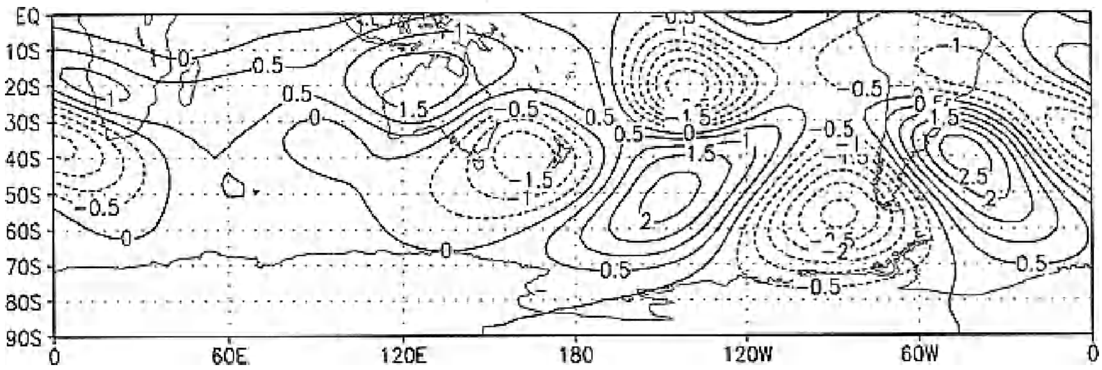


Figure 4.3 The PSA teleconnection pattern of positive and negative pressure anomalies (hPa) (after Mo and Higgins, 1998). Notice the chain of anomalies from Australia and the central Pacific and the 'centre of action' west of the Antarctic Peninsula.

4.2.1.4 The Antarctic Circumpolar Wave

The Antarctic Circumpolar Wave (ACW), first defined by White and Peterson (1996), is an apparent easterly progression of phase-locked anomalies in Southern Ocean surface pressure, winds, SSTs and sea-ice extent (Figure 4.5). It thus represents a coupled mode of the ocean-atmosphere system. The ACW has a zonal wavenumber of 2 (meaning a wavelength of 180°), and the anomalies propagate at a speed of 6-8 cm/sec such that they take 8-10 years to circle Antarctica, giving the ACW a period of 4-5 years. A similar feature has also been identified in sea surface height using satellite altimeter data (Jacobs and Mitchell, 1996). Given the much shorter response times of the atmosphere, these authors proposed that the ocean plays an important part in creating and maintaining the ACW. As there is little Antarctic multiyear ice, Gloersen and White (2001) suggested that the memory of the ACW in the sea ice pack is carried from one austral winter to the next by the neighbouring SSTs. White et al. (2004) showed a complex tropospheric response to sea ice anomalies that, for example, explained anomalous poleward surface winds and deep convection observed with negative sea ice edge anomalies.

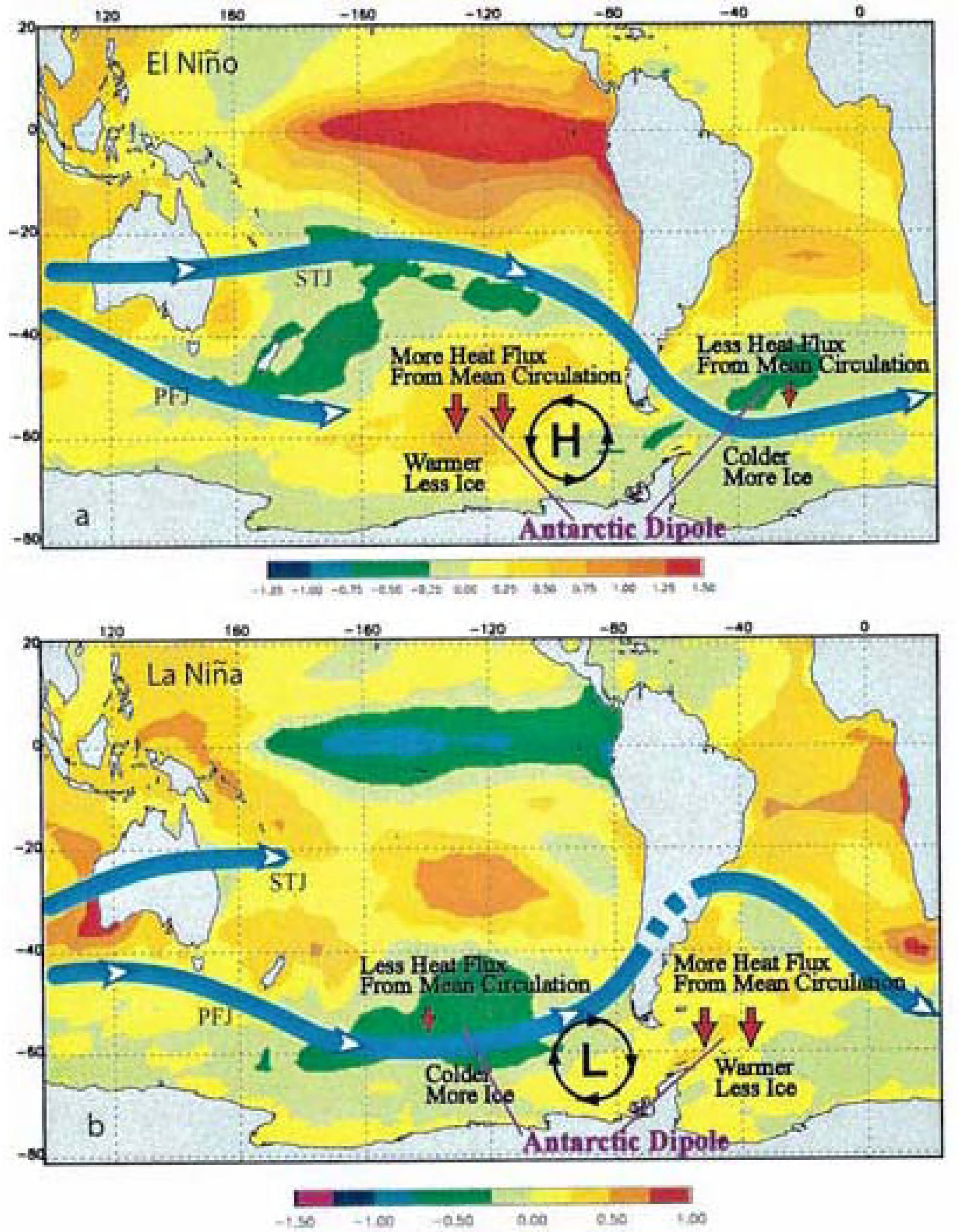


Figure 4.4 SST anomaly composites (°C) for (a) El Niño conditions and (b) La Niña conditions. Schematic jet streams (STJ is the subtropical jet and PFJ is the polar front jet), persistent anomalous high and low pressure centres, and heat fluxes are also marked. (Yuan, 2004).

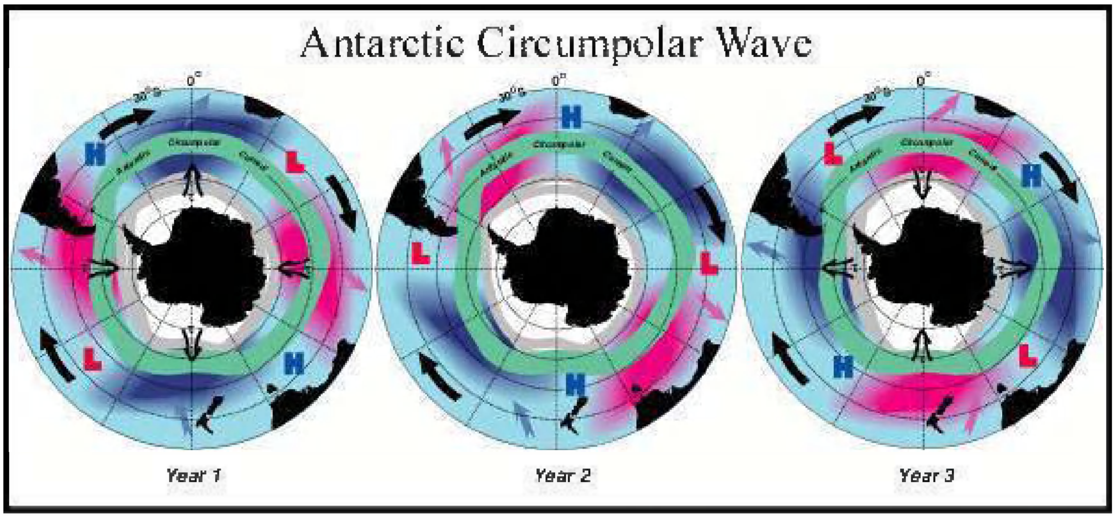


Figure 4.5 Simplified schematic summary of interannual variations in sea surface temperature (red, warm; blue, cold), atmospheric sea level pressure (bold H (high) and L (low)), and sea ice extent (grey line), together with the mean course of the Antarctic Circumpolar Current (green). Heavy black arrows depict the general eastward motion of the anomalies, and other arrows indicate communications between the circumpolar current and more northerly tropical gyres (White and Peterson, 1996).

Some authors have suggested that an ACW signal can be observed in an Antarctic ice core over the last 2000 years (Fischer et al., 2004). But, since the initial discovery of the ACW, others have questioned its persistence. Several observational and modelling studies have indicated that the ACW is not apparent in recent data before 1985 and after 1994 (e.g. Connolley, 2003), which is somewhat fortuitously the period that White and Peterson (1996) chose for their original analysis. In addition, there has been some discussion on one of its key characteristics, whether it really has a wave number 2 pattern. Many studies (e.g. Cai et al. 1999) have indicated that an ACW-like feature is apparent in GCM control runs but that it has a preferred wavenumber 3 pattern. Venegas (2003), using frequency domain decomposition, suggested that the ACW comprises two significant interannual signals that combine constructively/destructively to give the observed irregular fluctuations of ACW on interannual time scales, as also seen in GCM studies. The two signals comprise (i) a 3.3 year period of zonal wavenumber 3 and (ii) a 5 year period of zonal wavenumber 2, which was particularly pronounced during the period studied by White and Peterson.

However, most of the ACW debate has centred on its forcing mechanisms and, as a consequence, the very nature of its existence. For example, White et al. (2002) state explicitly that the ACW exists independently of the tropical standing mode of ENSO, and that its eastward propagation depends upon atmosphere-ocean coupling rather than on advection by the ACC: both these points have been refuted in the literature. The ADP (Yuan and Martinson, 2001), described previously, has the same wavelength as the ACW, but key differences are that the associated variability is twice that of the ACW, and that the dipole consists principally of a strong standing mode together with a much weaker propagating motion. Moreover, Yuan and Martinson (2001) state that the ADP is clearly associated with ENSO events. Several other authors have found that ENSO variability is important in driving a geographically fixed standing wave in Southern Ocean interannual variability, that is centred in the South Pacific and linked to the PSA pattern (Cai et al., 1999; Venegas, 2003; Park Y. et al., 2004).

Furthermore, several papers mention the ACC as the means of anomaly propagation. In an ocean model forced by realistic stochastic (random) anomalies, Weisse et al. (1999) showed that the ocean acts as an integrator of short-term atmospheric fluctuations (white noise) and turns them into a lower frequency signal (red noise); subsequently the average zonal velocity of the ACC determines the timescale of the oceanic variability and thus the propagation speed and period of the ACW (Cai et al., 1999; Venegas, 2003). The results of the study of Park Y. et al. (2004) indicated that any such propagating anomalies comprised only ~25% of interannual Southern Ocean SST variability and are often rapidly dissipated in the Indian Ocean, are intermittent in phase and frequently do not complete a circumpolar journey (Figure 4.6). Hence, they questioned the existence of the ACW, as originally described by White and Peterson (1996).

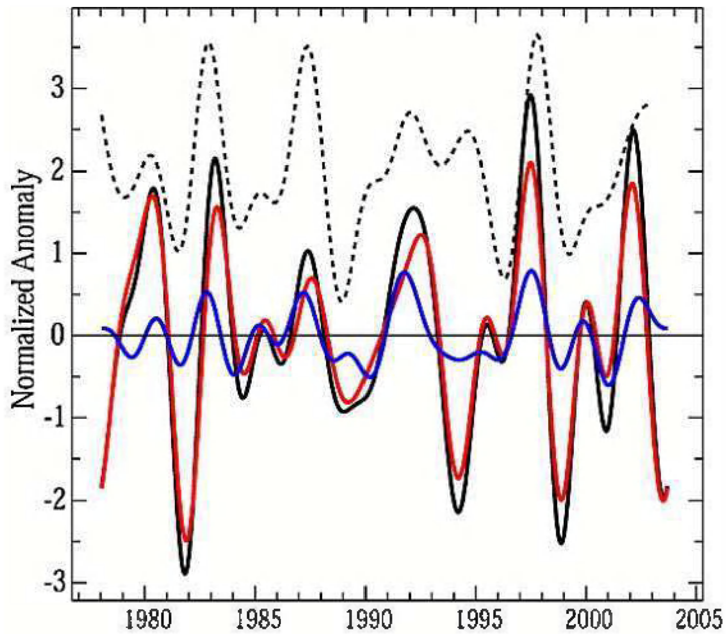


Figure 4.6 Temporal variations at 140°W of the total (black line), stationary (red line), and eastward (blue line) SST. Values are normalized by a standard deviation of the total SST. The Southern Oscillation Index (a measure of the phase of the El Niño – Southern Oscillation) is shown by a micro-dashed line, but with its sign being reversed and its zero axis being displaced by +2 to better compare with peaks of SST (Park Y. et al., 2004).

4.2.2 Depression Tracks

The paths of cyclones, known as depression or storm tracks, can be obtained using three broad methodologies. First, individual weather systems can be identified by their cloud signatures in satellite imagery or meteorological charts, and their subsequent paths can then be determined from further multitemporal imagery. The polar regions are well-suited to this kind of analysis because polar orbiting satellites provide many overlapping passes at high latitudes. Although attempts have been made to automate this process, most studies to date have been done manually, and the labour-intensive nature of this process means that the time periods covered are relatively short. For example, Turner et al. (1998) examined 12 months of AVHRR data in the Antarctic Peninsula sector during which 504 synoptic-scale lows were identified.

The availability of gridded data from numerical weather prediction (NWP) models and reanalyses allows entirely automated methods to be used that can provide climatological information on depression tracks, including trends. System-centred tracking is achieved by searching for local minima in certain fields, such as MSLP (e.g. Jones and Simmonds, 1993), or maxima in vorticity (e.g. Hoskins and Hodges, 2005). An example of the resultant tracks at high southern latitudes is shown in Figure 4.7a. A final methodology is to utilise Eulerian storm-track diagnostics by identifying the variance of vorticity at synoptic-timescales (2-6 days). Figure 4.7b is an equivalent figure to 4.7a using that method: note the significant difference in results in the circumpolar trough (CPT) with the system-centred approach finding a maximum, while the Eulerian method indicates relatively little cyclone activity at that site. The satellite-imagery study of Turner et al. (1998) reveals that there is indeed a maximum in cyclone activity at that site.

There is a marked difference between the main CPT storm track in austral summer and winter. In the summer it is nearly circular and confined to high-latitudes south of 50°S. In winter (Fig 4.7a) the storm track is more asymmetric with a spiral from the Atlantic and Indian Oceans in towards Antarctica, and a pronounced sub-tropical jet (STJ) - related storm track at ~35°S over the Pacific (e.g. Hoskins and Hodges, 2005). Simmonds and Keay (2000) found that the mean track length of winter systems (2,315 km) was slightly longer than in summer (1,946 km). The equinoctial seasons have depression track patterns intermediate between summer and winter. The CPT has a maximum in storm activity in the Atlantic and Indian Ocean regions at all times of year. While the previous description relates to the mean climatology, individual weather systems can behave differently; for example, many studies have shown cases where cyclones have a strong northerly track, particularly when associated with cold-air outbreaks from the Antarctic continent. The automated procedures can also be used for locating and tracking anticyclones. In such a study, Sinclair (1996) found that south of 50°S anticyclones were generally rare, but their tracks were associated with the main regions of blocking in the South Pacific as described previously.

Using gridded data, Simmonds and Keay (2000) showed that annual and seasonal numbers of cyclones have decreased at most locations south of 40°S during the 1958-97 period examined, and can be related to changes in the SAM. The latter is associated with a decline in pressure around Antarctica so there has been a trend to fewer but more intense cyclones in the circumpolar trough. One exception is the Amundsen-Bellingshausen Sea region where cyclonic activity has increased (Simmonds et al., 2003)

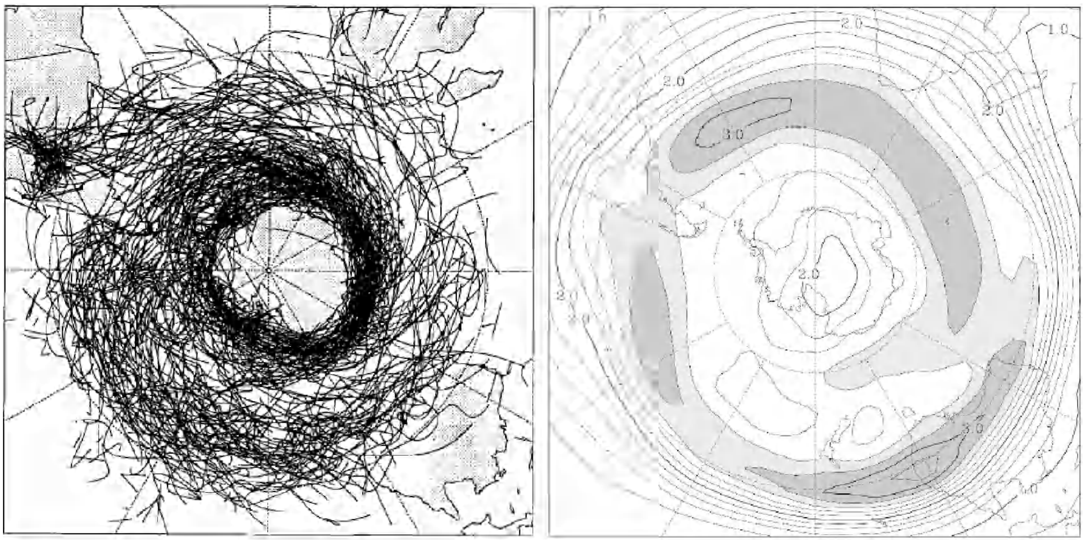


Figure 4.7 (a) Depression tracks for winter 1985-89. From Jones and Simmonds (1993). (b) Bandpass-filtered (2-6 day) variance converted to standard deviation for ξ_{250} for winter 1958-2002. This quantity provides a measure of storm activity and indicates the tracks of major storms. From Hoskins and Hodges (2005). The shaded areas show the large atmospheric variability associated with the frequent passage of depressions.

4.2.2.1 Depression formation and decline

Depression tracking studies can also show the locations where a weather system is first and last identified, respectively cyclogenesis and cyclolysis. By collating data from all cyclones examined, regions of preferred cyclogenesis and cyclolysis may be determined. Using data from the NCEP-NCAR reanalysis, Simmonds and Keay (2000) showed there to be a net creation of cyclones (i.e. cyclogenesis is greater than cyclolysis) north of 50°S and net destruction south of this. Most Southern Hemisphere cyclogenesis actually occurs at very high latitudes (~60°S) in the CPT but rates of cyclolysis there are even higher. Turner et al. (1998) found that about half the systems in the Antarctic Peninsula region formed within the CPT and half spiralled in from further north. Simmonds et al. (2003) also revealed the northern part of the Peninsula to be a region of high cyclogenesis. Some of these systems formed through lee cyclogenesis — a dynamical process associated with the passage of air over a barrier — to the east of the Antarctic Peninsula. Other important regions of cyclogenesis within the CPT are in the Indian Ocean sector, with a maximum at 65°S, 150°E (Hoskins and Hodges, 2005) at the edge of the sea ice. Cyclolysis is generally confined to the CPT, with maxima in the Indian Ocean and also the Bellingshausen Sea, where the steep, high Antarctic Peninsula often prevents weather systems passing further to the east: such areas are known as 'cyclone graveyards'.

4.2.2.2 Blocking events

Normally, low pressure systems do not intrude into the high inland area. Sometimes, in winter, blocking events occur and depressions from lower latitudes are channelled south over the ice sheet (Enomoto et al., 1998; Hirasawa et al., 2000; Pook and Gibson, 1999; Schneider et al., 2004). During these blocking events, an abrupt increase of surface temperature up to 40°C may occur within a few days, accompanied by a heavy snowfall, that significantly influences accumulation on the inland ice sheet. These events are not common, but may have a pronounced effect on the energy balance of inland areas and on the annual layer formation seen in ice cores.

4.2.3 Teleconnections

4.2.3.1 Atmospheric linkages

Section 4.2.1.3 described the teleconnections between the ENSO cycle of the tropical Pacific and the Antarctic. However, a problem is that while many of the El Niño and La Niña events give the atmospheric anomaly patterns described earlier, some of these events have resulted in very different conditions. For example, the 1982/83 El Niño event was one of the largest of the last century, but the region of positive MSLP/height anomaly was displaced towards the tip of South America so the Antarctic Peninsula had only average conditions rather than the low temperatures usually associated with El Niño events. In fact, the PSA is generally more variable than the Pacific North American Association, making it much more difficult to anticipate how the Antarctic atmosphere will respond as an El Niño event starts to become established.

In recent decades there has been a trend towards more frequent and more intense El Niño events. We could therefore expect MSLP values to have risen across the ABS and there to have been colder winter season temperatures across the Antarctic Peninsula and warmer conditions in the Ross Sea area. However, this has not happened, indicating that the winter warming on the western side of the Peninsula is not a direct result of changes in ENSO.

Bertler et al. (2004) noted a shift of the Amundsen Sea Low eastwards during some El Niño events and such a result would be consistent with the recent observed cooling in the Ross Sea area (Doran et al., 2002). However, this area has experienced jumps in the relationship between ENSO and West Antarctic precipitation (Cullather et al., 1996) indicating the variable nature of the teleconnections. In addition, the Antarctic Peninsula, far from cooling, has experienced the largest increase of temperature anywhere in the Southern Hemisphere (Turner et al., 2005a).

In summary, while the relatively short timeseries that we have of Antarctic meteorological observations and atmospheric analyses do suggest that tropical atmospheric and oceanic conditions affect the climate of the Antarctic and the Southern Ocean the connections do vary with time and are rather non-linear. Ice core proxies for El Niño are emerging to help fll this gap, notably the MSA proxy for the last 500 years of El Niño produced from a South Pole ice core (Meyerson et al., 2002). However, the teleconnections are not as robust as those in the Northern Hemisphere. In addition, many other factors in the Antarctic climate system, such as the variability in the ocean circulation, the development of the ozone hole and the large natural variability of the high latitude climate, all affect atmospheric conditions and can mask the tropical signals.

4.2.3.2 Oceanic Coupling

The atmospheric processes described above have strong impacts on the high latitude Southern Ocean, which can (via complex feedback mechanisms) produce changes to the large-scale coupled atmosphere/ocean/ice system. The atmospheric Rossby wave associated with ENSO (the PSA association) produces anomalies in SSTs across the Southern Ocean, via processes including changes to the surface heat budget associated with changes in clouds and radiation (Li, 2000). These changes are especially manifest in the South Pacific (Figure 4.4).

The SST anomalies so created lie largely within the domain of the ACC, and hence are subsequently advected eastward. A typical timescale for advection across the South Pacific and into the South Atlantic is 2 years. This eastward procession was previously associated with the concept of an Antarctic Circumpolar Wave (White and Peterson, 1996), whereby pairs of coupled anomalies in SST, sea ice concentration, winds etc propagate eastward around Antarctica. The potential ENSO trigger for such a wave was recognised early (e.g. Peterson and White, 1998), however it is now known that this phenomenon exists only during certain time periods, and is distorted or absent during others (Connolley, 2003).

Despite this, it is important to recognise that the SST anomalies that are advected around the Southern Ocean are subject to further air-sea interaction as they propagate, and will be magnified or diminished as a result. The southeast Pacific is one region very susceptible to ENSO forcing (see above), and Meredith et al. (2007) showed that the phasing of advection of SST anomalies across the South Pacific generally coincides with that of ENSO, such that positive reinforcement of the SST anomalies occurs here. This acts to sustain the anomalies as they propagate eastward. As ENSO-induced anomalies propagate eastward within the ACC, phased changes on comparable timescales occur further south within the subpolar gyres (e.g. Venegas and Drinkwater, 2001).

ENSO is not the only tropical/equatorial phenomenon with a teleconnection to high southern latitudes. The Madden-Julian Oscillation (MJO, Madden and Julian, 1994) is the dominant mode of intraseasonal variability in the tropical atmosphere, and is associated with large-scale convective anomalies that propagate slowly eastward from the Indian Ocean to the western Pacific with a period of approximately 30-70 days. Matthews and Meredith (2004) showed that during the southern winter the MJO has an atmospheric extratropical response that impacts on the surface westerly winds around the 60°S latitude band, which are the winds that drive the ACC. They were able to show that changes in the ACC were induced in response to tropical variability. The total timescale for MJO to influence ACC transport was of the order of 1-2 weeks.

In addition to teleconnections from lower latitudes influencing the Southern Ocean and Antarctica, it is now known that they can operate in the reverse direction also. Recent studies (e.g. Ivchenko et al., 2004; Blaker et al., 2006) have demonstrated that signals generated in regions such as the Weddell Sea by anomalies in the cover of sea ice or the upper-layers of the ocean can propagate through the Drake Passage to the western Pacific as fast ocean barotropic Rossby waves. The time scale for this propagation is just a few days, and subsequently the signal propagates as an oceanic Kelvin wave along the western boundary and the equator, reaching the equatorial western coast of South America after 2-3 months. This impact on the equatorial regions raises the prospect of a potential high-latitude influence on ENSO, and hence complex coupled ocean-atmosphere feedbacks between the equatorial Pacific and high latitude Southern Ocean. Ongoing research is investigating this possibility.

Studies with a coupled global climate model (Richardson et al., 2004) indicate a inter-hemispheric coupling on the time scale of 10 years. A freshwater anomaly generated in the Southern Ocean inhibits the ventilation of deep waters around Antarctica which causes the deep ocean to warm and the surface to cool. Cooling induces an increase of sea ice thickness and extent which causes cooling of the atmosphere. The cooling signal propagates in the Pacific to the Northern Hemisphere in less then a decade where it affects the North Atlantic Oscillation.

4.3 Temperature changes

4.3.1 Surface temperature

Surface temperature trends across the Antarctic can be determined using a number of different forms of data, including the in-situ observations, satellite infra-red imagery and ice core isotope measurements. In order to get a reasonable estimate of trends it is necessary to use all these data.

The in-situ observational record of Antarctic surface temperatures is rather sparse and sporadic before the IGY (see Appendix A in King and Turner, (1997)), although the Orcadas series from Laurie Island, South Orkney Islands began in 1903 and the Faraday Station/Argentine Islands record began in 1947. However, we are fortunate in having around 16 stations on the Antarctic continent or islands that have reported on a near-continuous basis since the IGY. In addition, a further six stations started reporting during the 1960s, so that we have around two dozen time series that allow the investigation of temperature trends. Unfortunately, the vast majority of the stations are in the Antarctic coastal region or on the islands of the Southern Ocean, with only Vostok and Amundsen-Scott Station being in the interior of the continent.

The in-situ record has been used by several workers to investigate temperature changes across the continent and Southern Ocean (Jacka and Budd, 1991; Jacka and Budd, 1998; Jones, 1995; Raper et al., 1984). Many of the records were scattered across a number of data centres and it was unclear as to the amount of quality control that had been carried out on the observations. SCAR therefore initiated the READER (Reference Antarctic Data for Environmental Research) project to bring as many of the observations together as possible, quality control the data and produce a new data base of monthly mean temperatures (Turner et al., 2004). The READER data base is available online at http://www.antarctica.ac.uk/met/READER/.

The READER data base has now been used in a number of studies concerned with the climate of the Antarctic, including that of Turner et al. (2005a), which considered changes since the start of the routine instrumental record. Here we will use the READER data base and the online meteorological data maintained by Dr. Gareth Marshall (http://www.antarctica.ac.uk/met/gjma/) to examine how Antarctic temperatures have changed over the period of the instrumental record.

Surface temperature trends from the station data since the early 1950s illustrate a strong dipole of change, with significant warming across the Antarctic Peninsula, but with little change across the rest of the continent (Figure 4.8a). The largest warming trends in the annual mean data are found on the western and northern parts of the Antarctic Peninsula. Here Faraday/Vernadsky Station has experienced the largest statistically significant (<5% level) trend of +0.53 ^{o} C/dec for the period 1951-2006. Rothera station, some 300 km to the south of Faraday, has experienced a larger annual warming trend, but the shortness of the record and the large inter-annual variability of the temperatures means that the trend is not statistically significant. 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 away from Faraday/Vernadsky, with the long record from Orcadas on Laurie Island, South Orkney Islands only having experienced a warming of +0.20^{o}C/decade. This record covers a 100-year period rather than the 50 years for Faraday. For the period 1951-2000 the temperature trend was +0.13^{o}C/decade.

Determining temperature trends across the interior of the Antarctic is difficult as there are only two stations with long records. However, attempts have been made to extrapolate the station trends across the rest of the continent. Chapman and Walsh (2007) produced estimates of annual tends (Figure 4.8b) and found the greatest warming over the Antarctic Peninsula, but with a small warming (~0.1° C/dec) across West Antarctica and much of East Antarctica. However, they also found cooling in a swath from the South Pole to Halley Station.

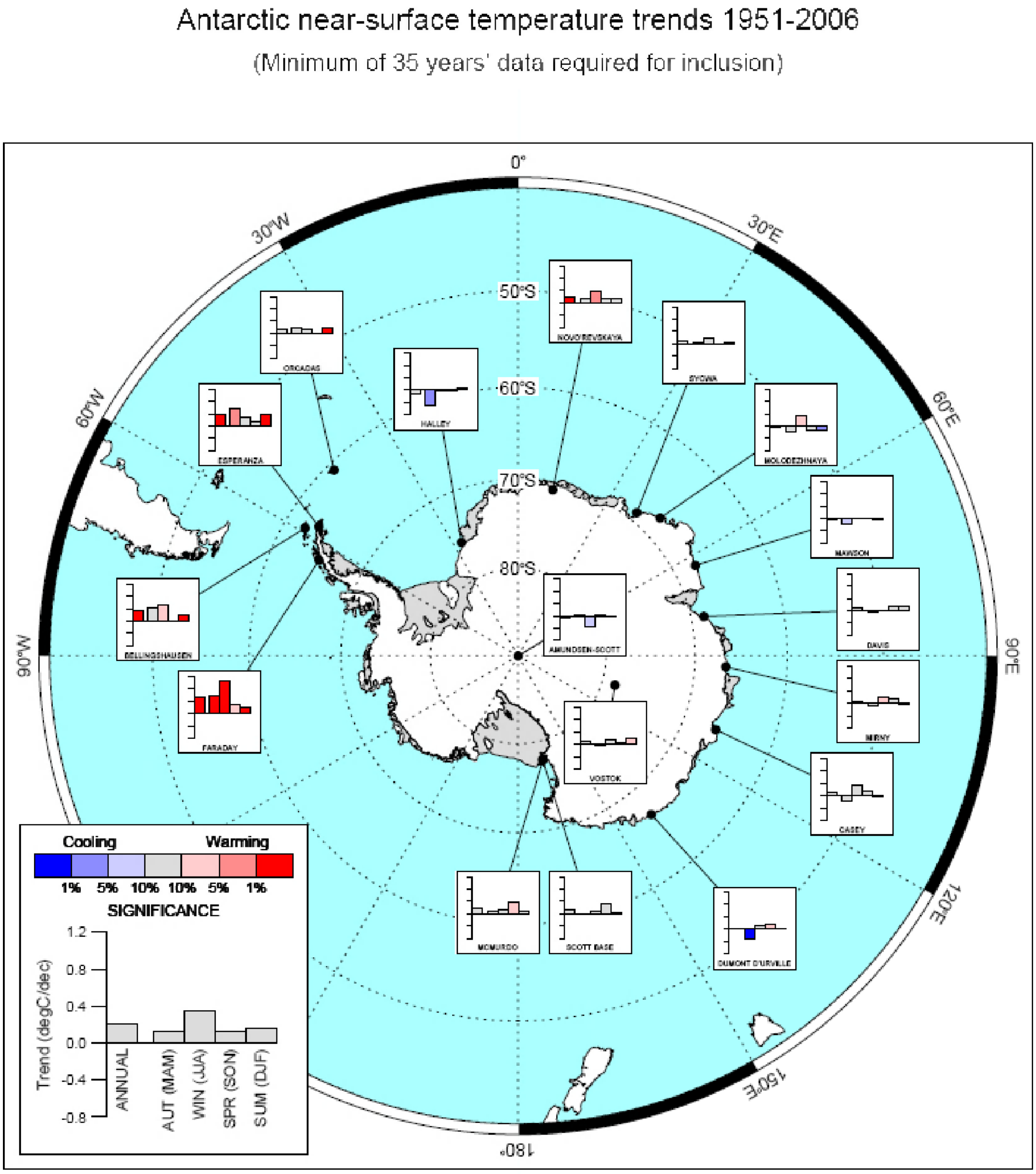
Steig et al. (2009) use statistical climate-field-reconstruction techniques to produce similar fields of trends for the seasons and the year as a whole. The annual trends (Figure 4.8c) show significant warming over most of West Antarctica with trends greater than 0.1° C/dec over the last 50 years. The trends are greatest during the winter and spring.

There has been a great deal of debate about the causes of the recent temperature changes across the continent. The summer warming on the eastern side of the Antarctic Peninsula has been shown to be a result of anthropogenic activity, and particularly the spring time loss of stratospheric ozone (Marshall et al., 2006). For the continent as a whole Gillett et al. (2009) carried out a formal attribution study to determine whether the observed changes were within the range of natural climate variability or whether they were a result of anthropogenic forcing. They found that recent changes were not consistent with internal climate variability or natural climate drivers alone, and were directly attributable to human influence.

Prior to the establishment of the research stations in the middle of the Twentieth Century we are reliant on ice core data to investigate surface temperature changes. Many studies have used single and multiple cores to investigate changes at selected sites or to investigate regional change. However, ‘stacking’ multiple cores can provide insight into Antarctic-wide change. Schneider et al. (2006) stacked several isotope records from ice cores to obtain a continental pattern of temperature over the past 200 years. The ice core stack was found to be well correlated with annual mean temperature and the data suggested warming of 0.2° C since the late nineteenth century. The paper suggested that recent Antarctic cooling is superimposed on longer term warming, with the more recent cooling being attributed to the SAM strengthening.

There is also evidence of climatic changes over the Southern Ocean. In recent decades, instrumental data recorded at the South African Weather Service station on Marion Island (46° 32' S and 37° 30' E) shows that the local climate of this island has undergone significant changes since the 1960s, mostly in the austral summer. These include a decrease in rainfall, an increase in non-rainy days, changes in wind speed and direction, and an increase in maximum and minimum local air temperature and in nearshore SST. Research suggests that the changes are linked to the well-documented shift of the semiannual oscillation and SAM after about 1980 (Rouault et al., 2005).

(a)



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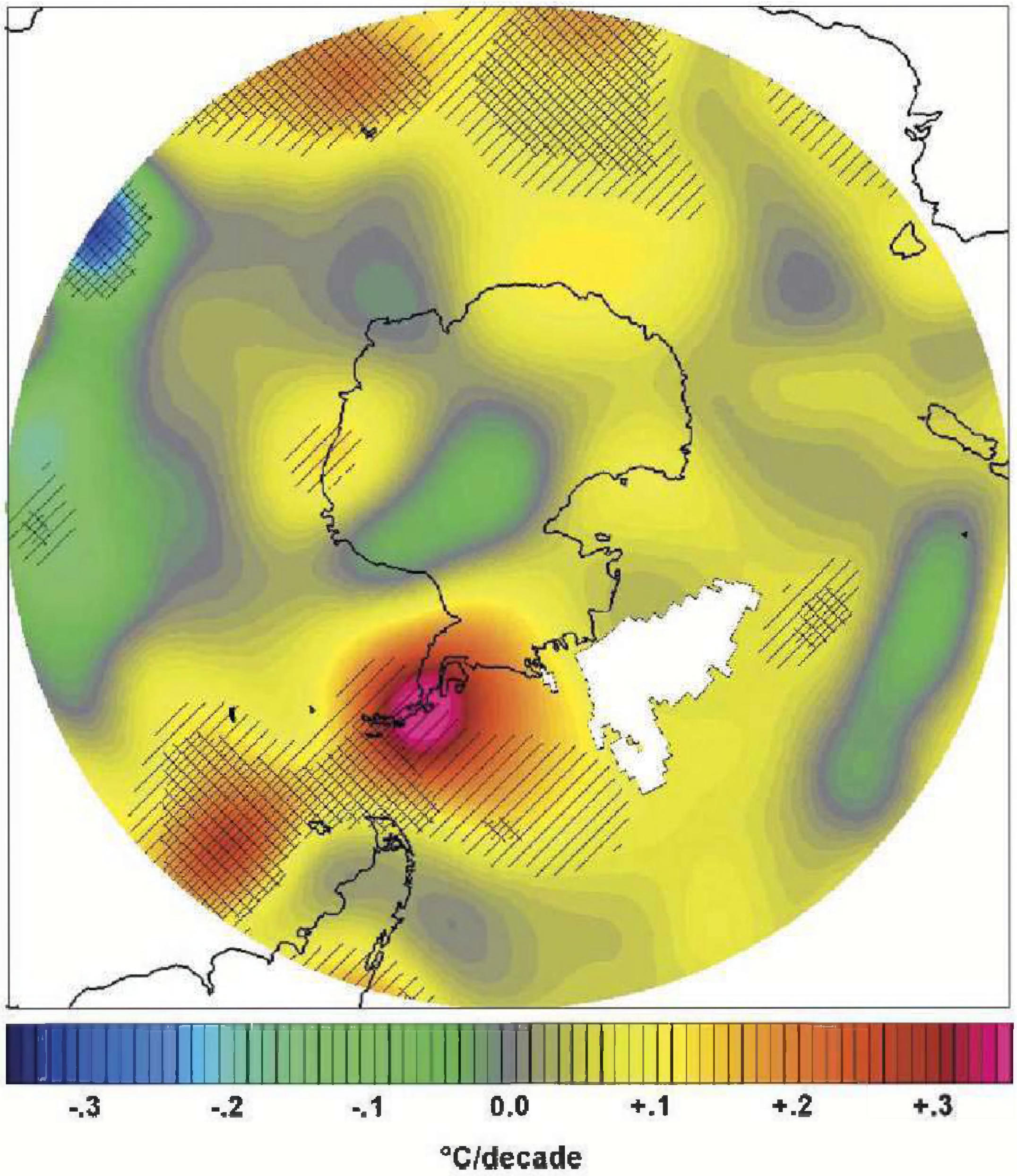

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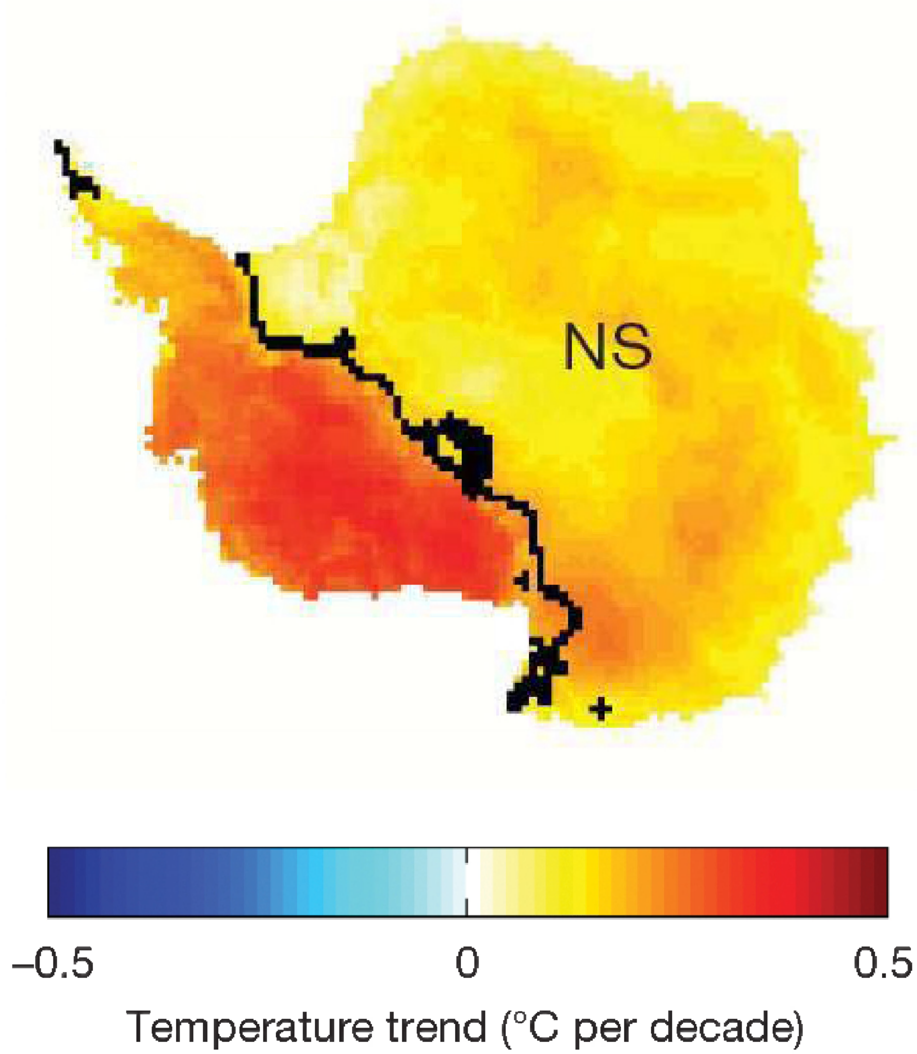


Figure 4.8. Estimates of Antarctic surface temperature trends. (a) Near-surface temperature trends for 1951-2006 based on station data. (b) Linear trends of annual mean surface air temperature (°C /dec) for the period 1958–2002. Greens and blues denote cooling; yellows and reds denote warming. Significant trends are indicated by hatching (95% = single hatching; 99% = crosshatching). From Chapman and Walsh (2007) (c) Winter season temperature trends reconstructed using infrared satellite data. NS indicates the trends are not significant in this area. From Steig et al. (2009).

Satellite-derived surface temperatures for the Antarctic have been used to investigate the extent of the region of extreme variability, since this was not possible with the sparse station data. King and Comiso (2003) found that the region in which satellite-derived surface temperatures correlated strongly with west Peninsula station temperatures was largely confined to the seas just west of the Peninsula. It was also found that the correlation of Peninsula surface temperatures with those over the rest of continental Antarctica was poor, confirming that the west Peninsula is in a different climate regime.

The warming on the western side of the Antarctic Peninsula has been largest during the winter season, with the winter temperatures at Faraday increasing by +1.03°C/decade over 1950-2006. In this area there is a high correlation during the winter between the sea ice extent and the surface temperatures, suggesting more sea ice during the 1950s and 1960s and a progressive reduction since that time. King and Harangozo (1998) found a number of ship reports from the Bellingshausen Sea in the 1950s and 1960s when sea ice was well north of the locations found in the period of availability of satellite data, suggesting some periods of greater sea ice extent than found in recent decades. However, there is very limited sea ice extent data before the late 1970s, so we have largely circumstantial evidence of a mid-century sea ice maximum at this time. At the moment it is not known whether the warming on the western side of the Peninsula has occurred because of natural climate variability or as a result of anthropogenic factors.

Temperatures on the eastern side of the Peninsula have risen most during the summer and autumn months, with Esperanza having experienced a summer increase of +0.41°C/decade between 1946-2006. This temperature rise has been linked to a strengthening of the westerlies that has taken place as the SAM has shifted into its positive phase (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 on the eastern side.

Around the rest of the Antarctic coastal region there have been few statistically significant changes in surface temperature over the instrumental period. The largest warming outside the Peninsula region is at Scott Base, where temperatures have risen at a rate of +0.29°C/decade, although this is not statistically significant. The high spatial variability of the changes is apparent from the data for Novolazarevskya and Syowa, which are 1,000 km apart. The former station has warmed at a rate of +0.25°C/decade between 1962–2000, which is significant at the 10% level, whereas the record from Syowa shows almost no change over this period.

One area of the Antarctic where marked cooling has been noted, at least over a relatively short time period, is the McMurdo Dry Valleys. Here automatic weather station (AWS) data for 1986-2000 shows that there has been a cooling of 0.7°C/decade, with the most pronounced cooling being in the summer (the December-February tend was 1.2°C/decade, statistically significant at the 2% level) and autumn (March-May trend of 2.0°C/decade, statistically significant at close to the 10% level). Winter (June-August) and spring (September-November) show small temperature *increases* of (0.6°C and 0.1°C/decade, which are not significant) (Doran et al., 2002). Bertler et al. (2004) suggest that this short-term cooling is associated with ENSO-driven changes in atmospheric circulation.

On the interior plateau, Amundsen-Scott Station at the South Pole has shown a small cooling in the annual mean temperature of -0.05° C/dec over 1958-2008, although this trend is not statistically significant. This small cooling is thought to be a result of fewer maritime air masses penetrating into the interior of the continent. The data show a cooling throughout the year, with the largest change being during the summer, however, only the annual change is statistically significant. The other plateau station, Vostok, has not experienced any statistically significant change in temperatures, either in the annual or seasonal data, since the station was established in 1958.

A recent analysis of the in-situ surface meteorological observations (Chapman and Walsh, 2007) gridded the available observations and analysed the trends. For the period 1958-2002 they found a modest warming over much of the 60°–90°S region, although the largest warming trends were over the Antarctic Peninsula. They also identified a zone of cooling stretching from Halley Station to the South Pole. They found overall warming in all seasons, with winter trends being the largest at +0.172°C/decade while summer warming rates were only +0.045°C/decade. For the 45 year period the temperature trend in the annual means was +0.082°C/decade. Interestingly the trends computed were very sensitive to start and end dates, with trends calculated using start dates prior to 1965 showing overall warming, while those using start dates from 1966 to 1982 show net cooling over the region. Because of the large interannual variability of temperatures over the continental Antarctic, most of the continental trends are not statistically significant significant through 2002.

The temperature records from the Antarctic stations suggest that the trends at many locations are dependent on the time period examined, with changes in the major modes of variability affecting the temperature data. Perhaps the largest change in climatic conditions across the high southern latitudes has been the shift in the SAM into its positive phase during austral summer and autumn (see Section 4.2.1.2). The SAM has changed because of the increase in greenhouse gases and the development of the Antarctic ozone hole, although the loss of stratospheric ozone has been shown to have had the greatest influence during Austral summer (Arblaster and Meehl, 2006). During austral autumn, the causality of the upward SAM trends is not well understood, as stratospheric ozone changes do not appear to play a major role (Fogt et al., In Press). As discussed at many points in this document, the changes in the SAM have influenced many aspects of the Antarctic environment over recent decades.

Thompson and Solomon (2002) considered the surface temperature trends over 1969-2000 and showed that the contribution of the SAM was a warming over the Antarctic Peninsula and a cooling along the coast of East Antarctica (Figure 4.9). They only considered the months of December to May, which was when the largest change in the SAM has taken place. They attributed the trends primarily to changes in the polar vortex as a result of the development of the Antarctic ozone hole. While the major loss of stratospheric ozone occurs in the spring, the greatest changes in the tropospheric circulation, such as the strengthening of the westerlies, has been in the summer and autumn. There is therefore a downward propagation of the vortex strengthening, with it starting in the spring in the stratosphere and moving down through the troposphere to the surface through the summer and autumn. As discussed earlier, the warming on the eastern side of the Antarctic Peninsula has been linked to the stronger westerlies associated with the changes in the SAM. However, the large winter-season warming on the western side of the peninsula appears to be largely independent of changes in the SAM.

It is very important to determine the surface temperature trends across the high Antarctic plateau, but as noted earlier, there are only two stations with long temperature records. Since the mid-1980s many AWSs have been deployed in the interior, filling important gaps in the observational network. These can provide valuable indications of temperature trends at remote locations, although few AWS systems have been maintained at the same locations since the 1980s and there can be gaps in the data when systems fail during the winter.

Another means of examining temperature trends is via the infra-red imagery from the polar orbiting satellites. Such imagery can only be used under cloud-free conditions, and provides data on the snow surface rather than at the standard meteorological level of 2 m above the surface, but with such high spatial coverage it provides a very valuable supplement to the in-situ observations. Comiso (2000) used the NOAA AVHRR imagery to investigate the trends in skin temperature across the Antarctic over the period 1979-1998. The satellite-derived temperatures were compared with the in-situ observations from 21 stations and found to be in good agreement with a correlation coefficient of 0.98. The trends showed a cooling across much of the high plateau of East Antarctica and across Marie Byrd Land (-0.1 to -0.2° C/yr), with the former being consistent with the trends derived by Thompson and Solomon (2002), although the Comiso trends are for annual data and the Thompson and Solomon study is concerned only with the summer and autumn.

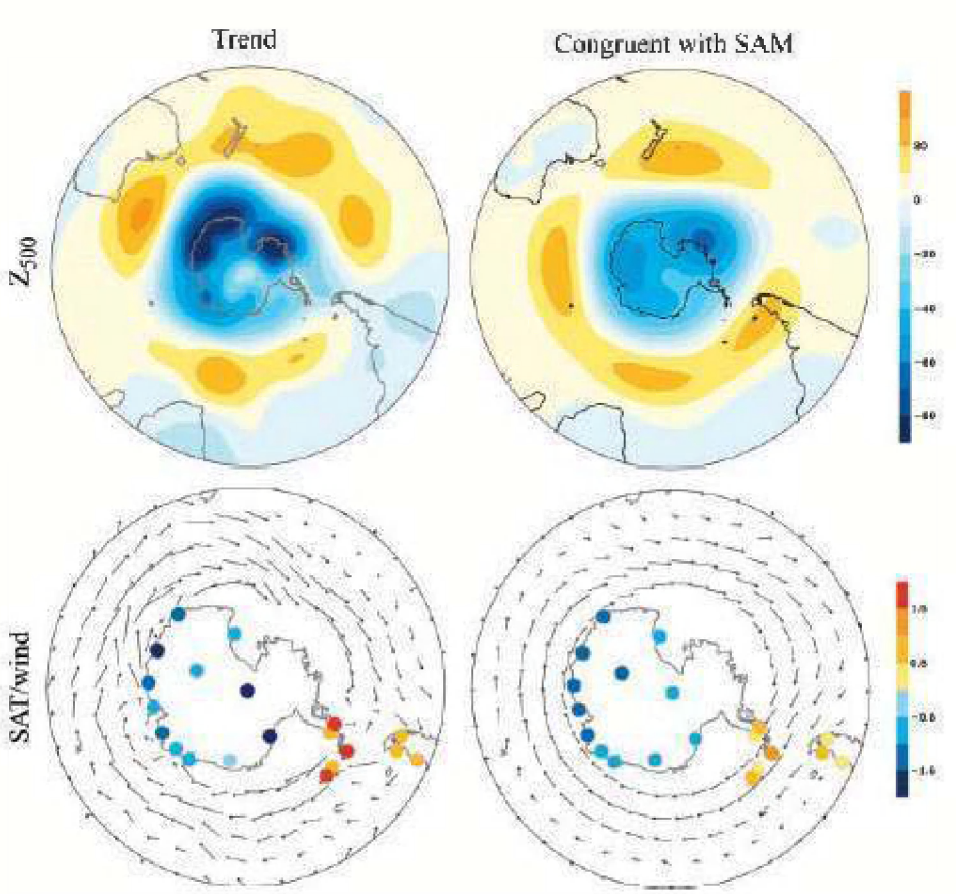


Figure 4.9. December-May trends (left) and the contribution of the SAM to the trends (right). Top, 22-year (1979-2000) linear trends in 500 hPa geopotential height. Bottom: 32-year (1969-2000) linear trends in surface temperature and 22-year (1979-2000) linear trends in 925 hPa winds. Shading is drawn at 10 m per 30 years for 500 hPa height and at increments of 0.5° K per 30 years for surface temperature. The longest vector corresponds to about 4 m/s. From Thompson and Solomon (2002).

In recent decades many relatively short ice cores have been drilled across Antarctica by initiatives such as the International Trans Antarctic Science Expedition (ITASE) (Mayewski et al, 2006). These provide data over roughly the last 200 years and therefore provide a good overlap with the instrumental data. Large-scale calibrations have been carried out between satellite-derived surface temperature and ITASE ice core proxies (Schneider et al., 2006). Their reconstruction of Antarctic mean surface temperatures over the past two centuries was based on water stable isotope records from high-resolution, precisely dated ice cores. The reconstructed temperatures indicated large interannual to decadal scale variability, with the dominant pattern being anti-phase anomalies between the main Antarctic continent and the Antarctic Peninsula region, which is the classic signature of the SAM. The reconstruction suggested that Antarctic temperatures had increased by about 0.2° C since the late nineteenth century. They found that the SAM was a major factor in modulating the variability and the long-term trends in the atmospheric circulation of the Antarctic.

4.3.2 Upper air temperature changes

Analysis of Antarctic radiosonde temperature profiles indicates that there has been a warming of the troposphere and cooling of the stratosphere over the last 30 years. This is the pattern of change that would be expected from increasing greenhouse gases, however, the mid-tropospheric warming in winter is the largest on Earth at this level. The data show that regional mid-tropospheric temperatures have increased most around the 500 hPa level with statistically significant changes of 0.5 – 0.7ºC/decade (Figure 4.10) (Turner et al., 2006). Figure 4.10 indicates warming at many of the radiosonde stations around the continent, but a clear pattern of winter warming is apparent around the coast of East Antarctica and at the pole.

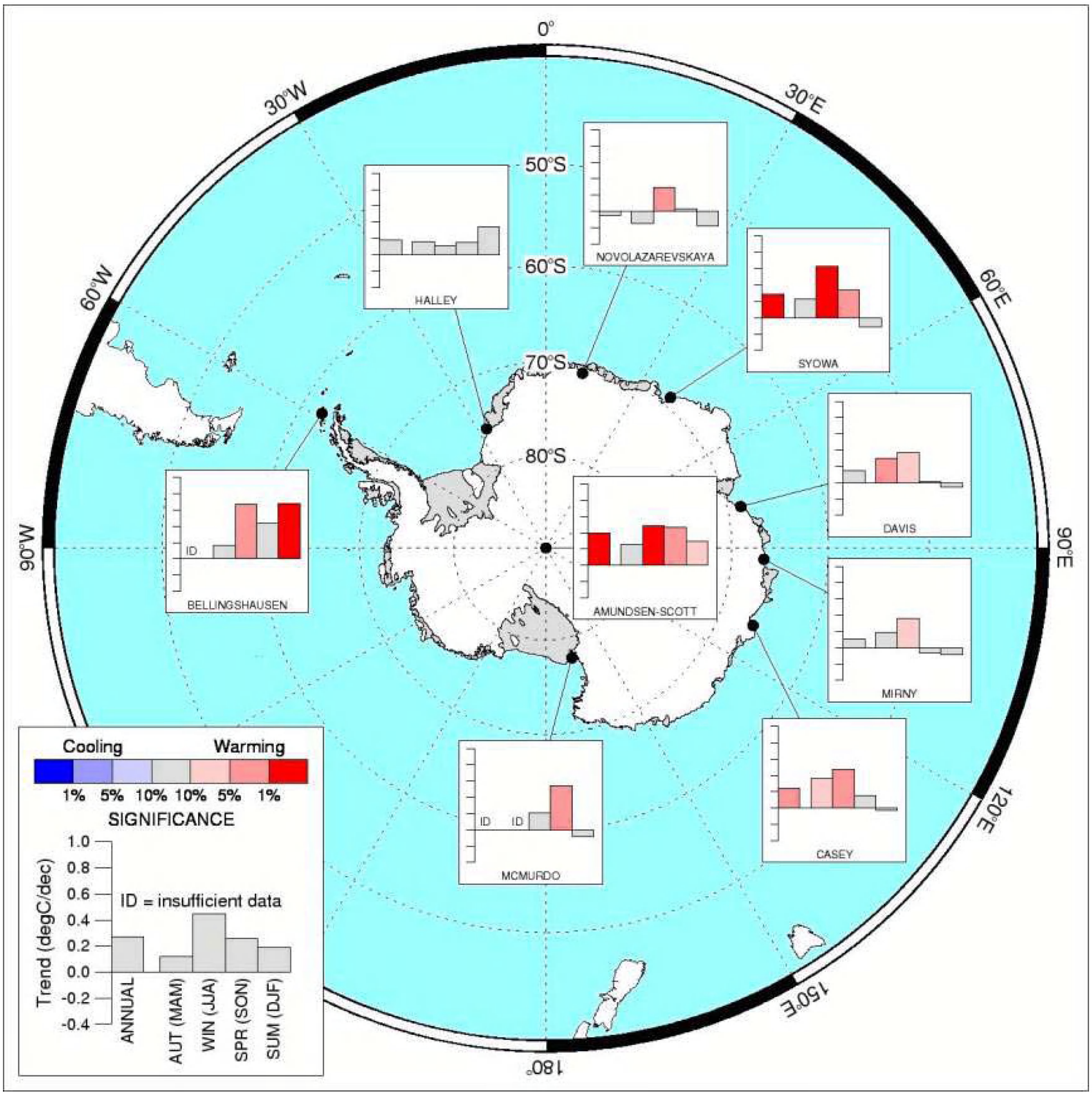


Figure 4.10. Annual and seasonal 500 hPa (approximately at 5 km above mean sea level) temperature trends for 1971-2003. From Turner et al. (2006).

The warming is represented in the ECMWF 40 year reanalysis, which is not surprising since the radiosonde ascents were assimilated into the system. In fact the warming trends are slightly larger than when computed from the radiosonde data, since there is a known slight cold bias in the early part of the reanalysis data set.

The exact reason for such a large mid-tropospheric warming is not known at present. However, it has recently been suggested that it may, at least in part, be a result of greater amounts of polar stratospheric cloud (PSC) during the winter (Lachlan-Cope et al., In press). PSCs are a feature of the cold Antarctic winter, forming at temperatures below about -78° C. However, the Antarctic stratosphere has cooled in recent decades because the greenhouse gas ozone is now missing from the lower stratosphere in spring, and the greenhouse gas carbon dioxide is concentrated in the troposphere and leads to further cooling of the stratosphere. Analysis of stratospheric temperatures in the reanalysis data sets suggest that over the last 30 years the area where PSCs might form in winter has increased in size, so promoting the formation of more PSCs. Once present, PSCs act like any other cloud, giving a warming below their level and cooling above. We have little data on the optical properties of PSCs, but modelling suggests that if the optical depth in the infrared is around 0.5 then a greater amount of PSCs could give a mid-tropospheric warming.

PSCs are not currently represented explicitly in climate models, but if further research shows that they are responsible for the large winter season mid-tropospheric warming they need to be represented more realistically in the models.

4.3.3 Attribution of change

Great advances have been made in our understanding of recent temperature changes across the Antarctic in the last few years. We now know that anthropogenic activity, and particularly the presence of the Antarctic ozone hole, has played a large part in the near-surface warming on the eastern side of the Antarctic Peninsula, and a formal attribution study (Gillett et al., 2009) found that that recent changes were not consistent with internal climate variability. However, we still do not know the reasons for the large winter season warming on the western side, although it is thought to be linked to a reduction in sea ice extent since the 1950s.

The recently discovered large mid-tropospheric warming above the continent in winter is not fully understood at present. If increasing amounts of PSCs are shown to be responsible this will be an interesting Antarctic amplification of the effect of greenhouse gas increases, along with being a side effect of the ozone hole. However, more research is needed to confirm this.

4.4 Changes in Antarctic Snowfall Over the Past 50 Years

4.4.1 General spatial and temporal characteristics of Antarctic snowfall

Snowfall accumulation, referred to as the surface mass balance (SMB), is the primary mass input to the Antarctic ice sheets, and is the net result of precipitation, sublimation/vapour deposition, drifting snow processes, and melt. Precipitation, which primarily occurs as snowfall, is dominant among these components (Bromwich 1988) and establishing its spatial and temporal variability is necessary to assess ice sheet surface mass balance. Comprehensive studies of snowfall characteristics over Antarctica are given by Bromwich (1988), Turner et al. (1999), Genthon and Krinner (2001), van Lipzig et al. (2002), Bromwich et al. (2004a), van de Berg et al. (2005), and Monaghan et al. (2006a). Snowfall is influenced to first order by the Antarctic topography, and ground-penetrating radar has shown that its spatial distribution is highly variable. Most of the snowfall occurs along the steep coastal margins (Figure 4.11) and is caused by orographic lifting of relatively warm, moist air associated with the many transient, synoptic-scale cyclones that encircle the continent (e.g. Bromwich et al. 1995; Genthon and Krinner, 1998). The synoptic activity decreases inward from the coast, and over the highest, coldest reaches of the continent the primary mode of snowfall is due to cooling of moist air just above the surface-based temperature inversion (Schwerdtfeger, 1970). This extremely cold air has little capacity to hold moisture, and thus the interior of the East Antarctic Ice Sheet is a polar desert, with a large area that receives less than 5 cm water equivalent of snowfall each year (e.g. Vaughan et al., 1999; Giovinetto and

Zwally, 2000). Large-scale atmospheric influences on Antarctic snowfall include the ENSO (Cullather et al., 1996) and the SAM (Genthon and Cosme, 2003; van den Broeke and van Lipzig; 2004). ENSO has an intermittent teleconnection with Antarctica (Genthon and Cosme, 2003) that especially impacts snowfall variability in West Antarctica (Cullather et al., 1996; Bromwich et al., 2000; 2004b; Guo et al., 2004). The response of Antarctic snowfall to SAM forcing is complex (Genthon et al., 2003), but may be linked to near-surface wind flow and temperature anomalies that are associated with the SAM (van den Broeke and van Lipzig, 2004).

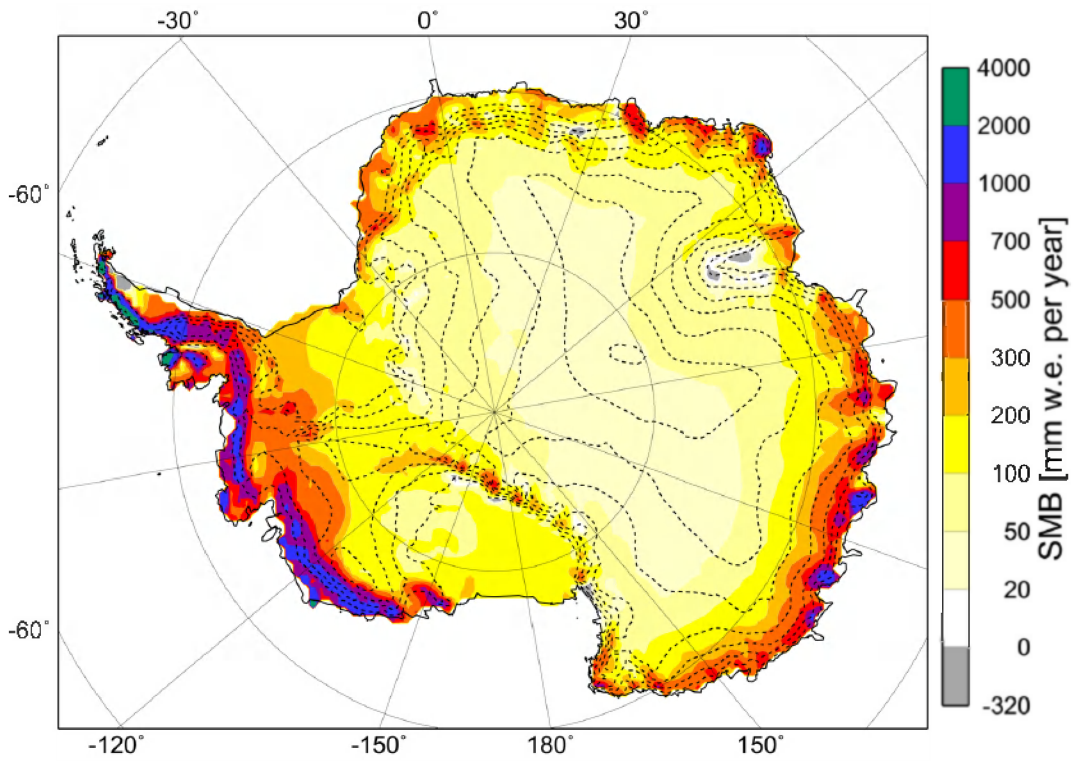


Figure 4.11. Annual Antarctic surface mass balance (mm/yr), from van de Berg et al. (2006).

4.4.2 Long-term Antarctic snowfall accumulation estimates

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

4.4.3 Recent trends in Antarctic snowfall

On average, about 6 mm global sea level equivalent falls as snow on Antarctica each year (Budd and Simmonds, 1991). Thus, it is important to assess trends in Antarctic SMB, as even small changes can have considerable impacts on the global sea level budget. The latest studies employing global and regional atmospheric models to evaluate changes in Antarctic SMB show that no statistically significant increase has occurred since ~1980 over the entire grounded ice sheet, WAIS, or the East Antarctic Ice Sheet (EAIS) (Monaghan et al., 2006a; van de Berg et al., 2005; van den Broeke et al., 2006). A validation of the modeled-versus-observed changes (Monaghan et al., 2006a) suggests that the recent model records are more reliable than the earlier global model records that inferred an upward trend in Antarctic SMB since 1979 (Bromwich et al., 2004a). The new studies also clearly show that interannual SMB variability is considerable; yearly snowfall fluctuations of +/-20 mm/yr weq, i.e., +/-0.69 mm/yr GSL (global sea level) equivalent, are common (Monaghan et al., 2006a), and might easily mask underlying trends over the short record.

In contrast to modeling studies, satellite altimetry measurements by Davis et al. (2005) suggest that increased snowfall has recently caused the EAIS to thicken, mitigating sea level rise by about 0.12 mm/yr between 1992-2003. Zwally et al. (2005) also found a thickening over EAIS from satellite altimetry for a similar period, but it was a factor of three smaller than the Davis study. Zwally et al. (2005) argued that their method more accurately accounts for firn compaction and the interannual variability of surface height fluctuations. The difference between the positive trends from the satellite altimetry studies and the zero trends in the modeling studies may be in part due to different temporal and spatial coverage (satellite altimetry does not extend past 81.6°S and has limitations along the steep coastal margins).

To extend the length of the Antarctic SMB record, Monaghan et al. (2006b) used the spatial information provided by atmospheric model precipitation fields from ERA-40 to extrapolate a suite of ice core SMB records in space and time. The resulting spatially resolved SMB dataset spans 1955-2004, approximately doubling the length of the existing model-based records. An updated version of the dataset (Monaghan and Bromwich, 2008), now adjusted to reflect the ERA-40 snowfall variance at interannual timescales, indicates that the 1955-2004 continent-averaged trend is positive and statistically insignificant (0.19 ± 32 mm/yr), and is characterized by upward trends through the mid-1990s and downward trends thereafter (Figure 4.12). The shape of the time series in Figure 4.12 suggests that a cyclic signal with a period of about 50 years may be impacting Antarctic snowfall, but is inconclusive without a longer time series. The continent-averaged trend is the net result of both positive and negative regional trends, which in some drainage basins are weakly ($p<0.10$) statistically significant (Figure 4.13). The positive SMB trends on the western side of the Antarctic Peninsula have been attributed to a deepening of the circumpolar pressure trough, which has enhanced ascent in the region (Turner et al., 2005b).

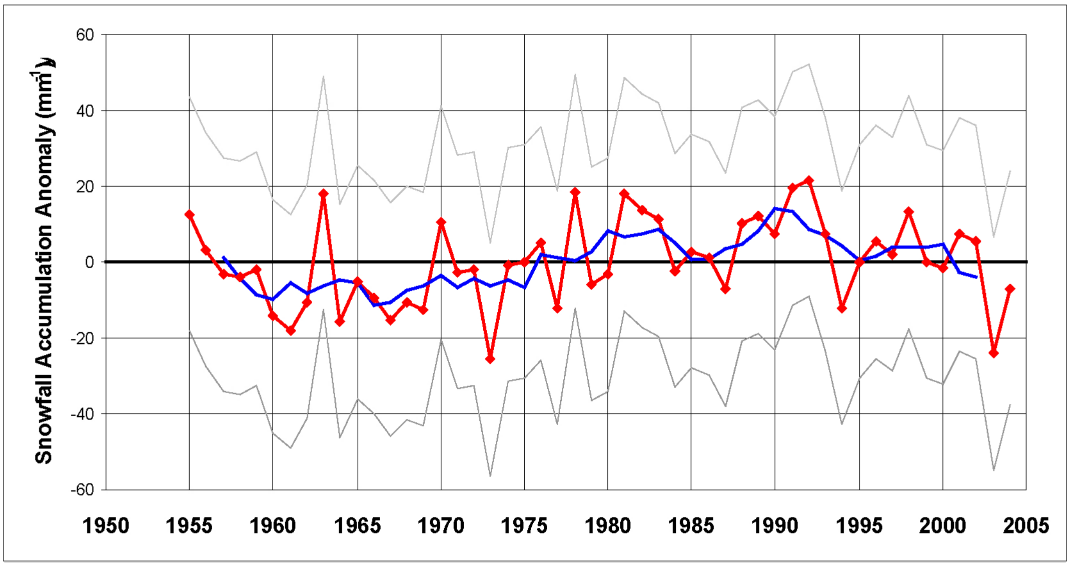


Figure 4.12 Annual Antarctic snowfall accumulation anomalies (mm/yr), in red) during 1955-2004, adapted from Monaghan et al. (2006b) and Monaghan and Bromwich (2008). Anomalies are with respect to the 1955-2004 mean. The five-year running mean is shown in blue. The ±95% confidence intervals for the annual anomalies are shown in grey. The statistical uncertainty accounts for the variance, as well as that due to the methodology and measurement error.

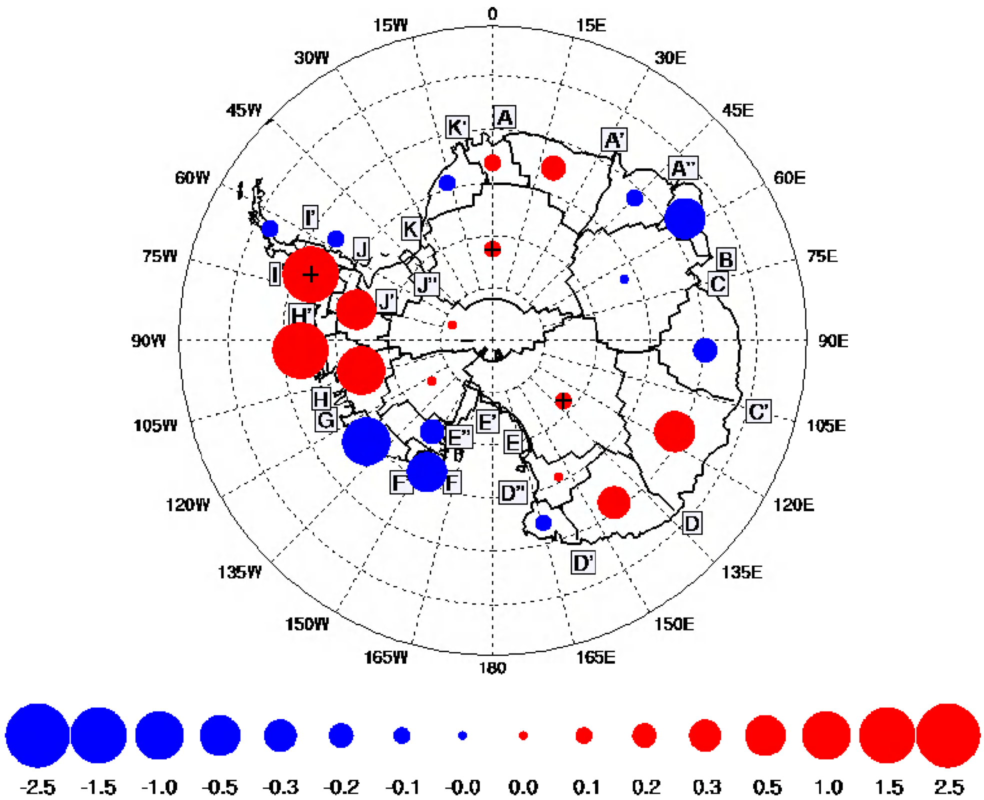


Figure 4.13 Linear trends of annual Antarctic snowfall accumulation (mm/year) for 1955-2004 in each of the Antarctic glacial drainage basins, adapted from Monaghan and Bromwich (2008). Statistical significance is represented by symbols: "+" is p<0.10; "*" is p<0.05; and "Δ" is p<0.01. The statistical uncertainty accounts for the variance, as well as that due to the methodology and measurement error.

4.5 Atmospheric Chemistry

4.5.1 Antarctic stratospheric ozone in the instrumental period

Historically ozone values were around 300 Dobson Units (DU) at the beginning of the winter (March), and similar at the end (August). The pattern began to change in the 1970s, following widespread releases of CFCs and Halons in the atmosphere (see below). Now, at the end of August values are about 10% less than they were in the 1970s, and decrease at about 1% per day to reach about 100 DU in October. Most of this loss occurs between 14 and 22 km altitude within the polar vortex, where virtually all ozone is now destroyed (Figure 4.14). Ozone values substantially recover with the warming in late spring, when the vortex dissipates and air from outside is mixed (Figure 4.15).

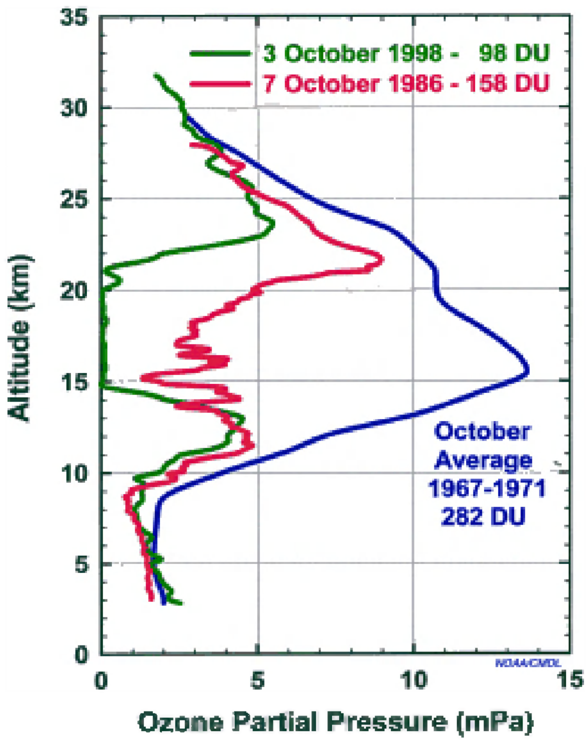


Figure 4.14 South pole ozone profiles, showing the progressive thinning of the ozone layer in late spring as the ozone hole developed during the 1980s and 1990s (courtesy NOAA/CMDL).

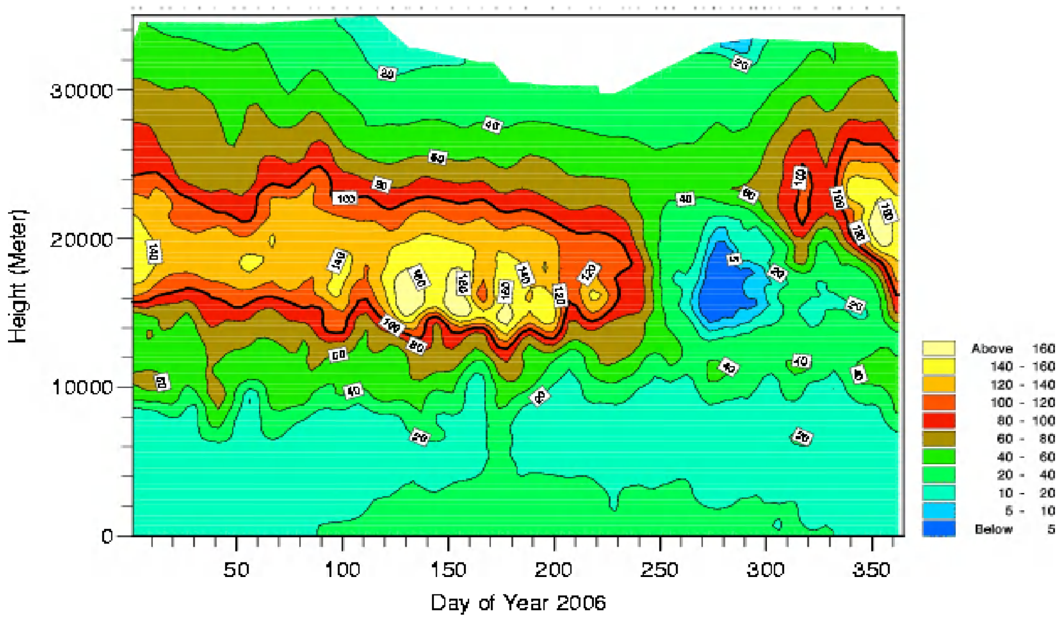


Figure 4.15 The annual cycle of ozone (nbar) in 2006 at Neumayer, 71°S (courtesy Alfred Wegener Institute).

The ozone hole is caused by reactive chlorine and bromine gases, formed from the breakdown products of CFCs and Halons, which were liberated in the troposphere from spray-cans, refrigerators and fire extinguishers. They have a typical stratospheric lifetime of 50 to 100 years. But the development of the ozone hole is also strongly linked to the dynamics of the polar vortex because it acts as a barrier (Figure 4.16a). During winter, lower stratospheric temperatures drop below –80°C, and at these temperatures clouds form despite the dryness of the stratosphere, initially composed of nitric acid trihydrate, but of ice if 5 to 10°C colder. On the cloud surfaces, the degradation products of the CFCs react to form chlorine gas:

$$HCl + ClNO_3 \text{ -> } HNO_3 + Cl_2$$

followed by photo-dissociation to chlorine atoms when sunlit, then reaction with ozone to form highly reactive ClO. Similar reactions take place involving the bromine compounds that result from degradation of halons. The highly reactive chlorine and bromine compounds then take part in a series of photochemical reactions such as:

$$Cl + O_3 \rightarrow ClO + O_2$$
$$ClO + O_3 \rightarrow Cl + 2O_2$$

in which in effect the Cl acts as a catalyst and gets recycled time and again. The net result of the catalysis is that:

$$2O_3 \rightarrow 3O_2$$

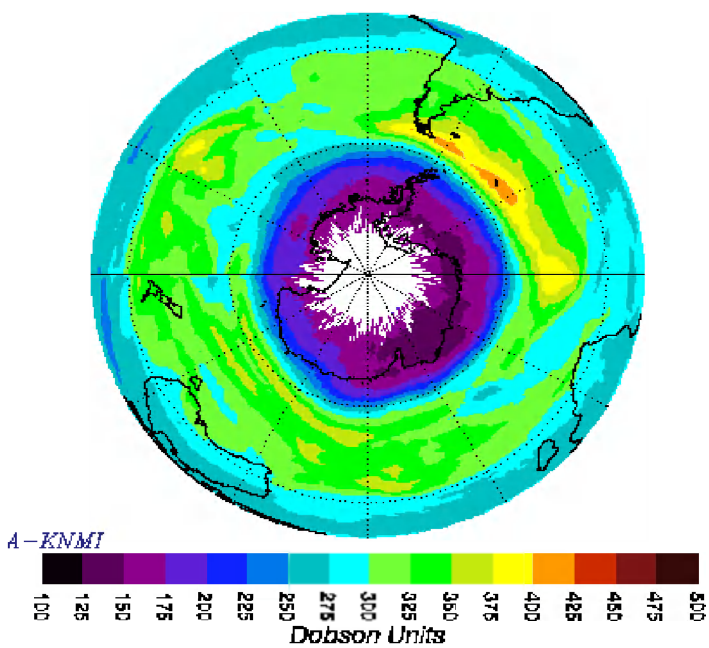
The process is helped by the absence of NO_2 (converted to HNO_3 and absorbed into the clouds), which would otherwise react with the ClO to recreate $ClNO_3$ and so remove reactive gases from the catalytic cycle.

As the vortex warms the clouds disappear, but the reactive chlorine and bromine compounds continue the ozone depletion for some weeks, until converted back to HCl and HBr. With further warming the vortex begins to break down and the sub-polar ozone-rich air sweeps across the continent.

The ozone hole is often offset from the pole towards the Atlantic, reaching as far north as 50°S. Ozone provides a screen against ultraviolet light of wavelength shorter than about 315 nm (UV-B), which can cause sunburn, cataracts and skin cancer in humans. Hence such an offset poses a serious health threat to the inhabitants of southern South America. It also poses a risk for flora and fauna, as UV-B can damage DNA, and can bleach chlorophyll that then becomes non-functional (see Section 4.12).

Another effect of the ozone hole is on the temperatures within the vortex, because ozone is a greenhouse gas that absorbs solar radiation. The absence of ozone has therefore resulted in significant reductions in temperature in spring (Figure 4.17), in November reaching a difference of up to 15°C in comparison with pre-ozone hole years.

The Montreal Protocol is an international agreement that has phased out production of CFCs, Halons, and some other organic chlorides and bromides, collectively referred to as Ozone Depleting Substances (ODSs). Because of its success, the amounts of ODSs in the stratosphere are now starting to decrease. However, there is as yet no convincing sign of any reduction in the size or depth of the ozone hole, although the sustained increases up the 1990s have not continued. Recent changes in measures of Antarctic ozone depletion have ranged from little change over the past 10 years (ozone hole area), to some signs of ozone increase (Bodeker et al., 2005). The halt in rapid ozone hole growth can be ascribed to the fact that almost all of the ozone between 12 and 24 km in the core of the vortex is now being destroyed (WMO, 2002), and is therefore comparatively insensitive to small changes in ODS amount.



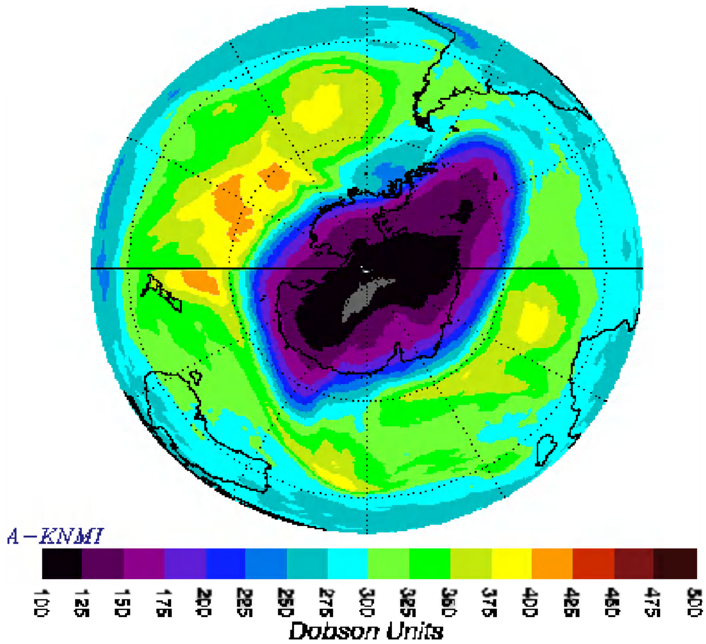


Figure 4.16 Measurements by the Ozone Monitoring Instrument on the Aura satellite, with a scale in DU. Top – on 14 September 2006, showing an almost symmetric ozone hole covering all of Antarctica. Bottom – on 10 October 2006, showing the ozone hole extending towards South America. (courtesy NASA/GSFC).

4.5.2 Antarctic Tropospheric Chemistry

Although often thought of as a single unit, the atmosphere is divided up into a number of regions. These are determined by the temperature gradient with height. The lowest region is referred to as the troposphere. In the troposphere, temperature generally decreases with height until a point is reached where this trend reverses. This upper limit is referred to as the tropopause. The height of the tropopause varies with latitude and is roughly at 8 km above hand or sea in the polar regions. The troposphere itself is nominally subdivided into layers; the lowest is the "boundary layer", a part of the troposphere that is directly influenced by the surface of the Earth in the exchange of heat, momentum and moisture. The height of the boundary layer is determined by physical constraints such as temperature and wind speed, and over Antarctica can vary considerably from tens to hundreds of metres. Above the boundary layer is the free troposphere, a region remote from the direct influence of the Earth's surface.

Over many regions of the world, the chemistry of the troposphere is studied in order to understand the effect of emissions from human activities. These might be direct emissions from industrial processes, or emissions associated with, for example, agriculture. Such activities release relatively reactive and short-lived trace gases into the atmosphere, and change it considerably from its natural state. Antarctica supports no major population centres and lies at a considerable distance from anthropogenic emission sources, so although some longer-lived pollutants do reach Antarctica, the Antarctic troposphere is a relatively unperturbed natural background atmosphere.

Compared with many other aspects of Antarctic science, the chemistry of the Antarctic troposphere has received relatively little attention. A primary reason for this was the perception that the chemical composition would be relatively uninteresting, with low concentrations of reactive trace gases like the hydroxyl (OH) or hydroperoxy (HO_2) radicals or the nitrogen oxides (NO and NO_2), and merely a sluggish chemistry dominated by unreactive reservoir gases that had been transported from distant source regions. Atmospheric chemists interested in studying a clean background atmosphere would naturally choose to work in a location that was more easily accessible and with more benign ambient conditions.

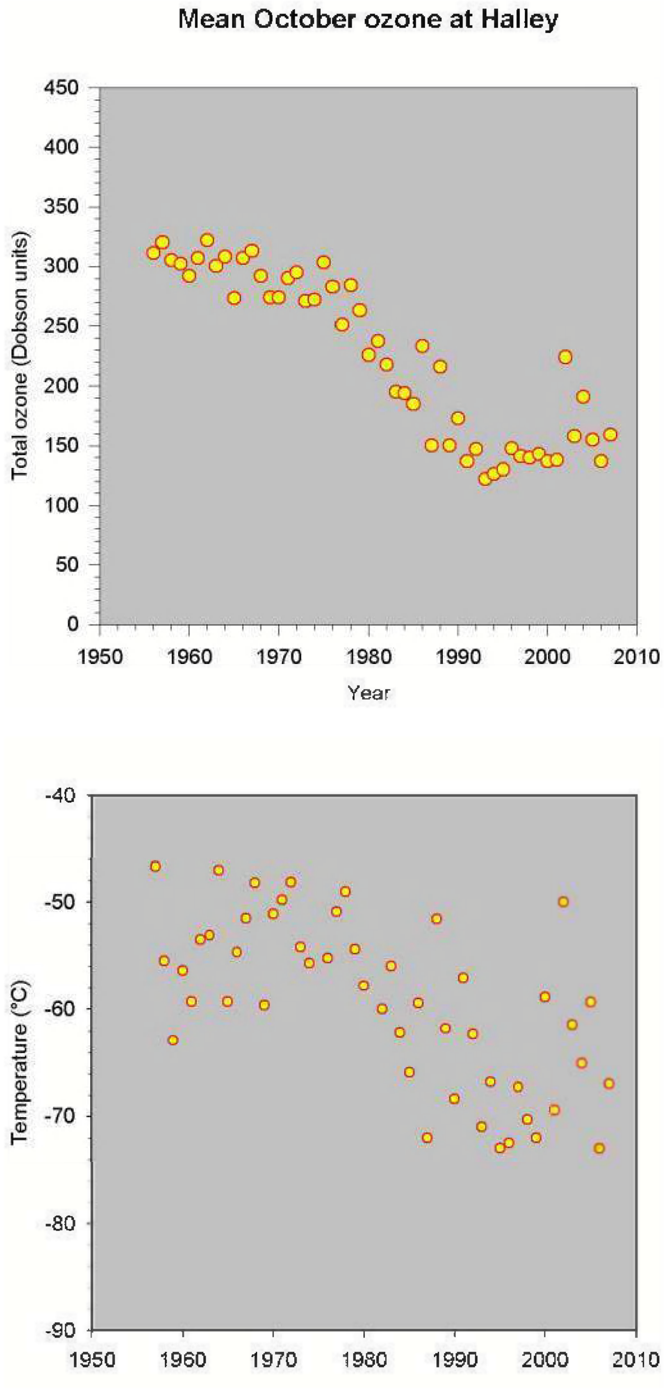


Figure 4.17 Time series of measurements at Halley (76°S) of (top) total ozone averaged over October of each year, and (bottom) temperature at 100 hPa averaged over November of each year (courtesy British Antarctic Survey). Although there is a lot of variability in the temperatures, they are now about 10°C colder than in the 1960s and 1970s, with the exception of 2002. The change in temperature is a maximum in November, later than the

maximum change in ozone, because there is more sunlight later in the year, and because the thermal time constant at 100 hPa exceeds a month. The anomalous results in 2002 were the result of an anomalous early breakdown of the vortex, they are not indicators of any large reduction in the chlorine and bromine gases that cause ozone loss.

Nevertheless, Antarctica is a significant part of the Earth system, and studies of Antarctic tropospheric chemistry have gradually become a recognised part of the work of national operators. A major driver has been the fact that deep ice cores are drilled from Antarctic ice sheets from which paleo-scientists strive to reconstruct changes in the Earth's atmospheric composition and climate through time. This work relies on analysing and interpreting changes in impurities held within the ice cores. A correct interpretation relies on knowing how the impurity entered the ice and any associated depositional or post-depositional effects. It would be cavalier to believe we could correctly reconstruct a past atmosphere from these chemical tracers without properly understanding the tropospheric chemistry of the present day.

The early studies of Antarctic tropospheric chemistry focused mainly on aerosols and long-lived radiatively and stratospherically important gases. Aerosols are important components of ice core impurities and can act as valuable proxies for environmental changes through time. For example, sea salt is a prime component of aerosol in coastal Antarctica, and sodium and chloride are both easily measured in ice cores. Studies of sea salt aerosol have been necessary to determine dominant sources and as a fingerprint for the behaviour of marine air masses. They have shown that, whereas for most of the globe the open ocean is the source of sea salt, in the polar regions most sea salt may be generated within and adjacent to the zone of newly-forming sea ice. This suggests that sea salt measured in ice cores might provide a proxy for assessing the extent of sea ice and how this varied under different climatic conditions. The records of long-lived gases have provided invaluable evidence of how the global atmosphere has recently changed. For example, the record of boundary layer carbon dioxide (CO_2), which has been measured at South Pole since 1957, has shown the massive rise in this potent greenhouse gas, and importantly, has bridged the gap between ice core records of CO_2 and present day ambient measurements. Also, systematic continuous measurements of the atmospheric CO_2 concentration at Syowa Station since 1984 revealed clear evidence for a seasonal cycle, a secular trend and interannual variations (Morimoto et al., 2003). The seasonal cycle varied from year to year, with especially large amplitudes in 1992 and 1998 and a large phase delay in 1993. A rapid increase in the CO_2 concentration was observed in 1987, 1994 and 1998 in association with ENSO events, and very low increase rate in 1991 to 1993, related to the Pinatubo eruption. From measurements of the stable carbon isotope ratio ($\delta^{13}C$) of atmospheric CO_2, it was found that the rapid increase of the CO_2 concentration was accompanied by a rapid decrease of $\delta^{13}C$. Considering the fact that the $\delta^{13}C$ values of terrestrial biospheric CO_2 are lower than those of atmospheric CO_2, the interannual variations of the CO_2 increase are ascribed primarily to changes of the CO_2 exchange between the atmospheric and terrestrial biosphere due to climate change in association with an ENSO event. It has been confirmed that long term continuous observations of greenhouse gases are essential in the Antarctic to monitor climate variations.

To clarify variations at the surface and in the transport of greenhouse gases, it was indispensable to know the vertical distributions of concentrations. Balloon-borne campaigns using a cryogenic sampler with a large balloon were carried out at Syowa Station in 1998 and 2003/04 to examine the vertical distribution of greenhouse gas concentrations up to 30 km in the stratosphere (Aoki et al., 2003). Together with bi-polar balloon-borne observations in the Arctic in 1997 and similar observations over Japan since 1985, stratospheric CO_2 above 20 – 25 km showed a secular increase with an average rate of 1.5 ppmv/yr, slightly less than the rate at the surface, and some increase of CO_2 age in the stratosphere. The O_2/N_2 ratio from these stratospheric air samples was analyzed to better understand the global carbon cycle

(Ishidoya et al., 2006). The vertical profile of O_2/N_2 showed a gradual decrease with height in the stratosphere, indicating the gravitational separation of molecules.

Beyond its role as an archive of global change, studies of the Antarctic troposphere have revealed a highly individual and active chemical system that is likely itself in the future to be an active player within a changing climate system. The rest of this section details this chemistry.

Of the more reactive trace gases, only surface ozone has historically been measured with any vigour. Year-round measurements were made at Halley station as early as 1958, but continuous records began considerably later, in 1975 at South Pole and in the 1980s at McMurdo/Arrival Heights and at Neumayer. Today, surface ozone is measured routinely at four coastal sites (those mentioned above plus Syowa) as well as at Sanae, some 170 km inland, and at South Pole on the Antarctic plateau (Helmig et al., 2007). The records show both interesting similarities and differences (see Figure 4.18). At nearly all stations, surface ozone reaches its maximum concentration during the winter months and is at its minimum during the summer. This is the classic seasonal cycle for a trace gas whose concentration is balanced by increases arising from air mass transport and destruction by the direct action of the Sun or by sunlight-initiated chemistry.

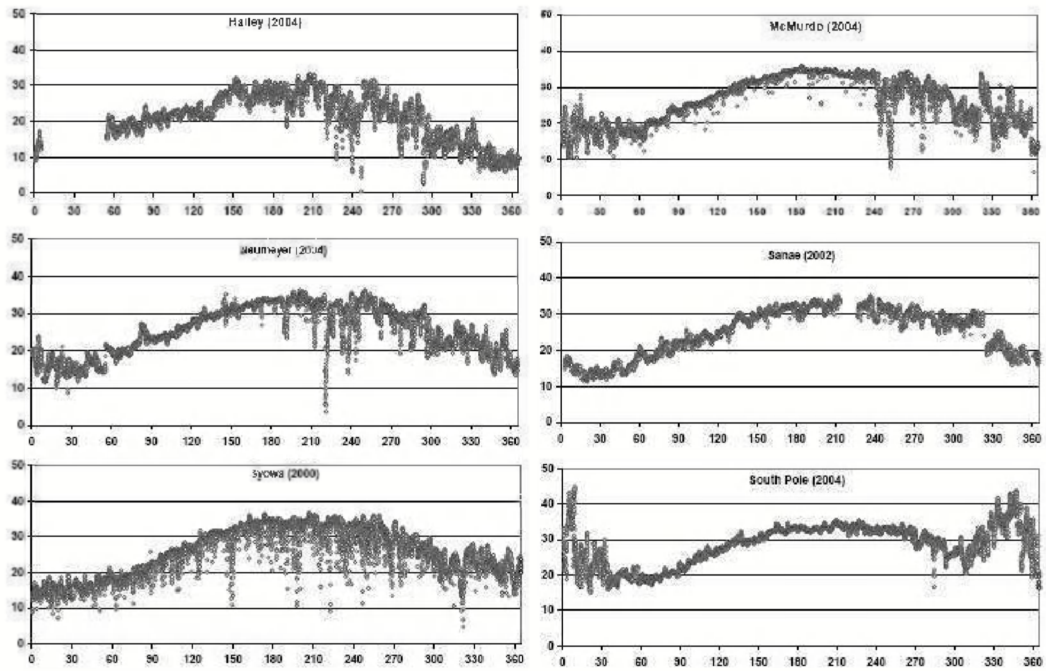


Figure 4.18 Annual records (against year calendar day) of surface ozone (in parts per billion by volume (ppbv)) for the Southern Hemisphere stations Halley, MacMurdo, Neumayer, Sanae, Syowa and South Pole, reproduced from Helmig et al. (2007).

During the Antarctic spring, however, significant differences are evident between the coastal sites and those lying inland. While at South Pole and Sanae, the decline from the winter maximum towards summertime values is essentially smooth, surface ozone at coastal sites exhibits extremely rapid and large episodic losses during the spring months (Wessel et al., 1998; Jones et al., 2006). These ozone depletion events (ODEs) can last for several days and ozone concentrations can drop as low as instrumental detection limits.

This behaviour is natural and occurs at coastal sites in both polar regions. The ozone loss is driven by reactions with halogen atoms, primarily bromine, in chemical cycles analogous to stratospheric ozone depletion. The following reactions were proposed to explain ODEs observed in the Arctic (Barrie et al., 1988):

$$2 \times (O_3 + Br \rightarrow BrO + O_2)$$
$$BrO + BrO \rightarrow 2Br + O_2$$
$$\text{net:} \quad 2\,O_3 \rightarrow 3\,O_2$$

Key to this process is the fact that bromine is recycled from bromine monoxide (BrO) to bromine atoms (Br) without the production of ozone. Ozone is therefore destroyed in a catalytic cycle whereby the bromine atoms responsible are regenerated and ready to react again with other ozone molecules. Other radicals, such as ClO, IO or HO_2 can also be involved in BrO recycling. For a full discussion see the review by Simpson et al. (2007).

Such recycling of course does not generate "new" halogens, so the important question is – what is the original source of the bromine atoms? Theoretical and laboratory studies have demonstrated that a series of reactions, widely referred to as the "Bromine Explosion" (shown schematically in Figure 4.19), are the source.

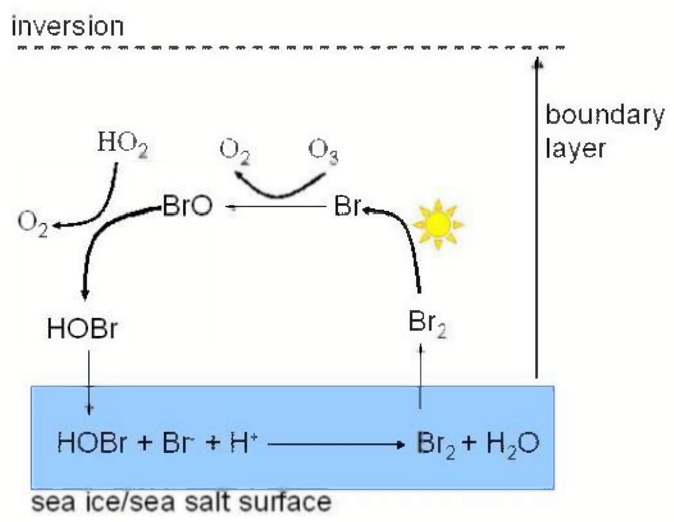


Figure 4.19 A schematic representation of the 'Bromine Explosion'.

Bromide (Br^-) is a ubiquitous ion found in sea water, albeit at concentrations ~280 times lower than chloride (Cl^-). Bromide derived from sea water reacts with HOBr, a molecule with a gas-phase source, and produces bromine molecules that are then released back into the boundary layer. The action of sunlight splits the bromine molecule into two bromine atoms, and ozone destruction can commence. The process is contained within the boundary layer. This reaction process requires a liquid (or "quasi-liquid") phase with concentrated sea salt brine, and the exact nature of this is still under debate. There is some evidence that this surface is associated with newly-forming sea ice (Rankin et al., 2002; Jones et al., 2006), and other evidence suggesting that sea salt-laden snow plays a role (McConnell et al., 1992; Simpson et al., 2007). Certainly, sea salt (*via* sea ice/aerosol etc.) is fundamental as the prime bromide source, and anything that might affect its availability is likely to affect the frequency of ozone depletion events.

The surface ozone records discussed above also reveal a second surprising feature. During the summer months at South Pole, surface ozone concentrations increase even above those measured during winter. This suggests that, rather than experiencing net destruction, ozone is being produced *in situ*. The only route for this within the troposphere is through photolysis of NO_2, but NO_2 is normally associated with polluted air and has low concentrations in pristine atmospheres. The questions then arise – what is the source of this ozone, and why, out of all the measurement locations, is it only evident at South Pole?

The answers lie within a new area of atmospheric science; that of snow photochemistry. The traditional view of polar snow was that it had an important influence on albedo, and also that it acted as a cap to exchange of trace gases between the boundary layer and the land or sea surfaces below. However, an active chemical role had not been anticipated. That view has now been overturned, and it has been shown through many chemical studies in the field, in the laboratory, and in modelling calculations, that snow is a major source of reactive trace gases to the polar boundary layer both through physical and photochemical release processes (see review by Grannas et al., 2007). For example, nitrate impurities within snow are photolysed to produce nitrogen oxides (NO and NO_2) (Honrath et al., 1999; Jones et al, 2000; Dibb et al., 2002; Beine et al., 2002a) that are released to the overlying boundary layer (Jones et al., 2001; Honrath et al., 2002; Beine et al., 2002b). Although this occurs across the Antarctic snowpack, the effects are particularly noticeable at South Pole because of the characteristically shallow boundary layer at this site. Emissions from snow are concentrated within a confined layer of the atmosphere, which accentuates the resultant chemistry (Davis et al., 2004). NO_2 released from snow, therefore, becomes a significant source of local ozone, accounting for the surface ozone measurements described above (Crawford et al., 2001).

The influence of the snowpack on boundary layer chemistry is enormous and, crucially, encompasses fast reactive photochemistry. As well as releasing NOx, the snow is a source of hydrogen peroxide (H_2O_2), formaldehyde (HCHO) and nitrous acid (HONO). These can be direct sources of OH, a highly reactive radical that reacts with numerous other trace gases thus driving tropospheric chemistry. Furthermore, enhanced concentrations of NO will generate OH through the reaction $NO + HO_2 \rightarrow NO_2 + OH$. As a result of snowpack emissions, the boundary layer above the snow-covered Antarctic is far from being quiescent, but contains a fast and reactive photochemical system.

Recent measurements at Halley station have shown that halogens are also major players in fast reactive photochemistry in the coastal boundary layer. As well as high concentrations of BrO measured during the spring, the seasonal cycle of iodine monoxide (IO) shows an equally high springtime peak, as well as significant concentrations during the summertime (Saiz-Lopez et al., 2007). Indeed, modelling studies based around field measurements have shown that at Halley, although snow photochemistry is active, it is the halogens that control the cycling of reactive radicals and hence the chemical pathways (e.g. Bloss et al., 2007). The origin of IO is not absolutely known, but the proposed source is from diatoms, marine phytoplankton that colonise the underside of sea ice.

The chemistry of the Antarctic troposphere is now known to be extremely complex and unusual. The sunlit months are characterised by fast photochemical systems with chemical origins associated with snow and sea ice. It is precisely this link, between the atmosphere and the cryosphere that makes present day tropospheric chemistry systems vulnerable to a changing climate.

4.5.3 Aerosol, clouds and radiation

Aerosol particles are known to cause significant effects on the radiative budget of the Earth, both directly, through scattering and absorption of shortwave and longwave radiation, and indirectly by acting as condensation nuclei. In polar regions, where the surface albedo can exceed 0.85 in areas covered by snow and ice, aerosols may produce appreciable warming at the surface due to multiple reflection if highly absorbing particles are suspended above these bright surfaces. It is important to determine the radiative properties of particles within the entire atmosphere column to evaluate accurately the radiative forcing. Large sets of radiometer (actinometer, pyrheliometer and sunphotometer) measurements have been carried out over the past 30 years at different Antarctic sites and examined to estimate ensemble average and long-term trends of background aerosol optical depth (AOD at 500 nm; Tomasi et al., 2007) (Figure 4.20). No significant trend was observed, except some large variation during the periods affected by the volcanic eruptions of El Chichon (1982), Pinatubo (1991) and Cerro Hudson (1971). To address topics related to the radiative forcing by polar aerosols in particular, a programme referred to as POLAR-AOD was conducted as a major project of the IPY 2007-2008.

Since the typical example of highly absorbing particles is black carbon (BC), behaviours of BC have been examined at several Antarctic stations. From the wintering intensive observations of aerosols at Syowa Station during 2004 and 2006, unique seasonal variation was obtained with a winter maximum, in contrast to the summer maximum obtained from observations at similar coastal stations as Halley or Neumayer (Wolff and Cachier, 1998), and in parallel to the results at Amsterdam Island (Pereirra et al., 2006). Also an interesting pathway of BC to Syowa was found in summer months, a high peak of BC concentration occurred with local katabatic wind maximum following trajectories coming from the continental side. As there would be no source of BC inland, those trajectories must have originated from South America or Africa in response to biomass burning, for example. These pathways were confirmed in the airborne campaign (ANTSYO-II) jointly conducted by AWI, Germany and NIPR, Japan with the AWI aircraft Polar 2, around Neumayer and Syowa stations in the 2006/07 season.

Clouds are an extremely important part of the Antarctic climate system and have a significant influence on the surface radiation balance and hence also on biological processes. Yet there are large uncertainties in quantifying their role, and there is little information on how cloud amount and cloud properties have changed in recent decades. Although the general distribution of clouds is well-known - with large amounts of cloud cover over sea ice or open water in the Southern Ocean, and low cloud amounts over the continent (King and Turner, 1997), we still have no reliable cloud climatology due to deficiencies in deriving cloud amounts from passive satellite sensor data over snow and ice surfaces (Yamanouchi and Kawaguchi, 1992; Kato et al., 2006). Determining trends in cloud amount from the station data is difficult because of jumps when the observers change. However, with careful assessment of the observations some trends have been estimated. The 50 years of observations from Syowa Station revealed a 10% increase in cloud cover over that period (Yamanouchi and Shudo, 2007), although no trend was detected in the observations from South Pole (Town et al., 2007).

The radiation budget is the key aspect of studies of Antarctic climate, and the general features of the surface radiation budget at the stations on the continent have been described (Liljequist, 1956; Kuhn et al., 1977; Yamanouchi, 1983; Yamanouchi and Kawaguchi, 1984). High precision surface radiation budget measurements continue at Neumayer and Syowa stations as part of the global Baseline Surface Radiation Network (BSRN; Ohmura et al., 1998). From the global measurements a decline in solar radiation over the land surface was apparent up to 1990 - a phenomenon known as “global dimming”. Widespread “brightening” has been observed since the late 1980s globally and in the Antarctic (Wild et al., 2003), although there is no consensus yet on the Antarctic measurements (e.g. see Yamanouchi and Shudo, 2007). As for the top-of-atmosphere radiation budget, following the pioneer work by Raschke et al. (1973) using satellite data in the early stage, several measurements have been conducted as part of the Earth Radiation Budget Experiment (ERBE) and Clouds and the

Earth's Radiant Energy System (CERES). As yet there is no clear conclusion about radiation trends at that level in the Antarctic (Yamanouchi and Charlock, 1997; Kato et al., 2006).

4.5.3.1 Spatial variations in atmospheric chemistry suggested from snow cores

Ice core chemistry includes a broad range of measurements such as: major soluble ions, trace elements, radionuclides, and organic acids. The following is a synthesis of some of the major findings related to the understanding of the chemistry of the atmosphere over Antarctica as understood from the examination of the chemistry of ice cores.

An updated compilation of published and new data of major ion (Ca, Cl, K, Mg, Na, NO_3, SO_4) and methylsulfonate (MS) concentrations in snow from 520 Antarctic sites is provided by the national ITASE programmes of Australia, Brazil, China, Germany, Italy, Japan, Korea, New Zealand, Norway, United Kingdom, United States of America, and the national Antarctic programme of Finland (Bertler et al., 2005). The comparison shows that snow chemistry concentrations vary by up to four orders of magnitude across Antarctica and exhibit distinct geographical patterns (see example Na in Figure 4.21). This Antarctic-wide comparison provides a unique opportunity to improve our understanding of the fundamental factors that ultimately control the chemistry of a snow or ice sample.

As expected, the East Antarctic interior shows significantly lower values (~2 ppb to ~30 ppb) than the coastal sites (~75 ppb to 14,680 ppb). However, high values have also been reported from Marie Byrd Land at high elevation, and low concentrations in the vicinity of the East Antarctic coastlines (Kaiser Wilhelm Land and Terra Adélie). Furthermore, the change from very low to very high concentrations seems to occur within a narrow band in the vicinity of the coast. While high Na deposition is readily explained in coastal areas due to high sea salt input, the narrow zone of marine air mass intrusions (mesoscale cyclonic activity) coincides with the rapid decrease of Na concentrations in the Antarctic interior. Here the katabatic wind streams, transporting Na-depleted air masses from the interior towards the coast, compete with the Na-rich coastal air masses. In contrast, the Antarctic Peninsula shows overall high values and no trends, caused by strong sea salt input all year round and a secondary non-sea-salt contribution from ice-free mountain peaks. Most of the data points located on the Antarctic Peninsula are surface samples representing winter snow. As Na peaks in most regions of Antarctica during winter, the higher Na concentrations reported from the Antarctic Peninsula are partially caused by this bias. This and similar maps for Ca, Cl, K, Mg, NO_3, SO_4) and MS are available in Bertler et al. (2005).

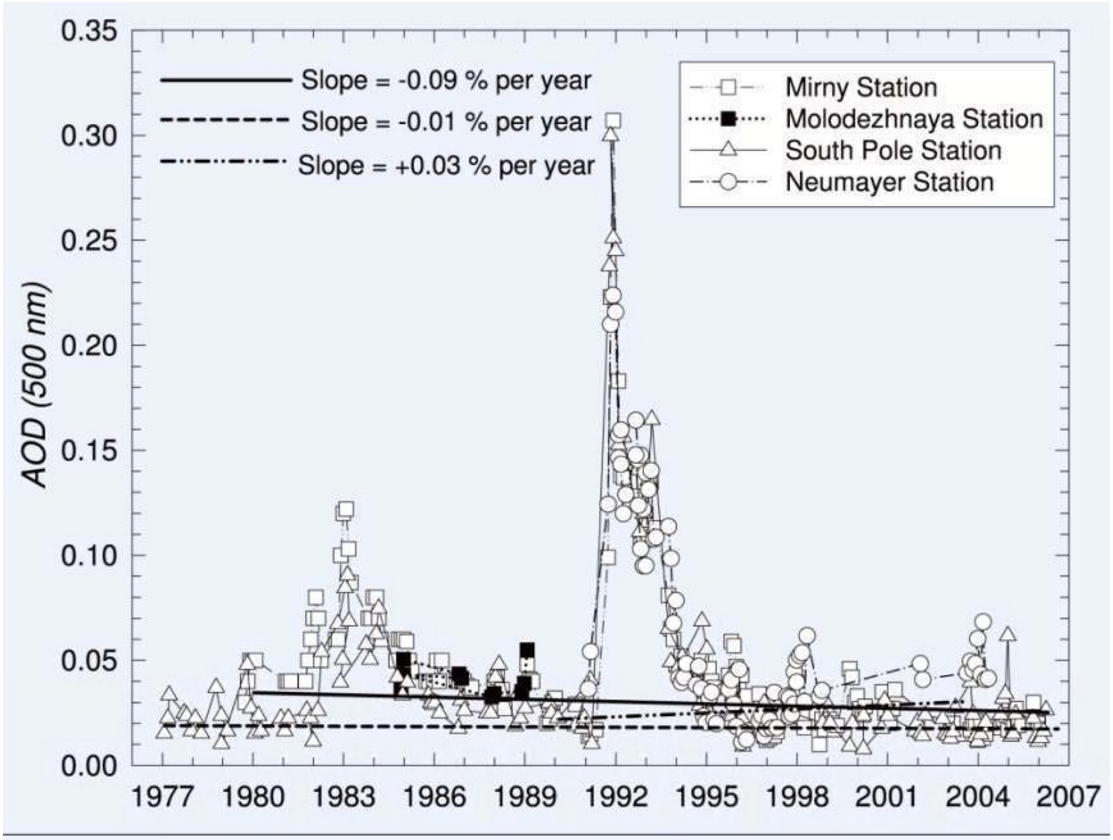


Figure 4.20 Time sequence of the monthly mean values of AOD (500 nm) derived from (1) filtered actinometer and sun photometer at Mirny from 1979/80 to 2005/06 (open squares), and at Molodezhnaya from 1985 to 1989 (solid squares); (2) filtered pyrheliometer and sun photometer at South Pole from 1977 to 2006 (open triangles); and from sun photometer measurements at Neumayer from 1991 to 2004 (open circles). The regression lines, defined separately for the Mirny, South Pole and Neumayer data sets without volcanic data are drawn to show the long-term trend of background aerosol extinction in Antarctica. (Figure 14 of Tomasi et al., 2007)

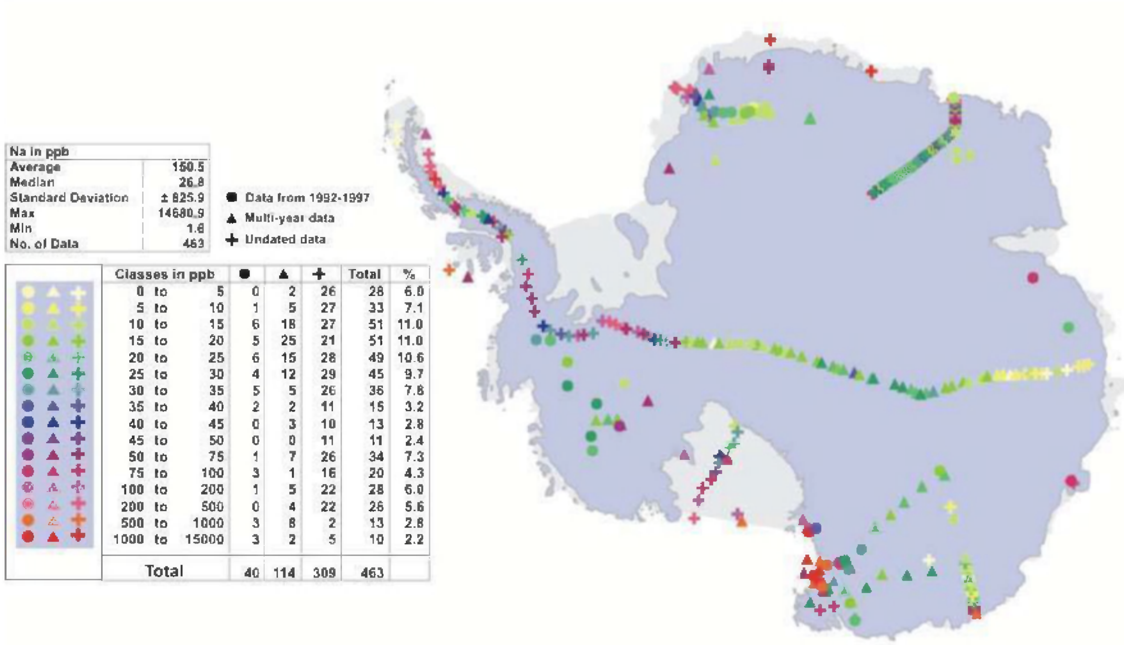


Figure 4.21 Spatial variability of Na^+ concentration measured in ppb. Solid circles represent data from 1992-1997. Solid triangles represent all other multi-year data. Crosses represent non-annual or undated samples. Spatial variability of Na^+ concentration ranges from 2 ppb to 14,680 ppb.

4.6 The Southern Ocean

4.6.1 Introduction

At first sight the Southern Ocean seems to be dominated by circumpolar symmetry mainly assured by the circumpolar current bands of the eastward flowing ACC covering mainly the mid latitudes (Figure 1.8) and the westward flowing Antarctic Coastal Current surrounding the continental margin in high latitudes. Superimposed on the zonal flow there are meridional circulation cells (Figure 1.9), the deep one related to sinking motion along the continental slope to form the Antarctic Bottom Water, and the shallow one forming Mode and Intermediate Waters in the mid latitudes. In spite of the current system linking the Atlantic, Indian and Pacific sectors, significant zonal differences are obvious and require a regional description of the variability and change in the Southern Ocean. Differences are related to the shape of the ocean basins and ridges which strongly affect the ocean currents, the shape of the continent with huge embayments, such as the Weddell and the Ross Seas giving rise to subpolar gyres (Figure 1.10), the distributions of ice shelves, the zonal structure of the forcing (storm tracks) and finally the different hydrographic conditions in the northward adjacent ocean basins.

The meridional circulation cells which can only schematically be represented in a two dimensional way (Figure 1.9), display clear zonal differences as well. They are obvious in the long term mean conditions, e.g. the horizontal distribution of the newly formed bottom water (Figure 4.22, Orsi et al., 1999) which result in the variable contributions of different sectors to the bottom water formation. It is estimated that 66% can be contributed to originate in the Weddell Sea, 25% in the Australian Sector and 7% in the Ross Sea (Rintoul, 1998). Due to the zonal differences in the hydrographic and forcing conditions, variations differ in the sectors as well, e.g. cooling and freshening of the newly formed bottom water in the

Australian sector (Rintoul, 2007) and warming and increase of salinity of bottom water on the prime meridian (Fahrbach et al., 2004). However, even within the basins different patterns are detected e.g. in the Weddell Sea (Fahrbach et al., 2004). To take into account the circumpolar differences in the observed variations the description of the Southern Ocean is arranged here zonally in different sectors. Since regional differences occur even on smaller scales than the three large ocean sectors, we focus on sub-areas where a particular type of variation is observed.

There is evidence that changes in the Antarctic Bottom Water propagate into the global ocean. Warming of the northward flow of Antarctic Bottom Water on a decadal time scale is observed in Vema Channel from the Argentine to the Brazil Basin (Zenk and Morozov, 2007). Large scale abyssal warming over decades is observed in the South Atlantic (Johnson and Doney, 2006) and the Pacific (Johnson et al., 2007).

Observation of change in surface waters is difficult to detect since there is an intensive seasonal cycle and only few observations. However, around South Georgia observations exist since 1925, which are frequent enough to resolve the annual cycle. They reveal significant trends in the upper 150 m, with 2.3°C of warming over 81 years, which are twice as strong in winter than in summer (Whitehouse et al., 2008).

Sub-Antarctic Mode Water is formed by winter cooling and convection just north of the Sub-Antarctic Front (McCartney, 1997, 1982). When it is subducted it becomes Antarctic Intermediate Water (Hanawa and Talley, 2001). This water mass is of particular interest due to its capacity to take up atmospheric CO_2 (Sabine et al., 2004). Significant warming (Wong, 2001; Gille, 2002; Aoki et al., 2003) and freshening was observed in these water masses (Wong, 1999; Curry et al, 2003; Bindoff et al., 2007; Böning et al., 2008) over the last decades. To understand the changes, the formation processes (e.g. Sallée et al., 2006) and the turbulent character of the field (Tomczak, 2007) need to be understood.

In a circumpolar view the ACC band has warmed in recent decades (Figure 4.23; Gille, 2002 and 2008; Levitus et al., 2000, 2005, Böning et al., 2008). The changes are consistent with a southward shift of the ACC (Gille, 2008). Some climate models suggest that the ACC shifts south in response to a southward shift of the westerly winds driven by enhanced greenhouse forcing (Fyfe and Saenko, 2006; Bi et al., 2002). A significant part of the changes in the wind system can be related to the positive trend in the SAM (e.g. Hughes et al., 2003). The poleward shift and intensification of winds over the Southern Ocean has been attributed to both changes in ozone in the Antarctic stratosphere (Thompson and Solomon, 2002) and to greenhouse warming (Fyfe et al., 1999). In addition to driving changes in the ACC, the wind changes have caused a southward expansion of the subtropical gyres (Cai, 2006) and an intensification of the Southern Hemisphere “supergyre” that links the three subtropical gyres (Speich et al., 2002, 2007). The “supergyre” provides the mechanism by which Sub-Antarctic Mode Water and Antarctic Intermediate Water is distributed between the ocean basins (Ridgway and Dunn, 2007). In the area of the formation of Antarctic Intermediate Water north of the Subantarctic Front freshening is observed. Surprisingly the changes are rather similar all along the ACC (Figure 4.23; Böning et al., 2008).

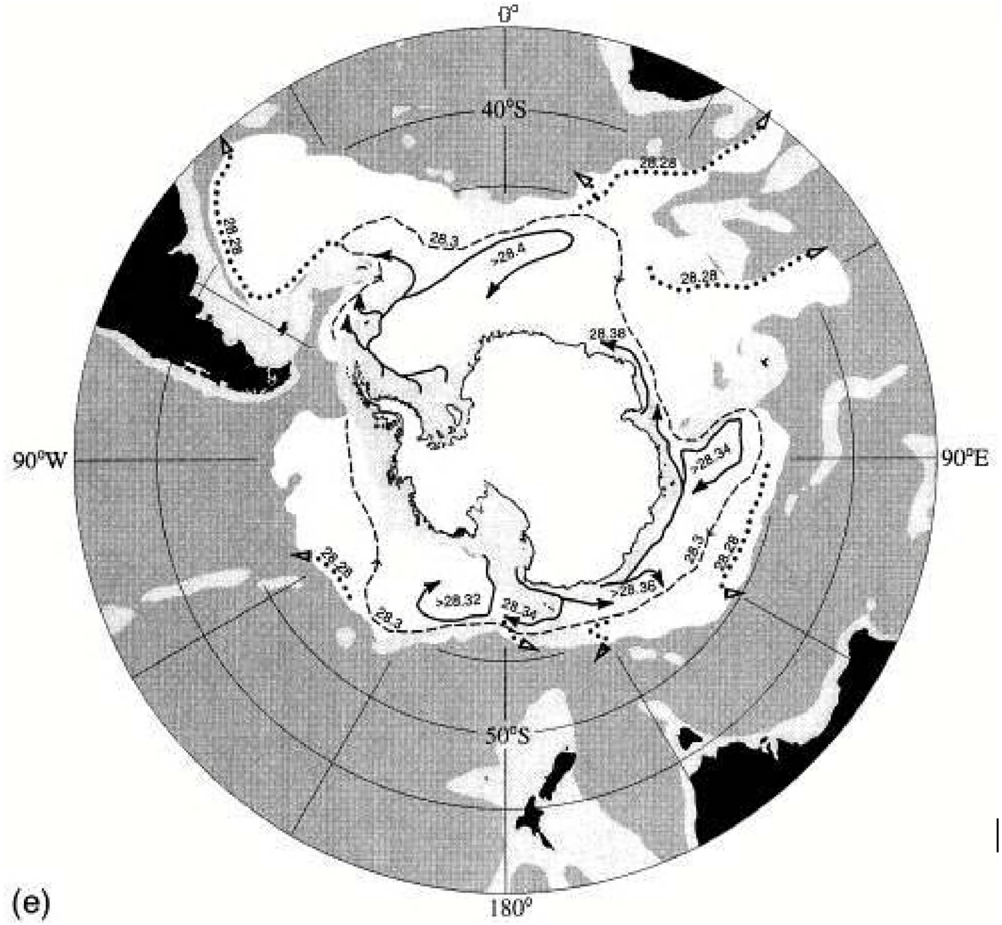


Figure 4.22 Schematic presentation of the formation areas and pathways of Antarctic Bottom Water as evidenced by water mass properties, in particular the concentration of the anthropogenic tracer CFC. It is injected into the ocean at the surface and indicates, by concentration maxima, recently ventilated water sinking to the sea bottom and spreading there from the formation areas for different densities here displayed as neutral density where the most dense water with neutral density > 28.4 kg/m^3 is found in the Weddell Sea (Orsi et al., 1999).

4.6.2 Australian Sector

Knowledge of the circulation in the Australian sector of the Southern Ocean has increased significantly in the last decade. Repeat hydrographic sections (Figure 4.24), moorings and satellite altimeter measurements have provided new insights into the structure, variability and dynamics of the ACC, water mass formation and the overturning circulation. The mean baroclinic transport of the ACC south of Australia is 147±10 Sv (Rintoul and Sokolov, 2001), consistent with recent estimates of the flow leaving the Pacific basin through Drake Passage (136±8 Sv, Cunningham et al., 2003) and the Indonesian Throughflow (Meyers et al., 1995). A multi-year time series derived from XBT sections and altimetry shows significant interannual variability (with a standard deviation of 4.3 Sv) but no trend in transport (Rintoul et al., 2002). High resolution hydrographic sections reveal that the ACC fronts consist of multiple jets, aligned with streamlines that can be traced using maps of absolute sea surface height (Sokolov and Rintoul, 2002, 2007). Eddy fluxes estimated from current meter moorings confirm that the eddies transport heat poleward and zonal momentum downward (Phillips and Rintoul, 2000). A cyclonic gyre lies between the ACC and the Antarctic continent, closed in the west by a northward boundary current along the edge of the Kerguelen Plateau (McCartney and Donohue, 2007).

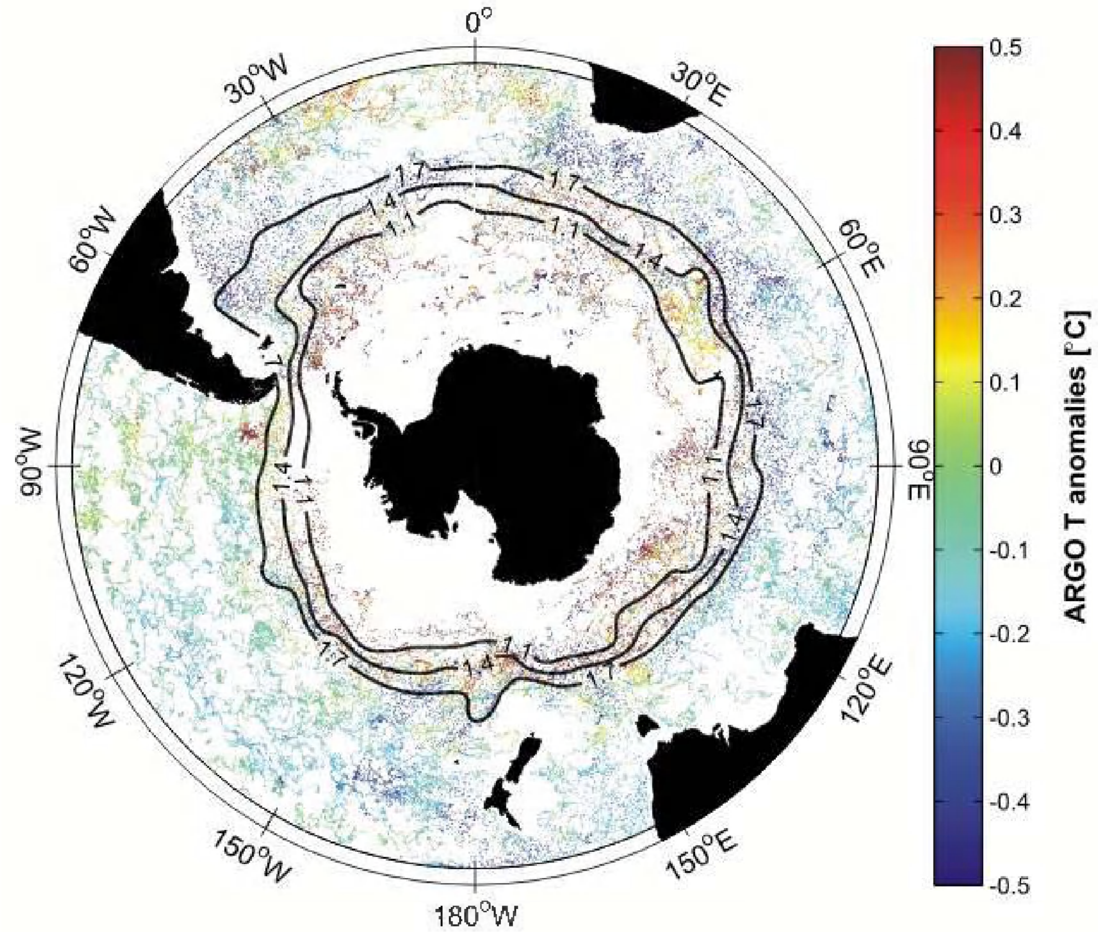


Figure 4.23 Temperature differences between measurements from Argo floats deployed after 2001, presented as dots, and the climatological mean from CARS in the density layers of the ACC show an increase of potential temperature of the water masses on the souhward flank of the ACC and a freshening on the northward side. On surfaces of constant densities freshing becomes visible as cooling, since the density loss by less salt is compensated by a density gain due to colder temperature. The major fronts representing ACC branches are displayed as dynamic height contours where the 1.4-m-dynamic-height contour is related to the Polar Front and 1.7-m one to the Subantarctic Front (Böning et al., 2008)

The ACC belt in the Australian sector has warmed in recent decades, as found elsewhere in the Southern Ocean (Gille, 2002, 2008; Levitus et al., 2000, 2005; Willis et al., 2004; Böning et al., 2008). The southward shift of the ACC fronts has caused warming through much of the water column, resulting in a strong increase in sea level south of Australia between 1992 and 2005 (Sokolov and Rintoul, 2003; Morrow et al., 2008). However, there is no observational evidence of the increase in ACC transport also predicted by the models (Böning et al., 2008). Recent studies suggest the ACC transport is insensitive to wind changes because the ACC is in an "eddy-saturated" state, in which an increase in wind forcing causes an increase in eddy activity rather than a change in transport of the current (Hallberg and Ganandesikan, 2006; Meredith and Hogg, 2006).

Changes have been observed in several water masses in the Australian sector between the 1960s and the present (Aoki et al., 2005). Waters north of the ACC have cooled and freshened on density surfaces corresponding to intermediate waters (neutral densities between 26.8 and 27.2 kg m^{-3}). South of the ACC, waters have warmed and become higher in salinity and lower in oxygen on neutral density surfaces between 27.7 and 28.0 kg m^{-3} (the Upper Circumpolar Deep Water –UCDW). The changes south of the ACC are consistent with a shoaling of the interface between the warm, salty, low oxygen UCDW and the cold, fresh, high oxygen surface water that overlies it. The pattern of water mass change observed in the Australian sector is consistent with the "fingerprint" of anthropogenic climate change in a coupled climate model (Banks and Bindoff, 2003).

The Antarctic Bottom Water (AABW) in the Australian Antarctic basin has freshened significantly since the early 1970s. Whitworth (2002) detected a shift toward fresher AABW after 1993, concluding that "two modes" of AABW were present in the basin, with the fresher mode becoming more prominent in the 1990s. More recent studies have documented a monotonic trend toward fresher, and in most cases warmer, bottom water between the late 1960s and the present rather than a bi-modal distribution. Aoki et al. (2005) used a hydrographic time series with nearly annual resolution between 1993 and 2002 to show a steady decline in salinity of the bottom water at 140ºE. By using repeat observations from the same location and same season, they could demonstrate that the trend was not the result of aliasing of spatial or temporal variability. Rintoul (2007) showed that the deep potential temperature – salinity relationship of the entire basin had shifted towards lower salinity between the early 1970s and 2005. The average rate of freshening at 115°E is 7 ppm/decade, which can be compared to a mean freshening rate of 12 ppm/decade in the North Atlantic, at similar distances downstream of the source of dense water (Dickson et al., 2002). These results suggest that the sources of dense water in both hemispheres have been responding to changes in high latitude climate.

The abyssal layers of the Australian Antarctic Basin are supplied by the two primary sources of bottom water that lie outside of the Weddell Sea: a fresh variety formed along the Adélie Land coast (144° E) and a salty variety produced in the Ross Sea (Rintoul, 1998). The changes observed in the Australian Antarctic Basin therefore reflect freshening of the AABW formed in the Indian and Pacific sectors of the Southern Ocean, which accounts for about 40% of the total production of AABW (Orsi et al., 1999).

The cause of the freshening of AABW in the Australian sector is not yet fully understood. Changes in precipitation, sea ice formation and melt, ocean circulation patterns, and melt of floating glacial ice around the Antarctic margin could all influence the salinity where dense water is formed. Oxygen isotope measurements in the Ross Sea indicate that an increase in the supply of glacial melt-water has contributed to the large freshening of shelf waters observed there in recent decades (Jacobs et al., 2002). The most likely source of the increased supply of melt-water is the rapidly thinning glaciers and ice shelves of the Amundsen Sea, including the Pine Island Glacier (Jacobs et al., 2002), where enhanced basal melt has been linked to warmer ocean temperatures (Rignot and Jacobs, 2002; Shepherd et al., 2004). While most of the ice sheet in East Antarctica appears to be gaining mass, glaciers draining parts of the Wilkes Land coast where the Adélie Land bottom water is formed have decreased in elevation (Shepherd and Wingham, 2007), and the floating ice in this sector thinned between 1992 and 2002 (Zwally et al., 2005). Therefore increased supply of glacial melt-water may have played a role in the freshening of both the Adélie Land and Ross Sea Bottom Water.

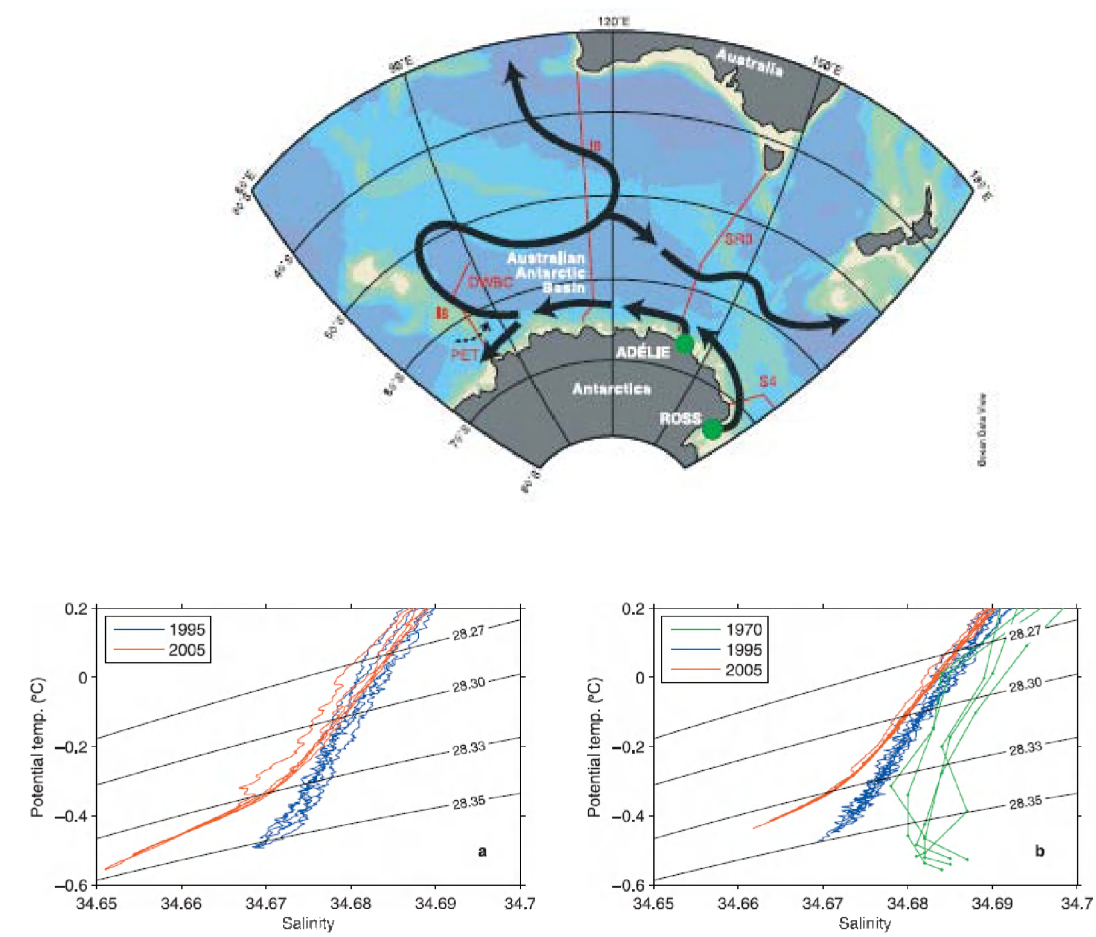


Figure 4.24 Top) Map indicating location of repeat sections along which changes in bottom water properties have been assessed. The Australian Antarctic Basin is supplied by two sources of Antarctic Bottom Water: the Ross Sea and Adélie Land. Lower) Changes in the deep potential temperature – salinity curves along 115°E, over the continental rise (61-63.3°S, left) and further offshore (56.5 - 61°S, right). From Rintoul (2007).

4.6.3 The Amundsen/Bellingshausen Seas

The southeast Pacific Ocean (70°W – 150°W) deserves enhanced scientific interest because of the significant alterations it faces or is supposed to face in a changing climate. Since 1951 annual mean atmospheric temperatures rose by almost 3°C at the Antarctic Peninsula (King, 1994; Turner et al., 2005a) which can be linked to changes in the Bellingshausen Sea, like sea ice retreat (Jacobs and Comiso, 1993), increased ocean surface summer temperatures of more than 1°C, enhanced upper-layer salinification (Meredith and King, 2005), the disintegration of smaller ice shelves (Doake and Vaughan 1991), and accelerated retreat of glaciers (Cook et al., 2005). The changes can be related to atmospheric variability including the Antarctic Circumpolar Wave (White and Peterson, 1996) or the SAM (Hall and Visbeck, 2002; Lefebvre et al., 2004), which both exhibit extreme values in the southeast Pacific Ocean. Different hydrographic conditions have a severe impact on marine species (e.g., the Antarctic krill) which use the Bellingshausen Sea for breeding and nursery before the larvae mainly drift eastward to the southern Scotia Sea/northwestern Weddell Sea (Siegel, 2005). A comprehensive field study on Antarctic krill in the Amundsen Sea has yet to be conducted.

Connected via the westward flowing, and in this part, weak coastal current, changes in the Bellingshausen Sea also influence the Amundsen Sea (100°W – 150°W) which is fringed to the south by the outlets of major ice streams draining the West Antarctic Ice Sheet. A possible collapse of the latter would result in a 5-6 m global sea level rise threatening many low-lying coastal areas around the globe including millions of their residents (Rowley et al., 2007), but is not likely in the next 100 years. The southernmost position of the ACC southern front (Orsi et al., 1995) together with a relative narrow continental shelf crisscrossed by numerous channels (Figure 4.25) allows Upper Circumpolar Deep Water (UCDW) with temperatures near 1°C to reach the ice shelf edges in the Amundsen and Bellingshausen Seas (Figure 4.26). This ocean heat could fuel melting of up to tens of metres per year at deep ice shelf bases. A linear relationship between melt rates beneath Antarctic ice shelves and ocean temperature is roughly linear at 1 m/yr per 0.1°C ocean warming derived from observations of 23 glaciers (Rignot and Jacobs, 2002). The change of ice melt rate to a change in ocean temperature may not follow this same relationship. In a simple box model Olbers and Hellmer (2009) confirm this rate to be consistent with the involved physical processes and investigate the sensitivity. Numerical modeling of ice-ocean interaction beneath Pine Island Glacier (Payne et al., 2007) concludes that the observed thinning at a rate of 3.9 $\pm$ 0.5 m/yr between 1992 and 2001 (Shephard et al., 2002) would correspond to a different rate of ~ 0.25°C warming of the UCDW underneath Pine Island Glacier. Such warming has not been observed on the Amundsen Sea continental shelf, but a 40-year long temperature time series from the nearby Ross Sea exhibits a warming of the off-shore temperature maximum (190-440 m depth) of ~ 0.3°C (Jacobs et al., 2002). Temperature variability correlated with changing freshwater fluxes due to basal melting is likely at the fringe of the West Antarctic Ice Sheet. Upper Southern Ocean temperatures increased since the 1950's (Gille, 2002; Böning et al., 2008), but the few oceanographic snapshots (e.g., Hofmann and Klinck, 1998; Walker et al., 2007) from the Amundsen/Bellingshausen Sea continental margin are insufficient to identify the time scales and strengths of variability or any trends.

4.6.4 Variability and change in Ross Sea shelf waters

The circulation in the Ross Sea is dominated by a wind-driven cyclonic gyre (Treshnikov, 1964) visible as a depression in the steric height transporting Circumpolar Deep Water to the south where by interaction with shelf and slope water Antarctic Bottom Water is produced. It is located south of the mid-ocean ridge between 170° E and 140°W (e.g. Gouretski, 1999) with its centre at about 68°S, 164°W shifting to the southeast with depth. The baroclinic transport of 8.5 Sv is significantly smaller than the one of the Weddell gyre. The eastern boundary is given by a southward deflection of the ACC due to the bottom topography. At the southern limb, westward flow transports water as warm as 1.6°C. From property maps Reid (1986) included the extension up to the Antarctic Peninsula. Antipov et al. (1987), Maslennikov (1987) and Locarini (1994) locate the eastern boundary at 140°W. The continental shelf area is relatively well sampled due to the normally weak ice cover in summer and the presence of several Antarctic stations (Jacobs and Giulivi, 1999).

Shelf waters' include a variety of low-temperature, ice-modified, high- and low-salinity water masses found below the ocean surface layers on the Antarctic continental shelf (Whitworth et al., 1998). Summer salinity profiles spanning about 20 years in High Salinity Shelf Water (HSSW) near Ross Island displayed gradual salinity increases below 200 m and interannual water column shifts that were several times the measurement accuracy and half the annual cycle in McMurdo Sound (Jacobs, 1985). Atmospheric forcing, sea ice production, HSSW residence time, ice shelf melting and intrusion of Modified Circumpolar Deep Water onto the continental shelf were considered as possible agents of change. Hellmer and Jacobs (1994) noted that salinity decreases could also result from fresher upstream source waters.

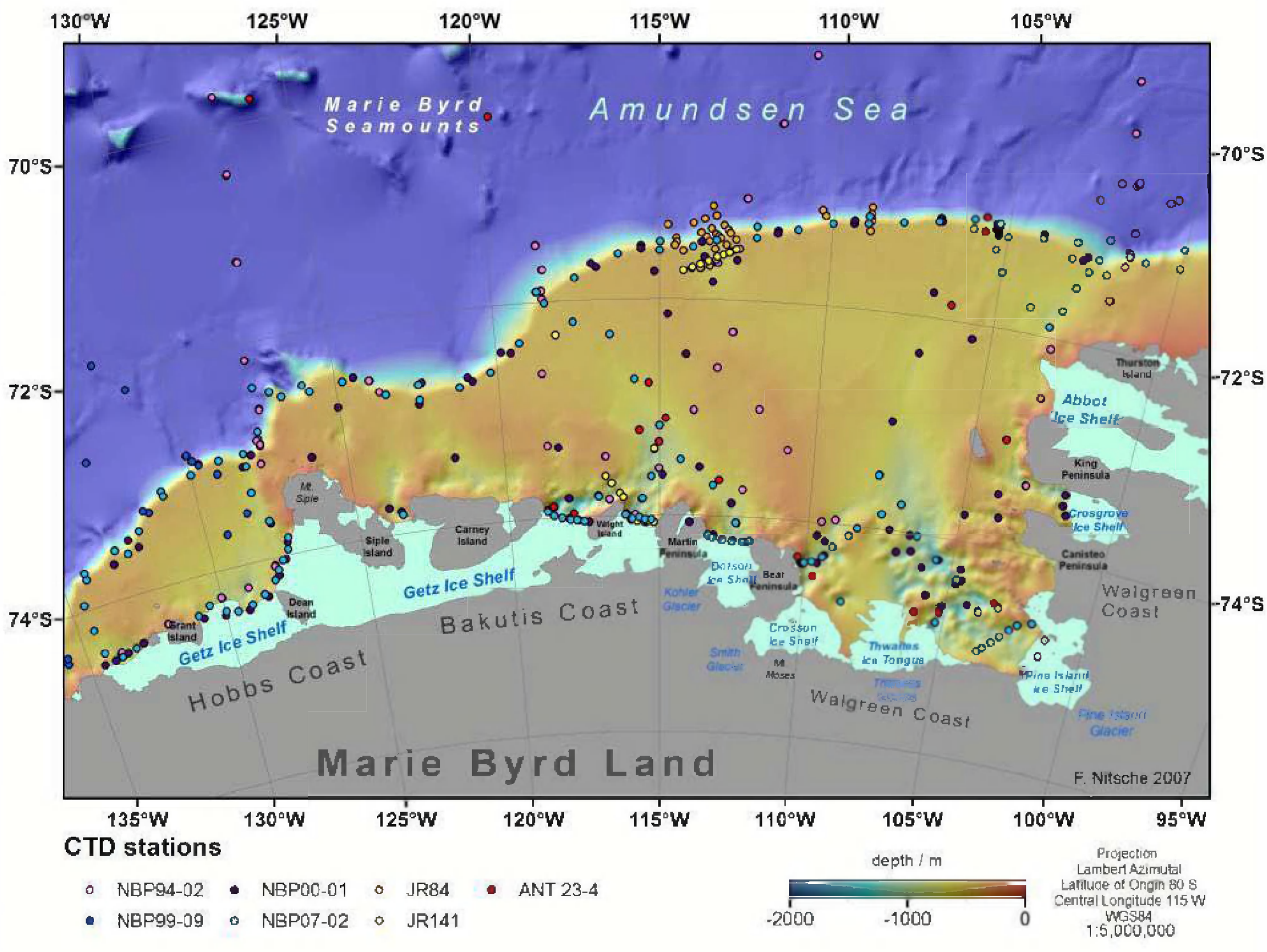


Figure 4.25 Bathymetric chart of the Amundsen Sea continental shelf and adjacent deep ocean spotted with the distribution of hydrographic stations of different cruises (colour coded). Fringing ice shelves and glaciers draining the West Antarctic Ice Sheet in light blue (Nitsche et al., 2007). The CTD station identifiers refer to the Nathaniel B. Palmer (NBP), the James Clark Ross (JCR) and the Polarstern (ANT).

Attempts to link observed HSSW changes to multiyear variability in regional sea ice extent, winds and air temperatures revealed the need for longer time series. The data base largely consists of sporadic summer measurements, and both modelers and observers have noted the possibility of aliasing in this shelf water record due to undersampling of a variable inflow. In addition, most measurements were in or near an HSSW eddy, another potential source of variability. Nonetheless, the deep HSSW trend in that area has closely tracked changes at depth along the Ross Ice Shelf and near 500 m throughout the western Ross Sea (Jacobs and Giulivi, 1998; Smethie and Jacobs, 2005).

Record low salinities at the site in February 2000 led to analyses that more strongly implicated changes in ice-ocean interactions upstream in the Amundsen and Bellingshausen Seas (Jacobs et al., 2002). Assmann and Timmermann (2005) successfully modeled averaged HSSW salinity profiles, and inferred that the freshening resulted from a Bellingshausen Sea thermal anomaly. Their periodic signal upwelled in the Amundsen Sea, reduced brine drainage near the sea ice edge and induced a subsurface salinity decrease that was advected into the Ross Sea. Interannual salinity variability is high, but the overall trend has been statistically significant and qualitatively consistent with freshening over a much wider area (Jacobs, 2006).

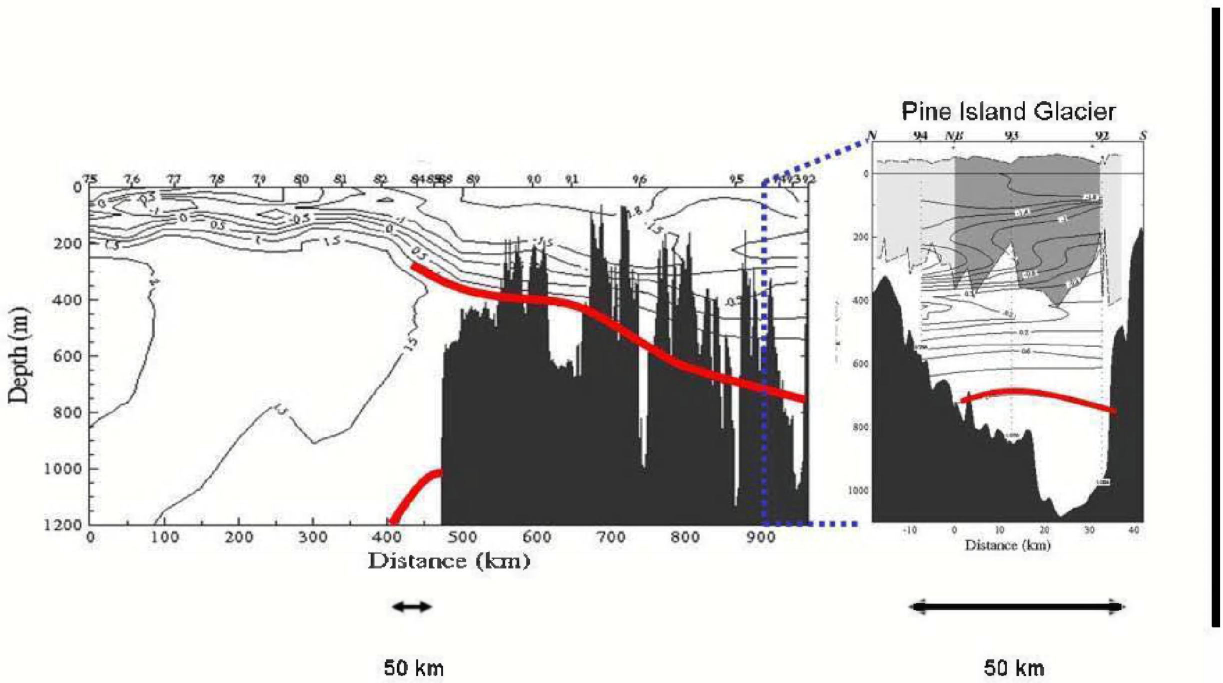


Figure 4.26 A cross-section showing penetration of warm water to the sub-ice shelf cavity in Pine Island Bay. Potential temperature of the upper 1,200 m along a band between 100°W and 105°W projected on a strait transect from the open ocean (left) to Pine Island Bay (right) measured during NBP9402 (see Figure 4.25 for station locations). Due to the ice conditions the stations could not be done along a straight line. The sea floor depth was extracted from the ship's 3.5 kHz echosounder data. Right figure depicts the temperature field in front of Pine Island Glacier with its draft shaded in gray. This short line is along the line of pink dots (sample stations) shown in Figure 4.25. The 1°C isotherm on the continental shelf and slope is marked in red (modified from Hellmer et al., 1998). The solid black area indicates the sea bed.

The record of summer shelf water thermohaline properties has recently been extended to 50 years (Figure 4.27), and the study area widened to include profiles in McMurdo Sound. The 50 year salinity trend continues to be near -0.03/decade, while slightly warming temperatures have remained consistent with HSSW formation by surface freezing in winter. HSSW near Ross Island thus serves as an index site to monitor change occurring in the Ross Sea and upstream (eastward) in the Antarctic Coastal Current. The salinity decline appears to derive mainly from increasing continental ice meltwater, and will subsequently change the properties if not the production rates of deep and bottom waters. Over regional areas the lower salinity has raised sea level via the halosteric component of seawater density.

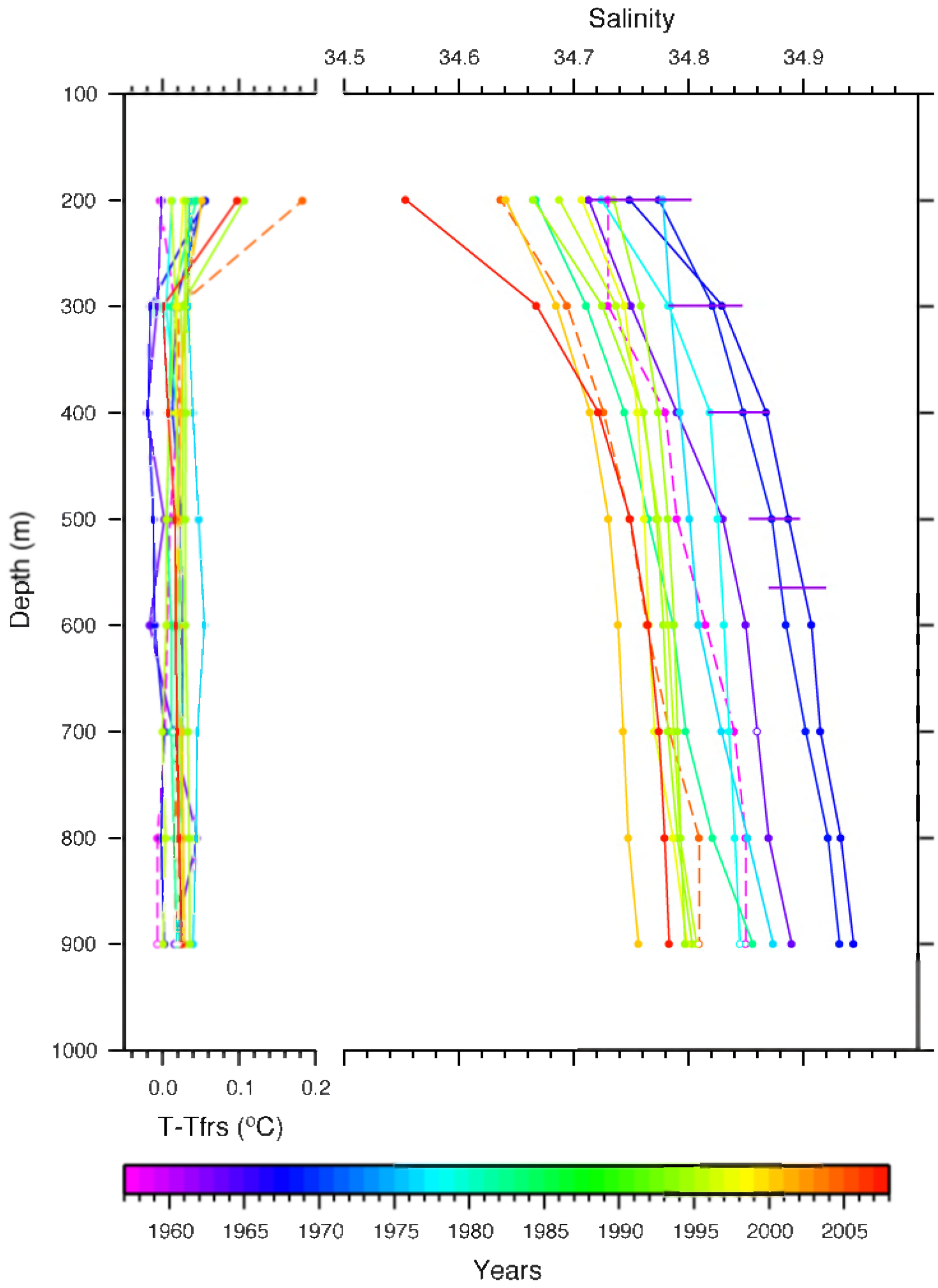


Figure 4.27 Summer temperature and salinity profiles over a 50 year period, some averaged from adjacent profiles, and dashed if more than 15 km from 168° 20'E, 77° 10'S, near Ross Island. Plotted values are 100 m averages/interpolations of CTD or bottle data, edited or extrapolated where shown by open circles. Horizontal lines depict the salinity ranges at six bottle casts in southern McMurdo Sound over 23 December 1960 – 9 February 1961 (Tressler and Ommundsen, 1963). Temperatures are referenced to the surface freezing point, ~-1.91°C at a salinity of 34.8. Tfrs is the surface freezing reference temperature.

4.6.5 The Weddell Sea Sector

The Weddell Sea hosts a subpolar gyre (e.g. Treshnikov, 1964), the Weddell Gyre (Figure 1.10), that brings relatively warm, salty circumpolar water (Warm Deep Water, WDW) south towards the Antarctic continent, and transports colder, fresher waters northward (Weddell Sea Deep and Bottom Waters, WSDW and WSBW, as well as surface waters). The transformation of this source water mass is one of the major climate-relevant processes in the Southern Hemisphere, affecting and involving ocean, atmosphere and cryosphere and a major contribution to the deep meridional overturning cell (Figure 1.9).

In respect to climate change, the most relevant property is the change in the intensity of the overturning circulation, i.e. the transport of shallow source water masses into the formation areas and the newly formed water masses leaving the Antarctic Systems into the world oceans. However, since these transport differences are relative small in comparison to the gyre scale circulation, the latter must be known to distinguish between recirculation and net transformation.

The cyclonic Weddell Gyre is bounded to the west by the Antarctic Peninsula, to the south by the Antarctic continent and to the north by a chain of roughly zonal ridges at ~60°S. The eastern boundary is less well defined but is generally agreed to extend as far east as ~30° E (Gouretski and Danilov, 1993). The flow associated with the Antarctic Slope Front, a boundary current tied to the steep topography of the Antarctic continental slope, contributes a major proportion of the gyre transport. The Weddell Gyre is primarily driven by the cyclonic wind field (Gordon et al., 1981), leading to a doming of isopycnals in the centre of the gyre. The volume transport is still an ongoing topic of research, partly because it is dominated by the barotropic component (Fahrbach et al., 1991) so it is difficult to measure. Integrating the wind field in a Sverdrup calculation revealed a volume transport of 76 Sv (Gordon et al., 1981) while the first attempts to reference the geostrophic shear to current meters yielded a transport of 97 Sv (Carmack and Foster, 1975). For a section across the Weddell Gyre from the tip of the Antarctic Peninsula to Kapp Norvegia, transports have, uniquely, been referenced to a long time series current meter array, and this has produced lower transport estimates of 20-56 Sv (Fahrbach et al., 1991; Fahrbach et al. 1994). At the Greenwich meridian, larger transports of the order of 60 Sv were obtained from referencing shear to shipboard ADCP (Schröder and Fahrbach, 1999). Current meter arrays subsequently yielded 45 – 56 Sv (Klatt et al., 2005). The larger values at the Greenwich meridian may be due to recirculations within the central gyre not being captured by the Kapp Norgvegia section. However, a recent high resolution section at the Antarctic Peninsula also yielded ~46 Sv (Thompson and Heywood, 2008), and it is suggested that referencing the geostrophic shear across the steep Antarctic continental slope to a relatively widely spaced current meter array may have underestimated the barotropic contribution to the total volume transport.

WSBW is formed on the Antarctic continental shelves where they are wide. In the Weddell Sea, the Filchner-Ronne Ice shelf is one such source region (e.g. Foldvik et al., 2004), and recent evidence suggests formation also on the eastern side of the Peninsula near the Larsen Ice Shelf. The water on the Antarctic continental shelves is typically fresher than its warmer, salty source to the north; it is believed to freshen by the addition of sea ice melt, glacial ice melt from the Antarctic ice sheet and floating ice shelves, and from precipitation. This freshening is a necessary precursor to the bottom water formation process, which involves salinification from brine rejection during sea ice formation, together with cooling. One contributing process to WSBW involves mixing with Ice Shelf Water, and the other process involves mixing with HSSW. The WSBW descends the continental slope and entrains ambient water as it goes (Baines and Condie, 1998). The descent of the WSBW affects the structure of a series of fronts along the rim of the Weddell Gyre, in particular the Antarctic Slope Front and the Weddell Front.

The WSBW is too dense to be able to escape the Weddell Sea. It can only escape by mixing with the water above it, becoming warmer, saltier and less dense, and forming WSDW, which is sufficiently shallow to flow out through the passages in the topography surrounding the Weddell Sea. WSDW outside the Weddell Sea has the water mass properties of Antarctic Bottom Water, of which it is believed to be the major contributor. Estimates of the proportion of Antarctic Bottom Water originating in the Weddell Sea range from 50 to

90% (Orsi et al., 1999). An inflow of water of WSDW properties from the east has been documented (Meredith et al., 2000) that may form in the region of the Prydz Bay Gyre, or may even originate in the Australian Antarctic Basin and enter the Weddell Sea through the Princess Elizabeth Trough (Heywood et al., 1999).

The export of WSDW to the world ocean is of the order of 10 +/- 4Sv (Naveira Garabato et al., 2002). This can escape through gaps in the ridges to the north and east of the Weddell Sea, and subsequently invades all ocean basins.

Measurements of water mass transports exist in relation to climate time scales only as snapshots and not over a long enough times to be able to detect changes. Therefore, one has to measure water mass properties and derive from their variations conclusions about changes of transport and formation rates. Carefully validated models play a significant role here. Even for water mass properties observations in sufficient spatial resolution and accuracy last only over one to two decades. Therefore it is still not possible to unambiguously distinguish between a trend and decadal to multidecadal variation. In spite of the fact that the variability in the deep water masses seems to be relatively small in comparison to those of the surface water masses, it is importance, because the large volume of the deep waters can store significant heat quantities or dissolved substances even if the changes in property is only minor.

Considering its remote and inhospitable location, the Weddell Sea was well observed during WOCE and subsequently through CLIVAR, with (largely summer-time) hydrographic sections across the Weddell Gyre onto the Antarctic continental shelf, and with arrays of moorings. These sections indicate a number of decadal-scale changes in water mass properties. The WDW warmed by some 0.04° C during the 1990s (Robertson et al., 2002; Fahrbach et al., 2004) and has subsequently cooled (Fahrbach et al., 2004). This was accompanied by a salinification of about 0.004 (note salinity does not have any units), just detectable over the decade. A quasi-meridional section across the Weddell Gyre occupied in 1973 and 1995 revealed a warming of the WDW in the southern limb of the gyre by 0.2° C accompanied by a small increase in salinity, whereas there was no discernible change in the northern limb of the gyre (Heywood and King, 2002). There has been debate as to whether these changes to the warm inflow to the gyre are caused by advection of warmer circumpolar waters, and/or by changes in the wind field (Fahrbach et al., 2004, 2006; Smedsrud, 2005).

During the 1970s a persistent gap in the sea ice, the Weddell Polynya, occurred for several winters. The ocean lost a great deal of heat to the atmosphere during these events. The polynya strongly affected the mode of overturning in the Weddell Sea, shifting it from shelf and slope processes to open ocean convection (Gordon, 1978) with consequences for the overturning rates. After this one event the large polynya did not show up again, but weak ice cover was observed in the Maud Rise area several times and interpreted as a sign of possible emergence of a new polynya, though no such polynya emerged. In a recent study, Gordon et al. (2007) attribute the occurrence of the Weddell Polynya to variations in the SAM.

Weddell Sea Bottom Water is observed on the western continental slope and within the basin up to the Greenwich Meridian. Whereas the bottom water was warming and getting more saline in the basin (Fahrbach et al., 2004) it became colder and fresher on the slope in the western Weddell Sea (Heywood et al, in preparation). It is a matter of debate if such regional changes in phase are caused by the time lag of 5-10 years involved when freshly formed bottom water spreads across the basin.

Because much of the Weddell Sea is covered with sea ice for much of the year, there is a lack of observations on the continental shelf and slope, especially in winter. This was a priority area during IPY and moored arrays were deployed together with hydrographic sections to fill the observational gaps. Measurements beneath the sea ice in the Weddell Sea were obtained for the first time by acoustically tracked floats and by instruments carried by marine mammals such as elephant seals.

4.6.6 Small-scale processes in the Southern Ocean

The importance of small-scale processes in surface, bottom and lateral boundary layers is well known. In ice-covered oceans the surface boundary layer is of particular interest (McPhee, 2009) since it controls the heat and momentum exchange between ice and ocean. These exchanges affect ice growth, melt and motion as well as the heat budget of the ocean and the currents (McPhee et al., 2008). Appropriate quantification of such processes is a prerequisite for successful modelling the large scale conditions.

For a long time small-scale processes related to ocean turbulence in the interior were almost neglected in the consideration of large-scale ocean conditions. This resulted from lacking appropriate observation techniques and theoretical concepts for relating small-scale to large-scale processes. Normally a simple energy cascade providing turbulent energy from larger scale ocean current shear to small-scale mixing was assumed, which could be applied in large-scale models by properly selected mixing parameters. However, improved measurements of turbulence suggest that the amount of turbulent energy is much larger in the interior than previously assumed. There, internal waves, not only generated at the continental slope, but as well over rough bottom topography, play a major role in generating intensive mixing in the interior away from the well-known strongly mixed boundary layers. The intensity of internal mixing is so high, that it has to be taken into account to quantitatively understand the meridional overturning circulation. The understanding of spatially and timely important variations in the intensity of mixing and its significant role for large-scale processes prompted research with the aim of detecting by observations whether changes in small-scale processes could affect large-scale conditions i.e. the oceans role in climate.

It has been estimated (e.g., Wunsch, 1998) that up to one-third of the energy required to drive the global ocean's overturning circulation (2-3 TW, see Wunsch and Ferrari (2004) for a review) stems from the work done by the wind on the ACC, and that the bulk of that energy is transferred to the mesoscale eddy field. Current conceptual and numerical models of the Southern Ocean overturning circulation (see Rintoul et al. (2001) and Olbers et al. (2004) for two reviews) unanimously highlight the eddies' crucial role in transporting water masses and tracers along the sloping isopycnals (surfaces of constant density) of the ACC — particularly in the upper overturning cell — as well as in transporting the momentum input by the wind to the level of topographic obstacles, such as ridges, where it may be transferred to the solid Earth.

Physical processes occurring on length scales on which earth rotation is of similar importance to ocean stratification (the baroclinic Rossby radius of deformation), which are smaller than 5-20 km south of 40°S, (see Chelton et al., 1998) play an important role in shaping and driving the circulation of the Southern Ocean. To represent ocean properties realistically measurements and models would need to resolve this scale. However this is normally not possible and the lack of resolution has to be compensated for by assumptions about processes at this scale and below.

For example, formation of the AABW filling the deepest layers of much of the global ocean abyss is crucially dependent on the convective (i.e. small scale) production of dense, high-salinity shelf waters over the Antarctic continental shelves during periods of sea ice growth (Morales Maqueda et al., 2004), as well as on the heat and freshwater exchanges between those waters and the adjacent ice shelves (e.g. Hellmer, 2004). The properties of Antarctic shelf waters are modified further by turbulent mixing associated with the dissipation of tidal energy taking place over the continental shelves fringing the continent (Egbert and Ray, 2003), including the vicinity of the ice shelf front (Makinson et al., 2006) and sub-ice-shelf cavities (Makinson, 2002). A variety of small-scale processes underlies the conversion of tidal energy into turbulence, most prominent are the breaking of internal gravity waves generated by tidal flows impinging on rough ocean-floor topography (e.g., Robertson et al., 2003), and the formation of thick frictional boundary layers (e.g., Makinson, 2002). The importance of these processes is accentuated by the proximity of the Antarctic continental shelves to the critical latitude (where the tidal frequency is equal to the one of inertial oscillations imposed on ocean flow by the Earth's rotation) of the dominant tidal constituent (M_2), near which the generation of internal tides and frictional boundary layers is most efficient (Robertson, 2001; Pereira et al., 2002). In addition to their role in moulding the properties of shelf waters, tidal fluctuations regulate the flow of those waters across the ice shelf front (Nicholls et al., 2004) and the continental shelf break (e.g. Gordon et al., 2004), often steered by local topographic features such as canyons. On descending the continental slope in broad sheets or narrow plumes, shelf waters tend to focus in largely geostrophic boundary currents that entrain ambient surface and intermediate waters and detrain ventilated fluid in the offshore direction (Baines and Condie, 1998; Hughes and Griffiths, 2006). The shelf waters' descent is promoted further by other smaller scale processes, such as double-diffusion (instabilities due to different diffusivities of heat and salt) and interleaving (layer formation) across the shelf break zone (Foster and Carmack, 1976; Foster, 1987), as well as nonlinearities in the equation of state (thermobaricity and cabbeling) and instabilities of the Antarctic Slope Front, which have been reported to generate cyclonic eddies effecting a net downward transport of shelf water (see Baines and Condie, 1998 for a review). In subsequent stages of its northward journey, the newly formed AABW navigates numerous topographic obstacles and, in so doing, undergoes profound modification due to vigorous turbulent mixing with overlying water masses (e.g. Heywood et al., 2002). A large fraction of this modification is likely driven by flows over small sills in confined passages (Bryden and Nurser, 2003) and mid-ocean ridge-flank canyons (Thurnherr and Speer, 2003), although the breaking of internal tides (Simmons et al., 2004) and internal lee waves (Naveira Garabato et al., 2004) must contribute significantly too.

While theoretical considerations (Marshall and Naveira Garabato, 2007) and altimetric observations of an energy cascade from smaller to larger motions in the Southern Ocean (Scott and Wang, 2005) endorse the view that much of the eddy field's energy is ultimately transferred toward the ocean floor and dissipated by viscous bottom drag, measurements of oceanic fine structure (Naveira Garabato et al., 2004; Sloyan, 2005; Kunze et al., 2006) suggest that a substantial fraction of the energy is dissipated via the generation and breaking of internal lee waves. These waves induce intense turbulent mixing (Figure 4.28) and, in so doing, contribute to driving the lower cell of the Southern Ocean overturning. It thus appears that the upper and lower overturning cells, often treated as largely independent entities in descriptions of the ocean circulation, may be strongly coupled by smaller scale processes. This proposition is consistent with the concurrent intensification of eddy-driven upwelling along inclined surfaces of constant water density, and turbulent mixing of superimposed water layers of different density (diapycnal mixing) in ACC regions of complex topography. In this view, eddy dampening - which is required to sustain a vigorous meridional circulation in the upper ocean - may be connected to the topographic generation of internal lee waves in the abyss (Naveira Garabato et al., 2007). Although the patchy indirect evidence available to date points to topographic generation as the key agent in the transfer of eddy energy to the internal wave field in the ACC (Figure 4.28), other mechanisms are likely to enhance this transfer e.g. interaction between internal waves and mesoscale eddies (Polzin, 2007) and the generation of internal waves by unstable mesoscale processes (Molemaker et al., 2005). The occurrence of these processes is indicated by altimetric evidence of an energy cascade at rather small scales in the Pacific sector of the ACC (Scott and Wang, 2005). Nonetheless, it appears that diapycnal mixing in these upper layers of the ACC is primarily driven by the breaking of downward-travelling near-inertial internal waves generated by the wind in the upper-ocean mixed layer (Alford, 2003).

The key point emerging is that relatively small scale processes in the Southern Ocean exert an important influence on the large-scale behaviour of global ocean circulation over a wide range of time scales of climatic significance. This is a highly significant conclusion when one considers that these processes and interactions are often absent, or parameterized with coefficients tuned to the present ocean state, in the models used to simulate the ocean's evolution (see e.g. Wunsch and Ferrari (2004) for a discussion).

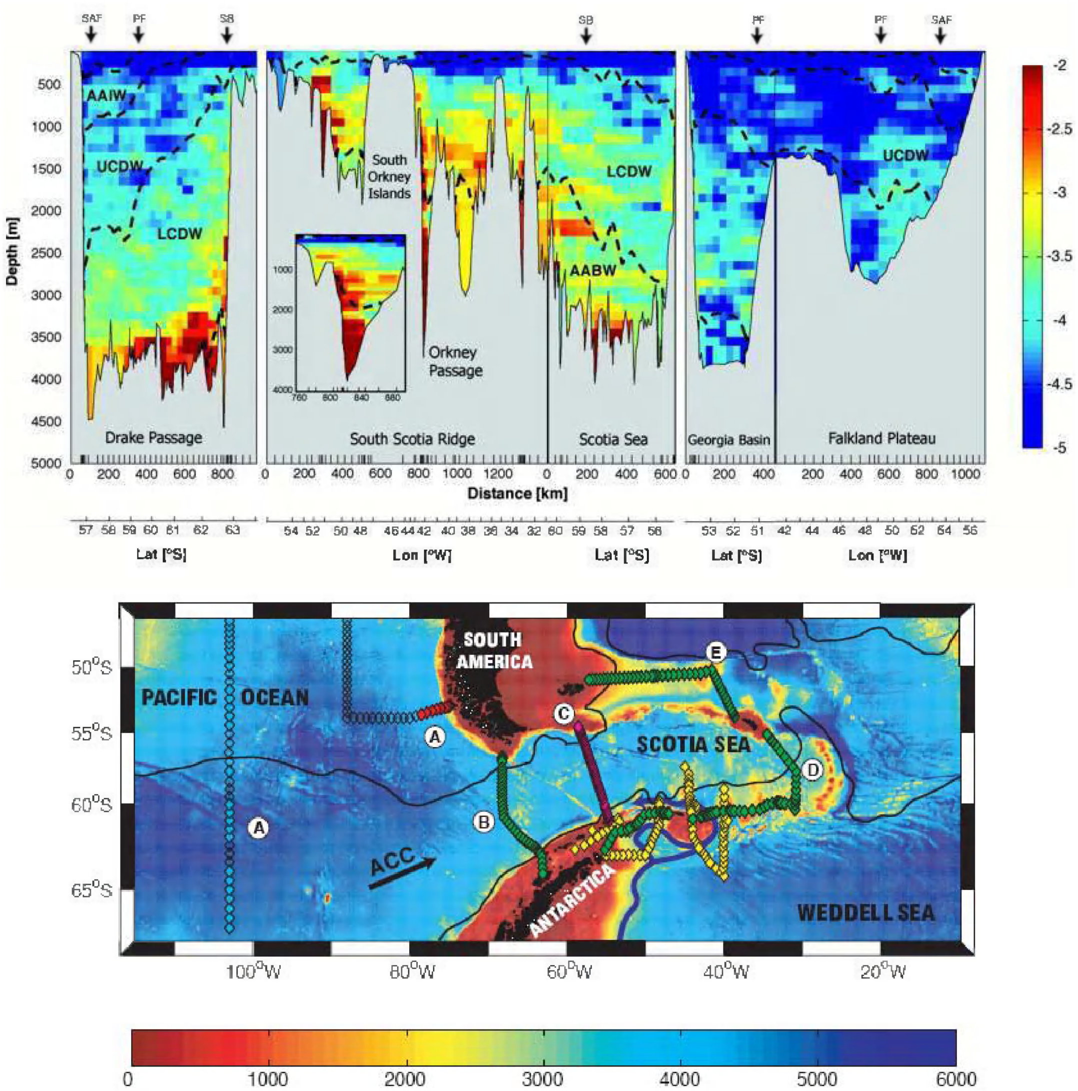


Figure 4.28 The level of turbulent energy is high over rough bottom topography up to a distance of several hundred kilometres and down into the deep ocean. As an indicator of turbulent mixing the vertical distribution of $\log_{10}$ of the turbulent diapycnal diffusivity (in m^2/sec) which can be measured (top), is displayed along a section following the rim of the Scotia Sea anticlockwise (green dots on the bottom graph). Density surfaces separating water masses (AAIW, Antarctic Intermediate Water; UCDW and LCDW, Upper and Lower Circumpolar Deep Water; AABW, Antarctic Bottom Water) are shown by the thick dashed lines. Crossings of the two main frontal jets of the ACC (SAF, Subantarctic Front; PF, Polar Front) and its southern boundary (SB) are marked in the upper axis. Station positions are indicated by tickmarks at the base of the topography. Reproduced from Naveira Garabato et al. (2004).

4.6.7 Dynamics of the circulation and water masses of the ACC and the polar gyres from model results

Observations are still not as frequent and spatially well distributed as needed to establish a well-proven view which would allow the unambiguous determination of forcing mechanisms that might be subject to change. To fill this gap, the dynamics are evaluated using results from models. However, the models themselves have shortcomings, so their results have to be considered with as much care as the ones derived from observations alone. Indeed, some models throw up contradictions that cannot be solved until new generations of models or observations become available.

A number of modelling studies, including both general circulation models (GCM) and simplified theoretical models, have been carried out in an effort to improve our understanding of the basic circulation of the ACC. These studies examined the controls on its transport strength and the mechanisms for import and export of various water masses. The GCM studies involved either the Southern Ocean only or fully global domains, and in most cases included the polar gyres, especially the cyclonic Ross Sea and the Weddell Sea Gyres. These gyres are intimately linked to the ACC, sharing their northern boundaries with the southernmost front of the ACC. Excluding the geographical areas covered by these two gyres, the ACC approaches close to the Antarctic continent (Orsi et al, 1995) and brings its relatively warm Circumpolar Deep Water (CDW) masses close to the edge of the Antarctic Ice Sheet.

A focus of most modelling studies has been to identify the forces that constitute the dynamical balance for the ACC. At the latitude of the Drake Passage, where the ACC is unbounded, the current receives a zonally-continuous momentum input from the dominantly westerly wind. Using the output from various OGCMs, efforts have been made to identify the leading dynamical process that can remove zonal momentum from the ACC at the same rate as the wind input. Given the unbounded nature of the current, it is perhaps not surprising that a classical Sverdrup balance is not able to balance the wind input (Gille et al., 2001), and that instead bottom form stress counteracts the input of momentum from the wind (Grezio et al., 2005). The amount of bottom stress is related to the representation in models of the bottom topography; models with a smoothed representation of topography produce less form stress and thus higher transport values for the ACC than those with rougher topography. Increasing the wind stress within the Southern Ocean increases the ACC transport (Gnandadesikan and Hallberg, 2000).

Eddies that result from hydrodynamic instability of the mean flow of the ACC also play a role in the momentum balance. At high latitudes the length scale at which oceanic flows are affected by the Earth's rotation is rather small - of the order of several kilometres - which means that energetic eddies are relatively small. Indeed they may be smaller than the typical grid cell in an OGCM. Most modelling studies have been carried out at a relatively coarse resolution, in which case they would not simulate eddies well. Some others, of higher resolution, do provide eddy-permitting simulations (e.g. Maltrud and McClean, 2005). The eddy-kinetic energy in OGCMs varies as a function of the model grid resolution, and this in turn has a significant influence on the simulated transport of the ACC. In some regions of the ACC, eddies cause an upgradient transfer of kinetic energy into the mean flow, while for the major part of the ACC the transfer is downgradient (Best et al, 1999). Because coarse-resolution models represent eddy processes and topography inadequately, they produce an unrealistic simulation of the ACC transport, and of the overall Southern Ocean circulation.

Under some scenarios of modelled climate change, there are significant changes in the wind field that drives the ACC. The response of the ACC to a change in the wind field occurs in the remarkably short period of two days (Webb and de Cuevas, 2006). The response is largely barotropic (induced by sea surface elevation) and controlled by the topography, with the changed wind stress quickly transferred by the barotropic flow into the bottom topography as form stress. This is an important finding in the context of climate change, as it suggests that changes in atmospheric circulation can be quickly transmitted into changes in ocean circulation. The ability of the ACC to respond quickly to the wind may explain the observed poleward shift of the ACC over recent decades. An analysis of an OGCM in which the observed poleward shift of the ACC was simulated lends support to the idea that human-induced climate change is currently influencing the ACC and will continue to do so over the coming century (Fyfe and Saenko, 2005).

Although numerical modelling of the ACC and adjacent polar gyres sheds some light on the behaviour of the ACC and its interaction with the gyres, at least two pressing questions remain poorly addressed. First, the transport volume of the ACC remains poorly constrained in different OGCMs, particularly so in models typically used in IPCC simulations, which show a wide discrepancy in transport values even where the external forcing is similar (Ivchenko et al., 2004). Analyzing the ACC transport in 18 coupled atmosphere-ocean models Russell et al (2006b) found that compared to the observed transport estimate of 135 Sv, the coupled models produced a spread ranging from a low of -6 Sv to a high of 336 Sv. They concluded that it is difficult, at present, to get the Southern Ocean “right” in coupled atmosphere-ocean models. This shortcoming reflects the lack of high resolution in many model simulations, and should be overcome as Southern Ocean eddy resolving ocean models become increasingly prevalent. Secondly it is currently difficult to model the interaction of the ACC and polar gyres with the edge of the Antarctic Ice sheet. To properly tackle this problem requires an OGCM to be coupled interactively to an ice sheet model.

4.7 Antarctic Sea Ice Cover during the Instrumental Period

4.7.1 Introduction

This section considers the variability and trend in Antarctic sea ice area and extent over the last century. This time splits into two very distinct periods. Since the 1970s microwave instruments on polar orbiting satellites have enabled sea ice observations to be made year-round and even during periods of complete cloud cover. Before that time, data were reliant on sparse ship observations that were mainly collected during the summer months.

4.7.2 Sea ice cover in the pre-satellite era

Detailed maps of the distribution of the sea ice cover at a good temporal resolution could not be made before the advent of the satellite era, because of the vast extent of sea ice in the region, general inaccessibility and adverse weather conditions. Observations of the sea ice cover were confined mainly to ship observations, such as those compiled by MacKintosh and Herman (1940). During the whaling period, the ships (which were not icebreakers) were normally located at or near the ice edge, the location of which was not precisely defined or consistently identified. The ice edge as defined using satellite passive microwave data is usually taken as the 15% ice concentration contour, and it is not known how well this would match the ice edge as observed by ships (Worby and Comiso, 2004). The whaling data were used by de la Mare (1997) to infer that the ice cover in the 1950s and 1960s was significantly more extensive than that seen in more recent times as revealed by satellite data.

A more direct comparison of the satellite monthly climatology for the period 1978 to 2007 with the MacKintosh monthly data set, as digitized from the original maps by Wadhams (personal communication, 2001), is presented in Figure 4.29. Assuming that the earlier ship observations provide a reasonably accurate representation of average ice edge locations, then the average locations of the ice edge during the satellite period were usually further south, showing that ice cover was more extensive in the 1950s and 1960s than later. These results are consistent with the observation that Antarctica has warmed during the same period (Steig et al., 2009). Interpretation of these results, however, has to be done in the context of the uncertainties as pointed out by Ackley et al. (2003).

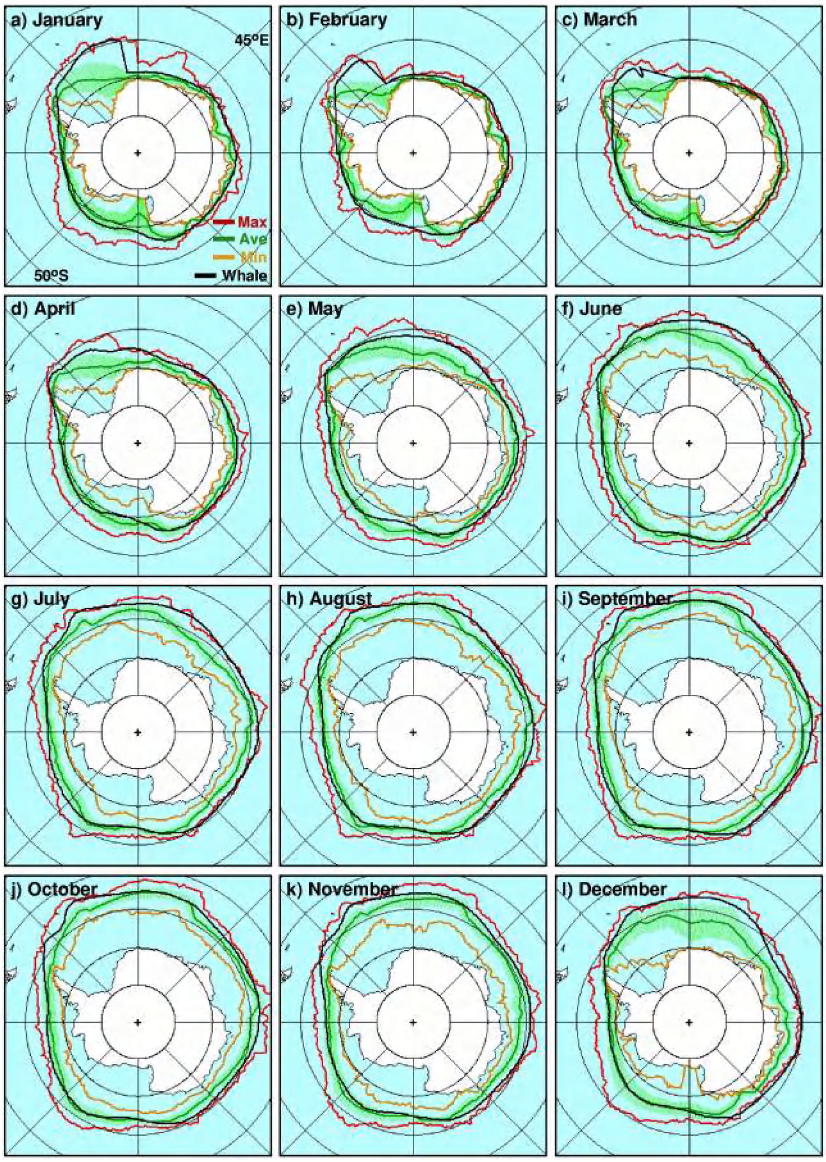


Figure 4.29 Average monthly location of the ice edge during the whaling period (1930s to 1940s) as compiled by MacKintosh (black), and as derived from satellite data for the period 1978 to 2007 (dark green). The standard deviation of the ice edge along different longitudes separated by one degree is shown in green, while the most northern and most southern locations are represented in red and gold, respectively. (Note: MacKintosh digital data was provided by P. Wadhams in 2001).

Simulations with the climate model LOVECLIM with simple data assimilation provides information on sea ice area during the Twentieth Century. It shows a decrease of 0.5 $\times 10^6$ km^2 from the 1960s to the early 1980s, but a slight increase from 1980 to 2000 (Goosse et al, 2008). Although we lack the sea ice data to confirm the early decrease, the model is consistent with available data from both atmosphere and ocean.

4.7.3 Variability and trends in sea ice using satellite data

The ESMR instrument flown on board the Nimbus-5 satellite was the first scanning microwave radiometer and therefore was first to provide the true spatial distribution of the ice cover at a reasonable time resolution. Although meant primarily for rainfall studies it found its best applications for sea ice studies because of the high contrast in the emissivity of sea ice and liquid water. While the data from this instrument are not used in the time series analysis presented below, the sea ice cover during this period was well documented (Zwally et al., 1983). The data led to the discovery of the existence of the large Weddell polynya about the size of California that persisted throughout winter and for three successive years from 1974 through 1976 (Zwally and Gloersen 1977). As discussed in Chapter 2, combining ESMR data with SMMR and SSM/I observations is difficult because (i) there is no overlap in data to ensure that the ESMR data provides the same ice edge and ice extent as in the other data sets; (ii) there are too many gaps including more than 3 months in succession in 1975; (iii) and inaccurate ice concentrations were recorded because of only one channel being available for the sensor, making it impossible to correct for spatial variations in emissivity and temperature. Qualitative analysis can be done: for example, a four-year average of the ESMR data can be compared with four-year averages of SMMR and SSM/I data during different periods, but, unless large changes are apparent, the comparison may be difficult to interpret. The following discussion concentrates on the period from 1978 when the SMMR data became available.

The large-scale seasonal variation of the sea ice cover in the Southern Hemisphere is depicted by the colour-coded multiyear monthly averages of the ice cover (Figure 4.30). The AMSR-E ice concentration data are gridded at 12.5 km to provide a more accurate spatial representation of the sea ice cover than the SMMR and SSM/I data, which are gridded at 25 km resolution. The maps were derived by taking the average of all data available for each month, using the dataset that starts in June 2002 and ends in December 2007. The monthly data set thus represents contemporary ice cover and could serve as a guideline for expected distribution of the Antarctic sea ice cover. The series of images starts in January, usually the warmest mid-summer month and the time of highest melt rate. The minimum extent usually occurs in late February or early March. There are two primary locations where ice survives in the summer: one in the Western Weddell Sea and the other in the Bellingshausen, Amundsen, and Ross Seas. The extents of the perennial ice in these two regions are comparable, but vary slightly in magnitude and location from one year to another. The period of most rapid growth is from April to June; by July it has reached close to its maximum extent, which is normally in September or October. Ice cover decays rapidly between November and January. Around the continent, sea ice advances the least from the coastline mainly in the Western Pacific (90°E to 150°E) and at the tip of the Antarctic Peninsula, where the coast is farthest to the north. The shape of the ice cover during ice maxima (September) is almost symmetrical around the continent, with a tendency to have a sharp corner at about 217°E, which is in part influenced by the shape of the bathymetry there.

Among the most distinctive features in the inner zone of the ice cover are the reduced ice concentrations (i) adjacent to the Ross Ice Shelf, (ii) at or near the Maud Rise (5°E), and (iii) near the Cosmonaut Sea (45°E). These are regions where short term (or transient) polynyas usually occur in mid-winter (Comiso and Gordon, 1996), and which have been associated with the initiation of deep ocean convection and the formation of the high salinity bottom water that drives the global thermohaline circulation. Because they normally appear during different times of the winter season and in different places they are not well represented in the winter climatological averages. The features are most evident in the images in late winter to spring (September to November) suggesting that the ice cover in these regions is vulnerable to divergence and melt because of relatively thinner ice and warmer water under the ice. Such regions have also been the scene of high productivity (Smith and Comiso, 2008) reflecting the possible influence of sea ice, as has been suggested previously (Smith and Nelson, 1986).

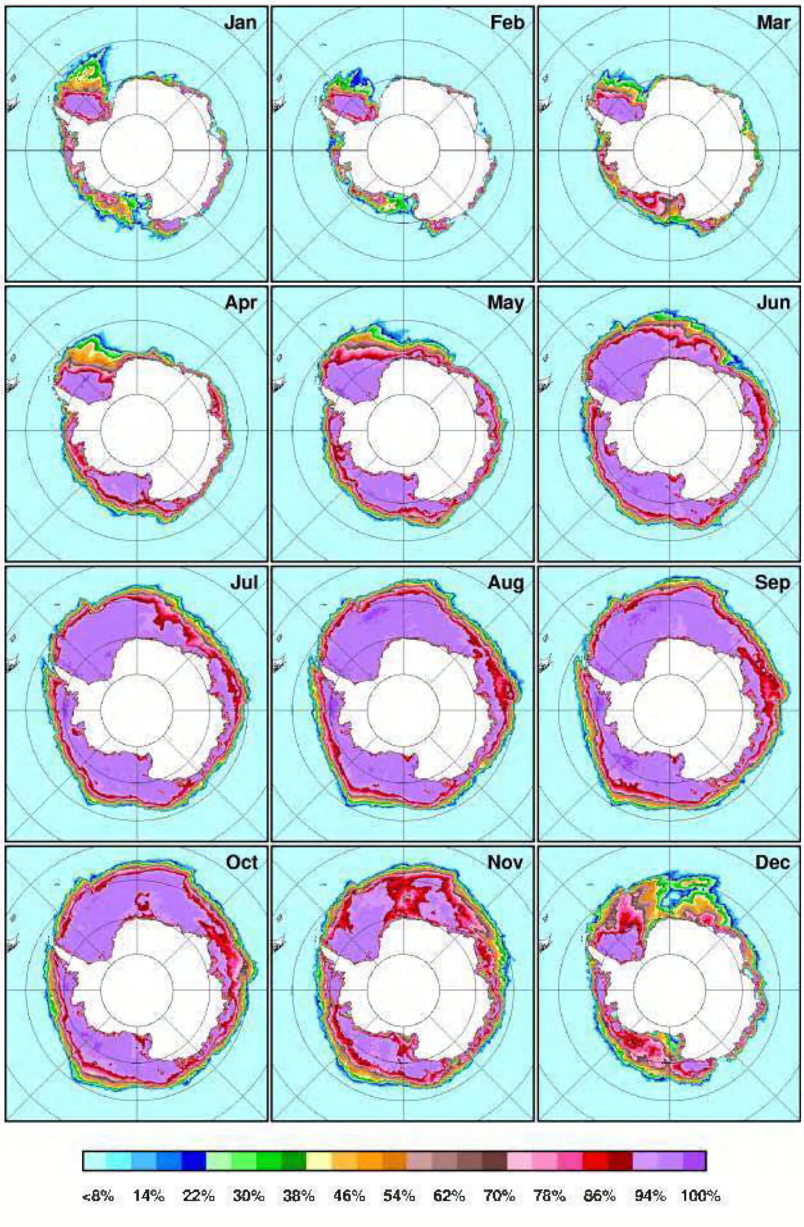


Figure 4.30 Monthly climatology of the sea ice concentration as derived from AMSR-E data (2002 to 2007). (derivation explained in text).

The parameters that have been used for quantifying the variability and trends in the sea ice cover are ice extent and ice area, both of which are derived from the passive microwave ice concentration data. Ice extent is defined as the integrated sum of the area of all the pixel elements with at least 15 % concentration, while ice area is the integrated sum of the area of each pixel multiplied by the ice concentration. Time series plots of monthly values of the ice area, ice extent and also average ice concentration as derived from SMMR and SSM/I data from November 1978 to December 2007 are presented in Figure 4.31. Over the 28-year period of consistently processed passive microwave data, the seasonal and yearly variations in the sea ice extent and ice areas appear to be very similar. The annual maxima and minima vary only slightly despite relatively large interannual variations in average ice concentrations, especially during the summer period. In the summer, the large fluctuation may be caused by the vulnerability of the ice cover to divergence due to winds, waves and other forcings, because of relatively smaller extents. For example, more ice could be advected further north where the water is warmer during some years when southerly winds are prominent. The average ice concentration is almost constant in the winter, at about 83%, while the average ice concentration in the summer ranges from 59% to 69%. The distributions of ice extents vary consistently with the ice area, with the latter being somewhat lower, as expected.

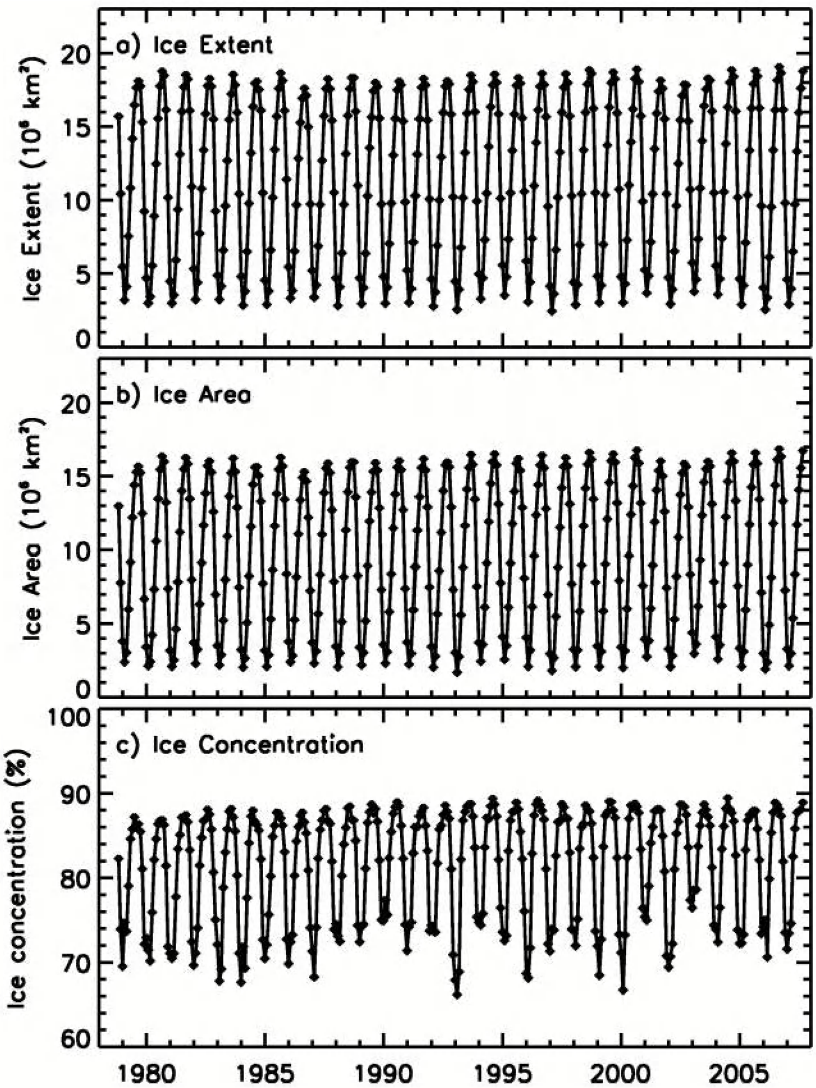


Figure 4.31 Monthly extent (the total area of sea ice and leads within the 15% sea ice concentration limit), area (the integrated area of sea ice within the 15% sea ice concentration limit), and average ice concentration in the Southern Hemisphere.

During the years when the monthly ice extent and area in the winter were most expansive (1980) or least expansive (1986), the corresponding values in the subsequent summer were not the most expansive or least expansive, respectively. Thus the decay patterns are not significantly influenced by the extent of ice during the preceding winter period. Conversely, the growth patterns are also not significantly influenced by the extent of ice during the preceding summer. This counter-intuitive phenomenon is more apparent on a regional basis. For example, in the Weddell Sea, anomalously extensive ice cover in winter is usually followed by anomalously low ice cover area in the summer and vice-versa (Zwally et al., 1983).

To assess interannual changes and trends in the sea ice extent and ice area, monthly anomalies are used, as in previous studies (e.g., Zwally et al., 2002a). The monthly anomalies were estimated by subtracting monthly climatologies from each month from November 1978 to December 2007. The monthly climatologies were generated by taking the average of all satellite data available for each of the months. For consistency only SMMR and SSM/I data during the period were used. The sea ice extent shows a positive trend of 0.9 ± 0.2% per decade for the Southern Hemisphere for the period January 1979 to December 2008. This is consistent with the 1.0 ± 0.4% per decade reported by Zwally et al. (2002a) for the 1979 to 1998 period. The trend in ice area is slightly more positive at 1.6 ± 0.2%/decade, in part because of a positive trend in ice concentration at about 0.93 ± 0.13%/decade. The errors cited are the statistical error as provided by the standard deviation of the slopes in the regression analysis. Unknown biases that may be associated with different calibration and resolution of the SMMR and SSM/I sensors are not reflected by this error. Assuming that the latter is small, as inferred from the overlap of SMMR and SSM/I data of about a month, the positive trend is small but significant. This is consistent with observed cooling in parts of the Antarctic during the period (Comiso, 2000; Kwok and Comiso, 2002).

Trend analysis of the ice extent in different Antarctic sectors (as defined in Zwally et al., 2002a) (see Figure 4.32) yields positive trends of varying magnitude in all except in the Bellingshausen/Amundsen Seas sector. The trend is least positive in the Western Pacific and the Weddell Sea sector at 0.7 ± 0.6 and 0.8 ± 0.5%/decade followed by the Indian Oceans sectors at 1.9 ± 0.6%/decade. The most positive is the Ross Sea sector at 4.3 ± 0.7%/decade; an increase that was simulated by the LOVECLIM model (Goosse et al., 2008). The trend in the Bellingshausen/Amundsen Seas sector is -6.8 ± 1.0%/decade, which serves to balance the relatively high trend in the Ross Sea. Since these two sectors are adjacent to one another, the opposite trends in the two sectors are in part caused by the advection of ice from one sector to the other. This pattern of change has been linked to the recent deepening of the Amundsen Sea Low as a result, primarily, of the loss of stratospheric ozone (Turner et al., 2009).

The Antarctic Peninsula adjacent to the Bellingshausen/Amundsen Seas sector is an area of marked warming, as described previously by King (1994) and Jacobs and Comiso (1997). Also, the variability of ice in the Ross Sea region is associated with the influence of ENSO (Ledley and Huang, 1997) and the continental area adjacent to it has been experiencing some cooling during the last two decades (Comiso, 2000; Doran et al., 2002). The positive trend in the Ross Sea, which is the site of a major coastal polynya, suggests increased ice production and a more important role of the region in bottom water formation.

Yearly maximum and minimum ice extent and ice area were estimated for each year using 5 day running averages of daily data and the results are presented in Figure 4.33. While interannual changes in wind circulation can be a significant factor, these parameters are basically associated with the thermal state of the ice-covered ocean and adjacent seas. The plots for maximum extents and areas are shown to be basically stable with very little interannual variability. Relatively high ice extents occurred in 1980, 1998, 2000, 2005 and 2006 while relatively low values occurred in 1986 and 2002. The high and low values in the ice area maxima are relatively the same but the variability varies slightly because of interannual changes in sea ice concentration. The interannual fluctuations in the ice minimum are much higher with relatively high values in ice extents occurring in 1982, 1983, 1986, 1987, 1994, 1995, 2001 and 2003 while relatively low values occurred in 1993, 1997, and 2006. The interannual changes in ice area minima are similar but there are significant differences, such as the changes from 1982 to 1983 and from 1986 to 1987 where the ice extent shows an increase from one year to the next while the ice area shows a decrease. Such a phenomenon is likely caused by divergences associated with wind conditions that can cause an increase in ice extent but not in ice area. Trend analysis of the data yielded results very similar to those of the overall trend, with the trends of maximum ice values being 0.9 ± 0.4%/decade and 1.5 ± 0.4%/decade for ice extent and ice area, respectively. The trends of minimum ice values are 0.6 ± 2.7 and 1.5 ± 3.0%/decade for ice extent and ice area, respectively. These results show that the trends of maximum ice extents and areas are similar to the corresponding minimum ice extents and areas but with higher statistical variability for the latter.

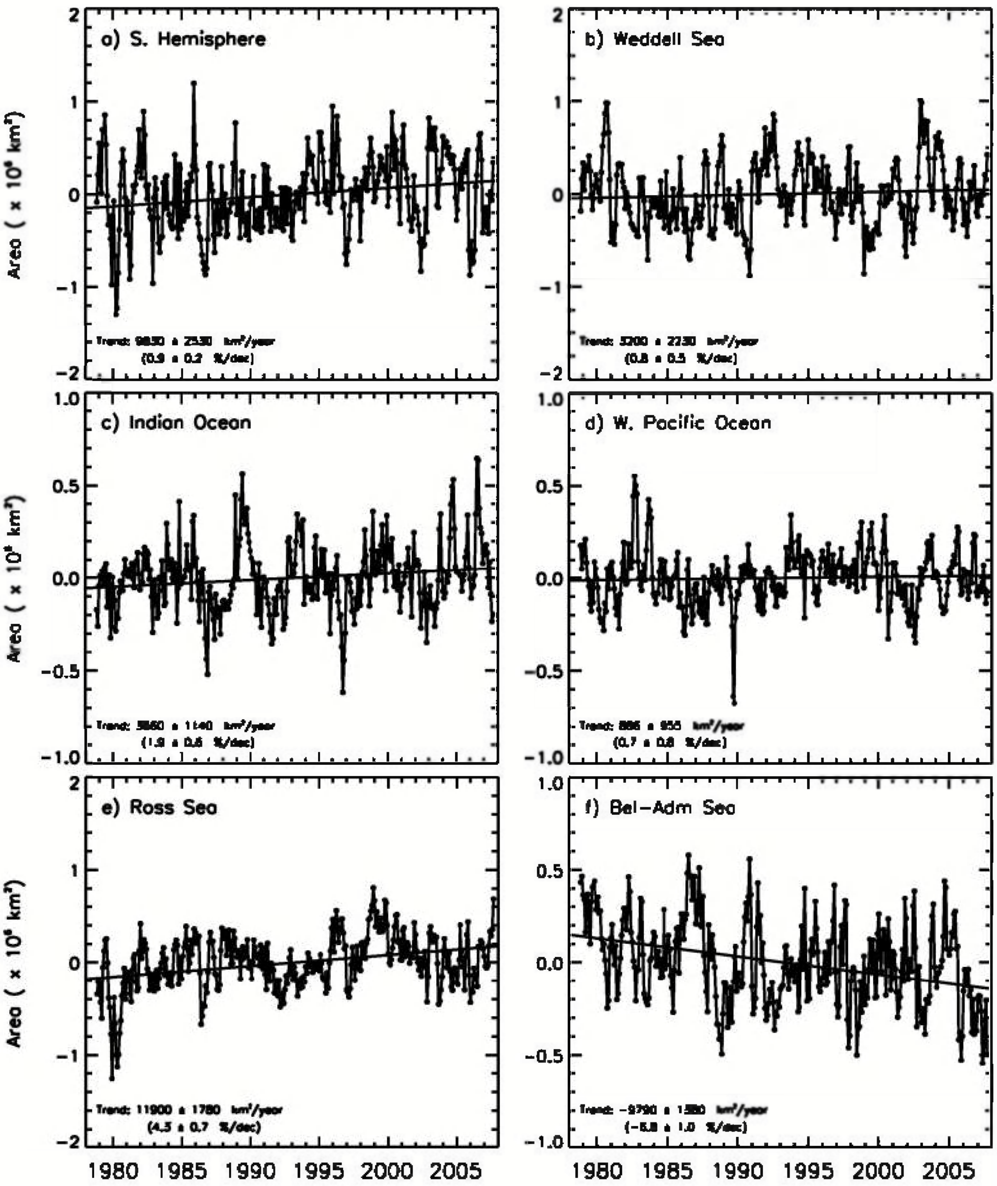


Figure 4.32 Ice Extent monthly anomalies and trend lines for (a) the Southern Hemisphere; (b) Weddell Sea; (c) Indian Ocean; (d) W. Pacific Ocean; (e) Ross Sea; and (f) Bellingshausen/Amundsen Seas Sectors

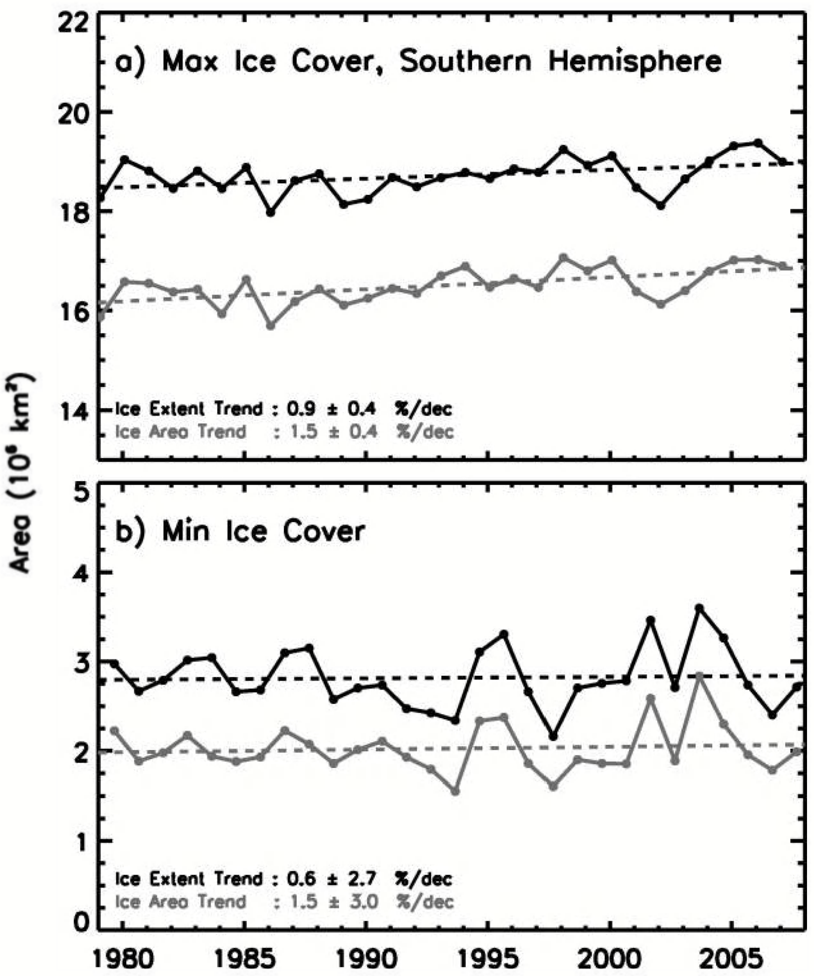


Figure 4.33 Yearly maximum and minimum extent and area of the ice cover in the Southern Hemisphere and trend results.

4.8 The ice sheet and permafrost

4.8.1 Introduction

Ice sheets are to some degree self-regulating systems. Increasing snowfall over the ice sheet will act to increase the ice thickness, but that will then increase the rate of ice-flow towards the coasts and thus remove the extra snowfall. Thus the ice sheet will evolve towards a shape and pattern of flow specific to the current climate, where flow exactly compensates the spatial pattern of ice accumulation (snowfall and frost deposition) and ice ablation (melting, wind erosion and calving).

This equilibrium state is a useful concept, but it is rarely realised; the driving environmental parameters of climate are themselves constantly changing, which continually modifies the equilibrium state sought by the ice sheet. At any given time, the changes in the ice sheet reflect responses to both recent and long-term changes in climate, and for this reason, areal extent, magnitude and duration of changes, as well as time scale must be carefully considered in any discussion of ice-sheet change.

Until the era of satellites, large-scale changes in the ice sheet were considered likely to unfold over thousands of years (Payne et al., 2006). Now, significant accelerations and decelerations of large outlet glaciers and ice streams have been observed on much shorter timescales (Bindschadler et al., 2003; Rignot and Kanagaratnam, 2006; Truffer and Fahnestock, 2007). In Antarctica, these changes affect a complex drainage system of ice streams and tributaries whose full extent has only become appreciated through satellite observations (Joughin et al., 1999; Bamber et al., 2000). Since the 1970s, there have been competing hypotheses concerning the influence of ice shelves on the flow rates of upstream glaciers. Satellites have confronted these hypotheses with empirical measurements for the first time. Another important contribution of satellites has been to identify two key regions of change in Antarctica; one near the northern tip of the Antarctic Peninsula, another within the rarely visited Amundsen Sea sector of West Antarctica.

Despite a good understanding of the complexity of the issues, its importance to understanding sea level rise has meant that measurement of the ice sheet's mass balance has been a primary goal of Antarctic science since the early efforts following the IGY (1957/58). Many of these were based on accounting methods involving calculations of the imbalance between net snow accumulation and outflow of ice over particular domains. Such efforts have always been hampered by the intrinsic uncertainty in measuring these parameters, and very few have produced measurements of ice-sheet imbalance that do not plausibly allow changes in a particular domain to be either positive or negative. So while there have been a few notable exceptions (e.g. Joughin and Tulaczyk, 2002; Rignot and Thomas, 2002; Rignot, 2008), and future efforts based on satellite data may prove to be valuable, our best measurements of change across the majority of the Antarctic ice sheet come not from accounting methods, but rather from those techniques that seek to measure the changing volume of the ice-sheet directly.

The most successful of these techniques of direct measurement has been the use of satellite altimetry (Figure 4.34). A number of research groups have evaluated data beginning in the early 1980s. Spanning data from multiple satellite altimeters, they have produced broadly consistent results (e.g., Wingham et al., 1998, 2006a; Davis et al., 2005; Zwally et al., 2005). These results illustrate that separate catchment basins within the ice sheet behave somewhat independently. What altimetry often fails to capture are the largest changes at the ice sheet margins, where steep slopes lead to large errors of measurement, and the large thickness changes on the floating ice shelves at the perimeter, where elevation changes represent only 1/8 of the full thickness change.

More recently, measurements of changes in the Earth's gravity field have been made using GRACE (Gravity Recovery and Climate Experiment). These satellites have confirmed that the Amundsen Sea sector is losing mass to the ocean (Velicogna and Wahr, 2006; Ramillien et al., 2006; Chen et al., 2006). The GRACE system works by tracking the range between two orbiting satellites: their differential accelerations provide a sensitive measure of how the distribution of mass across the Earth's surface is changing. The system is unable to distinguish separately the changes in ice, rock, and air; so the flows within the Earth's mantle and atmosphere must be removed to reveal changes in the ice sheets. So far this correction has been derived using models of the atmospheric and lithospheric mass flows, but observations of isostatic uplift measured in the field using GPS receivers mounted on exposed rock outcrops should provide a better constraint, especially if GRACE observations are combined with altimetric observations (Velicogna and Wahr, 2002). Chen et al. (2006) report a localised region of mass increase in East Antarctica. This could either be anomalously high snowfall, leading to growth of ice in this region, or an artefact of un-modelled post-glacial rebound. The system has only been in operation for a few years, and snowfall is variable from year to year, so the long-term significance of the available results can be questioned. However there is little doubt that longer records from these satellites will eventually provide extremely valuable information on the changes in the mass of ice sheets.

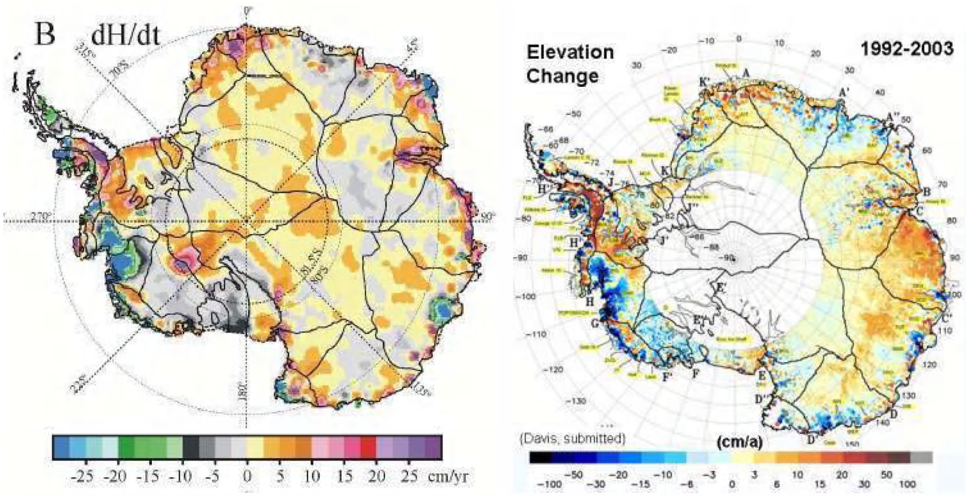


Figure 4.34 Elevation change Zwally (left) and Davis (right).

Antarctica's ice shelves and ice tongues ('ice tongues' are narrowly-confined ice shelves bounded by fjord walls) are particularly sensitive to climate change because both the upper and lower surface of the ice plate are exposed to different systems, each with a potential to cause rapid change: these are the atmosphere and the ocean. The ice-atmosphere system affects the ice shelf through changing surface accumulation, dust and soot deposition, and surface melt; the ice-ocean system controls basal melting or freezing, tidal flexure, and wave action. Moreover, it is now recognized that sea ice, which is not part of the ice sheet, but is viewed more properly as a component of the atmosphere-ocean system, has a large impact on the local climate and dynamics of the ice shelf front, controlling regional surface energy balance, moisture flux, and the presence or absence of ocean swell at the front.
Antarctica's ice shelves have provided the most dramatic evidence to date that at least some regions of the Antarctic are warming significantly, and have shown, what has been suspected for long time (e.g. Mercer, 1978), namely that changes in floating ice shelves can cause significant changes in the grounded ice sheet. This evidence has led to an ongoing re-assessment of how quickly greenhouse-driven climate changes could translate into sea level rise (Meehl et al, 2007).

Ice shelves in two regions of the Antarctic Ice Sheet have shown rapid changes in recent decades: the Antarctic Peninsula and the northern region of West Antarctica draining into the Amundsen Sea. Because there is such spatial variety in the observed changes, and because the causes of the changes likely vary regionally, changes over the past 50 years are discussed below according to province. A review of events in these two regions will introduce two main mechanisms by which climate change is thought to lead to changes in ice sheet mass balance: surface summer melt increase, leading to ponding, fracturing and disintegration: and basal melting of the ice shelves, leading to shelf thinning and grounding line retreat.

4.8.2 The Antarctic Peninsula

The Antarctic Peninsula north of 70°S represents less than 1% of the area of the entire grounded Antarctic ice sheet, but receives nearly 10% of its snowfall (van Lipzig, et al., 2004a). A third of this area lies close to the coast and below 200 m elevation, where summer temperatures are frequently above 0°C, so that this is the only part of continental Antarctica that experiences substantial summer melt. About 80% of its area is classed as a percolation zone (Rau and Braun, 2002), and melt water run-off is a significant component in its mass balance (Vaughan, 2006).

Beginning in the early 1990s, climatologists noted a pronounced warming trend present in the instrumental record from the Antarctic Peninsula stations (King, 1994; Vaughan and Doake, 1996; Skvarca et al., 1998). This region has the highest density of long-term weather observations in the Antarctic, dating back to 1903 for Orcadas Station. Rates of warming on the Antarctic Peninsula are some of the fastest measured in the Southern Hemisphere (~3°C in the last 50 years) (King, 2003; Vaughan, et al., 2003) and there has been a clear increase in the duration and intensity of summer melting conditions by up to 74% since 1950 (Vaughan, 2006).

A recent study has shown that circa 2005, the Antarctic Peninsula was contributing to global sea level rise through enhanced melt and glacier acceleration at a rate of 0.16 ± 0.06 mm/yr (which can be compared to an estimated total Antarctic Peninsula ice volume of 95,200 km^3, equivalent to 242 mm of sea-level) (Pritchard and Vaughan, 2007). Although it is known that Antarctic Peninsula glaciers drain a large volume of ice, it is not yet certain how much of the increased outflow is balanced by increased snow accumulation. One estimate of mass change due primarily to temperature-dependent increases in snowfall on the peninsula suggested a contribution to sea level of approximately -0.003 mm/yr (Morris and Mulvaney, 2004).

4.8.2.1 Glaciers

The ice-cover on the Antarctic Peninsula is a complex alpine system of more than 400 individual glaciers that drain a high and narrow mountain plateau. The tidewater/marine glacier systems in this region (excluding ice shelves and the former tributary glaciers of Larsen A, B and Wordie ice shelves) have an area of 95,000 km^2 and a mean net annual accumulation of 143 ± 29 Gt/yr (after van Lipzig et al., 2004a). Changes in the ice margin around the Antarctic Peninsula based on data from 1940 to 2001 have been compiled (Ferrigno et al., 2002, 2006; Cook et al. 2005). Analysis of the results revealed that of the 244 marine glaciers that drain the ice sheet and associated islands, 212 (87%) have shown overall retreat since their earliest known position (which, on average, was 1953). The other 32 glaciers have shown overall advance, but these advances are generally small in comparison with the scale of retreats observed.

The glaciers that have advanced are not clustered in any pattern, but are evenly scattered down the coast (Figure 4.35). Examination of the timing of changes along the peninsula indicates that from 1945 until 1954 there were more glaciers advancing (62%) than retreating (38%). After that time, the number retreating has risen, with 75% in retreat in the period 2000-2004. The results indicate a transition between mean advance and mean retreat; a southerly migration of that transition at a time of ice shelf retreat and progressive atmospheric warming; and a clear regime of retreat which now exists across the Antarctic Peninsula (Figure 4.36). The rapidity of the migration suggests that atmospheric warming may not be the sole driver of glacier retreat in this region. Glaciers with fully grounded marine termini exhibit unusually complex responses to changing mass balance because in addition to the normal forcings they are also subject to oceanographic forcing and subglacial topography. Future analysis of changes in all boundary conditions may reveal why the glaciers have responded in this way.

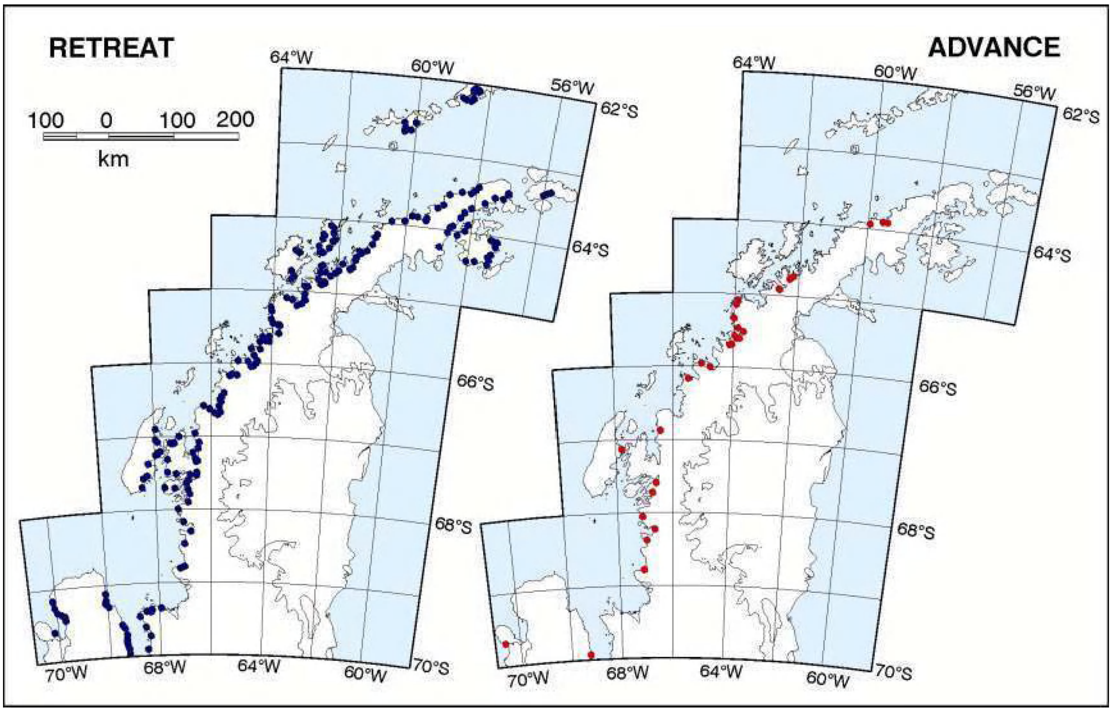


Figure 4.35 Overall change observed in glacier fronts since earliest records. (From Cook et al., 2005).

A recent study of flow rates of tidewater glaciers has revealed a widespread acceleration of ice flow across the Peninsula (Pritchard and Vaughan, 2007). This widespread acceleration trend was attributed not to meltwater-enhanced lubrication or increased snowfall but to a dynamic response to frontal retreat and thinning. Measurements were taken from over 300 glaciers on the west coast through nine summers from 1992 to 2005. They showed that overall flow rate increased by 10% and that this trend is greater than the seasonal variability in flow rate. A comparison of measurements between the years 1993 and 2003 only (with profiles tailored to optimize coverage in just these years) revealed a slightly greater overall acceleration of 12.4 ± 0.9%.

The loss of ice shelves (Section 4.8.2.2) has caused acceleration of the glaciers that fed them (Rignot et al., 2004a, 2005; Rott et al., 1996; Scambos et al., 2004) creating locally high imbalances in ice mass. Immediately after break-up, glaciers flowing into the now-collapsed sections of the Larsen Ice Shelf accelerated to speeds of 2 to 8 times the pre-disintegration flow rate (Rignot et al., 2004a; Scambos et al., 2004). The glaciers flowing into the Wordie Ice Shelf also accelerated following ice shelf loss, and have been losing mass to the ocean over the last decade (Rignot et al., 2005). One of these, Fleming Glacier, accelerated by about 50% during the period 1974-2003, and the region was losing mass at 18±5.4 Gt/yr. A field campaign carried out in December 2008 using GPS measurements and an airborne laser survey confirmed that the glacier maintains these high flow rates and experiences a pronounced ice thinning (Wendt et al., In Press). The ice flux increase may be partially offset by increased precipitation in the western Peninsula (Turner et al., 2005b), but both ice shelves (Fox and Vaughan, 2005) and glaciers in the west (Pritchard and Vaughan, 2007) continue to retreat. The combined estimate of mass loss (as of 2005) was 43±7 Gt/yr, but a more recent assessment of the region suggests this rate has slowed (28±45 Gt/yr, Rignot et al., 2008). In addition to the increase in flow rates, a recent study has revealed profound dynamic thinning of collapsed-ice shelf tributary glaciers flowing from the Antarctic Peninsula plateau to both east and west coasts (Pritchard et al., submitted). Analysis of ICESat laser altimeter data, processed along-track for the period 2003-2007, showed how surface elevation has changed over the whole of the Antarctic Peninsula. The high, central plateau and slow flowing ice caps thickened at rates as high as 1 m/yr. In contrast, some of the highest rates of thinning recorded either in Antarctica or Greenland (up to tens of metres per year) are occurring on glaciers that flowed into ice shelves that have now disappeared. Glacier tributaries feeding the intact but thinning ice shelves of Larsen C and remnants of Larsen B, plus George VI Ice Shelf and the little-studied Larsen D also thinned at rates up to several metres per year. This behaviour confirms that glaciers are very sensitive to ice shelf thinning as well as collapse, and that shelf collapse leads not just to short-term and localized adjustment but to sustained, widespread and substantial loss of grounded ice from tributary glaciers (Pritchard et al., submitted).

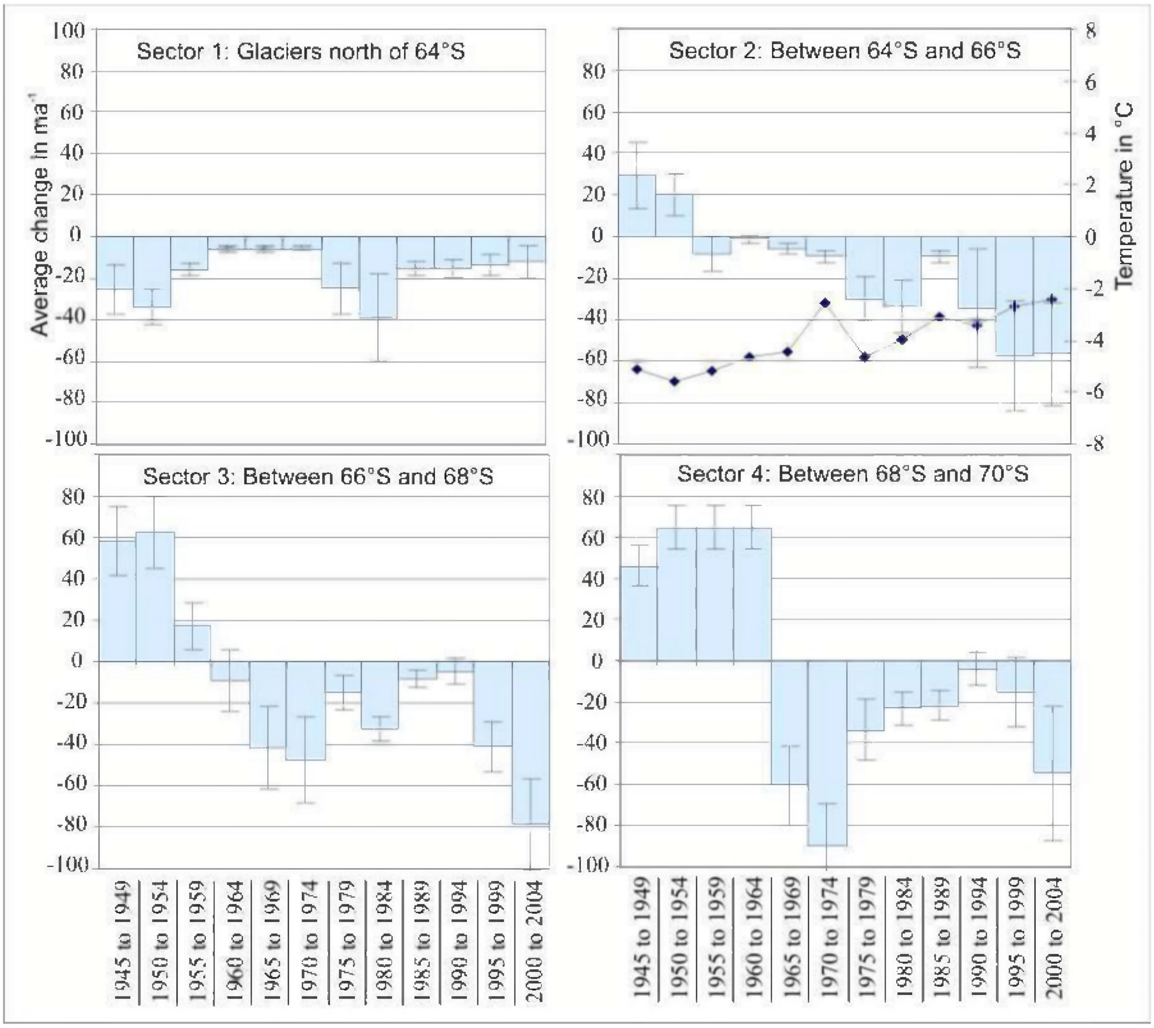


Figure 4.36 Change in Antarctic Peninsula glaciers over time and by latitude. Prior to 1945 the limit of glacier retreat was north of 64°S; in 1955 it was in the interval 64-66°S; in 1960 between 66-68°S and in 1965 between 68-70°S. (from Cook et al., 2005).

4.8.2.2 Ice shelves

Retreat of several ice shelves on either side of the Peninsula was already occurring when scientific observations began in 1903. Since that time, ice shelves on both the east and west coasts have suffered progressive retreat and some abrupt collapse (Morris and Vaughan, 2003; Scambos et al., 2000). Ten ice shelves have undergone retreat during the latter part of the 20th Century (Cooper, 1997; Doake and Vaughan, 1991; Fox and Vaughan, 2005; Luchitta and Rosanova, 1998; Rott et al., 1996, 2002; Scambos et al., 2000, 2004; Skvarca, 1994; Ward, 1995) (Table 4.1). Wordie Ice Shelf, the northernmost large (>1,000 km^2) shelf on the western Peninsula, disintegrated in a series of fragmentations through the 1970s and 1980s, and was almost completely absent by the early 1990s. The Wordie break-up was followed in 1995 and 2002 by spectacular retreats of the two northernmost sections of the Larsen Ice Shelf (termed Larsen 'A' and Larsen 'B' by nomenclature proposed by Vaughan and Doake, 1996) and the last remnant of the Prince Gustav Ice Shelf (Figure 4.37). A similar 'disintegration' event was observed in 1998 on the Wilkins Ice Shelf (Scambos et al., 2000), but much of the calved ice remained until 2008 when dramatic calving removed about 1,400 km^2 of ice. The ice bridge connecting the Wilkins Ice Shelf to Charcot Island disintegrated in early April 2009. In all these cases, persistent seasonal retreats by calving (Cooper, 1997; Skvarca, 1993; Vaughan, 1993) culminated in catastrophic disintegrations, especially for the Larsen A (Rott et al., 1996; Scambos et al., 2000) and Larsen B (Scambos et al., 2003).

The sequence of events leading up to the collapse of the Larsen B ice shelf suggests the processes responsible for the ultimate disintegration. In the 35-day period from 31 January 2002, satellite images recorded by the Moderate Resolution Imaging Spectroradiometer (MODIS) revealed a disintegration of a 5,700 km^2 section of the Larsen B ice shelf. The January MODIS images showed that prior to its disintegration, the Larsen B ice shelf was subject to more extensive ponding of meltwater than in previous years (Scambos et al., 2004). As this water drained into pre-existing crevasses, and filled them, the water pressure would have been sufficient to propagate the cracks through the entire thickness of the ice shelf (Weertman, 1973; Scambos et al. 2000). Satellite radar interferometry has been used with ice flow models to show that the ice shelf sped up considerably in the period before its final collapse because of weakening within its margins, perhaps as a consequence of this mechanism (Vieli et al., 2007). Once the Larsen B ice shelf had disintegrated into icebergs, the forces set up as they toppled against one another drove them rapidly apart (MacAyeal et al., 2003). A MODIS image taken on 7 March 2002 (Figure 4.37) shows a plume of icebergs being ejected, clearing the bay that was previously occupied by the ice shelf.

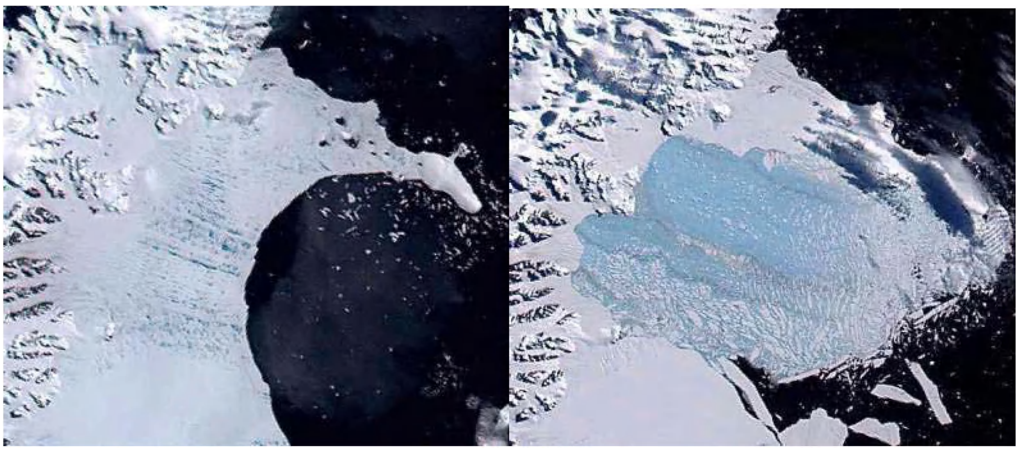


Figure 4.37 Rapid disintegration of Larsen B ice shelf. Image on left collected on January 31, 2002 and on right collected on March 7, 2002.

The latest results reveal an overall reduction in total ice shelf area on the Antarctic Peninsula by over 27,000 km^2 in the last 50 years. As discussed in the previous section, recent findings (and studies of similar events in the southern Greenland ice sheet; see Howat et al., 2007) have fostered new appreciation of the importance of floating ice on controlling ice flow, and the rapidity with which loss of floating ice could cause an acceleration in the contribution to sea level rise.

The direct cause of the Peninsula ice shelf retreats is thought by many to be a result of increased surface melting, attributed to atmospheric warming. Increased fracturing via melt-water infilling of pre-existing crevasses explains many of the observed characteristics of the break-up events (Scambos et al., 2000; 2003), and melting in 2002 on the Larsen B was extreme (van den Broeke, 2005).

Observations of northward-drifting icebergs support the theory that surface melt ponds or surface firn saturated with melt-water can rapidly culminate in disintegration of either ice shelves or icebergs (Scambos et al., 2004).

Ice Shelf	First recorded date	Last recorded date	Area on first recorded date (Km^2)	Area on last recorded date (Km^2)	Change (Km^2)	*% of original area* remaining	Reference
Müller	1956	1993	80	49	-31	61	*Ward (1995)*
Wordie	1966	1989	2,000	700	-1,300	35	*Doake and Vaughan (1991)*
Wordie	1989	2009	700	96	-600	5	*Wendt et al. (In Press)*
Northern George VI	1974	1995	~ 26,000	~ 25,000	-993	96	*Luchitta and Rosanova (1998)*
Northern Wilkins	1990	1995	~ 17,400	~ 16,000	-1,360	92	*Luchitta and Rosanova (1998)*
	1995	1998			-1,098	85	*Scambos et al. (2000)*
Jones	1947	2003	25	0	-25	0	*Fox and Vaughan (2005)*
Prince Gustav	1945	1995	2,100	~ 100	-2,000	5	*Cooper (1997)*
	1995	2000		47		2	*Rott et al. (2002)*
Larsen Inlet	1986	1989	407	0	-407	0	*Rott et al. (2002)*
Larsen A	1986	1995	2,488	320	-2,168	13	*Rott et al. (1996)*
Larsen B	1986	2000	11,500	6,831	-4,669	59	*Rott et al. (2002)*
	2000	2002		3,631	-3,200	32	*Scambos et al. (2004)*
Larsen C	1976	1986	~ 60,000	~ 50,000	-9,200	82	*Skvarca (1994) and Vaughan and Doake (1996)*

Table 4.1 Summary of changes observed in ten ice shelves located on the Antarctic Peninsula. The figures were obtained from references that recorded the measured area of a particular ice shelf on both the earliest and most recent dates available.

Specific mechanisms of ice shelf break-up are still debated. The role of subsurface waters circulating beneath the shelves in thinning and/or warming the ice remains undetermined. Others have suggested that a change to negative surface mass balance (Rott et al., 1998), or reduced fracture toughness due to a thickening temperate ice layer (Vaughan and Doake, 1996), or basal melting (Shepherd et al., 2002) caused the break-up. Recent modeling and observational studies have shown that the Larsen B, at least, was pre-conditioned to a retreat and breakup by faster flow, increased rifting, and detachment from the coast (Vieli et al., 2007; Glasser and Scambos, 2008); all these are consistent with a thinning shelf in the years leading up to disintegration.

The pattern of ice shelf retreat on the Antarctic Peninsula appears to be consistent with the existence of a thermal limit on ice-shelf viability (Morris and Vaughan, 2003; Vaughan and Doake, 1996) (Figure 4.38). The limit of ice shelves known to have retreated during the last 100 years is bounded by the –5°C and –9°C isotherms (calculated for 2000 A.D.) suggesting that the retreat of ice shelves in this region is consistent with the observed warming trend of 3.5 ± 1.0°C/century (Morris and Vaughan, 2003).

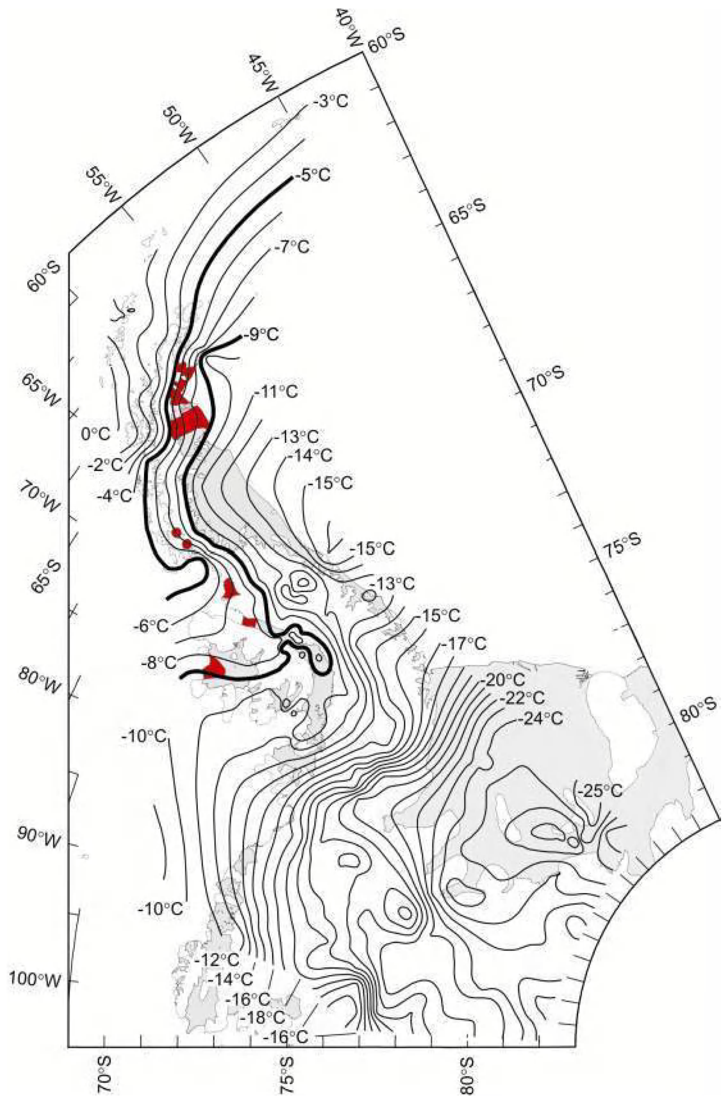


Figure 4.38 Contours of interpolated mean annual temperature. Currently existing ice shelves are shown in grey. Portions of ice shelves that have been lost through climate-driven retreat are shown in red (From Morris and Vaughan, 2003).

4.8.2.3 Sub-Antarctic Islands

Glaciers on the sub-Antarctic islands of Heard Island (53°S, 73°30'E), Kerguelen Islands (49°15'S, 69°35'E) and South Georgia (54°30'S, 36°30'W) have shown accelerating rates of retreat over the past half-century. The glaciers on Heard Island have shown extensive retreat since the 1940s (Allison and Keage, 1986; Kiernan and McConnell, 1999, 2002; Budd, 2000). After a period of advance between 1963-71, most of the recession occurred since 1970 (Allison and Keague, 1986). The total glacierized area has reduced by 11%, and several coastal lagoons have been formed as a result. The rapid glacier recession reflects a temperature rise on the island of about 1.3°C during the last 50 years (Budd, 2000). Of the twelve major glaciers and several minor glaciers on the island, current research includes two specific examples: Stephenson Glacier and Brown Glacier. Historical records, recent observations, and geomorphological evidence indicate that rates of retreat and downwasting of the tidewater Stephenson Glacier, and concurrent expansion of ice-marginal melt-lakes, has increased by an order of magnitude since 1987 (Kiernan and McConnell, 2002). In addition, Brown Glacier retreated 50 metres since 2000/01, contributing to a retreat of approximately 1.1 km since 1950 (a decrease in total volume of about 38%) (Australian Antarctic Division 2005: http://www.heardisland.aq/). Similarly at Kerguelen, glacier recession has accelerated since the early 1970s (Frenot et al., 1993, 1997).

Glaciers, ice caps and snowfields cover over 50% of the island of South Georgia. In a recent study, the changing positions of the 103 coastal glacier fronts on South Georgia were mapped using archival aerial photographs and satellite imagery dating from the 1950s to the present (Cook et al., in submission). Of these, 97% have retreated since their earliest recorded position (which, on average, was 1961). The majority (64%) of the glaciers retreated by between 0 and 500 m since their first observations. Two glaciers stand out as having retreated the most: Neumayer Glacier by 4.4 km since 1957, and the ice front fed by Ross and Hindle Glaciers, by 2.14 km since 1960. The rate of retreat for all 103 glaciers has increased from (on average) 8 m/yr in the late 1950s, to 35 m/yr at present, revealing an accelerating rate of retreat since the 1990s. The recent rapid increase in the average rate is largely due to large increases in retreat rates of glaciers in the north-east of the island, which are currently showing an average of 60 m/yr retreat. The glaciers along the south-west coast of the island, however, are significantly different in their rate of change, due to dissimilar weather patterns caused by orographic effects (Gordon et. al., 2008). They have been in retreat slowly since the 1950s, but this has remained at a constant rate of approximately 8 m/yr. This retreat rate may now be gradually increasing, although on a much smaller scale (currently 12 m/yr). The climate records from South Georgia (recorded at Grytviken from 1905 until 1988, and subsequently from 2001 until 2008) show that in the early 1900s the summer temperatures were relatively high, lower between the 1920s to the 1940s, and higher from the 1950s to the present (Gordon et al., 2008). The retreat of South Georgia glaciers over the past half-century coincides with the recent period of climate warming that began in the 1950s. Acceleration in retreat rates of glaciers on the north-east coast has occurred in the past decade as the climate has continued to warm, and although the glaciers on the south-west side have been slow to respond, their retreat rates may now also be on the increase (Cook et al., in submission).

4.8.3 West Antarctica

West Antarctica has received particular attention because it remains as the last "marine-based" ice sheet, a configuration that was suggested to be inherently unstable, fated to oscillate between fully extended to the edge of the continental shelf or completely lost, having suffered accelerating retreat (Weertman, 1974). Mercer (1968) suggested that full collapse of the West Antarctic ice sheet had occurred as recently as the last interglacial, 125,000 years ago and the concern driving much of the research of the West Antarctic ice sheet was whether such an eventuality was inevitable or even underway.

The West Antarctic ice sheet is conveniently divided into three sectors, each feeding ice into one of the major surrounding seas: the Ross Sea, the Amundsen Sea and the Weddell Sea. Being closest to the US research station at McMurdo, major US field research proceeded on the ice streams of the Ross Sea sector along the Siple and Gould Coasts. Meanwhile, UK field research was focused on the ice streams of the Weddell Sea sector that were similarly closer to their major station at Rothera. Ironically, the largest changes were observed by satellite to be occurring in the Amundsen Sea sector. Each area is discussed in the following sections, beginning with the area exhibiting the largest changes.

4.8.3.1 Amundsen Sea Embayment

The Amundsen Sea sector represents approximately one third of the entire WAIS. Recent observations have shown that this is currently the most rapidly changing region of the entire Antarctic ice sheet. Long before these observations were available the vulnerability and potential significance of retreat in this area was highlighted in a prescient paper by Hughes (1973).

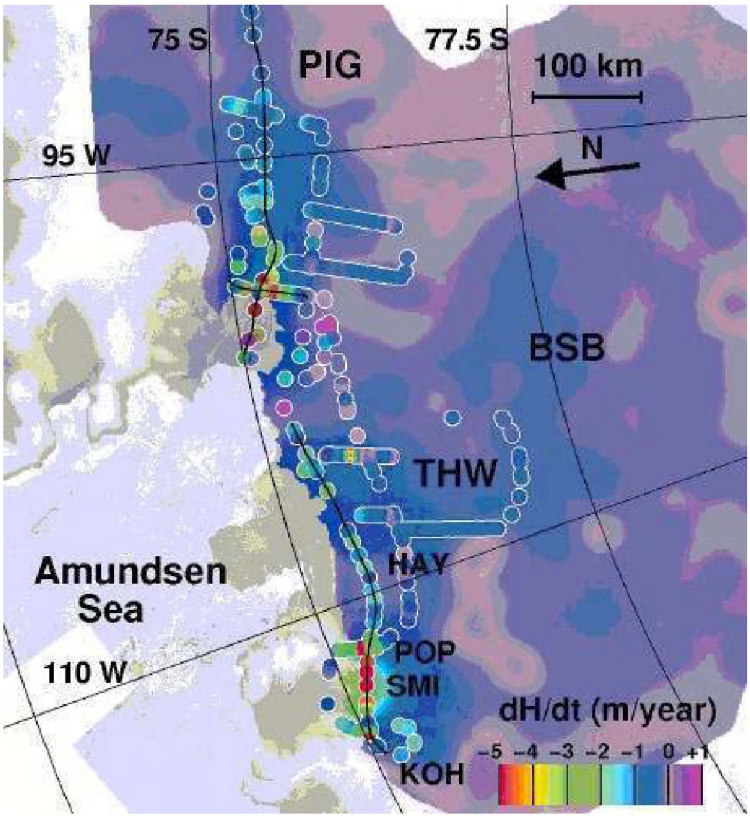


Figure 4.39 Recent and longer term thinning of the Amundsen Sea Embayment sector ice sheet. Regional colours represent rate of elevation change derived from two decades of satellite radar altimetry. Circled colours represent recent elevation changes derived from

airborne altimetry flown in 2002. The difference in the two surveys indicates a recent increase in the rate of thinning (from Thomas et al., 2004b)

Thinning of the ice in the Amundsen Sea sector occurred because of an increase in the discharge of several major outlet glaciers (Figure 4.39). Rignot (1998b) first reported flow acceleration and subsequent grounding line retreat of Pine Island Glacier, one of the two largest West Antarctic outlet glaciers draining into the Amundsen Sea. This retreat has been accompanied by thinning of the ice sheet at a rate of 10 cm/year averaged over a drainage basin twice the area of Great Britain (Wingham et al., 1998). Thinning rates reach well over 1 m/yr at the coast (Shepherd et al., 2001). This discovery of a 10% increase in flow speed in 4 years was anticipated based on oceanographic evidence of very high and increasing basal melt rates beneath the ice tongue fronting the glacier (Jacobs et al., 1996; Jenkins et al., 1997). Later direct measurement of elevation loss near the grounding line and an assumption of flotation at hydrostatic balance, revealed rates as high as 58 $\pm$ 8 m/yr with an ice shelf wide average of 24 $\pm$ 4 m/yr. (Rignot, 2006), exceeding the previously value of 15 m/yr (Shepherd et al., 2004). As basal melt increased, the grounding line retreated, possibly in two stages—during the 1980s and in 1994-96 - each leading to a separate increase in speed (Rignot, 1998b; Joughin et al., 2003). Most recently Rignot (2008a) has shown that the grounding line at Pine Island has retreated still further, with a simultaneous increase in both speed and acceleration. Pine Island Glacier is now moving at speeds nearly double those in the 1970s. No data are available to determine if an earlier period of acceleration occurred, however, a study of past images of Pine Island Glacier's ice shelf indicate that thinning, possibly by as much as 134 metres, occurred in 28 years with significant shifts to the lateral margins, including a major flow shift, beginning perhaps as early as 1957 (Bindschadler, 2001).

Other glaciers in the Amundsen Sea sector have been similarly affected: Thwaites Glacier is widening on its eastern flank, and there is accelerated thinning of four other glaciers in this sector to accompany the thinning of Thwaites and Pine Island Glaciers (Thomas et al., 2004a). Where flow rates have been observed, they too show accelerations, e.g., Smith Glacier has increased flow speed 83% since 1992.

Calculations of the current rate of mass loss from the Amundsen Sea embayment range from 50 to 137 Gt/yr with the largest number accounting for the most recent faster glacier speeds (Lemke et al., 2007; Rignot et al., 2008). Data sources and methodologies vary, but generally when uncertainties and the time intervals analyzed are considered, the estimates are consistent with accelerating rates of loss, in concert with the accelerations of the primary discharging glaciers. These rates are equivalent to the current rate of mass loss from the entire Greenland ice sheet. The Pine Island and adjacent glacier systems are currently more than 40% out of balance, discharging 280 ± 9 Gt/yr of ice, while they receive only 177 ± 25 Gt/yr of new snowfall (Rignot et al., 2008; see also Thomas et al, 2004b). The increasingly negative mass balance is confirmed by several recent radar altimetry assessments of reduction in surface elevation of the Pine Island catchment (e.g., Zwally et al., 2005; Rignot et al., 2008).

Summer temperatures in the Amundsen Sea embayment rarely reach melting conditions, and there is little reason to assume that atmospheric temperatures have had any strong role to play in the changes that have occurred there. Similarly, the patterns of thinning, which are very clearly concentrated on the most dynamic parts of the glaciers, indicate that the changes are not the result of anomalous snowfall. The most favoured explanation for the changes (e.g. Payne et al., 2004) is a change in the conditions in the sea into which this portion of West Antarctica flows (Figure 4.40). ITASE ice core research indicates that marine air mass transport in the Amundsen Sea sector of the WAIS has increased in intensity as of recent decades (Dixon et al., 2005. While there are no adjacent measurements of oceanographic change that can support this hypothesis, it appears to be the most likely option, and the recent observations of relatively warm Circumpolar Deep Water on the continental shelf and in contact with the ice sheet in this area suggest it is a reasonable one.

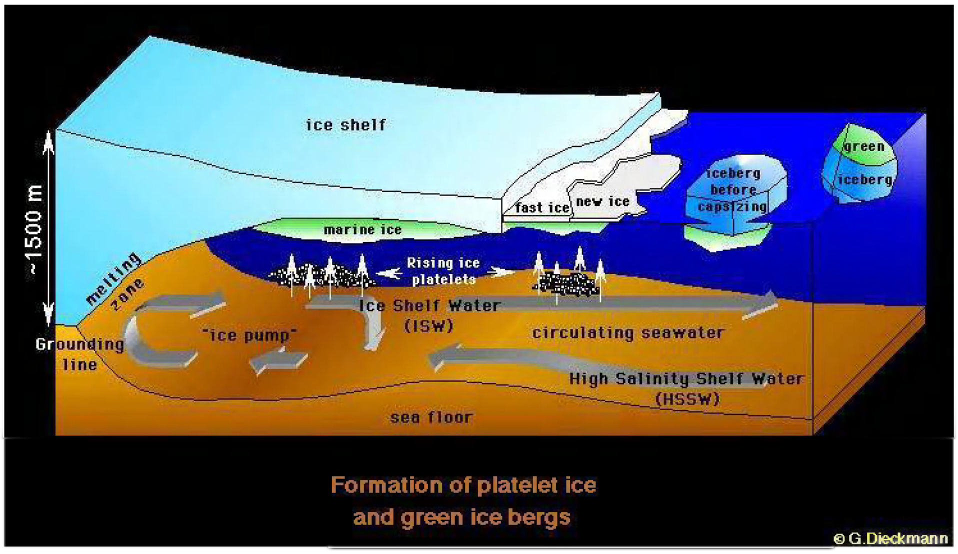


Figure 4.40 Water masses in the Antarctic coastal zone. (Dieckmann, unpublished)

4.8.3.2 Ross Sea Embayment

Elsewhere within West Antarctica, the changes are not as extreme. Among the ice streams feeding the Ross Ice Shelf, there is a rich history of change on millennial and shorter time scales. A major event approximately 150 years ago was the stagnation of Kamb Ice Stream (formerly ice stream C) (Retzlaff and Bentley, 1993). Since that time, ice upstream of the stagnated trunk has been thickening at a rate of nearly 50 cm/yr over an area tens of kilometres across. The next largest change is the gradual deceleration of the Whillans Ice Stream, immediately south of Kamb Ice Stream, at rates of between 1 and 2% annually (Joughin et al., 2002).

Aside from these two phenomena, the remainder of the ice flow in the region appears to be near equilibrium. Overall, Whillans and Kamb ice streams skew the cumulative mass balance calculations in the region to a net positive, indicating slight growth. An earlier estimate of 26.8 $\pm$ 14.9 Gt/yr by Joughin and Tulaczyk (2002) has only been slightly modified to 34 $\pm$ 8 Gt/yr recently by Rignot et al. (2008a), but the errors overlap, indicating consistency. The ocean-ice system of the Ross Sea is shown in Figure 4.41.

4.8.3.3 The Weddell Sea Embayment

This final third of the WAIS is about equal in size to the Amundsen and Ross Sea sectors, but appears to be more stable, at least for the past millennium. The ice streams are deeper than within the other sectors, but show few signs of flow rates or directions far out of the present equilibrium. The most recent calculation of its mass balance of -4 $\pm$ 14 Gt/yr (Rignot et al.,

2008) varies insignificantly from an earlier calculation of +9 ± 8 Gt/yr by Rignot and Thomas (2002).

Satellite altimeter records suggest, that there may be some areas within this sector (e.g. Rutford Ice Stream) where, in the last decade, there has been an excess of snow accumulation, although such records are too short to imply any likely ongoing change (Wingham et al, 2006a).

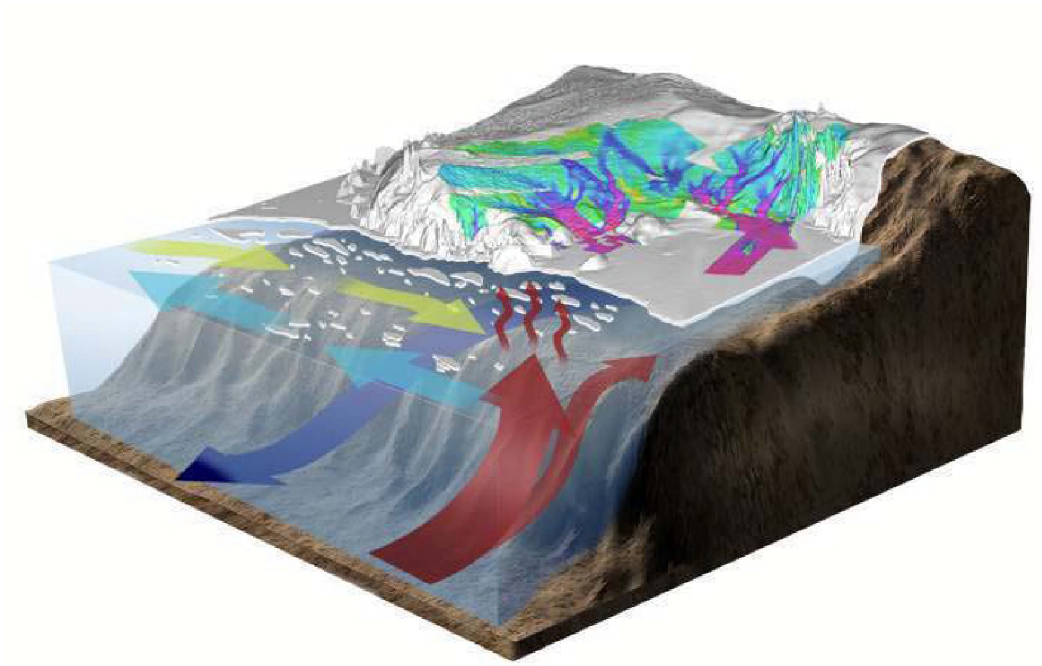


Figure 4.41 Pictoral view of the ocean-ice system of the Ross Sea. Colours on the ice sheet indicate flow speed with speed increasing from yellow to green to blue to magenta (see Figure 1.6). Colour in the ocean represents the major currents: light blue and yellow are the surface flows of the Antarctic Circumpolar Current and the opposing boundary current, respectively; red is CDW upwelling to reach the continental shelf and dark blue is the sinking Ice Shelf Water exiting from beneath the floating ice shelf. As CDW rises, some loses heat to the atmosphere (wavy vertical line) while the remainder circulated under the floating ice shelf causing basal melting. (illustration courtesy of National Geographic)

4.8.4 East Antarctica

Changes are less dramatic across most of the East Antarctic ice sheet with the most significant changes concentrated close to the coast. Increasing coastal melt is suggested by some recent passive microwave data (Tedesco, 2008). Satellite altimetry data indicate recent thickening in the interior that has been attributed to increased snowfall likely because of year-to-year and decade-to-decade fluctuations in snowfall (Davis et al., 2005), but ice core data do not show recent accumulation changes as significantly higher than during the past 50 years (Monaghan et al., 2006a). A resolution of this apparently conflicting evidence may be that there is a long-term imbalance in this area, which could possibly reflect a response to much more ancient climate changes. An alternate suggestion, based on direct accumulation measurements at South Pole, is that this thickening represents a short period of increased snowfall between 1992 and 2000 (Thompson and Solomon, 2002). The absence of significant atmospheric warming inland, distinct from the global trend of warming atmospheric temperatures, may have forestalled an anticipated increase in snowfall associated with the global trend.

The only significant exceptions to this broad-scale quiescence of the East Antarctic ice sheet occur on the Cook Ice Shelf and in the mouth of the Totten Glacier where thinning rates in excess of 25 cm/year have been measured (Shepherd and Wingham, 2007). It remains unknown whether these events are recent, or indeed, whether they are related to changing adjacent ocean conditions, as in the case of the Amundsen Sea outlets, or whether they are just longer-term responses of a regional dynamic origin. Both these areas are the outlets of the ice sheet occupying the two major marine basins lying beneath the ice sheet (Lythe et al., 2001).

The mass balance of the East Antarctic ice sheet has been calculated by many research teams with various sensors and methodologies: +22 $\pm$ 23 Gt/yr (Rignot and Thomas, 2002); -4 $\pm$ 61 (Rignot et al., 2008); 0 $\pm$ 56 (Velicogna and Wahr, 2006); and +15.1 $\pm$ 10.7 (Zwally et al., 2005). The results range from near zero to slightly positive with some of the variations dependent on the time interval investigated. One of the most significant factors giving rise to this uncertainty is that, at present, an *ad hoc* interpretation of the thickness changes must be made to determine whether they represent changes in snow surface accumulation, and thus changes in low-density snow and firn, or whether they are dynamic in origin and represent a change in ice, which has a much higher density.

4.8.5 Calving

Aside from the catastrophic ice shelf disintegration events already discussed, available data suggest that major rift-driven calving events have neither increased nor decreased on the major ice shelves (the Ross, Flichner-Ronne, or Amery). Rather there is ample evidence that their calving patterns continue to follow quasi-repetitive patterns extending back to the Nineteenth Century, when the ice fronts were first mapped (e.g. Jacobs et al., 1986; Keys et al, 1990; Lazzara et al., 1999; Budd, 1966; Fricker et al., 2005; see also Frezzotti and Polizzi, 2002, and Kim et al., 2007). So while the periodic calving of massive icebergs that appear to represent, in some cases, many decades of ice shelf advance, may appear dramatic, there is no reason to believe that they are not part of the normal fluctuations in a ice sheet that is, in the long-term, close to equilibrium.

4.8.6 Sub-glacial Water Movement

Regional surface elevation changes confined to areas of a few kilometres have been interpreted as manifestations of subglacial water movements (Gray et al., 2005; Wingham et al., 2006b; Fricker et al., 2007). These observations are interpreted to reflect the activity of a subglacial hydrologic system permitting faster ice flow that is more active than previously thought (Figure 4.42). While those earlier studies lacked simultaneous measurements of ice flow to accompany these likely shifts in water mass, a recent study of Byrd Glacier identified a period of 10% faster flow that fell within a period where the nearby subglacial lakes discharged water that probably exited the system by traveling underneath Byrd Glacier (Stearns et al., 2008). Improvements to numerical ice flow models are including subglacial water, but the specific nature of its role in ice sheet dynamics remains undetermined.

4.8.7 Other changes in the ice sheet

Much attention is focused on accelerating changes and instability, yet sediment deposited beneath fast moving outlet glaciers might provide some stability to ice sheet retreat driven by rising sea level. Recent observations and modeling suggest that wedges of sediment deposited near the grounding line may be important in stabilizing the ice sheet against sea level rise (Anandakrishnan et al. 2007; Alley et al, 2007). This stabilization is conditional on the position of the grounding line with respect to the crest of the wedge: should the grounding line retreat from the sediment wedge an unstable retreat analogous to that seen in tidewater glaciers could occur (Weertman, 1974; Schoof, 2007). It is becoming clear that rates of erosion and sediment transport can be large, as shown by the appearance of drumlin-scale sedimentary features underneath the ice sheet within just a few years (Smith et al., 2007a). Predictive models of the ice sheet will need to include sediment transport, as well as forces imposed by ice shelves, since these effects may compete to determine the stability or instability of the ice sheet margin.

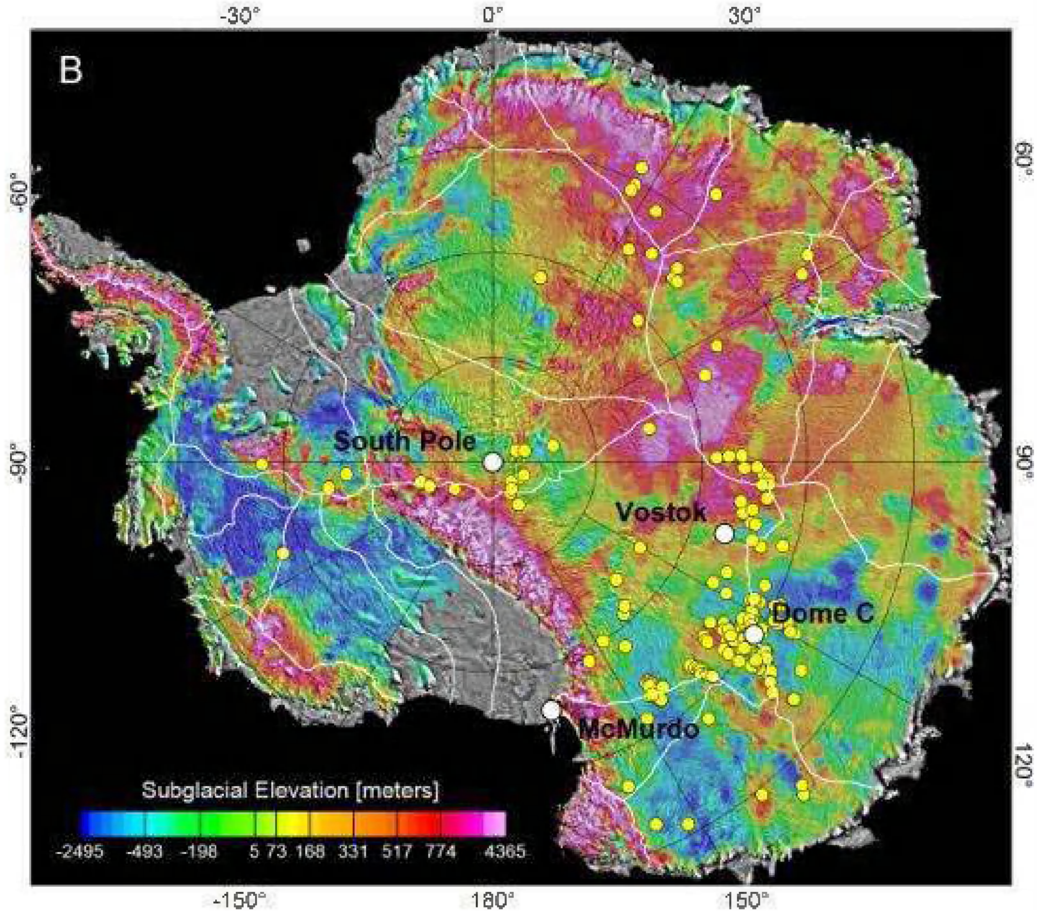


Figure 4.42 Locations of Antarctic sub-glacial lakes are indicated by the yellow circles. Over 145 subglacial lakes have been discovered with the majority clustered in the Dome C area of East Antarctica. Colours represent subglacial elevation. There is no clear correspondence between the subglacial topography and subglacial lake occurrence.

4.8.8 Attribution of changes to the ice sheet

The Antarctic ice sheet is known to respond slowly to large and sustained climate changes, but the new awareness that it can also respond rapidly to other changes makes it difficult to attribute a particular change in the ice sheet to a particular causal event or events such as recent/anthropogenic climate change. The inescapable fact is that ice sheet behaviour manifests itself as the superposition of multiple responses on multiple time scales to multiple environmental changes.

Further complicating the situation is that we do not know the changes that were occurring in the ice sheet prior to the period of satellite observations (which in the case of ice sheets began in earnest in 1992), let alone a century ago, and we have very little knowledge as to whether natural changes in the ice sheet over past millennia were smooth or step-wise and abrupt. The context for current changes must be understood by inference. Caution must be used in extrapolating current changes into the future.

The two areas of most rapid glaciological change are the Antarctic Peninsula ice shelves and the Amundsen Sea sector outlet glaciers. A possible trigger for the sudden collapse of ice shelves in the Antarctic Peninsula is pressurisation of crevasses filled with surface meltwater (Weertman, 1973; Scambos et al., 2000). The supply of meltwater increases with surface warming, placing a limit on the regions where ice shelves are viable (Vaughan and Doake, 1996). Some of the observed surface warming in the northern Antarctic Peninsula can be attributed to changes in global circulation, specifically to changes in the SAM (Marshall et al., 2006). The SAM exhibits considerable decadal variability, but 40-year trends similar to those observed are reproduced in global climate models forced by a combination of ozone depletion, and increasing greenhouse gas concentrations (Arblaster and Meehl, 2006). The stronger circumpolar westerlies bring warmer, northwesterly winds across the northern Antarctic Peninsula. In the eastern Peninsula, this trend is further amplified because northwesterly winds lead to downslope Föhn winds across the ice shelves, promoting warmer winter temperatures and longer melt seasons in summer (e.g. Van den Broeke, 2005). The loss of sea ice cover in the Amundsen and Bellingshausen seas (Jacobs and Comiso, 1997) may also affect the climate of the Antarctic Peninsula, and the viability of its ice shelves. Thinning from enhanced melting by ocean heat supply at the base of the ice shelf (Shepherd et al., 2004), or softening at the margins (Vieli et al., 2007; Khazendar et al., 2007), may have weakened ice shelves and predisposed them to collapse. While they are present, the ice shelves impart forces on the grounded glaciers that drain into them. Once ice shelves have disintegrated, these forces are removed, and the grounded glaciers accelerate (Rott et al., 2002; De Angelis and Skvarca, 2003; Rignot et al., 2004a) and start to thin (Scambos et al., 2004). This increases their discharge of ice into the ocean, and contributes to sea level rise.

The cause of the glaciological changes in the Amundsen Sea embayment is still an open topic of research. Ice sheet models have been used to show that changes similar to those observed can be caused by loss of basal friction in a small part of the ice sheet near the grounding line, perhaps caused by floatation as the grounding line retreats (Payne et al., 2004; Thomas et al., 2004a). Thinning of ice shelves by basal melt (Walker et al., 2007), softening of their margins (Vieli et al., 2007; Khazendar et al., 2007), or shortening by iceberg-calving (Dupont and Alley, 2005a) would also affect the force imposed on upstream glaciers. The ice shelves in the Amundsen Sea Embayment have been thinning (Shepherd et al., 2004; Bindschadler, 2002), and the grounding line of Pine Island Glacier has retreated as sections have thinned and gone afloat (Rignot, 1998b). This thinning, together with the observation that many independent ice streams are behaving similarly, has been taken to imply that changes in the heat supplied from the ocean are responsible for glaciological change in this sector (Payne et al., 2004). One mechanism for a variable supply of heat is episodic delivery of relatively warm CDW onto the continental shelf (Dinniman and Klink, 2004). CDW water occupies a deep layer, below the continental shelf break (700 to 1,100 m), but can be induced to upwell onto the continental shelf, especially via troughs (Walker et al., 2007), where it becomes available for basal ice shelf melting. Water layers in the depth of 700 to 1100 m are observed to be warming significantly further off shore (Gille, 2002), and there is evidence from salinity and other measurements over the past 40 years that increased melting is resulting from heat supplied by warmer CDW to northern West Antarctic ice shelves (Jacobs et al., 2002). According to coupled ocean atmosphere models, the upwelling of CDW is partly controlled by atmospheric circulation patterns (Hall and Visbeck, 2002). There is some correspondence between periods of enhanced heat supply predicted by a coupled model, and periods of observed glacier acceleration (Thoma et al., 2008). Direct observation of temporal changes in the delivery of circumpolar deep water to the ice shelf, and of the consequences for inland ice sheet flow are needed to test such models. To attribute the glaciological changes in the Amundsen Sea sector to a particular climate forcing will require a better understanding of the variability in ice sheet flow, how that flow is influenced by melting beneath ice shelves, and how the oceanic heat delivered to the ice shelves can change under the different atmospheric circulation patterns produced by various scenarios of radiative forcing.

4.8.9 Conclusions regarding the ice sheet

The Antarctic ice sheet is not behaving in a uniform manner – this is not surprising considering its enormous size, but the complexity and disparity of responses between different areas that has become observable in recent years might have been considered unlikely even a decade ago. The geographic extent of the ice sheet places different parts in markedly different positions within the global climate system and subjects them to different environmental drivers.

The most isolated portion is the East Antarctic ice sheet, primarily an extremely cold, high elevation plateau of ice, difficult for moisture-laden storms to reach. Recent atmosphere warming, pervasive throughout the rest of the planet, has not yet arrived, but slight increases in ice thickness are underway, likely an ongoing response on a centennial or millennial time scale to much older changes in climate.

Around the edges of the Antarctic ice sheet, the current state of the ocean is influencing the Amundsen Sea sector of West Antarctica and is likely also responsible for similarly behaving, but smaller, regions of East Antarctica. Here, the ocean appears to be the primary driving force, thinning the narrow fringing ice shelves, leading to rapid thinning and acceleration of the grounded ice. What happens at depth in the ocean matters to the ice sheet, and is itself strongly determined by the overlying atmospheric circulation, highlighting the complex climate interactions in this region. Other sectors of the West Antarctic ice sheet also contain fast moving ice streams, but aside from the now-stagnant Kamb Ice Stream and the decelerating Whillans Ice Stream, their current behaviour is far less extreme.

Ice on the Antarctic Peninsula is behaving quite differently from the rest of the continental ice sheet, in that it is engaged in an active interaction with a currently warming climate. Its north-south topography is the only barrier to the east-west atmospheric and oceanic circulation at these latitudes. Here, high rates of snow accumulation and melting drive a more vigorous glaciological regime. Recent observations have captured the sudden, and very likely recently induced, succession of ice shelf disintegrations followed by the dramatic acceleration of the glaciers that fed them. Perhaps more than any other single phenomenon, these events heighten concern about the near-future impact on global sea level of change in much larger ice reservoirs.

The attribution of ice loss on the Antarctica Peninsula to human-driven warming is now strong, and although not yet proved conclusively, there is a strong hypothesis that a similar case can be made for West Antarctic thinning. The thickening on parts of East Antarctica is an expected consequence of climate change, but it is not yet possible to make a satisfactory attribution of the changes there to any specifically observed climate change.

4.8.10 Changes in Antarctic permafrost and active layer over the last 50 years

4.8.10.1 Introduction

Permafrost temperatures and active layer depth are sensitive indicators of climate because they integrate different climatic factors (i.e. air temperature, seasonal snow cover, wind), which interact with each other and with the ground surface characteristics (i.e. vegetation, surface microrelief).

Permafrost temperatures and active layer thickness respond to the climate variations at different time scales because the permafrost thermal regime reacts: a) seasonally above the depth of zero annual amplitude (ZAA), b) annually at the ZAA, and c) from years to millenia at progressively greater depth. The active layer thickness responds seasonally to the climate input. Different methods are needed to monitor the permafrost thermal regime and the active layer thickness.

4.8.10.2 Changes in the last 50 years

Permafrost monitoring in Antarctica is a relatively new topic, although monitoring began in the 1960s at Signy Island in the South Orkney Islands.

More recently, new data were obtained on the thermal active layer (Cannone et al., 2006; Guglielmin et al., 2007). Comparing the new data with those collected at the same location four decades earlier, Cannone et al. (2006) show that the active layer thickness increased around 30 cm over the period 1963–90 (a period of warming on Signy Island), but then decreased by the same amount over the period 1990–2001 when Signy Island had a series of particularly cold winters.

The site of Boulder Clay (McMurdo Sound) represents the longest and most continuous data series of permafrost and active layer temperature (Guglielmin, 2004; 2006). Figure 4.43 shows the temperature recorded near the permafrost table (at a depth of 30 cm) and at the end of the borehole (360 cm deeep).

The permafrost temperature is stable at 360 cm, while at the permafrost table it shows a slight decrease of 0.1°C/year (Figure 4.43). This slight decrease (Guglielmin, In Prep.) is mainly related to the decrease of the air temperature and the decrease of the snow cover in the winter.

The decrease of ground surface temperature in relation to the decrease of surface air temperature confirms the pattern for the Dry Valleys by Doran et al. (2002) for the period 1986 to 1999 at Lake Hoare.

Five km north of Boulder Clay, at the MZS station, a borehole 15.5 m deep was drilled in bedrock in 1999 and monitored manually once a year and, since 2003, automatically all year round (Guglielmin, 2006). In the summer 2005/2006 a new borehole 30 m deep, just some metres away, was drilled as a template for the new IPY-ANTPAS monitoring network. The thermal profile obtained in the 30 m borehole (Figure 4.44) suggests at least two periods of cooling (around 14-15 m), following a previous period of warming. Guglielmin (2007, in prep) describes these quite short fluctuations of the ground surface temperature at MZS in the last 30 years.

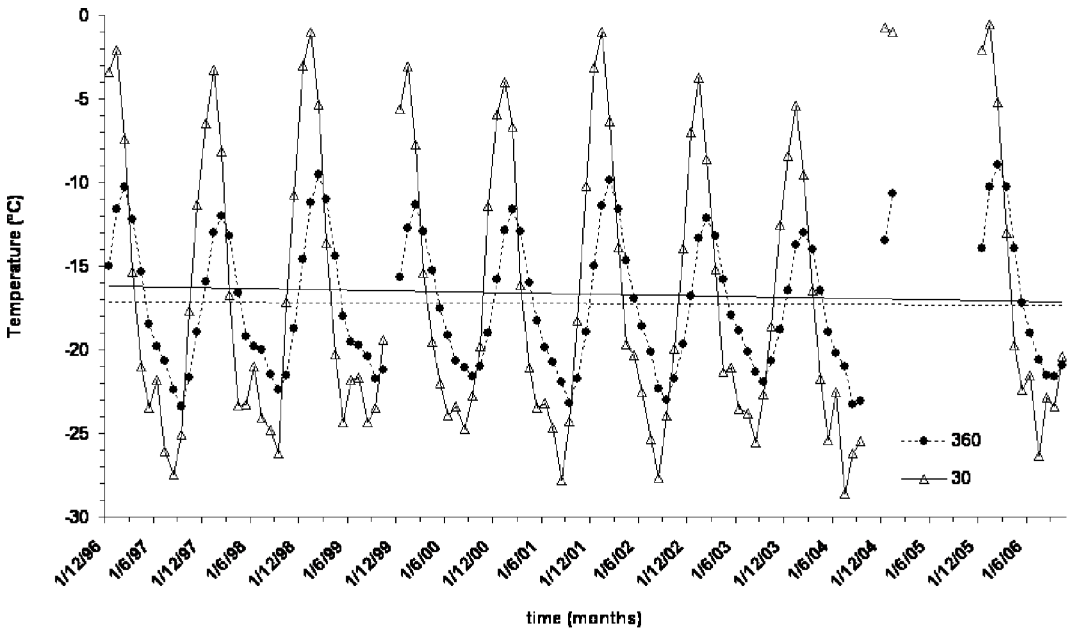


Figure 4.43 Monthly mean temperatures at depths of 30 and 360 cm at Boulder Clay since 1996. Note that the depth of 30 cm is very close to the permafrost table. The linear regression lines for 30 cm depth (solid line) and for 360 cm (dashed line) are also reported.

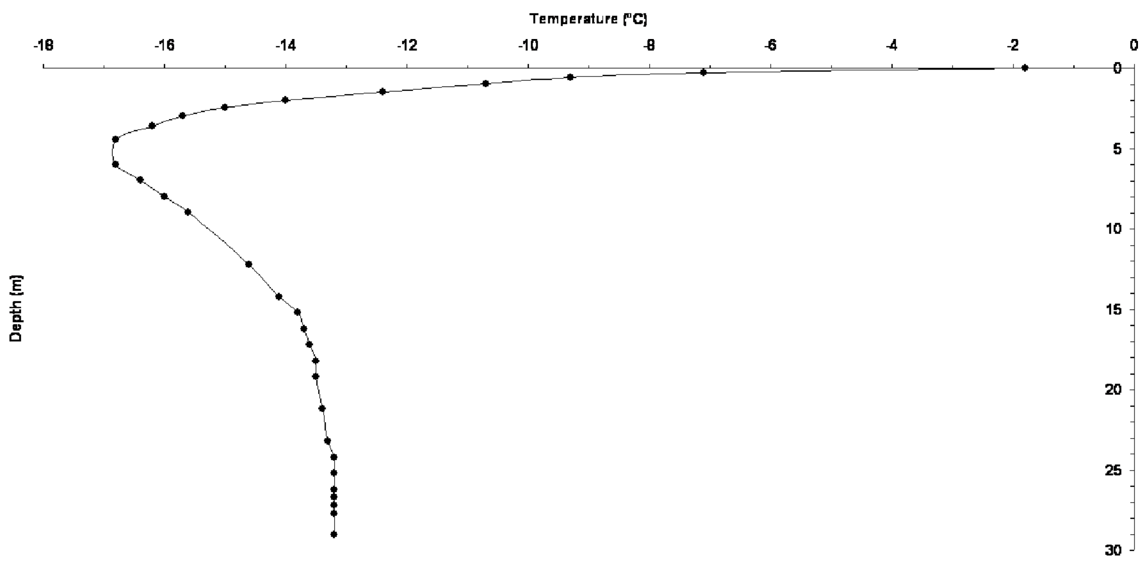


Figure 4.44 Permafrost profile recorded in the borehole at Oasi on 7 November 2006. The borehole was drilled through a homogenous granite outcrop on a gentle slope at 80 m a.s.l.

4.9 Long Term Sea Level Change

The first three assessment reports of the IPCC arrived at similar conclusions with regard to global sea level change during the Twentieth Century. For example, the third report (Church et al., 2001) concluded that global sea level had changed within a range of uncertainty of 1-2 mm/yr. Since then, there have been major workshops (e.g. World Climate Research Programme workshop on Sea Level Rise and Variability, Church et al., 2007), reviews by individual scientists (e.g. Woodworth et al., 2004), and, most recently, the publication of the ocean climate and sea level change chapter within the IPCC Fourth Assessment Report (Bindoff et al., 2007). A consensus seems to have been achieved that the Twentieth Century rise in global sea level was closer to 2 than 1 mm/year, with values around 1.7 mm/yr having been obtained for the second half of the last century in the most recent studies (e.g. Church et al., 2004; Holgate and Woodworth, 2004). However, it should be noted that the Antarctic contribution to sea level now is small compared to what it was following the LGM Transition and through the Holocene.

Fluctuations in the size of Antarctic and Greenland ice sheets during the glacial/interglacial cycles resulted in sea level variations of over 120 m. However, in spite of the enormous sea level-equivalent of the ice stored in the two ice sheets (Table 11.3 of Church et al., 2001), both seem to have played relatively minor roles in sea level change during the last two centuries. The major contributions to Twentieth Century sea level rise are believed to have originated from ocean thermal expansion and the melting of glaciers and ice caps. Antarctica's contribution appears to have been of the order of 0.1-0.2 mm/yr over the last few decades with some evidence for a slightly larger value in the 1990s (Bindoff et al., 2007).

The most recent data (i.e. from the 1990-2000s) from tide gauges and satellite altimeters suggest that global sea level is now rising at a rate of 3 mm/yr or more (e.g. Holgate and Woodworth, 2004; Beckley et al., 2007). The IPCC's 4th Assessment Report cites 3.1 mm/yr for 1993-2003 (IPCC, 2007). However, according to Cazenave et al., (2009), and based on GRACE and altimetric satellite data and Argo ocean float data from 2003-2008, the rate has slowed to 2.5 mm/yr; this reflects a significant slow down in the thermosteric component, balanced by an increase in ice contributions (half from mountain glaciers and half from ice sheets). This is still a higher rate than typical for the Twentieth Century. As pointed out by Milne (2009) the latest results are not the last word and we need longer time series to be confident in the magnitude of the trend.

Figure 4.45 shows a time series of annual mean sea level values from Vernadsky, suggesting an upward trend (uncorrected for local land movements) of 1.6 ± 0.4 mm/year, with a dip in the 1970s for which one has to be concerned about instrumental problems, and no evidence for recent acceleration. As an aside, one may note that observed Southern Hemisphere Twentieth Century sea level trends tend to be generally lower than Northern Hemisphere ones (e.g. see the long southern records studied by Hunter et al., 2003 and Woodworth et al., 2005).

4.10 Marine Biology

4.10.1 The open ocean system

The area covered by winter sea ice in the Southern Ocean has not changed significantly over the past decades suggesting that the impact of global warming on Antarctic ecosystems is not as severe as it is in the Arctic. There the sea ice cover is declining in both thickness and extent at a rapid rate, profoundly affecting the structure and functioning of Arctic marine ecosystems, particularly mammal and bird populations. For a comprehensive and most recent review with additional relevant literature see Nichol (2008). In the Antarctic, comparable shrinking of the winter ice cover has occurred only along the western side of the Peninsula and adjoining seas. This is a relatively small region but home to the well-known whale-krill-diatom food chain. Based on data shown in Figure 4.46, Atkinson et al. (2008) calculated that

70% of the total krill stock resides in the sector 0°-90° W, a region characterised by rapid regional changes both in water temperature (Meredith and King, 2005; Whitehouse et al., 2008) and winter sea ice cover (Parkinson, 2004).

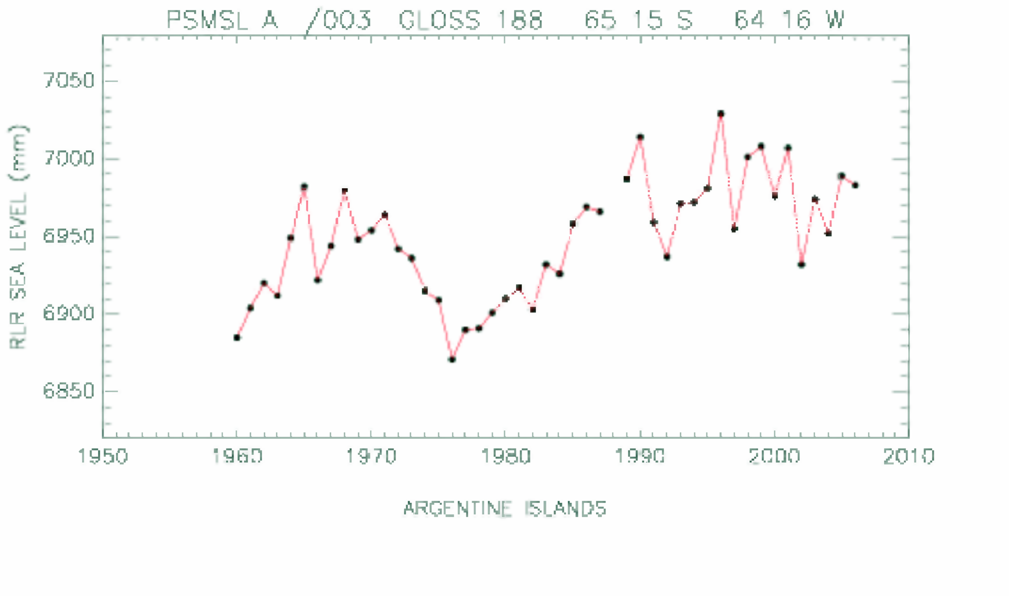


Figure 4.45 PSMSL Revised Local Reference (RLR) annual mean sea level time series for Vernadsky/Faraday (called Argentine Islands in the PSMSL data set)

Following near-extinction of the whale populations, the krill stock was expected to increase as a result of release from grazing pressure. Although predation pressure by seals and birds increased, the total biomass remained only a few percent of that of the former whale population. About 300,000 blue whales alone were killed within the span of a few decades equivalent to more than 30 million tonnes of biomass. Most of these whales were killed on their feeding grounds in the southwest Atlantic in an area of at most 2 million km^2 (10 % of the entire winter sea ice cover), which translates to a density of one blue whale per 6 km^2. Today's whale watchers would be thrilled. A 100 tonne blue whale (adults weigh 150 tonnes) contains about 10 tonnes of carbon, so the biomass of the whales on their feeding grounds would have amounted to 1.5 g C m^{-2} which is equivalent to the average coastal zooplankton biomass. Adding the biomass of krill estimated to have been annually eaten by the whales (150 million tonnes) to the m^2 calculation, we get 12 g C m^{-2} just for blue whales and their annual food intake. This number is equivalent to the biomass of an average phytoplankton bloom or, to take an example of more familiar grazers, to 240 cows of 500 kg each grazing on one km^2 of meadow.

The actual krill stock, from which the 150 million tonnes were being eaten, will have been at least three times higher prior to whaling. The magnitude of primary production required them to fulfil their food demands at the trophic transfer rule of thumb (10:1) would be around 300 g C m^{-2} yr^{-1} which is about that estimated for the North Sea, hence this does not leave much scope for other grazers such as protozoa and copepods. What percentage of the production was exported then from the surface through the mesopelagial habitat and ultimately to the deep-sea benthos, hence also sequestered as carbon, is an interesting but open question.

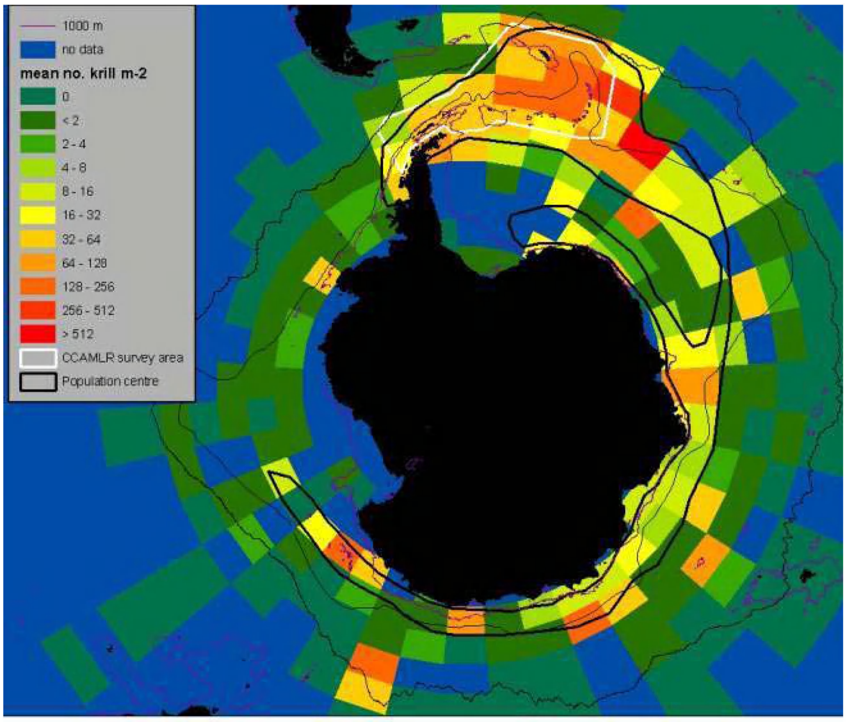


Figure 4.46 Circumpolar distribution of Antarctic krill, *Euphausia superba*. based on standardised data from KRILLBASE (8,789 stations). Black lines (from north to south) show Antarctic Polar Front and Southern Boundary of the Antarctic Circumpolar Current. Population centres drawn by eye, relative to the Commission for the Conservation of Antarctic Living Resources (CCAMLR) Survey (from Atkinson et al., 2008, © Atkinson et al., and InterResearch).

The above calculations indicate that krill stocks in the whale feeding grounds were close to the carrying capacity of the ecosystem prior to whaling. That being the case, when less krill was eaten by whales, enabling more krill to survive, many of the survivors would have starved. This would explain why a krill surplus, at least equivalent to the amount annually eaten by whales, was not recorded. At their former high stock sizes, and given the tendency of krill schools to appear at the very surface and discolour the water, they were commonly observed from ship decks, as noted for example by the scientists of the Discovery cruises (Hardy, 1967). Since that time, despite a significant increase in the numbers of observers, from cruise ships to research vessels, krill swarms are now rarely seen from ships decks. A thorough statistical assessment of all net catches has been carried out for different sectors of the Southern Ocean. The analysis suggests a 38 - 81% decline in krill stocks of the southwest Atlantic accompanied by an increase in salp populations (Figure 4.47, Atkinson et al., 2004; Ross et al., 2008). The extent of the krill decline and the underlying factors are under vigorous debate (Ainley et al., 2007; Nicol et al., 2007), because of difficulties in unravelling the effects of industrial whaling from those of sea ice retreat; there are also discrepancies between the abundances of krill as measured by net and acoustic methods, and enormous intra-annual as well as spatial variability has to be considered (Hewitt et al., 2003; Saunders et al., 2007). However, a significant negative correlation between krill density (30°W to 70°W) and mean sea surface temperature at South Georgia has been found for the period 1928-2003, which implies a large-scale response not only of krill but of the entire open ocean ecosystem to climate change (Whitehouse et al., 2008).

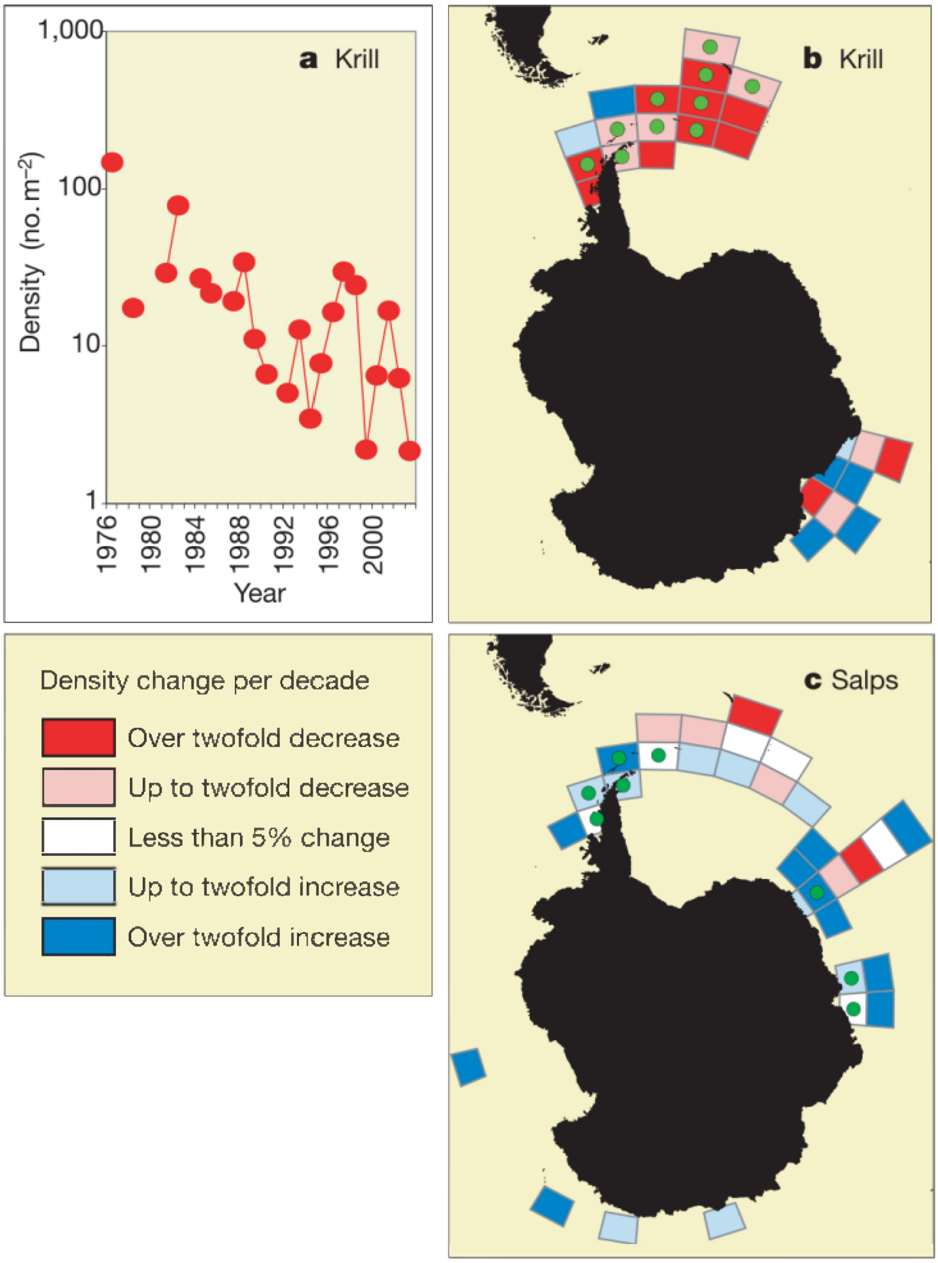


Figure 4.47 Temporal change of krill and salps. a, Krill density in the southwest Atlantic sector, 30° – 70°W. Illustrated temporal trends include b, post 1976 krill data from scientific trawls; c, 1926-2003 circumpolar salp data. Green spots denote grid cells useable in a spatio-temporal model, with data subdivided according to sampling method (Atkinson et al., 2004). This model revealed significant decreases in krill density within the southwest Atlantic sector since 1976 and a significant increase in salp densities at high latitudes since 1926. Reprinted by permission from Macmillan Publishers Ltd: Nature (Atkinson et al., 2004), © 2004

If former krill stocks were close to the carrying capacity provided by primary production, then a decrease in grazing pressure should have resulted in a "phytoplankton surplus", but there is also little evidence for that. Unfortunately, a comparison with phytoplankton stocks recorded during the Discovery era is not possible because the methods used at that time soon became obsolete. Nevertheless, the impression gained by the Discovery scientists is one of large diatom stocks: "...extremely rich production, which will probably be found to exceed that of any other large area in the world …" (Hart, 1934). The Discovery scientists were familiar with North Sea phytoplankton, which today has much higher biomass levels than those recorded for the Scotia Sea in recent decades. That comparison makes it likely that phytoplankton production has indeed decreased with that of krill stocks, a conclusion supported by the increase in the salp population. In contrast to krill, which are equipped to deal with the characteristically spiny and heavily silicified diatoms of the Southern Ocean, salps are adapted to feed on the lower biomass concentrations typical of the iron-limited microbial food webs. Their encroachment into the former krill habitat is an indication of declining phytoplankton, in particular diatom stocks.

A decline in phytoplankton concentrations can be explained by a corresponding decline in the supply of iron. There is reason to believe that the reduction in sea ice formation has resulted in a decrease in iron input from the continental margins of the Western Peninsula. In contrast, the simultaneous retreat of glaciers should have increased run-off, and possibly also iron input, from the land along the coasts of the Peninsula and adjacent islands. Comparisons of the chlorophyll concentrations recorded by the CZCS satellite of the 1980s with those from the current SeaWifs satellite indicate a decline along the Antarctic ice edge and particularly in the Scotia Sea, the only region of the globe where production has declined, but no major change along the coast (Gregg and Conkright, 2002). Production was found to have increased by 50%, off the Patagonian shelf, so it is also possible that the wind field transporting Patagonian dust from mud fields laid bare by retreating glaciers has changed, reducing the aeolian iron supply to the Scotia Sea. Also, the westerlies have intensified and would carry more dust. Whatever the mechanism, a reduction in phytoplankton biomass can only be explained by a corresponding reduction in iron supply combined with light limitation by deep mixed layers and heavy grazing pressure on phytoplankton stocks. Recently, five mesoscale, *in situ* iron fertilization experiments, carried out in the Pacific and southeast Atlantic Sectors of the Southern Ocean, have unambiguously demonstrated that plankton biomass is limited by iron availability. It follows that the higher productivity of coastal regions, including the southwest Atlantic, is maintained by input of iron supplied from land-masses, and from the sediments by deep mixing and upwelling along the continental margin. The presence of excess nutrients in these regions allows the assumption that iron supply limits productivity over most of the year throughout the Southern Ocean. The ramifications of this finding for the structure and functioning of Antarctic ecosystems have yet to be adequately explored, particularly because ongoing global change will affect coastal hydrography and hence the supply of iron.

An alternative but not mutually exclusive explanation for the phytoplankton decline can be a decrease in the rate of recycling of the iron entering the system. It is now well established that primary production in microbial food webs is based on recycling by grazers feeding on pico- and nano-phytoplankton, which are in turn eaten by predators such as ciliates and copepod larvae. The latter are the preferred prey of copepods, whereas filter-feeding salps consume all the components *en masse*. In the Southern Ocean the microbial community is characteristic of the iron-limited HNLC (high nutrient, low chlorophyll) area, where chlorophyll concentrations remain below 0.5 mg-Chl/m^2 throughout the year. These regions support a surprisingly high zooplankton biomass, comprising slow growing copepods and fast-growing salps, throughout the year, suggesting that they are an integral part of a recycling system that also regenerates iron in addition to ammonium (Barbeau et al., 1996).

Local increases in production above this level are invariably due to accumulation of diatoms and *Phaeocystis* colonies, and will be caused by input of new iron, whether from above or from below. The fate of these diatom blooms is under debate: are they consumed and their nutrients recycled in the surface or sub-surface layer, or does a significant portion sink to greater depths or to the sea floor? The latter fate is of particular interest in the light of proposals for large-scale ocean fertilization to sequester atmospheric CO_2.

Given the high densities of the former krill stocks, their rate of recycling will have been as effective as that in the microbial food web today, but on a much larger scale. Predation by whales will have contributed significantly to the iron recycling pool. That is because, except in the case of pregnant and lactating females, whales convert krill protein into blubber (hydrocarbons), so retaining energy but excreting nutrients, including iron. Whale faeces are liquid and rise to the surface where they are likely to have released iron, thereby increasing the efficiency of recycling. Krill also have a high (50%) lipid content, and krill excretion releases large amounts of iron (Tovar-Sanchez et al., 2007). It follows that the exceptionally productive ecosystem characterised by the food chain of the giants was maintained by the recycling of iron by krill and whale feeding. An alternative, but mutually inclusive hypothesis in which large whale stocks promoted the development of large stocks of their prey (krill) by dispersing them over a larger area has been recently suggested.

It is now acknowledged that large terrestrial herbivores (megafauna) condition ecosystems by promoting a vegetation cover conducive to their demands, e.g. grassland instead of forest by elephants. The removal of those herbivores leads to profound changes in landscape. Similarly, the top predators in lakes can determine the structure of the ecosystem down to the composition and biomass of the phytoplankton. Predation pressure on upper trophic levels is propagated down the food web by mechanisms known as trophic cascades. Although the effects of top-down control have been demonstrated for shallow benthic environments from many coastal regions, comparable mechanisms are only now coming to light from planktonic ecosystems, such as the reported worldwide increase in gelatinous plankton after removal of dominant fish stocks. Whether such changes can cascade down to the level of phytoplankton is not known. We simply do not know what effect the removal of whales (and seals) through hunting had on the ocean ecosystem around Antarctica. The belief that marine phytoplankton productivity is determined primarily by bottom-up driving forces is entrenched in the marine biological literature, but why the interaction between marine plankton and whales around Antarctica should be fundamentally different from that between their non-polar lake counterparts has yet to be addressed. The fact is that the processes driving annual cycles of phytoplankton production, biomass and species composition in the marine environment remain largely unknown. It is time to explore new approaches.

The linkages between phytoplankton and bacterioplankton in the Southern Ocean are not well understood. The timing of spring phytoplankton blooms and bacterioplankton activities are not necessarily linked, as they are in other pelagic systems, though there may just be a lag in this linkage that is longer than in lower latitude systems (Ducklow et al., 2007). Even less is known of the relationships between phytoplankton species and bacterioplankton species composition (e.g. whether particular bacteria are associated with diatom vs. cryptophyte phytoplankton). Considerations of phytoplankton primary productivity, linkages to CO_2 drawdown, and grazer populations are also linked to phytoplankton species composition. Phytoplankton species are susceptible to changes in sea ice duration and position of the ice edge, and shifts in water column properties such as the depth of the mixed layer. Shifts in phytoplankton species composition from diatom-dominated communities to more diverse communities dominated by cryptomonads and flagellates occurs following the ice edge retreat and spring diatom bloom in the Western Antarctic Peninsula (Moline and Prezelin, 1996), and will be (or already are) potentially more common occurrences as a result of warming in this region (Clarke et al., 2007). The linkages between phytoplankton and zooplankton populations are tight, as krill tend to dominate the zooplankton assemblages when diatoms are abundant, and zooplankton dominance can shift to salp dominance when the community is cryptomonad or flagellate dominated. One study in the Western Antarctic Peninsula recently reported that there has been a shift from krill-dominated waters to salp dominance since 1999 This trend may also be potentially representative of the longer term trends referred to above.

Acquiring a mechanistic understanding of the structure and functioning of the ecosystems surrounding Antarctica is a prerequisite for predicting their performance under the influence of global warming. Hypothetical conceptual frameworks of relevant mechanisms need to be developed that can be tested by comparing intact ecosystems with those where top-predators have been depleted both regionally and, where baseline data are available, temporally. Satellite data have vastly extended the scales accessible to such regional studies. Larger scale in situ iron fertilization experiments open up an exciting new avenue to study the effects of bottom-up versus top-down factors on higher trophic levels, and if carried out over several years, also on krill populations and on the underlying deep sea and benthos. Such experiments provide an ideal background to study the relationship between ecology and biogeochemistry at the species level, which in turn will improve interpretation of sedimentary proxies, in particular microfossils, for reconstruction of past climate change. Conceptual frameworks emerging from field studies and experiments can be explored, tested and refined with new generations of 4D mathematical models.

4.10.1.1 Iron fertilization experiments

It has been suggested that one way in which the rise of carbon dioxide in the atmosphere may be mitigated is to fertilise the ocean with iron so as to stimulate the production of plankton and hence the draw-down of carbon dioxide from the atmosphere into the ocean (Boyd et al., 2007a). These ideas are based on the results of a limited number of experiments in which different parts of the ocean, including the Southern Ocean, were seeded with iron (Boyd et al., 2007b). This current debate (e.g. see Oceanus, 24 June 2009, http://www.whoi.edu/oceanus/viewArticle.do?id=34167) may at some time shift its focus to exploring how to maximise the efficiency of the process and minimise harmful side effects. The hypothesis could be tested by a new generation of iron fertilization experiments carried out at larger scales and longer periods on the former whale feeding grounds. Given the apparent high rates of krill decline and a steady southward encroaching ocean warming there is a pressing need to develop an integrated understanding of how this ecosystem functioned not only in the recent past but also in the glacial ocean, in order to predict future changes in the pelagic and underlying benthic ecosystems around Antarctica. In situ iron fertilization experiments provide one methodology for testing ecosystem models, by enabling the study of interactions within ecosystems with a full complement of grazers and pathogens. The effect of iron fertilisation on higher trophic levels will depend on the locality and duration of the experiment. A regional survey of the underlying benthos prior to fertilization would yield a baseline to monitor possible changes in this ecosystem. Preliminary surveys of the deep-sea benthos of the Peninsula region have shown the presence of communities with high biomass and species numbers (Brandt et al., 2007) but their areal extent is not known. Deep carbon export flux has been shown to be above global average and to have a high regional variability in the Southern Ocean (Boyd and Trull, 2007; Sachs, 2008). Since fertilization will be carried out offshore, it is quite unlikely that shelf and coastal benthos are significantly affected.

An added incentive to carrying out such experiments is that they would offer an ideal training ground for the kind of large-scale international, interdisciplinary research taking a whole Earth System Science approach to investigating global change. Such large-scale experiments would not only provide a wealth of new insights into the structure and functioning of pelagic and underlying benthic ecosystems. They could also provide more reliable data for parameterising current and new coupled ecological-biogeochemical ocean-circulation models for use in assessing the Southern Ocean as a sink for anthropogenic CO_2. There is a concern that the incentive offered by the carbon credit market could result in excessive fertilization which could lead to unacceptable harm to Southern Ocean ecosystems (Chisholm et al., 2001). Three United Nation bodies, the Intergovernmental Oceanographic

Commission (IOC), the International Maritime Organization (IMO), and the Convention on Biological Diversity (CBD) agreed that proposals to use ocean fertilization to sequester carbon in the ocean give cause for concern due to unknown negative impact to the ecosystems (http://ioc3.unesco.org/oanet/OAdocs/INF1247-1.pdf). The two latter organizations recently argued that large operations are currently not justified and should not be allowed (www.cbd.int/decisions/?m=COP-09&id=11659&lg=0, www.maritime-connector.com/NewsDetails/2203/lang/.wshtml, www.ioccp.org). Scientists emphasized the necessity for independent research on small scale fertilisation studies (Buesseler et al., 2008). In addition it must be considered that once done such large global experiments with unknown outcomes would be very difficult or impossible to reverse.

4.10.2. Sea ice ecosystems

It is now clear that decreasing sea ice cover in the southwest Atlantic is due to a decline in ice production along the Peninsula. This can significantly reduce the supply of iron to the surface layer. The formation and presence of sea ice has several effects on the underlying water column and benthos. Deep convection due to brine discharge during ice formation can mix the entire water column down to the sea floor. Downward transported organic particles from the productive surface layer thus become available to benthic filter feeders. This mechanism will be a major source of food supply to the sponge-dominated fauna of Antarctic shelves and explains the apparent lack of gearing of reproduction of the shelf benthos to the ice–free period, when vertical particle flux from the overlying water column is at its maximum. Upward mixing of water that has contacted the sediment surface will bring iron to the surface layer. This mechanism of convective upward iron transport and subsequent fuelling of surface phytoplankton blooms has been reported from the Peninsula and around islands with shallow shelves where convective winter mixing is sufficient to reach the sediment surface without ice formation. It follows that the ongoing retreat of winter sea ice along the western Peninsula will result in declining depths of winter mixing and hence also in the supply of iron to the water column overlying deeper shelves, where surface warming reduces ice formation. A decrease in downstream spring productivity can be expected, which might be a factor contributing to krill decline in this region. However, it cannot be the only factor because the krill decline began before the retreat of the sea ice.

4.10.3 ENSO links and teleconnections to vertebrate life histories and population

Biological impacts with consequences for upper trophic levels in the southwest Atlantic region occur at the same time as the rapid warming observed to the west of the Antarctic Peninsula, in the Amundsen Sea and across the Bellingshausen Sea, and to the east of the Peninsula across the northwestern Weddell Sea, where the sea ice season has decreased (Parkinson, 2004). Food webs in these regions are dominated by euphausids, and in particular by Antarctic krill, *Euphausia superba*, which provide ecosystem structure and function, supporting the energetic demands of abundant predator populations (Croxall et al., 1988) and also a commercial fishery. As shown above, the relative density of krill across the region has shown a significant negative trend over the last 30 years, during which summer mean krill density correleated positively with the duration of sea ice the previous winter (Atkinson et al., 2004).

The strong connections between sea ice and ENSO variability across the southwest Atlantic result in correlations between ENSO variation and krill recruitment and abundance; quasi-cyclic sea ice variation is related to cycles in krill population (Fraser and Hofmann, 2003; Quetin and Ross, 2003). Recent studies have shown how ENSO related fluctuations in SST and winter sea ice extent affect the recruitment and dispersal of Antarctic krill. Delayed effects of sea ice conditions on abundance relate to the production, survival and development of the larval and juvenile krill. Same-year effects of spring and summer temperatures probably reflect a distribution and dispersal effect across the Scotia Sea and the degree of influence of cooler polar waters in northern regions around South Georgia (Murphy et al., 1998, 2007).

These effects cascade, bottom-up, to upper trophic levels. With few exceptions, periods of reduced predator breeding performance in this region are the result of low prey availability rather than direct local weather or oceanic effects (Croxall et al., 1988; Fraser and Hofmann, 2003; Forcada et al., 2005, 2006; Trathan et al., 2006; Hinke et al., 2007; Murphy et al., 2007).

For Adélie penguins breeding on the western Antarctic Peninsula, changes in sea ice extent during the breeding season covary positively with krill abundance and negatively with penguin foraging effort (Fraser and Hofmann, 2003). In the South Shetland Islands, krill recruitment covaries positively with Adélie penguin recruitment, which has declined dramatically (Hinke et al., 2007). The population trends of Adélie and Chinstrap penguins appear to be affected by a winter krill deficit. This deficit also affects the Gentoo penguins, but their population trends are positive (Figure 4.48, Ducklow et al., 2007). Similar results on population trends at the South Orkneys suggest the loss of buffering against the changing sea ice environment by the more abundant and ice-dependent Chinstrap and Adélie penguins, and positive population consequences through habitat improvement for the less ice-related Gentoo penguin (Forcada et al., 2006).

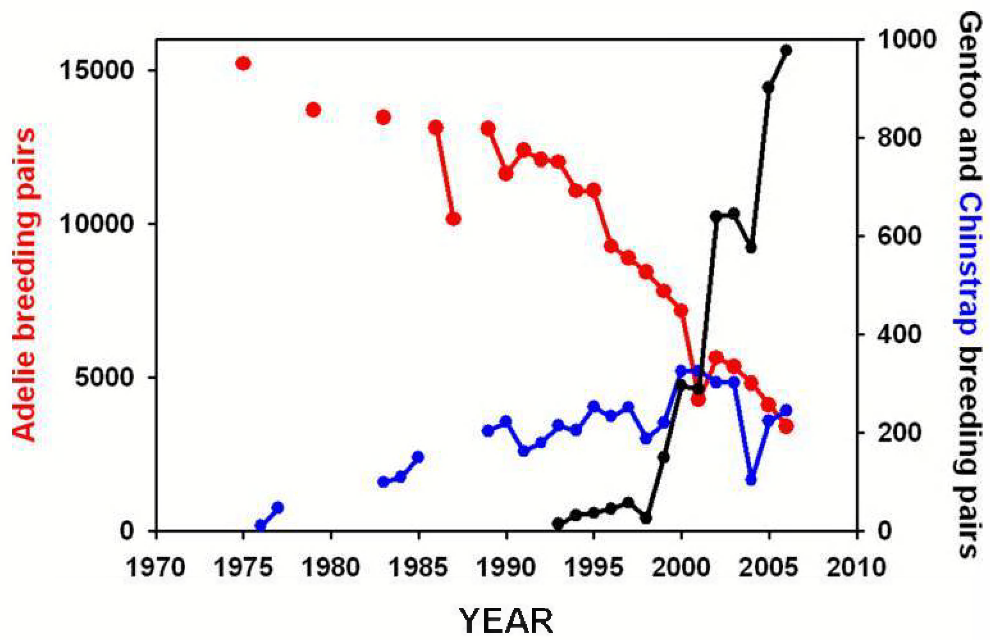


Figure 4.48 Population trends for three penguin species in the Anvers Island vicinity (West Antarctic Peninsula), 1975-2003, 1975-2006. Graph courtesy of W.R. Fraser and Palmer LTER, see also Ducklow et al., 2007)

Other krill dependent predators have shown comparable responses to the propagation of ENSO variability. Breeding outputs of Gentoo penguins and Antarctic fur seals at Bird Island, South Georgia (Forcada et al., 2005; Trathan et al., 2006) and southern right whales (*Eubaleana australis*), which feed in the waters around South Georgia (Leaper et al., 2006), correlate negatively with SST anomalies. Low chick and pup production and survival in penguins and fur seals result from a reduced food supply, usually of krill and krill-dependent fish species, such as mackerel icefish.

Signals of climate change related to ENSO variability are also observed elsewhere in the Southern Ocean. However, the consequences for ecosystem fluctuation are less well understood. In the Indian Ocean, and in particular in regions of the Antarctic continent, long-term reductions in sea ice extent have occurred in parallel with reductions of the Southern Oscillation Index (and an opposite trend in ENSO) (Barbraud and Weimerskirch, 2006). Long-term increases in sea ice season duration have possibly reduced the quantity and accessibility of the food supplies available in early spring for upper trophic levels. This would partly explain the delays in phenology observed in a guild of seabird species (Barbraud and Weimerskirch, 2006). While little is known about the marine ecosystem structure and food web interactions, there have been major fluctuations in the late 1970s suggesting a regime shift with repercussions for lower trophic levels (Hunt et al., 2001), but also for upper trophic levels, and in particular Emperor penguins, which suffered a major population decline (Weimerskirch et al., 2003). Long-term environmental effects in breeding success were further explored in long-term data sets of seabirds, including southern fulmars, snow petrels and emperor penguins in Terre Adélie. The observed population fluctuation in these species had a periodicity of 3-5 years, consistent with that in sea ice anomalies and the SOI (and hence ENSO). The observed cyclical patterns also indicated significant change since 1980, consistent with a regime shift, although these populations have remained stable since then.

In the Antarctic sector of the Pacific Ocean, and in particular the Ross Sea sector, trends in surface temperatures are less apparently correlated with ENSO, although there has been a decrease in the SOI. The impacts may be less evident or may even be the reverse of those observed in the Atlantic and the Indian Oceans. Annual estimates of breeding population size for Adélie penguin colonies on Ross Island, in the Ross Sea, show significant inter-annual variability, consistent with ENSO and the propagation of its variability through the ACW (Wilson et al., 2001; Ainley, 2005). Lagged correlations with a sea ice index and the SOI (ENSO) suggest demographic cycles, with reverse population trends to those observed in the Southwest Atlantic. However the consequences for Adélie penguins changed between the 1970s and the 1980s, with a concurrent increase in population trend and in the sea ice polynya providing improved penguin habitat. Like the Emperor Penguins in Terre Adélie, in the Indian Ocean, populations of Weddell seals at McMurdo Sound, in the Ross Sea, declined in the early 1970s, to become stable later on (Ainley et al., 2005). This suggests that ENSO impacts have not been so obvious in the Pacific sector as in the Atlantic Ocean and Indian Ocean sectors of the Southern Ocean. Based on long-term studies, Dayton (1989) and Barry and Dayton (1988) speculated about ENSO-related hydrodynamic processes supporting the explosive growth of fast growing sponges in McMurdo Sound with significant consequences for the entire benthic assemblage.

4.10.4 Invertebrate physiology

Decadal-scale variations in the coupled ocean-atmosphere system have an impact on animal communities and populations in marine ecosystems (Cushing, 1982; Beamish, 1995; Bakun, 1996; Finney et al., 2002). Present-day effects of global warming on the biosphere are associated with shifts in the geographical distribution of ectothermic (cold-blooded) animals along a latitudinal cline or with poleward or high-altitude extensions of geographic species ranges (Walther et al., 2002; Parmesan and Yohe, 2003; Root et al., 2003). Temperature means and variability associated with the climate regime can be interpreted as major driving forces on the large scale biogeography of marine water breathing animals. These relationships lead us to expect that climate warming will have a significant impact on ecosystems at both poles.

The marine Antarctic has cold temperatures that are close to freezing in several areas, and which have the lowest temperature variability at high latitudes (Clarke, 1998; Peck, 2005). Thus, Antarctic marine ectotherms live at the low end of the aquatic temperature continuum and within a narrow temperature window (making them highly stenothermal) (Somero and De Vries, 1967; Peck and Conway, 2000; Somero et al., 1996, 1998; Pörtner et al., 1999, 2000; Peck et al. 2002, 2006). Climate-induced changes in mean temperature and its variability should influence the survival of such organisms. That begs the questions – (i) to what extent has stenothermy in Antarctic species and phyla been overestimated? and (ii) how may species differ with respect to their respective levels of stenothermy and their capacities to acclimate to thermal change? Recent data show that Antarctic fish can undergo thermal acclimation and shift their physiological characters accordingly, e.g. in a zoarcid (Lannig et al., 2005) and a notothenioid (Seebacher et al., 2005).

A comparison of the mechanisms characterizing thermal intolerance between and within species of marine invertebrates and fish has led to the development of a unifying physiological concept of thermal limitation and adaptation. The first line of thermal intolerance in animals, which restricts performance in behaviour, growth and reproduction, is the limit on the capacity of oxygen supply mechanisms (Pörtner, 2001, 2002; Pörtner et al., 2001). Thermally induced reduction in oxygen supply capacity can take place at both high and low temperature extremes, before any biochemical stress indicators are affected. Between these extremes is a temperature window in which there is maximum scope for the aerobic activity and associated performance needed for successful survival in the wild. These thresholds were defined as critical temperatures (Tc) (see review by Pörtner, 2001). At more extreme low and high temperatures, there is a transition to anaerobic mitochondrial metabolism as the capacity for oxygen supply diminishes. This is seen in crustaceans (Frederich and Pörtner, 2000) and other invertebrate phyla (Pörtner, 2001, 2002), in temperate and Antarctic fish like zoarcids (temperate *Zoarces viviparus*, Antarctic *Pachycara brachycephalum*) and in sub-Arctic fish like Atlantic cod (*G. morhua*) (Mark et al., 2002; Zakhartsev et al., 2003; Lannig et al., 2004) (see also Van Dijk et al., 1999; Pörtner et al., 2004).

Recent evidence demonstrated the ecological relevance of oxygen limited heat tolerance through its effect at ecosystem level (Pörtner and Knust, 2007). Heat stress in the wild reduced performance (Pörtner and Farrell, 2008, Figure 4.49) and enhanced mortality even before critical temperatures were reached. These findings emphasize the early effect and crucial role of limitation in oxygen supply in compromising fitness.

Antarctic marine invertebrates may be more thermally sensitive than fish (Pörtner et al., 2007). The critical temperature in the infaunal bivalve *Laternula elliptica* lies at about 6°C (Pörtner et al., 1999; Peck et al., 2002). Before reaching that temperature this species develops systemic hypoxia (hypoxemia), which reduces whole organism performance. Early reductions in aerobic scope include a complete loss of ability to burrow in *L. elliptica* or to self-right in the limpet *Nacella concinna* at 5°C (Figure 4.50), and a 50% loss of capability at temperatures between 2°C and 3° C (Peck et al., 2004). The scallop, *Adamussium colbecki* was even more thermally constrained, being totally incapable of swimming when temperatures rose to 2°C. The early loss of performance and a progressive reduction in haemolymph oxygenation suggests that thermal thresholds are close to 0°C in *L. elliptica* (Peck et al., 2004; Pörtner et al., 2006). The thermally most sensitive Antarctic invertebrate to date is the bivalve *Limopsis marionensis* from the Weddell Sea, with a critical temperature of 2°C (Pörtner et al., 1999). Mortality tests confirm the fatal effect of oxygen limited thermal tolerance and the inability of invertebrates to acclimate to higher temperatures. In epifaunal scallops (*A. colbecki*), half of the specimens died after 19 days at 4°C (D. Bailey, pers.

commun.). In infaunal clams (*L. elliptica*), half died in 2 months at 3°C, and in the brittle star *Ophionotus victoriae* half died in less than 1 month at 3°C (L. Peck, pers. obs.). It appears that Antarctic stenotherms, especially among the invertebrates, live close to their thermal optimum, while others, like Antarctic fish, may live permanently below their optimum. For example, the Antarctic eelpout (*P. brachycephalum*) grows optimally at around 5°C, well above ambient temperatures (Brodte et al., 2006), in accordance with its ability to acclimate to warmth (Lannig et al., 2005).

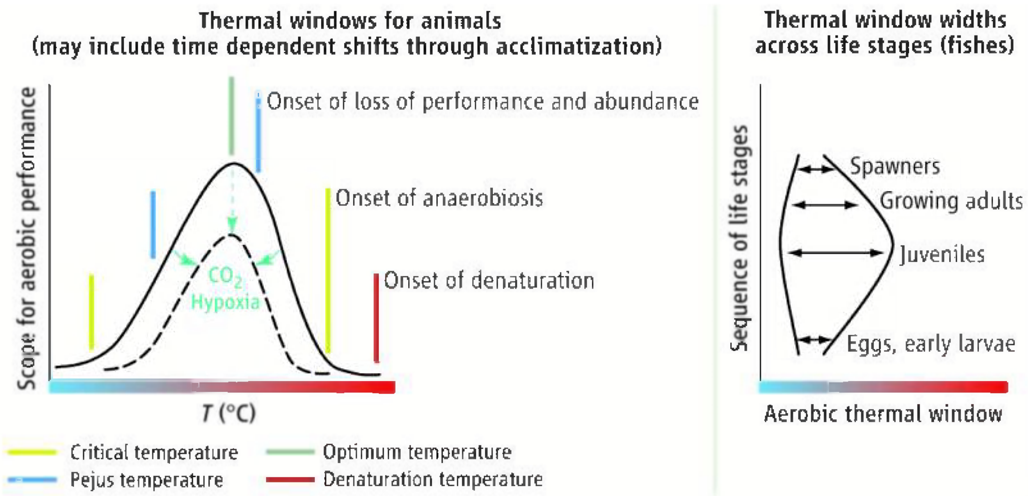


Figure 4.49 Temperature effects on aquatic animals. The thermal window of aerobic performance (left) display optima and limitations by pejus ("turning worse"), critical, and denaturation temperatures, when tolerance becomes increasingly passive and time-limited. Seasonal acclimatization involves a limited shift or reshaping of the window by mechanisms that adjust functional capacity, endurance, or protection. Positions and widths of windows on the temperature scale shift with life stage (right). Synergistic stressors like ocean acidification and hypoxia narrow thermal windows according to species-specific sensitivities (broken line), further modulating biogeographies, coexistence ranges, and other interactions. From: Pörtner, H.-O. and Farrel, A.P. (2008) Pysiology and Climate Change. Science 322:690-692; reprinted with permission from AAAS.

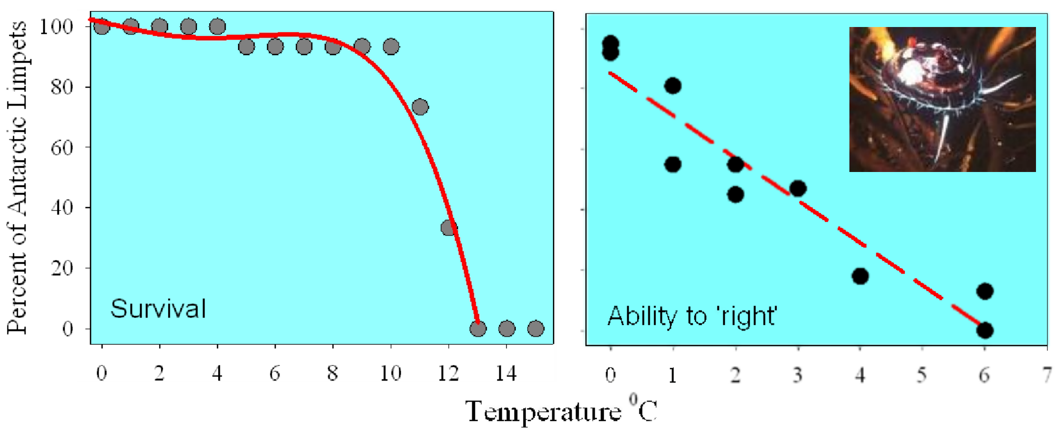


Figure 4.50 Acute temperature influences on Antarctic invertebrates. Survival and ability to 'right' (turn back over) of the Antarctic limpet *Nacella concinna*. Data from Peck (2005, unpublished).

Several Antarctic marine invertebrate species dwell in the intertidal zone and may experience elevated temperatures in summer. They likely use metabolic depression strategies and anaerobic metabolism to survive in response to temperature-induced hypoxemia (Pörtner, 2002). A role for hypoxia in metabolic depression was also evident during the winter season (Morley et al., 2007) and is known in temperate zone invertebrates (Grieshaber et al., 1994).

Under field conditions, the loss of aerobic performance capacity at higher temperatures limits the survival of invertebrates in warmer summers. Data from physiological and other laboratory studies suggests that further warming of the marine environment by as little as 1°C will exceed the thermal tolerance of some marine invertebrate species. These polar ectotherms clearly do not have the opportunity to retreat to cooler, i.e. higher latitude waters, which leads to the expectation of fatal consequences not only for individual species but possibly also for characteristic properties of ecosystem structure and functioning. While the apparent stability of Antarctic foodweb structures in response to potential species losses may at least temporarily buffer such changes (Jacob, 2006), it cannot prevent the potential loss of typical Antarctic fauna.

Functional or physiological aspects of meiofauna in general, and nematodes in particular, remain poorly known. Studies of the trophic position of meiobenthos in temperate and tropical areas have led to conflicting results (van Oevelen et al., 2006 and references therein) and studies in Antarctic and sub-Antarctic sediments are preliminary and restricted to a limited number of habitats (Moens et al., 2007). Biomarker analysis of bulk sediment organic matter and of nematodes in different regions and sediment types was carried out to assess the energy source of meiobenthic fauna in Antarctic shelf sediments (Moens et al., 2007). The results of this study suggested substantial selectivity of the metazoan meiobenthos for specific components of the sedimented organic matter, such as ice algae or flagellates, with this selectivity differing between sites and sediments. Laboratory experiments on a number of selected species from temperate regions showed that reproductive success, growth and metabolic activity of nematodes largely depend on temperature, the quality and quantity of food, and to a lesser extent salinity, with different species thriving under different conditions (Moens and Vincx, 2000a,b). A better understanding of the current functionality of the meiobenthic communities in different habitats is needed, and will allow for assessment of how these processes can be affected by changes in the environment. These changes might also impact structural aspects of the meiobenthic, and more specific the nematode communities, such as community composition and diversity, but also densities and biomass.

Examining physiological responses to temperature change is difficult for free-swimming pelagic organisms such as Antarctic krill, which do not behave naturally in captivity. For krill an alternative approach has been taken to look at temperature effects. Somatic growth rates can be determined in a manner largely free from laboratory artefacts, making them useable as a field-derived index of physiological health (Atkinson et al., 2006). By performing many such experiments across krill's temperature range during the summer growth period, the various factors affecting growth (chiefly food, krill size and temperature) can be teased apart. These results showed that, having adjusted for food and krill size, growth was highest at low temperatures and decreased substantially above 3°C (Figure 4.51). This stenothermy has implications for krill at their northern limits, for example at South Georgia. Surface layer temperatures have already warmed there by about 1°C in the last 80 years (Whitehouse et al., 2008) and future increases could make this region increasingly unsuitable for krill in summer.

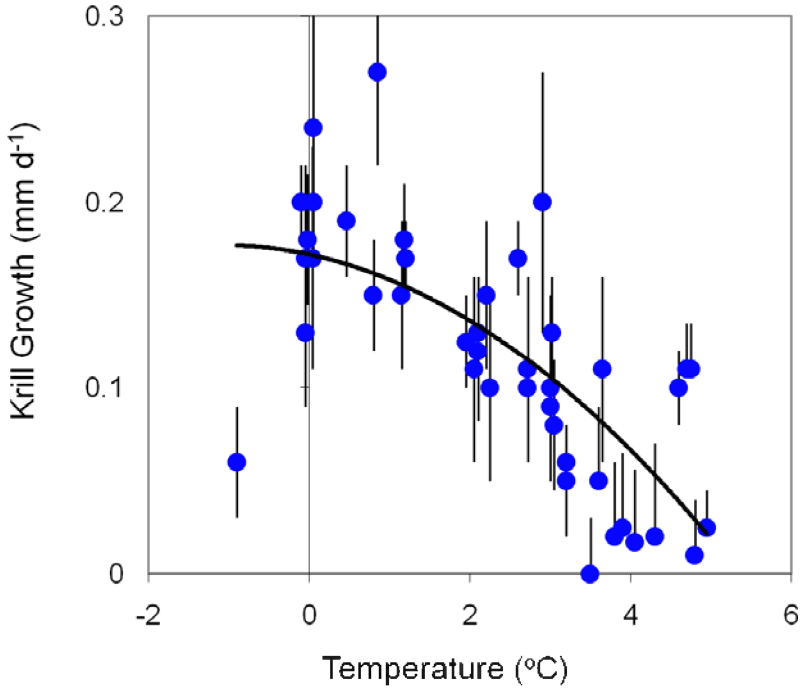


Figure 4.51 Daily growth rates of krill in the Scotia Sea/South Georgia region measured in January-February, and having been adjusted for effects of variable food availability and krill size via a mixed model (re-drawn from Atkinson et al., 2006). Each point represents a growth experiment derived for a swarm of krill, sampled across nearly the full range of summer surface temperatures that they occupy.

4.7.2.5 Seasonality Effect on the high Antarctic benthic shelf communities?

The Antarctic spring is one of the planet's principal episodes of oceanic primary production (Hense et al., 2003), reaching maximum values of 0.1 mg Chl/l in just a few weeks. As more than 10^7 km^2 of sea ice melts, it releases a huge trapped biomass (Thomas and Dieckmann, 2002). Sunlight continues to increase from spring to summer, driving notable changes within an ecosystem just emerging from a long, dark winter. This explosion of life is immediately followed by a growth spurt in the life cycle of the krill, the organism standing at the base of the food chains for nearly all Antarctic marine vertebrates. As winter approaches, the continental shelf and large areas of the open ocean pass back towards a seasonal coverage of ice more than a metre thick, which is why most of the large predators abandon the high Antarctic at the start of the long austral winter.

This pattern led to the conception of a long-lasting paradox – that the ocean around Antarctica experienced pronounced marine seasonality (Clarke, 1988), with a period of low activity in winter as a consequence of reduced food availability, despite the fact that the sea water temperature remained practically constant all year round.

While the marked environmental seasonality naturally does influence and condition life in the water column, the first inklings that the Antarctic paradox might not be entirely accurate arose after the discovery of the rich marine fauna that dwells on the continental shelves in the high Antarctic (Gutt et al., 1992). Over the past twenty years, the region has been shown to host one of the most diverse, high-biomass benthic communities in the ocean (Clarke and Johnston, 2003). Suspension feeders constitute the bulk of these communities, which depend on the particles settling down from the upper layers of the water column or laterally advected to them by currents. Due to low temperatures, a large number of species have slow metabolic rates, associated with a low energy demand, yet they still attain considerable age and size (Peck et al., 2006). This and other traits connected with reproduction patterns would at first glance appear to be in harmony with the tenets of the Antarctic paradox, rooted in the dormant state thought to prevail in winter. However, new features forcing the scientific community to reconsider the Antarctic Paradox have recently come to light. For instance, quite a few species exhibit reproduction rates similar to those in other regions, while others demonstrate higher growth rates than expected by quickly occupying areas scraped clean by icebergs (Teixidó et al., 2004). Experimental observations have furnished earlier selected results (Barnes and Clarke, 1994, 1995) supporting the assumption that the long Antarctic winter may not be as inactive as hitherto thought. These findings include:

1. The existence of "food banks" extending over hundreds of kilometres, offering a potential food source for numerous bottom-dwelling organisms (Mincks et al., 2005). This phenomenon also known as "green carpets" tends to form at the beginning of the austral spring, when the high primary production generated by melting ice is not immediately exploited by planktonic grazers and settles on the shelf seabed in a time span of hours to days (Gutt et al., 1998).

2. Widespread distribution of seabed sediment with high nutritive quality and grain sizes suitable for the anatomic structures of benthic suspension feeders. On average, measured concentrations of protein (3 mg/g) and lipids (2 mg/g) are higher than on other continental shelves and similar to the contents found in settling particles (Smith et al., 2006).

3. Tides acting as an incessant mechanism to resuspend the "food banks" and supply particles to suspension feeders throughout the year (Smith et al., 2006).

4. Benthic suspension feeders on Antarctic shelves feeding on small-sized particles in contrast to species from other latitudes that mainly ingest zooplankton (Orejas et al., 2003).

The new evidence of the physical-chemical conditions on the shelf seabed at Antarctic latitudes makes it necessary to reconsider the paradox that had formerly served as a cornerstone for understanding polar ecosystems. After the summer, resuspension by tidal currents and the high nutritional quality of the seabed sediment allow benthic trophic conditions to remain almost constant throughout the year, which provides the basis for a new model of Antarctic seasonality. The new model helps to explain the high diversity and high biomass of benthic communities around Antarctica, even when the food input from the euphotic zone becomes scarce when ice covers the ocean surface during the winter months. These new findings must be taken into account when planning future research on the Antarctic bottom-dwelling fauna. Special emphasis should be placed on carrying out studies during the austral winter, when processes occurring near the seabed (Figure 4.52) could be a key to understanding both the high productivity of the system in the early spring and the high biodiversity of the benthic ecosystem (Gili et al., 2006).

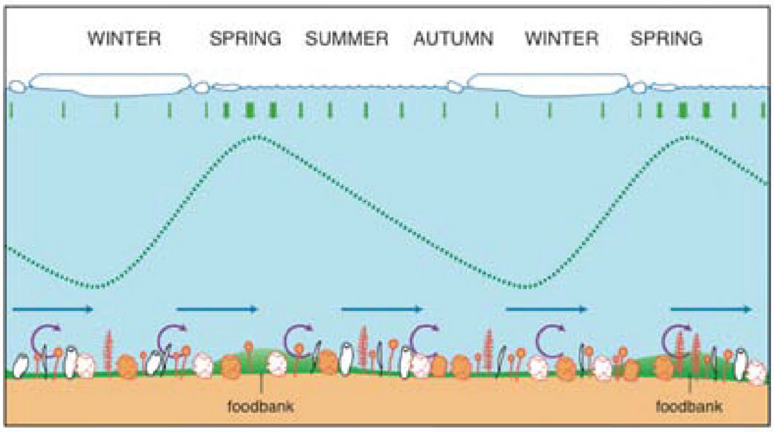


Figure 4.52 Synoptic view of the processes described in the text showing the seasonal vertical flux of new organic matter originated mainly at the beginning of spring (green line), the seasonal variation of food banks and the lateral and resuspension transport just above the seabed (arrows close the bottom).

4.7.2.6 Macroalgal physiology and ecology

Many Antarctic macroalgae, especially from the subtidal zone, are adapted to low temperatures (Wiencke et al., 1994, 2007), especially those endemic to the Antarctic. For example, the red algae *Georgiella confluens* (Figure 4.53), *Gigartina skottsbergii* and *Plocamium cartilagineum* grow only at 0°C and not at 5°C, and exhibit upper survival temperatures (USTs) as low as 7 to 13-14°C, respectively (Bischoff-Bäsmann and Wiencke, 1996). Other Antarctic red algae grow at temperatures up to 5 or 10°C and have USTs of up to 19°C. Macroscopic growth stages of endemic Antarctic brown algae grow up to 5°C and exhibit USTs of 11 to 13°C. Their reproductive microscopic stages grow up to 10 or 15°C and have USTs between 15 and 18°C (Wiencke and Dieck, 1989). Antarctic cold-temperate species (especially from the intertidal) are characterised by higher temperature requirements (Wiencke and Dieck, 1990).

This adaptation of Antarctic macroalgae to low temperatures is also reflected in photosynthesis. The photosynthetic capacities of endemic Antarctic species measured at 0°C are as high as in temperate algae measured at higher temperatures (Thomas and Wiencke, 1991; Wiencke et al., 1993; Weykam et al., 1996; Eggert and Wiencke, 2000). However, optimum temperatures for photosynthesis are clearly above the Antarctic water temperatures. The lowest temperature optima have been recorded in *Ascoseira mirabilis* (1 to 10°C). The other species tested so far are characterised by higher values between 10 and 20°C. Temperature optima for dark respiration are higher than those for photosynthesis (Drew, 1977; Wiencke et al., 1993; Eggert and Wiencke, 2000).

Due to global warming the geographic distribution of stenothermal cold water species will be limited in future to the coasts of the Antarctic continent itself. Species presently occurring today predominantly in the cold-temperate region and rarely in the Antarctic will become more established in the Antarctic region (Müller et al., submitted). This can cause unforeseeable changes in the ecology of Antarctic shallow water communities.

Light determines significantly the depth zonation and survival of macroalgae. Antarctic seaweeds are able to tolerate dark periods of up to one year without suffering damage, and their light requirements for growth are very low (Wiencke, 1990). In juvenile sporophytes of

Antarctic Desmarestiales, growth is light saturated at ≤ 20 µmol photons/m^2/s^1 (Wiencke and Fischer, 1990). The low light requirements for completion of the life cycle and for growth of Antarctic seaweeds are based on the low light requirements for photosynthesis (Wiencke et al., 1993; Weykam et al., 1996; Eggert and Wiencke, 2000). As a result, light compensation (I_c) and saturation (I_k) points are very low in Antarctic species and mostly range between 1 and 15 and between 14 and 52 µmol photons/m^2/s^1, respectively (Wiencke et al., 1993; Weykam et al., 1996; Brouwer, 1996; Eggert and Wiencke, 2000). These low light requirements allow Antarctic macroalgae to grow down to considerable depths. Theoretically, growth of some species is possible at water depths down to approximately 75 m (Wiencke, 1990; Wiencke et al., 1993).

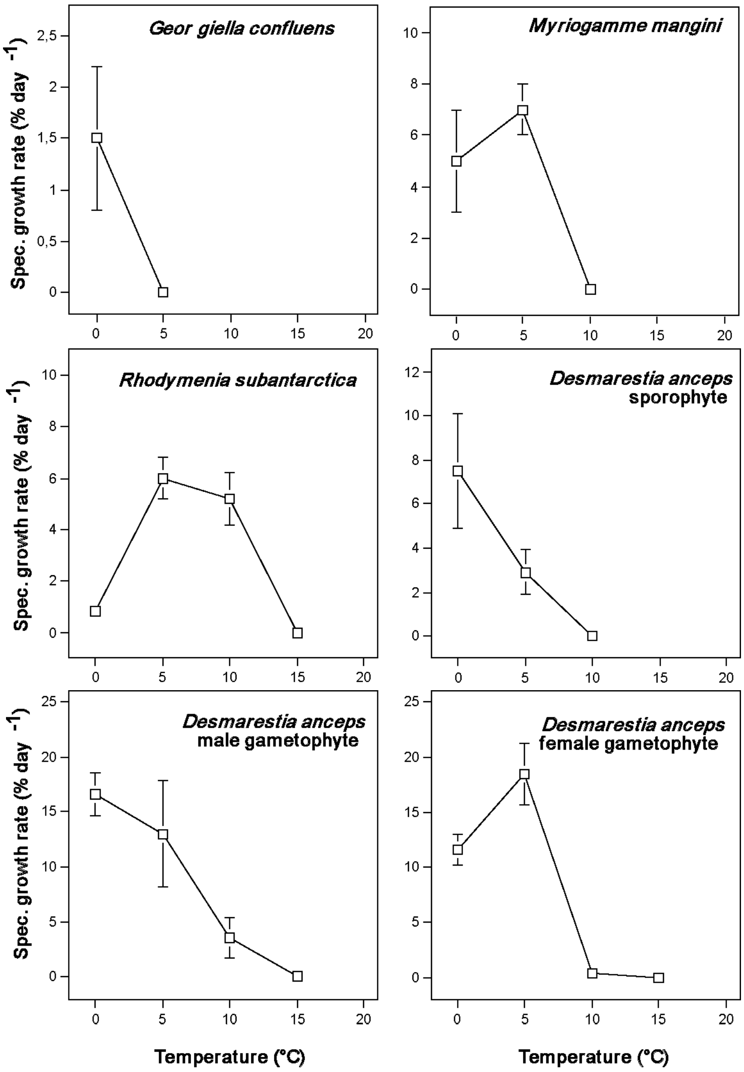


Figure 4.53 Physiological temperature adaptation of Antarctic macroalgae (after Wiencke and Clayton, 2002)

Antarctic algae are not only low light adapted but can also cope with high light conditions in summer due to their ability for dynamic photoinhibition, a photoprotective mechanism by which excessive energy absorbed is rendered harmless by thermal dissipation. The capability for dynamic photoinhibition is related to the depth distribution of the individual species. Species from the eulittoral, upper and mid sublittoral show a more or less pronounced decrease in photosynthetic activity during high light stress, and full recovery during subsequent exposure to dim light. In contrast, deep water and understory species recover only slightly and slowly indicating photodamage, especially after exposure to UV radiation. Recent studies indicate that the effect of UV radiation is not limited to the physiological and organism level. Rather, it influences the biodiversity, structure and function of macroalgal communities (Zacher et al., 2007; Campana et al., submitted). The effects on the community level depend on the UV susceptibility of the unicellular propagation units of the seaweeds (Roleda et al., In Press Zacher et al., 2009). So UV radiation represents an important factor for the determination of the upper distribution limit on the shore. The obtained results are of great importance for the estimation of the effects of enhanced UV radiation due to stratospheric ozone depletion. UV induced changes in algal zonation and biodiversity make changes in the trophic relations of coastal ecosystems very probable.

4.10.3 Marine/terrestrial pollution

Increasing concern about global changes and environmental protection is promoting international efforts to assess future trends and to mitigate the main causes and effects of climatic and environmental changes. In the last century the economic and industrial development of the Northern Hemisphere has had a dramatic impact on the global environment. Antarctica, the remotest continent in the Southern Hemisphere, has thus become a symbol of the last great wilderness and pristine environment. However, the world's future population growth and industrial development will occur in countries of the Southern Hemisphere. The rapidly changing global pattern of persistent anthropogenic contaminants and greenhouse gases may reduce the value of Antarctica and the surrounding Southern Ocean for research on evolutionary and ecophysiological processes and as an ideal archive of data for better understanding global processes (Bargagli, 2005).

Antarctica has a very small, non-native population and is protected by natural "barriers" such as the Antarctic Circumpolar Current and the zone of circumpolar cyclonic vortex, which reduce the entry of water and air masses from lower latitudes and the consequent import of propagules and persistent contaminants. Nonetheless, the recurring appearance of the "ozone hole" and the rapid regional warming of the Antarctic Peninsula indicate that Antarctica and the Southern Ocean are inextricably linked to global processes and that they are not escaping the impact of local and global anthropogenic activities.

Exploration, research, sealing and whaling have drawn people to Antarctica since the early 1900s, and the development of research, tourism and fishing in the last 50 years has driven a striking increase in human presence. Local impacts due to the presence of humans, such as contamination of air, ice, soil, marine sediments and biota through fuel combustion (for transportation and energy production), waste incineration, oil spillage and sewage, are inevitable, especially near operational bases. Another serious anthropogenic impact, especially in the sub-Antarctic islands (where climatic and environmental conditions are less extreme) is the introduction of alien organisms (Frenot et al., 2005). Although the number of tourists visiting Antarctica is usually two-three times greater than that of the logistic and scientific personnel, the latter reside for a longer time in permanent or semi-permanent stations. Most stations are located in coastal areas and until twenty years ago refuse was dumped into landfill sites or the sea or burnt in the open air. Several accidental spills of oil, lubricants and foreign chemical compounds have occurred at and around Antarctic stations; unfortunately, in the Southern Ocean there have also been significant oil spills, such as the release in January 1989 of about 550 m^3 of diesel fuel during the sinking of the Argentine supply ship *Bahia Paraiso,* near Anvers Island (Antarctic Peninsula).

4.10.3.1 Local-scale pollution

The IGY (1957-1958), with the involvement of 12 countries and over 5,000 persons occupying 55 stations in the continent and the islands of the Southern Ocean, marked the beginning of significant local detrimental impacts on the Antarctic environment. Although concern about local environmental pollution has been expressed since the 1970s (e.g. Cameron, 1972), the value of the Antarctic environment to science was only definitively acknowledged in 1991 with the adoption of the Madrid Protocol to the Antarctic Treaty. The Protocol marked the exclusion of Antarctica from the geopolitics of the Cold War period, territorial claims and the possible exploitation of natural resources. It provided strict guidelines for environmental management and protection, and established the obligation to clean-up abandoned work sites. Some countries began to document environmental pollution at abandoned stations and waste-disposal sites, and developed strategies for clean-up and remediation (e.g., Kennicutt et al., 1995; Deprez et al., 1999; COMNAP-AEON, 2001; Webster et al., 2003; Stark et al., 2006). Trace metals and several Persistent Organic Pollutants (POPs) such as Polycyclic Aromatic Hydrocarbons (PAHs) and other by-products of combustion and incineration processes, including Polychlorinated Dibenzo-p-Dioxins (PCDDs), Polychlorinated Dibenzofurans (PCDFs) and Polychlorinated Biphenyls (PCBs), were among the most common persistent contaminants in terrestrial and marine ecosystems within a few hundred metres of scientific stations (UNEP, 2002). However, despite existence of the Protocol, its translation into practice is patchy at best, and much remains to be achieved even to quantify direct human impacts on the Antarctic, let alone effective mitigation where required (Tin et al., 2009).

Most ice-free areas of continental Antarctica are cold desert environments with sparse biotic communities, comprising few species of microorganisms, cryptogams and invertebrates. Although there are several reports on the distribution of persistent contaminants in Antarctic soils, mosses and lichens (e.g. Bacci et al., 1986; Claridge et al., 1995; Bargagli, 2001), possible long-term (biotic and abiotic) effects of persistent contaminants in Antarctic terrestrial ecosystems are largely unknown. There is evidence that hydrocarbon spillage in soils can result in an increase in hydrocarbon-degrading microbes and concomitant decrease in the diversity of the soil microbial community (Aislabie et al., 2004). However, the "in situ" biodegradation rate is probably very low, because aliphatic and aromatic compounds can be detected in soils more than 30 years after a spill.

In striking contrast with the extremely reduced number of species in terrestrial biotic communities, most Antarctic marine ecosystems have a rich variety of species and a high biomass. Contaminants are introduced in the coastal marine environment through waste water, leachates from dump sites, and deposition of particulates from station activity and ship operations. The accumulation of metals and POPs has been reported in samples of water, sediments and organisms collected in the vicinity of several Antarctic stations (e.g. UNEP, 2002; Bargagli, 2005). Throughout the 1970s, wastes from McMurdo station were routinely discharged along the eastern shoreline of Winter Quarter Bay, which also provided docking facilities for ships. In 1988 the US National Science Foundation began a dumpsite cleanup and abatement programme and the bay became one of the most studied marine environment in Antarctica. Concentrations of PAHs, PCBs, Polichlorinated Terphenyls (PCTs) and metals such as Ag, Pb, Sb and Zn in sediments from the bay were much higher than in samples from outside the area (Risebrough et al., 1990; Lenihan et al., 1990; Kennicutt et al., 1995).

In Antarctica, evolutionary processes in isolation contributed to the development of rich communities of coastal benthic organisms such as sponges, hydroids, tunicates, polychaetes, molluscs, actinarians, echinoderms, amphipods and fish. These organisms are characterized by a high degree of endemism and ecophysiological adaptations to peculiar physico-chemical features of the Southern Ocean. As many species are long-lived, have low metabolic rates, lack the pelagic larval phase and need longer development time, Antarctic benthic organisms are more exposed to the long-term effects of environmental contaminants than are temperate or tropical species. Significant disturbance of benthic communities has generally been reported in the proximity of the most polluted coastal sites (e.g., Lenihan and Oliver, 1995; Conlan et al., 2000; Stark et al., 2003). Organism responses to the combined effects of toxic pollutants and organic enrichment from sewage disposal usually involve a decrease in the abundance and diversity of benthic fauna and an increase in resistant and opportunistic species. Research on polluted sediments near Casey Station (Cunningham et al., 2005) has revealed that benthic diatom communities are good indicators of anthropogenic metal contamination and may be useful in monitoring the success of environmental remediation strategies in polluted Antarctic sites.

In recent years there has been considerable interest in astrobiology and microbial forms living in extreme environments. Antarctica has become one of the most important places for research on bacteria that thrive on ice or in subglacial lakes. One of the main challenges for such research is the contamination of ice, especially by drilling fluids. These fluids are a complex mixture of aliphatic and aromatic hydrocarbons and foranes that coat the ice surface and may penetrate into the interior of the ice through micro-fissures. As a rule, the fluid is not sterilized during use, and Alekhina et al. (2007) isolated bacteria of the genus *Sphingomonas* (a well-known degrader of polyaromatic hydrocarbons) as well as bacteria attributable to human and soil sources from specimens of the deepest (3,400 and 3,600 m) ice borehole at Lake Vostok.

4.10.3.2 Global-scale contaminants

Today, human activity is one of the most important parameters affecting the chemical composition of the atmosphere. The most marked recent changes in the chemistry of the atmosphere over Antarctica are associated with greenhouse gases, notably the largely anthropogenically caused decrease in stratospheric ozone, and increases in anthropogenically sourced CO_2 (380 ppmv), CH_4 (1755 ppbv), and N_2O (319 ppbv). Anthropogenically sourced radionuclides stemming from above ground nuclear bomb testing are also present throughout Antarctica, as is evidence of the Chernobyl nuclear accident (Dibb et al., 1990), and have even been used to provide dating control within long-lived biological systems (Clarke, 2008).

Of particular concern also is how to unravel human-induced changes in the atmospheric cycles of heavy metals entering the Antarctic atmosphere. The chemical analysis of snow and ice deposited over time in polar ice caps gives unique information on the pollution of the atmosphere with heavy metals over the last centuries. It should however be emphasized that these Antarctic archives mainly reflect pollution emitted in the Southern Hemisphere, especially in South America and South Africa. This is because pollution emitted in the Northern Hemisphere is less likely to penetrate into the Southern Hemisphere and ultimately to reach Antarctica. Deciphering these frozen archives is very difficult. Antarctic snow is so clean that the utmost precautions are required to collect the samples without contaminating them. As an example, one thousand tonnes of Antarctic snow typically contains no more than a few milligrams of lead. Heavy metals, metals or metalloids associated with contamination and potential toxicity, are derived from both natural and anthropogenic sources. While the main natural sources of heavy metals include rock and soil, volcanoes, sea-salt spray, wild forest fires, and continental and marine biogenic activities (Nriagu, 1989), anthropogenic toxic metals come mainly from industrial and domestic activities such as the mining and smelting of metals, the combustion of fossil fuels and refuse incineration (Nriagu and Pacyna, 1988).

Local emissions of heavy metals from human activities in Antarctica have the potential to give rise to environmental contamination in local areas. Major anthropogenic emissions within Antarctica are related to fuel burning for operation of Antarctic stations, snowmobiles, heavy vehicles, ships and aircraft (Boutron and Wolff, 1989). Although local effects may be confined to an area close to the site of such activities, special attention has to be given to local pollution related to stations and regularly utilized traverse routes (Qin et al., 1999, because human activities within Antarctica are increasing in number, magnitude, and intensity. An inventory of these emissions is needed for assessing the discharges of heavy metals into the Antarctic environment.

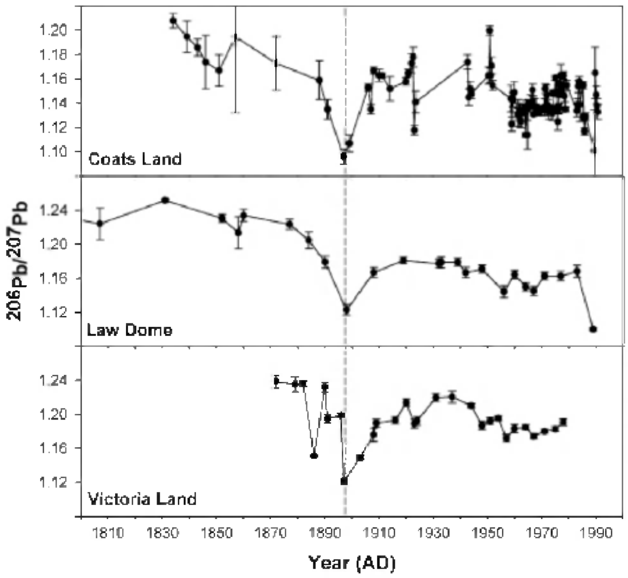


Figure 4.54 A comparison of $^{206}Pb/^{207}Pb$ ratios at Coats Land, Law Dome and Victoria Land over the past 200 years (Van de Velde et al., 2005).

Amongst the most spectacular results obtained so far is the reconstruction of past changes in lead pollution achieved by analyzing a series of snow samples collected in Coats Land, Victoria Land and Law Dome (Van de Velde et al., 2005). Lead isotopic profiles (Figure 4.54) show that lead pollution of Antarctica started as early as the 1880s. This early pollution was at least partially linked with non-ferrous metal production activities in South America, South Africa and Australia. Another important contribution was from coal-powered ships that crossed Cape Horn en route between the Atlantic and Pacific Oceans, as well as whaling ships and shore based stations along the Antarctic coast and Southern Ocean islands. This lead pollution then declined in the 1920s, correlated with the opening of the Panama Canal in 1914, which resulted in a pronounced decrease in ship traffic around the southern tip of South America. Lead pollution then increased again after World War II because of the very large rise in the use of leaded petrol in various countries in the Southern Hemisphere, combined with the continuous increase in non-ferrous metal production in South America, South Africa and Australia. Finally, lead concentrations strongly declined from the mid 1980s onwards, because of the fall in the use of leaded petrol in modern cars.

There is growing evidence from Antarctic snow and ice sampling highlighting the fact that Antarctica is already significantly contaminated with other metals such as Cr, Cu, Zn, Ag, 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)

POPs, pesticides and industrial chemicals, are ubiquitous anthropogenic compounds that are released around the world and found in the remote polar regions far distant from their primary sources, due to long-range atmospheric transport. Because they are hydrophobic, persistent and toxic, the trend of their distribution and their concentration are urgently needed to assess environmental contamination. The distribution pattern and levels of POPs such as PCBs and chlorinated pesticides such as Dichlorodiphenyltrichloroethane (DDT) and hexachloribenzene (HCB) are reported in the Antarctic environment (Fuoco et al., 1996; Carsolini et al., 2002; Weber and Goerke, 2003; Montone et al., 2003). Changing levels of some compounds of POPs in living organisms such as fish species reflect global redistribution and increasing transfer to Antarctic waters probably due to recent usage in the Southern Hemisphere and climate change (Weber and Goerke, 2003).

Despite increasing documentation, our ability to quantify environmental contamination from natural and industrial sources and to clarify the effects of various physico-chemical processes controlling atmospheric cycles of anthropogenic pollutants in the Antarctic is still limited because of weaknesses in the available scientific database.

Although some human activity in Antarctica, such as ship or aircraft transportation or the release of weather and research balloons can have widespread effects, scientists, tourists and fishermen are generally the main causes of local disturbance of the Antarctic environment. As pesticides have neither been produced nor applied in the continent, the discovery of DDT and its congeners in Antarctic marine biota in the 1960s and in the environment in the 1970s proved that persistent contaminants in the region come from other continents. Since then, HCB, Hexachlorocycloexanes (HCHs), aldrin, dieldrin, chlordane, endrin, heptachlor and other POPs have all been detected in Antarctica and the Southern Ocean. These chemicals are persistent, hydrophobic and lipophilic, accumulate in organisms and biomagnify in marine food chains (UNEP, 2002). While early ecotoxicological studies concentrated mainly on eggshell thickness and reproductive potential of birds, more recent evidence suggests that several POPs are also immunotoxic, endocrine disrupters and tumor promoters.

Although migrating species of marine birds and mammals may contribute to the southward transfer of POPs, their transport to the Southern Ocean and Antarctica mainly occurs through atmospheric and marine pathways. The Antarctic atmosphere loses more heat by radiative cooling than it gains by surface energy exchange and the deficit is balanced by atmospheric transport of heat, gases, moisture and aerosols from lower latitudes. The global transport of radiatively important trace gases such as CO_2, CH_4 and CFCs contribute for instance, to climate change and springtime ozone depletion events. Particles and reactive gases in air masses are partly removed by cyclonic storms in the belt of "westerlies" and are deposited in the Southern Ocean (Shaw, 1988). However, during the austral summer the circumpolar vortex disappears and long-term records of mineral dust, black carbon and ^{210}Pb at the South Pole and at some coastal Antarctic stations indicate an enhanced poleward transport of air masses (Wolff and Cachier, 1998). For instance, the contamination of Antarctic aerosol and snow by Pb and Cu has been widely documented (e.g., Barbante et al., 1998) and has usually been attributed to leaded petrol, non-ferrous metal mining and smelting and other anthropogenic sources in South America, Africa and Australia. Moreover, although during the last decades the deposition of Pb in Antarctic snow is decreasing, that of Cu, Zn and other elements is increasing (Planchon et al., 2002).

As the two hemispheres of the Earth have separate circulation systems through much of the atmosphere, there is probably a limited input of anthropogenic contaminants from the Northern Hemisphere to Antarctica. However, the profiles of radioactive debris deposition in Antarctic snow and ice during the 1950s and 1960s showed the presence of fission products released in the Northern Hemisphere (Koide et al., 1979). Under ambient temperatures POPs may volatilize from water, soils and vegetation into the atmosphere, where they are unaffected by breakdown reactions and may be transported over long distances before re-deposition. Volatilization/deposition cycles may be repeated many times and according to the theory of cold condensation and global fractionation (Wania and Mackay, 1993) the most volatile compounds such as HCHs, HCB and low-chlorinated PCBs can redistribute globally. Although their concentrations in Antarctic biota are usually below those documented to have reproductive effects in related species in temperate and Arctic regions, organisms living under harsh Antarctic conditions may be more stressed and more vulnerable to the adverse effects of pollutants than those in temperate regions (Bonstra, 2004; Corsolini et al., 2006). There is evidence that in some Antarctic marine organisms concentrations of Cd, Hg and other pollutants can be relatively high (e.g., Bargagli, 2001; Bustnes et al., 2007); furthermore, POP accumulation in marine sediments is increasing their availability and accumulation in benthic fish species feeding on benthic invertebrates (Goerke et al., 2004).

While there is serious concern related to protection of the Antarctic environment, major efforts in international legislation that will help to safeguard the Antarctic environment are underway. The Stockholm Convention on Persistent Organic Pollutants (2004) is a global treaty designed to protect the environment from chemicals that remain intact in the environment for long periods, such as POPs. This legislation follows in the footsteps of international legislation that protects Antarctica such as the Montreal Protocol (1989). The Montreal Protocol is designed to protect the ozone layer by phasing out the production of a number of substances believed to be responsible for ozone depletion.

4.10.3.3 Concern about the future impact of human activity

New classes of global pollutants are emerging, such as perfluorinated compounds (PFCs) that have a wide range of industrial applications. These compounds have been shown to be toxic to several species of aquatic organisms and to occur in biota from various seas and oceans, including the Arctic and the Southern Ocean (Yamashita et al., 2005). Catalytic converters in motor vehicles are increasing global emissions of Pt and other companion elements such as Pd, Rh, Ru, Os and In. Increased concentrations of Rh, Pd and Pt with respect to ancient Greenland ice samples were measured in surface snow from the Alps, Greenland and Antarctica (Barbante et al., 1999).

Climate change and global warming could enhance the transport and deposition of persistent contaminants in Antarctica. The oceanic transport of persistent contaminants is often considered to be much less important than atmospheric transport; however, models which combine the transport of semi-volatile chemicals in air and water, and consider continuous exchange between the two compartments indicate that the overall transport of POPs to remote regions is accelerated with respect to models treating air and water separately (Beyer and Matthies, 2001). The rapid regional climatic warming of the Antarctic Peninsula has also been detected in oceanic waters to the west (e.g. Meredith and King, 2005). The warming of surface water can affect POP volatilization and transport. In contrast to organisms in temperate and tropical seas, those in the Southern Ocean are well adapted to narrow ranges of water temperature close to the freezing point. Slight increases in temperature may have disproportionate influence on the properties of cell membranes and biological processes involved in the uptake and detoxification of environmental pollutants.

Although for continental Antarctica there is as yet no significant trend in meteorological temperature, a loss of ice shelves such as that in the Antarctic Peninsula (six ice shelves have largely disappeared in the last 50 years) could have dramatic effects on atmospheric precipitation (i.e. the deposition of contaminants) and the biogeochemical cycle of trace elements such as Hg. Mercury emitted by anthropogenic and natural sources occurs in the atmosphere mostly in the gaseous elemental form (Hg^0), which has a long lifetime in tropical and temperate regions. Once deposited in terrestrial and aquatic ecosystems the metal is partly re-emitted into air, thus assuming the characteristics of global pollutants such as POPs. The metal is now acknowledged to be one of the most serious contaminants in polar ecosystems because of the springtime Hg depletion events that have been reported in the high Arctic (Schroeder et al., 1998) and Antarctica (Ebinghaus et al., 2002). In polar regions, after sunrise, the globally-distributed Hg^0 undergoes photochemically-driven oxidation by reactive halogen radicals (from sea-salt aerosols) and rapidly deposits on snow and on terrestrial and marine ecosystems. Field evidence of enhanced Hg accumulation in soil and cryptogamic organisms from terrestrial ecosystems facing the Terra Nova Bay coastal polynya raises concern that Antarctica may become an important sink in the global Hg cycle (Bargagli et al., 2005). By changing the sea ice cover and increasing the availability of reactive halogens, warming could enhance the role of Antarctica as a "cold trap" for Hg through an increase in the out-gassing of the metal from continents and oceans. Furthermore, while the use of Hg and many POPs has declined or ceased in North America and Europe since the 1990s and earlier, the growing demand for energy, the burning of coal and biomass, the extraction of gold, intensive agriculture, the spraying of pesticides for disease vector control, and the lack of emission control technologies in South America, Africa and Asia will likely increase the atmospheric burden of Hg and many other persistent contaminants.

4.11 Biogeochemistry - Southern Ocean Carbon Cycle Response to Historical Climate Change

4.11.1 Introduction

The Southern Ocean, with its energetic interactions between the atmosphere, ocean and sea ice, plays a critical role in ventilating the global oceans and regulating the climate system though the uptake and storage of heat, freshwater and atmospheric CO_2 (Rintoul et al., 2001; Sarmiento et al., 2004a). The Southern Ocean is dominated by the eastward flowing ACC. The surface of the ACC is characterized by a northward Ekman flow creating a divergent driven deep upwelling south of the ACC and convergent flow north of the ACC. In the southern part, the upwelling of mid-depth (2-2.5 km) water to the surface provides a unique connection between the deep ocean and the atmosphere while in the northern part the downwelling provides a strong connection to water masses that resurface at lower latitudes e.g. Sarmiento et al. (2004a). These connections make the Southern Ocean extremely important in controlling the storage of carbon in the ocean and a key driver in setting atmospheric CO_2 levels (Caldeira and Duffy, 2000).

4.11.2 CO_2 fluxes in the Southern Ocean

The Southern Ocean carbon cycle response uptake can be described as a combination of seasonal and non-seasonal variability. In the Southern Ocean seasonal variability is the dominant mode of variability, and the mechanisms that drive the air-sea CO_2 fluxes are well known but their magnitudes remained poorly constrained (Metzl et al., 2006 and references therein). The seasonal cycle of CO_2 uptake is a complex interplay between the biological and physical pumps and can be described both in terms of its magnitude and phase. A recent climatological synthesis of more than 3 million measurements of surface pCO_2 measurements provides important insights into Southern Ocean behaviour at the seasonal timescales (Takahashi et al., 2009). During the austral summer biological production reduces surface ocean pCO_2 through photosynthetic activity and then exports part of this organic matter to the deep ocean. This reduction in surface pCO_2 is offset by the changes in the physical pump that reduce the capacity of the surface water to store CO_2 through upper ocean warming, which lowers solubility, thereby increasing the surface ocean pCO_2. The net result of this competition between the biological and physical pumps is that the Southern Ocean acts a sink of atmospheric CO_2 in the summer in the sub-Antarctic zone (SAZ; nominally 40°S – 50°S) and south of the Polar Front (PF; ~50°S; Figure 4.55). During the austral winter, a picture of two distinct zones separated by the PF is evident. South of the PF relatively little biological activity occurs in winter (deep mixing and light limitation); therefore surface pCO_2 values are set by the competition within the physical pump between deep winter mixing bringing up CO_2 water from the carbon rich deep ocean, leading to increased pCO_2 surface levels, and a cooling that increases the ability of the surface waters to store CO_2. As the deep winter mixing dominates in this region a net out-gassing of CO_2 to the atmosphere occurs. North of the PF during the austral winter the same cooling and mixing processes that occur further south exist, but with the addition of biological activity during the period, albeit at a reduced rate, reducing the pCO_2 in surface waters. The result of the combined responses of the biological and physical pumps means this region acts as a weak sink of atmospheric CO_2 during the austral winter.

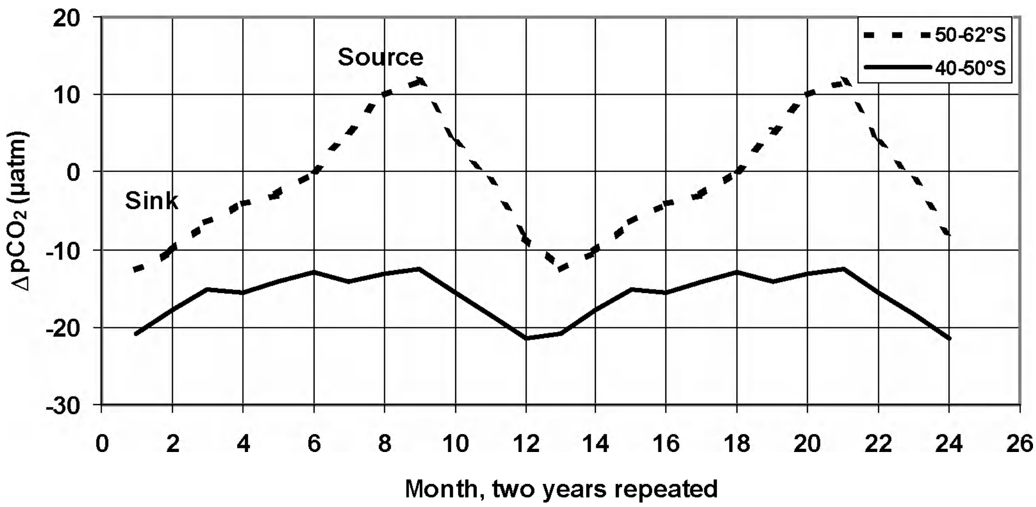


Figure 4.55 The annual cycle of ΔpCO_2 (pCO_2^{ocean}-pCO_2^{atm}; i.e. negative/positive = atmospheric CO_2 sink) in the Southern Ocean for the regions 40°-50°S (black line) and 50°-62°S (dashed line; Takahashi et al., 2009). The Sub-Antarctic zone (40°-50°S) acts a permanent CO_2 sink but at higher latitudes the ocean acts as an atmospheric sink during summer and a source during winter (Metzl et al., 2006).

When the observed winter and summer fluxes are integrated, the annual mean uptake is small south of 50°S (about -0.08 PgC/yr); conversely the SAZ (40°S-50°S) behaves as a strong sink (-0.74 PgC/yr) (Takahashi et al, 2009; see comments below). The response of the SAZ is consistent with other studies that suggest the SAZ is also a strong CO_2 sink approaching -1 PgC/yr (McNeil et al., 2007; Metzl et al., 1999) and an important region of mode and intermediate water formation and transformation. In comparison to the total uptake of 2 GtC/yr for the global ocean, the Southern Ocean south of 40°S takes up more than 40% of the total uptake. Note in these calculations we have used the gas transfer coefficent of Wanninkhof (1992) with the dataset of Takahashi et al., (2008).

Significant progress has been made in recent years in simulating the annual mean uptake of CO_2 by the Southern Ocean as precursor to understanding interannual to decadal variability. Figure 4.56 shows that the spatial pattern of the annual mean uptake is well represented in comparison with that simulated from the current class of ocean biogeochemical models (e.g. OPA/PISCES model; Aumont and Bopp, 2006)). Although different models do contain different representations of the magnitude of the seasonal cycle, the phase between each model and observations shows good agreement. This suggests that seasonal processes that drive this variability are well captured, and that as a result, we are in a better position to explore changes at interannual and longer timescales.

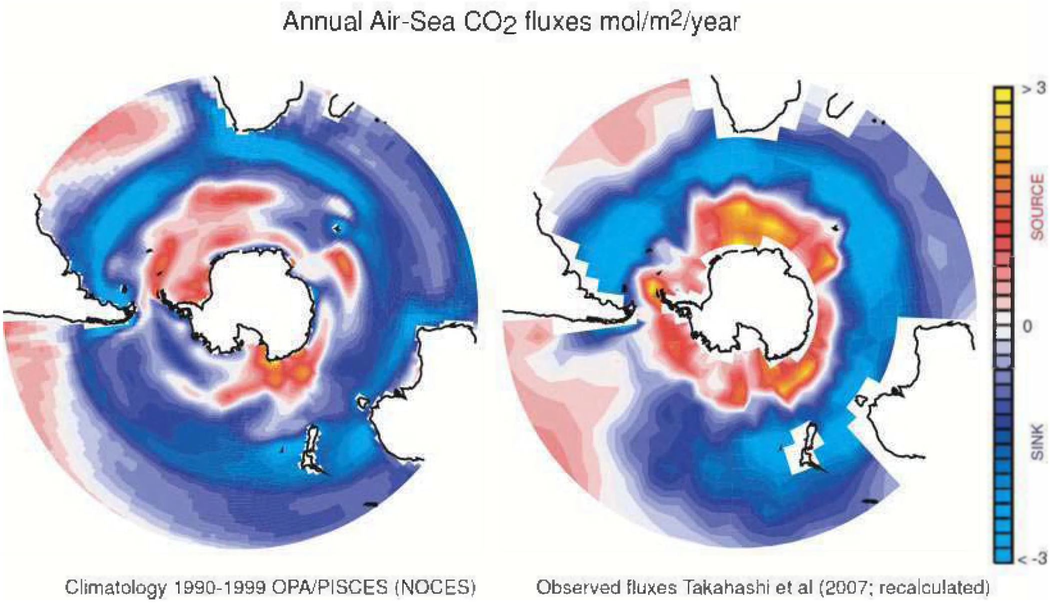


Figure 4.56 Annual mean uptake of air-sea CO_2 fluxes as calculated from OPA/PISCES 1990-1999 (Lenton et al., 2006) and that from the new climatology of Takahashi et al. (2009). The sub-Antarctic region (40-50°S) represents a strong sink (blue colors), whereas south of 50°S, large regions act as a CO_2 source for the atmosphere (red).

4.11.3 Historical Change - Observed Response

The Southern Ocean has undergone significant changes in response to climate changes; such as a net increase in heat and freshwater fluxes, and a poleward movement and intensification of winds e.g. Thompson and Solomon (2002). The major driver of these changes has been the SAM that is associated with changes in the Antarctic Vortex, primarily in response to depletion of stratospheric ozone and increasing atmospheric greenhouse gas concentrations (Arblaster and Meehl, 2006). While the SAM is a significant driver of variability it cannot explain all the variability present, as it has both linear and non-linear interactions with other climatic modes such as ENSO and Indian Ocean Dipole (IOD) that drives diverse responses in different regions. These physical changes impact directly on the physical pump and to a lesser extent on the biological pumps, therefore on the concentration of CO_2 in surface waters and the magnitude of both uptake and export of CO_2 from the atmosphere to the deep ocean.

In the Southern Ocean the interannual to decadal changes in biological production, ocean dynamics and thermodynamics that drive oceanic pCO_2 and air-sea CO_2 exchanges remain poorly understood and very undersampled. Decadal and interannual variations have been observed at high latitudes, but at only very few locations and during the austral summer (Jabaud-Jan et al., 2004; Brévière et al., 2006, Borges et al., 2008). Although these analyses provide important information on the response of the ocean to climate variability in the Southern Hemisphere, there is no clear detection of the decadal trends of oceanic CO_2 and associated air-sea CO_2 fluxes, unlike the situation in the north Atlantic or the north and equatorial Pacific where there are long-term time series of such data (e.g. Bates, 2001; Feely et al., 2002). Takahashi et al. (2009) have recently constructed a pCO_2 data synthesis, from which a significant increase of oceanic pCO2 during winter has been calculated, about +2.1±0.6 µatm/yr, which is close to or faster than the growth rate in the atmosphere (1.7 ppm/yr) over the period 1986-2007.

Repeat underway measurements of surface pCO_2 have been made regularly in the Southern Indian Ocean since the 1990s (e.g. Metzl et al.,1999). Although these measurements are quite often confined to regions where ships travel to resupply Antarctic and sub-Antarctic bases, they represent a valuable timeseries for exploring the evolution of the surface ocean. A recent study by Metzl (2009; Figure 3.68) in the South Western Indian Ocean calculated surface trends of pCO_2 between 1991-2007 and showed that oceanic pCO_2 increased at all latitudes south of 20°S (1.5 to 2.4 µatm/yr depending the location and season). More specifically, at latitudes of less than 40°S, they determined that oceanic pCO_2 increased faster than in the atmosphere since 1991, suggesting the strength of the oceanic sink decreased. In addition, when pCO_2 data are normalized to temperature, removing the effect of solubility on CO_2, this analysis showed that the system is increasing much faster in the winter than in the summer (Figure 4.57). These results suggest that the increase may be due to changes in ocean dynamics, given that the largest response occurs in the austral winter, when the winds are strongest. In the recent period (since the 1980s) the increase of pCO_2 appeared to be faster compared to the trends based on historical observations from 1969-2002 (Inoue and Ishii, 2005), suggesting that the Southern Ocean CO_2 sink has continued to evolve in response to climate change.

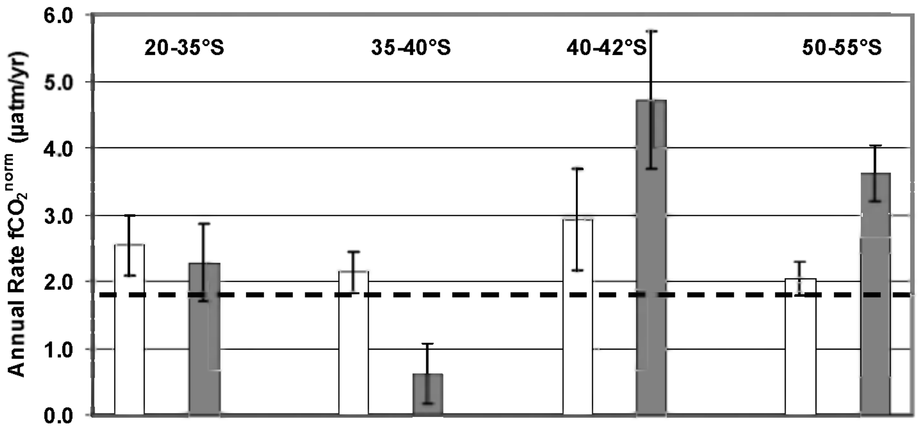


Figure 4.57 Annual mean trends of temperature normalized *f*CO_2 in 4 regions of the South-Western Indian Ocean (based on summer and winter observations in 1991-2007). The open bars indicate the growth rates estimated for summer and black bars for winter. Standard errors associated to each trend are also indicated. The dashed line indicates the atmospheric CO_2 annual growth rate (figure reproduced from Metzl, 2009).

As oceanic pCO_2 in recent years has been observed to be increasing close to, or faster than in the atmosphere, the signature of these changes in atmospheric CO_2 data should be detectable, as has been observed in the Equatorial Pacific during ENSO events (e.g. Peylin et al., 2005). At latitudes south of 40°S the ocean has a very large surface and it is expected that continental carbon source/sink variability has a low imprint in atmospheric CO_2 records (compared to the tropics and north hemisphere). This is clearly seen in the CO_2 record at La

Nouvelle Amsterdam Island (in the South-Indian Ocean), for example, where the seasonality of atmospheric CO_2 is very low. A recent study by Le Quéré et al. (2007) using a combination of atmospheric observations and inverse methods reported that in the period 1981-2004, the strength of the Southern Ocean CO_2 sink (south of 45°S) was reduced (Figure 4.58). Although this result remains controversial (e.g. Law et al., 2008), it does suggest that the observed increase in oceanic pCO_2 acts to reduce the strength of the air-sea CO_2 gradient (ΔpCO_2) and this in turn translates to reduction in the strength of the Southern Ocean CO_2 sink. This result is significant as it was expected in response to strengthening air-sea CO_2 gradient that the Southern Ocean CO_2 uptake would increase. These results are also consistent with a number of recent modeling studies, that have also suggested a decrease in uptake in the last decades and which attributed the upper ocean pCO_2 values to increases in the wind speed increasing the ventilation of carbon rich deep waters e.g. Lenton and Matear (2007).

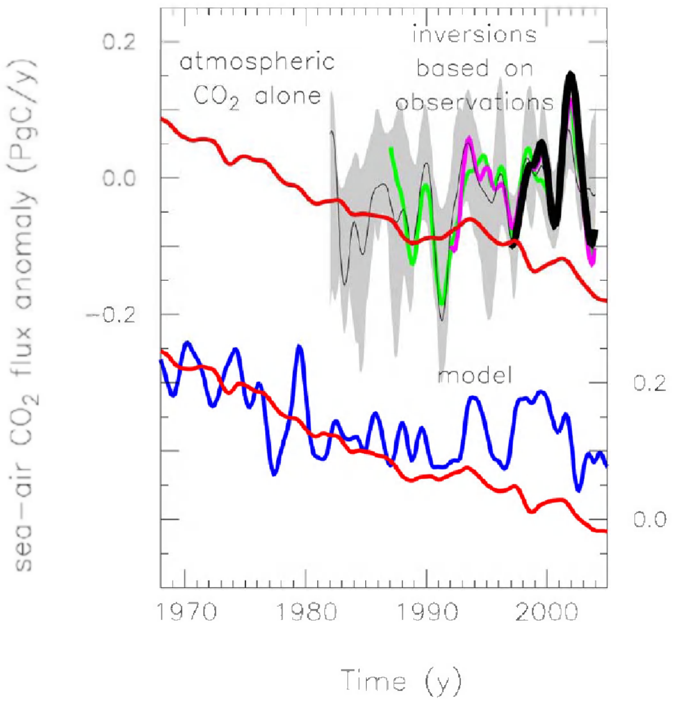


Figure 4.58 Sea-air CO_2 flux anomalies in the Southern Ocean (PgC/y, >45°S) based on atmospheric CO_2 data and inversed transport model (on top) and a global biogeochemical ocean model (bottom). Compared to experiments that do not take into account the climate variability (in red), both approaches suggest a stabilization or reduction of the ocean CO_2 sink since the 1980s (LeQuéré et al., 2007).

4.1.4 Historical Changes – Simulated View

To explore how changes in Southern Ocean air-sea CO_2 fluxes have responsed to historical climate change between 1948-2003. Matear and Lenton (2008) used a biogeochemical ocean model driven with observed changes (NCEP R1; Kalnay et al., 1996). They explored how the total carbon cycle as well as the natural and anthropogenic carbon uptake responded to the observed increases in windstress, heat and freshwater fluxes (Figure 4.59). Their results show a complex picture: when only either heat or freshwater fluxes increase, the total uptake is slightly greater than the total CO_2 flux response (with all fields increasing). In contrast, when only the wind speed increased, the total uptake was less than the total response. In addition, the anthropogenic response was always much smaller than the natural carbon cycle response and hence the natural response dominated the total response. The natural carbon sink dominates the total response over the last 50 years for two reasons: (i) because wind stress changes are larger than the corresponding changes in heat and freshwater flux, therefore changes in solubility are dominated by changes in ocean dynamics; and ii) the CO_2 gradient between the atmosphere and the ocean due to anthropogenic emissions is strong enough over this period to counter the increased CO_2 in surface waters caused by winds bringing up water rich in dissolved inorganic carbon from the deep ocean. Although there is some question of the validity of the changes in the pre-satellite era (1948-1979; Marshall, 2003), the largest changes occur in the later period 1979-2003, as seen in Figure 4.58.

As reanalysis products used in Matear and Lenton (2008) are assimilated products, all climate modes/variability are represented. Other studies using different ocean models and experiments to explore the response of the Southern Ocean air-sea CO_2 flux to the SAM alone e.g. Lenton and Matear (2007), Lovenduski et al. (2007) are very consistent with the view determined using all the superposition of all the climatic modes. This is not surprising given that these studies suggest that more than 40% of the total variance in CO_2 flux is be explained by the SAM in the recent period Lenton and Matear (2007).

Over the period only a small increase in primary production and export production is evident, suggesting only a weak link between atmospheric forcing and export production, despite the increased upwelling of deep waters in response to increased winds (i.e. supplying also macro and micro nutrients). The largest response was on the northern boundary of the High Nutrient Low Chlorophyll (HNLC) area of the Southern Ocean. Over the rest of the Southern Ocean, very little response was seen, demonstrating only a weak link between atmospheric forcing and production.

The increased ventilation of the Southern Ocean from simulations does not only alter the concentration of upper ocean CO_2 and the air-sea CO_2 fluxes, it also alters the carbonate chemistry of the upper ocean. These changes in carbonate chemistry affect the ability of the ocean to take up atmospheric CO_2, through changes in the Revelle factor (Revelle and Suess, 1957) and through the lowering of seawater pH. This pH reduction, or ocean acidification, reduces the ability of organisms that use calcite to build shells, potentially adversely impacting the marine ecosytem (Feely et al., 2004). In response to ocean acidification, a key carbon parameter is the aragonite saturation state (Ω_A), which influences the rate of calcification of marine organisms (Riebesell et al., 2000; Langdon and Atkinson, 2005). Simulations show that the observed increases and variability in heat, freshwater fluxes and in particular wind stress in the last 50 years, has moved the saturation horizon closer to the surface (Orr et al, 2005), potentially already impacting on the ecosystems in the Southern Ocean, although we do not yet have the observational evidence to support this hypothesis.

4.11.5 Changes in CO_2 inventories

In the previous sections, we focused on the changes of the air-sea CO_2 fluxes as observed and simulated over the last 50 years. When discussing the capacity of the ocean to reduce the impact of climate changes, the change in anthropogenic CO_2 in the water column since the pre-industrial era must be evaluated. This is important not only to estimate the global ocean's capacity to absorb anthropogenic CO_2 emissions, but also to detect the changes in carbonate saturation levels and the potential increase of ocean acidity, especially in the Southern Ocean (Feely et al., 2004; Orr et al., 2005). It has been estimated that in recent decades 30% of the total uptake since the preindustrial period has been taken up by Southern Ocean mode waters (Sabine et al., 2004). The anthropogenic CO_2 in the ocean (C_{ant}) cannot be directly measured, but under several assumptions, it can be derived from *in-situ* observations. This was first suggested by Brewer (1978) and Chen and Millero (1979), and in the last ten years several data-based methods have been investigated at regional and global scales (see a review in Wallace 2001; Sabine et al., 2004; Waugh et al., 2006; Lo Monaco et al., 2005). Comparisons of data-based methods (Lo Monaco et al., 2005) clearly show that all methods converge to estimate large inventories associated with mode and intermediate waters (Figure 4.60). Meanwhile in Southern Ocean uptake south of 50°S uncertainties in C_{ant} still exist, but these differences appear to reflect techniques that have been recognized to underestimate anthropogenic carbon in deep and bottom waters along the Antarctic coast (Lo Monaco, personal communication).

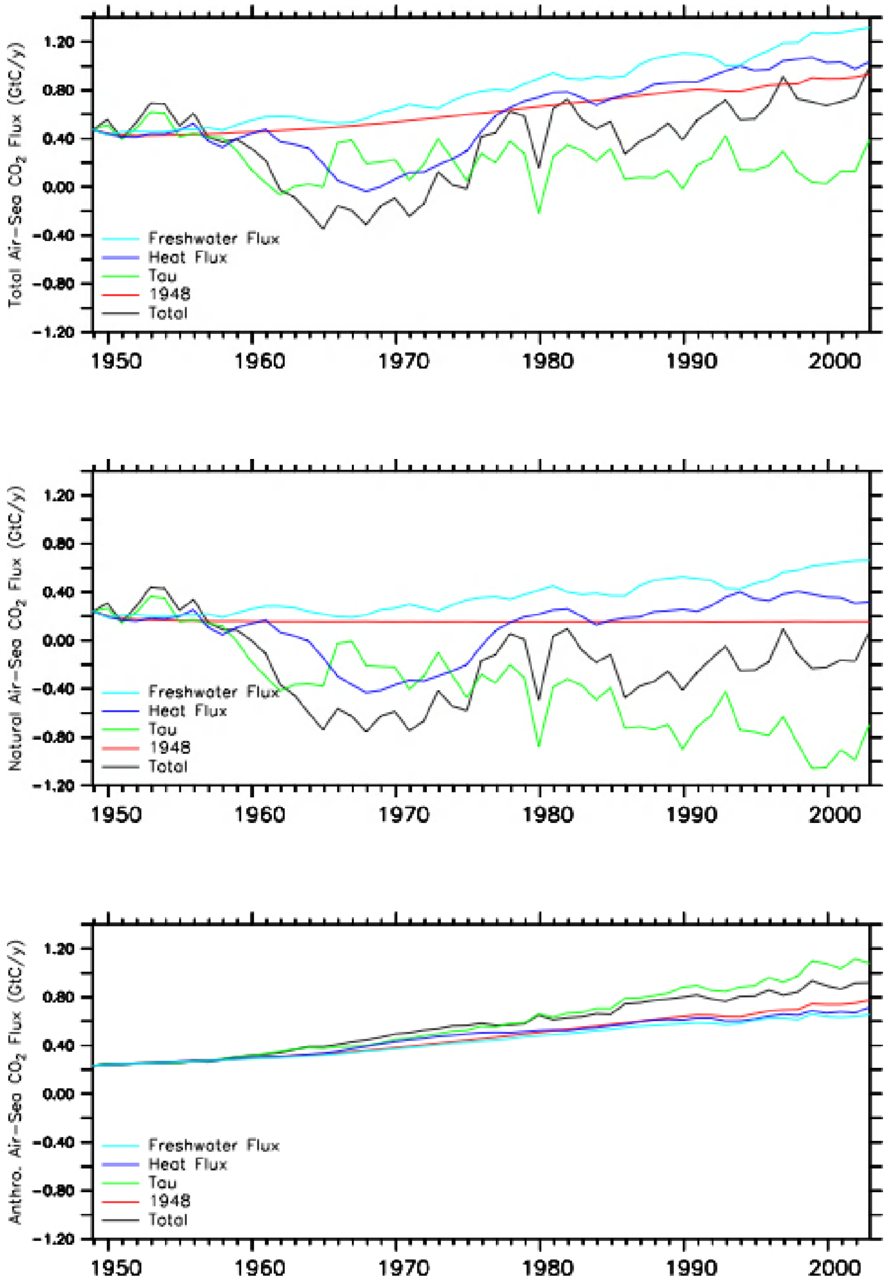


Figure 4.59 Annual-averaged Southern Ocean uptake of: a) total carbon; b) natural carbon; c) anthropogenic carbon. The different experiments use the following colour coding, total experiment (black line), 1948 (red line), wind stress (tau, green line), heat flux (blue line) and freshwater flux (cyan line).

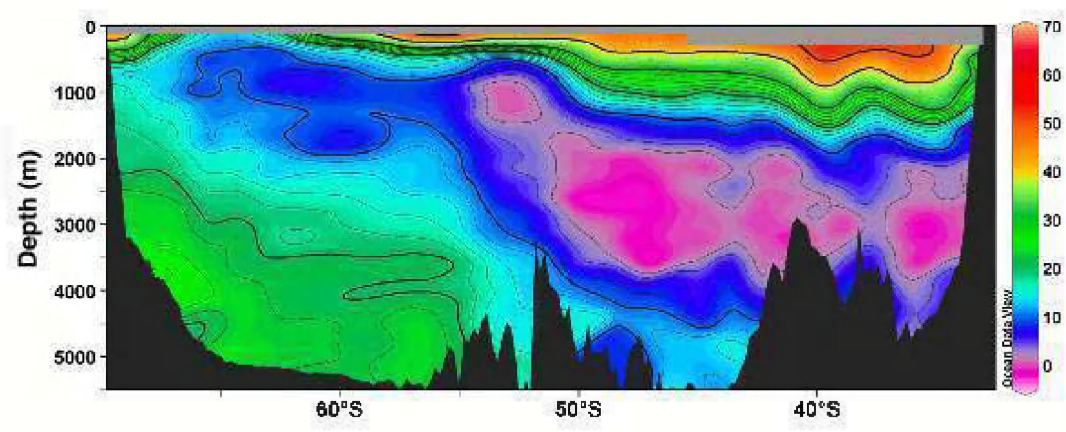


Figure 4.60 Anthropogenic carbon concentrations (colour scale, in umol/kg) derived using a back-calculation method in the Indian Ocean sector of the Southern Ocean between Antarctica (left) and Africa (right) (redrawn from Lo Monaco et al., 2005).

4.11.6 Concluding Remarks

The Southern Ocean plays a critical role in the uptake of atmospheric CO_2, accounting for more that 40% of the annual mean CO_2 uptake. Modelling and observational studies show that the Southern Ocean has undergone significant changes in the last 50 years; these views appear to be converging towards a coherent view. The largest change has been a reduction in the total CO_2 uptake in recent decades in response to the observed changes in climatic forcings, particularly changes in wind speed. The increased wind speed drives strong changes in the physical carbon pump, specifically through ocean dynamics, rather than through changes in solubility or in the Revelle factor; this view is further reinforced by only a weak response in the biological pump. In the future, in response to climate change, both the biological and physical pumps are expected to be impacted (see Section 5.8 for details).

4.12 Terrestrial Biology

Contemporary terrestrial and freshwater ecosystems within Antarctica occupy only 0.34% of the continental area (British Antarctic Survey, 2004), the remainder being permanent ice and snow. The combined land area of the isolated sub-Antarctic islands is likewise small. Individual areas of terrestrial habitat are typically 'islands', whether in the true sense of the word, being surrounded by ocean, or in the sense of being surrounded and isolated by terrain inhospitable to terrestrial biota in the form of ice (Bergstrom and Chown, 1999). While the most biologically developed and most studies of terrestrial exposures are found in coastal regions of the continent, particularly along the Antarctic Peninsula and in Victoria Land, terrestrial habitats exist in all sectors of the continent, and both in coastal and inland locations.

Terrestrial biological research within Antarctica has, however, been much more spatially limited, with major areas of activity restricted to the South Orkney and South Shetland Islands, Anvers Island, the Argentine Islands and Marguerite Bay along the Antarctic Peninsula/Scotia Arc, and the Dry Valleys and certain coastal locations in Victoria Land. Terrestrial and freshwater research along the continental Antarctic coastline has largely been limited to areas in the vicinity of the Schirmacher Oasis, Windmill Islands and Davis

Station, Casey Station, and mountain ranges in Dronning Maud Land. Sporadic biological records exist from more widely dispersed locations, but in most cases these relate to single short field visits to these locations, often by non-biologists or non-specialists. Indeed, there remain many instances where the only biological records available, or only species descriptions that exist, derive from the original exploring expeditions of the 'heroic era'. Even where terrestrial biological research is undertaken within a region or by a national operator, both taxonomic and process-based research coverage is extremely uneven across different regions or operators.

All Antarctic terrestrial ecosystems are simple in global terms, lacking or with low diversity in specific taxonomic or biological functional groups (Block, 1984; Smith, 1984; Convey, 2001). It is therefore likely that they lack the functional redundancy that is typical of more diverse ecosystems, raising the possibility of new colonists (arriving by both natural and, more recently, human-assisted means) occupying vacant ecological niches. Such colonists could include new trophic functions or levels, threatening the structure and function of existing trophic webs (Frenot et al., 2005, 2007; Convey, 2008). Responses of indigenous biota will be constrained by their typically 'adversity-selected' life history strategies, which have evolved in an environment where abiotic environmental stresses and selection pressures (i.e. properties of the physical environment) far outweigh in importance biotic stresses and pressures (i.e. competition, predation, etc.) (Convey, 1996).

The growth and life cycle patterns of many invertebrates and plants are fundamentally dependent on regional temperature regimes and their linkage with patterns of water availability (Convey et al., 2006). In detail, the interaction between regional macroclimate and smaller scale ecosystem features and topography define the microclimate within which an organism must live and function. There has to date been remarkably little effort to identify connections between macro and microclimatic scales, or to probe the application of large-scale macroclimatic trends and predictions at microclimatic scale. Distinct patterns in sexual reproduction are evident across the Antarctic flora and are most likely a function of temperature variation - indeed recent increase in the frequency of successful seed production in the two maritime Antarctic flowering plants (Convey, 1996) is proposed to be a function of warming in this region. In addition, phenology of flowering plants is cued to seasonality in the light regime. In regions supporting flowering plants, wind is assumed to play a major role in pollination ecology of grasses and sedges resulting in cross-pollination. The lack of specialist pollinators in the native fauna, combined with high reproductive outputs in non-wind pollinated species implies a high reliance on self-fertilisation.

The Antarctic biota shows high development of ecophysiological adaptations relating to cold and desiccation tolerance, and displays an array of traits to facilitate survival under environmental stress (Hennion et al., 2006). While patterns in absolute low temperatures are clearly important in determining survival, perhaps more influential is the pattern of the freeze-thaw regime, with repeated freeze-thaw events being more damaging than a sustained freeze event (Brown et al., 2004; Sinclair and Chown, 2005). How these patterns change in the future will be an area of major importance.

The response of Antarctic plants to increased UV-B radiation (280-315 nm) associated with the ozone hole, provides an illustration of another suite of ecophysiological/biochemical strategies. Reported responses vary widely between studies, ranging from negative effects on chlorophyll concentrations in tissues, on growth, and evidence of DNA damage in some species (e.g. Ruhland and Day, 2000; Xiong and Day, 2001; Robinson et al., 2005; Turnbull and Robinson, 2008), through little change (e.g. Björn, 1999; Lud et al., 2002; Boelen et al., 2006) to consistent positive effects, such as increased concentrations of UV-B screening pigments (Newsham et al., 2002). Previously it has been suggested that higher plants and bryophytes could differ in their abilities to synthesize UV-B screening pigments (Gwynn-Jones et al., 1999), but most recent data from Antarctic studies do not support this proposition (e.g. Newsham et al., 2002; Newsham, 2003; Newsham et al., 2005; Dunn and Robinson, 2006; Clarke and Robinson, 2008). The majority of Antarctic bryophytes studied have potential UV screening compounds inside their cells and/or attached to their cell walls (Clarke and Robinson, 2008), suggesting widespread UV screening potential in these species. A recent study has estimated that the cost of synthesising new protective pigment molecules on exposure to UV-B represents approximately 2% of the carbon fixated by a common Antarctic liverwort, analogous to estimates of 1-10% of biomass invested in cryoprotectants or desiccation protectants by many Antarctic terrestrial invertebrates and microbes (Snell et al., 2009).

	Entire sub-Antarctic	Maritime Antarctic	South Georgia	Marion	Prince Edward	Crozet	Kerguelen	Heard	Mac Donald	Macquarie
Dicotyledons	62	0	17	6	2	40	34	0	0	2
Mono-cotyledons	45	2	15	7	1	18	34	1	0	1
Pteridophytes	1	0	1	0	0	1	1	0	0	0
Total non-indigenous plants	108	2	33	13	3	59	69	1	0	3
Invertebrates	72	2-5	12	18	1	14	30	3	0	28
Vertebrates	16	0	3	1	0	6	12	0	0	6

Table 4.2 The occurrence of alien non-indigenous terrestrial species across Antarctic biogeographical zones (extracted from Frenot et al. (2005); see also Greenslade (2006) for a detailed description of established and transient alien species, and species recorded only synanthropically, from sub-Antarctic Macquarie Island).

A meta-analysis of the response of polar vegetation to UV-B radiation concludes that Antarctic bryophytes and vascular plants respond in a similar fashion to vegetation from other regions, with UV-B exposure leading to decreased above-ground biomass and height and increased DNA damage (Newsham and Robinson, 2009). Plants also appear able to protect themselves from elevated UV-B radiation through the induction of UV screening pigments (Newsham and Robinson, 2009), although this likely comes at a cost to biomass production (Snell et al., 2009). However this meta-analysis does suggest that the method by which UV-B radiation is applied to plants plays an important part in determining the strength of plant response to UV-B.

The final suite of life history traits includes elements relating to competition and predation. Their potential significance is illustrated by reference to ecosystem changes caused through the introduction of new predatory invertebrates to certain sub-Antarctic islands (e.g. Table 4.2). The introduction of carabid beetles to parts of South Georgia and Îles Kerguelen, where such predators were previously absent, is leading to extensive changes to local community structure, which threatens the continued existence of some indigenous and/or endemic invertebrates (Ernsting et al., 1995; Frenot et al., 2005, 2008). Regional warming has also been predicted to rapidly increase the impact of certain indigenous predators (Arnold and Convey, 1998). Providing an analogous impact within the decomposition cycle, detailed studies on Marion Island indicate that indigenous terrestrial detritivores are unable to overcome a bottleneck in the decomposition cycle, hence illustrating a further ecosystem service likely to be strongly influenced by recently introduced non-indigenous species (Slabber and Chown, 2002).

The lack of attention to these traits to date is unfortunate, particularly with respect to the understanding of alien species' impacts. It is already well known that Antarctic terrestrial biota possess very effective stress tolerance strategies, in addition to considerable response flexibility. The exceptionally wide degree of environmental variability experienced in many Antarctic terrestrial habitats, on a range of timescales between hours and years, means that predicted levels of change in environmental variables (particularly temperature and water availability) are often small relative to the range already experienced. However, as illustrated above with biochemical responses to UV-B exposure, any change in the balance of use of specific strategies carries a quantifiable cost, and carries implications for changes in the allocation of resources within the organism.

Given the absence of more effective competitors, predicted and observed levels of climate change may be expected to generate positive responses from resident biota of the maritime and continental Antarctic, and this is confirmed in general terms both by observational reports of changes in maritime Antarctic terrestrial ecosystems, and the results of manipulation experiments mimicking the predictions of climate change (Convey, 2003, 2007). Over most of the remainder of the continent, biological changes are yet to be reported, as might be expected given the weakness or lack of evidence for clear climate trends over the instrumental period. Potentially sensitive indicators of change have been identified amongst the biota of this region (e.g. Wasley et al., 2006), particularly in the context of sensitivity to changes in desiccation stress (Robinson et al., 2003). More local scale and short-term trends of cooling over recent decades in the Dry Valleys of Victoria Land have been associated with reductions in abundance of the soil fauna (Doran et al., 2002). The picture is likely to be far more complex on the different sub-Antarctic islands as, in addition to various different trends being reported in a range of biologically important variables, many also already host (different) alien invasive taxa, some of which already have considerable impacts on native biota (Frenot et al., 2005; Convey, 2007, Table 4.2).

The best-known and frequently reported example of terrestrial organisms interpreted to be responding to climate change in the Antarctic is that of the two native Antarctic flowering plants (*Deschampsia antarctica* and *Colobanthus quitensis*) (Figure 4.61) in the maritime Antarctic (Fowbert and Smith, 1994; Smith, 1994; Grobe et al., 1997; Gerighausen et al., 2003). At the Argentine Islands numbers of plants increased by two orders of magnitude between the mid 1960s and 1990 (Fowbert and Smith, 1994), although it is often overlooked that these increases have not involved any change in the species' overall geographic ranges, limited in practice by extensive ice cover south of the current distribution. These increases are thought to be due to increased temperature encouraging growth and vegetative spreading of established plants, in addition to increasing the probability of establishment of germinating seedlings. Additionally, warming is proposed to underlie a greater frequency of mature seed production (Convey, 1996b), and stimulate growth of seeds that have remained dormant in soil propagule banks (McGraw and Day, 1997). However, since 1990, there has been no further increase in the Argentine Islands populations, while there has also been no significant warming trend in either the annual or seasonal air temperature data record at this location over the period 1990-2008, which might suggest the link between environmental conditions and plant responses is even closer than initially thought (Parnikoza et al., in press).

Figure 4.61 The only two flowering plants native to the Antarctic continent are both restricted to the Antarctic Peninsula. The grass *Deschampsia antarctica* may develop into swards covering several to tens of square metres (left), while the pearlwort *Colobanthus quitensis* more typically is encountered as individual cushions (right) (photos P. Convey).

Changes in both temperature and precipitation have already had detectable effects on limnetic ecosystems through the alteration of the surrounding landscape and of the time, depth and extent of surface ice cover, water body volume and lake chemistry (with increased solute transport from the land in areas of increased melt) (Quesada et al., 2006; Lyons et al., 2006; Quayle et al., 2002, 2003. The latter authors highlight that some maritime Antarctic lake environmental changes actually magnify those seen in the atmospheric climate, highlighting the value of these locations as model systems to give ‘early warning’ of potential changes to be seen at lower latitudes. Predicted impacts of such changes will be varied. In shallow lakes, lack of surface ice cover will lead to increased wind–induced mixing and evaporation and increases in the diversity at all levels of the ecosystem. If more melt water is available, input of freshwater into the mixolimna of deeper lakes will increase stability and this, associated with increased primary production, will lead to higher organic carbon flux. Such a change will have flow–on effects including potential anoxia, shifts in overall biogeochemical cycles and alterations in the biological structure and diversity of ecosystems (Lyons et al., 2006).

Alien microbes, fungi, plants and animals, introduced directly through human activity over approximately the last two centuries, already occur on most of the sub-Antarctic islands and some parts of the Antarctic continent (Frenot et al., 2005, 2007; Greenslade, 2006; Convey, 2008, Table 4.2). The level of detail varies widely between locations and taxonomic groups (although at the microbial level, knowledge is virtually non-existent across the entire continent). On sub-Antarctic Marion Island and South Atlantic Gough Island it is estimated that rates of establishment through anthropogenic introduction outweigh those from natural colonization processes by two orders of magnitude or more. Introduction routes have varied, but are largely associated with movement of people and cargo in connection with industrial, national scientific programme and tourist operations. Although it is rare to have a record available of a specific introduction event, and there are undoubtedly instances of natural colonization processes resulting in new establishment, the impact of undoubted human-assisted introductions to some sub-Antarctic islands (particularly South Georgia, Kerguelen, Marion, Macquarie) is substantial and probably irreversible. Thus a range of introduced vertebrates and plants have led to large shifts in ecosystem structure and function, while in terms of overall diversity some islands now host a greater number of non-indigenous than indigenous species of plant. The large majority of aliens are European in origin.

Chapter 5

The Next 100 Years

Chapter Editor:

Colin Summerhayes

Authors:

David Barnes, Dana Bergstrom, Robert Bindschadler, James Bockheim, Greg Bodeker, Laurent Bopp, Tom Bracegirdle, Steve Chown, Pete Convey, Guido di Prisco, Eberhard Fahrbach, Jaume Forcada, Yves Frenot, Hugues Goosse, Julian Gutt, Dominic Hodgson, Ad Huiskes, Anna Jones, Rebecca Leaper, Wouter Lefebvre, Andrew Lenton, Amanda Lynch, Nicolas Metzl, Alison Murray, Siobhan O'Farrell, Lloyd Peck, Hans Pörtner, Howard Roscoe, Daniel Smale, Victor Smetacek, Colin Summerhayes, John Turner, Ann Vanreusel, David Vaughan, Cinzia Verde, Zhaomin Wang

5.1 Introduction

Determining how the environment of the Antarctic will evolve over the next century presents many challenges, yet it is a question of great interest for both scientists and policymakers concerned with issues as diverse as sea-level rise and fish stocks. A major problem is that we still have a poor understanding of the mechanisms behind many of the changes observed in recent decades. This is particularly the case in the ocean where we have few long time series of physical measurements and remarkably few spatially and temporally well separated observations of marine biota.

The evolution of the Antarctic climate over the next 100 years can only be projected through the use of coupled atmosphere-ocean-ice models. These have their limits in correctly simulating the observed changes that have taken place already, so there is still a degree of uncertainty about the projections from models, particularly on the regional scale. The models used in the IPCC Fourth Assessment Report (AR4) (IPCC, 2007) gave a wide range of projections for some aspects of the Antarctic climate system. In the case of sea ice extent that was not entirely surprising, since this is very sensitive to changes in both atmospheric and oceanic conditions. The IPCC report took a mean estimate from 20 Atmosphere-Ocean General Circulation Models (AOGCMs) without regard to how individual models performed globally or regionally. The model projections for temperature and precipitation were used to estimate how much sea level would rise under various greenhouse gas emission scenarios. In the following pages we show how that 'blunderbuss approach' can be improved upon. With a quantity such as near-surface air temperature it is possible to use the projections from the various models to derive various estimates of how temperature may change over land and ocean.

The atmospheric component of the next generation of climate models must have interactive chemistry if we are to obtain meaningful projections of the future extent and thickness of the ozone layer over Antarctica, which will clearly impact future surface climate in Antarctica, as shown by Perlwitz et al. (2008).

An extremely important question for both scientists and policy makers, since it determines how sea level may change, is how the Antarctic ice sheet will change over coming decades. Models can estimate how the precipitation onto the continent might change, but the current generation of ice sheet models cannot help us regarding possible dynamical changes that might occur in the flow of glaciers and ice streams, or ocean-induced melting at the ice margin.

Gauging how Antarctic terrestrial and marine biota might respond to rising temperature is also a major challenge, as is gauging the response of ocean biota to ocean acidification. Some laboratory experiments have been carried out into the survival of biota when subject to thermal stress, as have field manipulations mimicking some predicted elements of climate change, but conditions in the Antarctic involve many complex feedbacks and interactions that cannot be well replicated under either approach. Furthermore, numerically based biological models cannot yet approach the relative sophistication of models of the physics of the climate system, while physical models do not approach the high level of spatial scale or temporal resolution required for application to biological systems, providing yet a further limitation on what can be said about future change in the context of biotic and ecosystem responses.

In this chapter we consider how the physical environment of the Antarctic might change over the next century, and assess how the biota may respond.

5.2 Climate Change

5.2.1 IPCC scenarios

5.2.1.1 Introduction

Given the information about climate contained in the reports of the IPCC (IPCC, 2007) the degree to which the climate of the Earth will change over the next century will be heavily dependent on the success of efforts to reduce the rate of greenhouse gas (GHG) emissions. The Antarctic is a long way from the main centres of population, but greenhouse gases are well mixed through the atmosphere. Even with sharp reductions now in GHG production, it will take a long time for the levels of GHGs in the atmosphere to decrease. For instance, even if anthropogenic emissions of CO_2 were halted now, recent studies indicate that 25% of CO_2 from fossil fuels will persist in the atmosphere indefinitely (Archer, 2005; Mathews and Calderia, 2008), and Solomon et al. (2009) show that the climate change resulting from increases in CO_2 concentration in the atmosphere is largely irreversible for 1,000 years after emissions stop.

Future levels of GHG emissions will be determined by the complex interactions between many factors, such as changes in the energy mix of oil, gas, nuclear and renewables, the drive to greater fuel efficiency, the rise in population (which will increase energy demand and is likely to negate efficiency gains), the development and commercial application of new technologies (the so-called hydrogen economy; fuel cells etc), the extent of de- or re-forestation, and regionally varying social and economic developments - notably the growth of major economies in China, India and Brazil, and the gradual industrialisation and urbanisation of the developing world. The path of future evolution of GHGs and aerosols is therefore uncertain. Nevertheless, some assumptions have to be made for the purposes of determining how the Antarctic may be affected by climate change. To simulate the climate of the Twenty First Century through mathematical models requires first the selection of likely GHG emissions from a range of possible emission scenarios.

5.2.1.2 The IPCC greenhouse gas and aerosol emission scenarios

In the year 2000, the IPCC refined the emission scenarios that it had used until then as contributions to models of future climate change (Nakicenovic et al., 2000). The new scenarios were based, in IPCC terminology, on four storylines, which are narrative descriptions that highlight the main characteristics and dynamics of future economic and social development. The storylines (Nakicenovic et al., 2000) are:

- A1: a world of very rapid economic growth, global population that peaks in the middle of the Twenty First Century and declines thereafter, and rapid introduction of new and more efficient technologies.
- A2: a heterogeneous world with continuously increasing global population and regionally oriented economic growth that is more fragmented and slower than in other storylines.
- B1: a convergent world with the same global population as in the A1 storyline, but with rapid changes in economic structures toward a service and information economy, with reductions in material and energy intensity, and the introduction of clean and resource-efficient technologies.
- B2: a world in which the emphasis is on local solutions to economic, social, and environmental sustainability, with continuously increasing population (lower than A2) and intermediate economic development.

For each of the above storylines a 'family' of scenarios was developed, which include quantitative projections of major driving variables, such as possible population change and economic development. Six groups of scenarios were drawn from the four families: one group each in the A2, B1 and B2 families, and three groups in the A1 family. These characterised alternative developments of energy technology as A1FI (fossil intensive), A1T (predominantly non-fossil) and A1B (balanced across energy sources). Altogether IPCC developed 40 different scenarios, each considered equally valid and equally probable.

In the present report we use outputs from Scenario A1B, because climate model projections of the Twenty First Century using this scenario produce global mean near-surface warming that is about halfway between the output of simulations based on the warmest (A1F1) and coolest (B1) scenarios. Under this scenario emissions increase from the 1990 value of about 8 Gt/yr to a peak around the middle of the century and then decrease to about 14 Gt/yr by 2100. Concentrations of CO_2 are projected to rise from the current concentration of around 380 ppm to reach 720 ppm in the year 2100 (i.e. approximately a doubling of CO_2).

5.2.1.3 The emission scenarios in the 2007 IPCC AR4 models

In the IPCC's AR4 in 2007, climate model simulations were run using prescribed concentrations of GHGs (CO_2, CH_4 and N_2O) following the scenarios of Nakicenovic et al. (2000). As well, the scenarios for the future, two other emissions experiments were conducted: (i) a "climate of the Twentieth Century" (20C3M) modelling experiment, which was forced by observed GHG concentrations from the mid Nineteenth Century to the end of the Twentieth Century; and (ii) a "pre-industrial control" (PICNTRL) modelling experiment for which GHG and aerosol concentrations were kept at constant pre-industrial levels. For both the 20C3M and the "pre-industrial control" scenario runs the sulfate aerosol concentrations were calculated from prescribed sulfur dioxide (SO_2) emissions, but varied between models due to different methods of calculation. Both sets of runs also included inter-model differences of forcing by stratospheric ozone and volcanic aerosols, which are important for the Southern Hemisphere circulation due their role in forcing long-term changes of the Southern Annular Mode (SAM) (Shindell and Schmidt, 2004; Miller, et al., 2006). For the A1B scenario, most models were run with a gradual recovery of stratospheric ozone to pre-industrial levels over the Twenty First Century, but some did not include any stratospheric ozone forcing (see Miller, et al., 2006).

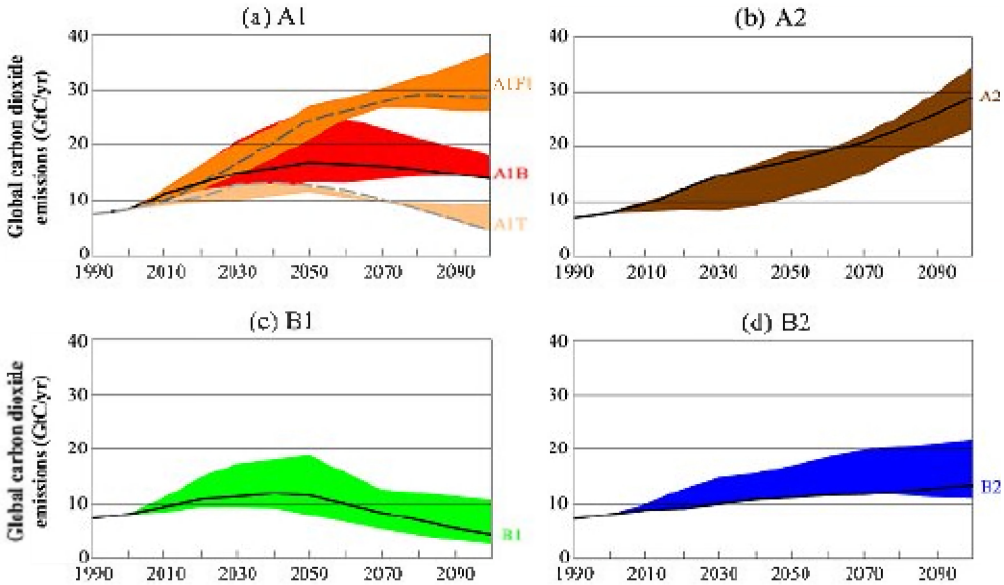


Figure 5.1 Total global annual CO_2 emissions from all sources (energy, industry, and land-use change) from 1990 to 2100 in Gigatonnes of carbon/year (GtC/yr) for the four emission families (A1, A2, B1 and B2) and the six scenario groups (A1F1, A1T, A1B and A2, B1 and B2) ($1Gt = 10^9$ metric tonnes). The 40 scenarios are represented by the four families (A1, A2, B1, and B2) and the six scenario groups. The fossil-fuel-intensive A1FI (comprising the high-coal and high-oil-and-gas scenarios) has the greatest emissions. Each coloured emission band shows the range of scenarios within each group. The solid and dashed lines within each scenario represent the paths of illustrative marker scenarios.

In this present volume (as explained below) we have weighted the model outputs to reflect the skill of each model at reproducing modern conditions in the 20C3M experiment. Although the weighted results include models that do not include stratospheric ozone recovery, four out of five of the most strongly weighted models do include representation of stratospheric ozone recovery. The consequences of neglecting stratospheric ozone recovery are significant only in the summer months when ozone concentrations are lowest.

We used 19 of the 25 AR4 coupled AOGCMs, which had the data required for weighting. The models used here were as follows: BCCR BCM2, CCCMA CGCM3, CNRM CM3, CSIRO Mk3, GFDL CM2.0, GFDL CM2.1, GISS EH, GISS ER, IAP FGOALS1, INM CM3, IPSL CM4, MIROC (hires), MIROC (medres), MPI ECHAM5, MRI CGCM2, NCAR CCSM3, NCAR PCM1, UKMO HadCM3, UKMO HadGEM1.

5.2.2 Climate models

Climate models are now being developed within the framework of Earth System Science, where the aim is to build a comprehensive picture of the feedbacks between the atmosphere, hydrosphere, cryosphere, biosphere and geosphere in the Earth System. Given the likely effects on global sea levels of melting ice at the poles, considerable attention is now being paid to modelling high latitude climate processes accurately. In that context a particular difficulty lies in our ability to validate models and model predictions, for which we require specific field and ocean observations that are commonly missing at these latitudes. Without those observations it is difficult to comprehensively verify whether or not the internal variability displayed by the different runs of the climate models is similar to that seen in reality. We face the further problem that because of the broad scale of climate model grids, we are not yet able to model with confidence variation at the regional to local scale, for example over the Antarctic Peninsula. Nevertheless, what models do provide, especially when tested against their ability to reproduce (or rather, simulate) modern conditions, is a reasonable, albeit crude, projection of what may happen given certain external forcings (e.g. changing carbon dioxide in the atmosphere).

The results from 24 climate models contributed by many modelling centres around the world were compiled into a central database for the IPCC AR4. This repository of model data forms the Coupled Model Inter-comparison Project phase 3 (CMIP3) multi-model dataset. Much of the discussion in this chapter is based on research that makes use of the CMIP3 dataset. This section therefore provides some background to these models and their strengths and weaknesses.

The CMIP3 climate models are coupled atmosphere/ocean/sea-ice/land-surface models. Each model is made up of sub-models that simulate either the atmosphere, ocean, sea ice, or land surface. These sub-models are coupled together to allow interaction between the systems. These interactions are complex and in some ways still not well understood.

5.2.2.1 Atmosphere models

Models of the atmosphere are founded on the fundamental laws of motion, thermodynamics and chemistry. The models use equations representing those laws to step forward in time from some initial state. The result is a representation of the evolution of the atmosphere in both space and time.

For the CMIP3 models the horizontal spacing of grid points is around 250 km and the vertical spacing is around 500m. This is sufficient to resolve features such as the typical atmospheric eddy – the extra-tropical cyclone – which can be around 1,000 km across. However, many smaller-scale phenomena, such as atmospheric convection and heat and momentum exchange at the lower boundary, are not resolved in these models. The effects of these small-scale processes on the larger scale must be parameterized (i.e. represented in the models statistically). Many of the parameterizations are developed and optimised for mid and low latitudes and therefore may not always be appropriate for the polar regions.

One example is parameterizations of the atmospheric boundary layer. These parameterizations perform poorly in the polar regions, where a very stable boundary layer is often present over snow and ice covered surfaces. Parameterizations based on observations of more weakly stable boundary layers at lower latitudes are often not suitable for such conditions. As a result, where conditions are very stable and stratified, the computed fluxes of momentum, heat and water vapour are often too small.

Climate models also do not include parameterizations for polar stratospheric clouds (PSC), which are the thin, tenuous clouds found in the stratosphere. Yet PSCs are known to be important for the radiation balance and temperature of the atmosphere between the mid/upper troposphere and the lower stratosphere.

Some of the CMIP3 models do not include the effects of stratospheric ozone, which has a significant impact on their representation of Southern Hemisphere wind patterns (Miller et al., 2006). Incorporating stratospheric ozone affects the strength of the circumpolar westerlies in climate models as expected, given that we now understand that the creation of the ozone hole helped to increase the strength of the circumpolar westerlies.

5.2.2.2 Ocean models

Ocean models also use a set of equations to represent the evolution of the ocean in response to climate change at specific time and space intervals. Baroclinic eddies in the ocean are smaller than those in the atmosphere, averaging around 100 km across, and to capture their variability the resolution of these models has to be higher than for atmosphere models. The grid spacing for the CMIP3 ocean models is around 100 km in the horizontal; as a result they do not simulate ocean eddy behaviour well. This is an important constraint given the key role of ocean eddies in north-south heat transport. This grid spacing is also too coarse to accurately represent the effect of narrow topographic ridges on the seabed in steering currents.

A particularly challenging phenomenon at high latitudes is the calculation of heat and moisture fluxes from the ocean to the atmosphere during cold air outbreaks from the continent. During such events very cold polar continental air can come into contact with relatively warm ocean surfaces. The flux parameterizations are difficult to verify due to the challenge of making observations in such conditions.

5.2.2.3 Sea ice

The formation and melting of sea ice is a complex process and has important feedbacks onto the ocean and atmosphere. Most sea ice models include simplistic representation of the thermodynamic energy transfer between ocean and atmosphere. The most notable difference between the sea ice models in the CMIP3 set is in their treatment of deformation and flow (rheology). The rheology used ranges from simple ocean drift models to more advanced Elastic Viscous Plastic (EVP) schemes. For more details see page 606 of the AR4 Working Group 1 report (IPCC, 2007).

5.2.2.4 Terrestrial ice and snow

Most global climate models (including the CMIP3 models) represent non-interactively the dynamics of the large ice sheets that cover much of Antarctica. Their presence is implicit in the land surface orography and surface albedo. There is also no explicit representation of glaciers. Some modelling studies covered by IPCC used ice sheet models run offline and forced by output from climate models to assess future mass balance evolution (e.g. Gregory and Huybrechts, 2006). However, these ice sheet models do not currently include the effects of moving ice sheets and glaciers. The discharge of ice into the ocean due to dynamical effects (mechanical breakup of ice sheets and glaciers plus lubrication of the bed by subsurface hydrological systems) is a large uncertainty at present, but must be addressed through models in due course because of its potential impact on sea-level rise (Rignot et al., 2005; Pfeffer et al., 2008).

5.2.2.5 Regional climate models

In order to examine regional climate, stretched-grid global climate models and nested regional models can be used to provide the benefits of high resolution modelling in a region of interest without the computational expense of running a climate model with high resolution globally. One key improvement that can be gained from a regional climate model is the representation of the effects of steep and high orography, especially in mountainous areas like the Antarctic Peninsula, or along steeply rising coasts. This can give improved regional detail of precipitation (Berg and Avery, 1995) and winds (van Lipzig et al., 2004b). Due to the sparse observation network over Antarctica it is difficult to determine whether or not regional climate models might improve Antarctic-wide precipitation projections.

5.2.2.6 Model evaluation

It is difficult to evaluate the performance of a climate model, because long-term projections cannot be checked against observations. As a substitute, the performance of climate models is assessed by having them simulate the climate of the Twentieth Century and comparing their output with observations (see chapter 8 of the AR4 report, IPCC, 2007). Even so, there is no guarantee that a climate model that does a good job of replicating the observed mean climate will be realistic in terms of its sensitivity to future forcing. At high latitudes differences in surface temperature between simulations and observations are significantly correlated with projected temperature change under future scenarios (Räisänen, 2007). This may be caused, at least in part, by sea ice biases (differences in sea ice concentration between simulations and observations of the present day), which have a strong impact on projected regional changes in the sea ice zone. This implies that for reliable projections of future changes an accurate simulation of current climate is particularly important at high latitudes.

5.2.2.7 Sea ice in the CMIP3 models

All models produce a seasonal cycle of sea ice with a peak in approximately the right season, though HadCM3 is a month late and NCAR CCSM two months early. IAP FGOALS has vastly over extensive ice, so that even the summer minimum would be off the scale used for the other plots. A more effective way to rank the relative success of the different models against observations is to use a measure based on the pointwise root mean square difference from the statistical average (the climatology). This shows that the MRI, CSIRO, HADGEM and MIROC_hires models are the best, although even the best scores are low. Clearly, a good simulation of Antarctic sea ice is a difficult challenge for a GCM. Most of the models use a viscous-plastic (VP) or EVP rheology; CSIRO uses cavitating fluid; and HADCM3 and MRI implement "ocean drift". Only the INM model has no ice advection. The best performing model is MRI, which has the most primitive "rheology". However, the MRI model is flux-corrected globally, and this is likely to strongly affect the sea ice simulation. The next best, CSIRO, uses the relatively simple cavitating fluid rheology. This illustrates the fact that many aspects of the model simulation besides sea ice model quality go into making up the simulation of the sea ice. If all else is equal, a more sophisticated and physically plausible scheme would be preferable. Parkinson et al. (2006) note that some of these models, especially CSIRO, show rather lower skills in the Northern Hemisphere, and suggest that there may be some tuning to one hemisphere or the other; we have only examined the Southern Hemisphere in this report.

The unweighted model average displays significantly higher skill (0.42) than any of the individual models (Connolley and Bracegirdle, 2007), presumably due to cancellation of errors (Parkinson et al., 2006). The implication is that it is better to use a multi-model average (even un-weighted) than to use just the single best model. This is akin to using the ensemble approach in weather forecasting. It is accepted that the models have some severe limitations (not least because of their coarse resolution), and that the process of averaging may not produce a more accurate result. Despite these caveats the models are likely to be improvements on simple projections of current trends, as they take a large number of parameters into consideration in an integrated way.

5.2.2.8 Temperature

Station observations (Turner et al, 2005a) show a maximum increasing temperature trend since 1958 on the west side of the Antarctic Peninsula, with smaller and generally non-significant changes around East Antarctica. Much of West Antarctica has no station observations and therefore satellite data have been interpolated to assess temperature changes in the region. Ongoing studies show significant warming trends that extend beyond the Peninsula region across most of West Antarctica (Monaghan et al., 2008; Steig et al., 2009). Since long-term observations are available for temperature trends, and because trends are less stable than means over shorter periods, we use data from 1960 onwards to evaluate the temperature trends. The un-weighted model average, for June – August (JJA) of 1960-1999, reproduces a maximum warming in winter over the Peninsula, but not extending over West Antarctica (Figure 5.2).

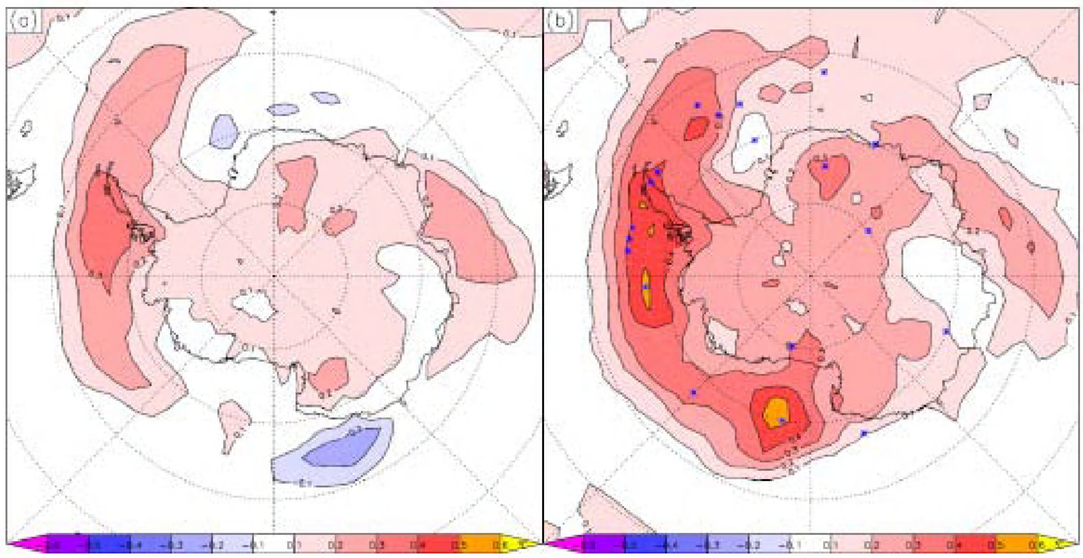


Figure 5.2 Temperature trends in ^{o}C/decade from 1960–2000 for winter (JJA). (a) Unweighted average of 19 models. (b) Weighted average, using weightings accorded to the individual models on the basis of their skill at reproducing modern conditions. Blue dots in (b) are the locations of the maximum trends from the individual models; four models position the maximum trend west of the Peninsula; two to the east of the Peninsula; four in the Weddell Sea; three in the seas around East Antarctica; three over the continent itself (although the magnitudes of these three trends are small); one on the Ross ice shelf and two in the Ross Sea (adapted from Connolley and Bracegirdle, 2007).

The weighted average maximum winter trend is most similar to observations and reconstructions, with generally more warming over West Antarctica and the Peninsula compared to East Antarctica (Figure 5.2a and b). The trend over the Peninsula (at 65°S, 70°W) is 0.38°C/decade (unweighted) or 0.45°C/decade (weighted), both of which are smaller than the observed value at Faraday/Vernadsky of approximately 1°C/decade. In this context, the observed winter temperature at Faraday/Vernadsky is highly variable and the trend depends strongly on the exact start date chosen. In the weighted average (Figure 5.2b) the trends around East Antarctica increase somewhat but remain fairly small; warming over the continent itself increases somewhat. Observations show small and insignificant cooling at the pole, and smaller and insignificant warming at Vostok.

There is a large scatter in the regional trends seen in different climate models (see Figure 2 of Connolley and Bracegirdle, 2007). However, there is also a large scatter within different runs of a single model, e.g. MPI ECHAM5 with the same external forcing; Figure 5.3 shows winter (JJA) surface temperature trends from four different ensemble runs of ECHAM5/MPI-OM all using the same Twentieth Century anthropogenic forcings. The differences are due to the internal variability of the model. Two runs (run 1 and run 3) show large winter warming to the west of the Peninsula as seen in observations. However, run 2 shows maximum trends over the Ross Sea and relatively small trends to the west of the Peninsula. There are also differences over the high interior of Antarctica, with some runs showing a slight warming and others a slight cooling. This indicates that the internal variability of both climate models and the real world may contribute to differences between observations and climate models on a regional scale. That makes it difficult to verify climate model simulations of regional-scale variability such as Antarctic Peninsula temperature changes. All the large model trends are over the sea ice rather than the continent, and are closely related to sea ice changes. In this they are behaving realistically, in that the observed winter trends around the Peninsula are believed to be reinforced by sea ice feedbacks (King, 1994).

5.2.2.9 Surface Mass Balance

Estimates of Antarctic surface mass balance (SMB) vary (Uotila et al., 2007). For this report we use a central value of 167 mm/yr water equivalent from Vaughan et al. (1999) with a spread of 30 mm/yr, which recognises the considerable variability in estimates from observations and models, and also allows for interannual variation. Typically, models indicate that this SMB is made up mostly of precipitation, with sublimation removing approximately 10-20%. Other studies show that blowing snow and melt (which are ignored here) are small on a continental scale. Nine models have SMB within 15 mm/yr of 167. IAP_FGOALS greatly overestimates (500 mm/yr final value); the GISS models (despite having a large value for sublimation) and MRI overestimate by about 100 mm/yr, but for different reasons: GISS's have a "central desert" area that is too small; whereas MRI does not simulate the very low values of SMB in the interior. Only the MIROC_medres model (116) and HADGEM model (131) substantially underestimate SMB. Of those models that do well on overall totals, two (BCCR and CNRM) produce SMB simulations that are implausible. They fail to produce large (> 500 mm/yr) SMB on and around the coast of East Antarctica.

5.2.3 Atmospheric circulation

The atmospheric circulation over Antarctica and the Southern Ocean is critical for the future evolution of global climate in a number of ways. Perhaps most important is the role of circulation in defining accumulation over the Antarctic ice sheet. Other aspects of importance include the role of circulation in the warming of the Antarctic Peninsula, the distribution of sea ice, and the seasonal to interannual variability of the Southern Hemisphere.

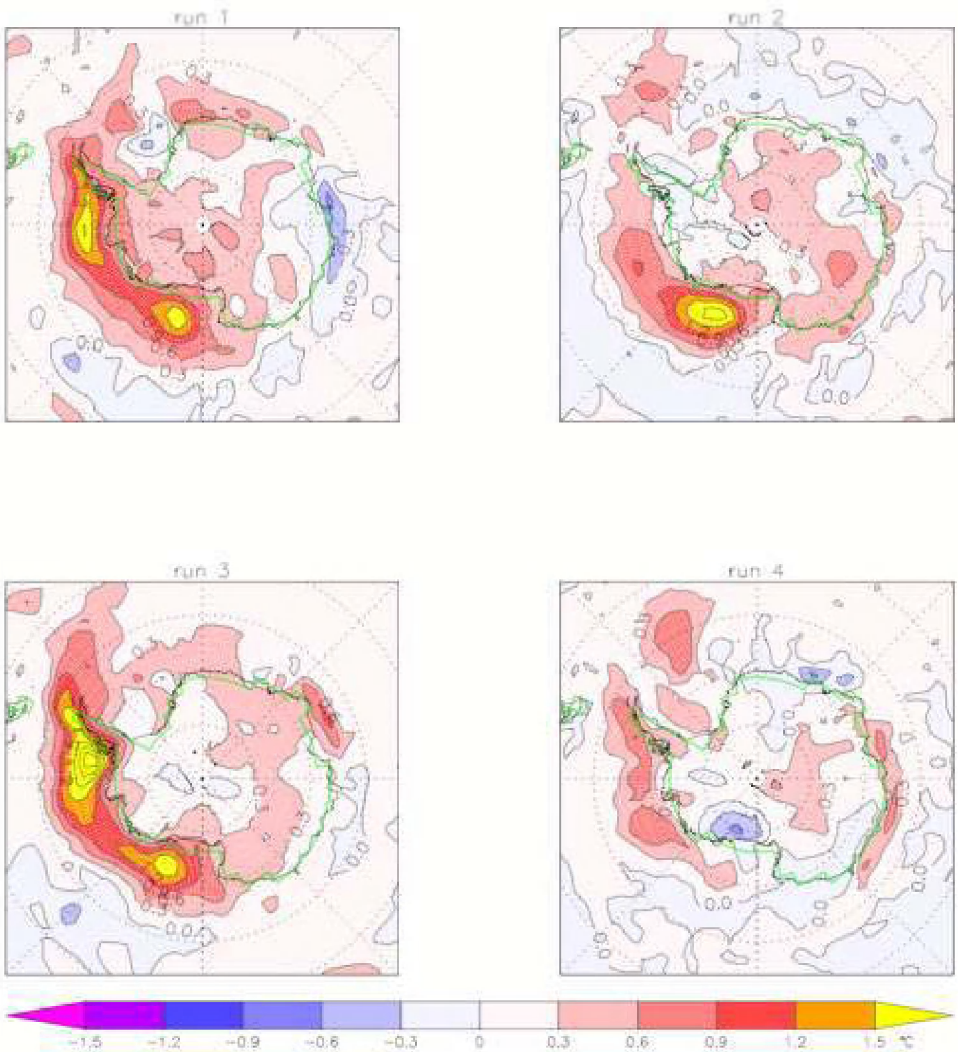


Figure 5.3. Surface temperature trends in °C/decade from 1960–2000 for winter (JJA) for the MPI ECHAM5 model. The four different runs all follow the IPCC climate of the Twentieth Century (20C3M) experiment and use the same anthropogenic forcings.

5.2.3.1 The role of modes of circulation variability

Mean sea level pressure changes associated with the long-term variability of circulation of the Southern Hemisphere have been reported for many decades (e.g. van Loon, 1967; Hurrell and van Loon, 1994). More recently, analyses of sea level pressure have revealed secular decreases over the Antarctic, associated with increases in mid-latitude westerlies, a poleward displacement of the polar front jet stream, and a more zonal circulation. The robustness of these observed changes is open to question, given that spurious trends are evident in reanalysis products (Bromwich et al., 2007). Further, there are significant differences between the ERA-40 and NRA re-analyses in the Southern Hemisphere, in which the ERA-40 has a tendency towards more intense cyclones in all seasons (Wang et al., 2006). Nevertheless, tendencies consistent with these changes in circulation have been demonstrated in station-based data and climate model realizations of the Twentieth and Twenty First Centuries.

Given the large inter-annual variability of the high southern latitudes, secular trends must be put in the context of responses to changes in the modes of variability that are dominant in the region, namely the El Niño Southern Oscillation (ENSO) and the Southern Annular Mode (SAM) (Turner, 2004; Sen Gupta and England, 2007). The long-term changes forced by increases of GHG concentrations have been found to possess a similar spatial pattern to the observed short-term (month to month) variability (Brandefelt and Källén, 2004). Of less significance, due largely to its contested nature (Park Y. et al., 2004), is the Antarctic Circumpolar Wave (ACW) (White and Peterson, 1996), a postulated pattern of variability with an approximately four-year period characterised by the eastward propagation of anomalies in sea ice extent.

The Southern Oscillation Index has a negative trend over recent decades, corresponding to a tendency towards more frequent El Niño conditions in the equatorial Pacific. This trend is associated with negative sea ice cover anomalies in the Ross and Amundsen Sea and positive sea ice anomalies in the Bellingshausen and Weddell Seas (Kwok and Comiso, 2002).

The SAM index has a positive trend over recent decades during the summer and winter, which reflects trends in the zonally averaged mid-latitude westerlies. The increases in the SAM index are associated with strong warming on the eastern side of the Antarctic Peninsula and low pressure west of the Peninsula (Orr et al., 2004; Lefebvre et al., 2004) – this reflects increased poleward flow, resulting in both the eastern Peninsula warming and reduced sea ice in the region (Liu et al., 2004). The SAM index trends are also related to the observed and projected (in the near term) East Antarctic surface cooling (Shindell and Schmidt, 2004; Marshall, 2007). However, changes in the SAM are not thought to be responsible for the large winter season warming on the western side of the Antarctic Peninsula.

Progress in simulating ENSO variability has led to significant improvements in representing the spatial pattern of sea surface temperatures in the equatorial Pacific (Randall et al., 2007). Uncoupled models have demonstrated similar ENSO variability to that observed (e.g. Marshall et al., 2007). However, serious discrepancies remain in the attempts of coupled models to represent the ENSO (Joseph and Nigam, 2006). Atmosphere-ocean interaction leads to inaccuracies in the sea surface temperature and in the structure of the thermocline (Cai et al., 2003; Davey et al., 2002). Furthermore, the timescale of variability in the coupled system is generally too short (van Oldenborgh et al., 2005), although in some models a peak at around 7 years is observed (Marshall et al., 2007). The interaction between climate change and ENSO variability is also subject to substantial uncertainty, with no coupled model consensus on the likelihood of a relationship between more frequent El Niño conditions and increasing GHG concentrations (van Oldenborgh et al., 2005; Collins et al., 2005; Wang, 2007).

Model outputs submitted to the IPCC AR4 simulate the SAM with a high degree of accuracy (e.g., Miller et al., 2006), generally with spatial correlations greater than 0.95 (Randall et al., 2007). The SAM signature in the surface warming anomaly over the Antarctic Peninsula is also captured by some models (e.g. Delworth et al., 2006). Other features, including the zonal structure and the temporal signal, exhibit large variance between outputs even in a single model (Miller et al., 2006; Raphael and Holland, 2006). Hence, the extent of discrepancies in the simulated SAM due to model shortcomings alone is difficult to gauge. As an added complexity, new evidence has emerged that ENSO variability can influence SAM variability in the southern summer (L'Heureux and Thompson, 2006).

5.2.3.2 Projected changes in modes of circulation variability

In the range of studies reported by the IPCC's AR4, it has been demonstrated that the ENSO response of climate system models in the Twenty First Century is highly model dependent (see Figure 5.4; (Meehl et al., 2007). This outcome has changed little in the most recent analyses (e.g. Yeh and Kirtman, 2007), and represents a major challenge in projecting Antarctic variability. Currently, there is some consensus that there will be little change in the magnitude of ENSO variability in the Twenty First Century, although some of the models that simulated Twentieth Century ENSO variability well do indicate Twenty First Century increases in the amplitude of El Niño events (Meehl et al., 2007). If such a trend is manifest, it would contribute to changes in sea ice cover and circulation over Antarctica of a similar sense to those observed in the last few decades, but the reliability of such a projection is confounded by the apparent decadal variability in the system (e.g. Fogt and Bromwich, 2006).

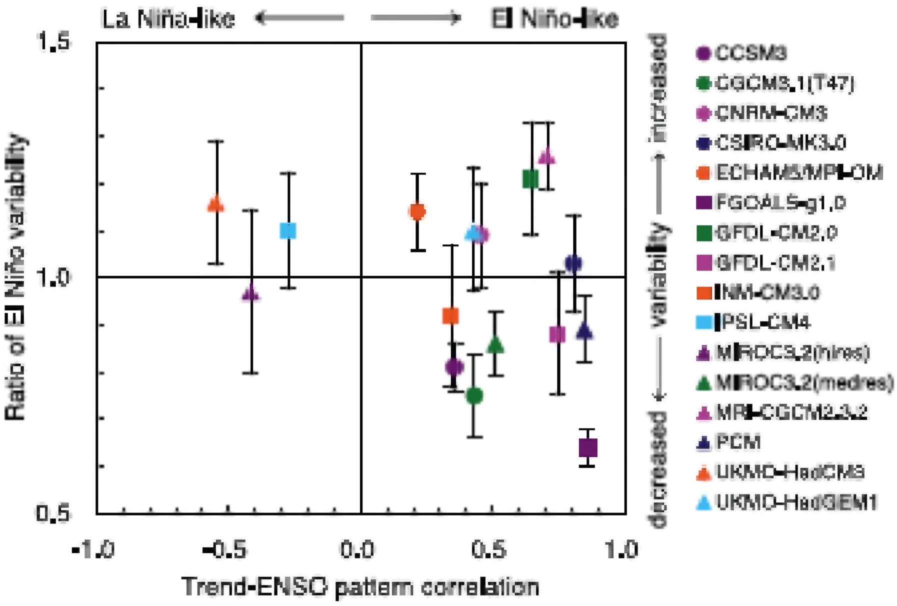


Figure 5.4 Base state change in average tropical Pacific SSTs and change in El Niño variability simulated by AOGCMs. The base state change (horizontal axis) is denoted by the spatial anomaly pattern correlation coefficient between the linear trend of SST in the 1%/yr CO_2 increase climate change experiment and the first Empirical Orthogonal Function (EOF) of SST in the control experiment over the area 10ºS to 10ºN, 120ºE to 80ºW (reproduced from Yamaguchi and Noda, 2006). The change in El Niño variability (vertical axis) is denoted by the ratio of the standard deviation of the first EOF of sea level pressure (SLP) between the current climate and the last 50 years of the IPCC A2 experiments (2051–2100), in the region 30ºS to 30ºN, 30°E to 60°W (reproduced from van Oldenborgh et al., 2005). Error bars indicate the 95% confidence interval; from IPCC (2007).

The future trend in the SAM, as characterized by the leading Empirical Orthogonal Function (EOF) of sea level pressure, has been reported from a number of model projections (e.g. GISSII - Shindell and Schmidt, 2004; CCSM - Arblaster and Meehl, 2006). Most models support a continuing positive trend in the SAM index, as manifest by a strong intensification of the polar vortex. The observed SAM index trend has been related to stratospheric ozone depletion (Sexton, 2001) and to greenhouse gas increases (Hartmann et al., 2000). Specifically, a larger positive trend was projected during the late Twentieth Century by models that included stratospheric ozone changes (e.g. Cai and Cowan, 2007; see Figure 5.5). Further evidence for this relationship is found in the fact that the signal is largest in the lower stratosphere in austral spring through summer (Arblaster and Meehl, 2006). Though somewhat uncertain, it is expected that ozone will continue to slow its decline in the Twenty First Century, as has been observed since 1997 (Yang et al., 2006). Hence, in future projections, the SAM index trends for simulations with and without ozone are comparable. However, the increase in greenhouse gases is also an important factor that supports a continued increase in the SAM index on an annual basis, forced by trends in the meridional temperature gradient (Brandefelt and Källén, 2004). Like the uncertainties surrounding the sources of model error in simulating the SAM in the Twentieth Century, the confounding element of future trajectories in stratospheric ozone concentration makes precise projection of the SAM more problematic than other elements of Antarctic climate; nevertheless the general trend is clear.

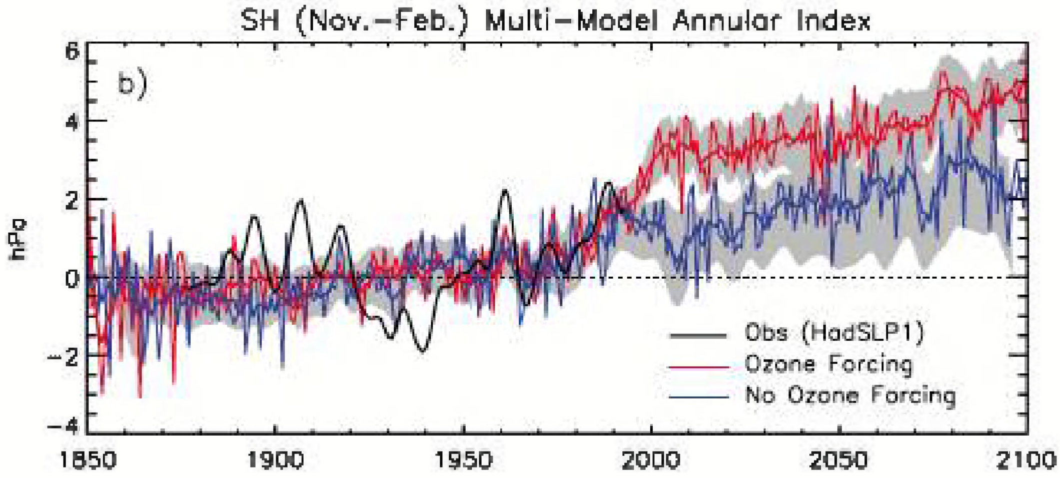


Figure 5.5 Multi-model mean of the regression of the leading EOF of ensemble mean Southern Hemisphere sea level pressure. The time series of regression coefficients has zero mean between year 1900 and 1970. The thick red line is a 10-year low-pass filtered version of the mean with ozone forcing; the blue line is without ozone forcing. The grey shading represents the inter-model spread at the 95% confidence level and is filtered. A filtered version of the observed SLP from the Hadley Centre (HadSLP1) is shown in black. Adapted from (Miller et al., 2006). From IPCC (2007).

5.2.3.3 Impacts on synoptic climate

Observed changes in weather systems are consistent with the trends in the dominant modes of variability, although these results too are strongly dependent on the quality of reanalysis products. In the Southern Hemisphere, cyclonic activity is strikingly different between the ERA-40 and NRA re-analyses (Bromwich et al., 2007). Nevertheless, consistent signals include a decrease in the frequency and increases in the size and intensity of extratropical cyclones in recent decades (e.g. Simmonds and Keay, 2000) with a moderate increase in frequency over the Southern Ocean (e.g. Fyfe, 2003).

A consistent result that has emerged recently from Twenty First Century projections is a tendency for a poleward shift of several degrees latitude in mid-latitude storm tracks (e.g. Fischer-Bruns et al., 2005; Bengtsson et al., 2006). Consistent with these shifts, Lynch et al. (2006) demonstrated increasing cyclonicity and stronger westerlies in high southern latitudes in a 10-member multi-model ensemble simulation of the Twenty First Century. One study (Fyfe, 2003) has suggested a reduction in sub-Antarctic cyclones of more than 30% by the end of the century. Fyfe (2003) did not definitively identify a poleward shift of storm tracks, but the relatively coarse grid of the CCCma climate model (T32, or approximately 600 km grid spacing) may not have been able to detect such a shift. These changes have been related to a simulated circumpolar signal of increased precipitation off the coast of Antarctica (Lynch et al., 2006; see also Section 5.2.5), which perhaps, though loosely, argues against the frequency trend being related to increased data availability, as suggested by Hines et al. (2000).

5.2.3.4 Impacts on accumulation

The Antarctic ice sheet constitutes the largest reservoir of freshwater on Earth, representing tens of metres of sea-level rise if it were to melt completely. Hence, the mass balance of the Antarctic ice sheet is an important contributor to the impacts of sea-level change over the next century. The circulation provides an important component of forcing, particularly through precipitation. The relationships between the major modes of variability and precipitation over the Antarctic continent have been studied, but while there seems to be a correlation between increased coastal precipitation and the SAM index (within the limits of the data), Noone and Simmonds (2002) have demonstrated that, in at least one climate model, eddy moisture convergence (i.e. associated with depressions) represents a large fraction of net precipitation. Synoptic activity intensification over the Southern Ocean would suggest a potential for increase in accumulation along the coasts (Sinclair et al., 1997). However, significant correlations with the SOI appear to be intermittent (e.g. Bromwich et al., 2000; Genthon and Cosme, 2003). More recently, a nonlinear interaction between the Southern Oscillation and the SAM that varies on decadal time scales has been identified as a possible reason for this irregularity (Fogt and Bromwich, 2006).

5.2.4 Temperature change over the Twenty First Century

A significant surface warming over Antarctica is projected over the Twenty First Century. The weighted average of the A1B scenario runs of the CMIP3 models shows an increase of the annual average surface temperature of 0.34°C/decade over land and grounded ice sheets (Bracegirdle et al., 2008) (Figure 5.6). All the CMIP3 models show a warming, but with a large range from 0.14 to 0.5°C/decade under the A1B scenario. The difference between the ensemble average projections of each scenario is smaller then the inter-model spread for any given scenario.

Due to the retreat of the sea ice edge induced by global warming, the largest projected surface warming occurs during the winter when the sea ice extent approaches its maximum, e.g. 0.51 ± 0.26°C/decade off East Antarctica (Bracegirdle et al., 2008).

Inland, away from coastal regions there is very little seasonal dependence of the warming trend, which in all seasons is largest over the high-altitude interior of East Antarctica according to the model average (Figure 5.7). Despite this large increase of temperature, the surface temperature by the year 2100 will remain below freezing over most of Antarctica and therefore will not contribute significantly to melting.

The pattern of warming for the next 100 years is different between simulations and observations of temperature change for the latter part of the Twentieth Century. The most notable difference is that the observed and simulated maximum of warming over the Antarctic Peninsula for the latter part of the Twentieth Century is not present in projections of change over the Twenty First Century. That is because although the Peninsula does continue to warm, other parts of Antarctica warm with it.

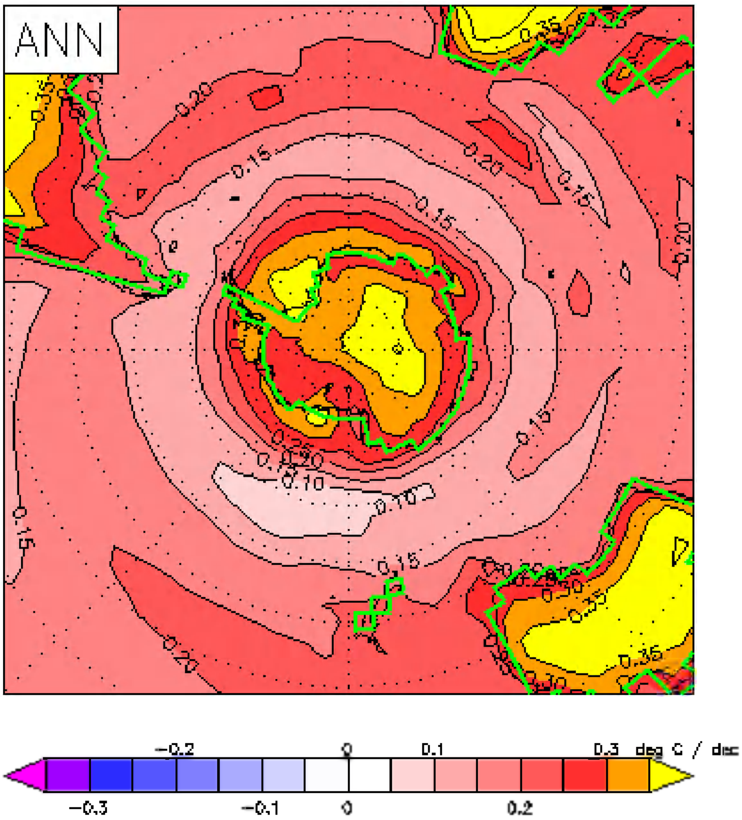


Figure 5.6 Skin temperature trend over Twenty First Century in °C/decade.

The model consensus for warming is strong for Antarctica as a whole, but there is large uncertainty in the regional detail. One way to measure the significance of a projected change is to calculate a signal to noise ratio of that change. Here the signal is the ensemble average change and the noise is the standard deviation of the inter-model spread. A change can be thought of as 'significant' if larger than the inter-model standard deviation, i.e. a signal to noise ratio of greater than one. In Figure 5.8 it can be seen that at most grid points the projected increases of temperature are larger than the inter-model standard deviation. In particular over the Antarctic continent the projected warming shown in Figure 5.7 is much larger than the inter-model spread (Figure 5.8). This demonstrates strong confidence that there will be a warming at the surface in these regions under the A1B scenario. There is less confidence in the large warming trends around the coast than in the smaller changes over the high interior. This is due to the large uncertainty over the sea ice and ocean projections.

According to the IPCC report the warming over the Antarctic continent is 0.5-1.0°C less than over most other landmasses around the globe (apart from south-east Asia and southern South America where increases are the same). The reasons for this are not known. Over the Southern Ocean projected surface warming is much smaller than the global average due to the large heat uptake by the ocean.

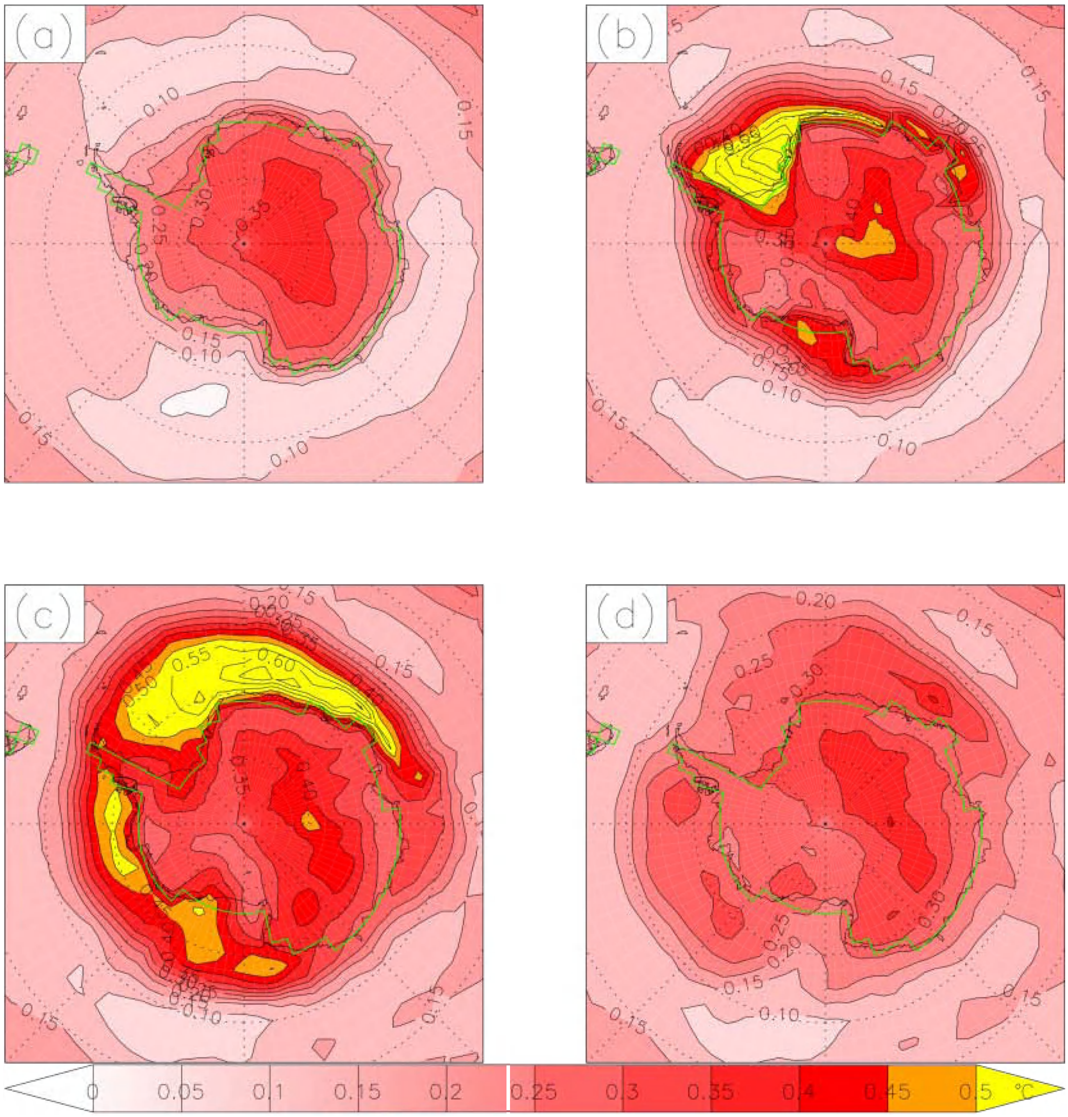


Figure 5.7 Skin temperature change over the Twenty First Century in °C/decade. (a) DJF, (b) MAM, (c) JJA and (d) SON.

In the mid-troposphere above the continent, the annual ensemble mean warming rate at 500 hPa of 0.28°C/decade is slightly smaller than the surface warming, with no evidence of the mid-tropospheric maximum that has been observed over the last 30 years (Turner et al., 2006). The mid-tropospheric warming at low latitudes is larger than over and around Antarctica, which increases the baroclinicity and seems to contribute to the southward migration of the storm tracks that is simulated by the CMIP3 models (Yin, 2005).

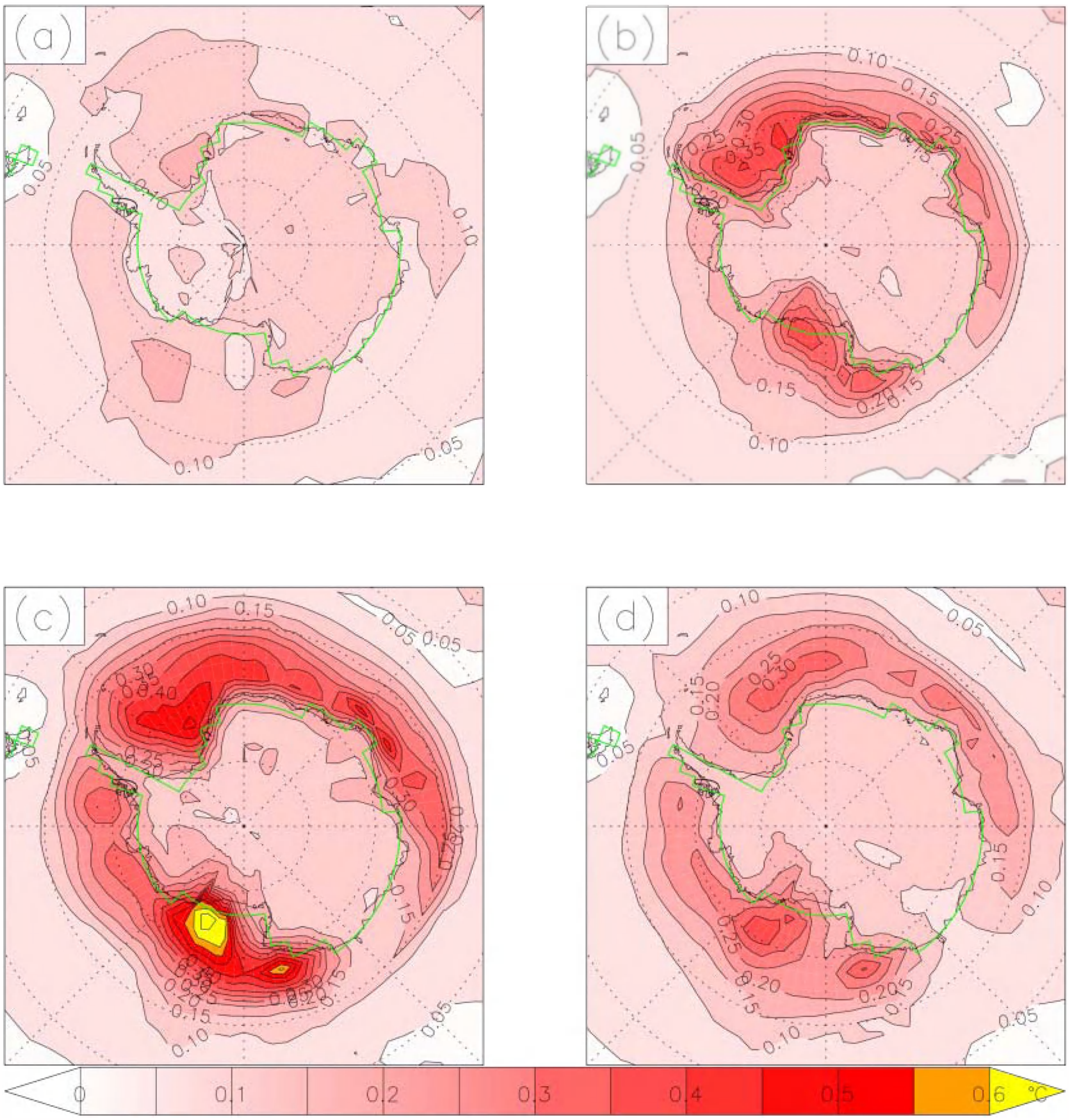


Figure 5.8 Inter-model standard deviation of Twenty First Century skin temperature change for individual grid points in ºC/decade. (a) DJF, (b) MAM, (c) JJA and (d) SON.

5.2.4.1 Extremes

Very little work has been done on changes to extremes over Antarctica. Temperature-related extreme indices are available from the CMIP3 archive and have been assessed by Tebaldi et al. (2006). The heat wave duration index (defined as the maximum period greater than five consecutive days with the daily maximum temperature greater than 5°C above the 1961-1990 mean daily maximum temperature) shows significant increases along coastal Antarctica, where melting can occur, but largest increases over the interior, where heat waves are not warm enough to cause melting. On even shorter timescales, the magnitude, timing and frequency of extreme events (e.g. temperature minima/maxima, freeze-thaw events) can be of considerable relevance to predictions of future trajectories in terrestrial ecosystems.

The extreme temperature range between the coldest and warmest temperature of a given year is projected to decrease around coastal Antarctica and show little change over most of the interior of the continent. Further assessment is required to determine the reasons for the projected pattern of changes. One possibility is that the models simulate a larger nighttime than daytime warming. This has been observed during the second half of the Twentieth Century over the Antarctic Peninsula (Hughes et al., 2007).

5.2.5 Precipitation change over the Twenty First Century

The net precipitation (in climate models this is precipitation less evaporation and ablation) over Antarctica is an important factor in the mass balance of the continental ice sheet. Surface sublimation and blowing-snow processes also help to determine the local mass balance of the ice sheet, but these factors have a limited contribution at scales larger than 100 km (Genthon, 2004) and are not included in climate models. Giovinetto et al (1997) estimated the various terms and found blowing snow to contribute about 6% to the surface mass balance. Hence, quantification of precipitation changes in high southern latitudes is paramount for resolving the uncertainty surrounding the future of the Antarctic ice sheet.

5.2.5.1 Skill in simulation of precipitation

Much of the validation in model performance with regard to precipitation focuses on net precipitation, since this quantity can be calculated using a range of methodologies, and, in particular, is less dependent upon problematic and sparse station measurements of snow accumulation. In an analysis used in the AR4, Uotila et al. (2007) found that only 5 of 15 global climate models examined were able to simulate long term average values of net precipitation consistent with the range of observations in the Twentieth Century. This range falls approximately between 150 and 190 mm/year, although recent analyses suggest the likely figure is at the high end of this range (Monaghan et al. (2006a) quoted 182 mm/year). Monaghan et al. (2006a) combined new records from the International Transantarctic Scientific Expedition (ITASE) with existing ice cores, snow pit and snow stake data, meteorological observations, and validated model fields to reconstruct Antarctic snowfall accumulation over the past 5 decades, and concluded that there had been no trend over that time, although a significant increase in the net precipitation has been reported in the immediate area surrounding the South Pole, based on in situ measurements (Mosley-Thompson et al., 1999) and on satellite altimetry (Davis et al., 2005). It is worth noting that this South Pole area is very small in comparison with the area studied by Monaghan et al. (2006a), and may well have been affected by the existence of the station there (a significant snow hill is also associated with Byrd's former station on the Ross Ice Shelf).

Observational uncertainty, including uncertainties in trend detection, means that an assessment of simulation skill remains particularly difficult. Nevertheless, it is known that specific deficiencies remain in the parameterizations of key processes that drive precipitation. This is particularly true of the polar regions, since polar cloud microphysics remains poorly understood. The importance of the moisture physics is further demonstrated by studies such as that by Turner et al. (2006), who found that local thermodynamic processes were a significant component of the climate change signal in the Antarctic winter. In the context of the SHEBA[1] experiment in the Arctic, much work has been done on the development of polar cloud physics parameterizations, but significant challenges remain even in fine scale models (Sandvik et al., 2007). In the Antarctic, there has been less focused development of the physical parameterizations needed to better represent clouds and precipitation, largely due to the absence of an intense field programme to provide the necessary data support for such an endeavour. The precipitation simulations in high southern latitudes by global models result

[1] SHEBA: Surface Heat Budget of the Arctic – an experiment carried out in the Beaufort and Chukchi Sea region of the Arctic in 1997-1998.

in significant biases (e.g. Covey et al., 2003), and this is also true, though not as severe, in high resolution, limited area models (e.g. Bromwich et al., 2004a; Van de Berg et al., 2005). Hines et al. (2004) found that in one global model, the simulation of Antarctic climate was highly sensitive to the mixing ratio threshold for autoconversion from suspended ice cloud to falling precipitation. Such sensitivity can only be resolved by measurements that are currently not available.

An important driving mechanism for precipitation in this region is the atmospheric circulation (e.g. Massom et al., 2004). The contributions of the multi-year and the synoptic time scales are roughly proportional over the coastal regions, but the synoptic time scale dominates the inland precipitation (Cullather et al., 1998). This component exhibits a close relationship with elevation and makes a positive contribution to transporting moisture from the ocean toward the pole. Global models exhibit significant biases in this regard also. Placed in the context of the available re-analyses, the multi-model ensemble created for the AR4 overestimates pressures over the ice sheet in summer and overestimates cyclone depths, particularly in the west Antarctic region, in all seasons (Lynch et al., 2006). Two issues of particular note in driving these deficiencies are surface forcing and spatial scale. With regard to the former, Stratton and Pope (2004) have noted that Arctic Model Intercomparison Project-style experiments (that is, model simulations with specified sea surface temperatures) do produce correctly located storm tracks, but often even these are more zonally oriented than is observed. Krinner et al. (2007) have found that errors in net precipitation on regional scales are moderated when observed sea surface conditions are prescribed. With regard to the latter, Bromwich et al. (2004b) noted that the reanalysis products, and therefore probably global models in general, underestimate Antarctic precipitation in the Twentieth Century. This is attributed to the smooth coastal escarpment in a coarse resolution model, which causes cyclones to precipitate less than they do in reality. Further, if the Antarctic Peninsula is not well resolved in a model, it produces too little lee cyclogenesis (Turner et al., 1998).

5.2.5.2 Projected changes in precipitation

The lack of a Twentieth Century trend in net precipitation in recent comprehensive analyses is particularly problematic in the context of Twenty First Century model projections. Almost all climate models simulate a continuing robust precipitation increase over Antarctica in the coming century (see Figure 5.9) The projected precipitation change has a seasonal dependency, and is larger in winter than in summer. In some models, there is also a phase shift, so that, for example, a narrow early winter peak in precipitation evolves to a broad winter peak (Krinner et al., 2007). Wild et al. (2003) reported that, in a simulation in which the CO_2 concentration in the atmosphere doubles, the annual net accumulation over Antarctica increases by 22 mm/year. Similarly, Huybrechts et al. (2004) analyzed results from an ice sheet model driven by a climate model simulation in which the CO_2 concentration doubles in 60 years, and found an associated increase of 15% to 20% in mean Antarctic precipitation. The increasing net precipitation in most climate models reflects warmer air temperatures and associated higher atmospheric moisture. Evaporation increases also, but does not keep pace with the precipitation increases. Two models, ECHO-G and GFDL-2.1, project a small decrease of net precipitation after the middle of the Twenty First Century. Interestingly, these models fall at either end of the quality scale in assessments of their ability to reproduce Antarctic synoptic climate, according to Uotila et al. (2007).

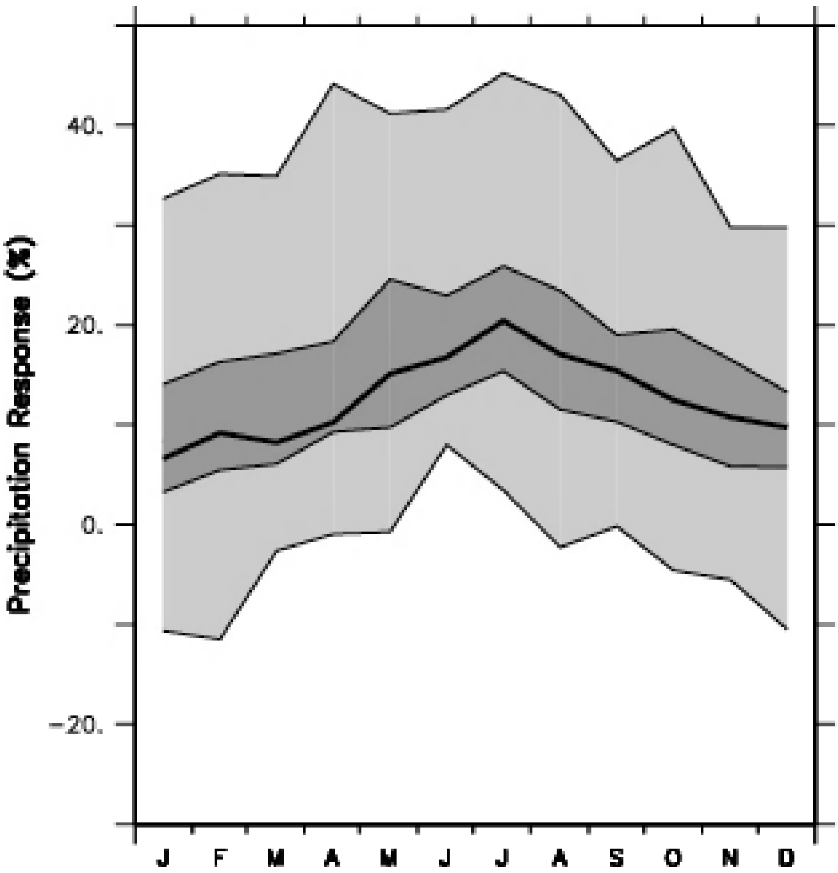


Figure 5.9 Annual cycle of percentage precipitation changes (averaged over the Antarctic continent) for 2080-2099 minus 1980-1999, under the A1B scenario. Thick lines represent the ensemble median of the 21 Multi-model Data Set models. The dark grey area represents the 25% and 75% quartile values among the 21 models, while the light grey area shows the total range of the models.

Only one study has attempted to address the projected changes in the intensity of precipitation. Krinner et al. (2007) calculated the difference in the number of days per year with daily precipitation exceeding five times the mean daily precipitation for the simulations using the LMDZ4 stretched grid atmospheric model (see Figure 5.10) This particular model experiment suggests that even using this relatively modest measure, which does not allow for the increase in total precipitation, the number of relatively strong precipitation events near the ice sheet domes and ridges increases, particularly in East Antarctica. This indicates an increased frequency of intrusions of moist marine air, in spite of a projected lower future cyclone frequency.

A separate, and biologically significant, requirement for inclusion in prediction of future precipitation patterns is the ability to differentiate between precipitation as snow and as rain. While this is not relevant across most of East Antarctica, it is a change that is already apparent along the western Antarctic Peninsula and Scotia arc archipelagos. Precipitation falling as rain is immediately available to terrestrial biota, and hence has some more direct impacts on terrestrial ecosystem changes than that falling as snow.

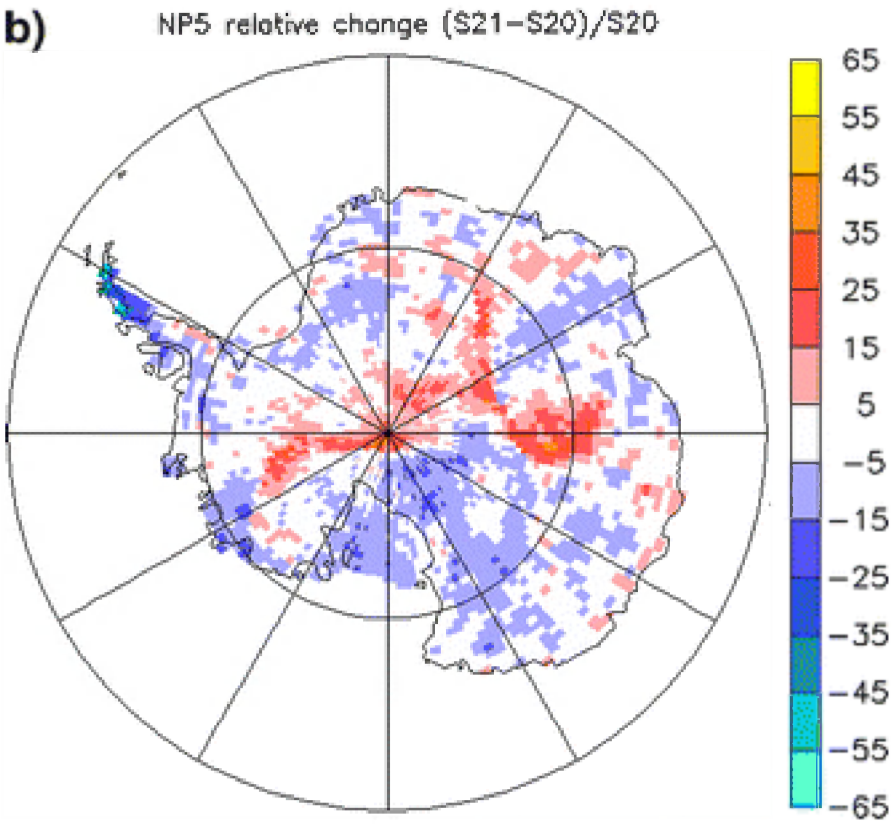


Figure 5.10 Number of days with precipitation exceeding five times the annual daily mean - relative change from the end of the twentieth to the end of the Twenty First Century (in percent). From Krinner et al. (2007).

The scatter among the individual models reported on in the AR4 is considerable. During the first half of the Twenty First Century, models using the A1B scenario predict anything from a maximum upward trend of 0.71 mm/year to a maximum downward trend of 0.13 mm/year. Some models project stronger net precipitation increases in the first half of the Twenty First Century, while other models project stronger increases after the middle of the century. Hence, the confidence in any conclusions arising from AR4 or this report regarding the future trajectory of Antarctic precipitation is extremely limited.

The study of Bracegirdle et al. (2008) estimated that by the end of the century the snowfall rate over the continent would increase by 20% compared to current values, which, if other effects such as melting and dynamical discharge are ignored, would result in a negative contribution to global sea-level rise of approximately 5 cm.

5.2.6 Atmospheric chemistry

5.2.6.1 Antarctic stratospheric ozone over the next 100 years

The success of the Montreal Protocol in constraining production of ozone depleting substances (CFCs, Halons, and organic chlorides and bromides) has meant that their amounts in the stratosphere are now decreasing at about 1%/year. Models predict the future evolution of ozone based on specific scenarios for future emissions of ozone-depleting substances. Both two and three-dimensional models have been used to predict future ozone loss, but because of the more complex stratospheric dynamics near the poles, polar ozone is best simulated by 3-D models. For comparison with past measurements, Chemistry Transport Models that use prescribed wind fields from past measurements have been very successful, but they clearly cannot be used for predictions of the future. Instead, coupled Chemistry-Climate Models are used. Those whose results are described below vary in their skill in representing the atmosphere, but there is sufficient general agreement between them and observations that we can have some confidence in their predictions.

Model	**Institute**	**Resolution**	**No. of levels**	**Top (hPa)**	**Reference**
AMTRAC	GFDL (USA)	2.0° x2.5°	48	0.0017	Austin et al. (2006)
CCSRNIES	NIES (Japan)	2.8° x 2.8°	34	0.01	Akiyoshi et al. (2004)
CMAM	MSC, UT, York (Canada)	3.75°x3.75°	71	0.0006	Beagley et al. (1997)
E39C*	DLR (Germany)	3.75°x3.75°	39	10	Dameris et al. (2005)
GEOSCCM	GSFC (USA)	2.0° x 2.5°	55	0.01	Bloom et al. (2005)
MAECHAM-4CHEM*	MPI Mainz (Germany)	3.75°x3.75°	39	0.01	Manzini et al. (2003)
MRI	MRI (Japan)	2.8° x 2.8°	68	0.01	Shibata et al. (2005)
SOCOL	PMOB, ETHZ (Switz)	3.75°x3.75°	39	0.01	Rozanov et al. (2005)
ULAQ	U L'Aquila (Italy)	10° x 22.5°	26	0.04	Pitari et al. (2002)
UMETRAC	UKMO, NIWA (NZ)	2.5° x3.75°	64	0.01	Struthers et al. (2004)
UMSLIMCAT	UKMO, Leeds (UK)	2.5° x3.75°	64	0.01	Tian and Chipperfield (2005)
WACCM	NCAR (USA)	4.0° x5°	66	4.5×10^{-6}	Park M. et al. (2004)

Table 5.1 Models whose results appear in Figures 5.11, 5.12 and 5.13; * bromine chemistry not included

In the figures below, four diagnostic quantities are discussed:

A. The minimum ozone in the Southern Hemisphere during September to October, which is a common diagnostic of the maximum depth of the ozone hole, and so of the maximum ozone loss.
B. The ozone mass deficit (OMD), which is the mass of ozone that would be required to elevate ozone columns above 220 Dobson Units (DU), as observed by satellite instruments (Huck et al., 2007). OMD can be calculated daily or averaged over some period, and it is the most accurate diagnostic of total Antarctic ozone loss.
C. The ozone hole area, which is the area with ozone less than 220 DU as measured by satellite instruments. Because they can only observe in sunlight, in early September there can be an unobserved area of more than 220 DU close to the pole, whose area is not subtracted from the area within the outer contour. Nevertheless this is a common diagnostic of ozone hole size.
D. The total inorganic chlorine (Cly), which equals the sum of chlorine in organic chlorine compounds entering the stratosphere, after degradation by UV light and reaction with oxides of hydrogen and nitrogen. Components of Cly are the comparatively stable compounds HCl and $ClNO_3$, as well as the reactive Cl, ClO, OClO, Cl_2O_2 and HOCl.

Stratospheric ozone is affected by a number of natural and anthropogenic factors in addition to reactive halogens: temperature, transport, volcanoes, solar activity, hydrogen oxides, and nitrogen oxides. In any discussion of future ozone, it is important to separate the effects of these factors, particularly if considering the future success or otherwise of the

Montreal Protocol. For example, when Cly has decreased to pre-ozone hole values, continued cooling of the stratosphere due to increased greenhouse gases warming the troposphere may lead to amounts of stratospheric ozone quite different to those of pre-ozone hole days. The net result is not that simple, however. Increases in greenhouse gases affect polar ozone via processes acting in opposing directions, making model predictions less certain near the poles than elsewhere:

1. Increased greenhouse gases act to cool the stratosphere, which will slow gas-phase ozone loss reactions and so tend to increase stratospheric ozone.
2. The same cooling acts to increase the amounts of PSCs, on which the reactions leading to ozone loss occur, so we can expect reduced ozone given the same Cly - opposite to the effect of gas-phase chemistry in 1 (above). This is particularly likely in the edge region of the vortex (Lee A et al., 2001), because PSCs are not ubiquitous there. Ozone loss in this edge region defines the ozone hole area diagnosed in Figure 5.12.
3. Increased GHGs act to increase the strength of the Brewer-Dobson circulation by changing the wave driving that causes it. In the short term, this increases the supply to the stratosphere of (a) CH_4 and so hydrogen oxides, (b) N_2O and so nitrogen oxides, and (c) CFCs and halons, and so reactive chlorine and bromine. Each of these helps to remove ozone, so this process also acts to reduce ozone in the short term. In the long term, the increased removal of CFCs and halons via the actions of the Protocol would more rapidly reduce ozone-depleting substances in the whole atmosphere, so that the effects from reduced CFCs and halons would oppose those arriving via CH_4 and N_2O.

Despite these complications, the general characteristics of future Antarctic ozone as shown in Figure 5.11 are similar in all models, and similar to projections in earlier models in World Meteorological Organisation (2003): minimum ozone occurs around 2000, followed by a slow increase. The increase is slow because the near-total destruction of ozone in the core of the ozone hole means that there is low sensitivity to Cly there, so only small changes in ozone hole depth are expected as Cly starts to decline. Larger sensitivity to changes in Cly is expected at the upper altitudes of the ozone hole (20-22 km) where ozone depletion is not complete, and this is a possible region in which to detect the onset of ozone hole recovery (Hofmann et al., 1997).

The minimum amount of ozone in Figure 5.11 differs widely between models, ranging from 60 DU to over 120 DU compared to the observed 80 DU, highlighting the difficulty of predictions of polar ozone by fully-coupled models. The fact that Chemistry Transport Models agree much better with observations and with each other suggests that it is transport in these fully coupled models that accounts for the differences and difficulties.

Similarly, the predicted values of maximum ozone mass deficit in Figure 5.12 vary widely between models (from 7 to over 33 million tons, compared to the observed 31 million tons), as do predictions of maximum ozone hole area. Note that because both ozone mass deficit and ozone hole area are below a 220 DU threshold, a bias in global ozone in any one model will create a bias of opposite direction in both diagnostics (e.g. the low bias in area and mass deficit in MAECHAM4CHEM is probably caused by a general high bias in global ozone).

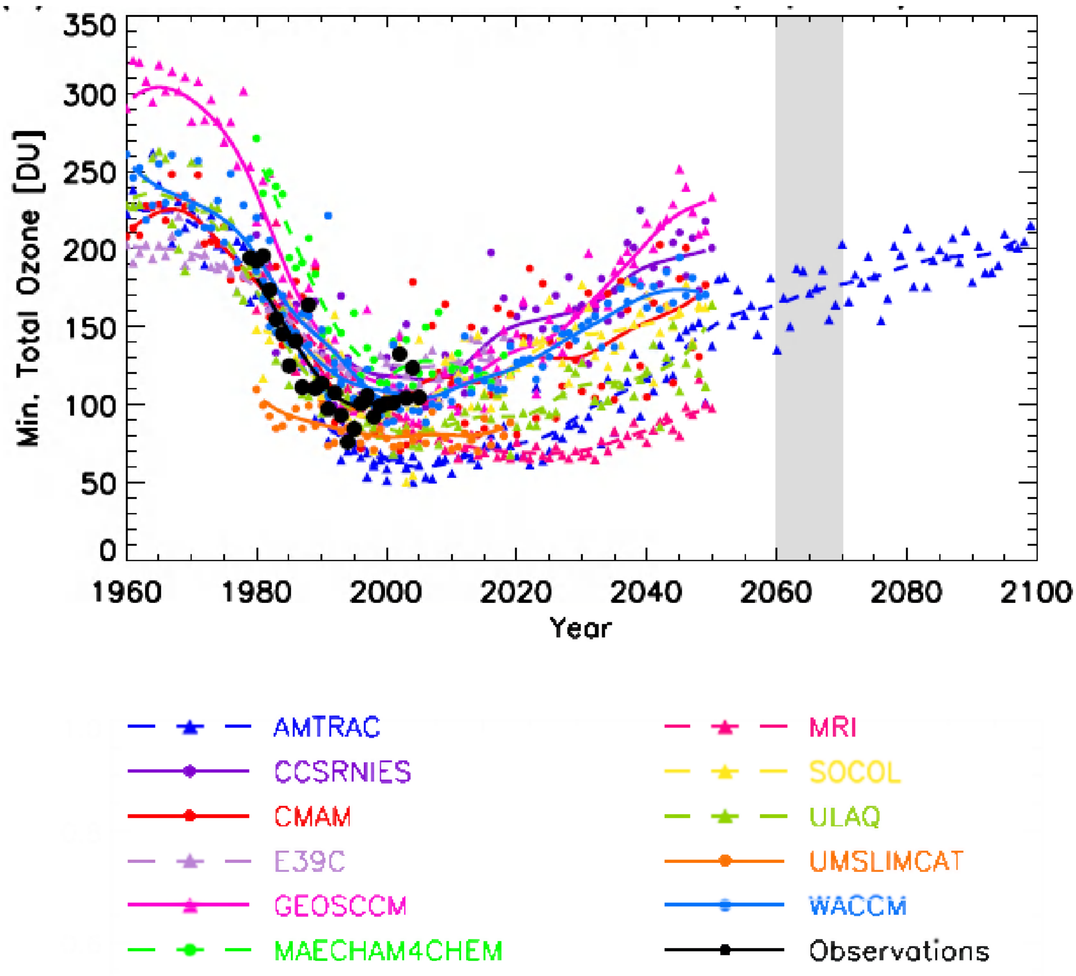


Figure 5.11 Minimum total column ozone in September to October predicted by various models, plus observations from the National Institute of Water and Atmosphere Research (NIWA) total ozone database in New Zealand (Bodeker et al., 2005). Solid and dashed curves show smoothed values. Light gray shading shows when CFCs and halons are expected to return to 1980 values. Adapted from World Meteorological Organisation (2006).

Some insight into the model differences can be obtained from comparisons of Cly in the models. As shown in Figure 5.13, there is a large spread in the simulated Cly, including in the maximum value and in the date at which it decreases to 1980 values. In several models, the maximum Cly is unrealistically low with the result that its return to 1980 values is too early, which is likely to ensure that model's return to 1980 values of ozone is too early. More weight should therefore be put on results from models with more realistic maximum Cly. AMTRAC matches the observations of Cly best, and predicts the latest return to 1980 values.

AMTRAC also predicts the latest recovery of ozone in Figures 5.11 and 5.12. This is almost consistent with the study of Newman et al. (2006), who used a parametric model of spring ozone amounts that includes Cly amounts and stratospheric temperatures. Figure 5.12 shows that they predicted that a return to 1980 ozone amounts would not occur until about 2070.

Despite the differences in models, extrapolating their results suggests that by 2100 Antarctic ozone will no longer be under the influence of CFCs and halons. However, it may not have reverted to 1980 values because of changes in stratospheric temperatures and dynamics caused by increased greenhouse gases.

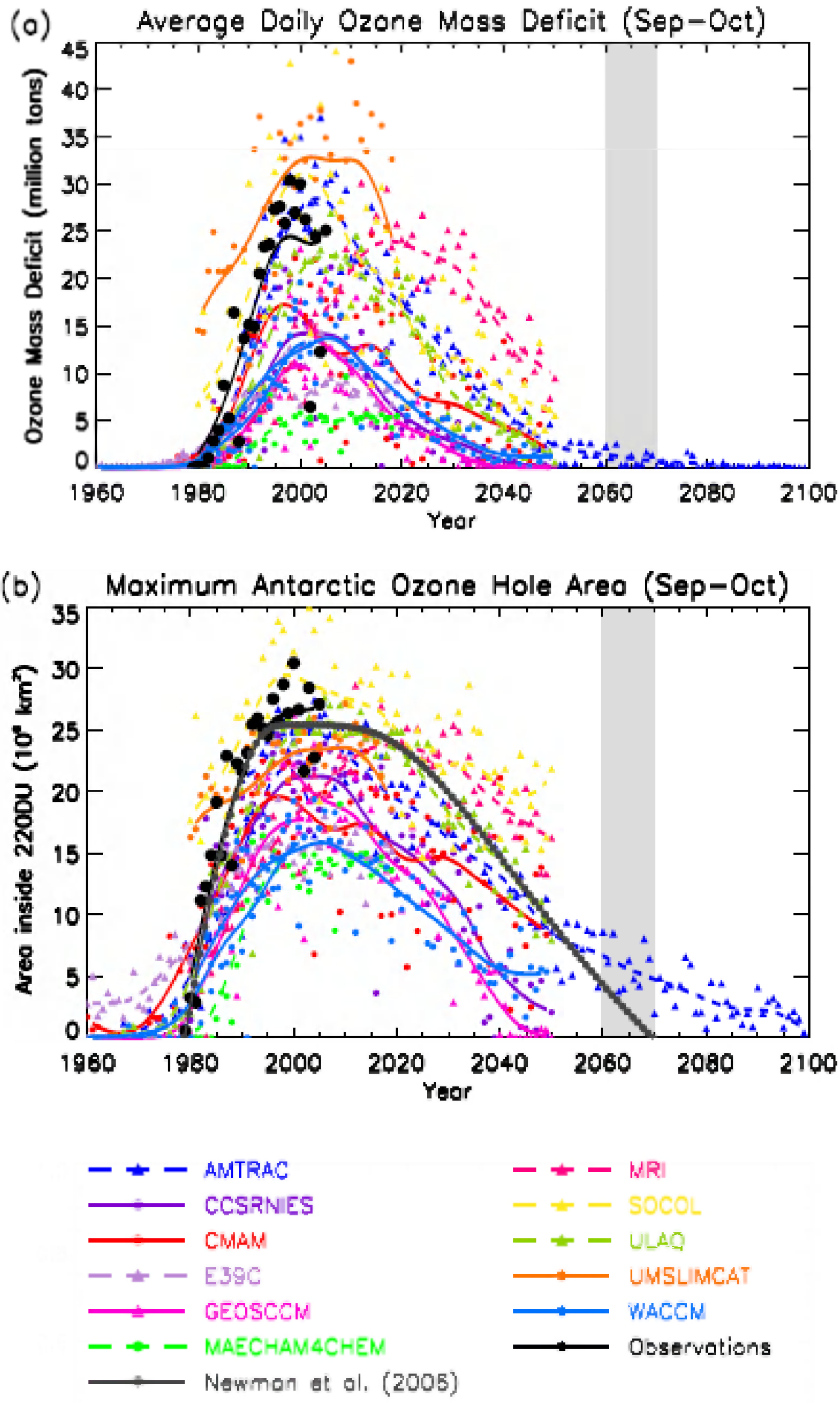


Figure 5.12 (a) September to October average daily ozone mass deficit, and (b) the maximum ozone hole area, for each year from each model. Curves, shading and source of observations are as in Figure 5.11. Adapted from World Meteorological Organisation (2006).

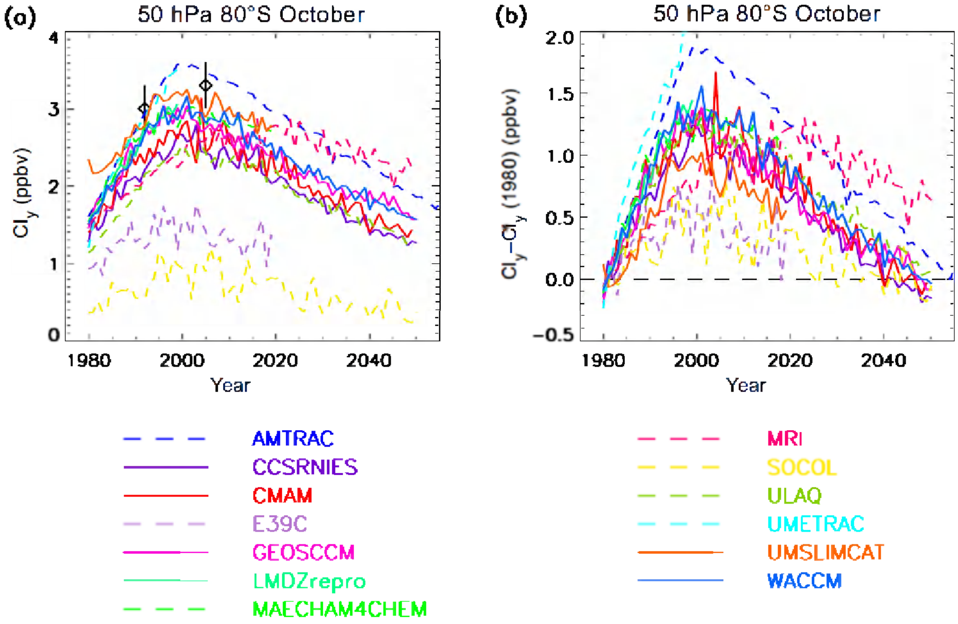


Figure 5.13 Zonal mean values of total inorganic chlorine (ppbv) in October at 50 hPa (20-25 km) and 80°S predicted by models: (a) total, (b) difference from 1980. Open black diamonds in (a) show estimates from measurements by UARS satellite in 1992 and by Aura satellite in 2005.

5.2.6.2 Antarctic Tropospheric Chemistry

In this section we discuss how the chemistry of the Antarctic troposphere is influenced by the cryosphere, and how that chemistry may change in a warmer world with a reduced cryosphere.

Trace gases in the atmosphere are split apart (photolysed) by solar radiation to generate highly reactive radicals. Over regions of snow and ice, incoming solar radiation is reflected back by the surface; the extent of this reflection is the surface albedo, which varies between 0.81 and 0.83 over snow in coastal Antarctica (Gardiner and Shanklin, 1989). The presence of snow on both land and sea ice surfaces thus significantly increases the pathway of solar radiation by reflection, so amplifying opportunities for photolysis of trace gases in the troposphere. Without this high albedo, the lifetime of trace gases would be increased and Antarctic tropospheric chemistry would be less reactive.

The cryosphere also limits tropospheric chemistry by acting as a cap to emissions from the underlying land or ocean. If warming were to reduce the extent of the snowpack, exposing land surfaces, it is likely that microbes would become active in the continental soil. Microbes are recognised sources of nitrous oxide (N_2O), which is a long-lived and powerful greenhouse gas. Emissions of N_2O from such freshly exposed soil would be likely to contribute to some extent to global warming.

Sea ice has a direct influence on boundary layer chemistry. Newly forming sea ice with its associated brine pools and frost flowers, as well as sea salt on snow, are all potential sources of inorganic bromine compounds. These trace gases are emitted into the boundary layer where they are potent destroyers of ozone. Under specific meteorological conditions, ozone depletion events (ODEs) are observed, where ozone concentrations can drop from a normal background amount to below instrument detection limits within a matter of minutes.

These extreme events are observed within the boundary layer using ground-based instruments. Vertical profile measurements of ozone have also been made using balloon-borne sensors to assess the influence of such halogen-driven ozone loss at higher altitudes. Studies at two different coastal sites in Antarctica have reported significant reductions in ozone up to around 3 km above the snow surface (Wessel et al., 1998; Kreher et al, 1997). Ozone is a radiatively important gas whose influence varies with altitude, being more important in the free troposphere where it is colder (Lacis et al., 1990). Roscoe et al. (2001) suggested that Antarctic boundary layer ozone loss could be sustained and mixed to higher altitudes where the reduction in ozone would exert a radiative cooling that would be significant on a regional scale. They further argued that in a warmer world with reduced sea ice extent, the natural process of BrO production and therefore ozone depletion would be reduced. Ozone in the free troposphere would consequently be sustained at higher concentrations, thereby exerting an additional warming influence. They estimated the additional warming to be of the order 0.05ºK, i.e. a small but positive feedback.

Plainly, sea ice also stops emissions to the atmosphere of trace gases with an oceanic origin. A loss of sea ice would therefore enhance emissions of such gases to the atmosphere. Tropospheric trace gases released from the oceans around Antarctica include: dimethyl sulphide (DMS); alkenes such as ethene (C_2H_4) and propene (C_3H_6); and bromocarbons such as bromoform ($CHBr_3$) and dibromomethane (CH_2Br_2), all generated by phytoplankton. DMS plays a critical role as a source of cloud condensation nuclei (CCN) via its oxidation to sulphate (SO_2) (see e.g. Finlayson-Pitts and Pitts, 1999). Changing the number of CCN alters cloud properties and albedo, so influences the Earth's radiation budget, surface temperature and climate. Feedback loops may exist with increased DMS emissions resulting in enhanced CCN, with consequent changes in climate that would then impact on DMS production and emission (Charlson et al., 1987). It has been suggested that chemical analyses of deep ice cores confirm Charlson's hypothesis, but this is not confirmed by recent (and apparently still unique) chemistry/climate modeling (Castebrunet et al, 2006).

Oxidation of DMS is predominantly driven by reaction with hydroxyl ions (OH) and proceeds via two channels:

$$OH + CH_3SCH_3 \rightarrow CH_3S(OH)CH_3 \text{ (addition)}$$
$$OH + CH_3SCH_3 \rightarrow CH_3SCH_2 + H_2O \text{ (abstraction)}$$

The proportion of DMS being oxidised by the respective channels depends partly upon ambient temperature; at temperatures below 12ºC the addition channel leading to dimethylsulphoxide (DMSO) is believed to dominate (Arsene et al., 1999). The abstraction channel, which dominates at warmer temperatures, ultimately results in production of SO_2, and hence CCN. In a warmer world, the proportion of DMS oxidising with OH via the abstraction channel could therefore increase, forming more CCN. However, that possibility may be countered by the influence of BrO. BrO oxidises DMS to DMSO rather than to SO_2, and Von Glasow (2002) showed that by including BrO reactions in an atmospheric chemistry transport model, global concentrations of DMSO increased by 63%. Around coastal Antarctica, even present day concentrations of BrO appear to significantly influence DMS (Read et al., 2008). Future concentrations of BrO might be affected in two ways. In a future warmer world, with the potential for additional oceanic emissions of bromocarbons, background BrO concentrations in coastal Antarctica could quite likely be higher than they are today. But the sea ice source leading to BrO might become less important, thus reducing BrO concentrations. Thus, even with enhanced levels of DMS in the future it is not clear that greater numbers of CCN would result around Antarctica. Whether or not the climate was influenced by these reactions would depend on the balance between increases in the oceanic sources, changes in BrO, and ambient temperature.

Iodine compounds are also emitted into the boundary layer. Significant concentrations of iodine monoxide (IO) have been observed in the boundary layer at Halley station on the Weddell Sea coast (Saiz-Lopez et al., 2007). The seasonal maximum occurred in spring, but even during the summer, concentrations were high enough to influence tropospheric chemistry processes. The IO is thought to originate from diatoms under the sea ice. It is a major source of new particles (marine aerosols and CCN) from which clouds originate. Aerosols and clouds scatter incoming solar radiation and so cool the atmosphere. The link between IO and sea ice suggests that in a warmer world with less sea ice, less IO will be produced, hence there will be fewer aerosol and CCN particles, and less radiative scattering, encouraging additional warming at the Earth's surface. But, if IO emanates directly from the ocean, and is not dependent upon the presence of sea ice, then reduced sea ice could increase IO emissions, hence CCN and aerosol production, encouraging cooling. To clarify whether the feedback would be positive or negative it is necessary to determine the source of boundary layer IO.

The snowpack itself also influences the atmospheric boundary layer by acting as a source of highly reactive trace gases. Over parts of Antarctica, such as the polar plateau, these significantly increase the oxidising capacity of the boundary layer above what is expected. In coastal regions, the effect is less pronounced. The differences reflect both the fetch of snow and the stability of the boundary layer. If all Antarctic snow disappeared, this source of trace gases would be removed, and the atmosphere would most likely move to a more sluggish state with longer-lived and less reactive chemical species. But, given that the East Antarctic Ice Sheet is unlikely to disappear for thousands of years or longer, emissions from snow will be an important driver of local tropospheric chemistry for years to come.

Clearly the cryosphere over and around Antarctica has a significant influence on tropospheric composition and chemistry. Although there are significant uncertainties, it is clear that a substantial change in the Antarctic cryosphere would alter Antarctic tropospheric chemistry from its present day state. Such an alteration might itself further influence the climate system. These factors need to be encapsulated in climate models.

5.3 Ocean circulation and water masses

5.3.1 Simulation of present-day conditions in the Southern Ocean.

Coupled General Circulation Models (CGCMs) used in the framework of the IPCC's AR4 have made significant progress in their representation of high latitude processes compared to earlier model versions (Randall et al., 2007). However, the Southern Ocean remains one of the regions where the largest differences are found between models and observations and between different models. In particular, the transport of the Antarctic Circumpolar Current (ACC) through Drake Passage simulated by those models ranges from –6 Sverdrup (Sv, where 1 Sverdrup $= 1 \times 10^6 m^3/sec$) (i.e. a westward transport) to more than 300 Sv (eastward transport). Only two models among the nineteen analysed by Russell et al. (2006a) were able to obtain transports that were within 20% of the value estimated from observation (135 Sv), while most of the simulated values were within 50 Sv of this estimate. Those strong biases have been attributed to different factors (Russell et al., 2006b). Some models tend to simulate too low a zonal wind stress in the Southern Ocean or to have a maximum in zonal wind stress that is located too far north compared to observations. As a consequence, the simulated wind stress in the latitude band of the Drake Passage is too low, resulting in a too weak ACC transport in those models (Russell et al., 2006a). The density contrast across Drake Passage appears to be another important driving factor of the ACC transport in models. The errors in the simulation of this density gradient, partly due to problems in estimating the export of

North Atlantic Deep Water (NADW) towards the Southern Ocean, could thus play a significant role in explaining the difference in transport between model and observations (Russell et al., 2006a).

The representation of Southern Hemisphere subpolar gyres in IPCC AR4 climate Models has been investigated by Wang and Meredith (2008). As shown in Chapter 1, the models reproduce three southern subpolar gyres: the Weddell Gyre, Ross Gyre, and Australian-Antarctic Gyre, in agreement with observations. Some models simulate the presence of a subpolar ''supergyre'', with strong connectivity between these three gyres. The gyre strengths and structures show a great range across the various models. The link between the gyre strengths and wind stress curls is weak, indicating that the Sverdrup balance (the theoretical relationship between the wind stress exerted on the surface of the open ocean and the vertically integrated meridional (north-south) transport of ocean water) does not hold for the modelled southern subpolar gyres; instead, the simulated gyre strengths are mainly determined by upper layer meridional density gradients, which are themselves determined predominantly by the salinity gradients. These findings suggest that a correct simulation of salinity is crucial for the simulation of Southern Ocean circulation.

When the temperature and salinity averaged over all the models is compared to observations (Figure 5.14), the zonal mean differences are relatively small. There is a tendency to have too warm and too salty water masses around 30-40°S in the depth range 500-1,000 m (Randall et al., 2007). This could be related to a too far northward site of formation of Antarctic intermediate waters (AAIW) in those models. Besides, on average, in the same depth range but in the latitude band 60-70°S, the models tend to simulate too cold and too fresh water masses. This bias could be caused by a too weak inflow of warm and salty waters from the north or by excessive exchanges with the surface. Overall, the representation of the vertical stratification for the ensemble mean seems a reasonable projection, although it conceals significant over and under estimates in several of the models (Russell et al., 2006a).

5.3.2 Projections for the Twenty First Century

There is no agreement for the projections of the ACC transports among 19 IPCC AR4 models (Wang and Meredith, 2008). The ACC transports are significantly increased in eight models, while they are significantly decreased in eight other models. A recent observational study on the evolution of the ACC transport over the past several decades revealed that that there is no increase in the tilt of isopycnals. This indicates that the ACC transport is not sensitive to the significant intensification of the Southern Hemisphere westerlies during the past several decades as suggested by some models, increasing our uncertainties in the future evolution of the ACC transport (Böning et al., 2008).

The Southern Hemisphere subpolar gyres over the Twenty First Century in IPCC AR4 models generally become significantly intensified (Wang and Meredith, 2008). This is a consequence of the wind forcing over the subpolar region becoming more cyclonic, associated with the intensification and southward shift of the circumpolar westerlies. Conversely, changes in freshwater forcing and in the transport of the adjacent ACC exert only minor influences. The strengthening of the subpolar gyres will likely have strong impacts on the mass balance of ice shelves and the stability of the Antarctic ice sheets, and could also impact strongly on the transformations of water masses within the subpolar gyres and the exports of dense waters to lower latitudes.

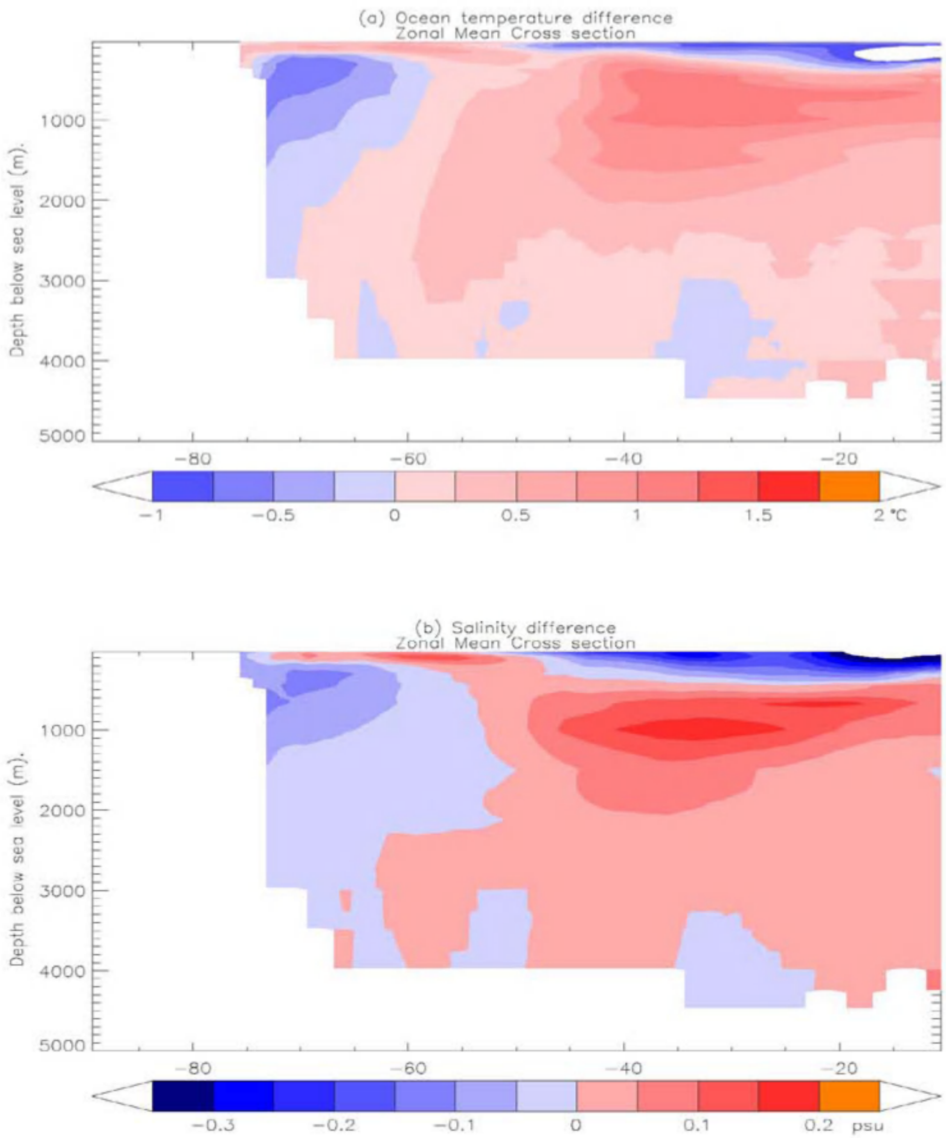


Figure 5.14 The zonal mean difference between observed (a) temperature (°C) and (b) salinity (psu) and the average of 19 CGCMS simulations for the period 1981-2000 (20C3M simulation), superimposed on the bathymetry. The observations are taken from World Ocean Atlas (2001, http://www.nodc.noaa.gov/OC5/WOA01/pr_woa01.html). The 19 models for which sufficient data for the 20C3M runs was available are CCCMA CGCM3.1, CCCMA CGCM3.1-T63, CNRM CM3, CSIRO MK3.0, CSIRO MK3.5, GFDL CM2.0, GFDL CM2.1, GISS_AOM, GISS MODEL E R, INGV ECHAM4, IPSL CM4, MIROC3 2 HIRES, MIROC3 2 MEDRES, MIUB ECHO G, MPI ECHAM5, MRI CGCM2-3_2A, NCAR CCSM3_0, NCAR_PCM1, AND UKMO HADCM3.

The changes of ocean temperature and salinity (means over 2091-2100 minus means over 2001-2010) are presented in summer-winter pairs in Figures 5.15(i) through 5.16(vi). For the purposes of averaging, winter is taken as the time of peak sea ice (August, September, October), and summer as the sea ice minimum (February, March, April). Output from 19 models is used in the calculations (see caption to Figure 5.14). This average over the ensemble of models is likely the most reasonable estimate of the future change presently available but, because of the large scatter between the projections performed with different models, the analysis of those future changes has to be taken with caution.

The SST changes are small compared with those observed in surface air temperature in the climate models, because the heat capacity of the ocean is much larger than that of the atmosphere. Nevertheless both the atmospheric and oceanic temperatures will have an effect on sea ice. Figure 5.15(i)(a) shows that south of 60ºS in summer the SSTs are likely to be warmer than at present in 2100 by between 0.5ºC and 1.0ºC, except in the Amundsen Sea where they are likely to be warmer by 1.0 to 1.25ºC. South of 60ºS in winter (Figure 5.15(i)(b)) the SSTs are likely to be close to what they are now, i.e. between up to 0.5ºC warmer or -0.25ºC cooler than they are at present, except far offshore off Dronning Maud Land, off West Antarctica and off Queen Mary Land, where they may warm to 0.5ºC to 1.0ºC.

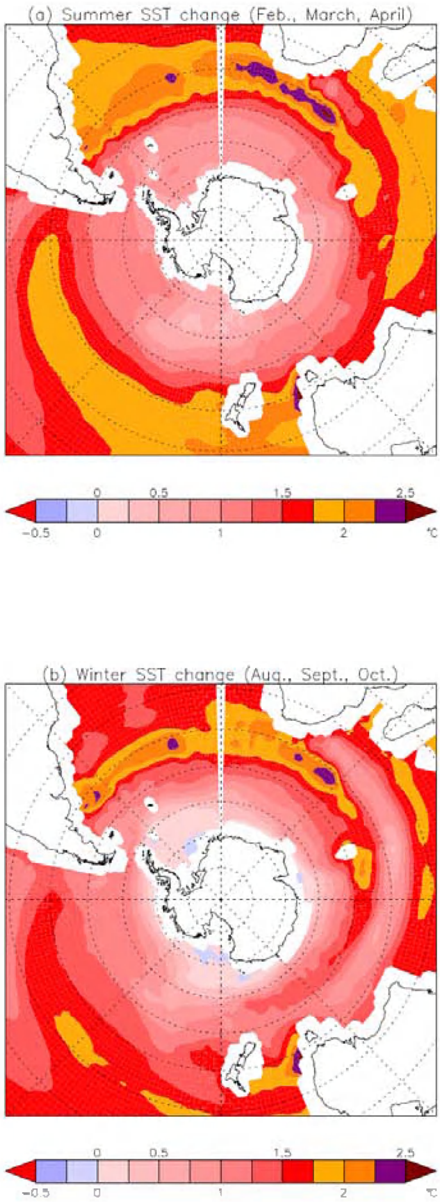


Figure 5.15(i). Sea surface temperature (SST) change for (a) summer and (b) winter between 2000 and 2100.

Figure 5.15(ii)(a) shows that to the south of 60ºS in summer, the surface waters will be fresher by 0.1 to 0.2 units, with local patches up to 0.3 fresher in the Weddell Sea, in the Ross Sea off Oates Land, and in a few patches elsewhere along the coast. Figure 5.15(ii)(b) shows that south of 60ºS in winter the pattern is very similar, but the surface waters are fresher over a larger area and there are salinities up to 0.3 units fresher west of the Peninsula.

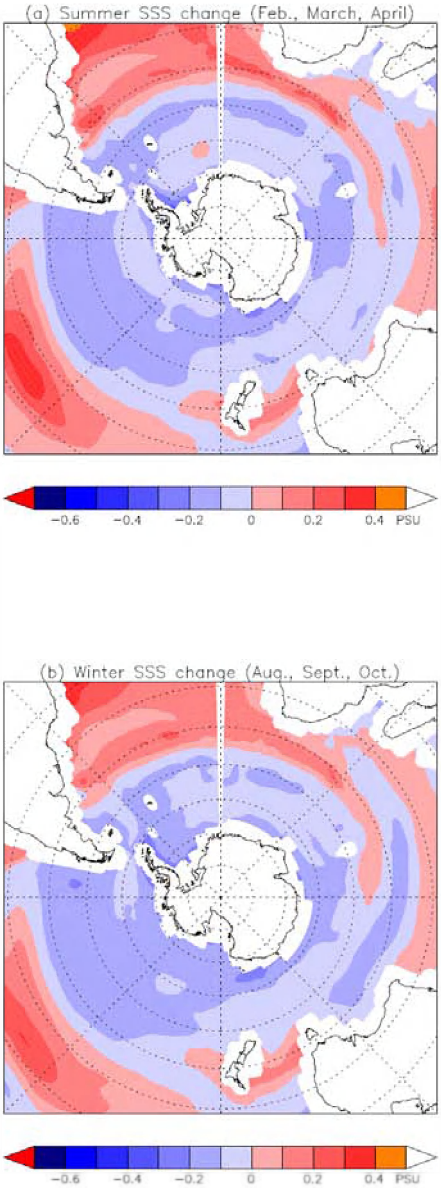


Figure 5.15(ii). Sea surface salinity (SSS) change for (a) summer and (b) winter, between 2000 and 2100.

Figure 5.15(iii) (below) shows that in both summer (a) and winter (b) the bottom water temperatures on the continental shelf at 200 m in 2100 are likely to be warmer by between 0.5ºC and 0.75ºC, except in the Weddell Sea where the warming is less (between 0ºC and 0.5ºC).

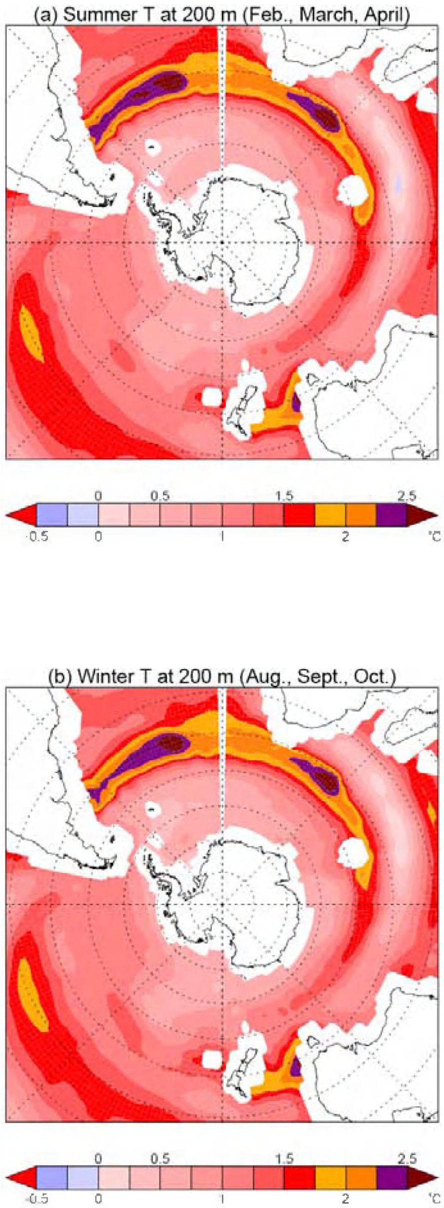
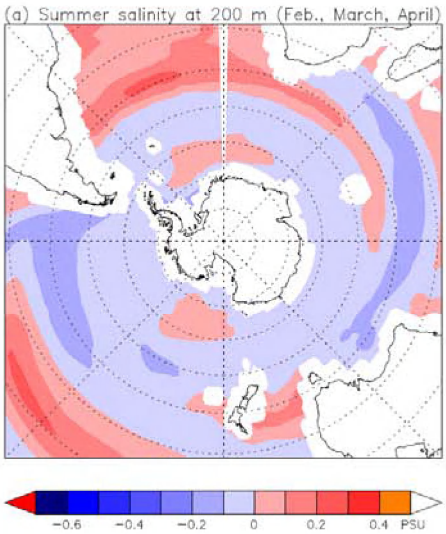


Figure 5.15(iii) Ocean temperature change at 200 m for (a) summer and (b) winter, between 2000 and 2100.

Figure 5.15(iv) (below) shows that in both summer (a) and winter (b) the bottom water salinities on the continental shelf in 2100 are likely to be up to 0.1 units fresher than they are now, and up to 0.2 fresher in the Weddell Sea.



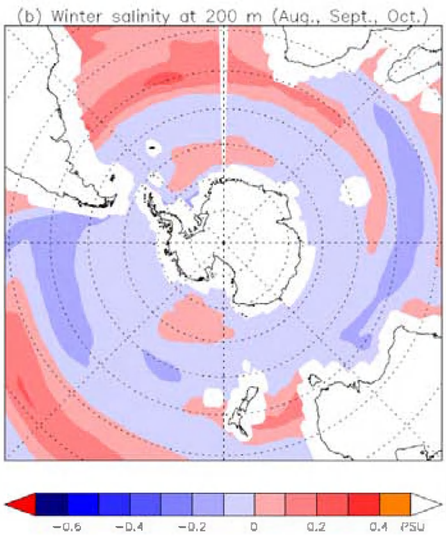
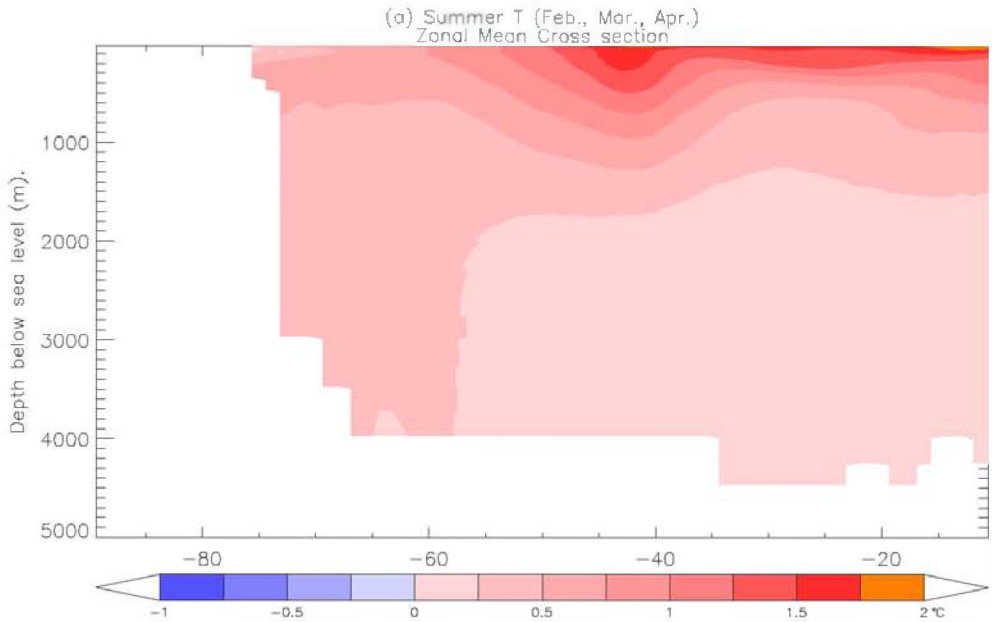


Figure 5.15(iv) Salinity change at 200 m for (a) summer and (b) winter, between 2000 and 2100.

As shown in Figure 5.15(v), regardless of season the bottom waters from the surface down to 4,000 m along the continental margin are expected to warm in winter and summer by around 0.25°C, with the possibility of warming by up to 0.5°C or slightly more at depths of 200 to 500m. The warming surface layers in both winter and summer are quite thin – largely less than 200 m. There is significant warming (0.75 to almost 2°C in all seasons) at the surface between 40 and 60°S, in the core regions of the Antarctic Circumpolar Current. The CGCMs reproduce quite well the mid-depth warming observed during the second half of the Twentieth Century, in particular if the effect of volcanic eruptions is taken into account in models (Fyfe, 2006). During the Twenty First Century, this warming is projected to continue, reaching nearly all depths when averaged over the ensemble of models. Close to the surface, the warming of the Southern Ocean during the Twenty First Century is weaker than in other regions. This is partly related to the large heat storage by the ocean, which removes a large amount of heat from the atmosphere in an area where the ocean covers nearly all the longitudes and where relatively deep mixed layers occur.



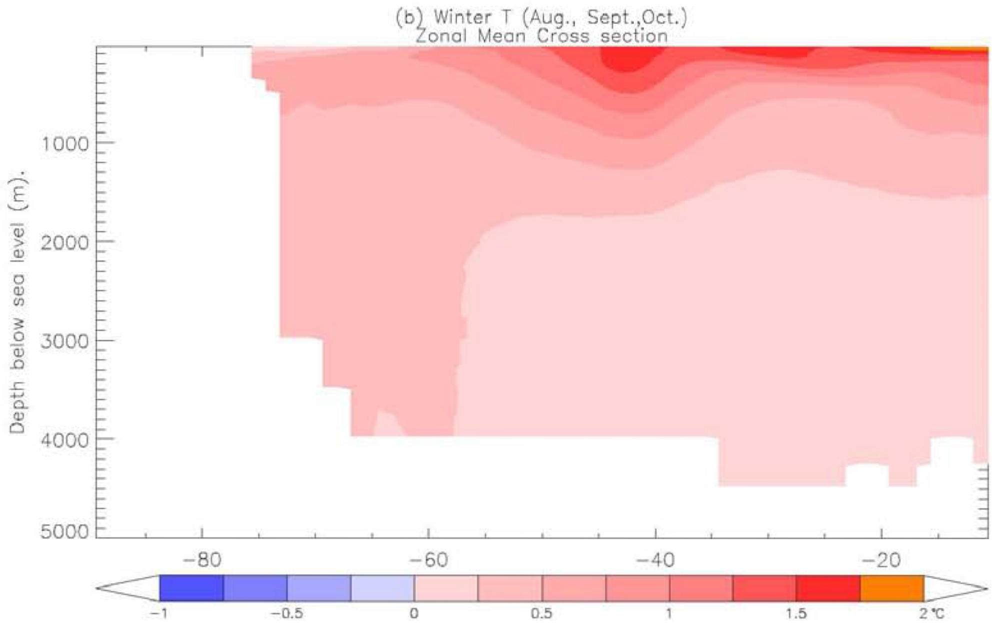


Figure 5.15(v) Zonal mean cross section of ocean temperature difference for (a) summer and (b) winter, between 2000 and 2100, superimposed on bathymetry.

As shown in Figure 5.15(vi), along the continental margin, there is no major change in salinity, except above depths of around 400m, above which surface waters are fresher than at present by up to almost 0.2 at the surface between the coast and about 45°S. The freshening is clearly very much a surface phenomenon, although the Antarctic Intermediate Water is also slightly fresher.

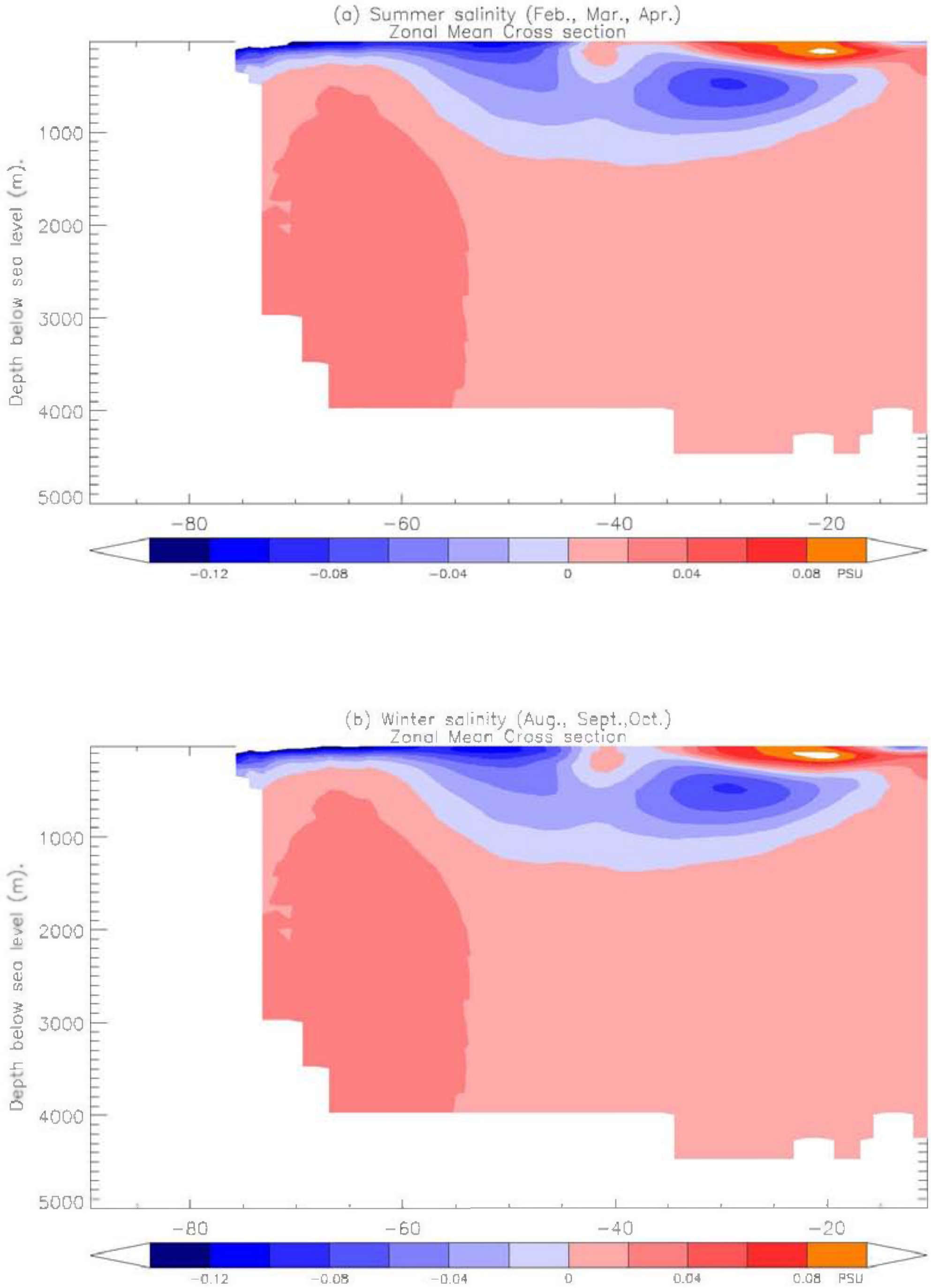


Figure 5.15(vi) Zonal mean cross section of salinity difference for (a) summer and (b) winter, between 2000 and 2100, superimposed on bathymetry.

The vertical stratification increases in most of the models during the Twenty First Century because of the surface warming and the freshening southward of 45°S (Figures 5.15(v) and 5.15(vi)). This surface salinity decrease, which has a dominant impact on the density changes at high latitude, is caused by the increase in precipitation minus evaporation at high latitude. Changes in freshwater transport by ocean currents and sea ice could also play a strong role as shown by Bitz et al. (2006) in their analysis of the results of the CCSM3 model. The enhanced stratification tends to reduce vertical exchanges and is responsible for a decrease in the vertical heat transfer to the surface from the relatively warm water at depth. This effect contributes to the moderate surface temperature increase simulated in the Southern Ocean during the Twenty First Century. Changes in the stratification alter the isopycnal diffusion in models by modifying the slope of the isopycnals, resulting in a reduction of the heat transport towards the surface (e.g., Gregory, 2000; Bitz et al., 2006). By contrast, through the same mechanisms, the stratification increase is responsible for a warming and an increase in salinity of the Southern Ocean at mid-depth in many models.

As a result of the surface density decrease, the ocean ventilation and in particular the formation of Antarctic Bottom Water decrease in many models although the magnitude of the changes in AABW is strongly variable among the models (e.g., Manabe et al., 1991, 1993; Hirst, 1999; Bates et al., 2005; Bitz et al., 2006; Bryan et al., 2006). This lower ventilation in response to the increase in greenhouse gas concentration in the atmosphere induces a positive feedback mechanism by reducing the heat storage in the Southern Ocean as well as reducing the uptake of carbon dioxide, more carbon remaining in the atmosphere during the Twenty First Century compared to a case with constant ventilation of the Southern Ocean (e.g., Sarmiento et al., 1998; Matear and Hirst, 1999).

Vertical mixing does not decrease over the Twenty First Century at all locations of the Southern Ocean in all models. Increased wind stresses could result in deeper mixed layers in some areas. The decrease in ice extent could induce stronger cooling in winter that leads to stronger mixing at the new ice edge. Shifts in convection patterns have also been noticed, with decreased mixing in some areas but increases in others (e.g., Bitz et al., 2006; Conil and Menéndez, 2006).

Ocean ventilation could be enhanced because of the surface divergence induced by the increase in the wind stress projected during the Twenty First Century. Russell et al. (2006b) argue that this effect could be significantly underestimated in some models because of westerlies located too far north in the control climate. By comparing two model versions of the GFDL model, they show that the model version displaying a too northerly maximum of the zonal wind stress simulates much lower heat (and carbon dioxide) storage in the Southern Ocean during the next centuries than does the version with more realistic winds. In the latter simulation, the effect of the enhanced wind-driven divergence is indeed strong enough to overwhelm the influence of the increased stratification at high latitude.

5.3.3 Long-term evolution of the Southern Ocean

Only a few studies have addressed the evolution of the Southern Ocean on timescales longer than a century. To do so, numerical experiments have been conducted in which CO_2 concentration is progressively increased during a few decades and then maintained constant over several centuries. All those numerical experiments show a strong warming of the Southern Ocean (Hirst et al., 1999; Bi et al., 2001; Goosse and Renssen, 2001; Bates et al., 2005; Petoukhov et al., 2005) as well as a strengthening of the ACC transport through Drake Passage (Bi et al., 2002; Bates et al., 2005). The ocean surface temperature changes and the sea ice shrinking obtained after a few centuries are in general as high or even higher in the Southern Ocean than the ones obtained in the Arctic. Indeed, the oceanic heat storage and the reduction of the oceanic heat transport towards the surface that are responsible for a weaker response in the Southern Ocean during the Twenty First Century are less operative during the third millennium**.** As a consequence, on multi-century timescales, the Southern Ocean is one of the regions of the globe that experiences the largest warming in those simulations.

The long-term evolution of the Southern Ocean is also associated with changes in ocean currents. As a first step, the general decrease found in AABW formation over the Twenty First Century in many models continues during the following centuries (e.g., Bates et al., 2005). This eventually leads in some simulations to a complete cessation of AABW formation (Hirst et al., 1999; Bi et al., 2001). However, in a second step, the long-term reorganisation of the Southern Ocean leads to an increase of AABW formation (Bi et al., 2001; Bates et al., 2005). This is due to a long term warming of the ocean at depth, which weakens the stratification and finally allows deep-water formation to start again or to be enhanced (Bi et al., 2001). The disappearance of sea ice could also play a role because very strong atmosphere-ocean interactions take place in the zones close to the Antarctic coast that become ice-free even in winter, leading to new sites of deep water formation and thus an enhancement of deep water production (Bates et al., 2005). It has been suggested that those long term changes in ocean circulation are associated with an increase in heat transport and play a significant role in the long-term warming simulated at high southern latitudes (e.g. Goosse and Renssen, 2001). Unfortunately, although this long term shift from a decrease of AABW formation to an increase is found in nearly all the available long-term simulations, the timing of the reversal is strongly depending on the model, ranging from a few centuries at most (Goosse and Renssen, 2001; Bates et al., 2005) to more than a millennium (Hirst et al., 1999; Bi et al., 2001). Furthermore, the freshwater flux from the melting of the Antarctic ice sheet, which could strongly influence stratification and deep water formation in the Southern Ocean on this time-scale, is included in none of those simulations, increasing our uncertainties in the long-term evolution of the system. A recent study (Swingedouw et. al., 2009) using a coupled climate-ice sheet model has shown that taking into account the long-term influence of the freshwater flux from the melting of Antarctic ice sheet contributes to the formation of a cold halocline in the Southern Ocean. This limits sea ice cover retreat under global warming and reduces local surface warming compared to simulations where this effect is not taken into account.

5.3.4 Conclusions

Forecasts of future conditions are made using coupled models. The wide range of results obtained by state of the art models indicates that there are still significant deficiencies in the different models - mainly due to inadequate representations of relevant processes (for example through inappropriate parameterisation), or to insufficiently high resolution. To a large extent model validation occurs through model intercomparison, simply because observations are scarce, which means that arguments may be circular. Surface data from satellites are available for some parameters with sufficient coverage in time and space, but in-situ data from the ocean interior are still patchy in space and intermittent in time. In the ice-covered areas of the Southern Ocean, in-situ data are still almost exclusively collected in the context of research expeditions determined in time and space by initiatives to study processes rather than to make repeat measurements, and limited by funding cycles. Operational observations of climate relevant data are still in their infancy, even though the Argo system is a big step forward. Although excellent technologies are available, with a wide range of autonomous devices, such as moored systems, floating systems, gliders, or sensor-carrying animals, the deployment and maintenance of these technologies in a sustained manner is still out of sight.

To reach that point a Southern Ocean Observing System (SOOS) has to be established. This will require the cooperation of agencies and research institutions in coordinating the use of their resources, but it requires at the same time further development of instrumentation in order to reduce the effort needed to obtain the required measurements with the appropriate time and space resolution. The provision of climate-relevant observations from an established network will release resources from the research community that can be used for process studies in order to improve the representation of those processes in the models.

5.4 Sea ice change over the Twenty First Century

The average of the CMIP3 model's sea ice extent compares well with observations, although there is a large inter-model spread (Arzel et al., 2006; Parkinson et al., 2006).

Models following the A1B scenario show that over the Twenty First Century the annual average total sea ice area is projected to decrease by 2.6×10^6 km^2, or 33% (Bracegirdle et al., 2008). There is strong consensus among the models for an Antarctica-wide decrease in sea ice; the inter-model standard deviation is low at 0.73×10^6 km^2 (9%). Arzel et al. (2006) assessed different measures of sea ice amount using a different subset of 15 of the CMIP3 A1B projections and found Twenty First Century deceases of 34% for sea-ice volume and 24% for sea-ice extent.

Most of the simulated ice retreat occurs in winter and spring when the sea ice extent is largest (Figure 5.16). The amplitude of the seasonal cycle of sea ice area will therefore decrease. The smaller amount of seasonal ice melting/freezing will affect the ocean due to changes in processes such as brine rejection.

There is strong confidence in the projected Antarctic-wide decreases of sea ice extent. At a more regional level, decreases of sea ice are less significant (e.g. Lefebvre and Goosse, 2008). One way to measure the significance of a projected change is to calculate the signal to noise ratio of that change. Here the signal is the ensemble average change, and the noise is the standard deviation of the inter-model spread. A change can be thought of as 'significant' if larger than the inter-model standard deviation, i.e. a signal to noise ratio of greater than one. In the regions where sea ice currently remains present throughout the summer, in particular the Weddell Sea, large reductions of sea ice extent are projected. On this there is quite a strong consensus between the models, with the model average reductions larger than the inter-model standard deviation (Figure 5.16). However, at the scale of one grid point the confidence of a reduction of sea ice over the Twenty First Century is not significant in many regions (i.e. the magnitude of the change is smaller than one standard deviation of the inter-model spread).

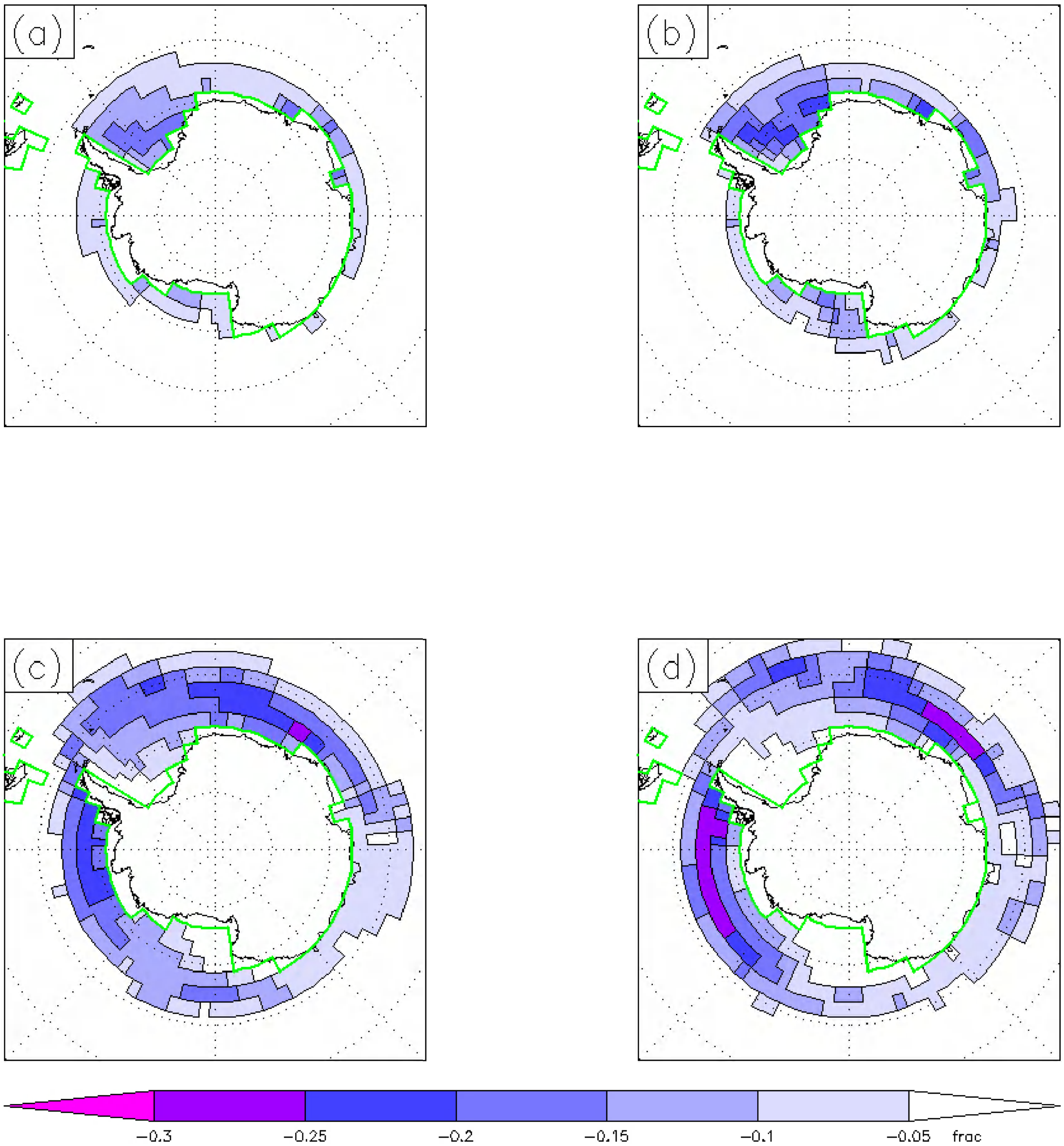


Figure 5.16 Twenty First Century sea ice concentration change for (a) DJF, (b) MAM, (c) JJA and (d) SON, showing the difference between the 2080-2099 mean and 2004-2023 mean. Changes are shown in terms of the fraction of the surface covered by sea ice, rather than sea ice percentage, since a spatial plot of sea ice percentage change would show infinite (or very large) increases where concentrations were initially zero (or very small).

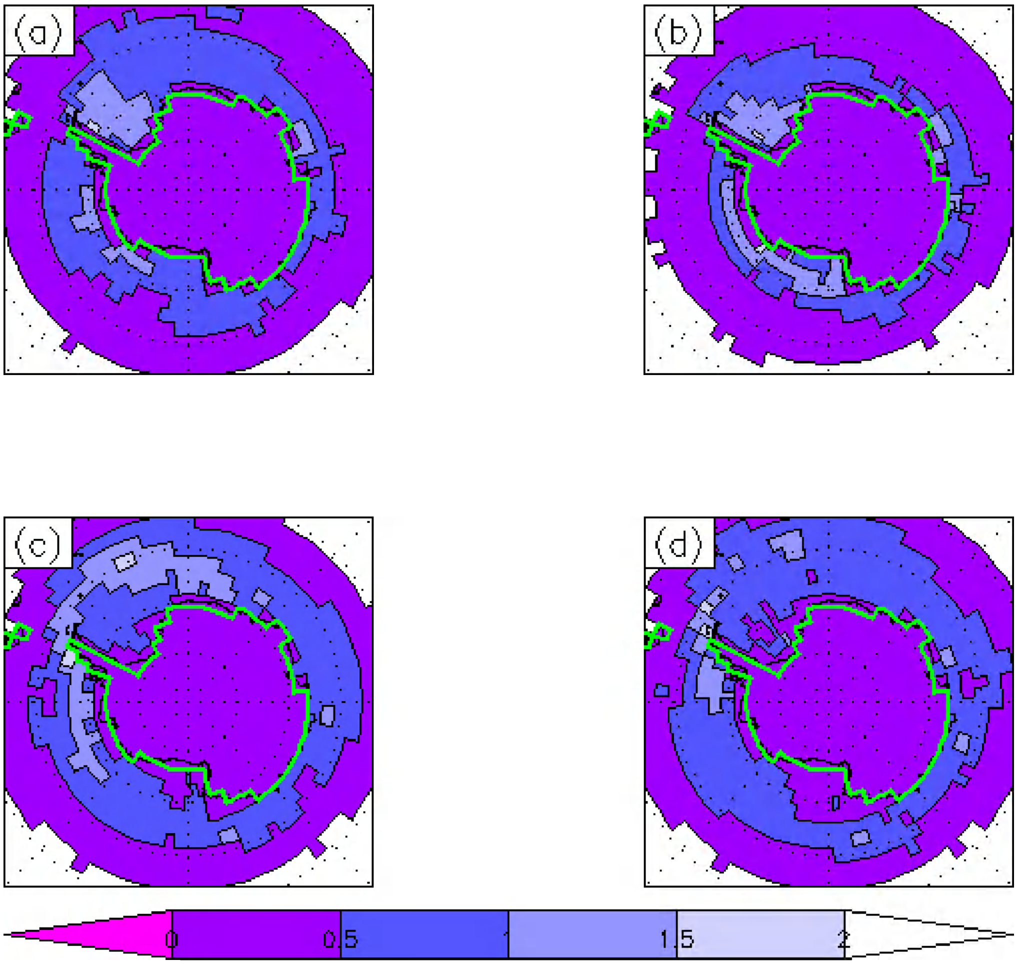


Figure 5.17 Signal to noise ratio of projections of sea ice reduction.

5.5 The terrestrial cryosphere

5.5.1 Introduction

Recent observations of ice sheet behaviour have forced experts to radically revise their view of ice sheet sensitivity to climate. Existing models, based primarily on earlier views of the essential governing dynamics of ice sheet flow, fail to reproduce observed behaviour, removing a primary tool for anticipating future changes. Predictions of the future state of ice sheets must, therefore, be based upon a combination of inference of past behaviour, extension of current behaviour, and whatever proxy data and analogues exist. The broad range of time scale response of the ice sheet guarantees that future behaviour will be composed of a superposition of the continuing gradual response to past climate change and the more rapid responses to present and future changes.

Currently there exists a patchwork of changes across the Antarctic ice sheet - growth in some areas, loss in others, some related directly to climate, some related to changes in oceans, driven by climate change. No single external driver, nor single aspect of climate change (atmospheric temperature, snowfall rate, or ocean conditions) will dominate in all areas. Rather the dominant effect in each area will depend on climate in that area, and the particular sensitivities of the ice sheet in that region.

Most certain is the expectation that in a warmer world there will be less ice and higher sea level. Five million years of paleoclimate data support this expectation (Naish et al., 2009;

Pollard and DeConto, 2009). Both data and modelling indicate the loss of the West Antarctic Ice Sheet during periods prior to one million years ago in which global temperature was ~3° warmer than pre-industrial levels. The data also demonstrate that sea-level change, and therefore the rate of ice loss, will be neither uniform nor monotonic. The implication is that the loss of ice and, thus, future ice-sheet behaviour, will be episodic. Translated into a picture of future ice sheet behaviour, this suggests there will be a combination of coherent activity and disjoint activity from different regions of the continent, among individual glacier basins and groups of glacier basins.

The most likely regions of near-future change are those that are changing most today (Antarctic Peninsula, and Pine Island Glacier). Most models agree that future warming in the Antarctic will be strong, and so it is likely that – depending on location - both snowfall and melt will increase later this century. Even small amounts of additional accumulation would cause significant volumetric growth due to the vast area of the ice sheet; average accumulation across all of Antarctica is only 15 cm/yr (water equivalent) and much lower in the cold, high interior of East Antarctica. But warming (especially around the margins) also leads to increased melting that not only would remove ice mass by runoff, but could cause the margin of parts of Antarctica to adopt some of the dynamic character of the present-day margin of Greenland, where surface meltwater penetrates to the ice sheet bed causing accelerated flow (Zwally et al., 2002b; Joughin et al., 2008). It must be borne in mind that models disagree over many other details of atmospheric climate change in Antarctica, and generally share little agreement in precise projections of future ocean warming, which might have the most significant influence of all on the ice sheets.

5.5.2 East Antarctic ice sheet

Present changes in the East Antarctic ice sheet (EAIS) are a patchwork of interior thickening at modest rates and a mixture of modest thickening and strong thinning among the fringing ice shelves (see Figure 4.34). As discussed in Chapter 4, the cause of the current slight interior thickening is probably a long-term dynamic response to a distant change in climate (e.g. a past cooling event), and not a response to recent increased snowfall. This effect is likely to continue and change only slowly. Eventual atmospheric warming over the East Antarctic interior is projected in most GCMs as ozone depletion is reversed, but there may not be a straightforward connection between the expected atmospheric warming and an increase in snowfall over the Antarctic continent. Many GCM projections show a similar degree of sensitivity; that a 1 °C increase in mean annual temperature would cause around 5 % increase in mean net surface accumulation (equivalent to 0.3 mm annual decrease of global sea level) (Meehl et al., 2007). In the most recent IPCC assessments, this contribution to sea level due to increasing snowfall was included, and was highly significant. Without its effect, sea level rise projections would generally be 5 cm higher.

Coastal changes are more difficult to anticipate. Most of the additional snowfall may be limited to the coastal areas, compensating for present processes responsible for the observed thinning of ice shelves, however the compensation will likely only be partial. Equally likely is an amplification of the present rapid thinning of the Cook Ice Shelf and the mouth of the Totten Glacier, spreading of ice shelf thinning to other coastal areas of the EAIS, and perhaps isolated initiations of summer acceleration of grounded coastal ice by lubrication of meltwater penetrating to the ice sheet base where summertime temperatures exceed 0°C.

There are marine basins beneath the East Antarctic ice sheet, especially in Wilkes Land (between 100°E and 160°E), and the potential for these to harbour even more dramatic and rapid ice loss remains unquantified. The general discussion of this potential is presented in the next section because it is already happening in West Antarctica.

5.5.3 West Antarctic ice sheet

The current loss of mass from the Amundsen Sea embayment of the West Antarctic ice sheet is equivalent to that from the entire Greenland ice sheet (50 Gt per year) (Lemke et al., 2007). This sector of the West Antarctic ice sheet is by far the most active, well ahead of the Ross Sea sector where the stagnation of Kamb Ice Stream dominates the positive mass balance contribution from that sector. Little net change is measured from the remaining Weddell Sea sector.

The future of West Antarctica is always cast against the backdrop of the "marine-based ice sheet instability", a concept first posed by Weertman (1974). Accelerating and irreversible retreat of marine-based glaciers resting on back-sloping beds has been confirmed by a more detailed, full-stress tensor, numerical analysis (Schoof, 2007). The observed doubling of the Pine Island Glacier's speed in less than 30 years demonstrates that marine based outlet glaciers are capable of dramatic accelerations in the relatively short period of a few decades (Joughin et al., 2003). Figure 5.18 illustrates one model's prediction of changes in the Amundsen Sea sector over the next millennium.

The process believed responsible for this dramatic behaviour is a gradual thinning of the fringing ice shelves seaward of the Amundsen Sea outlet glaciers - in effect, a slow motion version of the glacier acceleration that was observed to follow ice shelf disintegrations in the Antarctic Peninsula. The chain of events leading to this thinning begins with increased circumpolar circulation in the atmosphere above the Southern Ocean, driven by the increased pressure gradient between the ozone hole cooled Antarctic and the warmer Southern Hemisphere mid-latitudes. The surface waters drift northward due to the Coriolis effect, encouraging a greater upwelling of warm Circumpolar Deep Water. Now raised, these warmer waters are able to get onto the continental shelf, where their flow is directed toward the outlet glaciers by following the troughs carved by these glaciers in past glacial periods. Once they reach the floating ice shelves, they are responsible for extremely high melting rates of many tens of metres per year.

GCM predictions are for a continuation of the positive phase of the Southern Annular Mode, which will continue the stronger circumpolar circulations, thus continuing to upwell warmer waters onto the continental shelf in the Amundsen Sea. A doubled outflux in the glaciers in this sector would contribute to an extra 5 cm of sea-level rise per century. Ultimately, this sector could contribute 1.5 metres to global sea level, so a contribution from this sector alone of some tens of centimetres by this century's end cannot be discounted.

The relationship between sub-ice melt rates and ocean temperatures is just beginning to be explored. A recent modelling study by Pollard and DeConto (2009) suggests that the West Antarctic ice sheet will experience a complete collapse when nearby ocean temperatures warm by roughly 5°C. A corollary to this is that if ocean circulation patterns change and warmer portions of the ocean access the ice shelves, an equivalent collapse would unfold. Global climate and regional ocean modelling is needed to predict when and if future ocean temperatures and melt rates under the Antarctic ice shelves will increase to that extent, and improved ocean-ice models are required to explore whether the details of this interaction increase the thermal sensitivity of this system. Current models (described earlier in this chapter) suggest that there will not be a full collapse in this century.

Other sectors of the West Antarctic ice sheet are doing little to offset this potential contribution to sea-level rise. There is modest ice sheet growth within the Ross Sea sector as the basin of the near-stagnant Kamb Ice Stream continues to thicken through snowfall and the adjacent Whillans Ice Stream slowly decelerates. Continued deceleration of Whillans Ice Stream is likely—probably the result of freezing and stiffening of the subglacial till. Contributions to ice sheet growth are limited to the rate of snowfall, so even a second stagnant ice stream will not be able to offset the ice lost from the Amundsen Sea sector, and there is the intriguing probability that the Kamb Ice Stream will reactivate - an event that is not predictable, but one that has scientists' attention and that is being monitored by a host of sensors.

Measurements made in the Weddell Sea sector of West Antarctica do not raise alarms now or for the future. The open ocean is held well away from where ice streams first enter the Filchner-Ronne Ice Shelf. Like ice streams from the Ross Sea sector that feed the Ross Ice Shelf, the vast size of these ice shelves is probably the greatest insurance that sudden changes in the deep Southern Ocean will not greatly impact the flow rates of these major ice flows this century. Substantial thinning of even portions of these large ice shelves would require a re-evaluation of the potential impact of the considerable reservoir of ice held upstream.

The empirical observation that glaciers in the northern Peninsula flowed faster following the collapse of an ice shelf provides a possible analogue for the changes occurring in the Amundsen Sea sector of West Antarctica. There, the collapse of a substantial ice shelf has not been observed, but the surrounding ice shelves are thinning at rates in excess of 5 metres per year in places (Shepherd et al., 2004; Bindschadler, 2002). This thinning has perhaps caused ice to detach from the bed, or reduced the side-drag of ice shelves, decreasing resistance to the motion at the front of the Pine Island and Thwaites glaciers, in the same way that resistance from the Larsen ice shelf was removed by its disintegration. Models of ice flow suggest that this effect could provide an explanation for the acceleration of these glaciers (Payne et al., 2004; Thomas et al., 2004b).

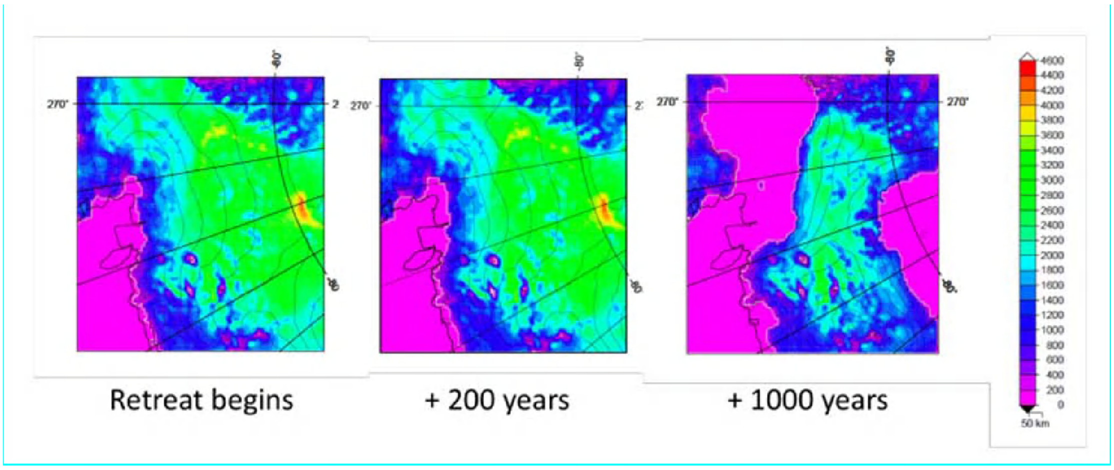


Figure 5.18 Prediction of the changes in ice thickness and surface elevation of the Amundsen Sea sector of West Antarctica based on the UMISM model. Colours represent ice thickness (in metres) and contour lines represent elevation. Initial state (left), and 200 yr projection (middle) and 1000 yr projection (right) assuming no ice shelf buttressing from the Amundsen Sea ice shelves (provided courtesy of J. Fastook).

The potential for counterintuitive behaviour of the Antarctic ice sheet does exist and must not be dismissed lightly given our limited understanding of recent ice-sheet behaviour. As one example, the supply of warm water that drives sub-ice-shelf melting beneath the Filchner-Ronne Ice Shelf is driven by sea-ice production in the Weddell Sea. Warmer or less windy conditions in the Weddell Sea would reduce sea-ice production, and could therefore decrease ocean overturning and reduce the delivery of warm water to the ice shelf, leading to less sub-ice-shelf melting. Another example is that in places along its ice front, the gap between the Ronne Ice Shelf and the seabed is less than 50 meters. Any substantial increase in thickness, for example caused by an increase in discharge upstream, could quickly cause new areas of ice shelf/seabed contact, decelerating ice-shelf flow, and resulting in thickening upstream and a local advance of the ice-sheet.

5.5.4 Antarctic Peninsula

Within the Antarctic Peninsula, recent climate changes have had a dominant impact on the behaviour of glaciers and ice shelves. That being the case, predictions of future climate can be used to help predict the future of this ice. Unfortunately, there is a demonstrable lack of skill in current GCMs in simulating recent changes on the Antarctic Peninsula, because they do not work well at the regional scale, and thus poor confidence in future projections of ice behaviour there at that scale.

Nevertheless, some mechanisms for recent warming are understood. Warming rates vary seasonally. Winter warming, related to a loss of sea ice to the west, exceeds summer warming, caused by an increase in the Southern Annular Mode. Both these trends cannot continue indefinitely (winter will never become warmer than summer). More precipitation occurs in winter than summer, so warmer winters will probably bring more snowfall. Yet the number of days of melt (temperatures above $0^{\circ}C$) is an important parameter tied to glacier growth or shrinkage and these days are projected to increase, offsetting some part of the increased snowfall.

Applying the concept of a thermal limit to ice shelf viability (Vaughan and Doake, 1996), increased warming will lead to a southerly progression of ice shelf disintegrations along both coasts of the Antarctic Peninsula. As in the past, these may well be evidenced and preceded by an increase in surface meltwater lakes. Prediction of the timing of ice shelf disintegration is not yet possible, but field programmes on some of the ice shelves thought to be most vulnerable to collapse now join a host of satellite sensors that monitor ice shelf health and watch for the expected signs of disintegration. These data should improve understanding of the causal mechanisms and lead toward a predictive capability. With a total volume of 95,200 km^3 (equivalent to 242 mm of sea level; Pritchard and Vaughan, 2007), roughly half that of all glaciers and ice caps outside of either Greenland or Antarctica, the mechanism of fast acceleration of these glaciers as an immediate consequence of ice shelf disintegration means this ice could impact global sea level faster than the more gradual melting of all other glaciers and ice caps.

Glaciers on the Antarctic Peninsula have been generally overlooked in global projections of sea-level rise (e.g. Meehl, et al., 2007). In assessing their likely future contribution to sea-level rise it is important to identify glacier volume changes directly, and when negative mass balance began. Other parts of the world have long-running glacier monitoring programmes. For the Antarctic Peninsula such datasets do not exist, although it is hoped that using photogrammetric measurements from historic photographs it may be possible to determine changes in glacier surface height, and hence volume and mass balance over past decades (e.g. Fox and Cziferszky, In Press).

5.5.5 Conclusions

Efforts to measure the mass balance of ice sheets have borne fruit in recent years with multiple methodologies that have provided a picture of increasing ice loss in the past one or two decades. The rapid rate of change makes it difficult to extrapolate these observations into the future. These same observations have revealed unsettling weaknesses in most time-dependent ice sheet models, reducing confidence in their predictions of the future of the Antarctic ice sheet.

Nonetheless, a conservative expectation of changes no greater than those already observed results in an increasing loss of ice in the Antarctic Peninsula and in the Amundsen Sea sector of West Antarctica, more than offsetting the slow growth in East Antarctica and the stagnant Kamb Ice Stream in West Antarctica. More disturbing is the possibility, albeit impossible to quantify at this time, of a number of climate influences that could amplify loss of Antarctic ice and accelerate future sea- level rise.

Most past predictive efforts on sea-level rise, such as those by the IPCC, have aimed at the predictable components, shying away from the less predictable elements, especially the response of continental ice sheets to climate change. Efforts are now being made to develop more integrated models that incorporate some of the ice-climate interactions that are now inferred as central to recent changes. In some cases, more field work is required before a sufficiently deep understanding of the process is possible. However, simplified schemes can be introduced to numerical models now for use in future assessments by the IPCC and others. For the moment there are no comprehensive, objective projections that can be cited, and the future evolution of the Antarctic ice sheet is better described through a more subjective, discursive approach.

5.5.6 Summary and needs for future research

The instrumental period has been a period of accelerating data collection culminating in the recognition of astonishingly rapid physical changes. These achievements have been fueled primarily by the array of extraordinarily capable satellite sensors. The availability of these data for research has continued to increase through a variety of data sharing agreements, further increasing their value. The research community has become increasingly reliant on their existence and availability. However the continuation of these exceptional resources cannot be taken for granted and many of the most valuable sensors, such as those on ICESat, MODIS and Landsat, are already beyond their design lifetime, with replacement missions either not planned or planned for launches so far in the future that prolonged gaps in coverage are probable. These gaps must be minimized and, if possible, eliminated. The prospect of slowly going blind to ice sheets at precisely the moment when their behaviour has suddenly become very dramatic carries with it the undesirable consequences that not only will we not be able to follow the continuing evolution of areas already changing dramatically, but we will not be able to detect new areas of change at an early stage, limiting our ability to understand the causes of these changes. Recommendations for satellite observations of the cryosphere are given in the Cryosphere Theme document produced for the Integrated Global Observing Strategy (IGOS) Partnership (http://www.eohandbook.com/igosp/cryosphere.htm).

As the IPCC's AR4 noted, climate models lack the ability to predict the future of ice sheets primarily because our understanding of ice flow dynamics cannot predict "future rapid dynamical changes in ice flow". Satellite data have identified the signatures of change in many regions and helped scientists suggest possible causes, however airborne and field studies are the only means to determine the root causes and quantify the sensitivity of the ice sheet to them. The major processes currently under investigation are: the disintegration of ice shelves and the consequent acceleration of feeding glaciers; ocean-ice interaction beneath the fringing ice shelves in both relatively warm and cold water environments; and the influence of a newly appreciated active subglacial hydrological system on ice flow. In all these areas, the emphasis is on directly accessing the active region, usually with attempts to maintain instruments that can survive to produce temporally extensive records of key parameters. These field studies are essential and must be pursued. Their success will be a quantitative understanding of these processes at a level that allows models to incorporate the key processes of ice flow, thus leading to improved predictions of future ice flow and ice sheet shape.

5.6 Evolution of Antarctic permafrost

Although we do not envisage a major reduction in permafrost area over the next 100 years, melting of ground ice may lead to subsidence over some 15,000 km^2 of Antarctica's ice-free regions. Areas that are particularly susceptible to this effect, known as thermokarst, occur in coastal areas including Casey Bay near Molodezhnaya Station (70.5°S, 12°E), the Pennell-Borchgrevink Coasts in North Victoria land (70.5-73°S, 165-171°E), the Scott Coast in the McMurdo Sound area (74-78°S, 165°E), and throughout the Antarctic Peninsula and its offshore islands (55-72°S, 45-70°W).

In recent years there have been significant changes in hydrologic and geomorphologic processes as a consequence of unusual local summer warming events. For example, in the McMurdo Dry Valleys (MDV), the mean summer temperature (December and January) during the period 1994 to 2003 was -0.19°C, and there were 30 days per year during that period in which the mean daily temperature was above 0°C. During December 2000-January 2001, the mean temperature was 1.5°C, and there were 43 days in which the mean daily temperature exceeded 0°C. This protracted warming event resulted in flooding of rivers and expansion of inland lakes. (see also Foreman et al., 2004). These extreme events may have long-lasting effects. For example, after the December 2001 "wet event" in the MDV, soil moisture in Taylor Valley remained about twice the level of the preceding 9 years for a 4-year period (Barrett, unpublished). Observations suggest that extreme events in the MDV may alter subsurface flow and increase the hyporheic zone (the region beneath and lateral to a stream bed, where there is mixing of shallow groundwater and surface water), cause flushing of salts from soils, and reactivate sand wedges and ice wedges.

About 90% of the year-round summer bases in Antarctica are in areas sensitive to thermokarst formation and mass wasting. For that reason, the effect of climate warming on melting permafrost should be of concern to the Council of Managers of National Antarctic Programs (COMNAP). Examples exist of unusual warming at McMurdo Station causing road failure and undercutting pilings for utility corridors.

In the Arctic and central Asian regions, the melting of permafrost carries with it the threat of the release of the greenhouse gas methane. There is no significant risk of that in the Antarctic, where the soils, particularly in interior regions, contain very small amounts of organic C (<0.05%).

5.7 Projections of sea level in Antarctic and Southern Ocean Waters by 2100

The two major reasons for sea-level rise are expansion of ocean waters as they warm (and an associated decrease in ocean density) and an increase in the ocean mass, principally by melting of land-based sources of ice (glaciers and ice caps and the ice sheets of Greenland and Antarctica). The amount of thermal expansion is non-uniform due to the influence of ocean currents and spatial variations in ocean warming. Global warming from increasing greenhouse gas concentrations is a significant driver of both contributions.

The IPCC provides the most authoritative information on projected sea-level change. The IPCC Third Assessment Report (TAR) (Church et al., 2001) projected a global averaged sea-level rise ranging from 9 to 88 cm between 1990 and 2100 using the full range of IPCC greenhouse gas scenarios, a range of climate models and an additional uncertainty for land-ice changes. For the IPCC's Fourth Assessment Report (IPCC, 2007), the range of sea-level projections, using a larger range of models, was 18 to 59 cm (with 90% confidence limits) over the period from 1980-1999 to 2090-2099 (Meehl et al., 2007). This rise in sea level is mainly a result of thermal expansion of the upper ocean and from glaciers and ice caps, with little contribution from Greenland and Antarctica. However, as mentioned earlier, ice sheet models are incomplete and do not allow for a rapid dynamic response of the ice sheets. To allow for ice sheet uncertainties, IPCC's AR4 increased the upper limit of projected sea-level rise by 10 to 20 cm, stating that 'larger values cannot be excluded, but understanding of these effects is too limited to assess their likelihood or provide a best estimate or an upper bound for sea-level rise.' The end result is that the upper bound of the IPCC TAR and AR4 projections from 2001 and 2007 respectively are similar (Figure 5.19).

Since 1993, there have been high-quality satellite-altimeter observations of sea level over most of the globe (about 65°N to 65°S), allowing accurate estimates of both global-averaged and regional sea-level change. Global correlation patterns (empirical orthogonal functions) estimated from the satellite altimeter record have been combined with coastal and island tide-gauge data (corrected for glacial isostatic adjustment) to estimate global-averaged sea levels since 1870 (Church and White, 2006). The results show that, from 1870 to the present, global sea level has risen by about 20 cm, at an average rate of 1.7 mm/yr during the Twentieth Century, with an increase in the rate of rise over this period (Figure 5.19). Jevrejeva et al. (2006) and Holgate and Woodworth (2004) have used quite different techniques of analyzing historical tide-gauge data and have found quite similar historical rates of sea-level rise. For the modern satellite period (since 1993), sea level has been rising more rapidly at an average rate of 3.1 mm/yr (IPCC, 2007). The latest estimates of sea level rise, for 2003-2008, based on GRACE space gravimetry measurements, show that that rate of sea level rise has now slowed to 2.5 mm/yr (Cazenave et al., 2009). Note that these rates of increase are an order of magnitude faster than the average rate of rise over the previous several thousand years, but significantly slower than the rates of rise at the end of the Last Glacial Maximum and at the end of the Younger Dryas event (see chapter 3).

About a third to a half of the sea-level rise during the first decade of the altimeter record can be attributed to thermal expansion due to a warming of the oceans; the other major contributions include the combined effects of melting glaciers and ice sheets. Changes in the storage of water on land (such as the depletion of aquifers and increases in dams and reservoirs) remain very uncertain.

Concern that the IPCC sea-level projections may be biased on the low side has been reinforced by the increase in the rate of rise of sea level since the early 1990s. The rate of 3.1 mm/yr noted in the IPCC (2007) report is faster than the central range of the IPCC projections and at the very upper end of the IPCC projections (Rahmstorf et al, 2007). This suggests that one or more of the model contributions to sea-level rise is underestimated (see inset diagram in Fig 5.19). Given these observations, Rahmstorf et al. (2007) projected a likely maximum sea-level rise of 1.4 m by 2100. Future sea level increases projected by the IPCC (2007) do not include possibly large contributions resulting from the dynamic instability of ice sheets during the Twenty First Century. Taking such variables into consideration, Pfeffer et al. (2008) estimate an upper bound of 2 m of sea-level rise by 2100.

Sea-level projections to and beyond 2100 are critically dependent on future greenhouse gas emissions, with both ocean thermal expansion and the ice sheets potentially contributing metres over centuries for higher greenhouse gas emissions. There is widespread acceptance that at peak past interglacial global warmings of 2-3°C above the temperature in 1900 led to rises of sea level of 4-6 m (e.g. see IPCC, 2007; Gregory and Huybrechts, 2006). We do not know enough about either modern processes or future rates to say how soon such rises in sea level might occur again (if at all), but they do not seem likely within the next few hundred years as far as we can tell from all that we know at present. Therefore the current target for mitigation should be a maximum of 2 m, with a likelihood of it being less.

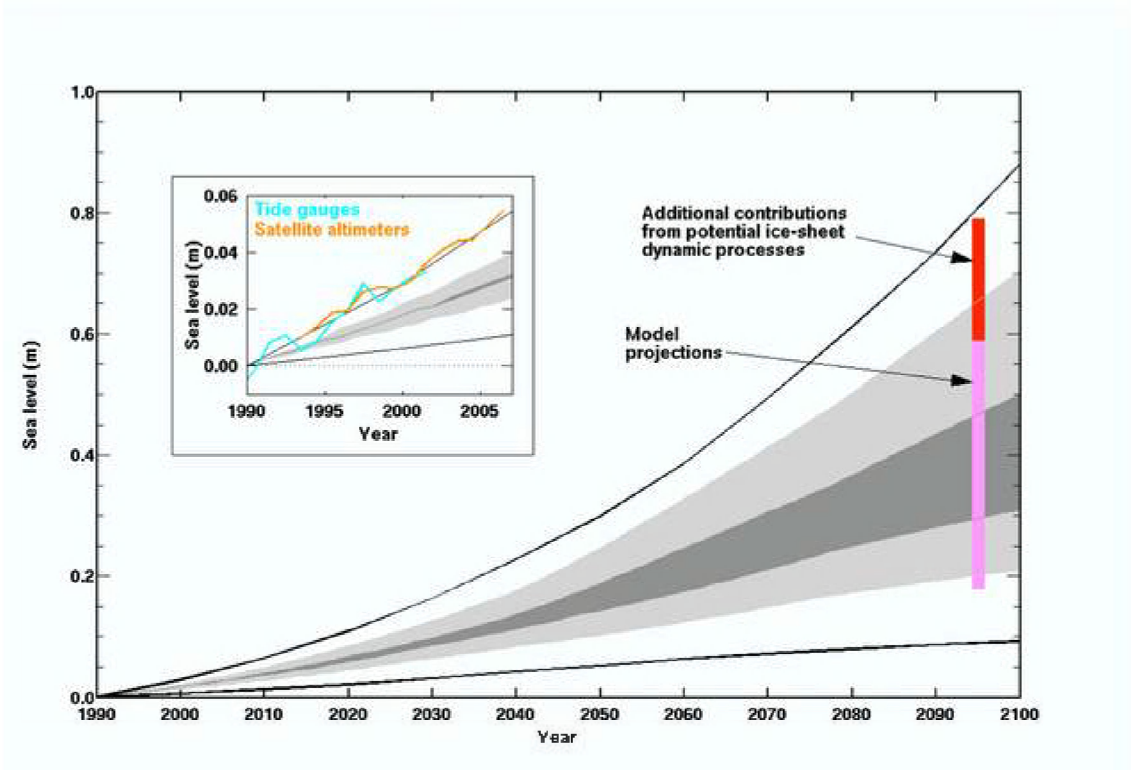


Figure 5.19 Projected sea-level rise for the Twenty First Century. The projected range of global averaged sea-level rise from the IPCC 2001 Assessment Report for the period 1990 to 2100 is shown by the lines and shading. (The dark shading is the model average envelope for all greenhouse gas scenarios, the light shading is the envelope for all models and all scenarios and the outer lines include an allowance for an additional land-ice uncertainty.) The updated AR4 IPCC projections (90% confidence limits) made in 2007 are shown by the bars plotted at 2095, the magenta bar is the range of model projections and the red bar is the extended range to allow for the potential but poorly quantified additional contribution from a dynamic response of the Greenland and Antarctic ice sheets to global warming. The inset shows the 2001 projection compared with the observed rate estimated from tide gauges (blue) and satellite altimeters (orange).

5.7.1 Regional Projections of mean sea-level rise

Overlying the global sea-level rise is a large regional variability, and sea-level rise during the Twenty First Century is not expected to be uniform around the globe. This is a result of changing atmospheric conditions, particularly surface winds (Lowe et al., 2006) and as a result of changes in ocean currents.

The strongest signatures of the spatial pattern of sea-level rise projections in the average of 16 coupled atmosphere-ocean models used for the IPCC AR4 are a minimum in sea-level rise in the Southern Ocean south of the Antarctic Circumpolar Current and a maximum in the Arctic Ocean. The next strongest features are maxima in sea-level rise at latitudes of about 30° to 40°N in the Pacific and to a lesser extent the Atlantic Ocean, and at about 40° to 50°S in all of the southern hemisphere oceans, at the poleward extremities of the subtropical gyres.

The minimum sea-level rise in the Southern Ocean is due to the thermal expansion coefficient being lower in cold waters than warmer waters. For instance if heat is added to the Southern Ocean region, the change in density and sea-level rise will be less than if the same amount of heat had been added in a warmer location (Lowe et al., 2006).

Sixteen of the models used for the IPCC assessment are shown in Figure 5.20 using the mid-range A1B scenario (stabilization of CO_2 equivalent at 720 ppm by 2100), following the approach of Church et al. (2008). The models show relative sea- level rise. The global mean sea-level rise has been removed to reduce some of the impact on the results from different models having different climate sensitivities. All the models show the Southern Ocean region has reduced sea-level increases compared to the global average of 15-20 cm. Four of the models show that in the coastal region close to Antarctica there is a small increase in sea level above the average of 5-10 cm. It is possible that freshwater changes to the local density or changes to local current systems have given rise to the higher sea levels in these coastal regions. These regional changes need to be investigated further in the individual models before firm conclusions can be drawn.

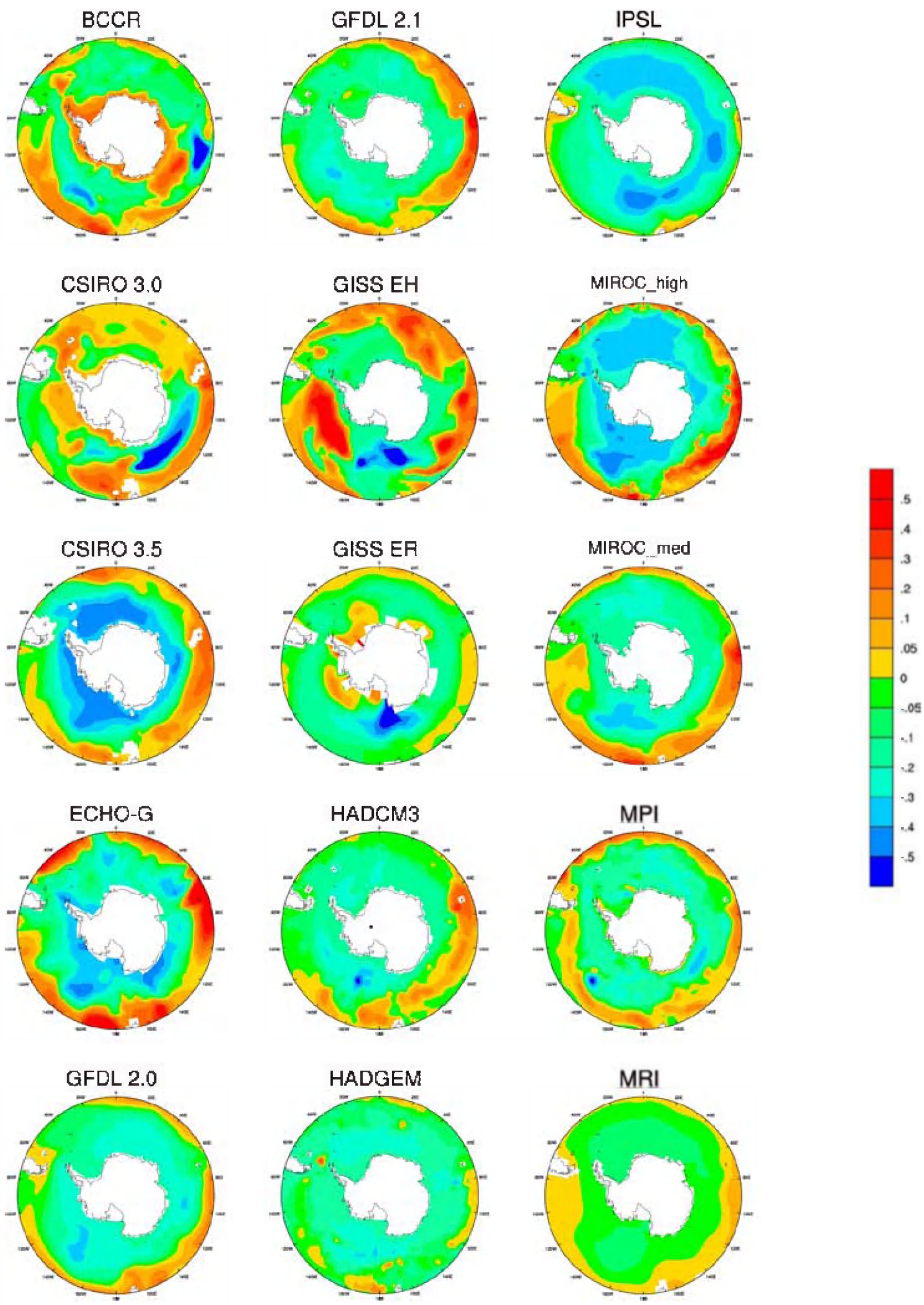


Figure 5.20 Regional Sea level rise projections for 15 IPCC AR4 models for the SRESA1B scenario at 2100 relative to the global mean, following the approach of Church et al. (2008)

Mid-latitude studies emphasize that extremes in sea level will increase in future as wind speeds and storm intensity increases. Less is known about storm surges in the Antarctic region, with only one study from the Ross Sea (Goring and Pyne, 2003) and two from long records in the Falklands and South Georgia (Woodworth et al., 2005; Hansom, 1981). The long record from the Falklands has an upward sea level secular trend of 0.70 ± 0.18 mm year since 1964, about half the observed global figure.

5.8 Biogeochemistry – response of the Southern Ocean carbon cycle to future climate change

5.8.1 Background

Atmospheric CO_2 emissions continue to increase at unprecedented rates. Already emissions have equalled or exceeded the previous worst-case scenarios for the IPCC (Raupach et al., 2007; Figure 5.21). Currently it is estimated that about 30% of the CO_2 emitted annually is taken up by the ocean (Sabine et al., 2004). The Southern Ocean plays a critical role in taking up this CO_2, with more than 40 % of the annual mean uptake of atmospheric CO_2 being taken up in the region south of 40°S (Takahashi et al., 2009). Clearly then the way in which the Southern Ocean responds to climate change will directly impact global atmospheric CO_2 levels and hence the rate of the earth's warming. Hence, determining the role of the Southern Ocean in CO_2 exchange must be a priority if we are to formulate effective global policies for stabilizing atmospheric CO_2 levels.

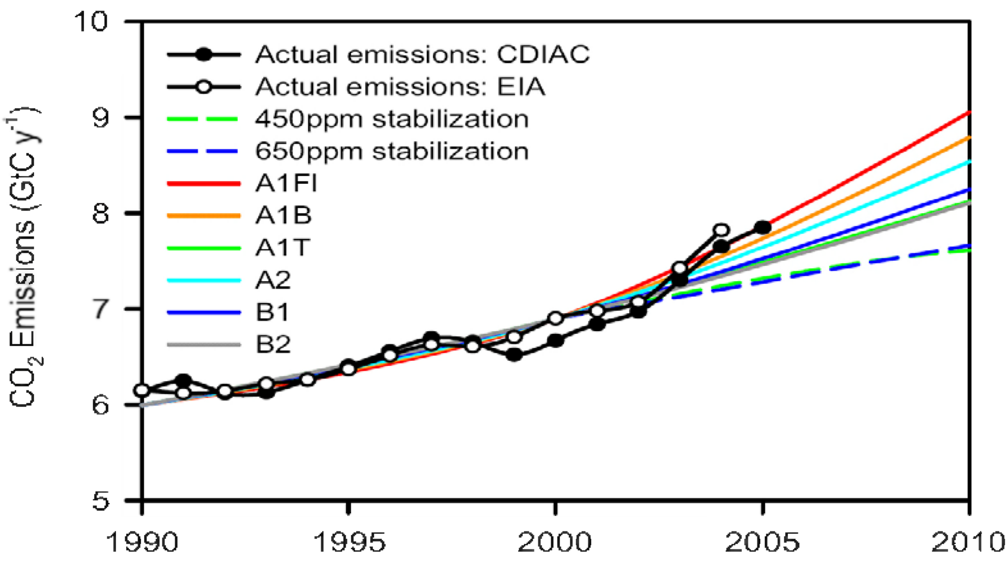


Figure 5.21 Observed CO_2 emissions over the last 25 years as compared with the different IPCC emission scenarios (Raupach et al., 2007) (Copyright (2007) National Academy of Sciences, U.S.A).

5.8.2 Future Southern Ocean carbon response

5.8.2.1 Response to increased winds

As the Southern Ocean becomes windier in response to global warming there will be an associated increase in the upwelling of carbon-rich water from the deep ocean. As this upwelled carbon reaches the surface it will increase the CO_2 content of surface water, reducing the pCO_2 gradient between the atmosphere and the ocean, and hence the uptake of CO_2 by the ocean from the atmosphere (Zickfeld, et al., 2007). At the same time atmospheric CO_2 concentrations will continue to rise in response to human activities. Eventually, as the atmospheric concentration of CO_2 exceeds the maximum deep-water pCO_2 value, estimated to be 430 microatmospheres (uatm) (McNeil et al., 2007), the Southern Ocean CO_2 sink will then change from a saturated CO_2 sink (Le Quéré et al., 2007) to a strengthening CO_2 sink. Model simulations suggest that this change will take place in the period 2020-2030 (under the IPCC A2 emission scenario) (Matear and Lenton, 2008; Zickfeld et al., 2008).

5.8.2.2 Response to ocean warming

As the Southern Ocean becomes warmer, its ability to store CO_2 through solubility changes is affected. Freshening associated with the warming will lead to a stratification of the upper ocean that will affect ocean carbon uptake through biogeochemical and physical changes. Warming will reduce the solubility of CO_2 in seawater, so reducing the ocean's ability to take up and store CO_2 (Figure 5.22a). A warming of 1% decreases oceanic pCO_2 by 4.23% (Takahashi et al., 1993). The result of these solubility changes will be to reduce the pCO_2 gradient between the atmosphere and the ocean, reducing the efficiency of ocean CO_2 uptake. Various studies of future CO_2 uptake suggest that the solubility effect will be significant (Matear and Hirst, 1999; Plattner et al., 2001; Sarmiento et al., 1998), although its magnitude remains poor quantified.

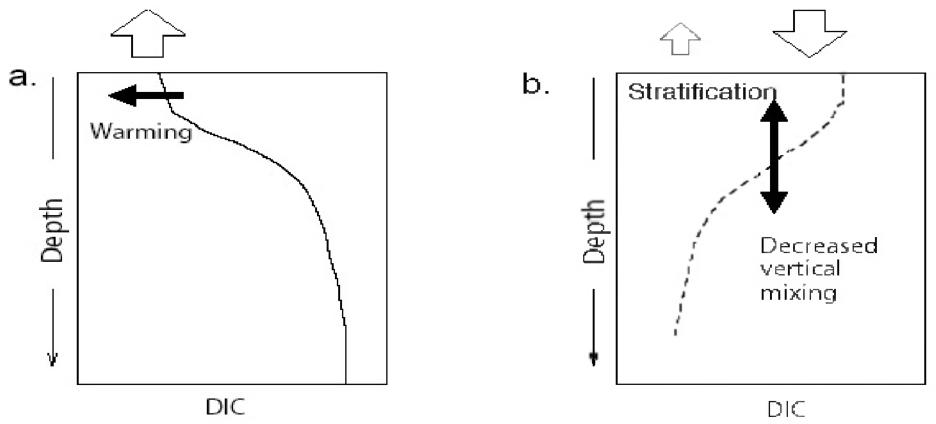


Figure 5.22 Schematic of how global warming is expected to impact on dissolved inorganic carbon (DIC) in the ocean and hence air-sea carbon fluxes of CO_2: (a) Sea surface warming decreases CO_2 solubility (solid horizontal arrow) and drives outgassing (Large open arrow); (b) Increased stratification impacts CO_2 through biological production; there is both enhanced biological production in response to warming and light supply, and reduction associated with reduced nutrient supply. The net effect is predicted to be an increase in CO_2 uptake (large downward open arrow).

The Southern Ocean is a high nutrient low chlorophyll (HNLC) region rich in the macronutrients (nitrogen, phosphate and silicate) needed by phytoplankton to grow, but poor in phytoplankton, most likely due to a lack of micronutrients, in particular iron. It also suffers from the low levels of light at high latitude, which may inhibit productivity. As the upper ocean warms, freshens and stratifies, conditions will favour an increase in productivity. That in turn will use up CO_2 in surface waters, which will increase the ocean – atmosphere pCO_2

gradient, thereby encouraging more uptake of CO_2 by the ocean from the atmosphere (Figure 5.22b). Increasing stratification of the surface ocean will also tend to limit the supply of nutrients from below, hence limiting productivity and increasing the uptake of CO_2. The amount of nutrients available in the stratified surface waters, hence productivity and CO_2 uptake, will also depend on the efficiency with which organic matter is exported out of the mixed layer by sinking (Figure 5.22b).

Bopp et al. (2001) explored the relationship between the different competing processes in a coupled climate carbon model and showed that because the Southern Ocean was nutrient-limited, the largest effect was from stratification, which increased the interaction between light and nutrients and led to a longer and more efficient growing season with a 30% increase in marine productivity and export production. Consistent with this, Sarmiento et al. (2004b) found that six different coupled climate carbon models showed increased primary production in the Southern Ocean between now and 2060 (Figure 5.23). The magnitude of the response was highly correlated with the strength of stratification, which in turn was related to changes in sea ice extent forced by continued global warming.

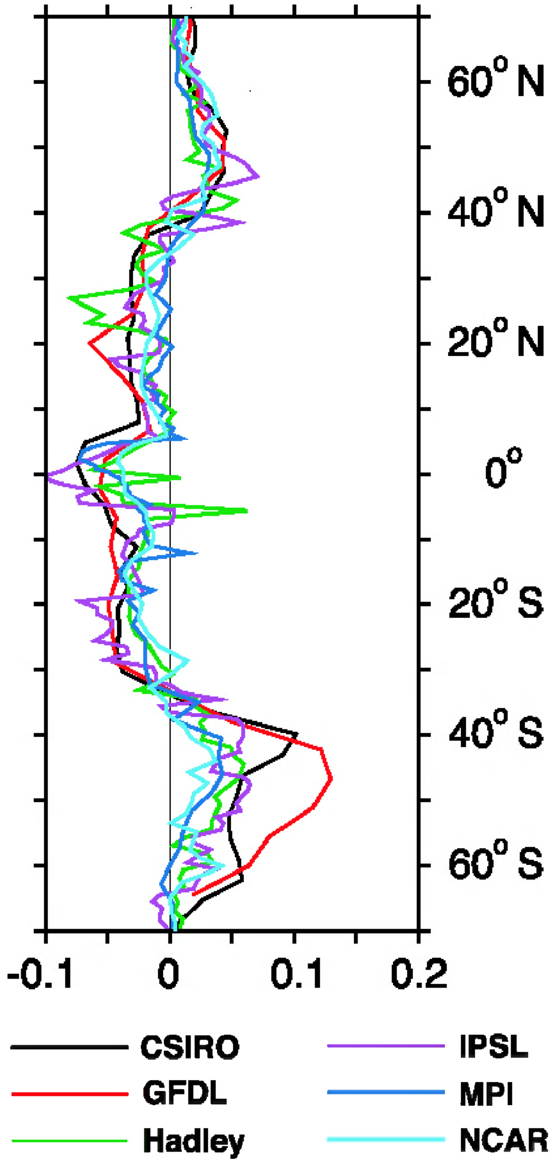


Figure 5.23 Primary Productivity (PP) changes PgC/degree calculated for the period averaged 2040-2060 using six different coupled climate carbon models (Sarmiento et al., 2004b). PP changes were calculated using Behrenfield and Falkowski (1997), and changes were assessed against control simulations that excluded global warming.

In addition to inducing biogeochemical changes, increased ocean stratification may also reduce uptake of CO_2 from the atmosphere through Southern Ocean density changes. Mignone et al. (2006) showed that the depth of the pycnocline is highly correlated with CO_2 uptake. As a result it is predicted that as stratification increases the pycnocline will shallow, so further reducing CO_2 uptake (Sarmiento et al., 1998).

5.8.3 Response to increased CO_2 uptake

The uptake of CO_2 by the Southern Ocean alters ocean carbonate chemistry, causing both a reduction in ocean pH (i.e. acidification) and a reduction in the ocean's ability to take up and store CO_2 through an increase in the Revelle Factor. The Revelle factor (Revelle and Suess, 1957), or buffer capacity, describes by how much the concentration of CO_2 in the ocean will change for a given increase in the partial pressure of CO_2 (pCO_2). The higher the Revelle (or buffer) Factor, the less able the ocean is to take up atmospheric CO_2. As shown in Figure 5.24, the Revelle factor in the Southern Ocean lies between 10 and 15. In the next 100 years it is predicted to increase to 17 or more, which will reduce the efficiency of the ocean to take up atmospheric CO_2.

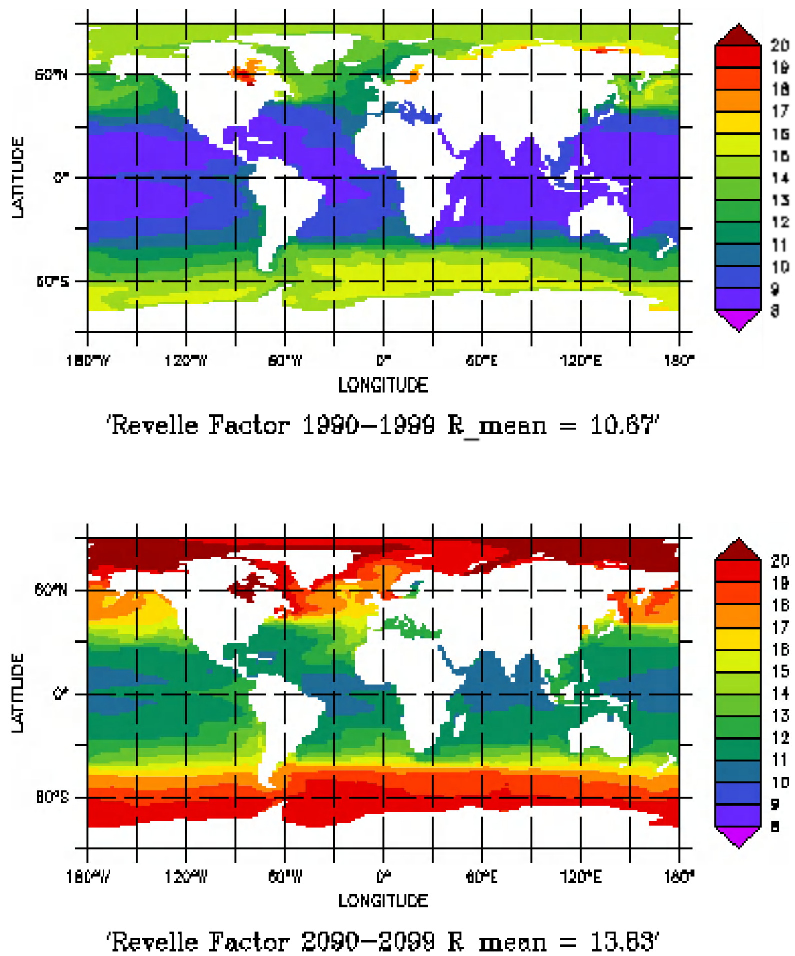


Figure 5.24 Surface values of the Revelle factor computed with the IPSL Coupled Climate Carbon Model forced by the A2 scenario and for the recent past and future (1990-1999 and 2090-2099) (Friedlingstein et al, 2006; Reprinted by permission from Macmillan Publishers Ltd: Nature doi:10.1038/nature04095, copyright (2005))

As the CO_2 levels in the ocean increase, the associated changes in ocean carbonate chemistry will lower the pH. Currently the upper ocean is supersaturated with respect to aragonite (used by important grazers like pteropods), and calcite (used by coccolithophores). As the ocean becomes more acidic (lower pH), the saturation states of both aragonite and calcite will be reduced until they drop below 1, when they pass through the saturation horizon - the point where the saturation state changes from super- to under-saturated. When the waters become under-saturated with respect to either aragonite or calcite it will no longer be possible for marine organisms to use these compounds to build calcium carbonate shells (Feely et al., 2004). Orr et al. (2005) used a suite of ocean models to show that by 2100 the saturation horizon will have shallowed significantly, the pH having dropped by an additional 0.3, and hence much of the Southern Ocean will be under-saturated with respect to aragonite (Figure 5.25). This acidification is expected to bring about a shift in marine ecosystems (Feely et al., 2004), and potentially a reduction in export production (Klaas and Archer, 2002), although it has been suggested that this effect may be offset by increased CO_2 uptake causing an increase in alkalinity (Heinze, 2004).

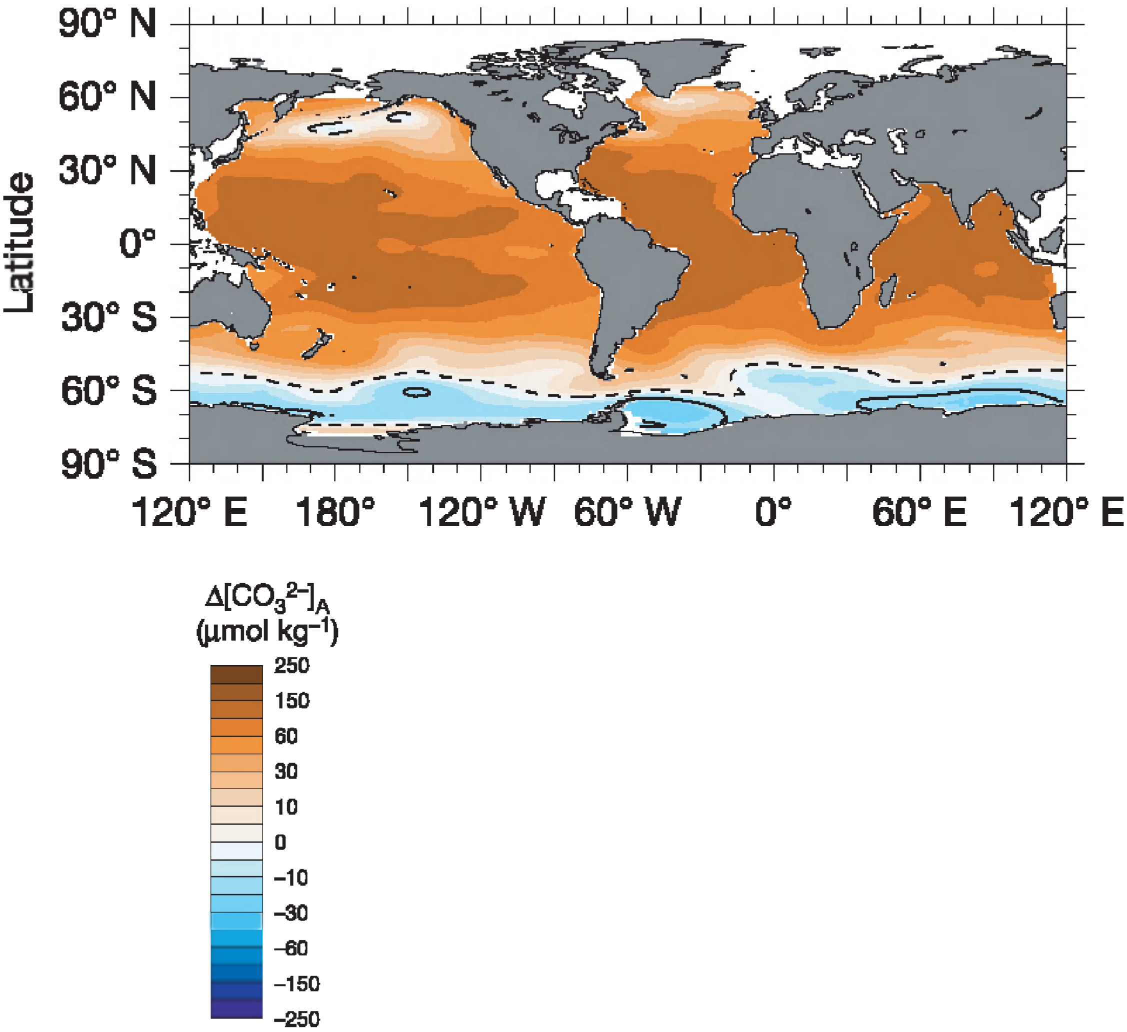


Figure 5.25 Predicted aragonite saturation state of the surface ocean in the year 2100 (Orr et al., 2005). The dashed line represents the saturation horizon and shows that much of the Southern Ocean (<50°S) is expected to be under-saturated with respect to aragonite by 2100.

While mean long-term changes in acidification and the Revelle Factor are significant, they are subject to large interannual variability (Matear and Lenton, 2008). This variability has the potential to perturb the system significantly, such that far reaching changes in the ocean ecosystem may occur well in advance of those expected by simply increasing CO_2 levels.

Studies of the global uptake of CO_2 from the ocean by the atmosphere suggest that over time the ocean will release sufficient CO_2 to amplify global warming (a positive feedback). As we have seen here the Southern Ocean is expected to evolve to act as a net sink for CO_2 over the next 100 years (a negative feedback). Nevertheless, the eventual behaviour of the Southern Ocean will depend not only on what happens around Antarctica, but also on what happens in the terrestrial biosphere. Changes in the uptake of CO_2 from the atmosphere by the terrestrial biosphere, particularly at mid-latitudes, could be large enough to change the pCO_2 gradient between the atmosphere and the ocean, thereby affecting the response of the Southern Ocean.

5.8.4 Concluding remarks

The magnitude of the response of the Southern Ocean to climate change remains uncertain. Simulations from coupled climate carbon models show a large range of responses (e.g. Friedlingstein et al., 2006), but do agree that the Southern Ocean will be an increased sink of atmospheric CO_2 in the future and that the recent reduction in CO_2 uptake will not continue. The magnitude of the total uptake is dependent on how the ocean responds to predicted increases in ocean warming and stratification, which can drive both increases in CO_2 uptake through biological and export changes, and decreases through solubility and density changes. The expected increased ventilation of carbon-rich deep water, combined with uptake of atmospheric CO_2, will increase the carbon content of the upper ocean, reducing the Southern Ocean's ability to take up more CO_2 in the future (through the Revelle or buffer Factor) and enhancing ocean acidification. This acidification is potentially worrisome because of its potential to impact the entire marine ecosystem. The various projections of future response are based primarily on coupled climate-carbon simulations, but such predictions need to be validated. It is therefore extremely important that strategies be developed to observe and detect change in the ocean carbon system (e.g. Lenton et al., 2006, 2009) to complement ongoing and planned physical measurements e.g. by Argo floats.

5.9 Biology

All organisms, both terrestrial and marine, are exposed to environmental change. Many polar species, more particularly in the thermally stable marine environment, are vulnerable to change because they are specialised to cope with a narrow range of conditions; in addition, they must face human-induced stresses. Species that fail to adapt to climate change requirements exceeding the organism's internal flexibility limits, or fail to migrate, will become extinct. The high vulnerability of Antarctic benthos (Peck, 2005) suggests that long-term survival of these organisms under conditions of rapid climate change may be compromised.

Nevertheless, we do have to bear in mind that many Antarctic species may have survived the massive climate changes from glacial to interglacial and back in the past, though geological and ice core records indicate that these have been primarily cooler episodes, warm interglacials having been relatively short. We also need to be aware that extrapolating laboratory results to the wild may involve unreasonable assumptions about the ability of biota to respond to change taking place over longer timescales than can be addressed by current technology and short-term funding regimes.

In addition we have to bear in mind that although the impacts of climate change tend to be larger in the polar regions than elsewhere, it is not easy to predict precisely what the effect of a change in one parameter, like temperature, may be, given the immense complexity of ecological systems and biological responses. This is particularly the case given the current lack of linkage between the description and modelling of environmental variables at macro- and biologically relevant micro-climatic physical scales. Summaries of the worldwide changes observed over the Twentieth Century are supplied by Anisimov et al. (2001), Walther et al. (2002), and in the 2007 IPCC report (Anisimov et al., 2007).

5.9.1Terrestrial Biology

Antarctica is extremely isolated and unusually cold as a result of its polar location and ice sheet. As a consequence, climate change will impose a complex web of threats and interactions on the plants, animals and microbes living in the ice-free areas of Antarctica. Increased temperatures may promote growth and reproduction, but may also contribute to drought and associated effects. Furthermore, high amongst future scenarios is the likelihood of invasion by more competitive alien species, easily carried there by humans seeking a place of unspoilt wilderness or chasing scientific knowledge. Such invasions are already a reality on many of the sub-Antarctic islands, with consequential and sometimes drastic consequences for the structure and functioning of native biota and ecosystems (Table 4.2) (Frenot et al., 2005, 2008; Convey et al., 2006). These invasions carry a clear warning for the future of terrestrial ecosystems on the Antarctic continent where, although a small number of alien species has already become established, none have yet become invasive (Convey, 2008). While dispersal and range changes are also natural processes, sub-Antarctic data indicate that human assistance outweighs the natural frequency of such events by two or more orders of magnitude. Furthermore, regional environmental change in the sub- and maritime Antarctic is likely to act in synergy with anthropogenic transfer, lowering the current barriers to both transfer and establishment that have previously protected Antarctica. Antarctica contains some of the only places on Earth where natural biological phenomena can be studied in their pristine state, but human visitation risks breaking Antarctica's isolation, and threatens Antarctica's unique legacy.

The consequences even of direct environmental changes might not always be easy to ascertain. For example, many sub-Antarctic islands show increases in mean annual temperature (Bergstrom and Chown, 1999). To date, there has been no suggestion that, even at the microclimate level, the increases are likely to exceed the upper lethal limits of most arthropods. However, in some areas, such as Marion Island, it is not only mean temperature that is predicted to change in line with current trends. Rather, the frequency of freeze–thaw events and occurrence of minimum temperatures are also predicted to increase, because of a greater frequency of cloud-free skies and a lower frequency of snow (which is a thermal insulator) (Smith and Steenkamp, 1990; Smith, 2002). An increase in the frequency and intensity of freeze–thaw events could readily exceed the tolerance limits of many arthropods, as recent work both on Marion Island and other south temperate locations has shown (Sinclair, 2001; Sinclair and Chown, 2005; Slabber, 2005). Other biota, such as continental bryophytes and lichens, may also be pushed beyond their tolerance limits if freeze–thaw frequency increases, especially given the physiological effects of this stress, such as the loss of soluble carbohydrates (Tearle, 1987; Melick and Seppelt, 1992). Thus, one of the major consequences of climate change might paradoxically not be an increase in upper lethal temperature stress, but rather an increase in stress at the other end of the temperature spectrum. How organisms are likely to respond to this kind of challenge has not been well investigated, though it is clear that lower lethal temperatures show substantial capacity for both phenotypic plasticity and evolutionary change.

In ice–dominated continental and maritime Antarctica, changes to temperature are intimately linked to fluctuations in water availability. Changes to this latter variable will arguably have a greater effect on vegetation and faunal dynamics than that of temperature alone (Convey, 2006). Future regional patterns of water availability are unclear, but increasing aridity is likely on the continent in the long-term (Robinson et al., 2003). Plant species which show high tolerance of desiccation, such as the moss *Ceratodon purpureus*, or others such as *Bryum pseudotriquetrum*, which have a high degree of physiological flexibility with respect to tolerance of desiccation, are more likely to persist under increased aridity than the relatively desiccation-sensitive and physiologically inflexible *Grimmia antarctici* (Wasley et al., 2006). Changes to water availability that cause an increased frequency of desiccation events are likely to negatively impact more strongly those species requiring hydrated habitats (hydric species) than those adapted to surviving shorter or longer periods of water stress (mesic or xeric species) (Davey, 1997).

In many respects, Antarctic terrestrial organisms are often well-adapted to the stresses of a highly variable environment, possessing features that should permit them to handle predicted levels of change that are often small compared with the natural variability already experienced. Indeed, with reference to temperature increase, resident biota will often be able to take advantage of reduced environmental stress, which will allow longer active periods/seasons, faster growth, shorter life cycles and population increase. Impacts of increased water availability are expected to be similar, although in both instances exactly the reverse consequences can be experienced locally, either directly as a result of decreased water input, or of interactions between increased temperature and water leading to greater evaporation and desiccation stress. Impacts of increased UV-B exposure associated with the spring ozone hole, while subtle, are expected to be negative.

With increases in the temperature component of current climate change in many locations of the Antarctic, many terrestrial species may respond positively by faster metabolic rates, shorter life cycles and local expansion of populations. But subtle negative impacts can also be predicted (and are perhaps being observed) with regard to increased exposure to UV-B, as this requires greater allocation of resources within the organism to defense and mitigation strategies, reducing resources available for other life history components (Convey, 2006; Hennion et al., 2006; Robinson et al., 2005; Snell et al, 2009). Changes in water availability will also impact on both terrestrial and more stable limnetic environments. Local reduction in water availability in terrestrial habitats can lead to desiccation stress (Convey, 2006) and subsequent changes in ecosystem structure, as has been reported from Marion Island where there have been dramatic changes in mire communities associated with a substantial decrease in rainfall (Smith, 2002).

The selective pressures experienced by Antarctic terrestrial biota over evolutionary time have resulted in adaptations with emphases on stress tolerance, plasticity and variation in life histories (Convey, 1996a). These adaptations have been at the expense of reduced competitive ability, leaving Antarctic ecosystems vulnerable to the impact of colonization by competitors that may be at more advantage under changed climatic conditions (Bergstrom and Chown, 1999; Convey et al., 2006b). These competitors may be either naturally dispersed or have ‘hitch-hiked’ with humans. As evidenced by the rapid increase in numbers and impacts of non-native species on the sub-Antarctic islands, the frequency of transfer by human agency (anthropogenic introduction) appears to far outweigh that by natural dispersal, not least as it overcomes the ‘dispersal barrier’ presented by the geographical isolation and survival of environmental extremes required in transit (Frenot et al., 2005, 2008; Whinam et al., 2005; Convey et al., 2006a; Convey, 2008). Furthermore, the combination of increased human visitation across the entire Antarctic region, and the lowering of dispersal and establishment barriers implicit through climate warming, are expected to act synergistically and result in a greater frequency of both transfers and successful establishments.

Changes in temperature, precipitation and wind speed, even those judged as subtle by climate scientists, will probably have profound effects on limnetic ecosystems through the alteration of their surrounding catchment, and of the time, depth and extent of surface ice cover, water body volume and lake chemistry (with increased solute transport from the land in areas of increased melt) (Quesada et al., 2006; Lyons et al., 2006; Quayle et al., 2002, 2003). Indeed, Quayle et al. (2002, 2003) show that some Antarctic lake systems magnify the already strong signal of regional climatic warming centered on the maritime Antarctic. Predicted impacts of these changes will be varied. A common factor is the changing influence of reduced lake ice and snow cover, which exert strong controls on the abundance and diversity of the plankton and periphyton (Hodgson and Smol, 2008). With increased warming, more of the lake is made available and production increases. Once the central raft of ice melts completely, the plankton and benthos can flourish, and diversity at all levels of the ecosystem increases. In shallow lakes, lack of surface ice cover will also lead to increased wind–induced mixing. In some areas of East Antarctica, longer periods of open water have led to increased evaporation and, together with sublimation of winter ice cover, have resulted in rapid increases in lake salinity in the last few decades (Hodgson et al., 2006c).

Increased inputs of melt water into the upper stratified layer of deeper lakes may also increase stability, and this, associated with increased primary production, will lead to higher organic carbon flux. Such a change will have flow–on effects including potential anoxia, shifts in overall biogeochemical cycles and alterations in the biological structure and diversity of ecosystems (Lyons et al., 2006). The predictions of Lyons et al. (2006) also serve to illustrate a profound paradigm shift in Antarctic biology that has occurred in the last 20 years. Although they and Convey (2006) state that we are not yet in a situation where we can develop a quantitative predictive model (or even models) that completely qualifies the response of Antarctic ecosystems to climate change, many of the predictions currently made are based on a foundation of long-term studies and monitoring, such as those at the McMurdo Dry Valleys LTER, or British Antarctic Survey sites on Signy Island. The importance of such long-term programmes cannot be overstated, particularly in national and global research funding environments increasingly predicated on 'short-termism'.

5.9.2 Marine Biology

Just north of the Antarctic Polar Front, the surface water temperature rises abruptly by about 3°C. The Front acts as a barrier for *Gene flow* - the movement of genetic information among populations within a species - in both directions, causing Antarctic evolutionary processes to occur in some degree of isolation. The number of established terrestrial alien species is much lower (and marine alien species very much lower) south of the Front, although there is evidence that some species exchange occurs in both directions (see Barnes et al., 2006).

Currently, benthic fish and cephalopods may have major barriers to dispersal, e.g. deep-water areas between continental shelves. The continental shelf areas and seamounts are typically separated by very deep (>4000 m) water. Such depths probably constitute insuperable barriers for the adults even in the larger fish species, but pelagic larvae may or may not be able to cross, depending on such factors as length of larval life, distance to be crossed, current direction and the existence of possible oceanographic barriers. The importance of the latter in influencing larval dispersal is becoming increasingly apparent, and indeed in a study of pelagic dispersal in the Antarctic Allcock et al. (1997) conclude that a previously suspected oceanographic barrier between Shag Rocks and South Georgia prevents transport between the two areas for the larvae of the octopus *Pareledone turqueti*.

The major unknown in the population genetics of Antarctic pelagic organisms is the extent to which they cross between water masses. It is unlikely that planktonic animals like salps, ctenophores and krill can cover significant (horizontal) distances under their own power; thus, to a large extent, they only move with the current systems. However, they may have considerable control of their own buoyancy and consequently can regulate the depths at which they float. With this capability they could possibly select currents at different depths to carry them in some "preferred" direction. The main water mass of the Southern Ocean rotates in clockwise direction around the Antarctic continent, but at a speed that would take perhaps several years for a complete circumnavigation. Nevertheless, this is the strongest and largest ocean current system in the world with a mean velocity of around 10-20 cm/sec between frontal jets and double that in the jets. These considerations would suggest that planktonic species are likely to have a high degree of genetic mixing over an extended time scale. *Gene flow* exerts a major influence on the rate and pattern of evolution. Spatial patterns of gene flow will play a central role in defining population structure in many polar species. Additional data are needed to understand the extent to which oceanography, coupled with the biology of individual species, may affect the genetic isolation of organisms here.

There is oceanographic evidence of separation of various component parts of the Southern Ocean. For example, the Weddell-Scotia Confluence seems to separate water to the west of the Scotia Arc from that immediately to the east, whilst other bodies of water show rotational movements (gyres) of long duration (e.g. perhaps a year or more in the Weddell Sea), which may lead to planktonic species being retained, and thus genetically isolated, within these bodies of water. Recent work on krill demonstrated a significant genetic break between the South Georgia and Weddell populations, underscoring the potential importance of oceanographic barriers to gene flow. Additional data are needed to understand the extent to which the oceanography of the Southern Ocean, coupled with the biology of individual species, may affect the genetic isolation of planktonic organisms. Palaeontologists are beginning to piece together a long history of temperate, cool temperate and cold climate marine biotas that have evolved in a mid- to high-latitude setting. Palaeobiological data will be used to assess the age of Antarctic habitats and species, and palaeontological information and data on modern forms will be used to compare diversity gradients in the fossil record and the living biota in the southern and northern polar regions. For many decades there was thought to be a fairly simple relationship between diversity and latitude, basically declining from the tropics to the poles. On land this is broadly true, with some notable divergences and exceptions. In the sea though it seems much more complicated, and very different in the north and south polar regions. Along the coast of North America there is evidence for a decline in molluscan, bryozoan and some other faunas towards the Arctic. A different pattern has emerged in the Southern Hemisphere, where what little information we have suggests a lack of uniformity to large-scale patterns in species richness. In the deep sea (the dominant area of the Southern Ocean) there is no obvious spatial trend at all (e.g. Brandt et al., 2007). On the continental slope there is simply not enough information to discern any trends. On the continental shelf the pattern is very taxon-dependent. For example in decapod crustaceans and hermatypic corals there is a strong latitudinal gradient, but there is a similarly strong longitudinal gradient (e.g. Stehli and Wells, 1971). In other taxa, e.g. pycnogonids, polychaetes and bryozoans, the Antarctic shelf is richer than average by area (e.g. Barnes and Griffiths, 2008; Munilla and Soler-Membrives, 2008). Gastropod molluscs are richer in Antarctica than at any southern latitude in the Atlantic, but poorer than at any latitude in the Indian Ocean (Linse et al., 2006). The problem is made more complex by the fact that the Southern Ocean is very poorly studied in many places (e.g. for the 40 degrees of longitude (length of the Mediterranean) of the Amundsen Sea), and rates of species description are low. In the 1970s it was suggested that whole fauna marine richness would be highest in the tropics and decline to lowest levels around Antarctica, but the only area where this notion has been tested, the South Orkney Islands, is richer that any southern hemisphere Atlantic or East Pacific archipelago (Barnes et al., 2008). The one area where a latitudinal cline is evident on the shelf is in the shallows, due to ice scour.

As well as understanding biodiversity and distribution, we are now starting to understand many aspects of the ecology, physiology, trophic and population dynamics of polar species. It is from this knowledge and understanding that concern about the future impacts of climate change (principally ice loss, warming and acidification) has rapidly emerged.

5.9.2.1 The impact of global climate change in polar marine environments

Given the differences in topography, substrates, freshwater input and glaciation history, the Antarctic and Arctic Oceans and their organisms are likely to respond differently to climate change. Laboratory experimental work has suggested that some of the more common species in the shallows are highly stenothermal (Peck, 2005). Small temperature differences (just a degree or two) may have great impacts on the physiology of stenothermal organisms as well as on the extent of sea ice, hence on the life history and biology of many species (but see Barnes and Peck, 2008). Although projected climate change should alter the situation, the polar regions currently offer an important opportunity to study species biodiversity and ecosystem functioning in environments largely undisturbed by humans. This is mainly the case in the Antarctic, where national territorial claims are still not applied, and international initiatives and organisations, e.g. the Antarctic Treaty System and the Convention on the Conservation of Antarctic Marine Living Resources (CCAMLR), prevent, or at least limit, commercial activities (exploitation of natural resources, industry, fishery, etc) with their consequent anthropogenic impacts. The main direct influences on the Antarctic marine ecosystem are likely to come from global climate change in the mid- to long term. It has to be recognised from the outset that the ecosystem has been radically disturbed by the effects of historic whaling and sealing, and that while much of the fur seal population may have recovered since then, whale populations are still severely depleted compared with former times. After a short phase of overfishing demersal fish stocks (rock cod and icefish) around the Antarctic Peninsula and Scotia Arc, the populations collapsed, but legal and illegal fishing continues, especially for Patagonian and Antarctic Toothfish as well as for krill.

Climate change is already having significant impacts on global marine and terrestrial systems (Hughes, 2000; Walther et al., 2002), and will continue to influence biological diversity. Many species are susceptible to climate change, and those of the marine environment are particularly vulnerable, even though warming is more evident in the air than in the sea as is evident from IPCC reports. The polar regions are undergoing more rapid environmental changes than elsewhere, in many instances due to the combined effects of natural climate change and human activity. However, these changes are much more evident in the Arctic than in the Antarctic except west of the Antarctic Peninsula.

Present patterns of biodiversity and distribution are a consequence of processes working on both evolutionary and ecological timescales. Among the physical and chemical factors controlling distribution and biodiversity of the modern polar marine fauna, the most important are ice scour, topography, substrate, temperature, currents, ice cover, oxygen, light, UVB, wind, and nutrients. Besides being largely interconnected, some of these factors are not constant, and vary over a range of temporal scales from less than daily through seasonal to inter-annual. Variability is of fundamental importance to ecosystem dynamics. The system may be disrupted if the pattern of environmental variability is upset.

The most important anthropogenic changes currently affecting the Antarctic are accelerated global warming and increased UV-B levels resulting from the ozone hole that develops in spring. Illegal and unregulated fishing and the introduction of alien species constitute further threats, although they are more limited in geographic scope. Pollution associated with scientific activities and ships, and visitor pressure from the growing tourism industry have very localised effects on community structure and diversity. Many of these changes have complex and interacting effects. For example, an impact on the lowest or highest level in a food web can propagate through to affect other taxa indirectly. Thus UV-B impact on primary producers may affect consumers and higher levels in the food web, while the extraction of the great whales undoubtedly had an effect that has cascaded down through lower levels.

Around the western side of the Antarctic Peninsula, which is currently subject to one of the fastest rates of climate change anywhere on the planet (Cook et al., 2005), there has been a considerable reduction of annual mean sea ice extent (reviewed in Clarke et al., 2006). There are indications that populations of *Pleuragramma antarcticum*, a key fish species of the trophic web, and whose reproduction is closely associated to sea ice, declined locally, to be replaced by myctophids, a new food item for predators (M. Vacchi, pers. com.; W.R. Fraser, Regional loss of Antarctic Silverfish from the western Antarctic Peninsula food web, in preparation). This change is thought to have been caused by seasonal changes in sea ice dynamics compromising reproduction processes.

Temperature trends elsewhere in Antarctica show little change or, in some places, a cooling that may be accompanied by local impacts, as in the lakes and soils of the Dry Valleys. There is no evidence of a continent-wide "polar amplification" similar to that predicted in the Arctic (Dyurgerov and Meyer, 2000; Oechel et al., 2000; Romanovski et al., 2002; Lemke et al., 2007). On balance this overall lack of change would not be expected to result in significant biological change, even in the open Southern Ocean, which has warmed by some 0.2°C. Sea ice, which has a significant relation to the ecosystem, has increased in the Ross Sea, and decreased in the Bellingshausen and Amundsen Sea, with local effects on the ecosystem. Acidification is beginning to occur in the ocean, slightly changing the chemistry, which on the basis of laboratory experiments is expected to first affect organisms with aragonite skeletons, such as pteropods, and ultimately to reduce the uptake of carbon dioxide from the atmosphere. As yet there is no evidence for any large-scale change in the Antarctic ecosystem associated with this effect. For many species, uncertainty in climate predictions leads to uncertainty in projecting impacts; however, continued warming and winter sea ice decrease are likely to affect reproduction cycles and the growth of fish, krill and benthos, possibly leading to declines in some populations and changes in their distributions.

In areas experiencing warming, increases have been recorded in sponge species (with extreme natural variability in recruitment and exceptionally fast growth) and their predators, and decreases in krill, Adélie and Emperor penguins and Weddell seals (Ainley et al., 2005). The reduction in krill biomass and the increase in abundance of salps (gelatinous pelagic organisms) have been suggested to be linked to regional decreases in sea ice (Loeb et al., 1997) that may also underlie recent changes in the demography of krill predators, e.g. mammals and birds (Fraser and Hofmann, 2003). Examination of growth of some seabed suspension feeders over the last two decades has revealed recent increases in annual growth rates in one species but little change or decreases in other similar species (see Barnes et al., 2007). In all of these cases it is hard to state that any change is definitively due to climate change, but evidence is mounting. There are signs that warming air temperatures have had negative impacts on the local biota on some sub-Antarctic islands. Signy Island and some sites at the West Antarctic Peninsula have witnessed an explosion of the fur-seal numbers that may be related to decreased ice cover resulting in increasing areas available for resting and moulting, but which may also be related to population increases on South Georgia; the growing seal population has had deleterious impacts on the local terrestrial vegetation.

The large reductions in the extent of cover and thickness of sea ice on the western side of the Antarctic Peninsula are potentially devastating to some species. The warming of the sea surface has been accompanied by an increase in phytoplankton in cooler regions (which might actually be beneficial to commercial fish stocks) and a decrease in warmer regions (Richardson and Schoeman, 2004; Montes-Hugo, 2009). If the sea ice cover continues to decrease, as models suggest, such responses to changes will widen, impairing predation processes and affecting community composition and levels of primary and secondary producers. For instance, marine ice algae would disappear due to loss of habitat. That may cause a cascade through higher trophic levels in the food web, diminishing the zooplankton that feed on algae, the fish that feed on zooplankton, and the sea birds and mammals that feed on the fish.

The responsiveness of species elsewhere to recent and future climate change raises the possibility that human influence may cause a major extinction event for some vulnerable species; we must consider such a possibility, though with due care. Thomas C D et al. (2004) analysed the global extinction risk from climate warming and concluded that many of today's species could be driven to extinction by climate change over the next 50 years. Such analyses provide compelling arguments for the development of policies aimed at reducing the impact of warming due to human activity. Greater and more rapid warmings have occurred around Antarctica before, during interglacial periods. Extinctions and radiations of species occur continuously, and most current species have probably survived through the climate changes of one or more glacial cycles (we can't be certain how many because the fossil record of many areas around Antarctica is very poor). However, projected warming and rates exceed those of the last eight interglacial warm periods. Given complete disappearance of sea ice we would expect extinction of those species that currently depend on it for survival. Climate models suggest that in the Antarctic such a reduction is unlikely within the next 100 years, when instead a 33% reduction in sea ice cover is projected.

5.9.2.2 Cold-adapted organisms and climate change: pathways of research

The development of research on biological adaptations to polar environmental conditions is relatively recent. Adaptations of the polar ichthyofauna in response to environmental change are commanding attention, and the effects of climate change on biodiversity are increasingly considered. There is ample evidence that recent climate changes already caused physiological problems to a broad range of species, drive evolutionary responses (Thomas C D et al., 2001, 2004; Walther et al., 2002), and produce micro-evolutionary changes in some species (Rodriguez-Trelles and Rodriguez, 1998). But species do not live in isolation and it is necessary to evaluate their responses at community and ecosystem levels. Ecologists and physiologists are thus faced with the difficult challenge of predicting the effect of warming not only on individual species, but also on whole communities. For instance, ice-shelf collapse increases the number of icebergs, enhancing the impact by scouring on benthic biodiversity as well as on the food web by altering regional and local current patterns. Although changes in sea temperature are as yet small, increased warming may cause sub-lethal effects on physiological performance and potential disruption in ecological relationships (Clarke et al., 2006).

Most of the work at the molecular and ecological levels in cold-adapted habitats has concentrated on Antarctic fish species. Understanding the impact of past, current and predicted environmental change on biodiversity and the consequences for Antarctic-ecosystem adaptation and function is a primary goal. Examination of Antarctic ecosystems undergoing change provides a major contribution to the understanding of evolutionary processes relevant to life on Earth. Key questions include: "How well are Antarctic organisms able to cope with daily, seasonal and longer-term environmental changes?" and "Will climate change result in either relaxation of selection pressure on genomes, or tighter constraints and ultimately the extinction of species and populations?"

Many of the same questions are being asked for the Arctic, not least through pressure from commercial fishing. Geography, oceanography and biology of species inhabiting Arctic and Antarctic polar regions have often been compared (e.g. Dayton, Mordida and Bacon, 1994) to outline the differences between the two ecosystems. The northern polar region is characterised by extensive, shallow shelf seas surrounding a largely land-locked ocean, whereas the southern polar region comprises a dynamic and open ocean and a very deep continental shelf (Smetacek and Nicol, 2005). The sea north of the Arctic Circle is almost completely enclosed and influenced by large human populations in extensively colonised terrestrial areas, as well as by industrial activities. The exchange of seawater through the passage between Greenland and the Svalbard Islands was not possible until 27 Ma (Eastman, 1997). The Arctic region was in a high-latitude position by the early Tertiary, but the climate remained temperate with water temperatures of 10-15°C. During the Miocene, about 10-15 Ma ago, Arctic land masses reached their current positions and it is thought that only at that time temperatures dropped below freezing ("unipolar ice-sheet mode)"]. Some studies indicate that glaciation events in the Arctic began 10-6 Ma ago, whereas in Antarctica glaciation started much earlier (Eastman, 1977). However, recent evidence revises the timing of the earliest Arctic cooling events, strongly supporting the indication of a "bipolar symmetry" in climate cooling. This suggests simultaneous evolution of ice at the poles and therefore a bipolar transition from "greenhouse" to "icehouse", pointing out the importance of greenhouse-gas changes in driving global climate patterns. According to this revision, the earliest Arctic cooling events are dated much earlier, to approx. 45 Ma.

Although high latitudes and cold climates are common to both the Antarctic and the Arctic, in many respects the two regions are more dissimilar than similar. The modern polar faunas differ in age, levels of endemism, taxonomic composition, zoogeographic distinctiveness, and in the range of physiological tolerance to various environmental parameters. Because of the isolating barrier of the Polar Front, the climatic features of the Antarctic waters are more extreme and constant than those of the Arctic, where the range of temperature variation is wider, thus facilitating migration and redistribution of the fauna.

In summary, the Arctic is the connection between the more extreme Antarctic oceanic system and temperate and tropical systems. Comparison of the ecosystems is likely to provide evolutionary insights into the relationship between environment and evolutionary adaptation. We have a remarkable opportunity to develop comparative studies on evolutionary differences between cold-adapted species and on how organisms from the polar habitats are affected by (and respond to) climate change. Comparing southern and northern polar processes may shed light on evolutionary pressures and provide insight into gene selection.

A detailed assessment of the impacts of climate change in the Arctic has been published (ACIA, 2005). ACCE will hopefully provide a similar contribution for the Antarctic.

Although the impacts of climate change on polar environments are exceeding those envisaged for other regions, and will produce feedbacks with global consequences, they remain difficult to predict because of the complexity of biological responses (Anisimov et al., 2007). Climate change may affect every aspect of an organism's biology, from cellular physiology and biochemistry to food web and habitat. Organisms must alter their physiology and biochemistry to cope with changes in enzyme activity and DNA damage, by means of phenotypic responses (occurring within the lifetime by enzyme activation/inhibition and induction/repression of gene regulation), and genotypic responses (occurring over a much longer timescale through the selection of beneficial mutations). Understanding the adaptation-response mechanisms in species living in both polar habitats may help also to understand change at lower latitudes.

In addition to adaptation, other key research themes include studies of life cycles (tactics and strategies for responding to environment features), micro-evolutionary processes driven by anthropogenic impacts, interactions between changing abiotic conditions (e.g. temperature, UV-B) and biotic responses, modelling interactions between environmental change and organism responses (to facilitate predictions of change), and development of conservation policies in relation to improved understanding of the response of ecosystems to change.

The challenge for the next decade will be to incorporate the physiological/biochemical viewpoint into the field of evolutionary biology. Such integration may provide more detailed answers than we can provide here to the question of how Antarctic and Arctic biota may respond to global warming, and the extent to which they will be able to adjust to it. Another challenge will be to determine the ability of polar species to repair the effects of changes induced by a wide variety of natural and anthropogenic processes, in the general framework of species and ecosystem responses to change.

Intensified communication between scientists should lead to a *multidisciplinary approach* to studying the ecosystem. Analyses of adaptive evolution across the biological spectrum from molecules to species must integrate physiology, biochemistry/molecular biology, morphology, taxonomy, biogeography, ecology, and ethology. Studying the response of evolutionary processes to changes in selection pressures demands collaboration between biologists, physical scientists and modellers. Investigating changes in the physical environment that have driven evolution over geological time requires collaboration with palaeontologists, paleoclimatologists, geophysicists, glaciologists and oceanographers. Statistical and molecular genetic approaches are needed to monitor changes in biodiversity. The multidisciplinary approach will allow links to be established between tectonics, climate change, glacial processes and evolution. For example, palaeobiological data can be used to assess the age of Antarctic habitats and species; these results can then be combined with molecular estimates of divergence time and disturbances of the mechanisms of adaptation.

The main aim of such cooperation is to identify key ecological processes and their physiological underpinning by using advanced multivariable community analyses and models, as well as mechanistic studies and mechanism-based models. This approach demands an intensive coordination, not only of ongoing, but also of future research activities. Only sound estimation of large-scale biodiversity as a result of evolutionary processes, as well as a synoptic mapping of the most relevant ecological parameters, such as currents, sedimentation, bottom topography, ice cover, will allow verification of different kinds of direct or indirect anthropogenic or natural impacts on benthic and pelagic communities. Physiological studies specify the sensitivities of marine organisms to environmental factors and thereby support a mechanistic understanding of the driving forces behind the patterns observed. Improved communication will help answer questions such as: How will shore systems develop when they no longer experience physical disturbance by sea ice? How will offshore systems change if pelagic algae replace ice-algae - will the zooplankton or the benthos benefit? What are the consequences for apex-predators, e.g. the polar bear in the Arctic?

This multidisciplinary approach will allow a series of important targets to be tackled including:

1. Links between tectonics, climate evolution, glacial processes and biotic evolution. In particular, we plan to continue to refine our understanding of how the present biota evolved, and why current patterns of biological diversity are what they are. Palaeobiological data can be used to assess the age of Antarctic habitats and species. These results can then be combined with molecular estimates of divergence time to provide a powerful approach to understanding Antarctic biotic evolution (as was used very successfully to determine the history of the notothenioid radiation by the ESF-funded Network on the Biology of Antarctic Fishes). Palaeobiological data can also determine the nature and origin of latitudinal diversity gradients.

2. Links between the physical environment and gene flow. Models of oceanic and atmospheric circulation can be used to predict transport of propagules into (and out of) the Antarctic. Such models can also be used to elucidate advective processes in the Southern Ocean and their impact on gene flow and population dynamics.
3. Links with northern polar studies. Comparison of southern with northern polar processes can elucidate significant evolutionary pressures and provide insight into gene selection.

Organisms have a limited number of responses that enhance survival in changing environments. They can:

(i) *Use the margins of internal physiological flexibility and capacity to sustain new biological requirements.* Species inhabiting coastal seabed sites around Antarctica are thought to have poorer physiological capacities to deal with change than species elsewhere (Peck, 2005; Peck et al., 2007; Pörtner et al., 2007). Experimental data suggest that they die when temperatures are raised by 5–10°C above the annual average, at which point many species lose the ability to perform essential functions, e.g. swimming in scallops or burying in infaunal bivalve molluscs when temperatures are raised only 2–3°C (Peck et al., 2004). In short, the margin range appears narrow, thus the efficiency of this ability may be poor. However, the rate at which the temperatures are increased in laboratory experiments is vastly faster than what occurs in nature, and behavioural change may be quite different at much slower rates of environmental change.

(ii) *Adapt to the new conditions and alter the range of biological capacity.* This strategy depends on the magnitude and rate of change, and aquatic habitats change temperature at a far slower rate than terrestrial ones, possibly creating fewer adaptation problems for most marine species. The ability to adapt, or evolve new characters to changing conditions depends on many factors including mutation rate, number of gametes produced per reproductive event, number of reproductive events and generation time. Antarctic benthic species grow more slowly than those from lower latitudes (Barnes et al., 2007; Peck, 2002) and develop at rates often 5 –10 times slower than similar temperate latitude species (Peck, 2002; Peck et al., 2006). They also live to great age, and exhibit deferred maturity (Peck et al., 2006). Data on numbers of embryos produced per reproductive event are scarce. However, there is a cline of increasing size in eggs with latitude (Clarke, 1992). This means fewer eggs are produced per unit effort. Fertilisation kinetic studies also reveal that around two orders of magnitude more sperm are needed for successful fertilisation of eggs of Antarctic marine invertebrates than in temperate species (Powell et al., 2001). From this and the egg data it is clear that fewer embryos are produced per unit reproductive effort by polar species. Longer generation times and fewer embryos reduce the opportunities to produce novel mutations, and result in poorer capacities to adapt to change than in similar species at lower latitudes.

(iii) *Migrate to sites where conditions are favourable for survival.* This depends on ability to disperse and availability of suitable sites. Intrinsic capacities to colonise new sites and migrate away from deteriorating conditions depend on adult abilities to travel over large distances, or for reproductive stages to drift for extended periods. Antarctic benthic species with pelagic (swimming or within the water column) phases have extremely long development times compared to lower latitude species (Peck, 2002). This means their larvae spend extended periods in the water column. However, the balance of

species with pelagic phases, compared with purely protected development, appears to be significantly lower in some Antarctic groups, especially molluscs. These groups (without pelagic dispersal phases) clearly have lower dispersal capabilities and capacities to migrate. The geographic context is also important here, and whereas most continents have coastlines extending over a wide range of latitude, Antarctica is almost circular in outline, is isolated from other oceans by the circumpolar current, and its coastline covers few degrees of latitude. In a warming environment this geographical constraint could be construed as supplying few places to migrate to. In contrast, Antarctic species do show unusually wide bathymetric ranges. With a deep shelf and strong connectivity with the continental slope, lack of latitudinal scope for migration may be compensated for by bathymetric migration possibilities. On all three major criteria, Antarctic benthic species can appear less capable than species elsewhere of responding to change in ways that can enhance survival. However, evidence of the vulnerability of Antarctic fauna to climate change is not yet clear cut (for arguments see Barnes and Peck, 2008).

Although the absence of wide latitudinal gradients in the Antarctic coastal region minimises the advantage of along-coast migration for survival, it highlights the importance of the sub-Antarctic as a critical research area, largely populated by (eurythermal) fish having a broad temperature tolerance. These fish live in a more variable environment, where changes might be faster and larger than in the High Antarctic. Historically, the sub-Antarctic may have been a site of long-term acclimation, because some cold-adapted notothenioids also inhabit sub-Antarctic waters (e.g. South Georgia, Bouvet), where in shallow waters the temperature may reach +4°C. The same concept can also be applied to invertebrates and warm-blooded animals. Because knowledge of their physiological performance is limited it is not known whether microbes behave in a similar way.

This discussion of latitudinal gradients ignores the fact that both the Ross and Weddell Seas penetrate to high latitudes (close to 85°S). However, that only becomes significant for most organisms in the event that the ice shelves occupying those shelf seas melt, which is unlikely in the 100 year time scale considered here.

In 2004, to address many of the current questions about the evolution of the biota in the face of climate change, the Scientific Committee on Antarctic Research (SCAR) launched an 8 year international scientific research programme “Evolution and Biodiversity in the Antarctic: the Response of Life to Change” (EBA). EBA integrates research across a wide variety of fields, from functional genomics and molecular systematics to ecosystem science and modelling, and draws on and contributes information to a wide rage of related fields, such as climate modelling and tectonics. Its main goal is to provide a platform for interactions between disciplines and researchers to improve understanding of the role of biodiversity in the Earth System and its responses to change, by offering the Antarctic context, and establishing cross-links with the Arctic, thereby enhancing the knowledge needed to support attempts to achieve a sustainable future for all life. EBA will provide SCAR and the international scientific community with the best possible estimate of the consequences for the Antarctic of continued environmental change.

New information, including the choice of suitable target species, long-term data sets and the concerted efforts from international multidisciplinary programmes, will help EBA to identify the responses of vulnerable species and habitats to climate change. This preliminary step is required to establish efficient strategies aimed at neutralising threats to biodiversity: in particular, before they become hopelessly irreversible, those that are essentially driven by anthropogenic contributions. EBA was selected by ICSU/WMO (the International Council for Science and the World Meteorological Organisation) as a “Lead Project” for the International Polar Year (IPY 2007-2008). This timely programme enabled the scientific community to address the increasing concerns expressed by the Antarctic Treaty Parties about the responses of Antarctic environments to natural and anthropogenic disturbances, and their request for information regarding ways in which these responses can be distinguished and mitigated to ensure long-term conservation of Antarctic environments and their biodiversity.

In summary, areas being investigated and internationally coordinated include:

- Cryptic species: to what extent may we have underestimated the diversity of the Antarctic biota and characteristics of Antarctic species?
- Radiations: when did the key radiations of the Antarctic taxa take place?
- Impact of glaciation at sea (evolutionary links between continental shelf and slope or deep-sea species).
- Phylogeography: geographical and bathymetric structure and relationships in the polar biomes.
- Population structure and dynamics in the context of evolutionary biology.
- Dispersal: immigration and emigration of organisms, dispersal, and role of humans as vectors.
- Genetic structure of populations, and extent to which population structures reflect past evolutionary history.
- The extent to which spatially separated populations of polar organisms interact at some level and, consequently, must be considered as metapopulations.
- The role of advective/transport processes in the gene flow and population structure of marine polar organisms.

In the broader sense such studies will enable biologists:

- To understand the evolution and diversity of life;
- To determine how evolution and diversity have influenced the properties and dynamics of present ecosystems and the global ocean system;
- To make predictions on how organisms and communities are responding and will respond to current and future environmental change.

5.9.2.3 Near-shore marine disturbances over the next 100 years

The Southern Ocean has the highest wind speeds and wave heights and the most icebergs of any water body. Since the formation of a massive ice sheet over the southern polar region, the biodiversity on the shelf of the Southern Ocean has been exposed to significant disturbance, mainly from scouring by the keels of moving icebergs and ice shelves that have sculpted the seabed and disturbed the benthic biota. As a result, the shelf environments of the Antarctic are amongst the most disturbed places on the planet, especially considering the slow tempo of polar coldblooded animals (ectotherms) and their inability to recolonise rapidly (Gutt and Starmans, 2001). In contrast the Southern Ocean also includes some of the world's least disturbed seabeds, both in the abyss and beneath the marginal ice shelves year round, and seasonally on the shelf below winter sea ice. Both in the shallows (Smale et al., 2007) and on the shelf down to depths of a few hundred metres (Gutt and Piepenburg, 2003) the seabed resembles a patchwork quilt of recovery since the last iceberg scour events. Despite the shallower parts of the continental shelf being very disturbed, they have the highest regional biodiversity as a result of continual processes of recolonisation (Gutt and Piepenburg, 2003). Newly scoured areas are dominated by rapidly spreading and rapidly growing pioneers, whereas less affected areas are occupied by slower growing, more competitively dominant species. Polar benthic communities can be highly hierarchical. Where there is no mechanical disturbance the major space competitors can monopolise space virtually unchallenged, producing a monoculture, as found mainly on the shallower parts of the shelf. Future changes in disturbance are likely to be highly variable regionally, in line with the regional development of warming. For example, we should expect considerable and continuing change in disturbance in the Antarctic Peninsula region, the area where the greatest warming and ice retreat is taking place (Cook et al., 2005).

Models of how the Southern Ocean will respond physically to the drastic and unprecedented, recent, rapid rises in CO_2 and temperature have high associated levels of error. Nevertheless, as shown above, extreme temperatures are not expected; models suggest a 0.5° to 1.0°C rise in the temperature of Southern Ocean surface waters in summer over the next 100 years, possibly increasing to 1.5°C in the Amundsen Sea, while winter SSTs remain within 0.5°C of where they are now, except locally where they may rise by 1.0°C. The temperatures on the shelf at depths of 200 m are likely to be much the same as those at the surface.

As well as warming, other physical responses in the Southern Ocean have already been detected including shrinking, duration and extent of seasonal sea ice (Zwally et al., 2002a), glacial retreat (Cook et al., 2005), rising acidification, and desaturation of aragonite levels (Key et al., 2004). Sea ice is expected to decrease by 33% overall, but with considerable regional variation, declining most in the Weddell Sea in December through May, and in the Amundsen and Bellingshausen Seas and in the South Atlantic and south Indian Ocean sectors in June through November (Figure 5.17). The gradual disappearance of ice shelves will open up new areas of ocean and seabed as habitat and for primary production.

Changes in disturbance with regional warming can be classed as acute (rapid change on short ecological time scales) or chronic (change on time scales spanning hundreds of years). Major elements of disturbance envisaged to alter with continued regional warming include:

Acute

1. Rapid ice-loading increases and decreases, caused by a rise in episodic scour by icebergs
2. Coastal sedimentation, caused by a rise in discharge from glaciers due to deglaciation on land
3. Freshening events, caused by increased seasonal melting of glaciers, ice sheets and ice shelves
4. Thermal 'events', caused by natural oscillations like El Niño events,

Chronic

5. Ice shelf disintegration, exposing new habitat
6. Long term ice scour decreases, from a decreasing supply of icebergs
7. Warming, at the whole Southern Ocean scale
8. Benthic response to changes in pelagic systems (e.g. decreasing the supply of food to the seabed)
9. Acidification
10. Deoxygenation

In the following subsections, the way in which each of these disturbance events may develop is considered as a 'best guess'. Acute events are considered first because they most easily conform to normal definitions of disturbance, but on evolutionary time scales chronic events may have just as much or more impact.

Rapid ice-loading increases and decreases from episodic scour by icebergs

Currently we are in an interglacial period and so should expect that some of the observed glacier and ice shelf retreat may be a continued response to the warming following the glacial maximum. There is evidence that a number of today's ice shelves were not present during the last interglacial, when it was warmer than now. Nevertheless, recent glacier retreat and ice shelf collapse in some regions, such as the Antarctic Peninsula, is almost certainly occurring faster than would be expected during natural cycles, whilst other regions have experienced either no change or small increases in ice cover. Many retreating maritime glaciers and ice shelf fronts have not yet retreated past their grounding lines and therefore calving ice will, at least over short timescales, continue to generate floating icebergs.

Ice shelves hold glaciers in check, so the disintegration of ice shelves adds to the production of icebergs and associated scour in two ways – first as chunks of ice shelf, and second from disintegration of the accelerated glaciers that were formerly held in check. An increased 'population' of Southern Ocean icebergs would lead to more ice scouring, with severe implications for shelf benthos. Although Antarctic icebergs can have draughts of up to 600 m depth, exposing most of the Antarctic shelf system to scour, most have draughts of 250 m or less, which generally restricts scour to the upper continental shelf.

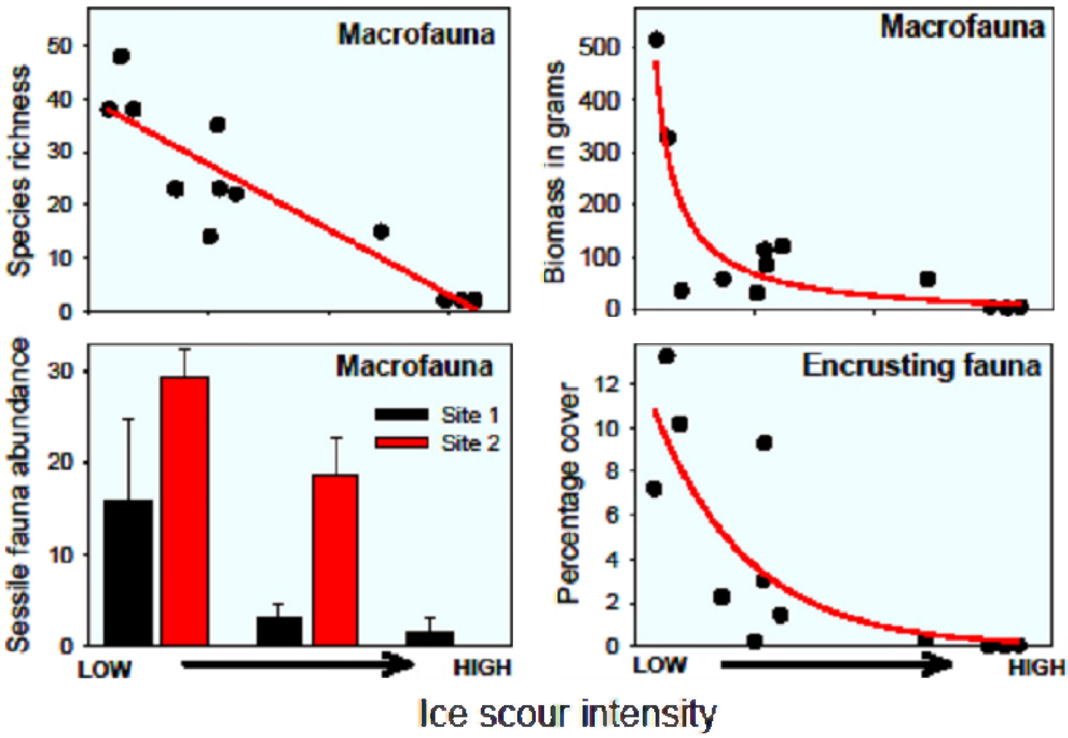


Figure 5.26 The influence of ice scour on fauna in the shallows at Adelaide Island, West Antarctic Peninsula. Patterns of faunal decrease with increasing ice scour frequencies in terms of macrofaunal richness (top left), biomass (top right) and abundance (bottom left) and cryptofauna (encrusting lithophiles) (bottom right). D Smale and D. Barnes, unpublished data.

Figure 5.27 When icebergs run aground they devastate any benthic life, as shown here from the Larsen B area east of the Antarctic Peninsula, and can produce irregularities in the intermediate sized to small-scale bottom topography. Width in foreground: approx. 1.3 m. Photograph: J. Gutt © AWI/MARUM, University of Bremen

Iceberg groundings are catastrophic and cause high mortality to benthic assemblages at scales of tens to hundreds of square metres, as in shallow waters around the South Orkneys, for example, where Peck et al. (1999) observed a 99.5% reduction in macrofaunal abundance following iceberg scour. Similar decreases have been found near the Antarctic Peninsula (Figures 5.26, 5.27). As a consequence, both in the shallows (Smale et al., 2007) and on the shelf down to depths of a few hundred metres (Gutt and Piepenburg, 2003) the seabed resembles a patchwork quilt of recovery since the last iceberg scour events. Despite the shallower parts of the continental shelf being very disturbed, they have the highest regional biodiversity as a result of continual processes of recolonisation. Newly scoured areas are dominated by rapidly spreading and rapidly growing pioneers, whereas less affected areas are occupied by slower growing, more dominant species. At greater depths it seems that iceberg scouring promotes and maintains biodiversity by increasing habitat heterogeneity and preventing monopolisation by dominant competitors (Gutt and Starmans, 2001). Polar benthic communities can be highly hierarchical. Where there is no mechanical disturbance the major space competitors can monopolise space virtually unchallenged, producing a monoculture, as found mainly on the shallower parts of the shelf. Because iceberg scour damages benthos, it promotes feeding opportunities for scavengers, which may be abundant at chronically disturbed locations. An enhancement to ice scouring in coastal waters will lead to benthic communities becoming dominated by pioneers (r strategists) and scavengers, whilst the species richness and the abundance of large and longer lived species will decline.

Increased iceberg scouring will also reduce the local densities and biodiversity of the meiofauna at an initial stage. Densities may recover quickly during recolonisation, but the community will remain impoverished for a long time after disturbance, especially at deeper shelf areas. The communities from shallow sub-tidal areas regularly affected by iceberg scouring seem better adapted to this process of physical disturbance (Lee et al., 2001a, b)

It seems likely that the numbers of icebergs will increase over the next hundred years or so, but the pattern of their seasonal and regional distribution may change with the projected reduction in sea ice cover. For example, we should expect considerable and continuing change in disturbance in the Antarctic Peninsula region, the area where the greatest warming and ice retreat is taking place (Cook et al., 2005). It is well known that the frequency of ice scouring in coastal waters is strongly linked to the persistence of seasonal fast ice (Smale et al., 2007). During the winter months icebergs are 'locked in' by fast ice (sea ice that is locked to the coast and lasts more than one year) that has formed around them, which restricts their movements and therefore their potential to cause disturbance. The extent and duration of fast ice has decreased in some regions during recent decades (Zwally et al., 2002a), and further reductions in sea ice duration would mean that icebergs will be more free to move around coastal waters for longer periods of the year. Thus, we foresee an increase in both ice scouring frequency and duration in coastal waters.

Coastal sedimentation from glaciers and due to deglaciation on land

Many maritime glaciers have dramatically retreated in recent decades along the Antarctic Peninsula, and the flow of ice into coastal waters is accelerating there and in the Bellingshausen Sea. The increased sedimentation associated with these glacial retreats is likely to have a considerable, but localised, effect on benthic communities adjacent to glacial termini. For example, a retreating Alaskan glacier may deposit up to 14 cm of sediment annually at its terminus (Cowan et al., 2006), whilst acute ice calving events considerably increase water column turbidity, and rapid glacier surges can lead to a 30-fold increase in seabed sedimentation for 4 km from the ice front (Gilbert et al., 2002). Directly beneath retreating ice fronts, where sedimentation rates are greatest, benthic fauna are completely smothered and the seabed is largely inhospitable. Studies on the effects of sedimentation on polar coastal benthos have been conducted almost exclusively at Spitsbergen in the Arctic, but the physical processes driving the observed patterns are likely to be consistent at both poles, despite considerable differences in the assemblages concerned.

In the most comprehensive study of its kind, Syvitski et al. (1989) sampled the benthos inhabiting ten Arctic fjords influenced by glaciers at differing stages of retreat. They proposed a general model of benthic community change during glacier retreat, which was principally driven by sedimentation rates. The seabed proximal to a retreating glacier is characterised by exceptionally high sedimentation and supports a pioneer assemblage of very few species (perhaps just one) of macrobenthic deposit feeders. This simple assemblage is likely to persist until the glacier front has retreated onto land, at which point sedimentation decreases to a moderate intensity and a more complex assemblage, still largely devoid of suspension feeders, can develop. The final stage in the faunal succession can occur once a glacier has retreated across land to expose an extensive valley floor, which filters sediment discharge and restricts the transport of sediment to the sea. The reduced sedimentation allows greater light penetration and causes minimal smothering, so that the seabed supports a diverse community, including suspension feeders and predators. This model has been supported by studies elsewhere, and Antarctic communities inhabiting the shelf adjacent to retreating glacier fronts might undergo similar change over ecological timescales. If, however, deglaciation continues on land and, consequently, increasing amounts of suspension are washed into the ocean by precipitation, as recently observed east of the Antarctic Peninsula (Figure 5.28), the benthos might even become poorer compared to a former ice shelf edge, and filter feeders will not recover not only at the former glacier termini but over large coastal areas. Instead only a reduced number of opportunistic deposit feeders or infauna would survive and perhaps benefit, and diversity would remain low.

Figure 5.28 Due to terrestrial deglaciation of the Antarctic Peninsula the amount of suspension washed into the ocean increases and affects planktonic and benthic biological systems. © 2007, S. Langner, AWI

Glacial retreat is likely to increase sediment loading in the shallow parts of steep sections of shelf, and could destabilise the substrate and promote slumping events. Such events could disturb assemblages over a considerable area, perhaps to depths of the order of hundreds of metres. Slumping of unstable sediments may smother considerable areas inhabited by benthos, and can result in high mortality of sessile species, reduced richness and significant community restructuring (see e.g. Gambi and Bussotti, 1999). Retreating ice fronts also deposit quite coarse material including boulders and 'drop stones'. Although small scale smothering by drop stones will damage sessile benthos, drop stones also represent an important source of hard substratum for colonisation.

Freshening events, from increased seasonal melting of glaciers, ice sheets and ice shelves

Compared with the Arctic (and lower latitude coastal environments) the influence of freshwater on marine benthos in Antarctica is minimal. Even so, during Antarctic summers melt water dilutes surface seawater (to depths of ~10 m), and is an important stressor on very shallow water benthos and intertidal biota (<10 m deep). The coastal waters of some Antarctic regions are likely to experience a considerable increase in freshwater input over the next hundred years from glacial retreat and melting ice sheets. IPCC model data (Figure 5.15(ii)) suggest that sea surface salinity around Antarctica will freshen by 0.1 to 0.2 units, with local values up to 0.3 in the Weddell Sea, in the Ross Sea, off Oates Land and in a few patches elsewhere along the coast; the winter pattern is similar to the summer pattern, but may freshen up to 0.3 units west of the Antarctic Peninsula. The model data show that these changes are restricted to surface waters. Stockton (1984) observed mass mortality in an epifaunal bivalve population following the summer formation of relatively fresh seawater at McMurdo Sound, but this is the only field report of its kind. Conversely, in the laboratory some (mostly) intertidal Antarctic algae exhibited rather broad salinity tolerances between 7 and 68 units for growth, photosynthesis and respiration (Wiencke et al., 2007). Both the intensity and frequency of biologically important freshening events are likely to increase in some regions in response to continued warming. Indeed, it is known that the salinity of the surface waters of the Ross Sea decreased during the late Twentieth Century, probably as a result of increased precipitation, reduced sea ice formation and continued melting of the West Antarctic ice sheet in response to global warming (Jacobs et al., 2002). Surface freshening on this scale can have a wide range of effects on both the water column and the seabed below it, including increased stratification and therefore reduced penetration of light and oxygen through the water column; these changes as well as the changes in salinity will have deleterious biological effects. Nevertheless it should be borne in mind that the modelled changes appear relatively small.

Short-term thermal events and natural oscillations such as El Niño events

Short periods of cold, such as 'ice winters' in temperate Europe, or unusually warm conditions in the tropics (associated with stronger El Niño Southern Oscillation (ENSO) events) have famously caused widespread mortality in the shallows. In the Antarctic, explosive growth of the exceptionally fast growing sponge *Homaxinella balfourensis* in McMurdo Sound has been related to the 1982/1983 El Niño event (Dayton, 1989). Even so, major thermal events and responses, are not common in the Southern Ocean. In winter that is because winter sea temperatures are already at their minimum, and because extensive vertical mixing induced by strong winds and waves limits any rise in sea temperatures in the shallows. The sole exception is close to active volcanoes, as at Deception Island. Most research into potential responses of biodiversity to climate change in Antarctica as elsewhere, have been dominated by considering the effects, in aquaria, of acute warming in very rapid rises (≥0.5°C per day) or rapid rises (e.g.1-3°C per week). We have already discussed these results, above. To date these experiments have largely been conducted on relatively few and arguably atypical species, the common and abundant types in the shallows, so how well these typify Antarctic biodiversity is unknown. There is also debate about the extent to which such short-term experiments and rapid temperature rises reflect true vulnerability to chronic regional warming, although certainly they do suggest the fauna is highly sensitive to acute warming. Current climate models suggest that acute warming is unlikely, as are average rises of more than 1.5°C (Figure 5.15).

Ice shelf disintegration exposing new habitat

Life below ice shelves is common in the Antarctic, since one third of the Antarctic continental shelf is covered by floating ice shelves. Due to their inaccessibility, the habitats under ice shelves belong to the least known on Earth. With an almost overnight ice shelf collapse, the change from an extremely oligotrophic ice shelf-covered ecosystem to a normal Antarctic shelf ecosystem, with high primary production during a short summer, is likely to be one of the largest ecosystem changes on the planet. First investigations in the Larsen B area following the collapse of that ice shelf showed that unique benthic species assemblages adapted to specific trophic conditions that might vanish (Figure 5.29) Krill (*Euphausia superba* and *E. crystallorophias*) and the pelagic fish *Pleuragramma antcticum,* appeared 5 to 12 years after the area became available, followed by seals (mainly Crabeaters), and Minke Whales (Gutt et al., 2008). In contrast, penguins, which inhabit rookeries about 60-100 km north of Larsen B had not yet entered the area. The sea-floor was dominated by very few fast growing species such as ascidians (Figure 5.30), sea-urchins with pelagic larvae, and young glass sponges.

Figure 5.29 Mega-epibenthic life under ice shelves can be extremely poor. Under the former Larsen B ice shelf there were locally only a few deposit-feeding holothurians. Width in foreground: approx. 1.3 m; 13 January 2007. Photograph: J. Gutt © AWI/MARUM, University of Bremen

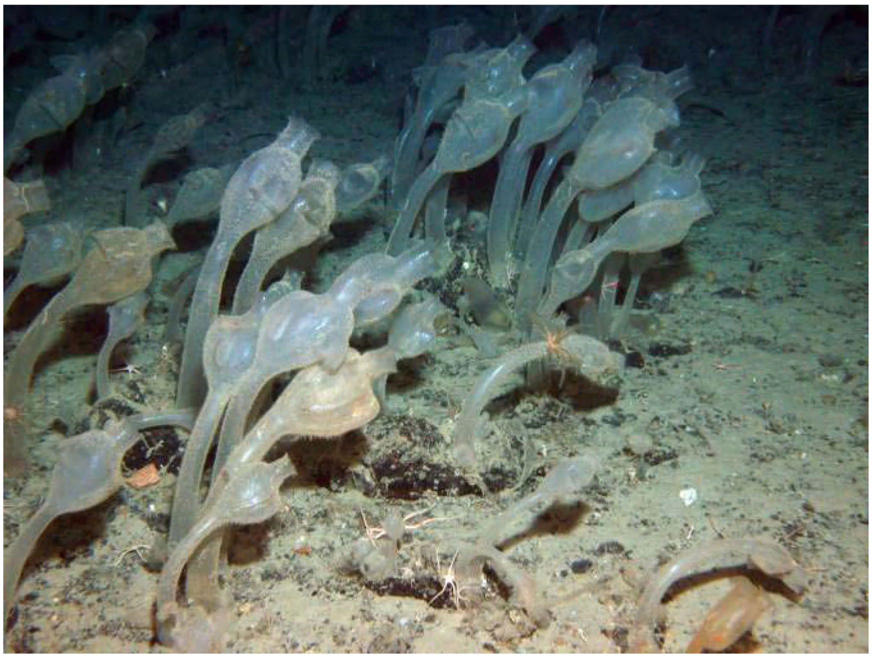


Figure 5.30 Massive population growth of ascidians may indicate a first step of a biodiversity-change after the disintegration of the Larsen B ice shelf. Width in foreground: approx. 1.0 m; 18 January 2007. Photograph: J. Gutt © AWI/MARUM, University of Bremen

Concerning the meiofauna, the area closest to the open Weddell Sea is characterized by much higher densities and percentages of nematodes than are found in the formerly ice shelf-covered bays. The outer stations might have close to a climax community, dominated by nematodes, whereas communities at the inner stations might have been sampled at an intermediate stage of succession, with lower densities but higher diversity.

Long-term ice scour decreases from a decreasing supply of icebergs

Whilst over short timescales (i.e. decades) it seems likely that ice loading and associated ice scour will increase in coastal waters, over longer timescales (i.e. centuries) it is likely that the frequency of ice scouring will diminish. Currently, about 55% of the seaward margin of the Antarctic Ice Sheet comprises floating ice, but this proportion will decrease as more ice shelves collapse. Once maritime glaciers and ice shelf fronts retreat onto land and past their grounding lines, calving ice will remain landlocked rather than being deposited into coastal waters where it can disturb marine benthos. Similarly, although the disintegration of an ice shelf can lead to rapid coastal ice loading by accelerating ice flows (Scambos et al., 2004), this can only be sustained for a finite period. So, whilst the major Antarctic ice streams will continue to transfer ice from land to sea over long time scales, the rate of ice deposition by small maritime glaciers and ice shelves will almost certainly reduce over time. The maximum depth at which ice scouring can occur is also likely to decrease, as the predicted retreat of the Antarctic ice sheet will result in thinner ice shelves and thus thinner calving icebergs. The current maximum draught of large tabular icebergs is rarely more than 600 m (Dowdeswell and Bamber, 2007). Relict scour marks, from when polar ice sheets were more extensive, have been detected in much deeper waters. If icebergs in the future have smaller draughts, the marine benthos inhabiting the deeper waters of the shelf will become unaffected by ice disturbance, so may become less diverse. Johst et al. (2006), using the Weddell Sea shelf with an average disturbance regime as a model system, found that a decrease in disturbance frequency was likely to be more detrimental to regional biodiversity than is a moderate increase (Figure 5.31).

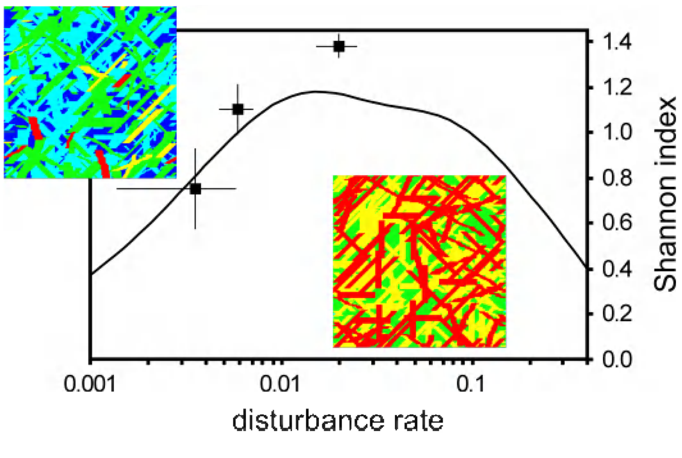


Figure 5.31 Benthic diversity (Shannon Index) on the Antarctic shelf predicted for a broad range of disturbance rates, using an analytical model based on different stages of recolonization (black line). Three values from empiric studies in the eastern Weddell Sea (small black quadrats +/-SD) and two scenarios from a similar spatially explicit model (large coloured quadrats) are superimposed; former ice scours (coloured lines) range from early (red/yellow) to intermediate (green) and climax stages (blue). The scenario in the upper left corner has a higher stage-diversity (more colours) than the lower, which represents a

completely disturbed community with only early and intermediate stages of recolonization, but no climax. For details see Johst et al. (2006) and Potthoff et al. (2006).

It is likely that disturbance regimes (and therefore community structure) in the shallowest waters will experience the greatest change. The shallow subtidal and intertidal zones around Antarctica are currently amongst the most frequently disturbed habitats on Earth. During summer, small icebergs and rafts of sea ice constantly scrape the shallow sea floor. In contrast, during winter, the icefoot - a narrow fringe of ice attached to the coastline - envelopes the seabed and associated fauna. As a result of intense disturbance, these habitats generally (but not always) support very simple assemblages, often dominated by opportunistic pioneers or mobile species that can retreat to deeper waters during winter. It seems likely that scouring by sea ice rafts and disturbance by the winter ice foot in shallow waters will eventually decrease with continued climate warming, as sea ice and ice foot formation and longevity will diminish under warmer conditions. The responses of shallow water and intertidal benthic communities to decreasing disturbance frequencies are difficult to predict, but it is evident that current disturbance holds community development at early successional stages in these habitats.

Warming at the whole Southern Ocean scale

Across much of the planet sea surface temperatures are warming but increases are much greater in some areas than others (Hansen et al., 2006). Although warming is difficult to detect in the surface layers of the Southern Ocean, due to a significant seasonal cycle and only few data, a warming signal of 2.3°C in August over the past 81 years, with about half the intensity in summer, is clearly visible in the upper 150 m around South Georgia (Whitehouse et al., 2008). In addition the surface of the Bellingshausen Sea has warmed rapidly by 1ºC over the last 50 years, but this signal rapidly diminishes away from the top hundred metres of the water column (Meredith and King, 2005). In intermediate layers (300 – 1000 m) warming by up to 0.3°C over almost 50 years (Böning et al., 2008) is significant but clearly smaller. Model outputs suggest that more widespread warming is inevitable, but that the most likely maxima are 0.5 to 1.0ºC, only locally reaching 1.5ºC (see Figure 5.15).

Whether populations or species will survive future temperature rises may be dictated by their ability to do critical activities such as feeding. Much of the evidence for the possible response to warming is based on laboratory experiments, which can only be performed on ecologically short timescales. That begs the question "Does the level of sensitivity exhibited in these short term experiments represent true vulnerability to likely environmental change, which will occur at a much slower rate?" Furthermore, in the natural environment sea temperatures vary with time, especially seasonally, and are only at maximum levels for a month or so – rather than continuous as in experiments, which leads to further doubt as to the meaning of the experiments. There are also questions as to how important the 'critical activities' measured to date in experiments are to survival, and how representative the species tested to date are of the wide biodiversity present. For example, the species that are convenient to sample and use as models, those in the shallows, are often atypical of their taxa in being highly dispersive (Poulin et al., 2002). Many other Antarctic taxa have current distributions that include sites or depths encompassing a much greater temperature range than 'typical Antarctic conditions'. South Georgia, for example, has populations of many otherwise typical Antarctic species, despite experiencing maximum summer temperatures 3°C warmer than some localities on the western side of the Antarctic Peninsula (Barnes et al., 2006b). Although Antarctic species tend to be endemic, in many taxa up to 40% of the species have ranges stretching into temperate waters where seasonal minimum temperatures exceed maximum values in the Southern Ocean. Such evidence implies that there may be a conflict between physiological (i.e. experimental) and ecological evaluations of vulnerability; ecological context may be a key missing ingredient. To gain more meaningful estimates of true vulnerability will require, for instance, investigations of populations across latitudes and depths, and at sites with differing temperature regimes, along with assessments of community level responses to changing conditions. This explains why biological response to predicted temperature rises in the Southern Ocean over the next 100 years is unclear despite experimental evidence indicating high levels of sensitivity.

Benthic response to changes in pelagic systems

The macrobenthos depends on a broad variety of food sources: suspended and deposited phytodetritus, other organic particles such as faecal pellets and aggregates, animals, macroalgae, and carrion. For the major primary food source, phytodetritus, it is possible to predict both an increase in supply due to an extended period for algal blooms in open water following a reduction in sea ice cover, and a reduction (see Thrush et al., 2006), at least at the regional level, due to there being less sea ice to melt and, as a consequence, a less well developed halocline.

It seems likely that most shelf-inhabiting suspension feeders may be adapted to the limited food supply typical of recent past glacial periods (Gutt, 2000). A further climate-induced increase in food supply could cause some species to suffer from a clogging of their feeding apparatus. The demise of such suspension feeders would also have potential consequences for the rich associated fauna, since their microhabitat or food source will be lost. Similar negative effects must be expected for large-scale geo-engineering operations to sequester anthropogenic CO_2 by fertilizing the Southern Ocean with iron (Smith et al., 2008). In contrast to ecological theory, fertilized systems with increased food supply are usually characterised by an increasing dominance of a few species only, and extinction of the majority of species (for examples from the deep-sea see Billett et al., 2001; Schewe and Soltwedel, 2003). As a consequence, in the Antarctic the extremely species-rich deep-sea fauna (Brandt et al., 2007), roughly estimated as 100 times richer than that of the Antarctic shelf, would be endangered. Since all deep-sea basins in the Southern Hemisphere are connected by the Southern Ocean, such negative effects may become widespread.

Due to fertilisation of surface waters by the upwelling of warm Circumpolar Deep Water over the Antarctic continental slope, increased sedimentation rates of phytodetritus may be expected anywhere in the Southern Ocean, and may have a similarly deleterious effect on the diversity of the benthic biota, even on the shelf. While Antarctic surface waters are zoogeographically isolated by the Polar Front, the deep-sea fauna is connected to the Antarctic slope and shelf fauna. Consequently, what happens on the slope and shelf could also affect the deep-sea part of the Antarctica's unique fauna and flora.

Only benthic systems living under strongly limiting food conditions (see Post et al., 2007; Riddle et al., 2007) might become richer in diversity and biomass, when food conditions improve, e.g. in areas where ice shelves disintegrate. An assemblage well adapted to extremely poor food conditions might suffer a decline rather than increase in diversity if the supply of food increased.

If the food supply reduces, Antarctic suspension feeders are generally not expected to suffer since they are able to survive extended periods of time (months to years) without food. Only communities that experience dramatic changes in local current conditions, e.g. due to a change in the shape of the ice shelf edge (Seiler and Gutt, 2007) or communities with seriously limited food supplies beneath ice shelves will disappear at the regional level if food supplies deteriorate further.

A shift in the pelagic realm from a retention-system (only few sinking microorganisms reach the bottom) to a loss-system (much organic material sinks rapidly to the bottom, e.g. in the form of faecal pellets (see Peinert et al., 1989)), caused by possible geographic shifts of large krill populations reaching the shelves (Atkinson et al., 2004), could also affect the benthos. How would the benthos respond? Because little is known about the food preferences of key benthic species (Orejas et al., 2003), this question can only be answered in general terms. Suspension feeders that prefer small food particles will suffer from a food overload that would benefit opportunists and those preferring larger particles. Deposit feeders might benefit more from an increase in the amount of food, but less from a change in food quality. Dense concentrations of deep-sea holothurians adapted to generally low-food supplies are typical of the Antarctic shelf (Gutt and Piepenburg, 1991). There is no reason to suppose that their food supplies will dramatically increase with warming of the ocean. The supply of food from the surface may change if increased UV-B radiation alters the composition of the plankton or causes there to be a change in the rate at which planktonic remains settle through the water column. If the ozone hole closes over the next 50 years the effect of UV-B radiation will diminish. Many of the predators target specific food items. That may make them very sensitive to changes in food availability, unless they can adapt their preferences rapidly.

Meiobenthic population structure and density seem to be regulated by variability in food input more than by changes in temperature. Vanhove et al. (2000) investigated a shallow bay at Signy Island (South Orkneys) over an 18 month period and found significant changes in faunal structure. The virtual lack of a feeding break during winter when the bay was covered by ice suggested that food was never limiting. This observation, in combination with the substantial selectivity of the metazoan meiobenthos for specific components of the sedimenting organic matter, such as ice algae or flagellates (Moens et al., 2007), suggests that changing food input in terms of cyclicity, quality and quantity has a significant impact on the meiofauna communities in the shallow waters of the Antarctic.

Acidification

About half of the anthropogenically emitted CO_2 has been absorbed by the global ocean (Key et al., 2004), making the global surface ocean more acid by on average 0.1 pH units. This seems like a small amount, but it is equivalent to a 30% increase in hydrogen ions because the pH scale is logarithmic. In both ecological and evolutionary contexts this constitutes a major disturbance. Orr et al. (2005) demonstrate that the rate of acidification will likely increase over the next few centuries, and that acidification will penetrate progressively deeper into the ocean for centuries after fossil fuels have been used up; surface ocean pH levels may become more acid by 0.2 to 0.3 units by 2100. Thus the magnitude of acidification will increase as an agent of disturbance both in absolute terms and spatially. Because levels and rates of acidification are greatest at the surface, organisms in the shallows and on the continental shelf will be influenced first and more seriously – this is likely to be a major problem for animals that secrete $CaCO_3$ skeletons. Nevertheless it should be borne in mind that pre-industrial ocean pH was around 8.2, which is slightly alkaline compared with that of neutral water (pH 7). So the current changes to pH levels of 8.1 and towards 8.0 or 7.9 could be said to be making the ocean less alkaline rather than actually acid (acids have pH<7).

As discussed earlier (section 5.8), under normal conditions calcite and aragonite are stable in surface waters because the carbonate ion is supersaturated there. As pH falls, so does the concentration of the carbonate ion. If it falls to the point that carbonate becomes under-saturated, then structures made of calcite or aragonite may begin to dissolve. Because the Southern Ocean has low saturation levels of $CaCO_3$, and because it is the location of much of the uptake of carbon dioxide from the atmosphere, it is at more risk than other areas of approaching under-saturation, particularly for animals that use the aragonite form of $CaCO_3$. Model projections of changing oceanic pH show that Southern Ocean saturation levels could become critical for organisms using aragonite within the next 100 years (Orr et al., 2005). Increased difficulty in synthesising skeletal material could pose major problems for the survival of many Southern Ocean species, particularly the aragonitic pteropods that are an important part of the plankton at the base of the food chain. Despite these fears there is evidence from the North Atlantic of increases in calcification of coccolithophores (calcitic plankton) over the past 200 years at the same time that the ocean has become less alkaline (Iglesias-Rodriguez et al., 2008). Bearing this conflicting evidence in mind, the jury remains out for the moment as far as the current effects of ocean acidification are concerned, though there is a worry about what the effect of the further predicted pH changes may be.

Deoxygenation

Antarctic Bottom Water (AABW) is the principal carrier of oxygen to the global deep sea environment. Very cold, dense and O_2 rich surface water sinks at various locations around Antarctica, especially the Weddell Sea, and flows into the deep Southern, Atlantic, Indian and Pacific Oceans. If the surface water is warmed it will become less dense, reducing the flow of surface water to the deep sea. A slower overall flow implies a decline in the flow of oxygen to the deep sea, which will be exacerbated because warm water dissolves less oxygen than cold water. A substantial reduction in the flow of oxygen to the deep sea could have a considerable effect on the global marine biota, given that 50% of the ocean is deep sea floor, which forms the planet's largest habitat. The biota of the deep sea are so poorly known that it is difficult to gauge the significance of deoxygenation in terms of loss of biomass, abundance, or species richness. That the deep sea biota of the Southern Ocean is very rich and contains species not found elsewhere is evident from the huge number of species new to science caught by recent cruises in the Weddell Sea (Brandt et al., 2007).

To date, the levels of warming detected in the Southern Ocean are too low to be significant in terms of oxygenation. Even if the Southern Ocean does warm significantly over the next 100 years it seems unlikely that the warming per se will be enough to cause significant deoxygenation. Instead, the warming of upper layers could stratify the water column more strongly and so reduce the amounts and rates of sinking, thereby reducing the transport of O_2 to the seabed.

5.9.2.4 Prospects for marine invasions by non-indigenous species

Establishment of non-indigenous species (NIS) is widely considered one of the greatest threats to biodiversity and endemic species. The threat these invaders pose once they have arrived in a new location, established themselves, and begun successfully to reproduce, is to outcompete native species for food or space, eat them or even hybridise with some. This has happened all over the planet, on all continents and most islands (even around Antarctica), across land, freshwater and marine habitats. The result is that over long time periods species have increased, decreased or otherwise altered the geography of their distributions. In the last few centuries, humans have radically altered organismal transport vectors, frequencies, journey times and survival prospects. Of the thousands of species travelling, probably only a small fraction survive and, of those, again only few establish themselves and yet fewer become 'invasive pests'. Even in these few aggressive invasive NIS, there is often a long lag phase between arrival and becoming a pest. Once established in terrestrial habitats, NIS have proved very difficult, often practically impossible to remove – except in the case of large mammals from small islands. In the sea, though, there is not a single case of an invading NIS being successfully removed. As a result, the global fauna is becoming more homogeneous as these few winners spread from port to port and the many losers will be native species with restricted distributions (endemics).

The potential for NIS arrivals to affect Southern Ocean biodiversity in the coming century is considerable because of: 1) historic Southern Ocean isolation and its domination by endemic species 2) lack of established NIS and the 'pristine' nature of the environment; 3) the slow response time of native organisms because of their extended generation turnover times; 4) the lack of durophagus predators (consuming prey with hard shells and bones); and 5) accelerating transport opportunities for NIS in a region of intense warming. These points are considered in turn below.

Firstly the Southern Ocean is the largest marine environment to be 'semi-isolated' for long periods of time - by surrounding deep water, the ACC and the Polar Frontal Zone (PFZ) (see Clarke et al., 2005). As a result most marine species and many genera are endemic to one or more regions within the Southern Ocean (Arntz et al., 1997). Their loss regionally thus would also mean a global loss. Experience of NIS arrival in regions with high endemism is that invasion tends to lead to drastic reduction and extinction of many endemic species, at least in terrestrial and fresh water environments. No NIS fauna have yet been found to be established in the Southern Ocean, although several species of assumed NIS algae now grow inside the caldera of Deception Island (Clayton et al., 1997; C. Wienke pers. com.). Recently the North Atlantic spider crab (*Hyas araneus*) has been recorded from the Antarctic Peninsula (Tavares and Melo, 2004), as have larvae of related subantarctic species (Thatje and Fuentes, 2003), but to date none are believed to be established. However, species have been found travelling in or close to the Southern Ocean in ballast water (Lewis et al., 2003) or fouling ship hulls (Lewis et al., 2006) or marine debris (Barnes and Fraser, 2003). Most recently, Lee and Chown (2007) found that mussels (*Mytilus galloprovincialis*) had survived a journey to and from the Southern Ocean on the ship Agulhas. This is an extremely aggressive invader that can smother coastal life in the absence of predation and that has been found to successfully breed at just 1°C (Lewis, unpublished data).

Macro-organisms native to the Southern Ocean tend to be characterised by slow development, growth and generational turn over (Arntz et al., 1994). The few species for which we have ample age spectra grow very old (over many decades) and many do not reach the age of first breeding for many years. This means that their ability to respond ecologically to competition for space or food is poor. Their long generation time also drastically slows rates of adaptation compared with potential invaders from temperate regions. Even the long lag times taken by many NIS take before they spread aggressively are well within the span of a single generation of some common Antarctic species. A closely linked (and fourth) point is the potential influence NIS durophagus (crushing) predators would have on Antarctic benthos. Southern Ocean shelf communities show many resemblances to community structure in Tertiary times, for example in lacking durophagus predators and having many shallow echinoderm suspension feeders (see Aronson et al., 2007). Invasion of even a few crushing predators could cause major changes to communities not adapted to such predation. Both the survival of such NIS and the ability of Antarctica's indigenous species to respond depend on the extent of regional warming.

The fifth point about the vulnerability of Antarctic marine biodiversity to NIS concerns increasing transport opportunities offered by rapid warming, both in the Scotia Arc and West Antarctic Peninsula region. This is the area most visited by tourist and scientific ships – which are amongst the most likely vectors for marine NIS (Lewis et al., 2003, 2006). It is also the area of most intense warming, and the only region within the Antarctic to show physiologically meaningful sea temperature increases to date (Meredith and King, 2005). Even the slight surface warming projected to occur in the next century, at least regionally, may decrease the ability of some native Southern Ocean fauna to function – e.g. to avoid predators (Peck et al., 2004). Conversely, rises in temperature might significantly raise the survival chances and competitiveness of temperate NIS. Crucially some crushing predators, like brachyuran crabs require slightly higher temperatures than those currently prevalent. Rising CO_2 levels in the ocean may also result in secretion of aragonite skeletons (shells) becoming more difficult, which would be more of a problem if NIS durophagus predators were to become established.

Potential invasions of NIS must be considered in the context of past species transport on various time scales. Species have moved into and out of the Southern Ocean both on evolutionary and ecological time scales (see Barnes et al., 2006). This will have occurred to some species even without them physically moving, simply because the Polar Front will have moved back and forth across their habitats between glacial and interglacial cycles. For example benthos on the shelf around the Kerguelen Islands will have been inside and outside the Polar Front multiple times (see Moore et al., 1999). It is likely that considerable movements of species occurred in response to glacial-interglacial cycles. It is therefore possible that there may be many species currently outside the Polar Front that will return, not as NIS, but as natives expelled during glacial maxima. The poor fossilisation conditions and destruction of potential fossils by advancing ice sheets during the onsets of glaciations makes it hard for us to know definitively which species are native. Another potential problem with recognising invading NIS is the patchy level of knowledge of Southern Ocean biodiversity. Arguably the most likely arrival areas of NIS do, however, coincide with the best-known regions (the shallow shelf, the Scotia arc, and the western Antarctic Peninsula). Despite the deep water and oceanographic barriers between Antarctic and temperate environments, there are many mechanisms for transport (Clarke et al., 2005; Barnes et al., 2006). Organisms may travel on floating debris, such as floating volcanic rock - pumice and driftwood. Some organisms may hitch-hike on megafauna (such as fur seals) and diseases can be introduced especially by highly mobile animals - such as avian influenza by albatrosses. Shipping has undoubtedly drastically increased opportunities for NIS, because of rapid travel from temperate ports (hotspots of invasive NIS) and across oceanographic barriers. Both Lewis et al. (2006) and Lee and Chown (2007) have found known invasive NIS associated with ships in the region – it seems that it is only a matter of time before the first established invading animal is found. How the native fauna will respond to such an invasion will depend on the chance nature of the invader identity, the area it arrives at (isolated island or Western Antarctic Peninsula) and the pace of climate change.

5.9.2.5 Marine picoplankton response to climate change

As a result of atmospheric circulation changes and sea surface temperatures rising in polar oceans, sea ice is expected to diminish, surface ocean stratification is expected to increase, and mixed layer depths are expected to reduce. The response of picoplankton (small eukaryotic protists and most bacteria) to such changes is largely unknown – but a few predictions can be made. Firstly, rising temperatures might enhance bacterial growth rates, cell numbers and secondary productivity. Increases in primary production, and the species and functional group composition of phytoplankton might strengthen microbial loop linkages (Falkowski et al., 1998; Falkowski and Oliver, 2007; Moline et al., 2000). Increases in dissolved organic matter production might fuel the growth of bacterioplankton, thereby creating a positive feedback to organic carbon cycling (Ducklow et al., in press). How these processes, and habitat changes (i.e. loss of sea ice) will impact microbial diversity and community structure is unknown. Studies are needed to consider different scenarios both experimentally and computationally to inform future models and enhance our understanding of Southern Ocean microbial ecology.

5.9.2.6 Biological response of birds and mammals

Biological responses to increasing climate-driven habitat and ecosystem fluctuation, and hence to climate change, can be articulated along four major axes: individual physiology and behaviour; species distribution; community structure; and ecosystem dynamics (Walther et al., 2002). Focussing on individual physiology and behaviour is important, because that is the level at which natural selection works, and other responses ultimately depend on physiology and behaviour.

The most observable changes in physiology and behaviour are the changes in phenology (Berteaux et al., 2004), which is the annual timing of life history events in populations (migration, arrival to and departure from breeding or feeding grounds, and reproductive events). These changes are thought to evolve by natural selection to match the environmental conditions and maximize the fitness of individuals (Futuyma, 1998). Given the extreme seasonality of high latitudes, phenology is a key aspect of the adaptation of Antarctic organisms and populations to change, and can be used to evaluate the match or mismatch in variability and trend between the rates of environmental change and of phenological response. A match between rates will be likely to occur with a stable optimum mean population fitness, and a mismatch to occur with decreasing fitness (Futuyma, 1998; Berteaux et al., 2004). Thus, measuring mean population fitness in long-term studies, which is more difficult than monitoring phenological changes, will be essential to identify and characterise responses of organisms to change.

Organisms will depend on their degree of phenotypic flexibility to cope with environmental change, or will adapt through changes in gene frequencies between generations, which at the population level is known as microevolution. Changes in gene frequencies are irreversible and will permanently modify their phenotypes. Though evolution is generally thought to be a slow process, microevolutionary changes can occur fast in response to climate change (e.g. Berteaux et al., 2004). Microevolution has been singled out as the main driver of Adélie penguin responses to extreme habitat changes caused by giant icebergs (Shepherd et al., 2005). This suggests that studies of fitness-related phenotypic traits collected over time and among related individuals of the same population may help detect evolutionary responses to climate change. Because microevolution occurs across generations, the typically long generation times of most predators are likely to affect their evolutionary responses (Rosenheim and Tabshnik, 1991). In predators, changes in optimum phenotypes of a trait from past adaptations will likely come with a fitness cost in survival or fecundity or in both. These short-term demographic costs may not be compensated fast enough by long-term adaptation, in which case populations are likely to decline (Stockwell et al., 2003).

Insights from long-term studies

The modification of the physical and biological environments around the Southern Ocean is accepted to be a direct or indirect consequence not only of changes in the mean climate, but also of the variance in climatic conditions at different spatial and temporal scales (Trathan et al., 2007). This variance has increased since the 1970s and may continue to increase with the frequency of extreme climatic events expected under many simulated scenarios (IPCC, 2007). The consequences for Antarctic birds and mammals are likely to be an increased variation in life history traits, which may have repercussions for fitness, population density and distribution.

Recent studies of phenological change in Antarctic seabirds with climate change indicate a multi-species delay in mean arrival and egg-laying dates in the Indian Ocean sector of the Southern Ocean (Barbraud and Weimerskirsch, 2006). Of the seabird guilds studied, the fitness of three populations - emperor penguin, snow petrel and southern fulmar - has also been evaluated (Jenouvrier et al., 2005a). These species show a stable population growth to date, and only southern fulmars show a significant phenological change, which is a delay in date of arrival to the nesting grounds. In southern fulmars, however, about 4% of the population growth rate may be caused by immigration, and without it the asymptotic growth rate (or mean fitness) becomes negative (Jenouvrier et al., 2003). This could indicate a mismatch between the rate of phenological change and of environmental change, leading also to an increased sensitivity of fitness to the environment. Indeed, this population, like those of the other two species fluctuates with the Southern Oscillation Index (and therefore with ENSO) and with increased sensitivity to sea ice concentration and high SST (Jenouvrier et al., 2005b). Because the propagation of ENSO effects to the southern Indian Ocean is most likely to have ecosystem-wide effects, as opposed to direct effects on the weather, increased ENSO variability is likely to bring more frequent food shortages. Extremely cold weather can sometimes affect the breeding success of emperor penguins or the nesting ability of snow petrels and have other adverse effects. However, these effects are less known to date. Snow petrels and emperor penguins are species that depend on the sea ice to complete their life cycles, and they show increased sensitivity to its loss. An important part of this sensitivity is likely to be the food shortage associated with increased sea ice extent and variability in sea ice duration (Jenouvrier et al., 2005a).

Most Antarctic marine predators are long-lived or relatively long-lived, and their average generation times are longer than the average interval between extreme climatic events. In most of them, rather than a direct impact of temperature or weather (precipitation, snowfall, or sea ice coverage) on individuals, the main impacts of climate are likely to be the changes in abundance, quality or stability of the food resources, which result from food web modifications related to changes in the physical environment. An increase in climate-related food shortages has caused population decline in the otherwise highly successful Antarctic fur seal at South Georgia (Forcada et al., submitted), which recovered from near extinction in the early Twentieth Century. The recent rapid increase in ENSO-related variability, with more frequent extreme events causing frequent food shortages, has led to a local decline in female fur seal fitness, and a loss of buffering of survival and propensity to breed in adult breeders due to the high rates of ecosystem fluctuation. These vital rates, most important to fitness, have increased in variability over the last 25 years (Forcada et al., submitted). When the temporal variation in these types of vital rates increases, the long-term fitness tends to decrease, causing loss of buffering against the environment (Morris and Doake, 2004).

In the fur seal case, changes in phenology were not apparent, which suggests that the loss of stability of the food supply and shortage in food, rather than other habitat constraints, are more likely to affect long-term fitness. An important difference between Antarctic fur seals and other species around Antarctica is their exposure to increasing ecosystem fluctuation derived from extreme climatic events, which are manifest near one of the Antarctic regions with most rapid warming, the Antarctic Peninsula. That makes it likely that the loss of buffering against environmental change occurs in species most sensitive to loss of their critical habitats, but also in those most affected by changes in stability and availability of food supply. In emperor penguins and other ice-dependent species, like Adélie penguins, the sea ice habitat is essential to complete the life cycle, and without it populations cannot persist. Sea ice-retreat is limited by the ice caps and other factors, and therefore there is likely to be little flexibility in phenological changes in these species. In that case, phenotypic plasticity may not be enough to ensure persistence, and it is likely that more successful species will replace those most sensitive to loss in critical habitats (e.g. Forcada et al., 2006). Microevolution may help long-term adaptation, but there is little hard evidence that it is occurring already.

Responses of baleen whales to climate change in the Southern Ocean

Predominately krill predators, baleen whales are a key component of the Antarctic ecosystem, but are currently at only fractions of their historical abundance due to exploitation to near extinction by whaling operations in the last century (Clapham and Baker, 2002). Contemporary data have shown that different species populations are recovering at different rates, both temporally and regionally, but long-term data sets on whale population dynamics are lacking (Leaper et al., 2008). Interpretation of the responses of baleen whale populations to climate change will therefore be especially difficult to disentangle from the effects of exploitation, and may not be detected for some time, as whales are such long-lived species. In the long term, given that some species are only in the early stages of recovery, such as the blue whale (Branch et. al., 2004), means that climate change impacts will almost certainly negatively affect the recovery potential of such species. In the short term, direct effects of temperature increases on baleen whales are unlikely because of their mobility and thermoregulatory ability. Instead, it is likely that climate change impacts will be mediated primarily through, (1) changes in sea ice dynamics that alter habitat characteristics and (2) changes in prey abundance and distribution (Nicol et al., 2008). In those areas of the Antarctic where sea ice is predicted to shrink (IPCC, 2007), pagophylic species such as blue and minke whales will be both directly and indirectly affected. As ocean productivity shifts with changes in sea ice extent, more northerly (oceanic) species, for example fin whales, may track changing prey availability and expand their range south, overlapping spatially with blue and minke whales. Prey availability may also have direct affects on whale demography, and this has already been demonstrated for southern right whales. Leaper et al. (2006) have shown that the breeding success of southern right whales feeding in South Georgia is driven by underlying relationships with the availability of krill, whose population fluctuations are correlated with changes in ocean climate, especially sea surface temperature (Trathan et al., 2006). Paradoxically the very fact that baleen whales depend upon the availability of suitable habitat in multiple locations (i.e. not just Antarctica) may also make them especially vulnerable to climate change, given that their low latitudes breeding areas may be subject to different impacts.

5.9.3 The Antarctic marine ecosystem in the year 2100

"The complexity of the Southern Ocean food web and the non-linear nature of many interactions mean that predictions based on short-term studies of a small number of species are likely to be misleading" (Clarke et al., 2007). The most critical points for biological predictions are (i) the enormous complexity of living communities, including hundreds to thousands of parameters (species and their life traits; for their general sensitivity see Barnes and Peck, 2008), (ii) our limited knowledge of these, and (iii) the lack of fine-scale detail in predictions of change in the physical environment, e.g. of the likelihood and extent of extreme events, and of fine scale spatial resolution. Nevertheless, physical predictions have improved considerably compared to a few years ago, because they now include some measurements of regional as well as global climate change.

Polar ecosystems are experiencing significant environmental changes. In the Antarctic the retreat of glaciers and the disintegration of ice shelves, with waters underneath being the least explored marine systems on Earth, provide the most obvious impacts on coastal marine areas. In offshore systems, a shift of pelagic communities towards the south is interpreted as a consequence of regional changes in sea ice dynamics, especially West of the Antarctic Peninsula, although the average sea ice extent has changed little. There are first signs of increased water temperature also along the coast. Because of an assumed high sensitivity of such biological systems to changes, inshore and offshore waters around the Antarctic

Peninsula and at the sea ice margin as well as along the Antarctic Convergence (Polar Front) are the main foci of ecological climate research.

The attempt made here to develop a scenario of how the Antarctic ecosystem might look in 2100 should be considered as a tool for identifying future research needs and recommendations to decision makers rather than as a true prediction. The most important constraint that militates against numerical forecasts is the reliability of a number of assumptions about the biological system and its environment. To make this "pre-stage" of a prediction as robust as possible, the concept is based only on coarse patterns. Air and ocean temperature are assumed to be the main climate-driven environmental parameters for the Southern Ocean ecosystem. In addition CO_2-triggered acidification and increased UV-radiation are likely to alter the ecosystem.

Substantially damage of the sea-floor ecosystem by grounding icebergs has been observed locally, following the climate-induced disintegration of the Larsen B ice shelf, and in very shallow water with a naturally high intensity of disturbance. By 2100 the Larsen C ice shelf and much smaller ice shelves west of the Antarctic Peninsula might also have collapsed, at least partly (Morris and Vaughan, 2003), and the northeast coast of the Antarctic Peninsula will have experienced a significantly elevated disturbance regime preventing the benthos from reaching a climax stage. After 2100, icebergs from the Filchner-Ronne ice shelf will continue to scour their way along the east coast of the Antarctic Peninsula, contributing to maintaining a patchwork of recolonization stages and, consequently, high benthic biodiversity. Iceberg calving events along the west coast of the Peninsula and around East Antarctica are also likely to increase as ice shelves collapse, so increasing benthic biodiversity, though not in so dramatic a manner as in the western Weddell Sea. Between now and 2100 elevated air and water temperature will not have risen to levels at which the large Filchner-Ronne and Ross Ice Shelves might disintegrate and cause large-scale circumpolar damage. A regional lessening or cessation of production of icebergs might cause local development of a benthic community in which competitive displacement leads to increased productivity but reduced biodiversity. An irreversible negative effect will be the loss of the unique marine biological system below the former ice shelves. These areas will serve as habitats of retreat for some vertebrate and invertebrate species that escape from warmed areas in the north and west of the Antarctic Peninsula, but these newly available areas will not completely compensate for habitat-loss elsewhere (J. Gutt, unpublished results).

The development of the sea ice plays a key role in the development of the pelagic as well as the open ocean ecosystem, and in intertidal communities; it has indirect effects on the sub-tidal benthos especially on the shelf but also in the deep-sea. Between now and 2100 all components of the ecosystem closely related to the sea ice will show a significant change in their ecological performance in response to the predicted 33% reduction of sea ice extent. A decline of krill by 38-80% in the Atlantic sector was already noticeable by 2004, having begun in the late 1970s (for most recent results and further references see Atkinson et al., 2008). Our 'forecast' is that this decline will not be compensated by increasing near-shore populations in the less affected "refuges" in East Antarctica between approximately 90°W and 105°W. In areas with an originally high krill population size the population is likely to stabilize at a low level. The implication is that over the long term all main krill consumers will experience a serious food limitation. The Minke Whale will lose 5-30% of its ice-associated habitat, while for Blue, Humpback, Fin and Sperm Whales a compressed foraging habitat along Southern Ocean fronts is suggested. This negative effect will be superimposed on the recently observed 9.6%/yr rate of increase in Humpback Whales and similar developments in other large species in recent decades, which runs counter to the observed decrease in krill. Due to the trophic complexity of this ecosystem it is impossible to foresee which animal group will suffer most, but some may suffer less than others while other may even become extinct, at least regionally. Colonies of Emperor and Adélie penguins, which are most closely adapted to a complex sea ice regime, but which are also affected by precipitation, will become extinct locally where their pack ice dominated habitat shifts to an open ocean system. They will be locally displaced by king, gentoo and chinstrap penguins. Long-term field data from an Emperor penguin colony in Terre Adélie (East Antarctica) showed very sensitive response to natural climate changes in both directions, when sea ice increased and decreased (Barbraud and Weimerskirch, 2001). A projection using predictions of warm events from IPCC climate scenarios and combined with data on the population dynamics of this East Antarctic colony forecast a 36% probability of a quasi-extinction by 2100, in an area with rather minor climate change impact (Jenouvrier et al., 2009). Since it will be difficult for this species to find sites for new colonies, the total net population will decrease significantly and the zoographic range will be compacted southward. Adélies will find new colonies where pack-ice becomes more divergent and at newly exposed coastlines when ice shelves disintegrate.

As in the case of penguins, the ice-bound Crabeater, Weddell, Leopard and Ross Seals will regionally become extinct due to changes in both, habitat and food web dynamics. The ice-tolerant Fur and Southern Elephant Seal will shift their geographical extent further south and regionally increase their population size – unless their food resources decline. The Antarctic toothfish also has an ice-related behaviour. A prediction of the effect of a 1.3°C temperature increase that could take place between the next 10-90 years shows a circumpolar decrease of population size. This species would only likely become extinct with an extreme warming coupled with a sea ice retreat of 2 km per year (Cheung et al., 2009). In essence most Antarctic species are adapted to changes in sea ice concentration and its extent. Species of any systematic group with a sufficient initial population size and circumpolar distribution are expected to survive at least in the Pacific sector south of Australia and New Zealand, where according to the predictions the sea ice is likely to remain relatively stable.

The effect of the future reductions of sea ice on primary production is likely to be more complex. Southernmost near-shore waters, which, under ‘normal’ conditions, have a low yearly primary production compared to areas like the Antarctic Peninsula, will become more productive as important components of the open ocean system move farther south (Whitehouse et al., 2008). This shift will coincide with a shift among unicellular algae from larger to smaller diatoms and generally from diatoms to *Phaeocystis* aggregations, both being less favourable food for krill. One side effect will be more emissions of precursors of dimethyl sufide (DMS) causing extra cloud cover. Using results from the “Climate System Model 1.4-carbon” and experiments on the ecology of phytoplankton, Boyde et al. (2007) assume that “the rate of secular climate change will not exceed background variability, on seasonal to interannual time-scales, for at least several decades…”. These results might underline the high relevance of the sea ice and its biota to the already observed and expected changes in the Antarctic ecosystem. At the sea floor, increased primary production may present a problem for filter feeders that are principally adapted to low food supply rather than decreased production. However, a few opportunists, e.g. among mobile deposit feeders, are likely to benefit and to extend their distribution range. This process will change the functioning of the benthic system, with significant consequences for cycles of organic and inorganic carbon as well as silicate remineralization. The increase in production will also act to reduce diversity in the deep-sea and on the continental slope, where a few species will benefit above the average, coming to dominate the entire assemblage. The extent of extinction of species between now and 2100 will depend very much on the number of at present largely undiscovered “cryptic” species, those that are visually not discernible from very close relatives, if they have only a limited range of occurrence. Populations of pelagic invertebrates and fish may collapse at the regional level, but due to their good dispersal capacity will survive in refuge areas. In contrast to the above scenario, in which phytoplankton growth increases in coastal waters, we may see in offshore waters especially close to the Antarctic Convergence (Polar Front) a spatially and temporarily hardly predictable decrease of primary productivity resulting from decreased stratification, vertical mixing, and increased microbial oxygen consumption. Such a development may negatively affect higher trophic levels of the food limited pelagic ecosystem.

Between now and 2100 water temperatures will not rise by much more than 1°C at the sea surface and down to 200 m depth in most areas. This is likely to have less of an effect on major components of the entire ecosystem than will the reduction in sea ice extent. It seems likely that only few species will become extinct as a direct effect of temperature increase, either because they proved unable to cope ecologically and physiologically with such an increase, or because they were restricted in their occurrence to an area with an above average temperature increase. Such biological changes are most likely at water depths down to 200 m, where a relatively high proportion of species is naturally used to environmental variability. Adaptation could result from a mobile life style, from effective dispersal of pelagic eggs and, or through ecological and physiological tolerance. The area with the highest benthic biodiversity and ecosystem functioning, at 150 – 300 m depth will remain much less affected since the predicted warming there is very low.

Most of the above projections do not consider the flexibility of organisms to “compensate” e.g. physiologically and ecologically for climate change impact. For example, two Emperor penguin colonies are known to live and reproduce even on land in contrast to their usual habitat, the sea ice. It can also not be excluded that adaptation through microevolution will become effective and have similar positive effects for a variety of species. While the potential for both mechanisms - flexibility and adaptation - might not be very high in long-lived and specialized species, it has to be recognized that the overwhelming majority of the species survived the last interglacial, when there was more warming than today. Nevertheless, some models of the future of the physical environment predict, at least regionally, temperatures that exceed those of the past interglacial. In this report we have chosen to use the average temperatures thrown up by IPCC 19 models, rather than any of the extremes.

The projected temperature changes are supposed to be insufficient to disrupt the thermal barriers that create the marine biological isolation of Antarctica, except locally where temperatures reach values significantly above average. As a consequence the invasion of species will likely remain restricted to areas where invaders can survive at their physiological limits. Under these circumstances it seems unlikely that invaders will have the potential to outcompete or diminish Antarctic species by predation pressure on a broad scale. In effect the Antarctic marine ecosystem will remain buffered from the direct effects of a global temperature-increase by the continued existence of the large and high ice sheet, which keeps the Antarctic cool and quite different from the Arctic. This assumption is confirmed by a newly developed bioclimate envelope model (Cheung et al., 2009), which predicts for 2050 an extremely sharp but non-continuous gradient between highly affected areas north of and close to the ACC and relatively stable conditions in high latitude Antarctic waters in terms of species invasion, local extinction, and turnover of metazoan species between. If, however, ocean temperatures increase by more than 2°C from today’s, and ecologically important species are less temperature tolerant in nature than in ecophysiological laboratory experiments, assemblages that cannot escape to colder regions are likely to become extinct by 2100 (Barnes and Peck, 2008). These may comprise thousands of invertebrate and vertebrate, pelagic and benthic species.

Our knowledge of the response of Antarctic calcifying organisms to acidification is extremely poor. Such species (coccolithorphorids and pteropods) can reach locally or regionally dense concentrations offshore, with pteropods also reaching high concentrations inshore at high latitudes. At the seabed, calcifying echinoderms, hydrocorals, bryozoans, and molluscs can shape local biodiversity or abundance hot-spots. Undersaturated conditions might become the most severe climate-induced negative impact on Antarctic benthic communities if key species cannot cope with increase pH, as in the case of the common Antarctic sea-urchin *Sterechinus neumayeri*; however, some species may even be winners in a more acid ocean, e.g. tunicates (Dupont and Thorndyke, 2009).

One of the most unpredictable parameters that shapes the Antarctic ecosystem is human behaviour. It has ranged from the enormous exploitation of natural resources in the past to the later protection not only of most of these but also of mineral resources. It still ranges on the one hand from more or less continuous global emissions of CO_2 to the atmosphere to proposed geoengineering projects for the Southern Ocean, both of which will or may cause irreversible ecological damage (Smith et al., 2008), to on the other hand the Madrid Protocol, one of the most strict international laws for protecting our global environment and to plans to identify, propose and declare High Seas Marine Protected Areas (HSMPAs) and Vulnerable Marine Ecosystems (VMEs).

Chapter 6

Recommendations

Chapter Editor:

John Turner

Authors:

Robert Bindschadler, Pete Convey, Guido di Prisco, Eberhard Fahrbach, Dominic Hodgson, Paul Mayewski, Colin Summerhayes and John Turner,

Great advances have been made in recent years into our understanding of Antarctic climate and environmental change. We now know that the climate system of the high southern latitudes is very complex and that there is variability on a range of time scales, with consequent effects on the terrestrial and marine biota. We also know that changes in the atmospheric and oceanic circulation around Antarctica, and the volume of the ice sheets, interact and influence climate at a global scale. Although a great deal of data are now available with which to investigate change – both in the past and over the next century, there are still major gaps in our knowledge and many areas where we require additional instrumental data gathering and model development.

The recommendations drawn together here summarise the conclusions we have reached in Chapters 4 and 5, which if treated separately would have led to some duplication. Other recommendations can be found at the end of Chapter 2, on specific observing system requirements, and at the end of Chapter 3, on research needed to improve the record of past climate change on geological time scales. In addition, scattered throughout the text there are many statements about additional research requirements.

- Collection of more *in-situ* data over the interior of the continent. AWSs are providing extremely valuable observations, but greater continuity in the observing programmes is needed and the maintenance of systems at selected sites for long periods. In addition to programmes describing Antarctic climatic variables at the macro-scale, there is an urgent need for the establishment of longer-term monitoring of biologically relevant microclimatic variables (as currently happens only during short-term biological studies), and the subsequent modelling effort to integrate patterns at the macro- and micro-scales.
- More traverses are required and more records extending back to at least 2000 years are needed at sites selected using information gained from ITASE efforts. Coastal ice core records not easily accessible by traverse need to be sampled, and there needs to

be further international collaboration to ensure that coverage over the continent is as complete as possible.

- Continuity of space-based measurements is absolutely essential, since these are the sole major source of data for the whole of the continent and its surrounding ocean, where measurements on the ground or on the sea are difficult, dangerous, and not normally made year-round. We must not go blind to ice sheets at the very moment when their behaviour has started to become highly significant in relation to changes in sea level. Recommendations for satellite observations of the cryosphere are given in the Cryosphere Theme document produced for the Integrated Global Observing Strategy (IGOS) Partnership (http://www.eohandbook.com/igosp/cryosphere.htm).
- Improved satellite systems are required to help estimate the mass balance of the Antarctic as we are at the limit of the present technology.
- More observations of temperature, salinity, biochemical properties, such as oxygen and flow in the Southern Ocean are required. It is essential for observing instruments to be developed and deployed in greater numbers throughout the year and for long periods as part of a Southern Ocean Observing System (SOOS) so that long time series of key oceanographic parameters can be obtained with sufficient spatial coverage. Ship borne observations have to be complemented by those from autonomous systems.
- Intensified use of highly sophisticated marine equipment such as ROVs, AUVs, crawlers, gliders, landers, and remote underwater laboratories would contribute to a significant enhancement in understanding Antarctic ecosystem functioning, and, consequently, provide the basis for improved predictions of the marine ecosystem response to climate and other environmental changes. AUVs and sustained underwater measurements are also key to understanding ocean-ice interaction, which has emerged as a primary driver of recent (and therefore likely future) large ice mass losses.
- The markedly different behaviour of Antarctic sea ice in comparison to that found in the Arctic requires special efforts to obtain *in-situ* observations of Southern Ocean sea ice properties.
- The Southern Ocean continues to be under sampled with respect to carbon cycle related properties. The international CO_2 community recommends for this region the construction of a CO_2 ocean data observing system and delivery of CO_2 ocean data products. Such data sets will provide valuable observational evidence for understanding historical climate change, and providing valuable insights into how the Southern Ocean may respond to further change. Such initiatives should be an integral component of the Southern Ocean Observing System required for monitoring and forecasting the ocean's role in Antarctic climate change.
- In terms of simulating the Southern Ocean response to historical climate change, some of the largest uncertainties in our results lie in the hydrological cycle (especially precipitation), the most poorly observed of all the forcing fields. To better understand and quantify the changes due to historical and future climate change, better estimates of the freshwater budget will be required.
- Although there are many observational programmes concerned with change in the Antarctic ice sheet, we still have little data on permafrost and the active layer. The lack of long-term monitoring data precludes drawing any definitive conclusions on

the impact of climate change on permafrost in Antarctica and this situation needs to be remedied through new observational initiatives.

- The last few years have also seen great advances in our understanding of terrestrial and marine ecosystems, and studies are now starting to address their resistance, resilience and adaptation to recent climate change. However, fundamentally important baseline biodiversity and biogeographic survey data are still lacking across most of the continent and parts of the surrounding Southern Ocean – those data and systematic and robust monitoring programmes across a network of representative locations are required to allow anything other than the current *ad hoc* and serendipitous approach to identifying biological responses to any aspect of environmental change in Antarctica. We also still require much more information on the links between the high latitude biota and broad-scale climatic factors, such as changes in the tropical atmosphere/ocean system (e.g. ENSO) and the modes of mid- and high latitude climate variability (e.g. the SAM).
- We recommend that the international community implement and monitor progress in the establishment of internationally recommended observing systems such as (a) CryOS (the Cryosphere observing system recommended for the IGOS Partners and adopted by the Group on Earth Observations), and (b) the Global Climate Observing System (GCOS).
- The Protocol on Environmental Protection to the Antarctic Treaty provides strict guidelines for the protection of the Antarctic environment and underscores its value to scientific research. Although rigorous application of the Protocol will help minimize the local impacts of both the tourism industry and national operators, constant vigilance is essential. Conservation measures should focus on achieving a better knowledge of the structure and functioning of Antarctic ecosystems and of the long-term effects of persistent contaminants in Antarctic organisms and food chains, and in developing continental-scale monitoring programmes based upon a network of carefully selected flagship sites.
- Higher horizontal and vertical resolution is needed in climate models to realistically represent many high latitude processes and their effects. Models must take into account in far greater detail than at present the complex orography in the coastal region, the behaviour of the atmospheric boundary layer, eddies in the ocean, and the effects of and sea ice. Sub-grid scale processes e.g. sea ice properties affecting atmosphere-ice-ocean interaction require improved parameterisations. Model outputs are also required at smaller physical scales relevant to the Antarctic habitats and communities, including the establishment and expansion of links between macro and microclimatic processes and trends.
- Climate model formulations need to be modified to recognise that parameters based on the behaviour of the atmosphere at low latitudes do not necessarily reflect processes operating in the Polar Regions, where the atmospheric boundary layer is commonly very stable. These models must include more sophisticated representation of the formation and melting of sea ice and its effects. In addition, the models need to be interactively coupled to ice shelf models so that the impact of changes in ocean circulation and water mass delivery below the shelf can be correctly simulated. This will lead to better predictions of sea level changes that might arise from interactions of the waters of the Southern Ocean with the periphery of the Antarctic Ice Sheet.
- Improved atmospheric chemistry needs to be included so that the models can better represent the effects of the ‘ozone hole’, including the important polar stratospheric

clouds. Greater spatial and temporal resolution are also imperative if biological processes, particularly on land, are to be integrated into future generations of climate models, and to permit objective tests of predictions of biological relevance. Advanced integrative and spatially explicit ecosystem modelling is needed to predict the future of the marine ecosystem. Such an approach demands widespread samples of ecological key species that are representative for ecological sub-systems, such as plankton, benthos or apex predators and long-term measurements of ecological key processes such as the response to acidification, warming and changes in ice cover and food regime.

- Realistic models are urgently required of the mechanical behaviour of the ice sheet and ice shelves in response to forcing by climate change, to underpin forecasts of likely sea-level rise and of the rates of change of ice sheet decay. To achieve this the next generation of ice sheet models must be able to account for rapid dynamical changes to the flow of glaciers and ice streams.
- Modelling efforts are also required to more fully understand the implications of Antarctic and Southern Ocean climate change throughout the Southern Hemisphere and globally, and *vice versa*.
- More observations are needed of permafrost, along with model predictions of permafrost change. It is import to expand the Circum-Polar Active Layer Monitoring (South) (CALM-S) network. To improve understanding of the development and evolution of permafrost under changing conditions in the Antarctic there needs to be an expanded Global Terrestrial Network for Permafrost sites (GTN-P) in Antarctica.
- A central location should be established for management of Antarctic permafrost, active-layer, and ground ice data.
- The PERMAMODEL should be applied to predict changes in permafrost distribution under different climate change scenarios, particularly along the Antarctic Peninsula and in maritime East Antarctica.
- Continued long-term and large-scale observations of functional and structural changes in ecosystems are essential to assess the sensitivity of ecological key species and to ground-truth predictive models. The establishment of a series of core long-term biological monitoring sites would be extremely beneficial both in documenting biological responses and trends, and allowing explicit tests of predictive hypotheses.
- More data on the marine biota are required for especially poorly studied areas like the Amundsen Sea, as the basis for the simulation of the impact of a warming ocean on marine biodiversity.
- Physiological and genomic studies currently interpreted as indicating vulnerability of certain Antarctic marine biota need placing in more ecologically realistic (longer term) timescales.
- Individual and species level responses (including resilience/resistance) to environmental variability and change require integration across communities, trophic webs and ecosystems.
- Biological colonisation routes and processes require identification and quantification in both terrestrial and marine environments, as does the relative importance of natural and human-mediated contributions to this process.

- Without a baseline biodiversity survey across much of the continent and Southern Ocean, objective documentation of future biological change and assessment of impacts will be impossible.
- Evidence should be sought for the possible effects of ocean acidification in Southern Ocean organisms.
- Comparisons should be made between southern and northern polar processes to shed light on evolutionary pressures and provide insight into gene selection.
- Many of the above recommendations will benefit from continued integration of cross-disciplinary expertise and approaches.
- Considerable improvement is needed in both the quantification of changes in precipitation (requiring an intense field programme), and the parameterization of the processes that drive precipitation. In due course, especially in the Antarctic Peninsula, biologists need to know what proportion of the precipitation is likely to fall as rain, since rain is immediately available to terrestrial biota.

- A better understanding of ecological driving forces within Antarctic ecosystems (terrestrial and marine) must serve as the basis for developing predictive models of the response of the Antarctic biota to climate change.

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