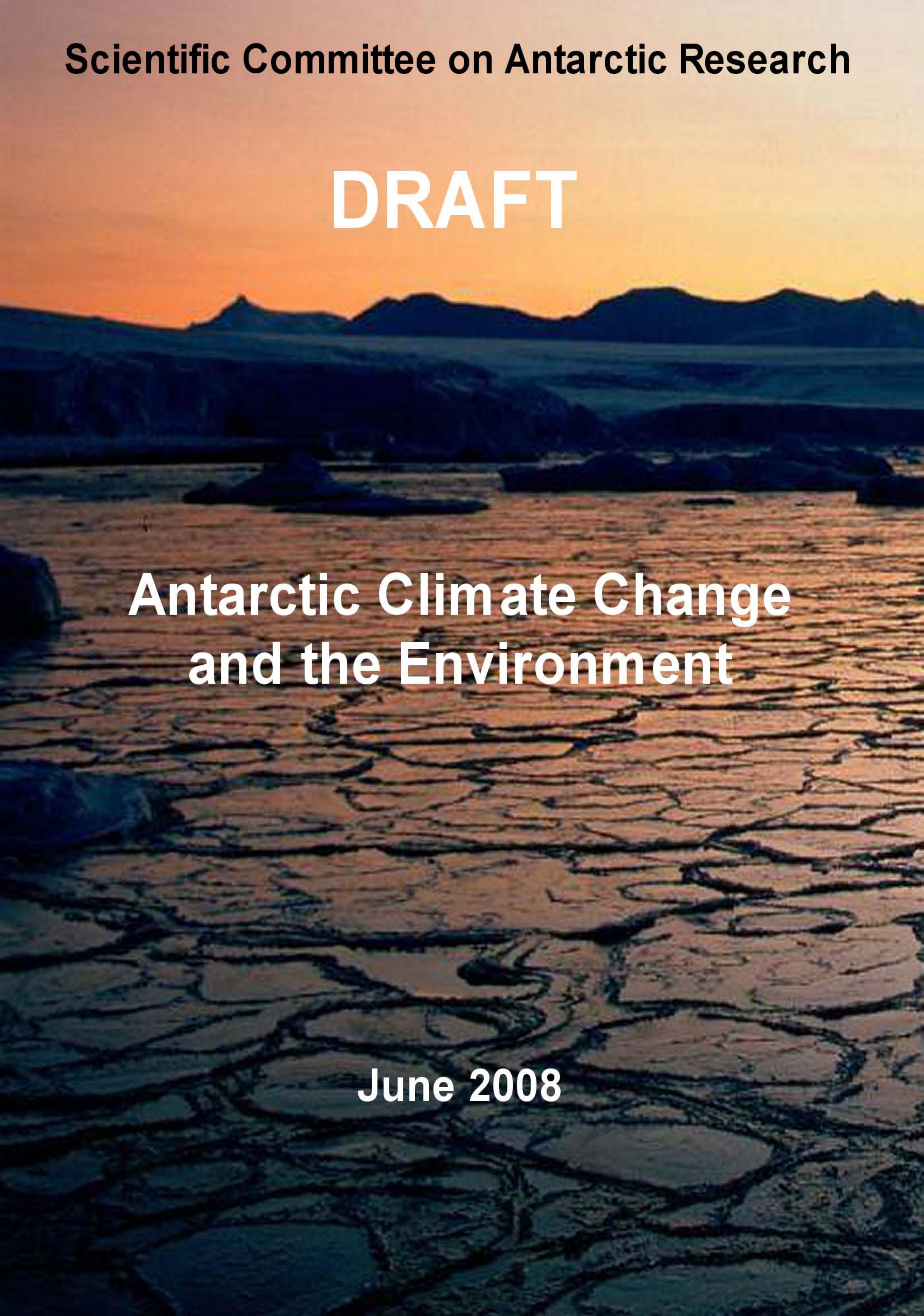
Scientific Committee on Antarctic Research

DRAFT



Antarctic Climate Change and the Environment

June 2008

Antarctic Climate Change and the Environment

Editors:

Dr. John Turner, British Antarctic Survey, UK
Dr. Pete Convey, British Antarctic Survey, UK
Prof. Guido di Prisco, Institute of Protein Biochemistry, Italy
Prof. Paul Mayewski, Climate Change Institute, University of Maine, USA
Dr. Dominic Hodgson, British Antarctic Survey, UK
Dr. Eberhard Fahrbach, Alfred Wegener Institute, Germany
Dr. Robert Bindschadler, NASA, USA
Dr. Colin Summerhayes, SCAR

Published by the Scientific Committee on Antarctic Research
Scott Polar Research Institute
Lensfield Road
Cambridge
CB2 1ER
United Kingdom

Contents

Executive summary

The Antarctic Environment in the Global System

The physical environment

The Antarctic is a major component of the Earth's climate system through its influences on the global atmospheric and oceanic circulations. Most of the heat from the Sun arrives in the tropics with there being a large poleward transport of heat in both hemispheres as a result of the large Equator to Pole temperature difference. Both the atmosphere and ocean play major roles in the poleward transfer of heat, with the atmosphere being responsible for 60 percent of the heat transport, and the ocean for the remaining 40 percent.

The orography of the Antarctic and high southern latitudes plays a extremely important part in dictating the climate of the continent. The Antarctic continent is located close to the Pole, with few other major orographic features being present in the Southern Hemisphere. The mean atmospheric flow and ocean currents are therefore very zonal (west – east) in nature. This means, for example, that the Antarctic Circumpolar Current (ACC), which is one of the major oceanic features of the Southern Ocean, can flow unrestricted around the continent isolating the high-latitude areas from more temperate mid- and low-latitude surface waters.

The isolation of the Antarctic close to the South Pole means that it has very low temperatures, especially on the high, interior plateau. In the coastal region temperatures are much less extreme, although at most of the coastal stations temperatures never rise above freezing point, even in summer. The highest temperatures on the continent are found on the western side of the Antarctic Peninsula where there is a prevailing northwesterly wind, and temperatures can rise several degrees above freezing during the summer.

The cold conditions at high latitude are responsible for the vast Antarctic ice sheet, which contains around 90% of the Earth's freshwater and covers around 99.6% of what we generally consider to be the Antarctic continent. In places the ice sheet is over 4500 m thick and the ice in the deepest layers is millions of years old. As the surface snows are buried by new snowfall, they are compressed and eventually transform into solid ice, a process that captures a chemical record of past climates and environments.

The continent is surrounded by the sea ice zone where, by late winter, the ice on average covers an area of 20×10^6 km^2, which is more than the area of the continent itself. At this time of year the northern ice edge is close to 60^o S around most of the continent, and near 55^o S to the north of the Weddell Sea. Unlike the Arctic, most of the Antarctic sea ice melts during the summer so that by autumn it only covers an area of about 3×10^6 km^2.

The Antarctic atmosphere, ocean and cryosphere are affected in non-linear ways by changes in atmospheric and oceanic conditions in the tropics and mid-latitude regions. For example, signals of the El Niño-Southern Oscillation can be found in Antarctic ice cores, but the Antarctic expression of near identical El Niño events can be different. In addition, there is increasing evidence that signals can also be transmitted in the opposite direction from high to low latitudes. The ozone hole is another example of extra-polar changes having a profound impact on the Antarctic environment, since the CFCs responsible for the 'hole' were emitted in the industrial areas, most of which are in the Northern Hemisphere. Most greenhouse gas emissions also come from such areas, yet have a profound effect on the radiation balance of the Antarctic atmosphere.

Marine and terrestrial biota

Levels of terrestrial biodiversity in the Arctic are strikingly greater than those even of the sub-Antarctic and much more so than the maritime and continental Antarctic. In comparison with about 900 species of vascular (higher) 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. However, despite the apparent ease of access to much of the Arctic, it is observed that a relatively low number of established alien vascular plants or invertebrates are known from locations such as Svalbard, in comparison with the *c*. 200 species introduced to the sub-Antarctic by human activity over only the last two centuries or so.

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 are barely represented 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. The recent anthropogenic introduction of vertebrate herbivores to most sub-Antarctic islands has led to considerable and negative impacts on megaherb-based communities

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, the microarthropod groups of mites and springtails, enchytraeids, earthworms, tardigrades, nematodes, spiders, beetles, flies and moths, with smaller representation of some other insect groups. 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. 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.

Climate Variability and the Antarctic

The Antarctic climate system varies on time scales from the sub-annual to the millennia and is closely coupled to other parts of the global climate system. On the longest time scales it has been found that the Antarctic ice sheet fluctuates on Milankovitch frequencies (20 ka (thousand years), 41 ka, 100 ka) in response to variations in the Earth's orbit around the Sun, which caused regular variations in the Earth's climate.

Proxy data that since the Last Glacial Maximum (LGM) at about 21K before present (BP) there have been a number of climatic fluctuations across the continent. One of the most marked has been the Mid Holocene (the Holocene itself began at ~11.7 K BP) warm period or Hypsithermal, which is present in various records from Antarctica. There is evidence for a Hypsithermal in East Antarctica but dating uncertainties are still high in some areas.

The instrumental period in Antarctic is only about 50 years in length, and since proxy data shows oscillations on longer time scales than this, it 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.

Reliable weather charts for high southern latitudes are only available since the late 1970s, but these have revealed a great deal about the cycles in the atmospheric circulation of high southern latitudes. The most pronounced climate cycle (or mode of variability) over this period has been the Southern Hemisphere Annular Mode (SAM), which is a flip-flop of atmospheric mass (as measured by a barometer) between mid-latitudes and the Antarctic coastal region. As a result of changes in the SAM there has been a marked drop in pressure around the Antarctic and an increase in mid-latitudes over the last few decades, so that the pressure gradient has increased across the Southern Ocean resulting in stronger surface winds. This has had implications for the distribution of sea ice and the oceanography. The changes in the SAM are thought to be primarily because of the loss of stratospheric ozone during the spring (the ozone hole), although increasing greenhouse gas concentrations, along with natural variability, have also played a part.

The weather charts have helped to establish the nature of the broadscale Southern Hemisphere high-low latitude climate linkages (teleconnections). However, there is evidence of decadal timescale variability in some of these links, 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. The ice core records have shed some light on teleconnections over the century timescale. The short cores can 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. Through most of the Holocene there has been a several hundred year time lag between Southern Hemisphere and Northern Hemisphere events, but in recent decades the Northern Hemisphere signal of rising temperature since about 1800 AD has paralleled the Southern Hemisphere one as depicted by the oxygen isotope signal at Siple Dome. 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.

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) requires decades to centuries to respond. At the other extreme, fast wave propagation in the ocean has timescales of just a few days.

On a year-to-year basis the Antarctic climate is more variable than conditions in the tropics or mid-latitude regions, as a results of feedbacks within the atmosphere-ocean-cryosphere.

Deep Time

Studying the history of Antarctic climate and environment is important as it provides the context for understanding present day climate and environmental changes both on the continent and elsewhere. Specifically 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. By knowing this natural variability we can accurately identify when present day changes exceed the natural state. The message from the palaeorecords is that change is normal and the unexpected can happen.

Levels of the greenhouse gas CO_2 in the atmosphere have ranged from roughly 3000 ppm (parts per million) in the Early Cretaceous (at 130 Ma (million years)) to around 1000 ppm in the Late Cretaceous (at 70 Ma) and Early Cenozoic (at 45 Ma), leading to global temperatures 6 or 7º C warmer than present. These high CO_2 levels were products of the Earth's biogeochemical cycles. In the Cenozoic temperatures gradually peaked around 50 Ma, with little or no ice on land.

Some 200 million years ago 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. As the Gondwanan fragments separated by sea-floor spreading between around 100 to 65 Ma, during the Cretaceous Period, Antarctica moved into a position over the South Pole. Up until the formation of a major ice sheet in the Oligocene (33-23 Myr BP), the terrestrial fauna and flora of Antarctica seem to have remained typical of south-temperate rainforest. Over time, South America, Africa, India, Australia and New Zealand moved away from Antarctica, opening the South Atlantic, Indian and Southern Oceans.

There is still uncertainty over when South America broke away from the tip of South America with the best estimate being that it was between 41 and 20 Ma.

The first continental-scale ice sheets formed on Antarctica in the Oligocene Epoch, around 34 Ma, and prior to the break of Tasmania from Antarctica. The development of the Oligocene ice sheet appears to have been a consequence of a decline in atmospheric CO_2 levels caused by reduced CO_2 outgassing from ocean ridges, volcanoes and metamorphic belts and increased carbon burial which dropped global temperatures at that time to levels around 4º C higher than today. The Oligocene ice sheets reached the edge of the Antarctic continent, but were most likely warmer and thinner than today's. Sharp cooling took place in the Miocene Epoch, at around 14 Ma, which was probably caused by the growing thermal isolation of Antarctica and related intensification of the ACC rather than by any change in CO_2. It 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 and into the Pleistocene at 1.8 Ma.

During the Pliocene, mean global temperatures were 2-3º C above pre-industrial values, CO_2 values may have reached 400 ppm, and sea levels were 15-25 m above modern levels.

Global cooling from around 3 Ma onwards led to the first ice sheets on North America and NW Europe around 2.5 Ma. These ice sheets enhanced the Earth's climate response to orbital forcing, taking us to the Earth's present "ice house" state , which for the last million years has been alternating between (i) long (40-100,000 years) glacial cycles, when much of the Northern Hemisphere was ice-covered, global average temperature was around 5º 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.

The establishment of the ACC and the Polar Front, created a barrier for migration of marine organisms between the Antarctic and lower latitudes, causing adaptive evolution to develop in isolation. 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 and began to diversify on the Antarctic continental shelf in the middle Tertiary, gradually adapting to the progressive cooling.

Over the past 30-40 Ma, in parallel with the diversification of the suborder, the physico-chemical features of the Antarctic marine environment have 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). With the local extinction of most of the temperate Tertiary fish fauna as the Southern Ocean cooled, the suborder experienced extensive radiation, dating from the late Eocene approx. 24 Ma, that enabled it to exploit the diverse habitats provided by a now progressively freezing marine environment. As temperatures decreased and ice appeared, Antarctic notothenioids acquired antifreeze glycoproteins

(AFGPs), an adaptation that allows them to survive and diversify in ice-laden seawater that reaches nearly -2° C.

The Last Million Years

The Antarctic ice core data show that the Earth's climate has oscillated through eight glacial cycles over the last 800,000 years, with CO_2 and mean temperature values ranging from 180 ppm and 10º C in glacials, to 300 ppm and 15º C in interglacials. 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, most likely in response to the increase in orbital eccentricity with time. The effect on global sea level was profound, with sea level dropping by 120 m on average during glacial periods. Ice cores from both Antarctica and Greenland show that temperatures were between 2-5º C higher than today in recent past interglacials. At the same time, global sea levels were 4-6 m higher than today's.

Diatom assemblages from 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. Such data have indicated 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. 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 4º, 5–10º, and 2–3º in latitude, respectively, compared to their present location. In the Atlantic and Indian sector, the sub-Antarctic Front would have shifted by around 4–5º and 4–10º in latitude, respectively.

The Holocene

The transition (Termination I) from the LGM (beginning about 21 ka BP) to the present interglacial period was the last major global climate change event. The main glacial to interglacial changes of the Pleistocene period, appear to be driven largely by orbital forcing, and in particular by the insolation of the Northern Hemisphere.

In general, geological evidence shows that deglaciation of the currently ice-free regions was completed earlier in East Antarctica compared with the Antarctic Peninsula, 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 around the continent, although there are some differences in the exact timing. In East Antarctica and the Antarctic Peninsula (excluding the northernmost islands) the Hypsithermal occurs somewhere between c. 4 and 2 K BP, whereas at Signy Island it spanned 3.6-3.4 – 0.9 K BP. Despite this there are a number of marine records that show a marine-inferred climate optimum between about 7-3 K BP and ice cores in the Ross Sea sector that show an optimum around 7-5 K BP, and the Epica Dome C ice shows an optimum between 7.5 and 3 K BP. The occurrence of a later Holocene climate optimum in the Ross Sea is in phase with a marked cooling observed in ice cores from coastal and inland locations. These differences in the timing of warm events in different records and regions points to 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 in particularly in the Dronning Maud Land region and particular regions of the Antarctic Peninsula 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.

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. 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.

The ice core record shows remarkable temporal detail that cannot be inferred from other proxy data. They show that the most dramatic changes in atmospheric circulation during the Holocene in the Antarctic are the abrupt weakening of the Southern Hemisphere westerlies at 5400-5200 years ago, and intensification of the westerlies and the Amundsen Sea Low starting around 1200 years ago.

With the exception of lake sediment studies, little terrestrial research has set out to examine changes in biodiversity, distributions and abundance over Holocene timescales. But it is becoming clear that across the continent and also the sub-Antarctic islands the contemporary biota is a result of vicariance (the separation or division of a group of organisms by a geographic barrier) and colonization processes that have taken place on all timescales between pre-LGM and pre-Gondwana-breakup. 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. 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. 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, then recolonising deglaciated areas), arrived through post-glacial dispersal from lower latitude islands and continents that remained ice free, or are present through a combination of both mechanisms.

On the oceanic islands, the biotas will have originally arrived via long-distance over ocean dispersal, with vicariance and terrestrial dispersal playing subsequent roles in shaping the biodiversity across glacial cycles.

The major impact of climate change on glacial timescales in the marine environment has been the glacial-interglacial expansion and contraction of the Antarctic ice sheet across the continental shelf and the consequent loss and recovery of benthic (bottom) marine habitats and the interglacial fluctuations in summer and winter maximum sea ice extent. Warm periods and expansions and contractions of the sea ice have also had an impact on marine mammal and seabird distributions. For example, changes in the Holocene distribution of marine birds can be tracked through the changing distributions of their nesting sites.

Changes During the Instrumental Period

The instrumental period began with the first voyages to the Southern Ocean during the seventeenth and eighteenth centuries, however, the greatest advance came with the International Geophysical Year (IGY) in 1957/58, which saw the establishment of many research stations across the continent. The ocean areas around the Antarctic have been investigated far less than the continent itself. Here we are reliant on ship observations which have mostly been made during the summer months. Satellite observations can help in monitoring the surface of the ocean, but not the layers below. And even here a quantity such as sea ice extent has only been monitored since the late 1970s when microwave technology could be flown on satellite systems.

The large scale circulation of the atmosphere

The largest change in the atmospheric circulation of the high southern latitudes has been the shift of the SAM into its positive phase, which has resulted in barometric pressures drops around the coast of the Antarctic and increasing surface pressure at mid-latitudes. This has increased the surface pressure difference between the Antarctic and the tropics, so increasing the westerly winds over the Southern Ocean by 15-20% since the late 1979s. This has also had oceanographic implications and influenced many other aspects of the Antarctic environment.

The SAM has changed because of the increase in greenhouse gases and the development of the Antarctic ozone, although the loss of stratospheric ozone has been shown to have had the greatest influence. The ozone hole is a phenomenon of the Austral spring, and at that time of year the loss of stratospheric ozone has resulted in a cooling of the Antarctic stratosphere, so increasing the strength of the polar vortex. However, during the summer and autumn the effects of the ozone hole propagate down through the atmosphere increasing the circulation around the atmosphere at lower levels. The greatest change in the SAM, which is indicative of surface conditions, has been during the autumn season.

The Antarctic Circumpolar Wave (ACW) is an apparent easterly progression of phase-locked anomalies in Southern Ocean surface pressure, winds, sea surface temperatures and sea ice extent. As such, it thus represents a coupled mode of the ocean-atmosphere system. The ACW has a zonal wavenumber of 2 (a wavelength of 180°) and the anomalies propagate at a speed (6-8 cm s^{-1}) such that they take 8-10 years to circle Antarctica giving the ACW a period of 4-5 years. While some authors have suggested that an ACW signal can be observed in an Antarctic ice core over the last 2000 years, since the initial discovery of the ACW others have questioned its persistence: a number of observational and modelling studies have indicated that the ACW is not apparent in recent data before 1985 and after 1994 somewhat fortuitously the period that was chosen for the original analysis. Whether the ACW is a persistent feature or just a transitory signal is therefore still under debate.

Mid-latitude low pressure systems (depressions) are a near-constant feature of the Southern Ocean and Antarctic coastal zone. The weather analyses available since the late 1950s allow us to track individual depressions, monitor depression formation/decay and examine movement of the main storm tracks. Using these data, it has been demonstrated 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 (the climatological belt of low pressure ringing the continent over 60-70^{o} S). One exception is the Amundsen-Bellingshausen Sea region.

In recent decades there has been a trend towards more frequent and more intense El Niño events. However, there is no evidence that this can be seen in the Antarctic. 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. However, the teleconnections (statistically significant linkages) 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.

Temperatures

Surface temperature trends across the Antarctic since the early 1950s illustrate a strong dipole of change, with significant warming across the Antarctic Peninsula, but with little change (or a small cooling) across the rest of the continent. 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 per decade 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, with the long record from Orcadas on Laurie Island, South Orkney Islands only having experienced a warming of +0.20^{o} C per decade. However, it should be noted that 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 per decade.

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. It was 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^{o} C per 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. However, the reason or reasons for this extensive sea ice are not known. The large winter season warming on the western side of the Antarctic Peninsula may therefore be a result of natural climate variability.

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^{o} C per decade over 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. 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 per decade, although this is not statistically significant.

On the interior plateau, Amundsen-Scott Station at the South Pole has shown a statistically significant cooling in recent decades that is thought to be a result of fewer maritime air masses penetrating into the interior of the continent.

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. Since the development of the ozone hole the trends have been towards a slight cooling around the coast of East Antarctica and a warming across the Antarctic Peninsula.

In recent decades many relatively short ice cores have been drilled across the Antarctic by initiatives such as the International Trans Antarctic Science Expedition, which provide data over roughly the last 200 years and therefore provide a good overlap with the instrumental data. The temperatures reconstructed from the cores 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^{o} 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.

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 (roughly 5 km above mean sea level) level with statistically significant changes of 0.5 – 0.7° C per decade. 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 during the winter as a result of the cooling stratosphere.

Snowfall

On average, about 6 mm global sea level equivalent falls as snow on Antarctica each year, so it is important to assess trends in Antarctic snowfall. However, measuring snowfall directly is difficult, and net accumulation (surface mass balance) is the quantity usually estimated. In recent decades, estimates of net accumulation over the Antarctic ice sheets have been made by three techniques: *in situ* observations, remote sensing, and atmospheric modelling.

The latest studies employing global and regional atmospheric models to evaluate changes in Antarctic net accumulation indicate that no statistically significant increase has occurred since ~1980 over the entire grounded ice sheet, West Antarctic Ice Sheet, or the East Antarctic Ice Sheet. Ice core net accumulation records can be extrapolated in space and time using the spatial information provided by atmospheric model precipitation fields from atmospheric reanalyses. Such data indicate that the 1955-2004 continent-averaged trend is positive and statistically insignificant (0.19 ± 32 mm per year), and is characterized by upward trends through the mid-1990s and downward trends thereafter.

The continent-averaged trend is the net result of both positive and negative regional trends, with positive trends on the western side of the Antarctic Peninsula. This, among other climate change factors, have been linked to observed decreases in Adélie penguin populations on the western side of the Antarctic Peninsula, by means of decreasing the availability of snow-free nesting habitat required by the birds.

The Antarctic ozone hole

Stratospheric ozone is an important constituent of the upper atmosphere above the Antarctic, but levels began to decline in the 1970s, following widespread releases of CFCs and Halons in 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 between 14 and 22 km altitude. 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 at about 1%

per year. However, there is little 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 (ozone mass deficit). 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, and is therefore comparatively insensitive to small changes in ODS amount.

Terrestrial biology

Terrestrial biological research within Antarctica has been rather 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.

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*) in the maritime Antarctic. At some sites numbers of plants have increased by two orders of magnitude in as little as 30 years, 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, and stimulate growth of seeds that have remained dormant in soil propagule banks.

Changes in both temperature and precipitation have already had detectable effects on lake 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.

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. 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 program and tourist operations.

The terrestrial cryosphere

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, that changes in floating ice shelves can cause significant changes in the grounded ice sheet. 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.

Retreat of several ice shelves on either side of the Antarctic 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. Ten ice shelves have undergone retreat during the latter part of the 20th Century.

Wordie Ice Shelf, the northernmost large (>1000 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) and the last remnant of the Prince Gustav Ice Shelf. A similar 'disintegration' event was observed in 1998 on the Wilkins Ice Shelf, but much of the calved ice remained until 2008 when dramatic calving removed about 14,000 km^2 of ice. 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, and melting in 2002 on the Larsen B was extreme. Specific mechanisms of ice-shelf break-up are still debated.

Some formerly snow-covered islands are now increasingly snow-free during the summer. Since the late 1940s, the total area covered by glaciers on Heard Island has reduced by approximately 11%, and several coastal lagoons have been formed as a result.

On the island of South Georgia, there are about 160 glaciers and of these, 36 have been mapped and analysed for changes over the past century. The results showed that two of the glaciers are currently advancing, 28 are retreating and 6 are stable or show a complex, ambiguous response.

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. Changes in the ice margin around the Antarctic Peninsula based on data from 1940 to 2001 have been compiled. 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.

The Amundsen Sea sector represents approximately one third of the entire West Antarctic Ice Sheet (WAIS). Recent observations have shown that this is currently the most rapidly changing region of the entire Antarctic ice sheet. Acceleration of flow due to basal melting of its ice shelf and subsequent grounding line retreat of Pine Island Glacier, one of the two largest WAIS outlet glaciers draining into the Amundsen Sea, was first reported in 1998. This discovery of a 10% increase 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. Later direct measurement of elevation loss near the grounding line revealed rates as high as 55 m per year, implying basal melting rates nearly 10 times the previously calculated value. As basal melt increased, the grounding line retreated, possibly in two stages—during the 1980s and 1994-96—each leading to a separate increase in speed. The most recent observations suggest the grounding line at Pine Island has retreated still further with a simultaneous increase in both speed and rate of speed increase. Pine Island Glacier is now moving at speeds 40% higher than in the 1970s.

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. 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 per year with the largest number accounting for the most recent faster glacier speeds. 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 per year of ice, while they receive only 177 ± 25 Gt per year of new snowfall.

Summer temperatures in the Amundsen Sea embayment rarely reach melting conditions, and there is little reason to consider that atmospheric temperatures have had any strong role to play in the changes that have occurred there. The most favoured explanation for the changes is a change in the conditions in the sea into which this portion of WAIS flows. 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. Elsewhere within West Antarctica, the changes are not as extreme.

Both the areas of most rapid change, the Antarctic Peninsula ice shelves and the Amundsen Sea sector outlet glaciers, are thought to be linked to greenhouse-forced changes in global circulation, specifically to changes in the SAM affecting the circulation around Antarctica, and the likely-related loss of sea ice cover in the Amundsen and Bellingshausen seas. A stronger (more positive) SAM, with increased westerly winds, also drives the Antarctic Circumpolar Current, pushing Circumpolar Deep Water against the western Peninsula coast and the northern coast of West Antarctica.

Changes are less dramatic across most of the East Antarctic ice sheet with the most significant changes concentrated close to the coast. Present changes in the ice sheet are a patchwork of 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 some recent passive microwave data. Satellite altimetry data indicate recent thickening in the interior that has been attributed to increased snowfall, but ice core data do not show recent accumulation changes are significantly higher than during the past 50 years. A resolution of this apparently conflicting evidence, may be that there is a long-term imbalance in this area, this could possibly be a response to much older 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. 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.

Sea level changes

The first three assessment reports of the IPCC arrived at similar conclusions with regard to global sea level change during the 20th century. For example, the third report concluded that global sea level had changed within a range of uncertainty of 1-2 mm per year. Since then a consensus seems to have been achieved that the 20th century rise in global sea level was closer to 2 than 1 mm per year, with values around 1.7 mm per year having been obtained for the second half of the last century in the most recent studies.

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 per year or more. This is at a higher rate than one might expect from IPCC projections, which has led to concern over possibly large ice sheet contributions during the 21st century, especially due to their dynamic instabilities. However, decadal variability in the rate of global sea level change makes it difficult to be confident that the apparent higher rates of the 1990s will be sustained, since high decadal rates have been observed at other times during the past century.

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 per year (which can be compared to an estimated total Antarctic Peninsula ice volume of 95,200 km^3, equivalent to 242 mm of sea-level). 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.

The Southern Ocean

The Southern Ocean plays a critical role in driving, modifying, and regulating global change. It is the only ocean that circles the globe without being blocked by land and is home to the largest of the world's ocean currents: the Antarctic Circumpolar Current.

Recent observational studies have suggested that the waters of the ACC have warmed more rapidly than the global ocean as a whole, with an increase of 0.17°C at depths between 700 – 1100 m over the 1950s to 1980s.

The ACC belt in the Australian sector has warmed in recent decades, as found elsewhere in the Southern Ocean. The changes are consistent with a southward shift of the ACC. Some climate models suggest the ACC will shift south in response to a southward shift of the westerly winds driven by enhanced greenhouse forcing. 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. However, there is no observational evidence of the increase in ACC transport also predicted by the models. 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. The poleward shift and intensification of winds over the Southern Ocean has been attributed to both changes in ozone in the Antarctic stratosphere and to greenhouse warming. In addition to driving changes in the ACC, the wind changes have caused a southward expansion of the subtropical gyres and an intensification of the southern hemisphere "supergyre" that links the three subtropical gyres.

Changes have been observed in several water masses in the Australian sector between the 1960s and the present. Waters north of the ACC have cooled and freshened on density surfaces corresponding to intermediate waters. South of the ACC, waters have warmed and become higher in salinity and lower in oxygen in the Upper Circumpolar Deep Water.

The Antarctic Bottom Water (AABW) in the Australian Antarctic basin has freshened significantly since the early 1970s, although the cause of this 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.

Considering its remote and inhospitable location, the Weddell Sea was well observed during recent major oceanographic campaigns, with (largely summer-time) hydrographic sections across the Weddell Gyre onto the Antarctic continental shelf, and with arrays of moorings. These indicate a number of decadal-scale changes in water mass properties. The Weddell Deep Water (WDW) warmed by some 0.04^{o} C during the 1990s and has subsequently cooled. This was accompanied by a salinification of about 0.004, 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^{o} C accompanied by a small increase in salinity, whereas there was no discernible change in the northern limb of the gyre.

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.

Biogeochemistry

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. A recent study in the South Western Indian Ocean calculated surface trends of CO_2 between 1991-2007. These results show that in the Southern Indian Ocean, the oceanic CO_2 increased at all latitudes south of 20°S. More specifically, at latitude poleward of 40°S, it was found that oceanic CO_2 increased faster than in the atmosphere since 1991, suggesting the oceanic sink decreased. In addition, when CO_2 data are normalized to temperature, the analysis shows that the system is increasing much faster in the winter than in the summer. These results suggest that the increase may be due to ocean dynamics given that the largest response occurs in the Austral Winter. In the recent period (since the 1980s) the increase of CO_2 appeared to be faster compared to the trends based on historical observations 1969-2002, suggesting that the Southern Ocean CO_2 sink has continued to evolve in response to historical climate change.

Sea ice

Ship observations have been compiled especially during the whaling period and these have been examined to estimate sea ice extent in the pre-satellite era. If we can assume that the ship data provides a meaningful representation of average ice edge locations during the period, it is apparent that the data from the previous several decades show relatively more extensive ice cover than the average from the satellite data. However, the ship locations are normally to the south of the farthest north the ice extended during the satellite era.

The sea ice extent data as derived from satellite measurements covering 1979-2006 show a positive trend of 0.98 ± 0.23% per decade for the entire Southern Hemisphere. Trend analysis of the ice extent in different sectors of the Antarctic region yields positive trends of varying magnitude in all except in the Bellingshausen Sea sector. The trend is least positive in the Weddell Sea sector at 0.8 ± 0.6% per decade followed by Western Pacific and Indian Oceans at 1.1 ± 0.8% per decade and 2.0 ± 0.6% per decade, respectively. The most positive is the Ross Sea sector at 4.5 ± 0.7% per decade.

Permafrost

Permafrost monitoring in Antarctica is a relatively new issue, however, some data on change are available. An analysis of data from Signy Island indicated that the active layer (the layer experiencing seasonal freeze and thaw) 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 endured a series of particularly cold winters.

The site at Boulder Clay (McMurdo Sound) represents the longest and almost continuous data series of permafrost and active layer temperature. Here there has been a substantial stability of the permafrost temperature at 360 cm, while at the permafrost table the temperature shows a slight decrease of 0.1° C per year. This slight decrease is mainly related to the decrease of the air temperature and the decrease of the snow cover in the winter, at least in this site.

Marine biology

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 environmental change on Antarctic ecosystems is not as severe as in the Arctic, where the sea ice cover is declining at an alarming rate. In the Antarctic, comparable shrinking of the winter ice cover has occurred only along the western Peninsula tip and the adjoining Scotia Sea. This is a relatively small region but home to the whale-krill-diatom food chain where more than 50 % of the Antarctic krill stock was concentrated.

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, their total biomass remained within only a few percent of that of the former whale population. About 300,000 blue whales were killed within the span of a few decades equivalent to more than 30 million tonnes of biomass. The majority of these whales were killed on their feeding grounds in the southwest Atlantic in an area of maximum 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.

There has been an 80% decline in krill stocks of the southwest Atlantic since the 1970s accompanied by an increase in salp populations. It is likely that phytoplankton production has decreased with that of the krill stocks, a conclusion supported by the increase in the salp population. A decline in phytoplankton concentrations can only be explained by a corresponding decline in the iron supply. There is reason to believe that the reduction in ice formation has resulted in a decrease in iron input from the shelf slopes of the Western Peninsula.

However, the extent of the krill decline and the underlying factors are under vigorous debate because of difficulties in unravelling the effects of industrial whaling from those of sea-ice retreat. Nevertheless, if the trend continues, recovery of the great whale populations will be jeopardised.

Contaminants

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 generally cause local disturbance of the Antarctic environment. As pesticides have neither been produced nor applied in the continent, the discovery of Dichlorodiphenyltrichloroethane (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, Hexachlobenzene (HCB), Hexachlorocycloexanes (HCHs), aldrin, dieldrin, chlordane, endrin, heptachlor and other POPs have been detected in Antarctica and the Southern Ocean. These chemicals are persistent, hydrophobic and lipophilic, accumulate in organisms and biomagnify in marine food chains

The Next 100 Years

Determining how the environment of the Antarctic will evolve over the next century presents many challenges, yet it is a crucial question that has implications for many areas of science, as well as for policymakers concerned with issues as diverse as sea level rise and fish stocks.

The evolution of the Antarctic climate over the next 100 years can only be predicted using coupled atmosphere-ocean-ice models. These have many problems in correctly simulating the observed changes that have taken place over the last few decades, so there is still a degree of uncertainty about the projections, particularly on the regional scale. 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, since it is very sensitive to changes in atmospheric and oceanic conditions.

The degree to which the climate of the Earth will change over the next century is 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 and fairly uniformly mixed across the Earth. Whatever happens it will take a long time for the levels of GHGs to drop. For instance, even if anthropogenic emissions of CO_2 were halted now it may take thousands of years for CO_2 concentrations to naturally return to pre-industrial amounts.

The IPCC developed a number of GHG emission scenarios, however, in the ACCE report we have concentrated on the most commonly used scenario, known as SRESA1B, which assumes a doubling of CO_2 and other gases by 2100. This seems a reasonable scenario considering GHG increases during the first decade of the century. In terms of temperature increase, the response to rises in GHGs in fairly linear, so that a quadrupling of gases would give temperature increases that were double those presented here.

Atmospheric circulation

Since about 1980 the SAM has moved rapidly into its positive phase. However, with the expected recovery of the ozone hole this forcing on the SAM should decline, but at the same time GHC concentrations will rise. Overall we expect the SAM to become more positive, but with a trend that is less rapid than has been observed over the last two decades. We can therefore expect to see further increases of surface winds over the Southern Ocean in the summer and autumn.

A consistent result that has emerged recently from 21st century model projections is a tendency for a poleward shift of several degrees latitude in the mid-latitude storm track.

Temperature

A significant surface warming over Antarctica is projected over the 21st century. The average of the SRESA1B scenario runs of the IPCC models shows an increase of the annual average surface temperature of 0.34°C per decade over land and grounded ice sheets. All the models show a warming, but with a large range from 0.14 to 0.5° C per decade. Due to the retreat of the sea-ice edge, the largest warming occurs during the winter when the sea ice extent approaches its maximum, e.g. 0.51 ± 0.26° C per decade off East Antarctica.

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. 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 from simulations and observations of temperature change for the latter part of the 20th century. The most notable difference is that the observed and simulated maximum of warming over the Antarctic Peninsula for the latter part of the 20th century is not present in projections of change over the 21st century.

Model consensus for warming is strong for Antarctica as a whole. However, there is large uncertainty in the regional detail. There is less confidence of the large warming trends around the coast than the smaller changes over the high interior. This is due to the large uncertainty over the sea ice and ocean projections.

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 warming is much smaller than the global average due to the large heat uptake.

The annual mean warming rate at the 5 km above sea level is 0.28° C per decade, which is slightly smaller than the surface warming, with no evidence of a mid-tropospheric maximum, which has been observed over the last 30 years.

Very little work has been done on changes to extremes over Antarctica. 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.

The IPCC warming of 3°C over the next century is considerably faster than the fastest rate of rise observed in Antarctic ice cores (4° C per 1000 years). But it is comparable to or even slower than the rapid rates of temperature rise typical of Dansgaar-Oeschger Events during glacial times in Greenland, and of the Bolling Allerød warming in Greenland at 14,700 ka BP, and the warming in Greenland at the end of the Younger Dryas around 11,700 ka BP.

Precipitation

Almost all climate models simulate a robust precipitation increase over Antarctica in the coming century. The projected precipitation change has a seasonal dependency, and is larger in winter than in summer. If the IPCC model output is weighted according to the skill of the models in simulating recent change, they suggest that by the end of the century the snowfall rate over the continent would increase by 20% compared to current values. If other effects such as melting and dynamical discharge are ignored, this would result in a negative contribution to global sea-level rise of approximately 5 cm.

With the expected southward movement of the mid-latitude storm track we can expect greater accumulation in the Antarctic coastal region.

The ozone hole

Stratospheric ozone is affected by a number of natural and anthropogenic factors in addition to reactive halogens: temperature, transport, volcanoes, solar activity, hydrogen oxides, nitrogen oxides. In considering future ozone concentrations, it is important to separate the effects of these factors. Recovery of stratospheric ozone amounts will be slow. However, models suggest that by the end of the 21st Century, Antarctic stratospheric 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.

Tropospheric chemistry

Over the ocean, a loss of the sea ice would enable emissions of trace gases with an oceanic origin. Various gases, such as dimethyl sulphide (DMS), that are measured in the troposphere are known to be released from the oceans around Antarctica. Such gases are generated by phytoplankton. The seasonal cycle in the atmosphere of these trace gases is closely linked to the extent of sea ice but also to the Sun. In a warmer world with a reduced sea ice extent, emissions from the ocean would likely increase. The dominant control would then be the Sun, so that a seasonality with a wintertime minimum but an extended summer maximum could be expected. DMS plays a critical role as a source of cloud condensation nuclei (CCN) via its oxidation to sulphate. Changing the number of CCN alters cloud properties and albedo with a consequent influence on the Earth's radiation budget, surface temperature and climate.

Terrestrial biology

Climate change will impose a complex web of threats and interactions on the plants and animals living in the ice-free areas of Antarctica. Increased temperatures may promote growth and reproduction, but may also contribute to drought and associated effects. High amongst future scenarios is the likelihood of invasion by more competitive alien species, easily carried there by humans.

Many sub-Antarctic islands are showing increases in mean annual temperature. 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. But an increase in the frequency and intensity of freeze–thaw events could very readily exceed the tolerance limits of many arthropods.

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. Future regional patterns of water availability are unclear, but climate models predict an increase in precipitation, especially in the Antarctic coastal region.

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 that available for other life history components.

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.

The terrestrial cryosphere

Recent observations of ice sheet behaviour in Greenland and Antarctica have forced experts to radically revise their view of ice sheet sensitivity to climate change. Existing ice-sheet models, do not properly reproduce this observed behaviour, casting doubt on the value of these models at predicting future changes. Predictions of the future state of ice sheets must, therefore, be based upon a combination of inference from past behaviour, extension of current behaviour, and interpretation of proxy data and analogues from the geological record.

The most likely regions of near-future change are those that have been shown to be changing today. However, most models agree that, although warming was not observed everywhere in Antarctica in the last 50 years, it will be strong in the coming century, and so it is likely both snowfall and melt will increase during this century. Even the relatively small increases in the rate of snowfall that are expected to parallel this warming, would cause significant volumetric growth due to the vast area of the ice sheet. But warming 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.

The East Antarctic ice sheet has experienced interior thickening, probably a long-term dynamic response to a distant change in climate and not recent increased snowfall. This effect is likely to continue and change only slowly. 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.

There is currently a loss of mass from the Amundsen Sea embayment of the West Antarctic ice sheet, thought to be a result of a progressive thinning of the fringing ice shelves seaward of the Amundsen Sea outlet glaciers as warmer waters get onto the continental shelf. A continuation of the positive phase of the SAM, 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 0.75 m to global sea level if the area of ice lost is limited to that where the bed slopes downward toward the interior of West Antarctica, so a contribution from this sector alone of some tens of centimeters by this century's end cannot be discounted.

In an assessment of expert glaciological opinion made in 2000, it was determined that a group of leading glaciologists believed that within the next 200 years there remained a 30% probability that loss of ice from the West Antarctic ice sheet could cause sea level rise at a rate of 2 mm per year, and a 5% probability it would cause rates of 1 cm per year. Since that opinion was gathered there has been enormous scientific progress made in understanding the ice sheet. However, little has been discovered that would cause a reduction of the levels of risks expressed in that expert judgment. Conversely, the recent observations that inland ice sheets can be impacted by the loss of floating ice shelves and the continued acceleration of ice-sheet thinning and glacier-flow in the Amundsen Sea embayment are now firmly understood to be the result of glacier-acceleration, and can no longer be argued to result from a few years of unusually low-snowfall rates (as might have been interpreted at the time that the opinion was gathered, but can be argued to support the hypothesis that the ASE could already be entering a phase of collapse). Similarly, recent improvements in numerical analysis of the stability of marine ice sheets, which are supported by many other ice-sheet modellers, appear to reinforce earlier concerns that a potential instability does exist in marine ice sheets that could lead to deglaciation of parts of WAIS.

Measurements made of the conditions in the Weddell Sea sector of West Antarctica do not raise alarms now or for the future of the terrestrial cryosphere, since the open ocean is held well away from where ice streams first enter the Filchner-Ronne Ice Shelf.

Around the Antarctic Peninsula increased warming will lead to a southerly progression of ice shelf disintegrations along both coasts. As in the past, these may well be evidenced and preceded by an increase in surface meltwater lakes, and/or progressive retreat of the calving front. Prediction of the timing of ice shelf disintegration is not yet possible.

Although at present most of the effects leading to loss of ice are confined to the more northerly section of the Antarctic Peninsula (which contains a few centimetres of global sea-level rise), the total volume contained on the Antarctic Peninsula is 95,200 km^3 (equivalent to 242 mm of sea-level). This is roughly half that of all glaciers and ice caps outside of either Greenland or Antarctica. The mechanisms that have led to recent acceleration of these glaciers as an immediate consequence of ice shelf disintegration, and glacier retreat, which could progress further south in the coming century, suggests that this ice could impact global sea level relatively rapidly, and perhaps be comparable to the more gradual melting of glaciers and ice caps in other mountainous environments across the Earth.

Sea level

For the IPCC's Fourth Assessment Report, 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. This took into account the full range of emission scenarios and climate models, but did not include a contribution from the dynamic instability of Greenland. However, sea levels might be expected to rise around Antarctica itself at rates lower than in other parts of the world, owing to the important role of the ocean adjustment to climate forcings and, if the rise stems from Antarctic melt, then also due to the elastic response of the solid earth.

Overlying the global sea-level rise is a large regional variability and sea-level rise during the 21st century is not expected to be uniform around the globe. This is a result of changing atmospheric conditions, particularly surface winds 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 minimum sea level rise in the Southern Ocean is due the thermal expansion coefficient being lower in cold waters than warmer waters.

Biogeochemistry

The response of the Southern Ocean to climate change remains highly uncertain and simulations from coupled climate carbon models show a large range of responses.

Patterns have emerged from future scenarios that suggest that the Southern Ocean will be an increased sink of atmospheric CO_2 in the future and that the recent decrease in uptake will not continue into the future. The magnitude of the total uptake is very 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 changes and density changes. The increased absorption of atmospheric CO_2 and upwelling of deep-water impacts the ability of the ocean to store CO_2.

Studies of the future global uptake in coupled and uncoupled simulations show that the global marine ocean carbon cycle acts as a positive feedback on climate change i.e. amplifying climate change. This highlights the importance of the Southern Ocean as it is expected to act as a negative feedback in the 100 years. We note also that Southern Ocean uptake changes are vulnerable to changes in the terrestrial biosphere. Changes in uptake, particularly at mid-latitudes, can be very large and therefore can significantly impact the response of the Southern Ocean in the future by impacting the gradient between the atmosphere and the ocean.

The ocean circulation and water masses

An intensification of the ACC is generally simulated in response to the southward shift and intensification of the westerly winds over the Southern Ocean simulated over the 21st century. Averaged over all the model simulations performed in the framework of the IPCC AR4, the increase in transport projected for the end of the 21st century reaches a few sverdrups in the Drake passage. The enhanced winds induce in addition a small (less than 1° in latitude on average) but robust southward displacement of the core of the ACC. Furthermore, nearly all the models simulate regional changes in the horizontal oceanic circulation, but the patterns of the projected changes vary significantly between the models. One exception is probably the Ross Sea where a large number of models simulate a cyclonic anomaly at the end of the 21st century compared to the late 20th century.

During the 21st century, the observed mid-depth warming of the Southern Ocean is projected to continue, reaching nearly all depths. However, close to the surface, the warming of the Southern ocean during the 21st century is expected to be weaker than in other regions.

The ocean ventilation could be enhanced because of the surface divergence induced by the increase in the wind stress projected during the 21st century.

Sea ice

With the SRESA1B scenario, over the 21st century the IPCC models suggest that the annual average total sea-ice area is projected to decrease by 2.6×10^6 km^2, which is a 33% decrease compared to current values. Most of the ice retreat is expected to be in winter and spring when the sea ice extent is largest. The amplitude of the seasonal cycle of sea ice area will therefore decrease.

There is strong confidence in the Antarctic-wide decreases of sea ice extent. However, it is more difficult to produce regional predictions. But in the regions where sea ice currently remains present throughout the summer, in particular the Weddell Sea, large reductions of sea ice extent are expected. On this there is quite a strong consensus between the models.

Permafrost

Given the possible warming scenarios in Antarctica, it is anticipated that over the next 100 years there will be a reduction in permafrost area, subsidence of ground surface, changes in hydrology and mass movements.

The areas most susceptible to permafrost degradation lie within the zones of discontinuous and sporadic permafrost, which includes the northern Antarctic Peninsula and the South Shetland and South Orkney Islands and possibly coastal areas in East Antarctica.

There are approximately 80 year-round and summer bases in Antarctica, of which about 90% are in areas that are sensitive to thermokarst and mass wasting. There are therefore risks to infrastructure.

Marine biology

How the rich biodiversity of the Southern Ocean will respond to medium/long term warming is unclear. Certain changes over decades have been found in pelagic and benthic populations of some Southern Ocean species but none definitively linked to climate change. Most of the evidence currently being discussed for how benthos might cope with temperature rises is experimental. Being typically ‘stenothermal’ (capable of living or growing only within a limited range of temperature) is considered a key trait characterising Antarctic marine animals and if this is the case they would be highly sensitive to predicted climate change. Experiments have shown that most species, including a wide range of invertebrates, had upper lethal temperatures below or near to 10° C. Some can survive just a 7° C temperature window. Even in the most severe warming scenarios a rise of this magnitude in the Southern Ocean is very unlikely by 2100, but organisms can be critically affected at lower temperatures long before lethal levels.

So whether populations or species will survive future temperature rises may be more dictated by their ability to do critical activities e.g. feeding.

There will be acute agents of disturbance to the marine biota, including 1) rapid increases in both ice-loading and coastal concentrations of large icebergs from iceshelf collapses; 2) coastal sedimentation associated with ice melt, smothering benthos and hindering feeding; 3) freshening of surface waters leading to amongst other changes, stratification.

The chronic impacts of climate change considered important are 1) long term decreases in ice scour probably leading to increased local scale, but decreased regional biodiversity; 2) the physiological effect of direct warming leading to reduced performance at critical activities and thus geographic and bathymetric migrations; 3) increased acidification leading to skeletal synthesis and maintenance problems, and; 4) slight deoxygenation of surface waters but disruption of currents and downwelling could ultimately lead to more serious deoxygenation for deeper layers.

Concluding remarks

The climate of the high latitude areas is more variable that 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 expected increases of GHGs over the next century, if they continue to rise at the current rate, will be remarkable because of their speed. We can make reasonable broad estimates of how quantities such as temperature, precipitation and even sea ice extent might change, and consider the possible impact of marine and terrestrial biota. However, we cannot say with confidence how the large ice sheets of Antarctic will respond, but recent rapid changes observed give cause for concern.

PREFACE

The Antarctic Climate Change and the Environment (ACCE) project is an initiative of the Scientific Committee on Antarctic Research (SCAR), which aims to provide an up-to-date assessment of the climatic changes that have taken place on the continent and across the Southern Ocean, give improved estimates of how the climate may evolve over the next century and examine the possible impact on the biota and other aspects of the environment.

The project considers the meteorology and climatology of the Antarctic, the oceanography of the Southern Ocean, biogeochemistry, the ice sheet, the ice shelves, ice caps and glaciers around the Antarctic Peninsula and sub-Antarctic islands, sea ice, frozen ground environments, and terrestrial and marine biota.

The focus of ACCE is on climatic, environmental and biological changes that have taken place over roughly the last 10,000 years (The Holocene), with particular emphasis on the last 50 years when more in-situ data are available. To set recent changes in a broader context we also include a brief summary of changes that occurred during Deep Time. A particular target was to cover attribution of past climatic changes and make useful statements about the changes that have taken place as a result of natural climate variability and the possible impact of anthropogenic factors.

For prediction of future climate change the most important factor is how greenhouse gas emissions will change over the coming decades. Within this aspect of the project we have used the various greenhouse gas emission scenarios of the Intergovernmental Panel on Climate Change (IPCC) (www.ipcc.ch). We have also drawn extensively on the series of climate model integrations that were carried out for the IPCC Fourth Assessment Report (AR4). While the IPCC was a tremendous initiative that provided a major advance in our understanding of natural climate variability, the role of anthropogenic factors in recent climate change and the best projections yet for how climate may evolve over the next century, it's scope was global and there was limited space to consider the Antarctic. The ACCE review was therefore conceived as a study that could consider in more depth than was possible within IPCC the complex environment of the Antarctic. It's goals are somewhat similar to the Arctic Climate Impact Assessment (ACIA) (http://amap.no/acia/) in that there was a need for a detailed regional review of past and possible future climate in a climatically-sensitive region.

The plan for ACCE was formulated at the SCAR Executive meeting in Sofia, Bulgaria in July 2005 and approved by the SCAR Delegates at their meeting Hobart, Australia during July 2006.

The members of the SCAR Antarctica and the Global Climate System (AGCS) programme were asked to lead the preparation of the ACCE report, but many other scientists were involved. A small editing team was established that were responsible for issuing the invitations to write the various sections of the report, collate the contributions and carry out an initial editing. The members of the editing team are:

Dr. John Turner, British Antarctic Survey, UK (J.Turner@bas.ac.uk)
Dr. Pete Convey, British Antarctic Survey, UK (P.Convey@bas.ac.uk)
Prof Guido di Prisco, Institute of Protein Biochemistry, Naples, Italy (g.diprisco@ibp.cnr.it)
Prof. Paul Mayewski, Climate Change Institute, University of Maine, USA (Paul.Mayewski@maine.edu)
Dr. Dominic Hodgson, British Antarctic Survey, UK (D.Hodgson@bas.ac.uk)
Dr. Eberhard Fahrbach, Alfred Wegener Institute, Germany (Eberhard.Fahrbach@awi.de)
Dr. Bob Bindschadler, NASA, USA (robert.a.bindschadler@nasa.gov)
Dr. Colin Summerhayes, SCAR (cps32@cam.ac.uk)

This draft version of the ACCE report is being circulated widely to the Antarctic science community. It is hoped that the draft will be discussed extensively at the SCAR/IASC Open Science Conference in St Petersburg and the SCAR Delegates' Meeting in Moscow, both of which will be held in July 2008. Comments on the draft can be sent to any of the editors listed above. It is planned to publish the final report later in 2008.

The ACCE review is very cross-disciplinary in nature and every effort has been made to try and integrate the physical and biological aspects of the study into a document that is accessible to the widest possible readership. Jargon and more technical terms have been kept to a minimum, yet copious references have been provided for those seeking more detailed information. A summary has also been provided for wide dissemination.

The ACCE review is the first of its type and tries to place Antarctica in its global context. It has the potential to make an important contribution to our understanding of the role of the Antarctic in the Earth system and we would like to thank all the scientists who contributed to this report, as well as those who reviewed various sections. We would also like to thank Ms. Gill Alexander who assembled and formatted the report.

Chapter 1

The Antarctic Environment in the Global System

Coordinating Lead Author:
John Turner

Lead Authors:
John Turner, Pete Convey, Eberhard Fahrbach

Contributing Authors:
Dominic Hodgson, Robert Bindschadler, Peter Barrett

Review Editors:

1.1 Introduction

The Antarctic is the region south of 60° S and includes the Antarctic continent, a number of the sub-Antarctic islands and a large part of the Southern Ocean. The whole continent lies within the Antarctic Circle, except for the northern part of the Antarctic Peninsula.

The continent covers an area of 14 x 10^6 km^2, which is about 10% of the land surface of the Earth. This figure includes the area of the ice sheet, the floating ice shelves and the areas of fast ice. The continent is dominated by the Antarctic Ice Sheet, which contains around 30 million km^3 of ice or about 70% of the world's fresh water. The Antarctic has the highest mean elevation of any continent on Earth, and reaches a maximum elevation of over 4,000 m in East Antarctica.

The ice sheet is made up of three distinct morphological zones, consisting of East Antarctica (covering an area of 10.35 x 10^6 km^2), West Antarctica (1.97 x 10^6 km^2) and the Antarctic Peninsula (0.52 x 10^6 km^2). The orography rises very rapidly inland from the coast and the continent has a domed profile, with much of it being above 2,000 m in elevation.

West Antarctica is lower than East Antarctica, with a mean elevation of 850 m. However, some areas do reach more than 2000 m in height, with exposed mountain peaks rising above the ice sheet to more than 4000 m. East and West Antarctica are separated by the Transantarctic Mountains, which extend from Victoria Land to the Ronne Ice Shelf and rise to a maximum height of 4,528 m.

The Antarctic Peninsula is the only part of the continent that extends a significant way northwards from the main ice sheet. It is a narrow mountainous region with an average width of 70 km and a mean height of 1500 m. The northern tip of the peninsula is close to 63° S, so that this barrier has a major influence on the oceanic and atmospheric circulations of the high southern latitudes.

1.2 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, so creating a large Equator to pole temperature difference. The atmospheric and oceanic circulations respond to this large horizontal temperature gradient by transporting heat polewards. 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.

Both the atmosphere and ocean play major roles in the poleward transfer of heat, with the atmosphere being responsible for 60 percent of the heat transport, and the ocean the remaining 40 percent. In the atmosphere, heat is transported by both low pressure systems (depressions) and the mean flow. The 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 contrasting orography of the two polar regions is very important in prescribing the atmospheric and oceanic circulations of the Northern and Southern Hemispheres. The Antarctic continent is located close to the pole, with few other major orographic features being present in the Southern Hemisphere. The mean atmospheric flow and ocean currents are therefore very zonal in nature. For example, the Antarctic Circumpolar Current, which is one of the major oceanographic features of the Southern Ocean, can flow unrestricted around the continent isolating the high latitude areas from more temperate mid- and low-latitude surface waters. This has only been the case since about 30 million years ago when the Drake Passage, which connects the South Atlantic Ocean to the South Pacific Ocean, opened up. Prior to that time the ocean currents had more of a meridional component that allowed greater penetration poleward of more temperate water masses.

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 back to space, aided by the relatively cloud-free atmosphere that also contains little water vapour. This is one of the important feedback mechanisms found in the polar regions where a cooling can be enhanced much more than with the unfrozen ocean or bare ground.

1.3 The climate of the Antarctic

The plateau of East Antarctica experiences very low temperatures because of the very high elevation, the lack of cloud and water vapour in the atmosphere, and the isolation of the region from the relatively warm maritime airmasses found over the Southern Ocean. The high plateau of East Antarctica is located slightly away from the South Pole, which has climatological implications for the atmospheric circulation over the Southern Ocean.

The very cold temperatures in the interior of the Antarctic, coupled with its isolation from warm, moist airmasses mean that precipitation amounts here are very low, with only about 5 cm water equivalent falling per year. The Antarctic is therefore a desert and the driest continent on Earth. But the low temperatures mean that there is very little evaporation and sublimation, so that even this small amount of precipitation builds up year by year to form the major ice cap.

In the Antarctic coastal region temperatures are much less extreme than on the plateau, although at most of the coastal stations temperatures never rise above freezing point, even in summer. The highest temperatures on the continent are found on the western side of the Antarctic Peninsula where there is a prevailing northwesterly wind, and here temperatures can rise several degrees above freezing during the summer.

At stratospheric levels above the Antarctic in winter there are very cold temperatures because of the lack of incoming solar radiation. This results in a strong temperature gradient between the continent and mid-latitudes (see Fig. 1.1 below), along with very strong winds around the Antarctic – stronger than around the Arctic because the pole to Equator temperature difference is larger. This pool of cold air above the Antarctic and strong surrounding winds is known as the polar vortex. It plays a very important part in determining the atmospheric circulation of the high southern latitudes, as well in the formation of the ozone hole, where this ‘containment vessel’ allows CFC to build up during the winter (see section 3.6.1).

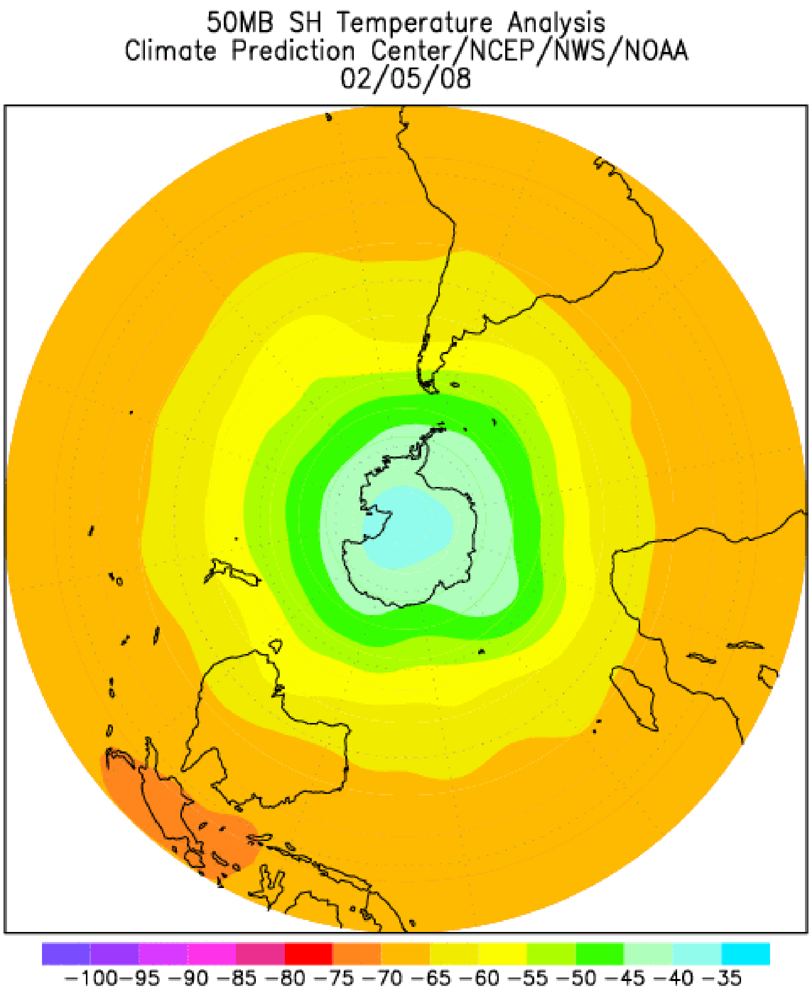


Fig. 1.1 The polar vortex above Antarctic as seen via the 500 hPa (roughly 5 km elevation above mean sea level) temperatures.

1.4 The Antarctic cryosphere

The Antarctic ice sheet comprises the vast contiguous coverage of glacial ice that rests on the Antarctic continent and surrounding seas (Figure 1.2). It is the single largest solid object on the surface of the planet containing around 90% of the Earth's freshwater and covering around 99.6 % of what we generally consider to be the Antarctic continent. The ice sheet is nourished at its surface by deposition of snowfall and frost, which, because of the year-round cold environment, does not melt but accumulates year-on-year. As the surface snows are buried by new snowfall, they are compressed and eventually transform into solid ice, a process that captures a chemical record of past climates and environments. In places the ice sheet is over 4500 m thick and the ice in the deepest layers is millions of years old. Glacial flow produces a relentless movement of ice towards the sea, and eventually it either calves away in icebergs or melts directly into the coastal waters – the entire Antarctic ice sheet can be considered as an immense conveyor belt, transporting ice from the interior to the coast at a rate of around 2000 billion tonnes per year. This conveyor is naturally regulated to balance loss by calving and melt with nourishment from snowfall. The continuation of this balance is important because any substantial imbalance will have a noticeable impact on world sea levels.

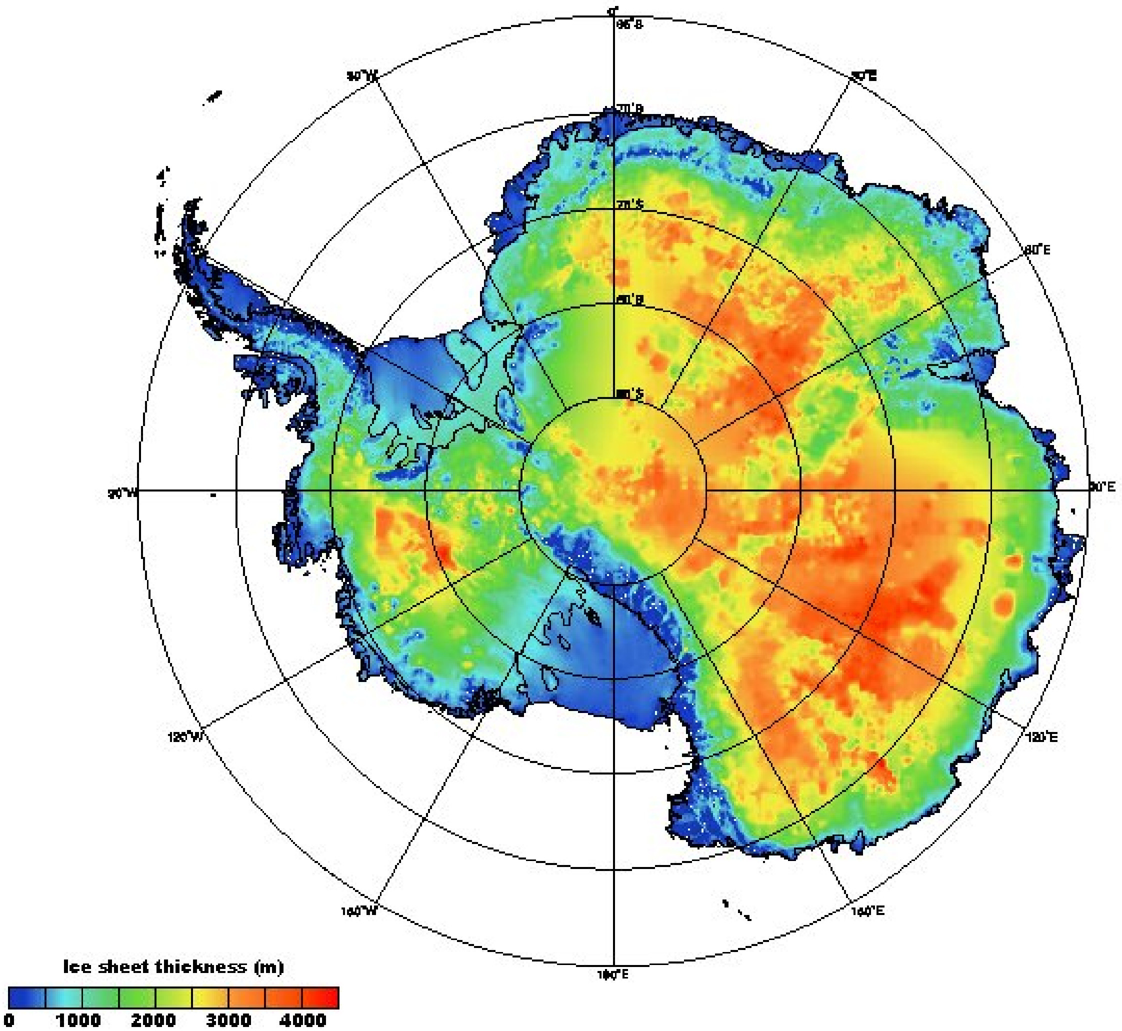


Fig. 1.2. The thickness of the Antarctic ice sheet.

The Antarctic ice sheet has a maximum thickness of about 4700 m and this huge mass of ice gradually flows down to the edge of the continent in a number of ice streams that move at speeds of up to 500 m per year. Once the ice streams reach the edge of the continent they either calve into icebergs, which move northwards, or start to float on the ocean as ice shelves. The ice shelves constitute 11% of the total area of the Antarctic, with the two largest shelves being the Ronne-Filchner and the Ross Ice Shelves, which have areas of 0.53×10^6 km^2 and 0.54×10^6 km^2 respectively. The ice shelves are several hundreds of meters thick and the ocean areas under them are important for the formation of cold, dense Antarctic Bottom Water, which is discussed below.

The continent is surrounded by the sea ice zone where, by late winter, the ice on average covers an area of 20×10^6 km^2, which is more than the area of the continent itself. At this time of year the northern ice edge is close to 60^o S around most of the continent, and near 55^o S to the north of the Weddell Sea. Unlike the Arctic, most of the Antarctic sea ice melts during the summer so that by autumn it only covers an area of about 3×10^6 km^2. Most Antarctic sea ice is therefore first year ice, with the largest area of multi-year ice being over the western Weddell Sea. Consequently most sea ice around the continent is relatively thin, with an average thickness of about 1-2 m.

Permafrost is whatever earth material that remains below 0 ºC for two or more consecutive years while the overlying layer which, seasonally, can experience temperatures above 0°C is called “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 part of the ice sheet and the glaciated areas, and even in some spots of the sea floor in Continental Antarctica. It is therefore found across parts of the Antarctic Peninsula and is also widespread 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 Dry Valleys and along the narrow coastal zone of East Antarctic. The Dry Valleys are a particularly interesting area where liquid water is rare, yet there is extensive ground ice glacial sediments.

Permafrost thickness ranges from 240 to 900 m in continental Antarctica, 15 to 180 m 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.

1.5 Observations for studies of environmental change in the Antarctic

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 availability of summer observations in some fields, such as when investigating oceanographic conditions using ship-board systems. It is still the case that few operating nations carry out year-round studies or even monitoring of biological systems, either marine or terrestrial.

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, with many of these still operating today. Year-round capability allowed the monitoring of many aspects of the Antarctic environment, such as meteorology, geospace, coastal sea ice conditions and sea level. With these instrumental records in many cases now being 50 years in length, these are very 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 in the interior. The advent of satellite remote sensing in the 1960s was therefore crucial for obtaining data in the data-sparse interior of the continent. Initially only visible and infra red imagery were available, but increasingly advanced instruments have been developed that have provided a rich source of data for studies in glaciology, geology and observing the surface of the ocean.

There has also been a progressively greater use of autonomous instruments in remote parts of the Antarctic. Automatic weather stations have been deployed in many isolated locations since the early 1980s, with the data used for weather forecasting and climate change studies. Such instrument deployments have sometimes been coupled with equipment to provide upper atmosphere and geospace observations, which is very operationally efficient. There have also been an increasing number of biological studies, including environmental monitoring at smaller, biologically relevant scales, although such datasets are typically of short duration. Generally there has been few attempts to link environmental observations across different scales of measurement. The ocean regions have long been a data void for sub-surface oceanographic observations, but new autonomous systems are providing frequent measurements and they 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 resolution coverage through the last few hundred years. Deeper ice cores, such as collected at Vostok and Dome C, have provided new insights into the glacial cycles over much of the last million years. Even earlier periods can be investigated with sediment cores from ocean areas, although the resolution here is coarser.

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

1.6 **Antarctic climate variability**

The Antarctic climate system varies on time scales from the sub-annual to the millennia and is closely coupled to other parts of the global climate system. On the longest time scales it has been found that the Antarctic ice sheet fluctuates on Milankovitch frequencies (20ka, 41ka, 100ka) in response to variations in the Earth’s orbit around the sun, which caused regular variations in the Earth’s climate.

Proxy data shows that since the Last Glacial Maximum (LGM) at about 21K before present (BP) there have been a number of climatic fluctuations across the continent. One of the most marked has been the Mid Holocene warm period or Hypsithermal, which is present in various records from Antarctica. There is evidence for a Hypsithermal in East Antarctica but dating uncertainties are still high in some areas.

The instrumental period in Antarctic is only about 50 years in length, and since proxy data shows oscillations on longer time scales than this, it 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.

Reliable weather charts for high southern latitudes are only available since the late 1970s, but these have revealed a great deal about the cycles in the atmospheric circulation of high southern latitudes. The most pronounced climate cycle (or mode of variability) over this period has been the Southern Hemisphere Annular Mode (SAM), which is a flip-flop of atmospheric mass (as measured by a barometer) between mid-latitudes and the Antarctic coastal region. Normally atmospheric pressures are quite low around the Antarctic coast, but increase as you move northwards into mid-latitudes. Over the last few decades there has been a marked drop in pressure around the Antarctic and an increase in mid-latitudes, so that the pressure gradient has increased across the Southern Ocean resulting in stronger surface winds. This has had implications for the distribution of sea ice and the oceanography.

One of the reasons that the SAM has changed in recent decades is the Antarctic ozone hole. Stratospheric ozone is an important constituent of the upper atmosphere above the Antarctic, but levels began to decline in the 1970s, following widespread releases of CFCs and Halons in 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 between 14 and 22 km altitude. Although the chemical reaction takes place in the stratosphere during the spring, the effects propagate downwards in the atmosphere, changing the surface layers during the summer and autumn. One effect of the ozone hole has been to accentuate the SAM signal giving stronger winds around the Antarctic during the summer and autumn.

The ozone hole is one example of extra-polar changes having a profound impact on the Antarctic environment, since the CFCs responsible for the ‘hole’ were emitted in the industrial areas, most of which are in the Northern Hemisphere. Most greenhouse gas emissions also come from such areas, yet have a profound effect on the radiation balance of the Antarctic atmosphere.

The Earth’s climate system, comprising the atmosphere, cryosphere and the oceans, exhibits a high degree of coupling between different parts of the system on a range of spatial and time scales. Although the Antarctic is far removed from where most of the 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 the Antarctic and the Southern Ocean. In addition, there is increasing evidence that signals can also be transmitted in the opposite direction from high to low latitudes.

Statistically significant high-low latitude linkages (or teleconnections as they are known) can take place via the atmosphere and ocean, however, 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 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) requires decades to centuries to respond. At the other extreme, fast wave propagation in the ocean has timescales of just a few days.

About 30 years of reliable atmospheric reanalysis fields are available for the high southern latitudes, so it has been possible to establish the nature of the broadscale Southern Hemisphere teleconnections. However, 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. The ice core records have shed some light on teleconnections over the century timescale. The short cores can 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. Through most of the Holocene there has been a several hundred year time lag between southern hemisphere and northern hemisphere events, but in recent decades the northern hemisphere signal of rising temperature since about 1800 AD has paralleled the southern hemisphere one as depicted by the oxygen isotope signal at Siple Dome. 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.

1.7 Terrestrial and marine biota of the Antarctic

Levels of terrestrial biodiversity in the Arctic are strikingly greater than those even of the sub-Antarctic and much more so than the maritime and continental Antarctic. This is the case even relative to the superficially environmentally extreme and isolated High Arctic Svalbard and Franz Josef archipelagos approaching 80°N. In comparison with about 900 species of vascular (higher) 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. However, despite the apparent ease of access to much of the Arctic, it is observed that a relatively low number of established alien vascular plants or invertebrates are known from locations such as Svalbard (Rønning 1996; Coulson in press), in comparison with the *c.* 200 species introduced to the sub-Antarctic by human activity over only the last two centuries or so (Frenot *et al.* 2005, 2007). 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 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 (Muerk *et al.* 1994; Mitchell *et al.* 1999; Fell 2002; Shaw 2005; Convey *et al.* 2006a). These plants present an unusual combination of morphological and life history characteristics (Convey *et al.* 2006a), 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).

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, spiders, beetles, flies and moths, with smaller representation of some other insect groups (Gressitt 1970; Convey 2007b). 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).

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

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.

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 1.2. 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 (in press), Maslen and Convey, in press. ND - number of representatives of group unknown; * - large changes likely with future research due to current lack of sampling coverage, expertise and/or synonymy.

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 1.3. The occurrence on alien non-indigenous species across Antarctic biogeographical zones (extracted from Frenot et al. (2005); see also Greenslade (in press) for a detailed description of established and transient alien species, and species recorded only synanthropically, from sub-Antarctic Macquarie Island).

	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
Monocotyledons	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 1.4. Species diversity in some of the main representative terrestrial groups of the sub-Antarctic biogeographical zone, in comparison with three Arctic locations. Data sources: Svalbard - Barr (1995), Elvebakk, & Hertel (1996), Frisvoll & Elvebakk (1996), Rønning (1996), Coulson & Resfeth (2004), Coulson (in press); Franz Josef Land - Barr (1995); Greenland - Jensen & Christensen (2003); sub-Antarctic - Convey (2007a).

Group	Sub-Antarctic	High Arctic (Svalbard)	High Arctic (Franz Josef Land)	Greenland
Rotifera	> 59	154		
Tardigrada	> 34	83		
Nematoda	> 22	111		
Platyhelminthes	4	10		
Annelida (Oligochaeta)	23	34		
Mollusca	3/4	0		2
Crustacea (non-marine)	44	33		65
Insecta	210	237		631
Collembola	> 30	60		41
Araneida	20	19		60
Acarina (free-living)	140	127		127
Myriapoda	3	0		1
Mammalia	0	3	2	8
Aves	0	17	6	39
Flowering plants	0	164	57	515
Bryophytes	26	373	150	612
Lichens	150	597	> 100	950

Chapter 2

Antarctic climate and environment history

Coordinating Lead Author:
Dominic A. Hodgson

Lead Authors:
Nerilie Abram, John Anderson, Peter Barrett, Mike Bentley, Peter Convey, Valerie Masson-Delmotte, Paul A. Mayewski, Robert Mulvaney, Regine Röthlisberger, Colin P. Summerhayes, Elie Verleyen, Wim Vyverman, Guido di Prisco and Cinzia Verde

Contributing Authors:
Mark Stevens, Steve Chown, Xavier Crosta, Guillaume Massé, Luca Bargelloni

Review Editors:

This chapter should be cited as:
Hodgson DA, Abram N, Anderson J, Barrett P, Bentley M, Convey P, Masson-Delmotte V, Mayewski PA, Mulvaney R, Röthlisberger R, Summerhayes CP, Verleyen E, Vyverman W, Stevens M, Chown S, Crosta X, Massé G (200?) Antarctic climate and environment history. In: Turner J, Convey P, di Prisco G, Mayewski PA, Hodgson DA, Fahrbach E, Bindschadler R (eds) Antarctic Climate Change and the Environment

2.1 Introduction

This chapter provides a review of Antarctic climate and environmental history from deep time through to the present. Modern climate over the Antarctic 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 to understanding the dynamics of the Earth's past, present, 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 both on the continent and elsewhere. Specifically 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. By knowing this natural variability we can accurately identify when present day changes exceed the natural state. We can also study, for example, past warm periods to determine the processes in the oceans, atmosphere, cryosphere and biosphere that caused them and the effects that they had on the environment at those times.

Although we have no analogue for a high CO_2 atmosphere during the past million years, there are periods in the geological record when atmospheric CO_2 concentrations ranged between 1000 and 3000 ppm, and we can analyse those records to see what sorts of environments prevailed then. We can also examine how Antarctica was affected by the decline in atmospheric CO_2 from the end of the Cretaceous Period (65 Ma BP), in what is called the 'greenhouse world' to the cold, low CO_2 world we now live in, which is called the 'ice-house world'. For these earlier times we have no samples of fossil air to draw on and must infer CO_2 concentration from proxy data. In contrast, for the past 850,000 years of the 2 million year long Quaternary Period, the relationship between CO_2 and atmospheric temperature can be analysed in detail using a range of chemical data along with bubbles of air trapped in ice cores.

In Antarctica, past climates and environments can be reconstructed from a range of palaeoenvironmental records, each of which provides different, yet complimentary, information on the patterns, processes and mechanisms of change. Instrumental data from satellites, ground-based instruments and oceanographic surveys span much of the last four decades. However, for records that pre-date this 'instrumental period' it is necessary to use geological records on land and the stratigraphic records contained in ice cores, marine sediments and lake sediments. Ice cores provide a continuous high-resolution record of environmental change in the interior of the continent and at some coastal locations. Here water molecules, particles and gas bubbles trapped in the ice provide valuable archives of change in atmospheric composition. Analyses of ratios of hydrogen and oxygen isotopes in water molecules, and of concentrations of methane, carbon dioxide and other gases in trapped bubbles have enabled scientists to infer past temperatures, precipitation and other environmental parameters. The longest ice core records date back to 850,000 years (EPICA 2004), and aerial geophysical surveys are continuing to find ice up to 1 Ma.

Geological records on land are generally not accessible in Antarctica due to the thick continental ice cover. However in a few places, like the Antarctic Peninsula, there are well-exposed shallow marine sedimentary strata that predate the first ice sheets. Seymour Island in particular has yielded a great variety of plant and animal fossils from Late Cretaceous and Early Cenozoic times (100-34 million years ago, Ma). In addition, deep drilling into marine sediments on the continental margin of Antarctica has provided records of Antarctica's climate during the tens of millions of years that it has been ice covered. On the margin, thousands of meters of sediment have accummulated as a result of glaciers and (in warmer times) rivers carrying gravel, sand and mud from the interior to the coast. That process of accumulation has been periodically disrupted by ice sheet oscillations, the ice margin expanding at times as far as the edge of the continental shelf, and then contracting again to leave behind a layer of compact glacial debris or till. During interglacial times when the ice front retreated, fine-grained mud accumulated. Coastal sediments can also record changing water depth as sea level rises and falls, because waves winnow out the mud when the water is shallow but let it settle when the water is deep water. Until now, deep drilling has been restricted to only a few locations (e.g, ocean drilling in Prydz Bay, or drilling from platforms on ice in the Ross Sea). In contrast, the collection of relatively short piston cores and vibracores of sediments from the seas around Antarctica has been quite widespread. These cores, typically not more than 10 metres long, but on occasion up to 40 metres, 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. 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 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. On land, the drilling of sediments that accumulated in lakes in the ice-free regions of Antarctica provides geological records of climate that typically span the Holocene and occasionally beyond. 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. 2004). Glacial geomorphological records such as moraines and trimlines, coupled to cosmogenic isotope dating of exposed rocks, also provide information on past ice sheet extent and thickness, and the timing of glaciation and deglaciation.

Using these records, this chapter describes the geological setting of Antarctica and its evolution into the glaciated 'ice-house' continent, moves on to focus on the transition between the Last Glacial Maximum (LGM, c. 21 ka BP) and the present Holocene interglacial period, which began about 11.7 ka BP. It then describes the development of Holocene conditions and natural variability right up to the present day. Finally, it briefly considers the impact of these environmental changes upon the terrestrial and marine biospheres.

2.2 **Deep Time**

2.2.1

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. The message from the palaeorecords is that change is normal and the unexpected can happen (Bradley et al. 2002). 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.

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. As the Gondwanan fragments separated by sea-floor spreading between around 100 to 65 Ma, during the Cretaceous Period, Antarctica moved into a position over the South Pole (Fig. 2.1). A rich record of plant and animal fossils from Late Cretaceous and Early Tertiary times has been found on Seymour Island (Francis et al. 2008). Up until the formation of a major ice sheet in the Oligocene, the terrestrial fauna and flora of Antarctica seem to have remained typical of south-temperate rainforest (Francis & Poole 2002). Over time, South America, Africa, India, Australia and New Zealand moved away from Antarctica, opening the South Atlantic, Indian and Southern Oceans (Fig 2.2). The western margin of Antarctica, which extended from South America through the Antarctic Peninsula to Marie Byrd Land and New Zealand behaved differently from the rest. As it rode over the proto-Pacific ocean floor, which was subducted at a trench formerly situated along the western margin of the Antarctic Peninsula, it developed into an active margin characterized by volcanism and the lateral movement of fragments of formerly continuous Gondwana terrain (see for example McCarron and Larter (1998)). This western arc broke apart first at around 85 Ma south of New Zealand, and then at around 41 Ma between South America and the Antarctic Peninsula to create the Drake Passage (Scher & Martin 2006) (Fig 2.2). There is some dispute about the latter date; some thinking the Drake Passage may not have opened until 20 Ma (e.g. see Siegert et al, (2008)). If the earlier date is accepted, then the last Gondwanan fragment to leave Antarctica was Australia, with the deep water gap between Tasmania and Antarctica opening at around 31 Ma (Wei 2004).

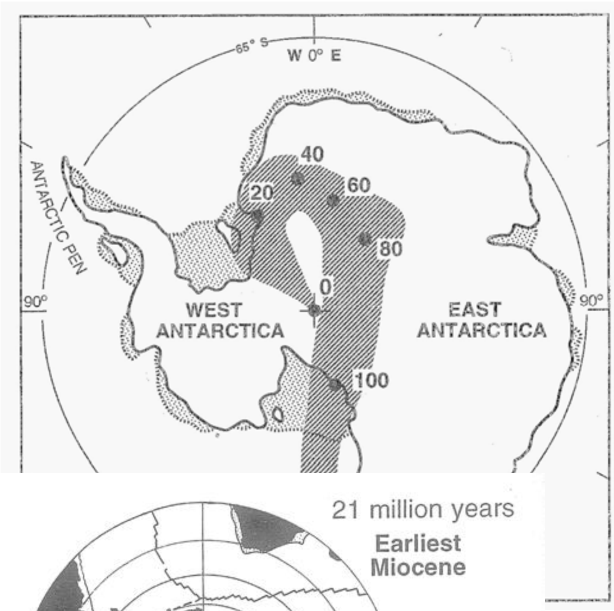


Figure 2.1. Apparent polar wander path for East Antarctica over the last 120 Ma (modified from DiVenere et al., (1994)). The shaded area represents the A_{63} error envelope.

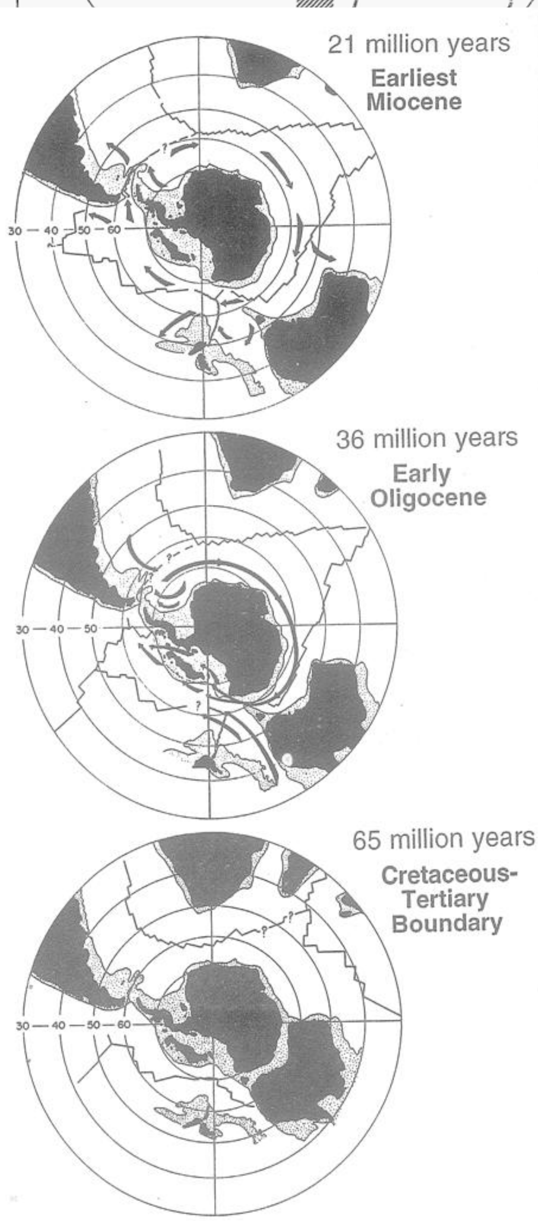


Figure 2.2. Three maps showing the progressive separation of Antarctica and the other Southern Hemisphere continents, leading to the opening of "ocean gateways" that allowed the development of the Antarctic Circum-polar Current from some time soon after 34 Ma. Modified from Kennett (1978).

Antarctica's thermal isolation by the atmospheric West Wind Drift and the oceanic Antarctic Circumpolar Current could not take place until the break between Tasmania and Antarctica widened and deepened, well after 31 Ma. Yet the evidence from changing sea levels through time suggests that there may have been small ice caps on Antarctica even back into the Cretaceous. Levels of the greenhouse gas CO_2 in the atmosphere ranged from roughly 3000 ppm in the Early Cretaceous (at 130 Ma) to around 1000 ppm in the Late Cretaceous (at 70 Ma) and Early Cenozoic (at 45 Ma) (e.g. Berner (1990); Andrews et al., (1995)), leading to global temperatures 6 or 7ºC warmer than present. These high CO_2 levels were products of the Earth's biogeochemical cycles, including the weathering of silicate minerals, sedimentary burial of organic matter, and emanations from volcanic processes. In the Cenozoic these temperatures gradually peaked around 50 Ma (Fig. 2.3 and Crowley and Kim, (1995)), with little or no ice on land, and subsequently declined in parallel with CO_2 (Fig. 2.3; and Berner (1990); see also Fig. 6.1 in Jansen et al, (2007)). Superimposed on the high CO_2 world of the Early Cenozoic, deep-sea sediments provide evidence of the catastrophic release of more than 2000 gigatonnes of carbon into the atmosphere from methane hydrates around 55 Ma ago, at the Paleocene/Eocene boundary, raising global temperatures a further ~4-5°C, though they recovered after about 100,000 years (Fig. 2.3 and Table 2.1) (Zachos et al. 2003, Zachos et al. 2005).

The first continental-scale ice sheets formed on Antarctica in the Oligocene Epoch, around 34 million years ago (Zachos et al. 1992), and prior to the Tasman break-up. The development of the Oligocene ice sheet appears to have been a consequence of a decline in atmospheric CO_2 levels (DeConto & Pollard 2003, Pagani et al. 2005, Siegert et al. 2008) caused by reduced CO_2 outgassing from ocean ridges, volcanoes and metamorphic belts and increased carbon burial (Pearson & Palmer 2000) which dropped global temperatures at that time to levels around 4ºC higher than today (Fig. 2.3). The Oligocene ice sheets reached the edge of the Antarctic continent, but were most likely warmer and thinner than today's.

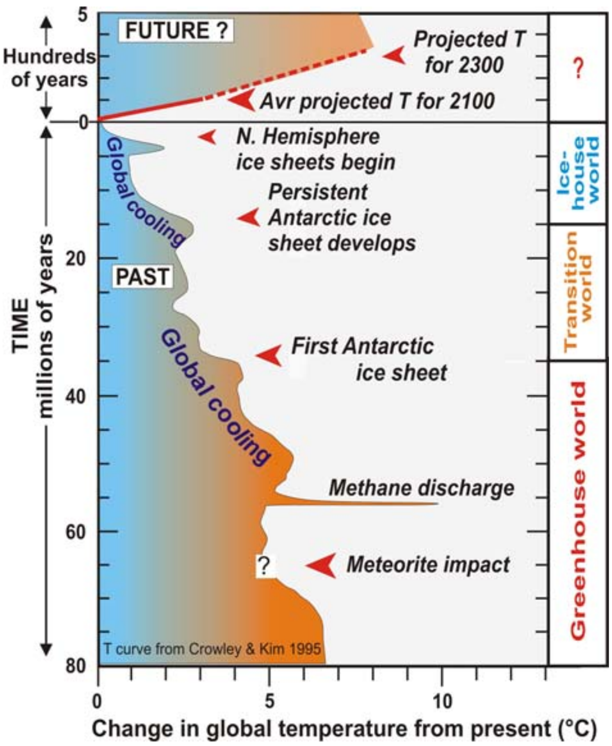


Figure 2.3. Change in average global temperature over the last 80 Ma, plus future rise in temperature to be expected from energy use projections and showing the Earth warming back into the 'greenhouse world' typical of earlier times more than 34 million years ago (Barrett 2006). The temperature curve of Crowley and Kim (Crowley & Kim 1995) is modified to show the effect of the methane discharge at 56 Ma (Zachos et al. 2003, Zachos et al. 2005) and the early Pliocene warming (Ravelo et al. 2004) (from AGCS, in press, (2008)).

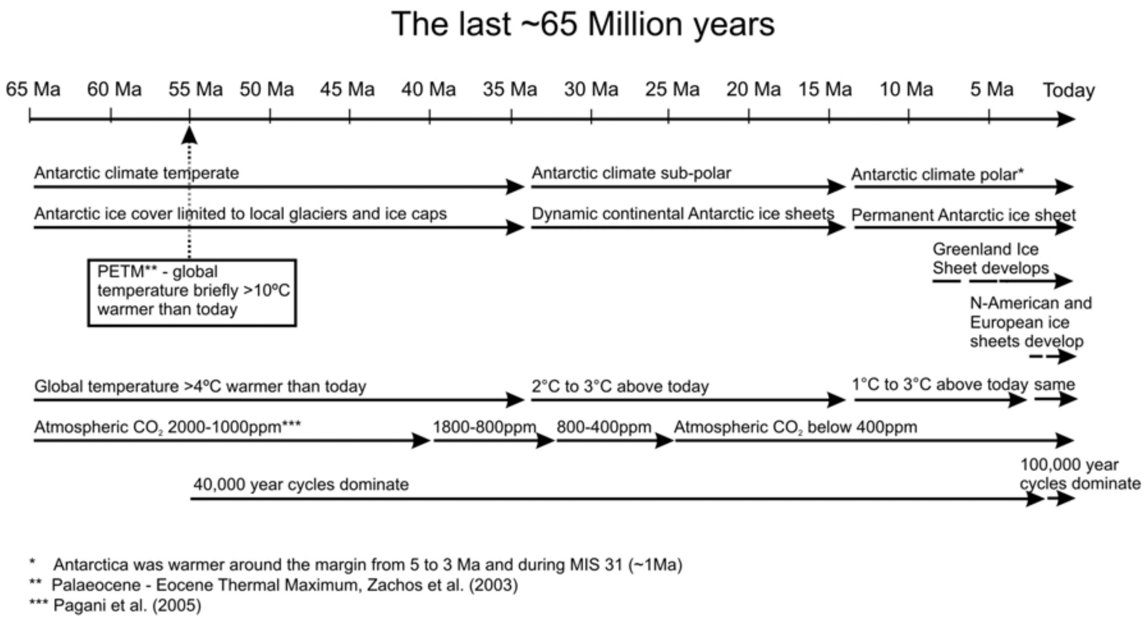


Table 2.1: **Main climatic events of the last 65 million years; the Antarctic context**

The fossil record suggests that cool temperate and tundra communities survived the formation of the major ice sheet in the Oligocene, persisting until the Middle Miocene; whether they persisted later in some places is a matter of debate (Convey et al. 2008). Further sharp cooling in the Miocene Epoch, at around 14 Ma (Fig 2.3), was probably caused by the growing thermal isolation of Antarctica and related intensification of the Antarctic Circumpolar Current rather than by any change in CO_2 (Shevenell et al. 1996). It thickened the ice sheet to more or less its modern configuration (Flower & Kennett 1994), which is thought to have persisted through the early Pliocene warming from 5 Ma to 3 Ma (Kennett & Hodell 1993, Barrett 1996) and into the Pleistocene at 1.8 Ma.

During the Pliocene, mean global temperatures were 2-3ºC above pre-industrial values (Haywood et al. 2000), CO_2 values may have reached 400 ppm (see Jansen et al, (2007) and references therein), and sea levels were 15-25m above modern levels (Jansen et al, (2007) and references therein). Pliocene temperatures several degrees warmer than today around the Antarctic margin are implied by coastal sediments (Harwood et al. 2000) and offshore cores (Whitehead et al. 2005), and from diatom ooze from this period recently cored from beneath glacial sediments under the McMurdo Ice Shelf (Naish et al. 2007). The occurrence of the ooze indicates that the core site in the Ross Embayment was essentially free of ice at this time.

Global cooling from around 3 Ma onwards (Ravelo et al. 2004) (Fig. 2.3) led to the first ice sheets on North America and NW Europe around 2.5 Ma (Shackleton et al. 1984). As noted by the Scientific Committee for Antarctic Research science programme, Antarctica and the Global Climate System (AGCS 2008), these ice sheets enhanced the Earth's climate response to orbital forcing, taking us to the Earth's present "ice house" state (Fig. 2.3), which for the last million years (Table 2.2) has been alternating between (i) long (40-100,000 years) glacial cycles, when much of the Northern Hemisphere was ice-covered, global average temperature was around 5ºC 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.

Marine continental drill cores show the cyclic expansion and contraction of the Antarctic ice sheet from its inception at 34 Ma almost to the middle Miocene transition, with sea level varying on a scale of tens of metres (Naish et al. 2001, Barrett 2007, Siegert et al. 2008). The drilling shows that the Antarctic ice sheet was dynamic, fluctuating on Milankovitch frequencies (20 ka, 41 ka, 100 ka) in response to variations in the Earth's orbit around the sun, which caused regular variations in the Earth's climate. However, the Antarctic ice sheet is not simply responsive to external forcing. By virtue of its size and the amount of water it contains it has become one of the major drivers of Earth's climate and sea level change. Fluctuations in the ice sheet lead to significant variations in global sea level.

2.2.2 **Marine Deep time**

The polar habitats, and the respective ichthyofaunas

The two polar regions are more dissimilar than similar. Geography, oceanography and biology of species inhabiting the polar regions have often been compared (e.g. Dayton et al. 1994) to outline the differences between the two ecosystems. The main differences between the Arctic and the Antarctic regions are the greater age and isolation of the latter. The Antarctic has been cold longer, with ice sheet development preceding that in the Arctic by at least 10 million years.

Throughout the Cenozoic, regional tectonic and oceanographic events played a key role in delimiting the two polar ecosystems and influencing the evolution of their faunas, whose composition and diversity are strongly linked to geological history. The modern polar ichthyofaunas differ from each other in age, endemism, taxonomy, zoogeographic distinctiveness, range of physiological tolerance to various environmental parameters, and biodiversity (Eastman 1997). During the fragmentation of Gondwana, Antarctica played a key role in altering ocean circulation and forcing climate toward cooling. With the opening of the Drake Passage 41-20 Ma, separation of Antarctica from South America was complete. This event allowed the establishment of the Circum-Antarctic Current and of the Polar Front, a roughly circular oceanic system running between 50°S and 60°S. Just north of the Polar Front, the surface water temperature has an abrupt rise of ca. 3°C. Because of this, the Polar Front acts as a barrier for migration to both directions, causing adaptive evolution to develop in isolation.

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. In the Arctic isolation is less stringent, and the range of temperature variation is wider, both in the ocean and on the surrounding lands, thus facilitating migration and redistribution of the fauna. The colonised terrestrial portions are extensive; they are directly linked to temperate areas, producing wide and complex terrestrial mechanisms of feedback to the climate, which add to those originating from ocean and atmosphere circulation. The anthropogenic impact on the environment is also far greater.

The differences in the two polar environments correspond to different impacts of historical and current climate changes. Hence, comparative studies on the ecosystems provide powerful evolutionary insights into the relationship between environment and evolutionary adaptation.

A detailed assessment of the impacts of climate change in the Arctic has been published (ACIA 2005). No similar report for the Antarctic has so far been available.

Fishes of the southern ocean - Notothenioidei

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.

Over the past 30-40 million years, in parallel with the diversification of the suborder, the physico-chemical features of the Antarctic marine environment have 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). With the local extinction of most of the temperate Tertiary fish fauna as the southern ocean cooled, the suborder experienced extensive radiation, dating from the late Eocene approx. 24 mya (Near 2004), that enabled it to exploit the diverse habitats provided by a now progressively freezing marine environment.

As conditions of extreme temperature developed, fish evolved physiological and biochemical mechanisms of adaptation to survive in the cold. Antarctic notothenioids are highly 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 (Somero 2005). Thus, the question to what extent Antarctic fish may adapt to environmental change becomes a very important issue.

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 of the ancestral stock. Similar to typical vertebrates, seven families have hemoglobin (Hb)-containing erythrocytes circulating in the blood, whereas Channichthyidae (the crown group) are devoid of Hb (Ruud 1954; Cocca et al. 1995; Zhao et al. 1998). Three families, i.e. 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 corresponding to New Zealand, Australia and high-latitude South America. As temperatures decreased and ice appeared, Antarctic notothenioids acquired antifreeze glycoproteins (AFGPs), an adaptation that allows them to survive and diversify in ice-laden seawater that reaches nearly -2°C (DeVries & Cheng 2005). 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 AFGP gene sequences in their genomes (Cheng et al. 2003).

The Antarctic fish fauna includes 322 species grouped into 50 families (Eastman 2005), whereas the Arctic fauna includes 416 species (58 are freshwater) grouped into 96 families (Andriashev & Chernova 1995). In Antarctica, endemism of the benthic fauna is 88% and rises to 97% when only the dominant suborder Notothenioidei is considered. In the Arctic, endemism for marine fish is 20-25%. The Arctic fauna has no endemic higher taxonomic category equivalent to the Antarctic notothenioids, and there has been no comparable adaptive radiation of any fish group. Arctic fish are generally more eurythermal and euryhaline (i.e. they can tolerate relatively large changes in ambient salinity) than their Antarctic counterparts, and the North Atlantic and North Pacific character of the marine fauna reflects the continuity of shelf areas between the Arctic and boreal regions.

Specialisations, cold adaptation and evolution

Two study cases of adaptive specialisations will be briefly mentioned here: globin genes and antifreezes. Polar fish are the only vertebrates endowed with these two adaptations. 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.

Globin genes

Specialised hematological features are among the most striking adaptations developed by the Antarctic ichthyofauna during evolution. The dominant suborder Notothenioidei is the best characterised group of fish in the world, from this and other viewpoints. Seven of the eight notothenioid families have Hb (usually a single major component, Hb 1), but in lesser amount and multiplicity than other fish. Work is currently in progress on the structure/function of Hbs from bony and cartilaginous Antarctic, sub-Antarctic, Arctic and temperate fish.

The coastal waters of Antarctica are cold and oxygen-rich. The metabolic demand of polar fish for oxygen is relatively low, the solubility of oxygen in their plasma is high, but the energetic cost associated with circulation of a highly corpuscular blood fluid is large. With the selective pressure for erythrocytes and oxygen carriers relaxed and with the cells posing a rheological disadvantage in the temperature-driven increase in blood viscosity, notothenioids have evolved reduced hematocrits, Hb concentration/multiplicity and oxygen affinity. Oxygen binding is generally favoured at low temperature.

A unique evolutionary specialisation in adult vertebrates, first reported by (Ruud (1954)), is typical of the 16 “icefish” species of the notothenioid family Channichthyidae (Fig 2.4), namely their colourless blood is devoid of Hb. Icefishes maintain normal metabolic function by delivering the oxygen physically dissolved in the blood to tissues. Reduction of the hematocrit to near zero appears advantageous because it diminishes the energetic cost associated with circulation of a highly viscous, corpuscular blood fluid (Wells et al. 1990; di Prisco et al. 1991; Eastman 1993).

The globin-gene status in notothenioids has been characterised, and potential evolutionary mechanisms leading to the Hb-less phenotype have been evaluated (Cocca et al. 1995; Zhao et al. 1998; di Prisco et al. 2002). Expression of adult globins was abrogated by a single, large-scale deletional event in the ancestral channichthyid, that removed almost the entire notothenioid globin gene complex with the exception of the 3’ end of the α-globin gene of major Hb 1. These inactive remnants, no longer under positive selection pressure for expression, subsequently experienced random mutational drift, without, as yet, complete loss of sequence information. These genetic remnants should prove useful as tools for development of molecular phylogeny of icefishes and for calibration of a vertebrate mutational clock free of selective constraints.

Similar to channichthyids, red-blooded Antarctic fishes can carry the necessary oxygen dissolved in the plasma (di Prisco 2000). Resting icefishes, as well as red-blooded fishes, do not seem disadvantaged by their alternative oxygen-transport strategy; nevertheless they are vulnerable to stress. Antarctic fish may lack the heat-shock response (Hofmann et al. 2005). This would limit the ability to respond to stress factors, such as additional thermal challenges.

Antifreeze compounds in Antarctic and Arctic fish

The evolution of AFGPs in Antarctic and Arctic fish is a classical example of adaptation developed independently at both poles. To avoid death by freezing at sub-zero temperatures, fish in these environments have evolved antifreeze molecules secreted at high concentrations into their body fluids. The evolution of AFGPs meets the criteria for a "key innovation" (Eastman 2000). Because of the constant freezing temperatures, Antarctic notothenioids synthesise AFGPs constitutively, whereas in Arctic fish AFGPs exhibit seasonal patterns of biosynthesis. This is an efficient energy-saving strategy, that avoids costly biosynthesis when freezing is not a danger. At the same time, it suggests that even a small increase in environmental temperatures will not pass unnoticed in biosynthesis control.

Studies have addressed the the evolution of AFGP genes (Cheng 1998). In Antarctic notothenioids the AFGP gene evolved from a functionally unrelated pancreatic trypsinogen-like serine-protease gene. Analysis of AFGP in the Arctic cod (*Boreogadus saida*) showed that the genome of this species (phylogenetically unrelated to notothenioids) contains genes which encode nearly identical proteins. This would suggest a common ancestry. However the genes of the two fish groups have not followed the same evolutionary pathway. Assuming an endogenous, yet unknown genetic origin, the cod AFGP genes have evolved from a different, and certainly not trypsinogen-like, genomic locus. This is one of the most powerful examples of convergent evolution at the molecular level yet established.

Figure 2.4. An icefish: *Champsocephalus esox*

2.3 **The last Million years**

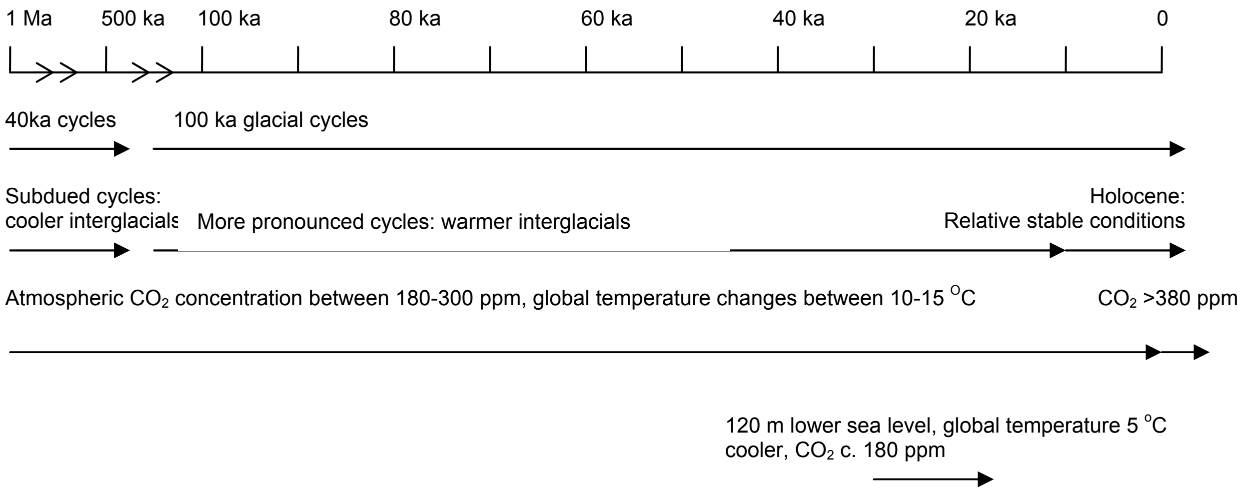


Table 2.2. Main climatic events of the last 1 million years; the Antarctic context

Ice cores spanning up to the past 800,000 years 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 (Fig. 2.5). Unlike today, the CO_2 signal follows the temperature signal (EPICA 2004), signifying that changes of CO_2, as well as of 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) use an elegant carbon cycle box model to provide a convincing explanation for the variation in the CO_2 signal and its relation to temperature in these cores.

The Antarctic ice core data show that the Earth's climate has oscillated through eight glacial cycles over the last 800,000 years, with CO_2 and mean temperature values ranging from 180 ppm and 10ºC in glacials, to 300 ppm and 15ºC in interglacials (Severinghaus et al. 1998). The pattern has changed through time, with a fundamental reorganisation of the climate system at 900-600ka from a world that for the preceding 33 million years had been dominated by 41ka oscillations in polar ice volume, to a 100ka climatic beat (EPICA 2006), most likely in response to the increase in orbital eccentricity with time. The effect on global sea level was profound, with sea level dropping by 120m on average during glacial periods (e.g. Fig. 6.8 in Jansen et al, (2007)).

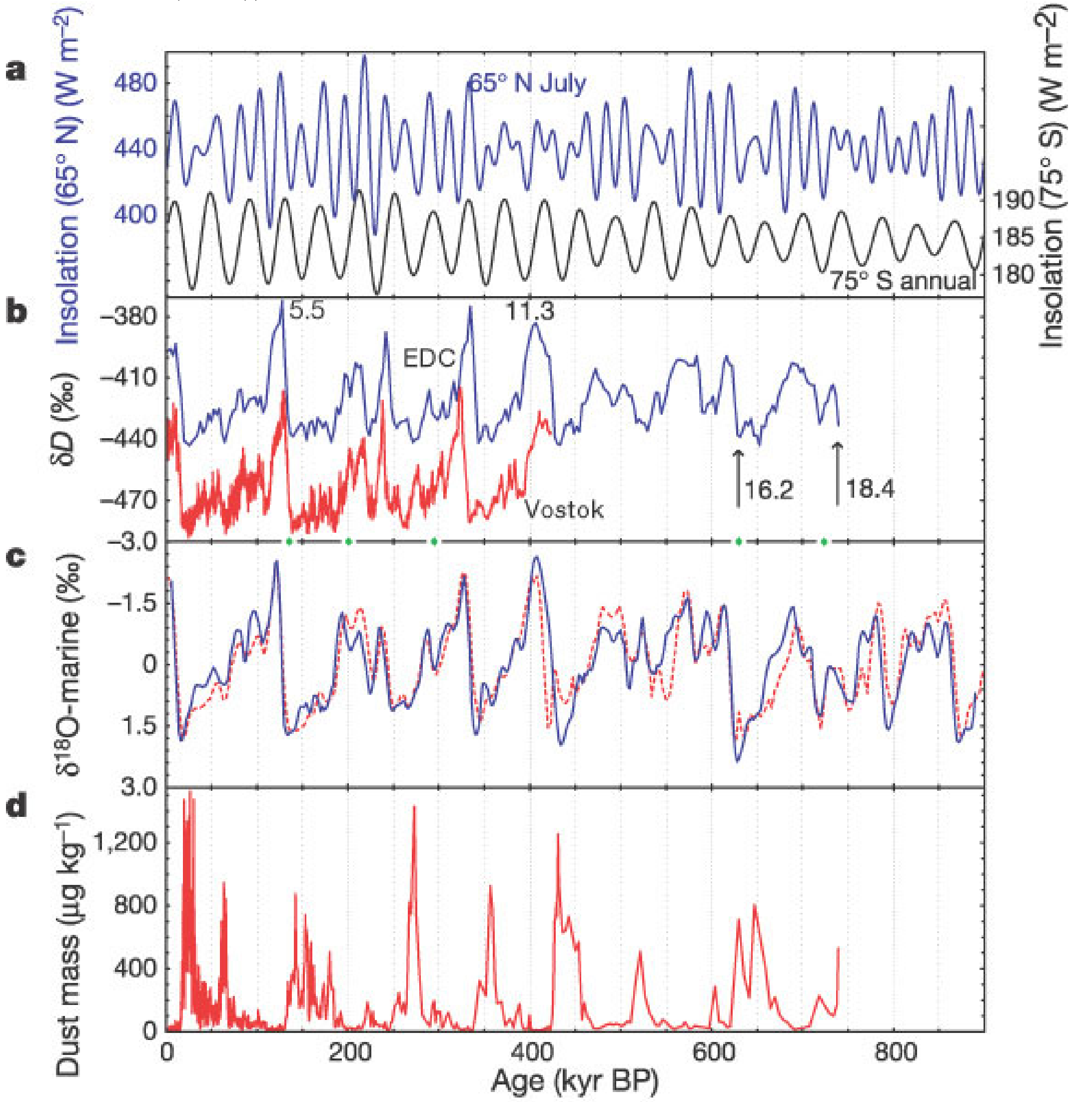


Figure 2.5. a, Insolation records[4]. Upper blue curve (left axis), mid-July insolation at 65° N; lower black curve (right axis), annual mean insolation at 75°S, the latitude of Dome C. **b**, δ*D* from EPICA Dome C (3,000-yr averages). Vostok δ*D* (red) is shown for comparison[1] and some MIS stage 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 site MD900963 and ODP677[3]; 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 r but consisting only of ODP site 677 for the earlier period[2]. Both records have been normalized to their long-term average. **d**, Dust from EPICA Dome C (EPICA 2004).

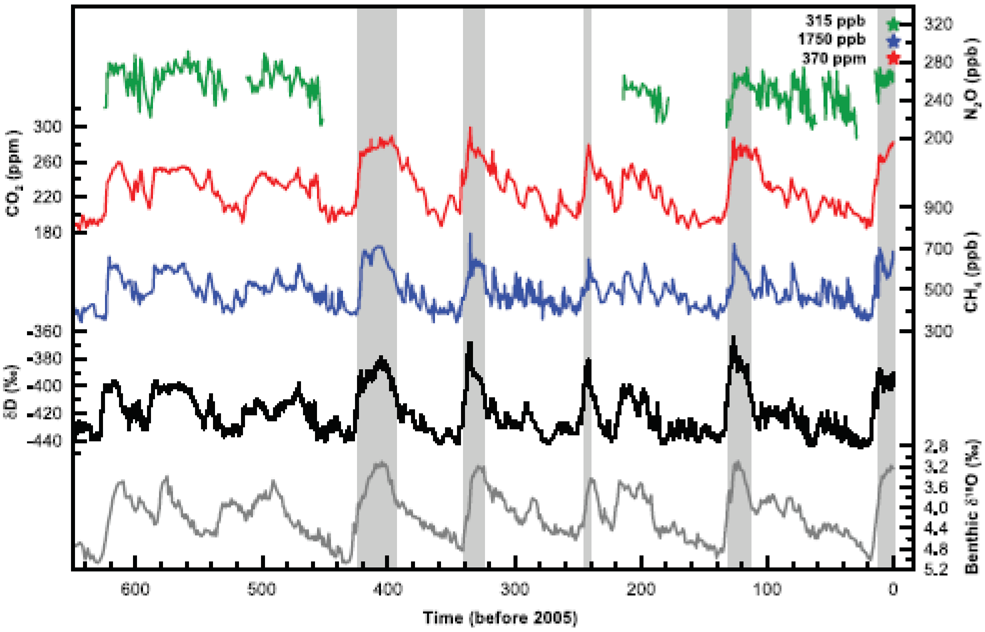


Figure 2.6. Variations of deuterium (δD; black), a proxy for local temperature, and the atmospheric concentrations of the greenhouse gases CO2 (red), CH4 (blue), and nitrous oxide (N2O; green) derived from air trapped within ice cores from Antarctica and from recent atmospheric measurements (Petit et al. 1999b, 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 r. The stack of 57 globally distributed benthic δ18O marine records (dark grey), a proxy for global ice volume fluctuations (Lisiecki & Raymo 2005), is displayed for comparison with the ice core data. Downward trends in the benthic δ18O 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 & Raymo 2005). The stars and labels indicate atmospheric concentrations at year 2000. (From IPCC)

Ice cores from both Antarctica and Greenland show that temperatures were between 2-5ºC higher than today in recent past interglacials (Jansen et al. 2007). This is matched by the lake sediment record (Hodgson et al. 2006c). 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 15-25m 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 30,000 years – see Jansen et al, (2007)). Given that modern CO_2 levels of around 380 ppm are higher than at any time in the last 850,000 years (EPICA 2004) (Fig. 2.6 and Table 2.2), 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 knowing 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 (Fig. 2.7, and EPICA, (2006), but exhibit a see-saw relationship. Antarctic warm events correlate with but precede those in Greenland, and vice versa (Fig. 2.7). Warming in the Antarctic begins when Greenland is at its coldest during the so-called Dansgaard/Oeschger (D/O) events (Fig.2.7). One theory 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 (Fig. 2.7). 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 & Wright 1991, Knorr & 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 in section 2.8.

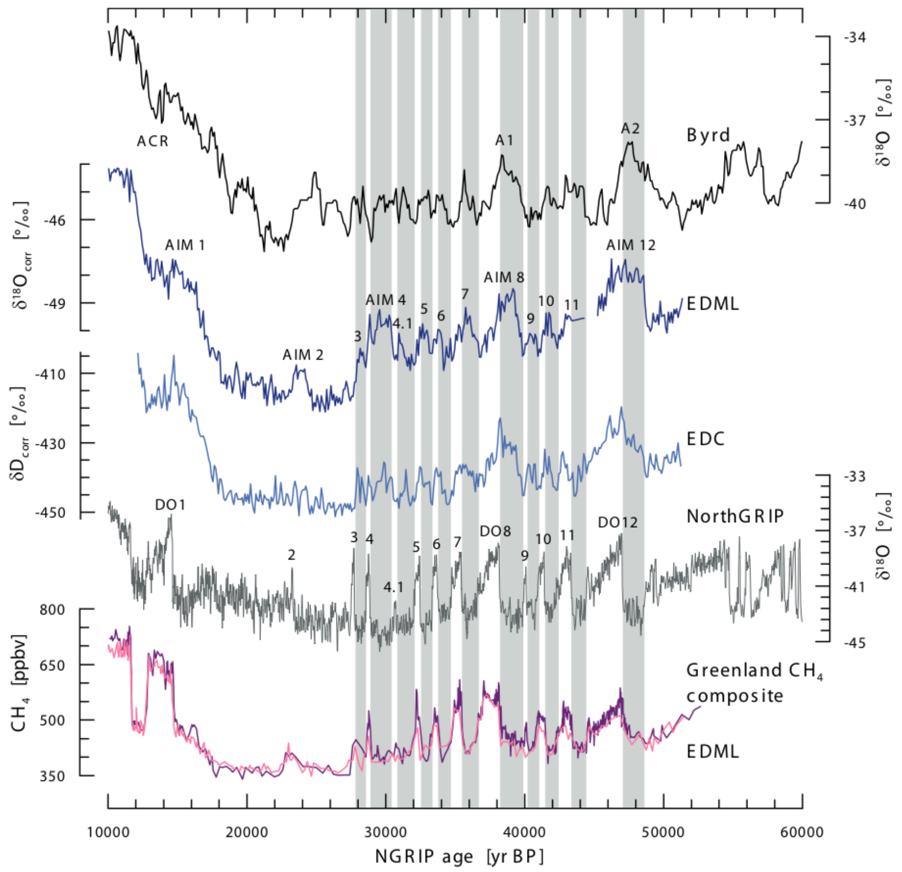


Figure 2.7 Methane (CH_4) synchronization of the ice core records of $\delta^{18}O$ as a proxy for temperature reveals one-to-one association of Antarctic warming (AIM) events with corresponding Greenland cold (stadial) events (D/O) covering the period 10, 000-60,000 years ago. EDML = EPICA core from Dronning Maud Land Antarctica, Byrd = core from West Antarctica; EDC = core from East Antarctica; NGRIP = core from North Greenland. Gray bars refer to Greenland stadial periods. Figure modified from EPICA (2006) by H. Fischer (from AGCS, in press (2008)).

While the changes from glacial to interglacial states (Fig. 2.6) 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 (Fig. 2.7). Equally, millennial to centennial scale variability observed within the past 12,000 years 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 most likely lie in poorly understood aspects of the global ocean circulation. 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 10ºC within a very few decades (Fig. 2.7); comparable changes in Antarctica were much slower.

2.4 Sea Ice at the Last Glacial Maximum

Sea ice is a crucial part of the Earth’s climate system. 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)).

Diatom assemblages from 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). Given enough cores it is possible to map the position of the sea ice edge back through time and to determine where it was in summer and in winter. This has been done for the Southern Ocean at the LGM by Gersonde et al, (2005) see Fig. 2.8. 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 (CLIMAP 1981).

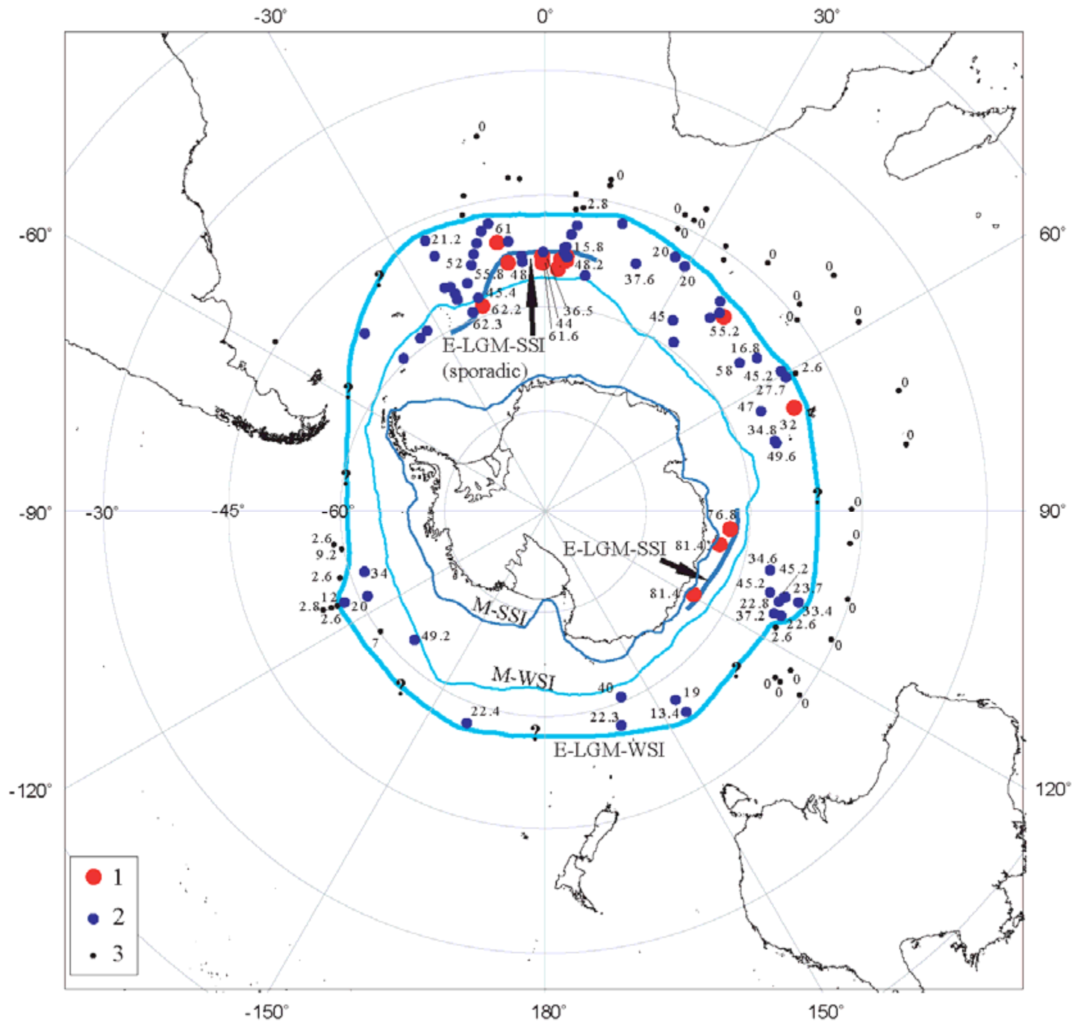


Fig. 2.8. Sea ice distribution at the Southern Ocean EPILOG-LGM (E-LGM) time slice. E-LGM-winter sea ice (E-LGM-WSI) indicates maximum extent of winter sea ice (September concentration >15%). Modern winter sea ice (M-WSI) shows extent of >15% September sea ice concentration according to Comiso (2003). 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 *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)

According to Gersonde et al., (2005) the LGM sea ice edge in the Atlantic and Indian sectors reached close to 47ºS (Fig 2.8), 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 is 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 4º, 5–10º, and 2–3º in latitude, respectively, compared to their present location. In the Atlantic and Indian sector, the Subantarctic

Front would have shifted by around 4–5º and 4–10º in latitude, respectively. The Subtropical Front displacement would have been minor, by around 2–3º and 5º 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.

Given records of high enough resolution, fluctuations in the position of the sea ice edge through Late Quaternary and Holocene time can be ascertained for comparison with geological and ice core data over about the past 220,000 years (Crosta et al. 2004) (Fig 2.).

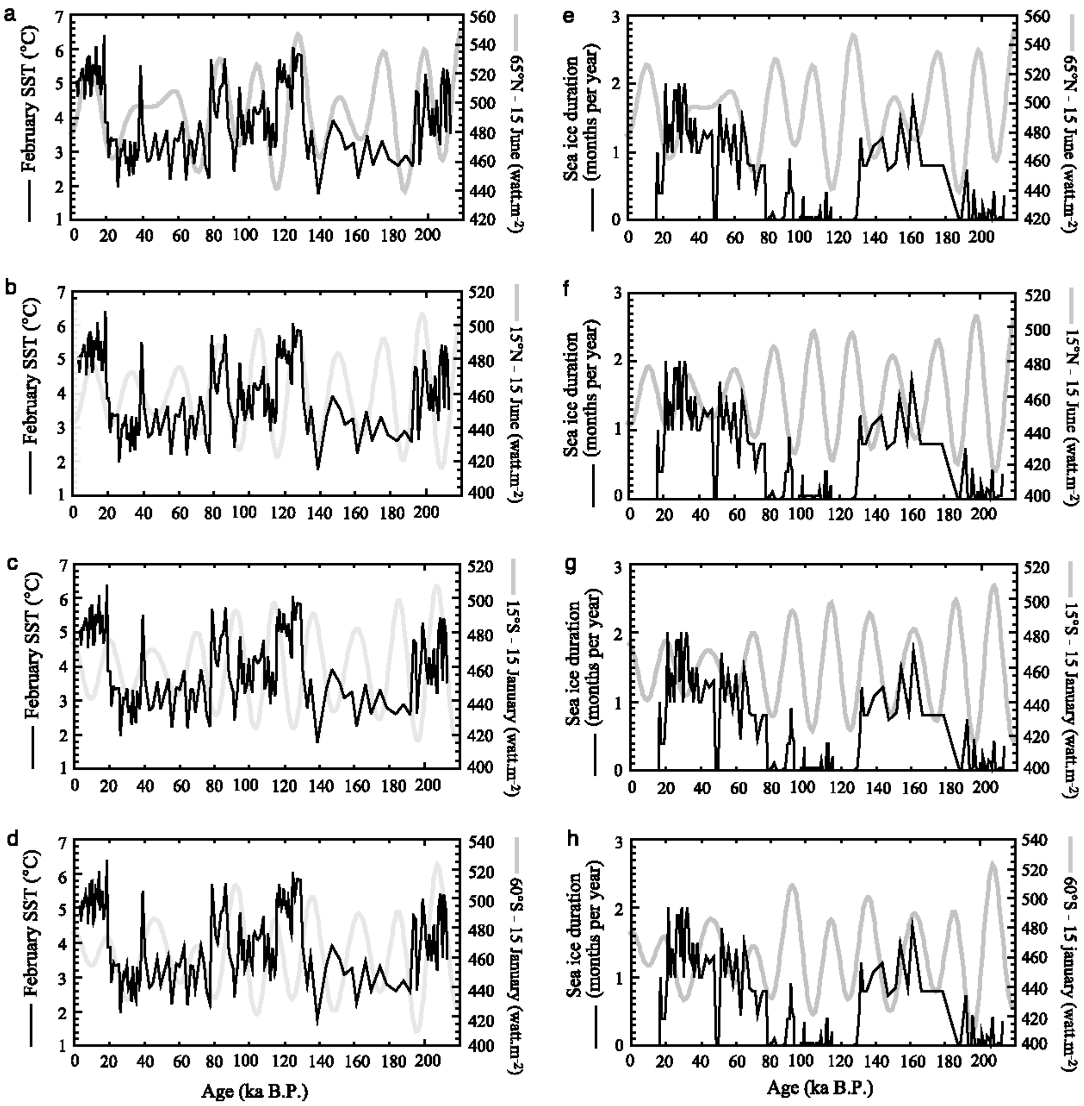


Fig. 2.9. Comparison of February SSTs and sea ice duration in core SO136-111 with insolation curves. Parameters estimated by Modern Analog Technique 5201/31 are represented by black lines; insolation curves are represented by grey lines. (a) SSTs versus insolation at 65°N for the 15th of June, (b) SSTs 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).

Microfossils have another attribute. It so happens that assemblages of planktonic organisms exhibit a relationship to the temperature of the surface waters in which they live that can be expressed mathematically as a so-called transfer function. Knowing this relationship at the present day, we can use it to transform assemblage data from the recent past into estimates of seawater temperature (Crosta et al. 2004). The validity of the data can be ascertained by comparison with oxygen isotope records were these are available. The data indicate that for cores well offshore, sea ice formation lags temperature decline at the onset of glaciations by about 1000 years (probably reflecting the time for Southern Ocean temperatures to become cold enough for sea ice to form), but that warming and sea ice retreat are simultaneous during glacial terminations (the transitions from glacial to interglacial) (Crosta et al. 2004).

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

The transition (Termination I) from the Last Glacial Maximum (beginning about 21 ka BP) to the present interglacial period was the last major global climate change event. Profiles across Termination I in Greenland and Antarctica are depicted in some detail in Fig. 2.10. The locations in Antarctica where long ice cores and other core data have been obtained are given in Fig. 2.11

The main glacial to interglacial changes of the Pleistocene period, like those depicted in Fig 2.6, appear to be driven largely by orbital forcing, and in particular by the insolation of the northern hemisphere represented in Fig 2.10 by the curve of insolation at 60ºN in June and in Fig. 2.5.

One of the first things to note is that northern and southern hemisphere temperatures respond differently to the forcing – they are out of phase, as we discuss below. Early indications of the system's response to forcing at high latitudes come from far away, being seen in signals of the character of deep water in the tropics. For instance a comparison of isotopic data from benthic and near-surface zooplankton in sediment from a tropical Pacific Ocean core (Stott et al. 2007) suggests that during the last deglaciation increasing insolation at high southern latitudes caused a retreat of sea ice and led to warming of the southern high-latitude surface waters from which deep water was derived around Antarctica. This warm deep water arrived in the western tropical Pacific about 1500 years before the Pacific surface waters warmed (Stott et al. 2007) and preceded the global rise in atmospheric CO_2.

Looking at the ice core record over the last transition in Antarctica (Fig 2.10) we find that warming starts first, at about 19000 ka BP, followed after a few hundred years by an increase in atmospheric CO_2 concentration (Monnin et al. 2001). Much the same applies to another greenhouse gas, Methane (CH_4)(Fig 2.10), which also lags the initial rise in Antarctic temperature. Comparing CH_4 with temperature in the Greenland record we see a good match; but there is no such match in Antarctica. As we see in more detail later, the sequence of events as recorded by temperature differs from north to south, the difference being a function of oceanic and atmospheric circulation, including see-saw effects related to changes in the thermohaline circulation, and biogeochemical cycle feedbacks.

The precise role of insolation in determining the onset of terminations remains uncertain (Schulz & Zeebe 2006). Similarly, and despite the strong correlation between CO_2 and Antarctic temperature seen in Fig 2.10, the mechanisms governing glacial-interglacial changes in greenhouse gas concentrations and their lags in relation to Antarctic temperatures are still a challenge for modellers (Kohler et al. 2005), although recent developments appear to provide an elegant solution (Köhler & Fischer 2006).

As shown in Fig 2.10, the northern hemisphere experienced more dramatic changes than the southern hemisphere during the deglaciation, starting with an abrupt warming (the Bølling-Allerød) at 14.7 ka BP followed by a return to colder conditions (the Younger Dryas episode) and a final rapid warming leading to the Holocene at 11.7 ka BP (Rasmussen et al. 2006).

Because Termination I is the most recent transition between a glacial and an interglacial period, the ice core record is less compressed than in older ice, giving us higher resolution data and a more robust age control (EPICA 2004, Rasmussen et al. 2006). Ice core records can be synchronised both regionally, using aerosol tracers like dust and calcium from continental aerosols, or sulphate from volcanism) (Severinghaus et al. 1998), and globally, using well-mixed atmospheric gas records such as methane or oxygen isotopes extracted from bubbles in the ice (Blunier et al. 1997, Morgan et al. 2002, Blunier et al. 2007). The accuracy of the Greenland ice core's layer-counted age scale (named 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 to synchronise the northern and southern hemisphere records, is estimated to be at most 250 years for the Younger Dryas period (an abrupt cooling interrupting the northern hemisphere deglaciation). For these various reasons, the Greenland and Antarctic ice core records are widely used to determine the magnitude and timing of climatic and environmental changes 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 (Fig. 2.9). 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. 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 therefore offer a range of different temporal resolutions. The records are thinnest at coastal locations. At inland locations, especially those not located on domes, the records may be affected by changes in ice sheet elevation, or changes in ice origin from upstream areas. Drilling and handling may damage the quality of the ice core samples when the transition is located in the brittle zone, between 400 to 1000 m depth.

These records offer the potential to compare local signals (Antarctic site temperature and accumulation), regional signals (sea salt and marine biogenic sulphur concentrations), hemispheric signals (Antarctic dust concentration) and global signals (greenhouse gas concentrations in the atmosphere), which are described in more detail below. The stable isotopic composition of hydrogen (expressed as δD) or oxygen (expressed as or $\delta^{18}O$) in ice is normally used to quantify past Antarctic surface air temperature changes, because there is a strong linear relationship between snowfall isotopic composition and temperature (Masson-Delmotte et al. 2004). This relationship is created 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 the relationship remains stable in central Antarctica through both glacial and interglacial times, leading to uncertainties of no more than 20-30% in reconstructed temperatures (Jouzel et al. 2003). Although 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 is limited in inland Antarctica. Nevertheless, past changes in evaporation conditions may have affected Antarctic snowfall isotopic composition. To counter this effect, the deuterium excess parameter ($d = \delta D - 8\delta^{18}O$), which is mainly affected by a kinetic fractionation effect, has been used to quantify more precisely past changes in moisture source and site temperatures (Jouzel et al. 2001); (Vimeux et al. 2002). The moisture source temperature is an integrated quantity that may correspond either to changes in sea-surface-temperature in the dominant evaporation area, or to geographical changes of the main moisture origin at sea with time.

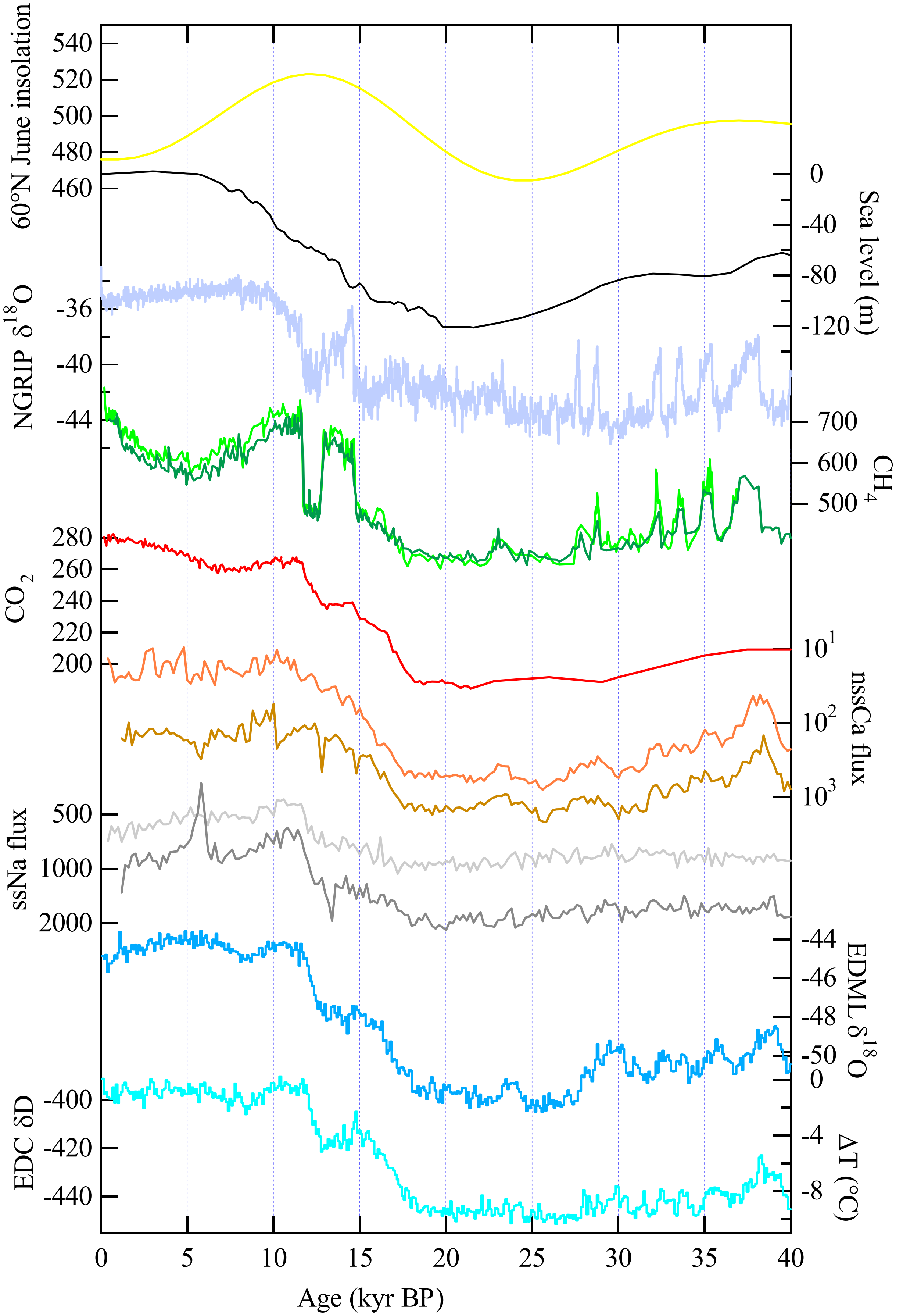
The water stable isotope changes between glacial and interglacial periods are of comparable magnitude from one Antarctic ice core to the next, with absolute differences likely to reflect differences in ice sheet elevation. It is estimated that the temperature difference between glacials and interglacials over the past 400,000 years averaged around 9°C with an uncertainty of +/-2°C (Jouzel et al. 2001). In central Antarctica, the coherency between the various temperature records based on stable isotopes is very high (within 1°C) (Watanabe et al. 2003). As we see from Fig 2.8, Antarctic temperatures derived from stable isotope measurements appear to have been more or less constant from 19 to 23 ka BP during the "Glacial Maximum" (i.e. the temperature minimum). Warming associated with Termination 1 in Antarctica began at about 19 ka BP. It was interrupted by a 1.5°C cooling from 14 to 12.5 ka BP, a period labelled the "Antarctic Cold Reversal" (Jouzel et al. 1995). Warming began again at 12.5 ka BP (Fig 2.8), which led to the beginning of the early Holocene, with the fastest temperature rise detected in the 850 ka EPICA Dome C record – a rate of ~4°C per 1000 years (Masson-Delmotte et al. 2004). It is noteworthy from Fig 2.8 that the Antarctic warmings at 15 ka BP and 12.5 ka BP preceded their counterparts in Greenland, and that the Antarctic Cold Reversal, which seems to be the southern equivalent of the cold Younger Dryas period in the northern hemisphere, preceded that event, suggesting that the controls on Termination I were probably operating much like those that controlled the Dansgaard-Oeschger events referred to earlier (see Fig. 2.7), with delays in heat transfer modulated by ocean circulation on the one hand and inertia in northern hemisphere ice sheets on the other hand.

In the 4th Assessment Report of the Intergovernmental Panel for Climate Change (IPCC, 2007) coupled ocean-atmosphere-sea-ice climate models run under conditions typical of the Last Glacial Maximum and of the present-day tend to underestimate the range of glacial-interglacial temperature change in central Antarctica (Masson-Delmotte et al. 2006). Given a situation in which the elevation of the ice sheet was unchanged, the models simulated a glacial-interglacial temperature change for central Antarctic surface of 4.5°C rather than the 9°C observed. Clearly some refinement is needed in the models.

It is worth noting that analysis of the IPCC-AR4 climate forecasts (Bracegirdle et al. 2008) shows a rise of around 3°C over the century to 2100. This is considerably faster than the fastest rate of rise observed in Antarctic ice cores (4°C per 1000 years) (see also (Masson-Delmotte et al. 2004). But it is comparable to or even slower than the rapid rates of temperature rise typical of Dansgaar-Oeschger Events during glacial times in Greenland (Fig 2.7), and of the Bolling Allerød warming in Greenland at

14,700 ka BP (Fig 2.10), and the warming in Greenland at the end of the Younger Dryas around 11,700 ka BP.

Across the transition in Termination I, samples from ice cores show that the concentration and flux of most aerosol species falls from high levels in the glacial to lower levels in the Holocene (e.g. see Na and Ca levels in Fig 2.10). The non-sea-salt Ca is considered to represent mineral dust of continental origin, the likely source for which is Patagonia. The higher concentration of dust in Dronning Maud Land (brown curve in Fig 2.10) indicates greater input of aeolian dust to the Weddell Sea region from the nearby Patagonian source (Fischer et al. 2007). The flux of non-sea-salt calcium decreases into the Holocene by a factor of 30 at Dome C and 10 in Dronning Maud Land (Fig 2.10), probably again reflecting proximity to the Patagonian source. Models show that there is surprisingly little change in transport strength and atmospheric residence time between South America and Dome C between glacial and interglacial periods, so the large reduction in flux seen at both sites during the transition, which began at about 18 ka BP, must primarily result from changes in Patagonia, the dust source region (Wolff et al. 2006, Fischer et al. *2007*).



60°N June insolation
540
520
500
480
460
Sea level (m)
0
-40
-80
-120
NGRIP δ18O
-36
-40
-44
CH4
700
600
500
CO2
280
260
240
220
200
nssCa flux
10^1
10^2
10^3
ssNa flux
500
1000
2000
EDML δ18O
-44
-46
-48
-50
EDC δD
-400
-420
-440
ΔT (°C)
0
-4
-8
0
5
10
15
20
25
30
35
40
Age (kyr BP)

Figure 2.10. Climate and environmental records of Termination I. From top to bottom: yellow, 60°N June insolation (W/m²) (Berger & Loutre 1991a); black, global sea level change (m) (Waelbroeck et al. 2001) (Lambeck & Chappell 2001); 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 2004) (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 (ug/m²yr) (Wolff et al. 2006, Fischer et al. 2007); sea-salt sodium flux, a proxy for marine aerosols, from EDML (light gray) and EDC (dark gray) (ug/m²yr) (Wolff et al. 2006, Fischer et al. 2007); blue, EDML $\delta^{18}O$ (‰), a proxy of EDML temperature (EPICA 2004); and finally, in light blue, EDC δD (‰), a proxy of EDC temperature (Jouzel et al. 2007).

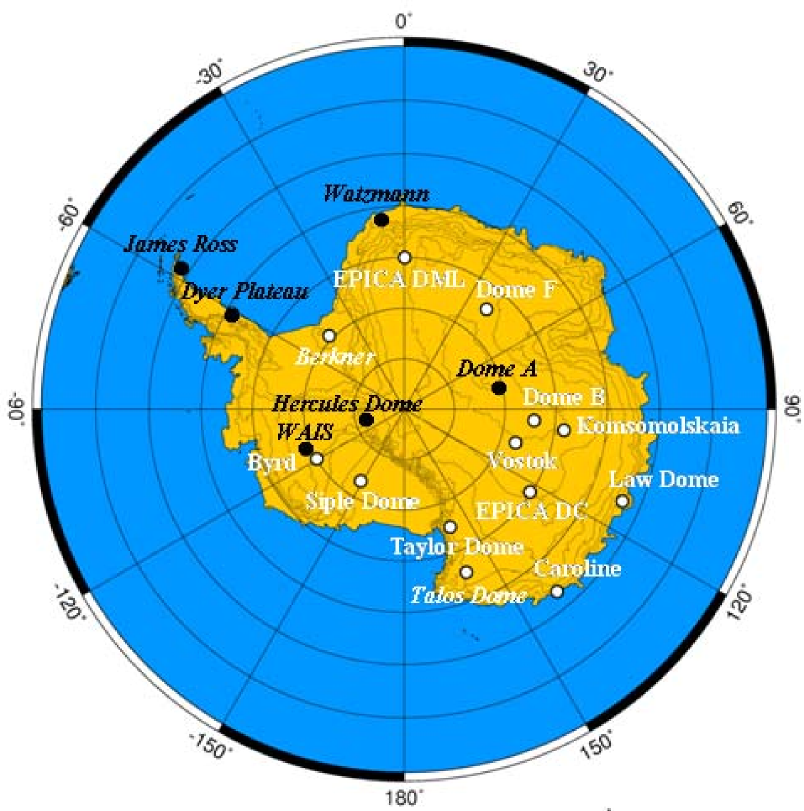


Figure 2.11. Map of Antarctica (created using http://www.aquarius.ifm-geomar.de/omc/) showing the deep ice core sites where climate records have been obtained back to the Last Glacial Maximum and beyond: Byrd (Hammer et al. 1994) (Blunier et al. 1997), Caroline (Yao et al. 1990), Vostok (Petit et al. 1999a), 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-community-members 2004), Dome F (Watanabe et al. 2003), Siple Dome (Brook et al. 2005) and EPICA Dronning Maud Land (EPICA 2004) . The surface elevation is represented as grey contours (100, 200, 500, and each 1000 m). Locations of existing 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. Future ice core drilling projects are displayed in black.

The flux of sea-salt Na is considered to reflect the extent of sea-ice cover in the vicinity. Here the signal is the opposite to that of Ca, in that the flux of Na is higher at Dome C than in Dronning Maud Land (DML) (Fig. 2.8). This could reflect the proximity of the DML site to the large expanse of the commonly ice-covered Weddell Sea, which lies up-wind. The flux of sea-salt Na decreases into the Holocene, reflecting diminution of the sea ice as warming proceeds, but by a larger amount at Dome C than at DML (Fig. 2.10), which again may reflect the persistence of a large expanse of sea ice in the Weddell Sea close to DML (see discussions in (Wolff et al. 2006, Fischer et al. *2007*)).

Although the broad features of the last glacial termination are now well documented in Antarctica, there is room for refinements to be made in the sequence of events through 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. To make such refinements calls for further high-resolution ice core records of the last deglaciation, and a comparison of records from ice drilling sites representative of different oceanic basin catchments. Recently drilled ice cores at Berkner Island and Talos Dome, spanning the last deglaciation and beyond, and future new deep ice cores (black dots in Fig 2.11) will help to progress understanding of the responses of climate and biogeochemical cycles to external forcings, of the variability of climate in and around Antarctica, and of the reaction of the Antarctic ice sheet to the major climate changes of the past.

2.6 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 (IPCC, (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 Last Glacial Maximum 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. 2003).

The configuration of the Antarctic Ice Sheet during the Last Glacial Maximum (LGM, c 21 ka BP) was modelled in the CLIMAP reconstruction by Stuiver and colleagues (1981) and later revised by Denton and colleagues (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).

2.6.1 **EAIS 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 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 20,000 years, indicating that the ice sheet retreated from the continental shelf prior to the LGM (Anderson & Andrews 1999).

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 ~ 9000 yr BP, and that the retreat of the grounding line to its present coastal position was complete by ~ 2000 yr 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 8000 and 5000 yr 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 11500 yr BP. A

Mid-Holocene re-advance of the Lambert/Amery system occurred between 7000 and 4000 yr 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 21,200 and 17,000 yr 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 19,000 yrs BP (see Figs. 2.7 and 2.8). 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 7000 years 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 ~ 7000 and 5000 yr BP in this region (Denton et al. 1991).

2.6.2 **WAIS 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 & 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 & 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 10,000 cal yr BP (ka BP) (Lowe & Anderson 2002).

2.6.3 **Antarctic Peninsula Ice Sheet Expansion and Retreat**

The Antarctic Peninsula Ice Sheet (APIS) advanced to the edge of the continental shelf during the Last Glacial Maximum. The expanded ice sheet there was more than double the size of the landmass (Ó Cofaigh et al. 2002, Dowdeswell et al. 2004, Heroy & 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 ~14,000 cal yr BP on the outer shelf off Marguerite Bay (Heroy & Anderson 2005). Steps in the data occur at ~14,000 and possibly 11,000 cal yr BP, coincidental with global melt water pulses MWP 1a and 1b, and show that rapidly rising sea level at those times destabilised the marine ice sheet (Heroy & Anderson 2005).

By ~10,000 cal yr 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 & Anderson in press). Also, as the ice sheet vanished 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 dynamical processes.

2.7 Antarctic deglaciation and its impact on global sea level

In 1996 the Intergovernmental Panel on Climate Change noted that of all of the terms in the sea level change equation, the largest uncertainties relate to the Earth's major ice sheets (Houghton et al. 1996) with the Antarctic Ice Sheet having least evidence to constrain its past volumes (Houghton 2001). This is important, as better understanding of how global ice sheets have contributed to sea level across Termination I and into the Holocene is the key to predicting how relatively subtle changes in these ice sheets could influence sea level in the future. When global continental ice sheets grew before the Last Glacial Maximum (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 10,000 years. 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 meters 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 & Lambeck 1988) to 6-13m (Bentley 1999). More recently, an improved three-dimensional thermomechanical model has suggested a contribution of 14-18m (Huybrechts 2002). This variability is significant, because when the calculated contributions of the Last Glacial Maximum 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 (Weaver et al. 2003); (Bassett et al. 2007a), or the Laurentide and Fennoscanadnavian Ice Sheets (Peltier 2005). Geological evidence does not rule out a contribution from Antarctica (Verleyen et al. 2005, Bassett et al. 2007b). 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 ice thickness and maximum marine limits during the LGM (Huybrechts 2002).

Observations of relative sea-level changes provide 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 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. The interaction of these two factors can result in complex curves of RSL change over time. Typically, these curves are determined for the period since the end of the Last Glacial. 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 provides 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 isolation basin case, 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 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 & Orombelli 1994, Hall et al. 2003, Bentley et al. 2005a).

2.7.1 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 75-105º E and 30-45º 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) (Fig. 2.12). The Ross Sea curves show exponential RSL fall owing to deglaciation of the coast by the retreating West Antarctic Ice Sheet (Conway et al. 1999a). The amount of RSL fall (~ 30 m) is low compared to other 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.

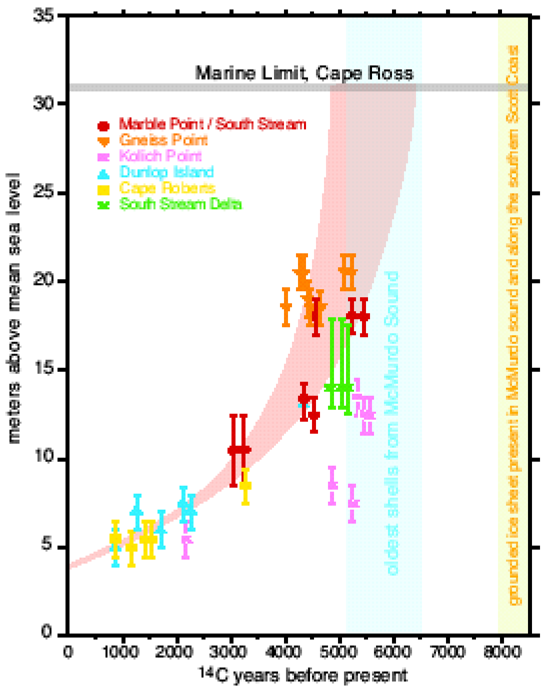


Figure 2.12. Relative sea level curve for the Ross sea based on radiocarbon dates of elephant seal skin along with shells and penguin remains and guano, in raised beaches on the southern Scott Coast and Terra Nova Bay. This gives valuable information regarding the timing of deglaciation (Hall et al. 2003).

Zwartz et al. (1998) and Verleyen et al (2005) determined RSL curves for the last 15,000 years for the Vestfold Hills and Larsemann Hills in East Antarctica by dating a series of marine-lacustrine transitions in insolation 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 B.P., 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. 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.

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 ~+14.5-16m at about 4000 ^{14}C yrs BP. In Marguerite Bay (south-central Peninsula) the RSL curve shows exponential RSL fall since deglaciation at ~ 9000 ^{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. 2005).

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 Last Glacial Maximum (Bentley 1999); testing the hypothesis that Meltwater Pulse 1B was wholly or partly derived from Antarctica (Bassett et al. 2007a); and determining the timing and style of deglaciation around the ice sheet margin (Conway et al. 1999a). 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 & James 2005) and presence/absence of meltwater pulses (Bassett et al. 2007a), 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 untapped 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.

2.8 Changes in sea ice extent through the Holocene

Despite the fundamental importance of sea ice to many aspects of climate (see section 2.4), 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 only began in the 1970’s, 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. 2002). 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.

Most 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 (Crosta et al. 2004, Gersonde et al. 2005). 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 (Zielinski & Gersonde 1997, Diekmann et al. 2000, Armand et al. 2005, Crosta et al. 2005). 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. 1998), see Fig 2.13 from (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 (Fig.2.13) (Hodell et al. 2001, Nielsen et al. 2004, Crosta et al. 2007, Crosta et al. 2008). 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 ~50ºS 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 ~50ºS. This was then followed by a late Holocene warming that saw sea ice retreat southward of 50^{o}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^{o}S (Hodell et al. 2001).

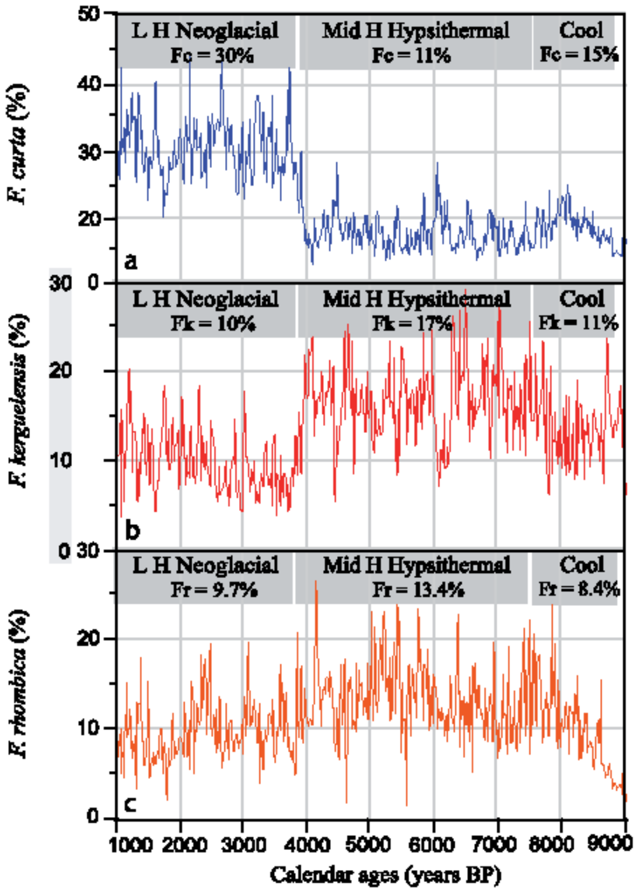


Figure 2.13. Relative abundances of (a) the Fragilariopsis curta group and (b) *F. kerguelensis* and (c) *F. rhombica* versus calendar ages in core MD03-2601 off Adélie Land, East Antarctica. Mean abundances of the three species are reported for each climatic period. Fc, F. curta; Fk, F. kerguelensis; Fr, F. rhombica; LH, late Holocene; Mid H, mid Holocene (Crosta et al. 2007).

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 particu;ar ecological responses at 0.5°C, and a more gradual change in *F. rhombica* (Fig. 2.12) the latter giving a signal that is in good agreement with model output (Renssen et al. 2005b). This lasted until at least ~1 ka BP (the top of the core) (Fig. 2.14)(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 Fig 2.10), 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 in the Southern Ocean may instead be consistent with forcing by southern summer insolation (Renssen et al. 2005a). 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 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 (Fig. 2.14). 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. 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 so far no quantitative calibration has been made (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

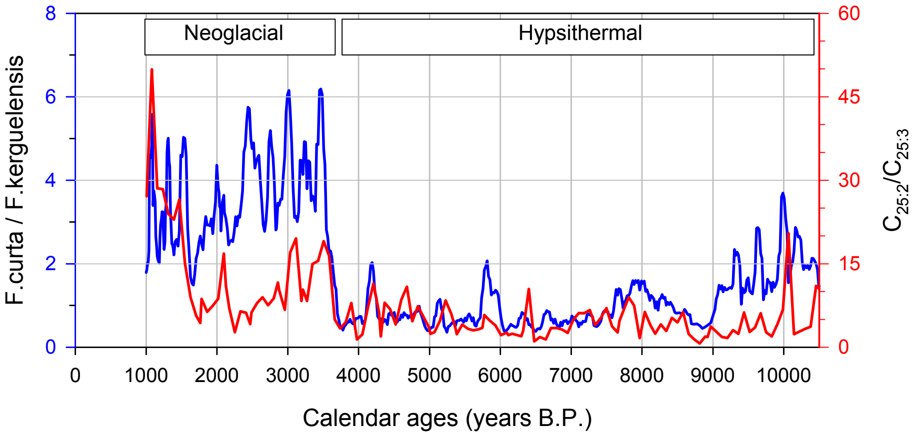


Fig. 2.14. Diatom (blue) and highly-branched isoprenoid (red) records from a sediment core retrieved from the Adelie Trough off Wilkes Land, East Antarctica. High values of the *F. curta*/*F. kerguelensis* ratio (left axis) and the $C_{25:2}/C_{25:3}$ ratio (right axis) during the Neoglacial indicate 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. 2007a). 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 during the 20th century (Curran et al. 2003, Abram et al. 2007a). Understanding the factors that caused these regional variations in sea ice decline may help to better constrain the regional responses of Antarctic sea ice to future climate change, and the far-reaching impacts that these may in turn have on Earth's climate.

2.9 Holocene climate changes – regional perspectives

In this section we synthesise 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), Ingólfsson et al. (2003), Ingólfsson et al. (2003), Ingólfsson and Hjort (2002), Ingólfsson (2004), Jones et al. (2000) and Hodgson et al. (2004) 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 2000 years with a focus on Neoglacial cooling, the presence of warm periods, the possibility of Little Ice Age (LIA) like event, and the recent rapid climate changes.

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 1300 yrs following the Antarctic standard (Berkman et al. 1998) 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)).

2.9.1 East Antarctica

Deglaciation history and the Pleistocene-Holocene transition

The widespread Antarctic early Holocene optimum between 11,500 and 9000 yr 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. 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 paleolimnological 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 & 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).

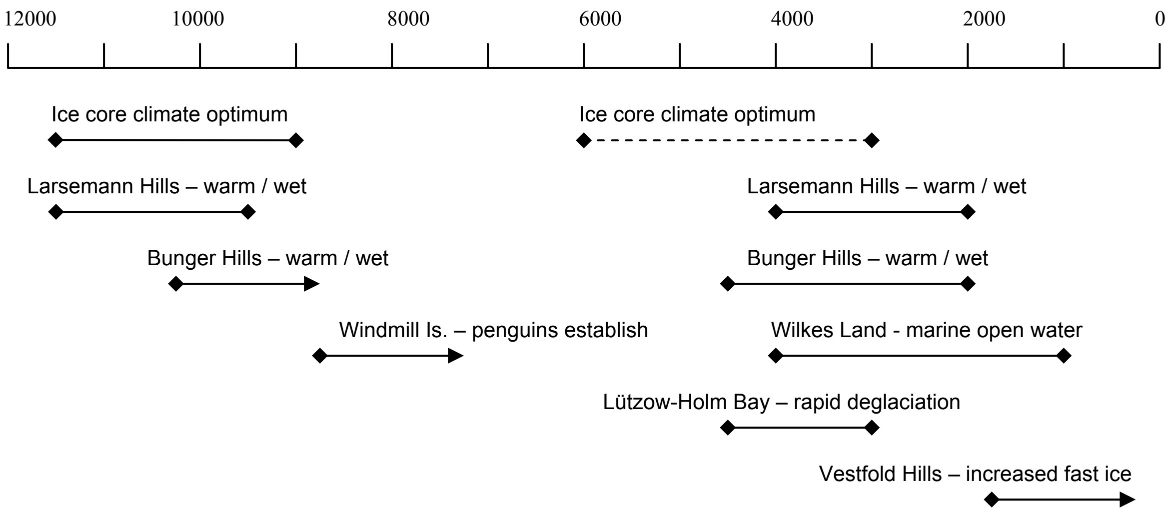


Table 2.3. Selected Holocene climate changes – East Antarctica

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 & 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 & 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 warm period 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 11,500 and 9000 yr. BP (Masson et al. 2000).

After the early Holocene

All eastern Antarctic sites show a weak climate optimum between 6000 and 3000 yr 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 since 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 paleoclimatological inferences from being made for the period after the early Holocene optimum, particularly in lower altitude lakes (Fulford-Smith & Sikes 1996, Roberts & McMinn 1999b).

In Wilkes Land, open water conditions were inferred in 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 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 & 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 & 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. (2004) 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 2.8 - 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. 2006c), 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 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 & 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 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. 1000 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 East Antarctica, but dating uncertainties are still high in some areas. Because the MHH acts as one of the historical analogues for the present day 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 2000 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 or 1100-700 BP), and Little Ice Age (LIA) (consitsting of cool intervals beginning about 1650, 1770, and 1850 AD, each separated by slight warming intervals) 300-100 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 BP). The search in Antarctica for climate signals like these that are apparent in the northern hemisphere is an important element in understanding how the Earth's climate system works. One limitation in such a comparison is the dearth of southern hemisphere records.

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 2 ka 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 2000 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 & Pickard 1986). Cold conditions were inferred between c. 1.3 ka BP and 250 yr BP (Bronge 1992) and low precipitation could tentatively be inferred from c. 1.5 ka ^{14}C BP (c. 1.3-1.5 ka BP, but dating uncertainty is high (Roberts & McMinn 1999a). Meltwater input into the lakes gradually decreased from 3 ka BP onwards (Fulford-Smith & 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 at 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. 2006b). Instead, we see a very rapid salinity rise during the past few decades that has been bought about by increased evaporation rates. 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 & 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). Detailed paleoclimatic 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. There is no 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. In the last few decades a rapid salinity increase has been recorded in lakes in the Windmill Islands region.

2.9.2 **Antarctic Peninsula**

The pattern and mechanisms of Holocene palaeoenvironmental change in the Antarctic Peninsula region have recently been reviewed (Bentley et al. 0000).

Deglaciation history and the Pleistocene-Holocene transition
The early Holocene climate optimum detected in ice cores lasted to around 9,000 years BP (Masson et al. 2000, Masson-Delmotte et al. 2004) occurred during the continued deglaciation of the continental shelf near the Antarctic Peninsula (Bentley 1999, Ingólfsson et al. 2003, Bentley et al. 2006). Ice sheet retreat around the Peninsula probably began c. 14-13 ka BP (Evans et al. 2005, Heroy & Anderson 2005), and continued through the Holocene. 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 & Taylor 2002, Taylor & 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, Ingólfsson et al. 2003).

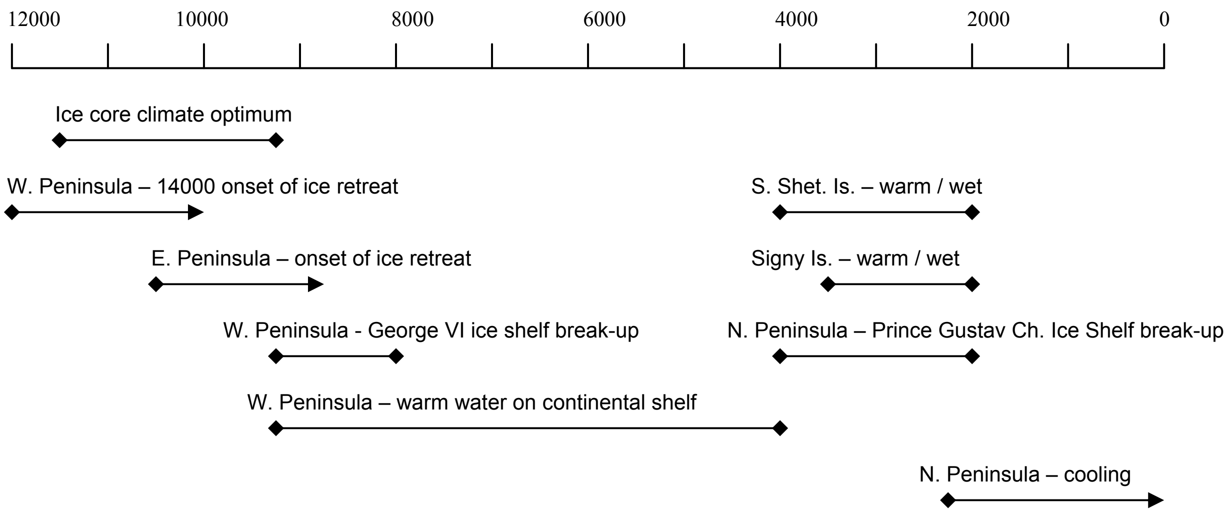


Table 2.4. Selected Holocene climate changes – Antarctic Peninsula

After the early Holocene

The period after the early Holocene optimum shows complex patterns in the Peninsula region. Ice shelves on the western side had collapsed, whilst those on the east were still stable (Hodgson et al. 2006a). The onset of marine conditions in epishelf lake sediments on Alexander Island shows that King George VI Ice Shelf collapsed at c. 9.6 ka BP, immediately following 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 Peninsula, indicate a dramatic increase in the presence of warmer surface waters over the Peninsula'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 et al. 2007). The ice shelf reformed from c. 7.9 ka 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 Antarctic Peninsula, shows that it was stable throughout the Holocene [from 11,500 years 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 Peninsula in the early Holocene, with a steepening of the thermal gradients to the north and west (Figure 2.14) (Hodgson et al. 2006a). 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 (~13,300 years B.P. at Palmer Deep and 15,700 years BP at Lafond Trough) and the east of the Antarctic Peninsula (~10,600 years BP at Erebus and Terror Gulf, 10,700 years B.P. at Greenpeace Trough, and 10,500 years BP at Larsen B embayment, all using conventional carbon-14 ages) (Domack et al. 2005, Heroy & 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.

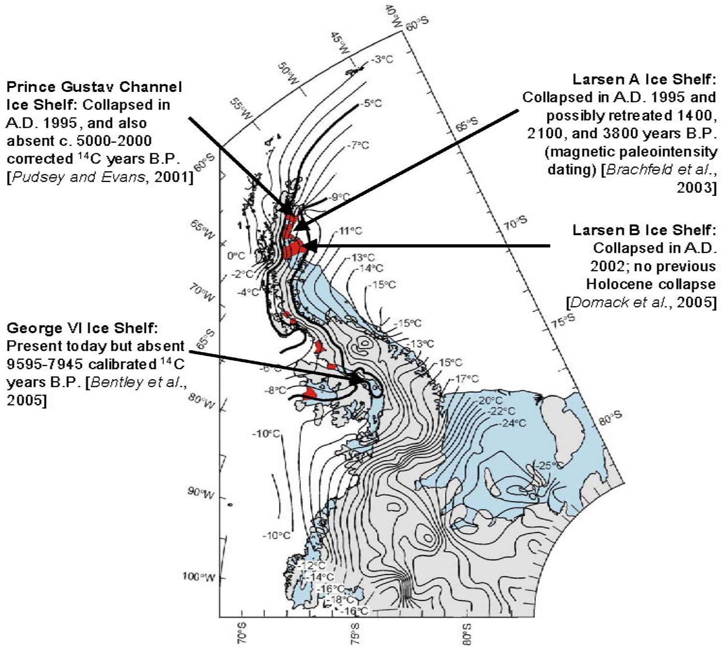


Fig. 2.15. Recent retreat of Antarctic Peninsula ice shelves shown together with the –9ºC mean annual isotherm that marks the southern limit of ice shelf stability. Ice shelves that have retreated or collapsed are marked in red and extant ice shelves in blue. Locations of ice shelves studied are marked together with known Holocene retreat events. (Map courtesy of D. Vaughan, British Antarctic Survey.)

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 Peninsula 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 & Zale 2006). In general, on the west side of the Peninsula 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 (Hansom & Flint 2004), 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, Björck et al. 1991b, Björck et al. 1996) and cold marine waters with extensive sea-ice cover in the fjords 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 Peninsula, 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 Peninsula 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 Peninsula. This interval is reviewed in detail in Hodgson et al. (2004). 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 Antarctic Peninsula region (Björck et al. 1991a) or c. 3.3 to 1.2 ^{14}C ka BP (c. 3.6-3.4 to 1.2 ka BP-0.9 ka BP) just to the north of the Peninsula (Jones et al. 2000, Hodgson & 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 et al. 1996) and James Ross Island (Björck et al. 1996) to maritime Antarctic islands such as Signy Island (Jones et al. 2000, Hodgson & Convey 2005). Sites in the northern Peninsula 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 Peninsula between c. 5 and 2 ka BP (Pudsey & 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 Antarctic Peninsula is common to both the early and the mid-Holocene (Figure 2.15).

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 Peninsula region as in East Antarctica, there is a need for further well-dated, high resolution sedimentary records.

Past 2000 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 (6000 years), this date is far from certain (Pudsey & 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 & 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 in Ardley Island and Barton Peninsula (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 (Liu et al. 2006).

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 Peninsula. In proxy records, the warming is seen in increased sediment accumulation rates in some Peninsula lake cores (Appleby et al. 1995), and some high-resolution marine cores. 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). 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 is observed in various regions of the Peninsula, with the lakes in Signy Island showing a remarkably rapid response in ecosystem functioning.

2.9.3 Ross Sea Region, Victoria Land Coast and Trans Antarctic Mountains

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 recession was slow until c. 10.8 ^{14}C ka BP (c. 12.7-12.9 ka BP) (Denton & Marchant 2000, Hall & Denton 2000). 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 & 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 & 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 & 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).

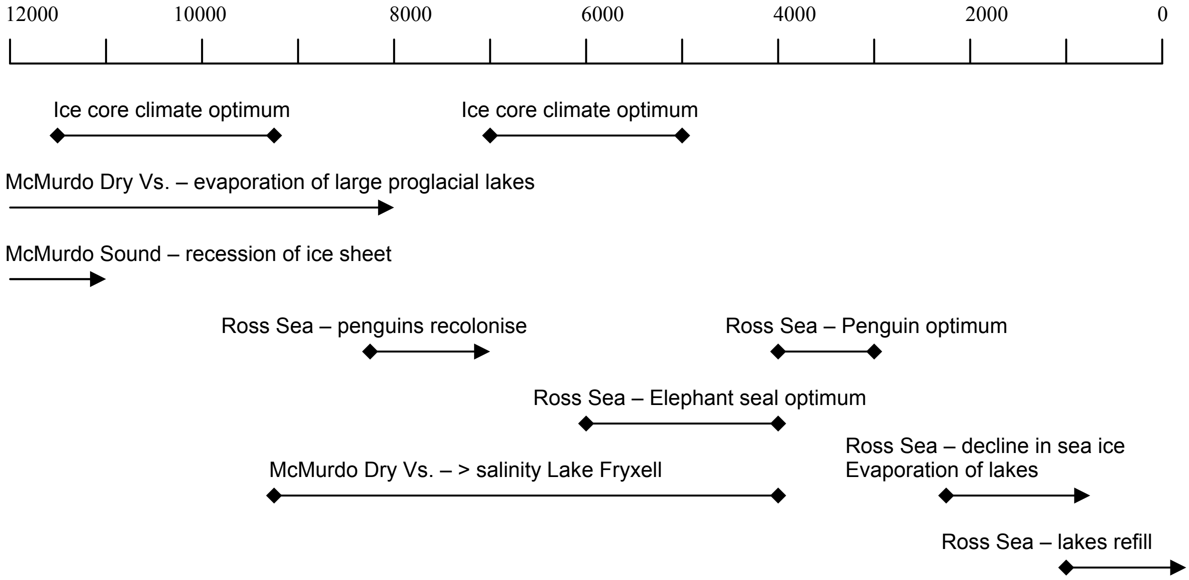


Table 2.5. Selected Holocene climate changes – Ross Sea

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 Ross Sea region until c. 8 ka BP, after an absence of c. 19,000 years (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 & Denton 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 & 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 & Denton 2000)). This corresponds with a secondary climate optimum detected in the ice cores of the Ross Sea sector, between 7000 and 5000 yr BP (Masson et al. 2000).

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 & Orombelli 1994, Emslie & 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 & Denton 2000)).

The past 2000 years – Late Holocene warm period and recent rapid climate change

In contrast to the Peninsula, the warmest conditions in the Ross Sea 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 6000 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 & 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 is (Hall & 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).

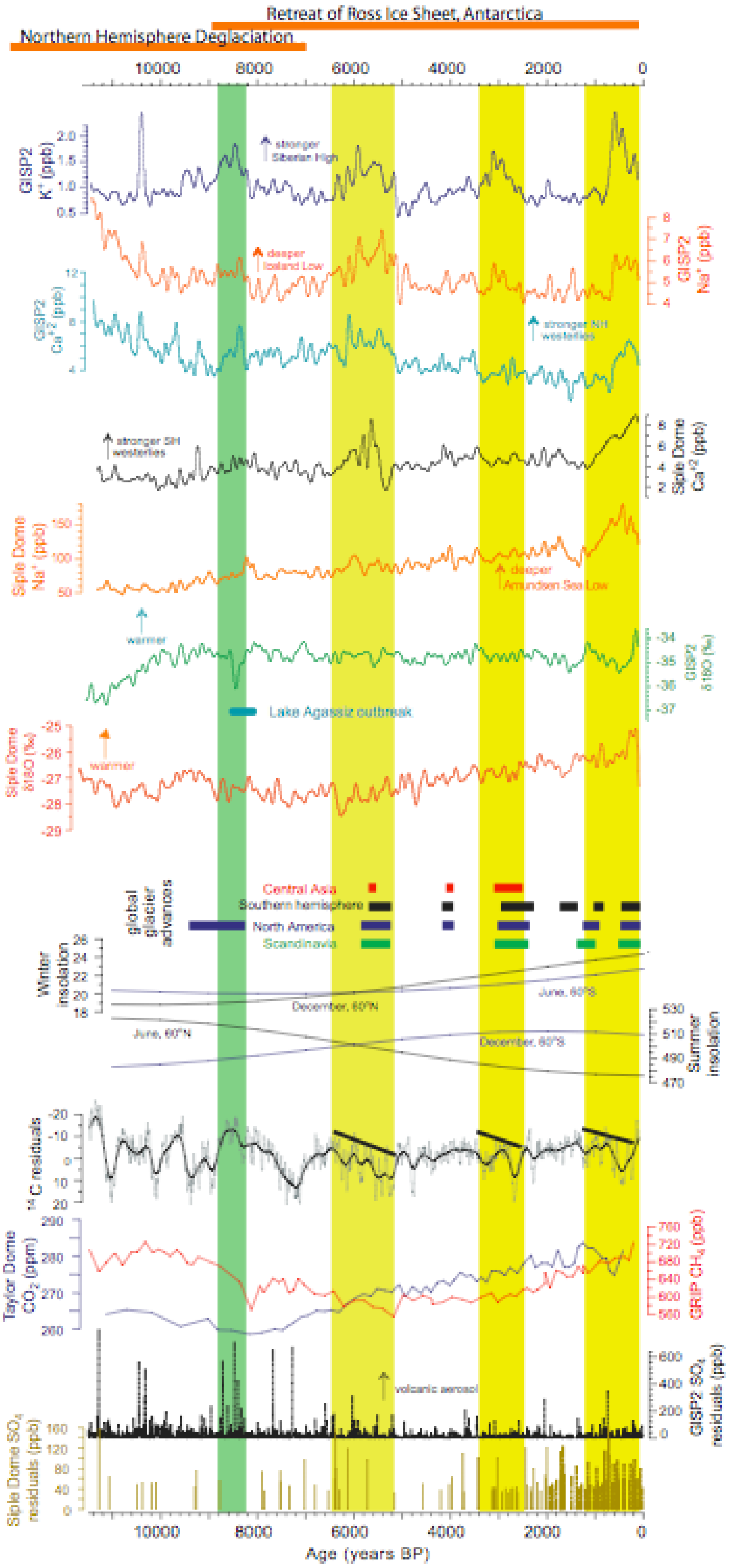
2.8.4 Summary

In general, geological evidence shows that deglaciation of the currently ice-free regions was completed earlier in East Antarctica compared with the Antarctic Peninsula, 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 East Antarctica and the Antarctic Peninsula (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 Ross Sea sector that show an optimum around 7-5 ka BP, and the Epica Dome C ice shows an optimum between 7.5 and 3 ka BP. The occurrence of a later Holocene climate optimum in the Ross Sea 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 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 in particularly in the Dronning Maud Land region and particular regions of the Antarctic Peninsula 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.

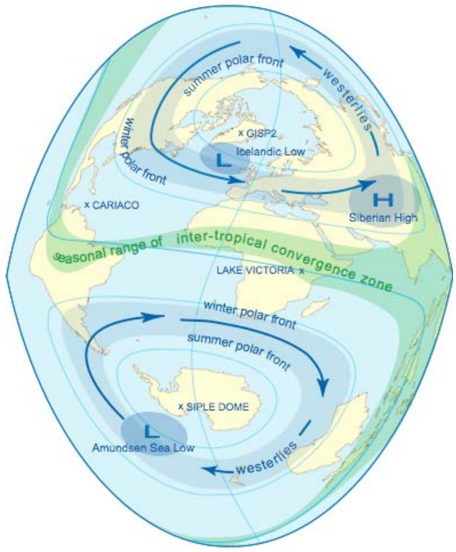
2.10 Holocene climate changes (The Ice Core Record)

As we have seen in sections 2.2 and 2.4, above, ice cores provide a robust reconstruction of the variability in past climate. For the Holocene period we tend to have higher resolution than at older times, because the ice is less compressed and it is easier to count individual annual layers, and because there are many more short cores penetrating all or part of the Holocene than there are long cores extending further back in time. The higher resolution and greater geographic spread of cores allows us to get clear continent wide and regional pictures of changed in temperature, and precipitation, and in other properties from the collection of which we can determine changes in atmospheric circulation such as the local impact of the El Nino-Southern Oscillation (ENSO), the strength of the westerlies, the intensity of the Antarctic Oscillation or Southern Annular Mode (SAM), the extent of sea ice, marine productivity, and the chemistry of the atmosphere. These signals can be compared with those from ice cores in Greenland to help us to understand the extent to which the system is forced by the sun or by ocean circulation.

Ice core data from Fig. 2.16 are based mainly on a comparison of the West Antarctic Siple Dome core and the Greenland GISP2 core, locations of which are shown on the map adjoining the figure. Profiles of properties from the Siple Dome core (e.g. δ18O representing temperature) confirm that over the past 11,000 years representing Holocene time the climate of Antarctica has been relatively stable. The Siple Dome



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



summer polar front
westerlies
winter polar front
x GISP2
Icelandic Low
x CARIACO
H
Siberian High
seasonal range of inter-tropical convergence zone
LAKE VICTORIA x
winter polar front
summer polar front
x SIPLE DOME
L
Amundsen Sea Low
westerlies

Figure 2.16 Examination of potential controls on and sequence of Antarctic Holocene climate change compared to 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 & Mayewski 2002); GISP2 Na^+ proxy for the Icelandic Low (Meeker & Mayewski 2002); GISP2 Ca^{++} proxy for the Northern Hemisphere westerlies (Mayewski & 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 & Stuiver 1997); Siple Dome $\delta^{18}O$ proxy for temperature (Mayewski et al. 2004); timing of the Lake Agassiz outbreak that may have initiated Northern Hemisphere cooling at ~8200 years ago, (Barber et al. 1999); global glacier advances (Denton & 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 & Loutre 1991b); summer insolation values (W m^{-2}) at 60°N (black curve) and 60°S latitude (blue curve) (Berger & Loutre 1991b); 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 (Indermuhle 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. 1999b). Green bar denotes the 8800-8200 year ago event seen in many globally distributed records associated with a negative $\Delta^{14}C$ residual. Yellow denotes 6400-5200, 3400-2400, and since 1200 year ago events seen in many globally distributed records associated with positive $\Delta^{14}C$ residuals. Map insert shows location of GISP2, Siple Dome, Icelandic Low, Siberian High, Amundsen Sea Low, Intertropical Convergence Zone, and westerlies in both hemispheres.

Oxygen isotope curve shows early Holocene warming between 10000 and 9000 years ago, cooling that peaked around 6400 years ago, then continued slight but steady warming with some mild ups and downs, concluding with a recent prominent warming event. These ups and downs include a broad warming typical of the Mid Holocene Hypsithermal, and which peaked around 2000 years ago, then a Neoglacial cool event peaking around 1700 years ago, followed by further warming to a peak around 1400 years ago, a slight cooling and then a significant rise in recent decades. By contrast, and with few exceptions, the GISP2 isotope curve is more or less flat with some modest ups and downs, concluding, as at Siple Dome, by warming higher than before. The overall warming since 6000 years ago at Siple Dome may relate to the increasing summer insolation at 60ºS (Fig. 2.16); CO_2 also began to increase slightly during this period, as indicated by data from the Taylor Dome on the eastern edge of the Ross Sea (Fig. 2.16).

For the past 1500 years or so these curves can be related to global temperature patterns derived from temperature proxies and the instrumental record. However, while that exercise can be carried with a fair degree of confidence for the northern hemisphere, where there are abundant temperature proxies (e.g. IPCC, 2007, Chapter 6, Fig 6.10), it is more difficult for the southern hemisphere, where the proxies are far fewer (IPCC, Chapter 6, Fig. 6.12). What we can see, from Fig 2.17, is that the Siple

Dome oxygen isotope curve diverges markedly from the Law Dome isotope curve from the East Antarctic coast. Since around 1750AD conditions have warmed at Siple Dome and cooled at Law Dome. It seems likely therefore that the Antarctic ice cores must reflect regional conditions to some extent, particularly those near to the coast.

What we can also see in the ice core data is the same kind of thermal time lag between the northern and southern hemispheres that we see in pre-Holocene times. For instance in Fig 2.18 the Medieval Warm Period peaking at 1000AD and Little Ice Age peaking at 1600AD start a few hundred years earlier in the south. This is also clear from the comparison between the Siple Dome and GISP2 oxygen isotope data in Fig. 2.16. The delay is thought due to the time taken by the meridional overturning circulation to translate warming or cooling signals from south to north.

Some climate events seen in the oxygen isotope records were relatively abrupt, if short-lived. For instance short-term sharp cooling events in the Siple Dome core 8600 and 8800 years ago precede a short sharp cooling around 8400 years ago in Greenland. Conceivably these southern and northern events are related, but offset by a

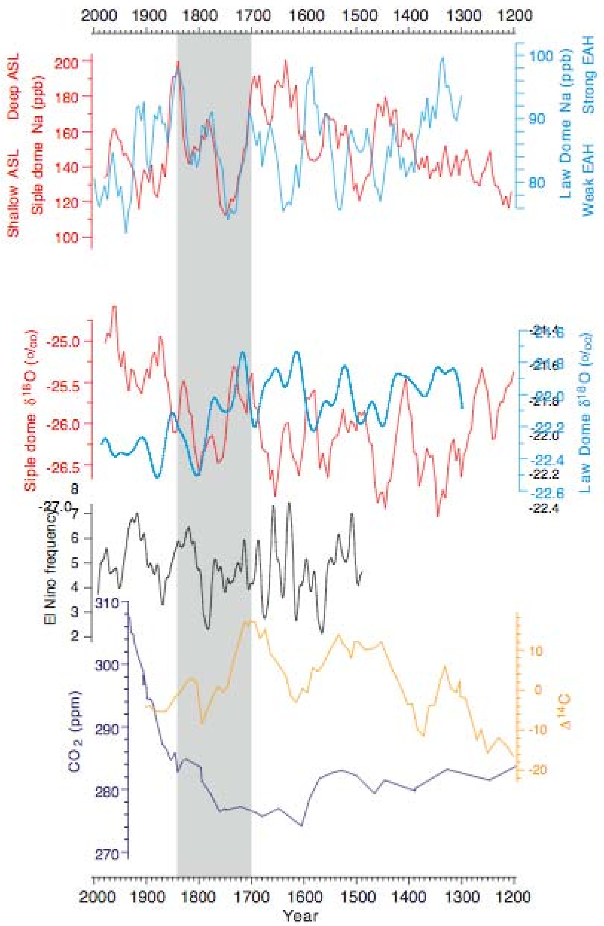


Figure 2.17. 25 year running mean of SD (Siple Dome (red)) and DSS (Law Dome (blue) Na^+ (ppb) used as a proxy for the ASL (Amundsen Sea Low) and EAH (East

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). 25 year running mean SD (red) and DSS (blue) $\delta^{18}O$ (^{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, 1997; Steig et al., 2000). Frequency of El Nino polar penetration (black) based on calibration between the historical El Nino 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 1700-1850 year era climate anomaly discussed in text (references cited in (AGCS 2008)).

lag of a few hundred years much as we see with other climatic events in ice cores, due to the ocean circulation effect. It is somewhat counterintuitive that they occur at about the same time as a marked negative $\delta^{14}C$ proxy for solar intensity indicative of warming, which is extended across all records in Fig 2.16 by a green band. However, this also happens to be the time of the outbreak of Lake Agassiz into the North Atlantic, which substantially changed oceanographic conditions around Greenland (Fig. 2.16). In addition, this was a time of increased volcanic activity in the northern hemisphere, as recorded by SO4 in the GISP-2 core (Fig. 2.16), and volcanic aerosols can have a cooling effect.

The short sharp cooling 6400 years ago in Siple Dome similarly precedes a short sharp cooling 5500 years ago in Greenland (Fig. 2.16). These coolings are associated with a positive excursion in the $\delta^{14}C$ proxy for solar intensity, indicative of solar cooling, and marked by the middle yellow band across the figure. There is no obvious change at Siple Dome associated with the $\delta^{14}C$ proxy for solar intensity indicative of cooling at around 3000 years ago, but slight cooling is associated there with the $\delta^{14}C$ proxy for solar cooling after 1700. Strong cooling occurs in Greenland in association with this marked $\delta^{14}C$ marked decrease in solar intensity (Fig. 2.16).

The correspondence between solar cooling suggested by $\delta^{14}C$ data and cooling as suggested by oxygen isotope data is clearly not 1:1. For example, on the one hand the Δ14C data suggest pronounced solar cooling at 7000 years, when there is no oxygen isotopic change, and on the other hand the oxygen data suggest a pronounced cooling event at 4600 years BP, when there is no comparable $\delta^{14}C$ signal.

Chemical data from the ice cores can tell us about changes in wind strength. For instance an increase in Ca indicates increasing transport of dust, while a corresponding increase in Na suggests increasing transport of sea salt. The two increasing together at Siple Dome after 1000 AD (Fig. 2.18) suggest strengthening of the Amundsen Sea Low and the Southern Hemisphere westerlies (Mayewski and Maasch, 2006), in association with southern hemisphere cooling (e.g. Masson et al., 2000). We see much the same in the northern hemisphere after AD 1600 (Fig. 2.18), indicative of strengthening of the Siberian High, the Iceland Low, and the associated westerlies. The 600 year lag reflects the delayed northern hemisphere response to what is probably common external forcing by the sun, which had its effect first in the southern hemisphere before being transferred north through the ocean.

Similar increase in wind strength associated with solar cooling are indicated for Greenland by the K, Na and Ca data from the GISP-2 core centred on 3000 and 6000 years ago (Fig.2.16), as well as at around 8400 years ago during the Lake Agazssiz event. There is some evidence for similar strengthening of the Amundsen Sea Low and associated westerlies in the Siple Dome core 6400-5200 years ago and 8000 years ago, associated with solar cooling, but not at 3000 years ago. It is not clear why the solar-forced cooling at 3000 years ago affected Greenland but not Antarctica.

The most dramatic changes in atmospheric circulation during the Holocene in the Antarctic are the abrupt weakening of the Southern Hemisphere westerlies at 5400-5200 years ago, and intensification of the westerlies and the Amundsen Sea Low starting around 1200 years ago.

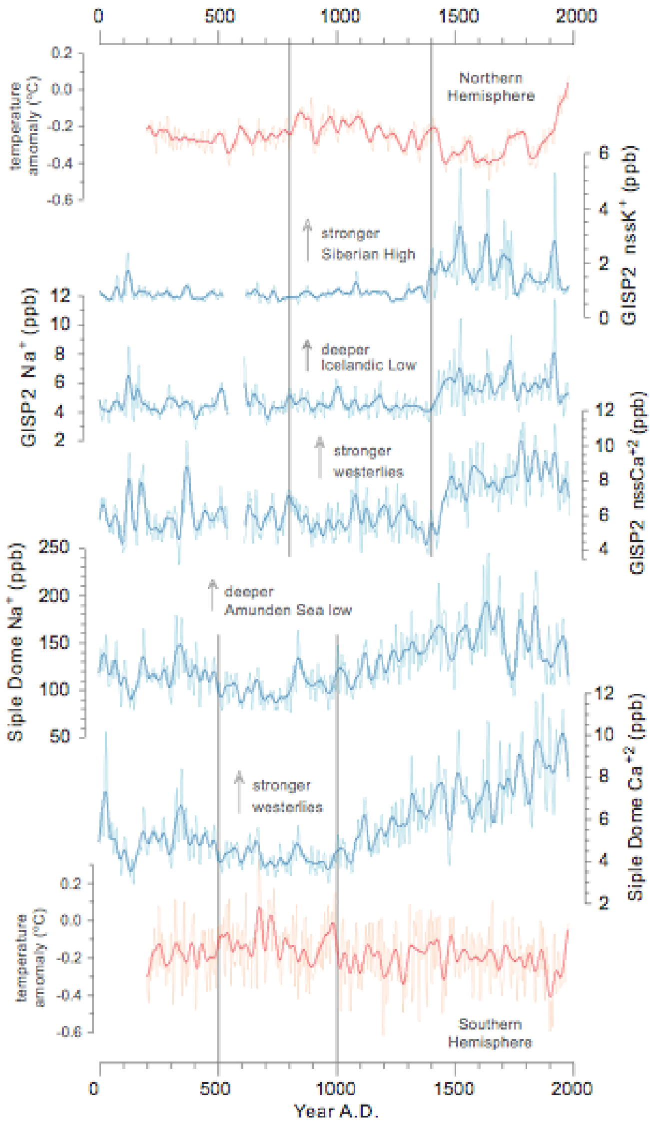


Figure 2.18. Northern and Southern Hemisphere reconstructed temperatures (in red from Mann and Jones, 2003) and ice core reconstructed atmospheric circulation systems (in blue from Mayewski and Maasch, 2006)) (Icelandic Low, Siberian High, Northern and Southern Hemisphere westerlies, and Amundsen Sea Low). Data is 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 analogs for naturally warm and cool temperatures, respectively). Figure taken from Mayewski and Maasch (2006). References cited in (AGCS 2008).

Evidence for abrupt climate change over the last 700 years can be seen in high resolution ice core records from East Antarctica (Law Dome) and West Antarctica (Siple Dome) (Fig. 2.17). As mentioned earlier, these two regions have operated inversely with respect to temperature and to the strength of atmospheric circulation on multi-decadal to centennial scales (Mayewski et al. 2004); temperatures decreased at Law Dome while increasing at Siple Dome, and the East Antarctic High weakened at Law Dome while the Amundsen Sea Low increased at Siple Dome. The exception is a climate change event (grey shading in Fig 2.17) that began around 1700 AD and ended around 1850 AD, during which time circulation and temperature behaved in the same way in both areas. The δ18O data show that this was a period of cooling in both places, which coincided with a period of decreasing solar output ($\delta^{14}C$). The close of this cooling event coincides with the onset of the steep modern rise in CO_2, and with the beginning of a significant warming event at Siple Dome that led to the warmest temperatures of the last 1800 years as represented by oxygen isotopes (Fig. 2.17) (Mayewski et al. 2005). The close of this event at around 1850 AD coincides with a major transition from zonal to mixed flow in the North Pacific (Fisher et al. 2004), suggesting a global scale association between Antarctic and North Pacific climate for this event.

In addition to offering high resolution temporal records, ice cores now also offer the novel prospect of examining in some detail the association between change in temperature and change in atmospheric circulation, especially during the last 2000 years, the period for which annually resolved dating is most accurate. This period holds particular interest as it is characterized by ice, ocean, and atmosphere conditions similar to those of today, as well by conditions during a past warm event – the Medieval Warm Period - and a past cool event – the Little Ice Age. Bipolar ice core records for this period 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 (Thompson & Wallace 2001, Bertler et al. 2004, Thompson & Wallace Regional climate impacts of the northern hemisphere annular mode); (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 2000 years (Fig. 2.19). The strength of the westerlies of recent decades in both hemispheres (green dots) falls within the range of variability of the strength of the westerlies of the Little Ice Age (blue dots), although those of the northern hemisphere in the late 1980s approach the range of variability of the westerlies of the Medieval Warm Period (red dots). The separation of the wind/temperature fields of today from those of the past confirm that the modern day atmosphere is not an analog for that of the Medieval Warm Period. The smaller amount of differentiation of today's conditions from those of the past in the southern hemisphere compared with the northern hemisphere may reflect the thermal isolation of the Southern Ocean from the global ocean by the increased strength of the Southern Annular Mode.

Taken together these various ice core data show that while the link to variations in solar output is strongly suggestive, as deduced also from the studies of sea-ice mentioned earlier, a fully satisfactory explanation for the forcing of Holocene Antarctic climate changes remains elusive.

A key finding is that whereas through most of the Holocene there has been a several hundred year time lag between southern hemisphere and northern hemisphere events (also seen in past glacial periods, as documented in section 2.2 and Fig. 2.7), in recent decades the northern hemisphere signal of rising temperature since about 1800 AD (top of Figure 2.18) has paralleled the southern hemisphere one as depicted by the oxygen isotope signal at Siple Dome (see Fig. 2.17). 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. This recent synchronicity is not shown as clearly in the Mann and Jones (2003) southern hemisphere data (Fig 2.18) perhaps because they were based on just three data sets. Recent re-analyses of northern and southern hemisphere proxies by the IPCC (Chapter 6 figures 6.10 and 6.12 in IPCC, 2007) confirm that the post-1800 AD temperature signals in the two hemispheres are more or less synchronous.

The extent to which these and other variations in Antarctic climate since Termination I affected Antarctic ecosystems is unclear, but this natural variability must have had its effect on life in the Antarctic and be taken into account in understanding Antarctica's modern environment and the potential for future change.

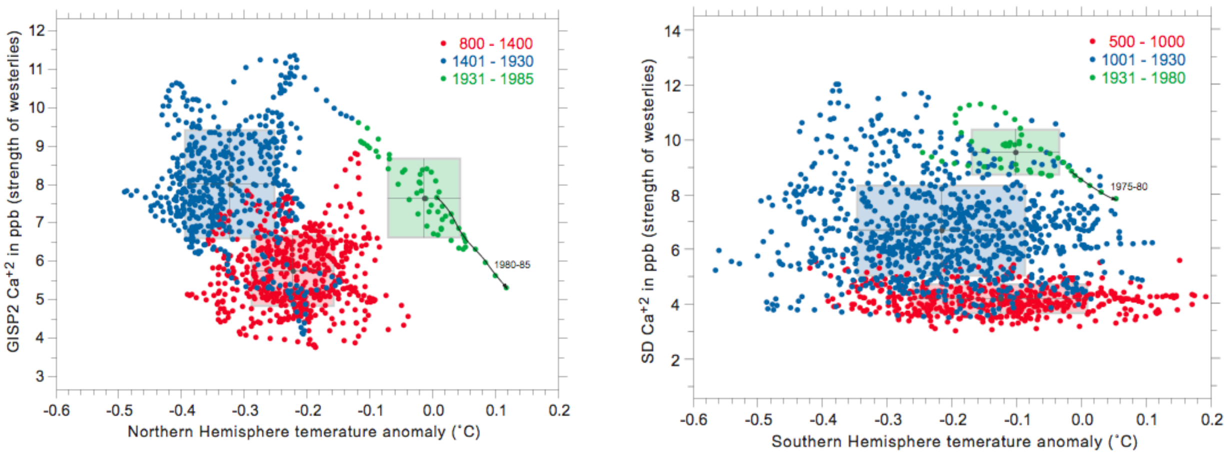


Figure 2.19. (Left side) Phase diagram for Northern Hemisphere temperature versus ice core proxy for Northern Hemisphere westerlies (time lines from Fig. 2.9.3). (Right side) Phase diagram for Southern Hemisphere temperature versus ice core proxy for Southern Hemisphere westerlies (time lines from 2.9.4). 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. Figures from Mayewski and Maasch (2006).

2.11 **Biological responses to climate change**

2.11.1 **The terrestrial environment**

With the exception of lake sediment studies (Hodgson et al. 2004), little terrestrial research has set out to examine changes in biodiversity, distributions and abundance over Holocene timescales. There has been a widely held 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 & Stevens 2007, Convey et al. 2008, Pugh & Convey In press), and it is becoming clear that across the continent and also the sub-Antarctic islands the contemporary biota is a result of vicariance (the separation or division of a group of organisms by a geographic barrier) and colonization processes that have taken place on all timescales between pre-LGM and pre-Gondwana-breakup. 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 & Chown 1999). Exceptions to this generalization are provided first by the continental Antarctic 'Dry Valleys' of Victoria Land, providing several thousand square kilometers 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.

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 – (Convey et al. 2008, Pugh & Convey In press), 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 & 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 are 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 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 (Convey et al. 2008), Fig 2.20. 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. 2006c). 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, 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 & Pugh 1996, Marshall & Coetzee 2000, Stevens & Hogg 2003, 2006), (Allegrucci et al. 2006, Cromer et al. 2006, Gibson & Bayly 2007) and dispersal (Hodgson et al. 2006c) for a variety of different species, and is based on the level of cosmopolitanism (dispersal model) or endemism (vicariance model) (Gibson & Bayly 2007), on direct palaeolimnological evidence, or, most recently, on molecular phylogenetic and evolutionary studies (Skotnicki & Selkirk 2006, Stevens & Hogg 2006)

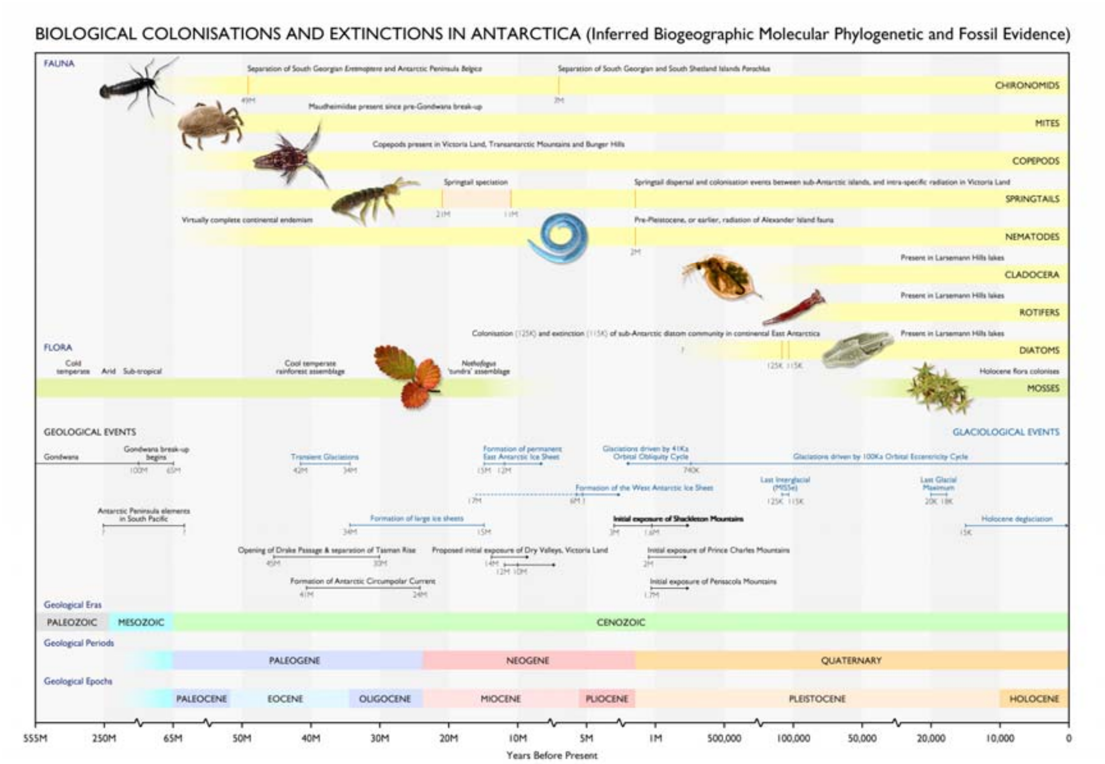


Figure 2.20. Biological colonisations and extinctions in Antarctica since the break-up of Gondwana based on molecular, phylogenetic and fossil evidence. The upper panel shows schematic timelines for the survival of different species on the continent. The lower panel shows the major geological and glaciological 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 will 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 & Convey 2004). 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 (Chown et al. 1998). For aquatic species, at least some groups such as the diatoms, diversity is controlled by the ‘connectivity’ among habitats with the more isolated regions developing greater degrees of endemism.

Changing species distributions, abundance and biodiversity

Through the Holocene the changing environmental conditions described above have caused marked changes in species distributions particularly for those species that have well defined ecological ranges. This is best recorded in palaeolimnolgical 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 has been recently observed in the Antarctic Peninsula region where increasing temperatures in the last 50 years have resulted in the southward migration of a number of plant and animal species and the establishment of new species that appear not to have survived on the continent before.

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 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 specialist 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 is the pattern of 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 area of major importance to ecosystems.

In contrast to many 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. 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.

2.11.2 **Biological responses to climate change: the marine environment**

The major impact of climate change on glacial timescales in the marine environment has been the glacial-interglacial expansion and contraction of the Antarctic ice sheet across the continental shelf and 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 (66ºS, 110ºE), in East Antarctica, where the expanding ice sheet resulted in habitat elimination followed by recolonisation and succession (Hodgson et al. 2003). Maximum expansion of the Antarctic ice sheet to the continental shelf edge at the Last Glacial Maximum was diachronous so it is likely that at any given time refuges were available for the biota. 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.

Warm periods and expansions and contractions of the sea ice have also had an impact on marine mammal and seabird distributions. For example changes in the Holocene distribution of marine birds can be tracked through the changing distributions of their nesting sites. Peaks in penguin populations indicate periods of open water in the Ross Sea between 4000 and 3000 corr.^{14}C yr BP (Baroni & Orombelli 1994), and 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) linked with a significant decline in sea ice and expansion of the elephant seal colonies (Hall et al. 2006). In the Bunger Hills isotopic concordance between the a marine sediment core and snow petrel mumiyo, and a significant correlation between mumiyo δD and δ13C, suggest that past δ13C variation in plankton was transferred through diet to higher trophic levels and ultimately recorded in stomach oil of snow petrels (Ainley et al. 2006). Divergence in signals during cold periods may indicate a shift in foraging by the petrels from 13C-enriched neritic prey to normally 13C-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. 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, some in opposite directions at different stages of breeding and life cycles.

Adaptation and evolution (marine vertebrates)

In the study of adaptive evolution, the role of molecular analyses is exponentially growing. The new powerful aid recently provided by molecular biology allows us to explore a number of important aspects, including the function of individual genes and genome sequencing. Nowadays molecular phylogeny has become essential for studying evolution, at the level of both protein and nucleic acid sequences. Hemoglobin (Hb), antifreeze glycoprotein (AFGP), myoglobin (Mb), are some relevant examples (Cheng 1998; Bargelloni et al. 1998, 2000; Near 2004; Sidell & O'Brien 2006; Verde et al. 2006a, 2006b; Giordano et al. 2007).

Notothenioid fish: a case study of speciation in the marine realm

Speciation is the evolutionary process by which new biological species arise. One peculiar form of speciation process is adaptive radiation, which describes the rapid diversification of one or few species into several distinct species filling many different ecological niches. A classical example of adaptive radiation is represented by the

Galápagos finches, described by Charles Darwin in the journal of his voyage onboard the Beagle. As for the Darwin's finches, other cases of adaptive radiation occurred in isolated environments (islands, newly formed lakes), where isolation releases the pressure of competition. In general, mass extinctions precede adaptive radiations vacating most ecological niches. A third element that characterises this speciation process is the presence of key innovations, which enable the species to diversify. As mentioned above, examples of adaptive radiation have been found in isolated ecosystems, both terrestrials (e.g. islands) and aquatic (lakes). Conversely, the marine realm, where barriers to movement of organisms are limited and water currents favour dispersal over long geographic distance, offers little opportunity for radiations to occur.

One possible exception is the evolution of a group of teleost fish, the Notothenioidei, in the waters surrounding the Antarctic continent. 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 notothenioid 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. Antarctica is separated from other continents by large and deep water masses, with no shallow water connections (sea mountain ridges, plateau). The presence of an oceanographic barrier, the Antarctic Polar Front, further reduces the exchanges between Antarctic and sub-Antarctic waters. This isolation was established after the opening of the Drake Passage (22-25 million years ago (mya) (Barker 2001) or 41 mya according to (Scher & Martin 2006) and the Tasmanian gateway, 35-33 mya (Stickley et al. 2004). A dramatic reduction in sea water temperatures, with presence of sea ice, occurred in the last 35 million years, to reach the present day -2°C. The subzero water temperatures and the progressive extension of ice sheets likely determined the extinction of most demersal fish species, leaving space to those species that evolved some adaptation to the freezing conditions.

This is the case of Notothenioidei, the majority of which possess specific "antifreeze" glycopeptides (AFGP) that prevent the formation of ice crystals in biological fluids. Analyses based on molecular markers confirmed that this was a "key adaptation", which evolved once in the ancestor of most notothenioid species (Bargelloni et al. 1994, Chen et al. 1997, Cheng et al. 2003). In turn, this enabled the ancestral notothenioid to fill the many ecological niches emptied by ice-driven extinctions. Indeed, estimates obtained using "molecular clocks" indicate that the main notothenioid diversification started 23-15 mya (Bargelloni et al. 1994, Near 2004), in parallel with the establishment of permanent sea ice. Isolation and extinctions, together with the evolution of AFGP provided the opportunity for nototheniods to radiate, but alone cannot explain the great number and diversity of notothenioid species. Other adaptations or modifications led to notothenioid radiation. The ancestral notothenioid was a benthic fish, yet acquisition of neutral buoyancy allowed the repeated colonisation of the pelagic habitats (pelagic, epi-pelagic, cryopelagic). Even for truly benthic species, diversification occurred through partition of depth range, from very shallow waters (0-30 m) to great depths (2950 m, *Bathydraco scotiae*). Variation in body size (from the few centimeters of *Pleuragramma antarcticum* to close of two meters of the large pelagic predator *Dissostichus mawsoni*) and diverse feeding habits further contributed to diversifying 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 circum-polar 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 nice example of how all the above described factors shaped the evolution of notothenioids is provided in a study by Chen et al. (1998). The authors mapped ecologies and geographic distributions of the species belonging to a notothenioid family, Channichthyidae, onto a molecular phylogenetic tree. It clearly emerged that speciation was always associated either with a shift in ecological habits (feeding behavior, depth range) or with disjunct geographic distribution. In conclusion, the evolution of notothenioids represents an extraordinary example of adaptive radiation in a marine environment, with all the canonical characteristics of this mode of speciation (isolation, mass extinctions, key adaptations, ecological shifts and habitat fragmentation).

The oxygen transport

Red-blooded Notothenioidei

The hematological features of many Antarctic Notothenioidei have been extensively investigated in the past few decades. Recently, the comparison of the biochemical and physiological adaptations of cold-adapted Antarctic notothenioids with sub-Antarctic and temperate species has been a tool to understand whether (and to what extent) an extreme environment has required specific adaptations (Verde et al. 2006b; di Prisco et al. 2007). Red-blooded notothenioids differ from temperate and tropical species in having fewer erythrocytes (one order of magnitude lower, and three orders lower in channichthyids, than in temperate fish) and reduced Hb concentration and multiplicity (none in channichthyids) (Eastman 1993). This may be advantageous in coping with increased viscosity of body fluids at low temperature, and finds partial compensation in the increased blood volume and higher cardiac output (Egginton et al. 2002). The erythrocyte Hb content is variable and in some species seems positively correlated with life style (Eastman 1993).

The family Channichthyidae (icefishes) and the reduced role of hemoglobin

The evolutionary development in icefishes of an alternative physiology based on Hb-free blood may adequately work in the cold for notothenioids in general. The benefits due to this loss include reduced costs for protein synthesis, simplified metabolic pathways and lower amounts of oxygen radicals. However, as pointed out by Pörtner et al. (2007), the shift from Hb-mediated oxygen transport to mechanisms based on diffusion may account for higher vulnerability of icefishes, and of notothenioids in general, to warmer temperatures. Functional specialisation to permanently low temperatures implies reduced tolerance of high temperatures, as a trade-off. Gene expression patterns and loss of genetic information, especially for Mb and Hb in Channichthyidae, reflect the specialisation of Antarctic organisms to a narrow range of low temperatures.

The loss of Mb and Hb in icefish, together with enhanced lipid membrane densities (accompanied by higher concentrations of mitochondria), becomes explicable by the exploitation of high oxygen solubility and low metabolic rates in the cold, where an enhanced fraction of oxygen supply occurs through diffusive oxygen flux. Icefish developed compensatory adaptations that reduce tissue oxygen demand

and enhance oxygen transport (e.g. modest suppression of metabolic rates, enhanced gas exchange by large, well-perfused gills and through a scaleless skin, large increases in cardiac output and blood volume). Oxygen delivery to tissues occurs by transport of the gas physically dissolved in the plasma. Recent studies highlight how the loss of Hb and Mb and, in particular, of their associated nitric oxide (NO)-oxygenase activity, may have favoured the evolution of these compensations in icefishes (Sidell & O'Brien 2006). Channichthyids diverged from other Antarctic notothenioids approx. 7-15 mya, but radiation of species within the icefish clade appears to have been confined to the last one million years (Bargelloni et al. 1994).

Why have icefishes alone taken such a radical course leaving the other Antarctic families with only partial reductions in Hb? Does Hb 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?

In high-Antarctic benthic notothenioids, a single Hb present in limited amount may be regarded as the consequence of its reduced role as oxygen carrier, possibly in relation to the fish life style and metabolism, and to the peculiarity of the environmental conditions, which comprises high stability/constancy of physico-chemical factors and high oxygen content (Verde et al. 2006b). The loss of Hb expression in icefish is the result of gene deletion, and the decrease of Hb content/multiplicity in red-blooded notothenioids is probably due to a down-regulation of an existing gene. The lack of multiple globin genes is not lethal in thermostable environments ("loss without penalty"), as shown in modern notothenioids.

Hemoglobins of Arctic fish

Historically, studies on the molecular mechanisms underlying fish biodiversity and thermal adaptations in cold extreme environments had found their natural scenario in the most extreme habitat on Earth, the Antarctic ocean. In recent years, the urge to extend these studies to the North Pole has become stronger. Polar research is becoming more and more focussed on the extent of warming occurring in the polar regions (Anisimov et al. 2007). There is already compelling evidence for widespread alterations in ecosystems due to climate change.

Arctic fish share some cold-water features with Antarctic fish, such as aglomerular kidneys and identical AFGPs (Chen et al. 1997a,b); other features, such as the neurophysiological function, do not show full cold adaptation. These differences may be due to the geological histories and shorter evolutionary time scale at polar temperatures. Unlike the vast majority of notothenioids, the Arctic ichthyofauna has high biodiversity and many species display Hb multiplicity. The possession of multiple Hbs provides a strategy for finely tuned regulation of oxygen transport in response to environmental variability and/or variations in metabolic demands.

The molecular evolution of hemoglobins in polar fish

The growing knowledge of the phylogenetic relationships among the notothenioid families 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 mya is the time of widespread continental glaciation and sharp drops in Southern-ocean surface temperatures (Eastman 1993; Eastman & Clarke 1998)].

The different phylogenetic histories of polar fish depend on the differences in the respective habitats. As a result of the isolation of Antarctica, the genotype of Notothenioidei diverged with respect to other fish groups in a way interpreted as a giant species flock (Eastman & McCune 2000). 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. The two diversities are reflected in phylogeny, as shown for instance by the amino-acid sequences of globins. Showing low identity with temperate species, the globin sequences in the Arctic liparid *Liparis tunicatus* and zoarcoid *Anarhichas minor* are consistent with species history that places cottoids (liparids) and zoarcoids close to notothenioids in the teleost phylogeny (Dettaï & Lecointre 2004, 2005). On the other hand, in one of the main Arctic families, Gadidae, the branching order of the globins does not reflect the expected phylogeny, and the sequences occupy different positions in the two phylogenetic trees with regard to temperate and Antarctic sequences (Verde et al. 2006a). The life style of benthic species such as *L. tunicatus* and *A. minor* is similar to that of Antarctic notothenioids, and such a similarity is mirrored to some extent by Hb evolution. The phylogenetic analysis shows that the constant physico-chemical conditions of the Antarctic ocean and the inactive life style of non-migratory Arctic species, such as *A. minor* and *L. tunicatus*, contribute to grouping their globins with the Antarctic globin sequences, allowing to recover teleost phylogeny (zoarcoids with notothenioids, gadids as sister-group to both) (Verde et al. 2002, 2006a; Giordano et al. 2007). In contrast, the variations typical of the Arctic ocean, in conjunction with the pelagic and migratory life style of gadids, correspond to high divergence of their globin sequences.

2.12 Concluding statement

To be inserted after SCAR feedback

Draft notes: Add in comments from P. Barrett / J. Francis / E. Steig

Chapter 3

The Instrumental Period

Coordinating Lead Author:
John Turner

Lead Authors:
John Turner, Jonathan Shanklin, Mike Sparrow, Rob Arthern, David Bromwich, Andrew Fleming, Gareth Marshall, Tony Worby, James Bockheim, Guido di Prisco, Cinzia Verde, Pete Convey, Zhaomin Wang, Howard Roscoe, Anna Jones, David Vaughan, Philip Woodworth, Ted Scambos, Alison Cook, Anthony Lenton, Adrian Jenkins, Josifino Comiso, Mauro Gugliemin

Contributing Authors:
Colin Summerhayes, Mike Meredith Alberto Naveira-Garabato, Steve Chown, Stevens, Adams, Roger Worland, Françoise Hennion, Ad Huiskes, Dana Bergstrom, Dominic Hodgson, Robert Bindschadler, Roberto Bargagli, Nicolas Metzl, Todd Arbetter, Kees van der Veen, Andrew Monaghan, Kevin Speer, Steve Rintoul, Hartmut Hellmer, Stan Jacobs, Karen Heywood, David Holland

Review Editors:

3.1 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 meteorology of the continent, ocean conditions, the sea ice extent and the terrestrial and marine biology. However, the collection of data was sporadic and it was not possible to venture to any great extent into the inhospitable interior of the Antarctic.

At the start of the Twentieth Century stations were first operated year-round and this really started 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 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 50 year records on 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 which have mostly been made during the summer months. Satellite observations can help in monitoring the surface of the ocean, but not the layers below. And even here a quantity such as sea ice extent has only been monitored since the late 1970s when microwave technology could be flown on satellite systems.

3.2 Observations, data accuracy and tools

3.2.1 Introduction

In this section we consider the various types of data and models that are available to investigate climatic and environmental change in the Antarctic over roughly the last 50 years. We examine the availability and accuracy of in-situ physical and biological data, along with the information that can be obtained from satellite systems that have become of increasing importance over the last couple of years.

Mathematical models have been applied increasingly in Antarctic research. Initially their applications was in the more physical areas of climate, ice sheet, sea ice and ocean modelling. However, increasingly they are being applied in more biological area – a trend that seems set to continue in the future.

3.2.2 Meteorological Observing in the Antarctic

The IGY provided a big impetus towards setting up continuously operated stations in the Antarctic and over forty were established, of which over a dozen are still operating today. 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 at coastal sites, primarily so that stores can easily be transported ashore. This means that in some ways their weather is not a true representation of the continent as a whole, as they are much milder due to the influence of the sea. 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, however there is increasing reliance on automatic systems during the "night". Surface temperature, humidity, sunshine, pressure, wind speed and direction are largely measured by automated instruments but an observer is needed to estimate the 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 keep note of the weather: rain, snow, fog, gale etc. as well as 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. At the Russian Vostok station special suits were worn for outdoor work under these conditions.

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.

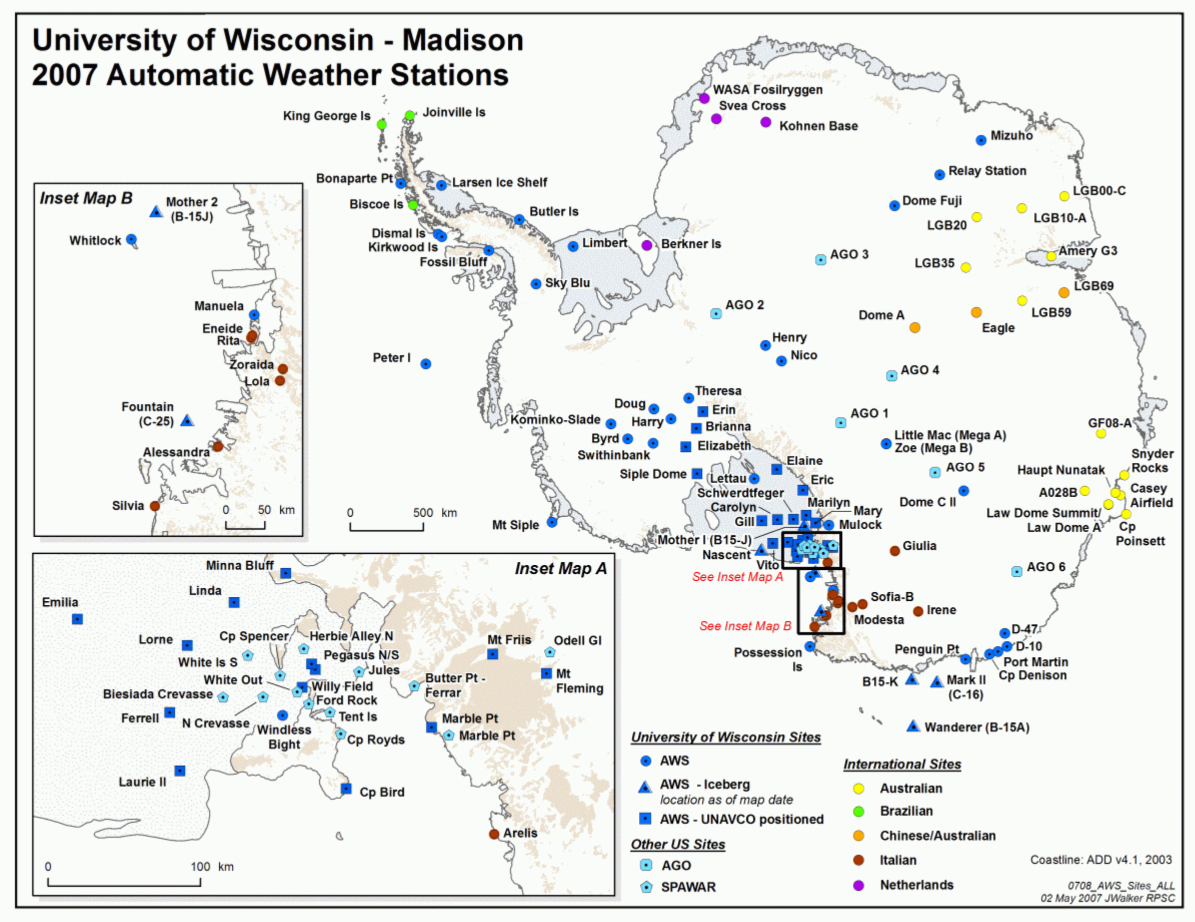


Fig 3.1. Location of AWS sites [Source: University of Wisconsin – Madison]

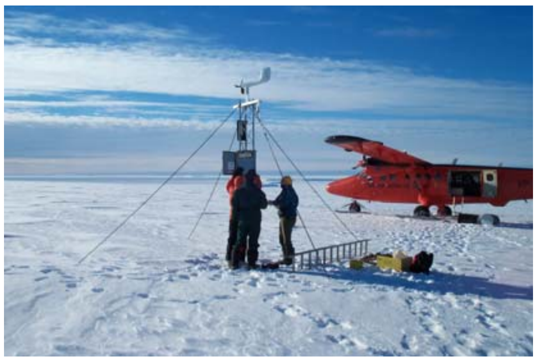


Fig 3.2 Servicing the AWS at Butler Island (Jon Shanklin)

Automatic stations (Figs. 3.1 and 3.2) generally measure a reduced range of parameters, usually just pressure, temperature and wind, although some may measure humidity and have house-keeping data such as snow depth. 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 give a big improvement in the measurements, and can also monitor precipitation falling from clouds.

The Antarctic atmosphere is very clear, as there are few sources of pollution. On a fine 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, however some have difficulty in discriminating variation in visibility above 20km.

At around 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 the temperature, humidity and pressure to an altitude of over 20 km, with wind speed and direction found by tracking the package with GPS sensors. One particular problem affects some balloons during winter: the combination of low ambient temperature and darkness 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 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.

The ozone hole was discovered via ground-based observations from Antarctica, and most manned stations continue with long term measurements of the ozone column. Ground based sensors include the traditional Dobson ozone spectrophotometer, the Brewer spectrometer and the SAOZ (Système d'Analyse par Observation Zenithales) spectrometer, or variants of these. All use the Sun as a source and measure the differential absorption of light as it passes through the ozone layer. At a few stations ozone sondes are flown, which give precise profiles of ozone in the atmosphere. These bubble air through a cell, which generates a current that is proportional to the amount of ozone present.

Observational problems

Although many of the 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 what 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, and hence under recorded wind speeds. Propeller type combined vanes and anemometers suffer less from rime, but can suffer mechanical failure. 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 however suffer problems, particularly in conditions of heavy blowing snow, when the pinging of snow grains on the sensor saturates the detector.

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 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 round both problems and allow continuous recording, although 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.

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 that have, where possible, been obtained from the operators who run the Antarctic stations. The observations have been highly quality controlled and are on the web. For more details see http://www.antarctica.ac.uk/met/READER/.

3.2.3 **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).

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.

Fig. 3.3 Water sampler-CTD package being lowered into the water (source 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 in the ocean when the ship stops on station (Fig. 3.3). 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, transmisomiters 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 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.

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 250m 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* (2004) were able to determine the three-dimensional structure of hydrographic, chemical and biological variables in the Southern Ross Sea. Other towed systems such as the Continuous Plankton Recorder are also used routinely in the Southern Ocean (e.g. *Hunt and Hosie, 2003*).

Autonomous vehicles

The types of autonomous vehicle vary widely, depending in part on the nature of propulsion, and the sensor payload. AUV's (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.
As an example, the AUV AutoSub-2 (Fig. 3.4) 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.

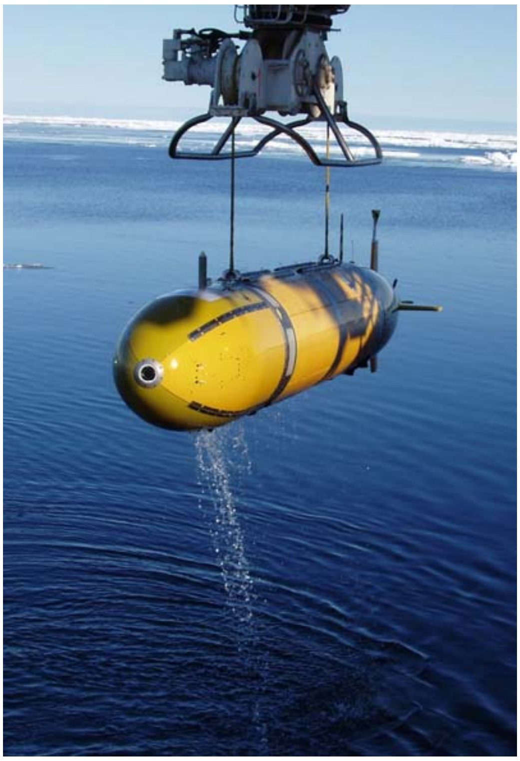


Fig. 3.4. Autosub (source Gwyn Griffith)

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).*

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 *(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.

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 this has meant experimenting with the use of 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).*

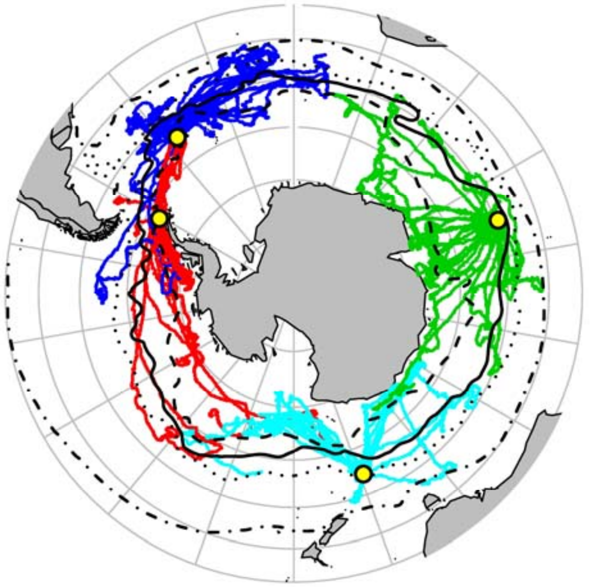


Fig. 3.5 Tracks of elephant seals tagged with instruments to record temperature and salinity profiles in the Southern Ocean. (source Biuw et al., 2007)

Other methodologies

Satellites underpin a vast amount of modern oceanography, and their utility is nowhere greater than data-poor regions of the Southern Ocean. 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.

The use of novel techniques, such as the deployment of small sensors on wide-ranging marine animals (Fig. 3.5) 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*).

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 center 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. Fig. 3.6). 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 choices of parameters.

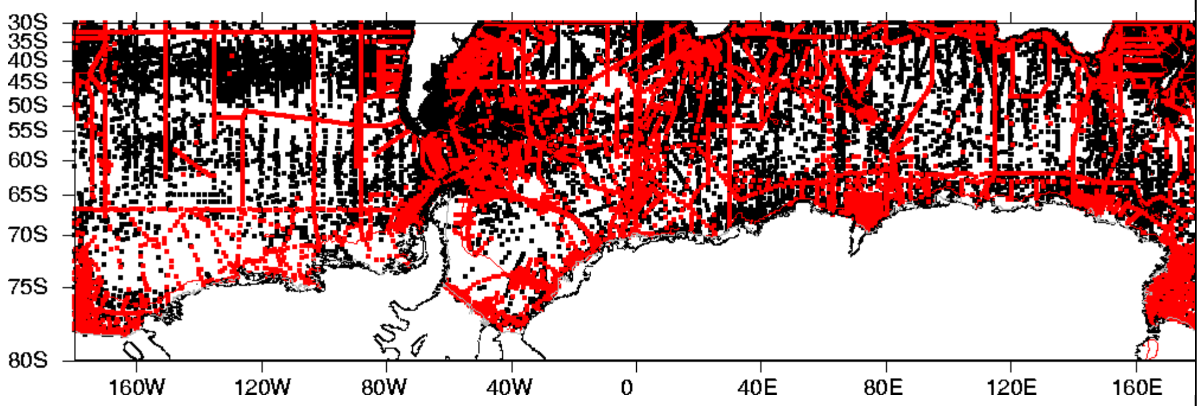


Fig. 3.6. Example figure produced using the Southern Ocean Atlas (http://woceatlas.tamu.edu/). Dots show all data from 1900 to present, including bottle data. Red dots are CTD data only.

Southern Ocean READER (REference Antarctic Data for Environmental Research) 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/). Links to the data and a suite of visualization resources are available (see Fig. 3.7).

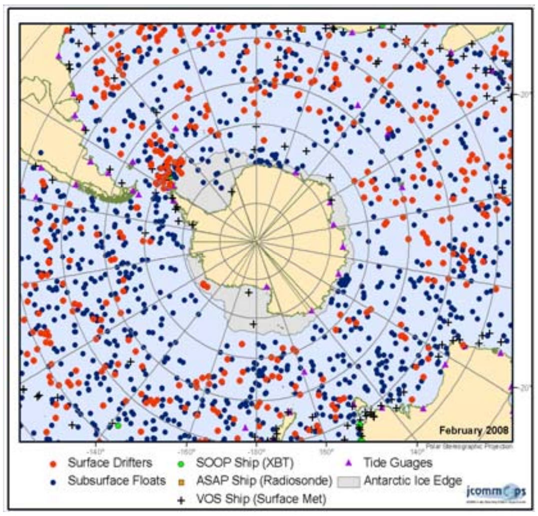


Fig. 3.7a. Observations available in the Southern Ocean region from JCOMMOPS (as of Feb. 2008).

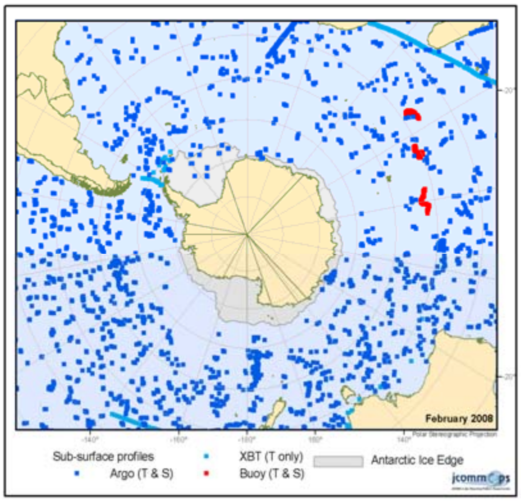


Fig. 3.7b.Sub-surface profiles available in the Southern Ocean region from JCOMMOPS (as of Feb. 2008).

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.

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-thethered profilers and the inclusion of sensors on marine mammals that forage under the ice. Argo is currently also limited to the upper 2000m. 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, 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 - 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.

3.2.4 **In-situ glaciological observations**

Transition from exploration to science

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, has continued to the present International Polar Year, 2007. In the intervening decades, glaciological fieldwork adopted new technology, and revealed behaviour that could not have been observed by any other means. Some measurements can now be made over wider 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 cause.

The geometry of the ice sheet.

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; Siegert et al. 2005).

Airborne radar, sometimes flying from remote field camps, is now the principle method of determining ice thickness. 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. A recent compilation places the total volume of the Antarctic ice sheet 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).

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 modelling the evolution of the ice sheet.

The glaciological budget and sea level

One of the central questions of Antarctic glaciology is how rapidly the ice locked up in the ice sheet can be transferred to the oceans to affect sea level (Nerem et al. 2006). At the surface of Antarctica, few places are warm enough for melting and runoff, even in summer. The snow that falls is simply buried and compressed as fresh snow accumulates upon it. Eventually the individual snow grains sinter together to form glacier ice, which flows to the coast. Much of this ice creeps along a few tens of metres each year as a thick, slow-moving sheet. Some flows faster through mountainous terrain as narrow outlet glaciers, and some drains rapidly as ice streams, flanked by slower moving ice. Ice streams and outlet glaciers reach speeds of several kilometers per year, making these the most dynamic components of the ice sheet.

When it reaches the coast, the ice flows directly into the ocean. There it either remains attached, floating as an ice-shelf, or breaks off as icebergs. Any imbalance between the loss at the coast and the replenishing snowfall inland will cause the ice sheet to grow or shrink over time, with consequences for sea level worldwide.

A measure of how quickly ice is transferred from ice sheet to ocean is gained by summing the total accumulation of snow upon the ice sheet, then subtracting losses to icebergs, floating ice-shelves, or coastal melt. Typically this budget is applied to ice grounded on rock or sediment, because gain or loss of floating ice does not affect sea level directly. Despite this, there can be an indirect effect: because floating ice shelves impart forces on grounded glaciers and ice streams, they slow the delivery of ice to the ocean. 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. The focus of many glaciological investigations is to observe and quantify the ice budget, the effect of ice shelves, and predict how these will evolve in future.

In-situ observations of ice-flow can be made by repeatedly surveying the position of marker poles carried along by the ice. Nowadays this surveying makes use of precise locations derived from Global Positioning System (GPS). Point velocities measured in this way provide constraints to calibrate remote sensing methods such as feature tracking or satellite radar interferometry (discussed in section ??). Modern dual-frequency GPS receivers and post-processing of data can give positional accuracy of centimetres or better (King, 2004). Ice motion is almost everywhere greater than a metre per year, so surface velocities are measured within one percent in a single year. For faster flowing ice, the relative accuracy is even higher.

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 order kilometres, the depth error from radar surveys is of order 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 order ten percent of the turnover (e.g. Fricker et al. 2000; Rignot and Thomas 2002).

The vertical component of position derived by GPS is less accurate than the horizontal. Nevertheless, direct estimation of the rate of thickening or thinning can be derived from repeated GPS surveys (Smith et al. 1998; Hamilton 2005). 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.

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 (Bindshadler 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.

Ice streams slide on water-saturated glacial sediment. This can provide extremely good lubrication; drag from the edge can be comparable to drag from beneath, even when the width of an ice stream greatly exceeds its thickness. In recent decades, glaciologists have made important discoveries concerning the rapidity of ice stream flow, and its inconstant behaviour (e.g. Shabtaie and Bentley 1987); the widespread and filamentary nature of ice streams and their tributaries (Joughin et al. 1999; Bamber et al. 2000); the extent of coastal thinning and acceleration in the Amundsen Sea sector and the Larsen Ice Shelf systems (Shepherd et al. 2001; Rott et al, 2002), and the rapidity of response to various forcings (Truffer and Fahnestock, 2007). Much of this dynamism can ultimately be traced to the slipperiness of basal sediment pressurised by meltwater. A directed program of fieldwork has illuminated the causes of this fast and changeable flow (e.g. Alley and Bindschadler, 2001; Bindschadler 2006).

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 floatation (Kamb 2001). This lessens the load on the underlying sediment and allows it to deform easily, lubricating the sliding. 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). This variability makes it difficult to predict their discharge to the oceans.

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 fluidised, or whether they are lodged and strong enough to support large shear stress retarding the flow of ice (Smith, 2007b).

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 linkages between and flow through 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).

Sediment can be transported beneath ice streams (Alley et al. 1987). Recent observations and modelling suggest that wedges of sediment deposited near the grounding line may be important in stabilising the ice sheet against sea level rise (Anandakrishnan et al. 2007; Alley et al, 2007). This stabilistation 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.

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 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.

Rate of 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. 2005; Frezzotti et al. 2005; Monaghan et al. 2006). 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. 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. 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 (e.g. Vaughan, D. G., Bamber, J. L., Giovinetto, M., Russell, J. and Cooper, A. P. R. 1999. Reassessment of net surface mass balance in Antarctica. Journal of Climate 12: 933-946) of the mean annual snow accumulation rate. 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). In future, ice penetrating radar flown on aircraft or satellites may allow more complete coverage of the ice sheet (Hawley et al. 2006).

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.

The historical flow of ice

Radars that operate at low frequency can see deep into the ice sheet. At depth the layering within the ice is affected by the history of how it has deformed over time. Comparing these deeper layers with models of ice deformation provides a view to events in the distant past: different candidates for the ice evolution can be tested against the observed layer architecture until a good match is found. This method works well for slow moving ice and small surface slopes found near ice-divides (the equivalent of watersheds). In certain places, a picture of how the ice sheet retreated after the last ice age can be pieced together from this type of measurement (Conway et al. 1999; Martín et al, 2006). More recent changes, such as transient changes in the flow of ice, or episodic floatation, can also be inferred from field-based radar measurements (Catania et al. 2006). Other clues are provided by dating the exposure of boulders discarded by the retreating ice, through the effect of cosmic rays upon their isotopic composition (this and other geological methods are discussed in section ??). If similar methods can be used to find the flow history most consistent with the large number of layers seen in airborne radar surveys (Siegert et al. 2005), this would place a valuable constraint on the historical behaviour of the ice sheet.

Field observations specific to ice shelves

The flow of ice-shelves, their integrity, and the forces they exert on upstream glaciers depend upon their mechanical properties and temperatures as well as their geometry. Drilling allows temperature profiles through the ice to be recorded, and samples of the ice to be recovered for mechanical tests (e.g. Rist et al., 2002). 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 satellites.

3.2.5 **Met Analyses**

3.2.6 **Satellite Observations**

Sea Surface Temperature

Sea surface temperature (SST) 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 insight into atmospheric and oceanic circulation patterns and anomalies (such as El Niño). On a 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 datasets. Observations began in November 1981 using data collected by the Advanced Very High Resolution Radiometer (AVHRR) instruments on the NOAA Polar-orbiting Operational Environmental Satellites series. 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.

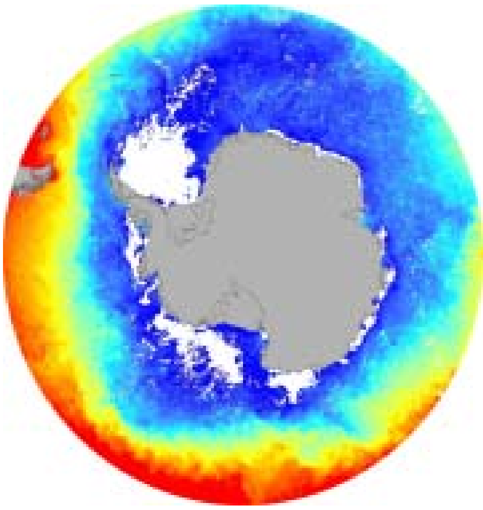


Figure 3.8. Southern Ocean averaged SST for January from Aqua MODIS sensor

The direct broadcast capability of instruments such as the AVHRR series 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 improves the coverage for the Southern Ocean that sees little data due to significant cloud and ice cover.

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 (<10km) SST products to the operational oceanographic, meteorological, climate and general scientific community.

Satellite SST data is also available from a range of other satellites including the European Space Agency ATSR instruments and more recently the MODIS instruments on the NASA Terra and Aqua satellites (Fig. 3.8).

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

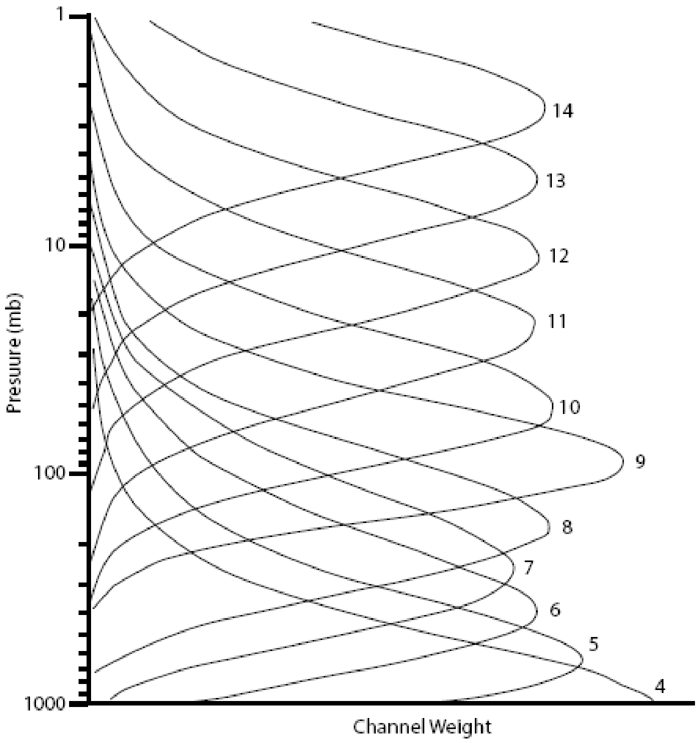
In addition to SST data derived from thermal infrared measurements, observations are also made using passive microwave measurements. Accuracy and resolution is poorer for SST derived from passive microwave measurements, however the advantage gained with passive microwave is that radiation at these longer 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 Advanced Microwave Scanning Radiometer (AMSR) instrument on the NASA EOS Aqua satellite.

Satellite Sounding

Satellite sounding instruments measure radiation at infrared or microwave wavelengths that have 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 made a huge difference in improving our knowledge of the climate of the polar regions and hence the forecasts of Numerical Weather Prediction (NWP) models. Today, regions such as the Southern Ocean, previously a huge data void, are filled by gridded data with a spatial resolution of tens of km. 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 different wavelengths that have different absorption characteristics a complete profile of temperature through the atmospheric column can be derived (cf. Fig. 3.9).

The principal satellite sounding instruments are the TIROS-N operational vertical sounder (TOVS) sensor system, which has flown on polar-orbiting satellites since 1978, and its 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, while HIRS/3 replaces the HIRS/2. Significant improvement in forecast accuracy has come from utilisation of 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). However, similar problems have been identified in the ERA40 reanalysis prior to 1996 (Bromwich *et al.*, 2007). In the Antarctic there is the additional problems that the high 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).



Weighting functions for channels 4-14 of the AMSU-A instrument.

Figure 3.9. AMSU-A weighting function diagram

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) (Fig. 3.10). Satellite altimeters provide long-term observations of mesoscale circulation patterns in the Southern Ocean and the variability of features such as the Antarctic Circumpolar Current (ACC). Satellite altimetry also provides information that is of use in mapping sea surface wind speeds and significant wave heights.

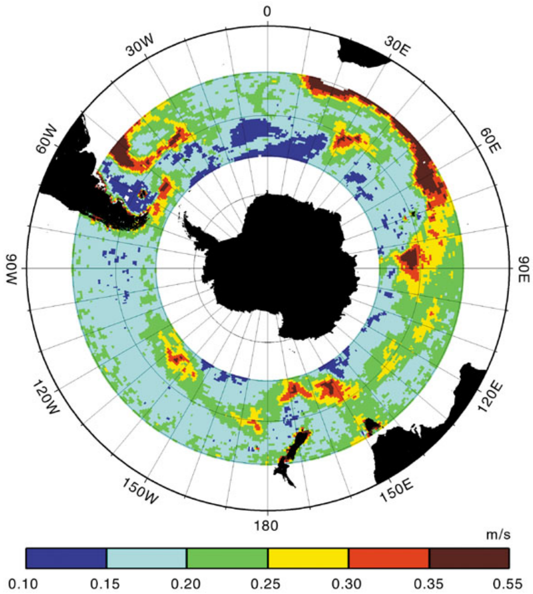


Figure 1.10. Southern Ocean eddy kinetic energy from satellite altimetry

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 smaller freeboard measurements of dominant first year ice in the Antarctic (Giles, K. A. 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 has been collected since 1978 from a variety instruments including Seasat (1978), Geosat (1985-1990), 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 (planned for 2008 launch) and the ESA Sentinel satellites. ESA's CryoSat mission will also provide new data from the SIRAL instrument specifically designed for ice sheet and sea ice observations.

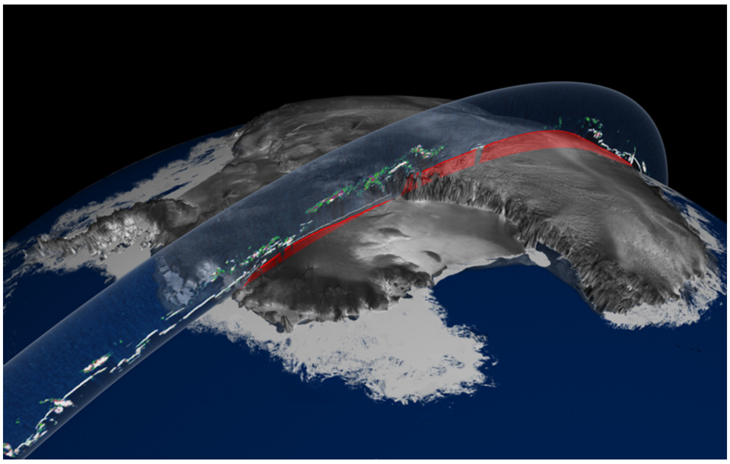


Figure 3.11. Icesat orbital track indicating cloud observations.

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) (Fig. 3.11). The main objective of the ICESat mission is to measure ice sheet elevations and changes in elevation through time. Secondary objectives include measurement of cloud and aerosol height profiles, land elevation and vegetation cover, and sea ice thickness (Schultz et al., 2005).

Ocean colour

Satellite ocean colour sensors provide data on the concentration of chlorophyll in the ocean. 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 (Fig. 3.12) 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 Nino/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).

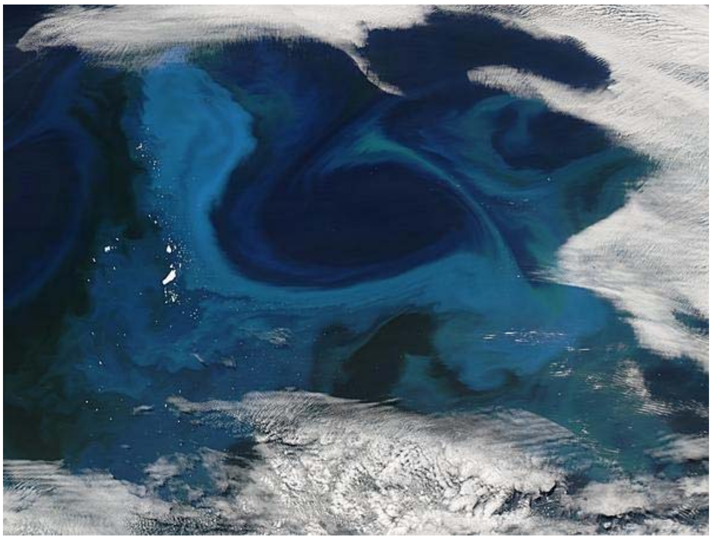


Figure 3.12. Phytoplankton bloom off South Georgia in the Southern Ocean acquired by the Aqua MODIS instrument

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 highlights 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. Therefore satellite measurements of chlorophyll help obtain a comprehensive understanding 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 under-estimate 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.

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. The latter is 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, 1997b; Marshall and Turner, 1999; McMurdie *et al.*, 1997) and coastal katabatic winds off the Ross Ice Shelf (Marshall and Turner 1997b).

Figure 3.13, 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 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.

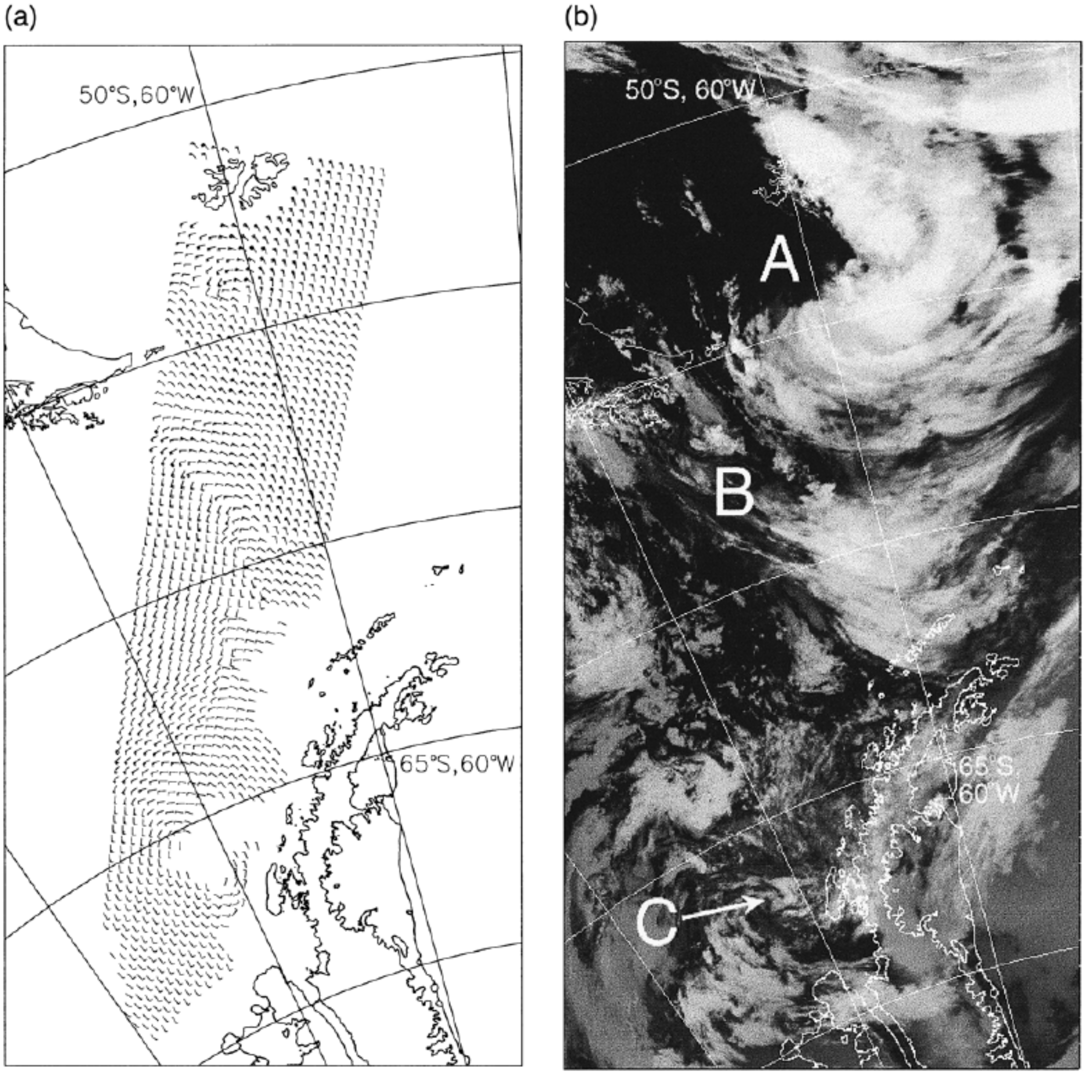


Figure 3.13. Near-coincident data over the Southern Ocean on 15 Jan 1995. (a) ERS-1 scatterometer swath acquired between 13:09 and 13:15 UTC; (b) Thermal infrared Advanced Very High Resolution Radiometer (AVHRR) imagery obtained at 13:11 UTC. Wind feathers point in the direction the wind is blowing; half a barb represents 2.5 ms^{-1}, and a full barb 5.0 ms^{-1}. Labelled features are described in the text. (Marshall and Turner, 1999; courtesy of AMS).

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 (Fig. 3.14). 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 & snow and the oceans.

Sea ice studies benefit from information about 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 for them to assist ship navigation in sea ice.

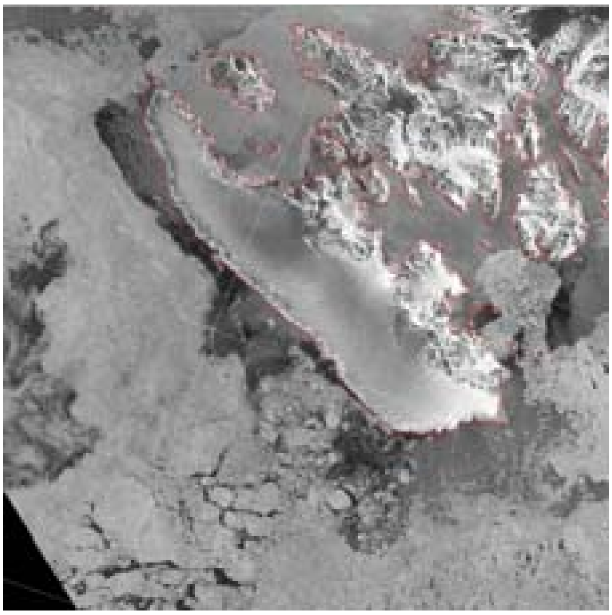


Figure 3.14. Envisat ASAR image showing sea ice around Adelaide Island, West Antarctic Peninsula

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.

Much information about land ice and snow are also 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 (Fig. 3.15). These new techniques have provided information to monitor glacier discharge (Pritchard & Vaughan, 2007), map grounding lines (Gray et al, 2002), contribute to mass balance studies (Rignot & Thomas, 2002), determine glacier & ice sheet motion (Fishcer et al, 2003) and monitor propagation of ice shelf rifts (Fricker et al, 2002).

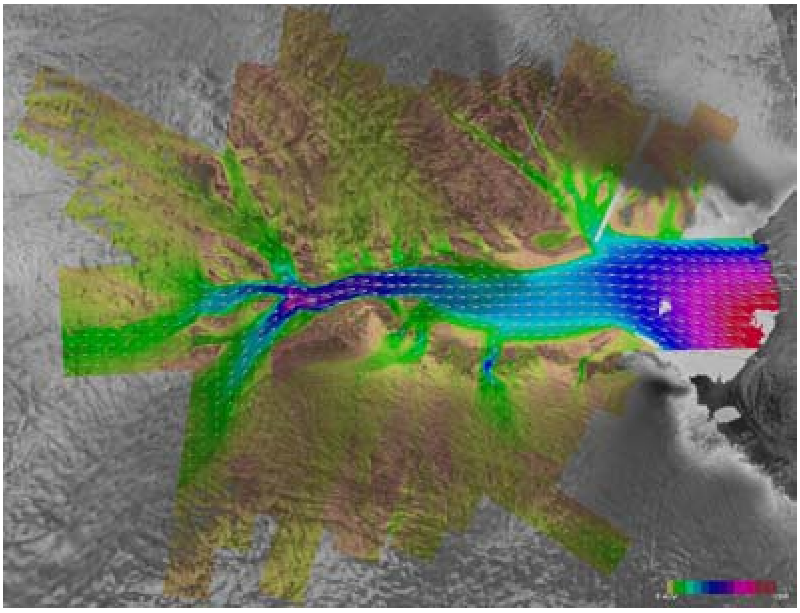


Figure 3.15. Velocity of the Lambert glacier, Antarctica, from RADARSAT Synthetic Aperture Radar (SAR) imagery from the 2000 Antarctic Mapping Mission

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. This resulted in the RAMP dataset (http://bprc.osu.edu/rsl/radarsat/radarsat.html) which has subsequently been updated in 2002 with the MAMM (Modified Antarctic Mapping Mission) dataset in 2000. Recently other SAR systems with different wavelength radars have been launched including the Japanese ALOS and TerraSAR-X systems. Into the future there are plans for continuity of the European SAR systems from the ESA Sentinel series, 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.

Optical imaging systems

Landsat and similar optical satellite instruments have provided images of the Antarctic ice sheet since the 1970s. In recent years newer instruments such as ASTER and Quickbird have provided more imaging opportunities and the opportunity for higher spatial resolution. These images reveal important details about the Antarctic ice sheet, geology and even limited vegetation cover. Recently the Landsat Image Mosaic of Antarctica (LIMA) (http://lima.usgs.gov) has provided the first consistently processed surface reflectance dataset of the continent from Landsat ETM+ images.

Certain optical imaging systems provide stereo image data of the surface from which elevation data can be extracted using established photogrammetry techniques. Limitations in availability of suitable stereo data have been partially solved with the advent of along-track stereo imaging systems such as the ASTER and SPOT5 instruments. Elevation data from these sources may be used on studies of geology, geomorphology and potentially in ice sheet surface changes.

Future EO missions relevant to the Antarctic

A number of other satellite missions are scheduled for launch in coming years, many of which will have direct benefit to 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 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
Scheduled to launch in 2008, 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. Additionally these data will 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.

3.2.7 **Satellite Observations – Covering comparable satellite data that we can use to investigate mass balance.**

The value of satellites

The availability of satellites has transformed the study of the Antarctic ice sheet (Payne et al. 2006). Until the era of satellites, large-scale changes in the ice sheet were considered likely to proceed over thousands of years. Now, significant accelerations and decelerations of large outlet glaciers and ice streams have been observed on much shorter timescales (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 on 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.

Glacial retreat of the Amundsen Sea sector

In the Amundsen Sea sector, satellites have been used to measure a retreat of the grounding line of Pine Island glacier (Rignot 1998). This retreat has been accompanied by thinning of the ice sheet at a rate of 10 cm per year averaged over a drainage basin twice the area of Great Britain (Wingham et al. 1998). Thinning rates reach well over a metre per year at the coast (Shepherd et al. 2001). This rapid thinning was detected by satellite radar altimetry, which records the travel time of a microwave pulse from satellite to surface to give the range to the surface; orbital tracking, or GPS, providing a height reference for the satellite. Repeated observations can detect elevation changes, and hence thickening or thinning of the ice sheet.

Similar measurements can be obtained using satellite laser altimetry (Smith et al. 2005; Csatho et al. 2005). The laser 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 thinning of the ice in the Amundsen sea sector occurred because of an increase in the discharge of several major outlet glaciers. From 1974 to 2000 Pine Island Glacier accelerated by about 20%, undergoing two distinct accelerations separated by a period of steady flow (Rignot et al. 2002; Joughin et al. 2003). A widening of the fast-flowing portion of Thwaites glacier increased its ice discharge, by about 4% between 1996 and 2000 (Rignot et al. 2002). These changes were detected using the technique of satellite radar interferometry (Goldstein et al. 1993), which has allowed widespread observations of flow speed in Antarctica. The method relies on an interference effect, brought about by differences in the path length travelled by microwave radiation from satellite to surface and back. This path length changes as the scatterers in the snow are carried along with the ice flow. A geometric correction that depends on the shape of the surface and the known position of the satellite between revisits can be removed, allowing the relative motions within the image to be recovered to a precision of a few centimetres. If a fixed velocity reference within the image is available the absolute velocity field can be recovered. Mosaics of such images allow the velocity field across a substantial fraction of the ice sheet to be mapped. Where thickness measurements are available at the coast, the fluxes leaving the ice sheet can be determined using interferometry, and compared with snowfall inland to give the ice budget (Rignot and Thomas, 2002).

More recently, measurements of changes in the Earth's gravity field have been made using the 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.

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 (we have 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 with 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 entres 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. 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. Meanwhile East Antarctica seems close to balance, or thickening slightly, according to these independent assessments.

The collapse of the Larsen Ice shelf

Identifying the chain of events that has caused the glaciers that drain into the Amundsen Sea Embayment to speed up is still an active area of research, but new measurements from this region and elsewhere in Antarctica provide compelling evidence that these changes have their origins at the coastal margins. Some inferences can be drawn from a sequence of events observed over 1000 km further north. 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, 2003). A MODIS image taken on March 7th 2002 shows a plume of icebergs being ejected, clearing the bay that was previously occupied by the ice shelf. Subsequently, glaciers inland that had flowed into the Larsen-B ice shelf accelerated, almost along their entire length, shown by satellite radar interferometry using the ERS-1, ERS-2 and Radarsat-1 satellites (Rignot et al. 2004), and by tracking the motion of features using imagery recorded by the Landsat-7 satellite (Scambos et al. 2004). A similar effect had been observed following the collapse of the Larsen-A ice shelf (Rott et al. 2002). After the Larsen-B collapse, Hektoria, Green and Evans glaciers flowed eight times faster than they had previously (Rignot et al. 2004).

The question of whether inland glaciers would speed up following the removal of an ice shelf has been a concern of glaciologists, since it was suggested in the 1970s that the junction between a floating ice shelf and a grounded ice sheet could become unstable (Hughes 1973; Weertman 1974). Mercer (1978) suggested that such a process could lead to complete collapse of West Antarctica, if the ice shelves around its periphery fell victim to rising temperatures.

The role of the oceans

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; Bindshadler, 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. 2004).

Other changes catalogued from space

The Amundsen Sea sector and the Larsen Ice Shelf on the east coast of the Peninsula are the most obvious regions of change in Antarctica, but changes have also been observed elsewhere. Several major outlet glaciers in East Antactica are thinning at their coastal margins, in particular Cook and Totten glaciers (Shepherd and Wingham, 2007). In contrast, the interior part of East Antarctica thickened slightly from 1992-

2003, most likely because of year-to-year and decade-to-decade fluctuations in snowfall (Davis et al. 2005).

Satellites provide images of the ice sheet in the visible band of the electromagnetic spectrum (Landsat, MODIS, Spot), the infra-red (AVHRR, MODIS), and in microwave bands sensitive to thermal emission form the surface (SMMR, SSM/I, AMSR-E), or active illumination (Scatterometers and Synthetic Aperture Radar). These satellite images can show how the coastline of Antarctica has changed (e.g Ferrigno et al. 1998). In the Antarctic Peninsula 87% of 244 glacier termini surveyed retreated between 1940 and 2001 (Cook et al. 2005). Most of the glaciers on the west side of the Peninsula have also accelerated, which has been revealed by tracking the motion of crevasses using satellite radar imagery (Pritchard and Vaughan, 2007). The reason for this acceleration could be loss of friction at their base as they have thinned and approached floatation (Vieli et al. 2001). In the same way that radar layers within the ice can be used to infer variations in the historical behaviour, flow stripes observed from space can provide a similar constraint on past events, particularly the activation and deactivation of ice streams (Bindschadler and Vornberger, 1998; Hulbe and Fahnestock, 2007).

Satellite observations using radar altimetry, laser altimetry and radar interferometry have shown that large amounts of water can move beneath the ice sheet in weeks or months (Gray et al 2005; Wingham et al. 2006; Fricker et al. 2007), and this may also affect the dynamics of the ice flow by altering the lubrication at the base.

Satellite observations of snow properties

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 part of the electromagnetic 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 et al. 2000). This temperature information has proved useful in glaciological applications, including modelling the snow compaction (Li et al. 2007).

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 using 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 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.

3.2.8 Sea ice observations

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 sea ice edge. The region close to the ice edge is rich in food and many whales congregate here, attracting the whaling fleets. Most of these observations are for the summer season. De la Mare (Nature 1997 389 p57) 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 move southward by 2.8 deg of latitude between the mid 1950s and the early 1970s. He inferred this as meaning that there had been a decrease in the area covered by sea ice of 25%. However, there is a great deal of debate over how the locations of whale catches can be translated into ice edge estimates that are comparable to those 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 sea ice extent decrease between the 1940s and 1970s.

Coastal stations have also made valuable sea ice observations, albeit from a very limited number of locations. However, the ice edge is over the Southern Ocean for much of the year, well removed from most coastal observatories making direct observation difficult. 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 (Ref).

Recently sea ice extent has started to be determined from ice core data collected in the coastal region of the Antarctic. An analysis of the Law Dome methanesulfonic acid (MSA) record has proved a proxy record of the sea ice extent in the 80 E to 140

E sector extending back to 1841 (Curran et al 2003 Science p1203). The good correlation ($P < 0.002$) between MSA and satellite-derived ice extent in the 22 year overlap period indicates the reliability of the technique at this location. Unfortunately, such MSA/sea ice relationships may not be so robust at other locations around the Antarctic and the techniques may have to be tuned for different sectors of the continent.

Satellite observations of sea ice extent and concentration

From the 1960s it was possible to obtain a broader-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. It has therefore not been possible to investigate broadscale sea ice variability before the 1970s using visible or infra-red imagery.

The extent and concentration of Antarctic sea ice is only known with confidence since the early 1970s when reliable satellite passive microwave observations became available.

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 of the growth and decay patterns of sea ice for the entire Antarctic region.

It was hoped that there would be further developments in sea ice monitoring was the launch in 1975 of Nimbus-6/ESMR-2 with its dual polarized 37 GHz radiometer. However, the instrument did not perform well and no useful data was obtained. But 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 the multifrequency capability, ice concentrations were derived at a much better accuracy with this data than with ESMR data (Gloersen et al., 1993; Comiso 1995). SMMR lasted for about 9 years and luckily, 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 are the same set used for generating ice concentration maps from SMMR. The sensor is also conically scanning with similar resolutions but a wider swath and 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 also 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. Secondly, it is a horizontally scanning radiometer going from nadir to around 50^o with varying resolution and with different incident angles. Third, there were lots of missing bits in the data stream causing the elimination of a large fraction of the data and therefore 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 xxx, 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 will 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.

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 spend 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 [*World Meteorological Organization*, 1970] 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 1663 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://www.aadc.gov.au).

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.

3.2.9 Observations on Antarctic Permafrost

Permafrost distribution and form

Permafrost is defined as ground that remains below 0ºC for two or more years in succession (van Everdingen, 1998). Since permafrost is defined as a material and not a condition, it may include unconsolidated sediments, bedrock, and bodies of ice (excluding glaciers). Permafrost is a sensitive indicator of a changing climate, primarily from its change in distribution, form, and properties (Intergovernmental Panel on Climate Change, 2001; 2007; Arctic Climate Impact Assessment, 2005; Jorgensen et al., 2006).

Permafrost exists throughout ice-free areas of Antarctica, which comprise only 0.35% or 49,500 km^2 (Fig. 3.16). The rest of Antarctica, which features the massive land-based East Antarctic Ice Sheet (12 million km^2) and the marine-based West Antarctic Ice Sheet (1.7 million km^2), is covered by ice, which is not considered to be permafrost. However, sediments and rock beneath the glaciers in Antarctica may be frozen where the glacial ice cover is thin (grey areas on Fig. 3.2.9.1). In other areas, particularly in locations depicted with a cross (Figure 3.2.9.1), sub-glacial lakes have been identified in Antarctica. These lakes exist in areas where the ice is sufficiently thick to induce pressure melting (Herterich, 1988). It is not known whether or not submarine permafrost exists around the Antarctic continent, but results from the Dry Valley Drilling Project from the 1970s (McGinnis, 1981) and the ongoing Andrill project (http://www.andrill.org/) suggest that unlike the arctic, permafrost is not present below McMurdo Sound.

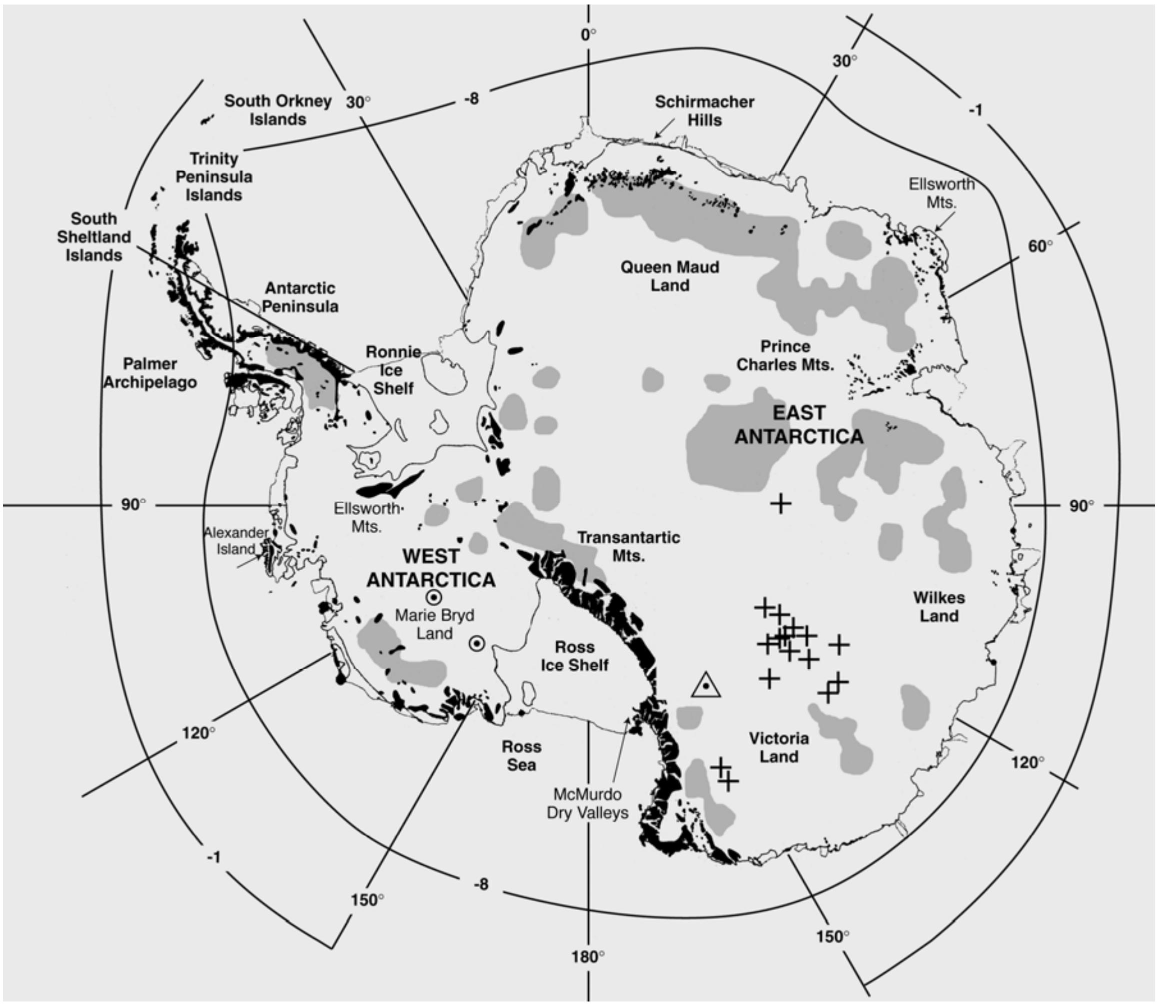


Figure 3.16. Distribution of permafrost in the Antarctic region. The northern limit of permafrost corresponds roughly with the –1ºC isotherm for mean annual air temperature in the sector 45 to 70 W longitude. Continuous permafrost likely exists in

areas with a mean annual air temperature of –8ºC and lower. Permafrost exists throughout the ice-free areas (shown in black). Subglacial permafrost beneath the Antarctic ice sheet may be restricted to the areas shown hatched. Subglacial lakes are depicted with the symbol † (Bockheim, 1995).

Based on limited data, the northern limit of discontinuous permafrost in Antarctica corresponds roughly with the –1º C isotherm for mean annual air temperature (MAAT) for the sector between 45º and 70º W longitude (Fig. 3.16). Continuous permafrost likely exists in areas with a MAAT of -8ºC or colder, corresponding to trends reported in the Northern Hemisphere (Brown and Péwé, 1973).

Figure 3.17 shows permafrost distribution in the McMurdo Dry Valleys (MDV), which represent the largest ice-free region (6,700 km^2) in Antarctica. In the MDV the MAAT ranges from –18 to –30ºC. The figure shows three forms of permafrost: buried ice or ground ice is shown in blue; ice-cemented permafrost is depicted in green, and "dry-frozen" permafrost is shown in red. The latter form of permafrost may be unique to Antarctica and represents material that is well below 0ºC but has insufficient interstitial moisture to be cemented. In the MDV about 55% of the permafrost is ice-cemented, 43% is dry-frozen, and ground/buried ice comprises at least 2% of the area (Bockheim et al., 2007). The distribution of ground ice in Antarctica is of particular interest. Although small bodies of ground ice exist in the MDV, widespread ground ice in the form of buried glacier ice and ice wedges occurs in the Northern Foothills of North Victoria Land (Guglielmin and French, 2004) and probably elsewhere in coastal areas of Antarctica.

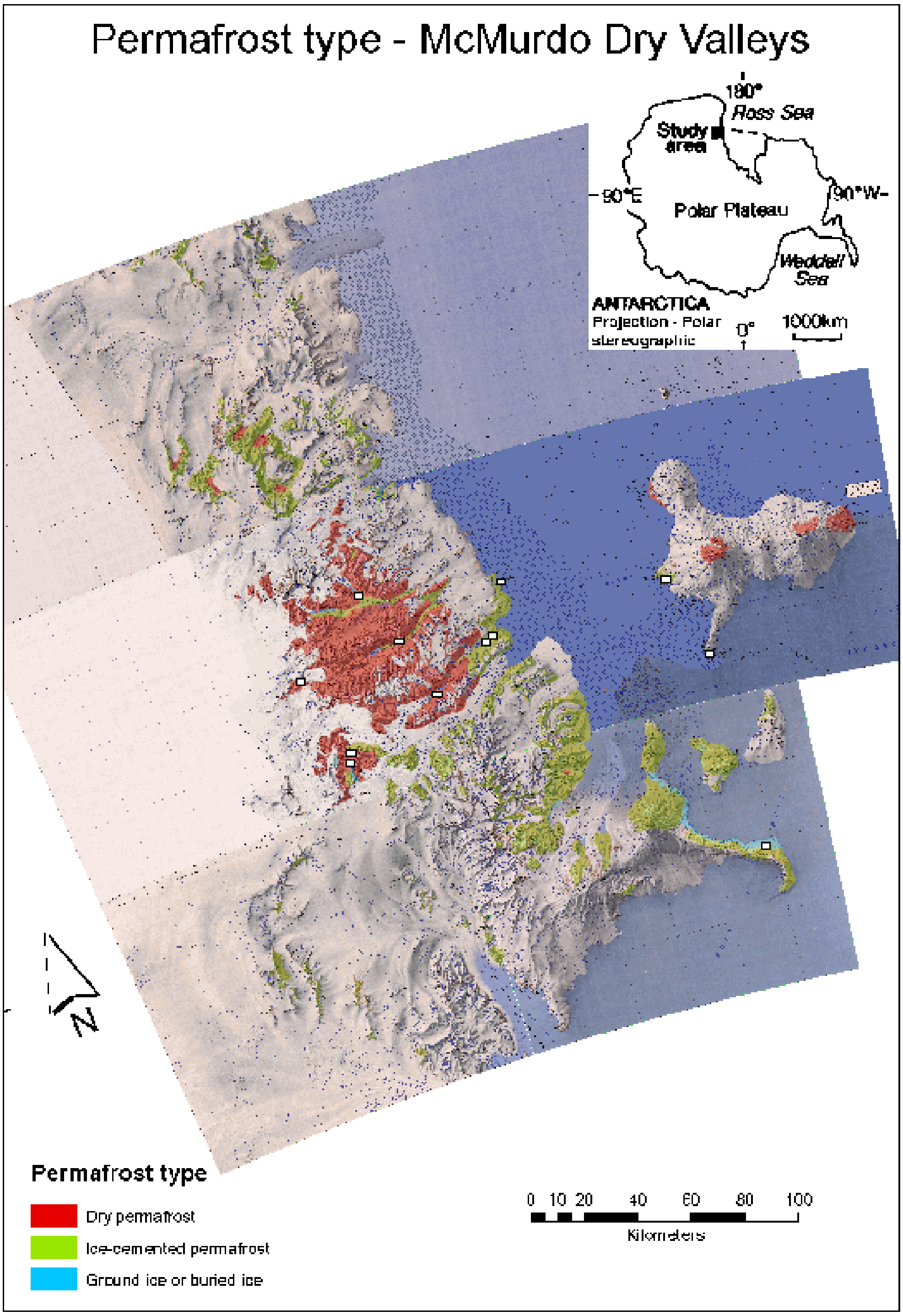


Fig. 3.17. Reconnaissance map (1:2 million scale) showing permafrost distribution by form in the McMurdo Dry Valleys (Bockheim et al., 2007). Squares show Circumpolar Active-Layer Monitoring – South (CALM-S) sites.

Permafrost properties

The thickness of permafrost in Antarctica varies with region (Bockheim and Hall, 2002). In the MDV of interior Antarctica (77ºS, 161-166ºE), it ranges from 240 to 970 m (Decker and Bucher, 1977). In North Victoria Land (74ºS, 164ºE), 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 and South Orkney Islands and occurs only at the higher elevations (Ramos and Vieira, 2003). Permafrost temperatures, normally measured at a depth of 50 m, range from -14 to -24ºC in continental Antarctica (Decker and Bucher, 1977). The temperature of permafrost at a depth of 10 m in NVL ranges between -12 to -17ºC (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 (Fig. 3.17), 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 (77.4ºS, 163.8ºE), the seasonal thaw or active layer thickness varied from 30 cm in 2001 to 60 cm in 2002 (Fig. 3.18); 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).

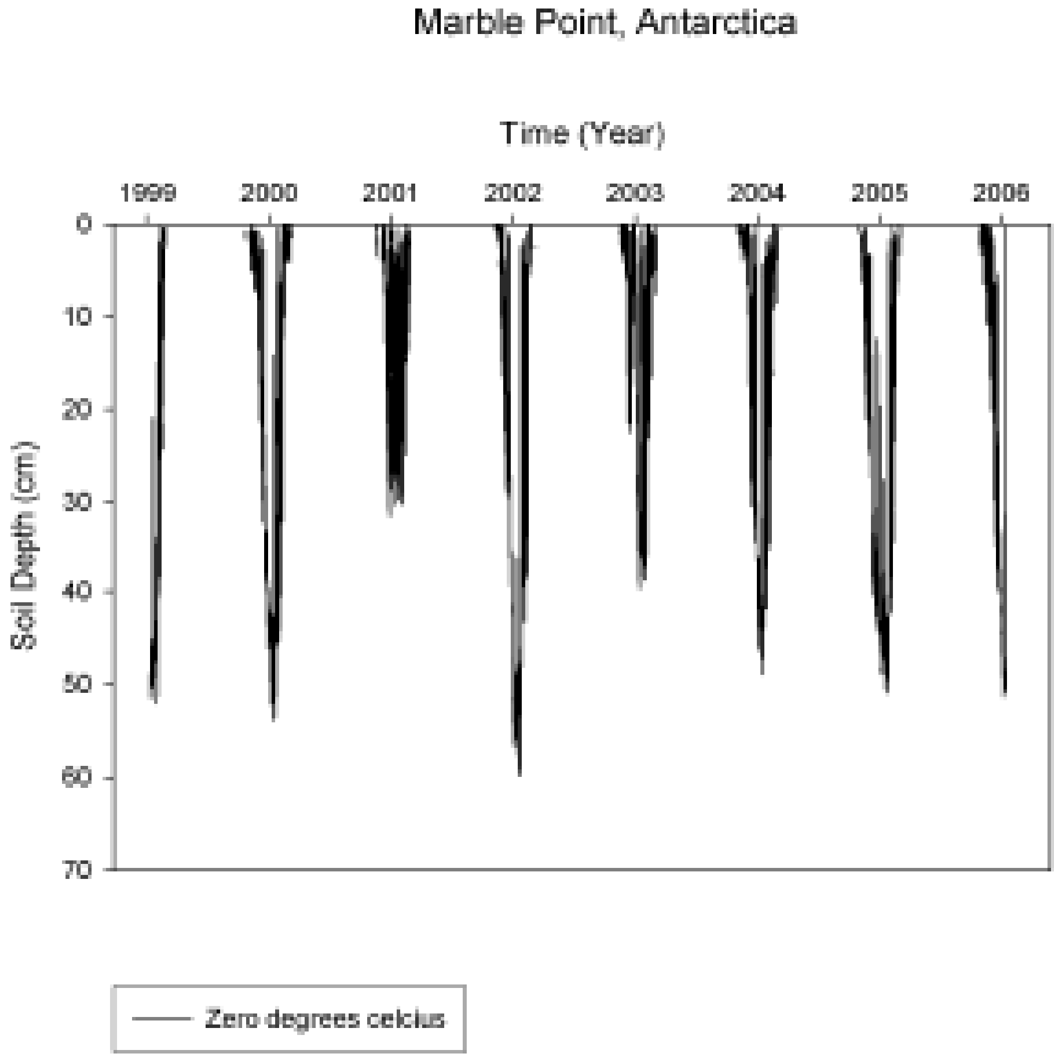


Fig. 3.18. Variation in active layer thickness at Marble Point (77.4ºS, 163.8ºE), McMurdo Dry Valleys, Antarctica during the period 1999 to 2006 (Seybold, unpublished).

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. Dataloggers 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.

3.2.10 **Marine Biology**

Specimens of vertebrates and invertebrates are collected both in the proximity of coastal stations and during wide-range oceanographic campaigns, such as ICEFISH (**I**nternational **C**ollaborative **E**xpedition to collect and study **F**ish **I**ndigenous to **S**ub-Antarctic **H**abitats) and CAML (**C**ensus of **A**ntarctic **M**arine **L**ife), both part of the SCAR/IPY programme Evolution and Biodiversity in the Antarctic. The Response of Life to Change (EBA). Otter, mid-water, Blake and MOCNESS trawls, plankton nets, beach seining, tide pooling, and traps are routinely used. A wide variety of nets are used, for bottom, pelagic and bentho-pelagic trawling; gill nets and traps are set at different depths near shore; in many instances, hook and line is a very useful strategy, e.g. from holes and cracks in the sea ice, often in fish huts, where lines with many hooked arms reach the bottom at many hundred meters and are very effective in capturing big pelagic fish. Many Antarctic stations (e.g. McMurdo, Mario Zucchelli, Palmer) are equipped with aquaria and running seawater, where it is possible to study live marine organisms for long periods of time. Many ships have tanks with running seawater and carry life specimens to stations, and also to non-polar institutions.

Much of our knowledge of the effect of the environment on vertebrate physiology and evolution has come from fishes, 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.

A new powerful aid to evolutionary biology and ecology has been recently provided by *molecular biology*, which allows us to explore the function of individual genes. Nowadays molecular phylogeny is essential for studying evolution, and important insights come from protein (e.g. hemoglobin) and nucleic acid sequences (mitochondrial genes: e.g. 12S and 16S rRNAs; nuclear genes: e.g. 28S rRNA, genes encoding globins, antifreeze glycoproteins, tubulins, myoglobin, etc), as well as from large-scale chromosome change (Stam et al. 1997, 1998; di Prisco 1998, 2000; Bargelloni et al. 2000; Pisano et al. 1998, 2003; Papetti et al. 2007).

Over the last twenty years, important advances have been achieved in understanding the molecular mechanisms involved in evolutionary adaptation to temperature, and important progress is likely to occur, as scientists exploit new molecular approaches and become concerned about environmental changes in polar regions. Increasing knowledge of protein structure and function is progressively assembling the puzzle of cold adaptation. However, 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.

Most of the studies of protein adaptation to temperature take advantage of one of two approaches. The structural/mutational approach produces a detailed portrait (although often controversial) of the interactions stabilising the high- and low-temperature-adapted proteins. However this approach, due to much labour and costs, restricts analyses to a limited number of proteins, leading to a potentially biased view of thermal stabilising mechanisms. The second approach, less expensive yet more comprehensive, uses sequence comparisons of families of homologous high- and low-temperature-adapted proteins. In the past, the latter approach has been hampered by limitations, due for instance to paucity of genomic sequences from extremophile organisms. Nowadays, the possibility to sequence whole genomes may provide the necessary amount of data, allowing to reject or accept some of the classical hypotheses currently invoked to explain protein thermal adaptation.

Sequencing genomes of model organisms is a great challenge for biological sciences. In the last decade, scientists have developed a large number of methods to align and compare sequenced genomes. The analysis of a given sequence provides much information on the genome structure, but only to a lesser extent on the function. Comparative genomics are a useful tool for functional and evolutionary annotation of genomes. In principle, comparison of genomic sequences may allow to identify the 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.

The fast development of genomic technologies has brought biological sciences to the threshold of a revolution. The availability of sequences greatly enhances our understanding of the structure, content, and evolution of genomes. Future research in genomics will not only depend on the new technologies, but also on the integration with physiology and biochemistry.

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

To date, there is not much knowledge on the capacity of the polar marine fauna to respond to on-going climate changes. Given the amount of information available on cold adaptation, the study of fish of the polar seas will provide invaluable clues to the development, impact and consequences of climate change, and hence on the future of the Earth.

3.2.11 **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). These geographic regions are also meaningful biogeographical regions (see Pickard and

Seppelt 1984, Smith 1984, Longton 1988, Chown and Convey 2006, 2007, Huiskes et al. 2006; Convey 2007).

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 (Hughes et al. 2006). 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 & 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 1996; 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 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 (ref from Sands et al), 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, in press). 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 (ref from Sands et al), 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 (check), 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 (list refs from Pugh & Convey in press, Convey et al in press), starting to generate a baseline against which future changes can be compared.

3.2.12 **Models**

Coupled Atmosphere-Ocean Models

Predictions of future climate changes can only be made by using coupled global climate models. These are also primary tools with which to simulate the past and present-day climates. 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, coupled atmosphere-ocean models mean three-dimensional coupled atmospheric and oceanic general circulation models.

In the early 1990s, there were only a few coupled global climate models; in the fourth assessment report of the Intergovernmental Panel on Climate Change (IPCC AR4), around 24 such models joined the model inter-comparison project. Here we give a brief description of a coupled atmosphere-ocean model. The strengths and weaknesses of the coupled models are also discussed. We then present some results that are relevant to the Antarctic from IPCC AR4 models.

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

The atmospheric component consists of a dynamical core that time-steps the equations of momentum and thermodynamics that are a set of partial differential equations, along with many physical processes. These partial differential equations are usually 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 pressure, height or a terrain following coordinate. Typically, there are 20-30 levels in the vertical direction. Sub-grid processes, such as cumulus clouds, have to be parameterized using the quantities that the model can resolve, and it is well known that there are many uncertainties in these parameterizations. The water cycle and cloud-radiation interactions play very important roles in the climate system, but they are two of the largest areas of uncertainty in the current generation of climate models.

The oceanic component has many similarities to the atmospheric element. However, due to the existence of side boundaries, 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, the Miami Isopycnal Coordinate Ocean Model). Usually 20-40 levels are used in the current generation of models. Due to the much larger density of sea water than the air, the ocean current is much slower than the wind and generally much higher spatial resolution is needed to resolve the processes with the same time scale as those processes in the atmosphere. But due to the much more complicated physical processes in the atmospheric component, an atmospheric model requires much more computing resources per grid than an ocean model. The typical spatial scale of mesoscale eddies in the ocean is around 10 km, which is too fine to be resolved by current ocean models. However, eddy processes are actively involved in the buoyancy and momentum transport across the Antarctic Circumpolar Current and in the boundary layers.

The land-surface component 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 surface roughness and albedo. Land surface processes can also strongly affect the oceanic heat and freshwater fluxes.

Snow over land and sea ice are two important elements that are explicitly resolved in coupled models. Glaciers and ice shelves are usually not resolved; large scale ice sheets 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 and more coupled models employ sea ice models with more complicated thermodynamics and with sea ice dynamics of various complexity.

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 sheets, are usually prescribed as 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 from the present-day climate. The use of flux adjustments was due to the 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, it is not necessary for many coupled climate models to use flux adjustments, because of model improvements.

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 drift of atmospheric and oceanic heat transport are the largest in the southern high latitudes (see Fig 16 of Gordon et al. 2000). It is difficult to represent small-scale topography feature over the Antarctic and in the Southern Ocean. Shelf processes are believed to be very important in the Antarctic deep/bottom water formation. Melting and freezing at the base of ice shelves are completely unknown in the current generation of coupled climate models; the role played by sea ice in the southern high latitude water cycle is unknown because of the lack of observed sea ice volume and sea ice drift velocity. Models have large uncertainties and the simulated water fluxes into the ocean are highly varied. All models cannot resolve the Antarctic ice sheet and ice shelves and their effects on the ocean circulation.

Ocean circulation simulation is still very variable across the range of models. ACC transport is significantly biased in most IPCC AR4 models (Russell et al. 2006). The simulated strength of the largest sub-polar gyre - the Weddell Gyre, has a very different value in different models (Fig. 3.19). 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 Fig. 3.20. Sea ice extents simulated by AR4 models are also very different (see Parkinson et al. 2006).

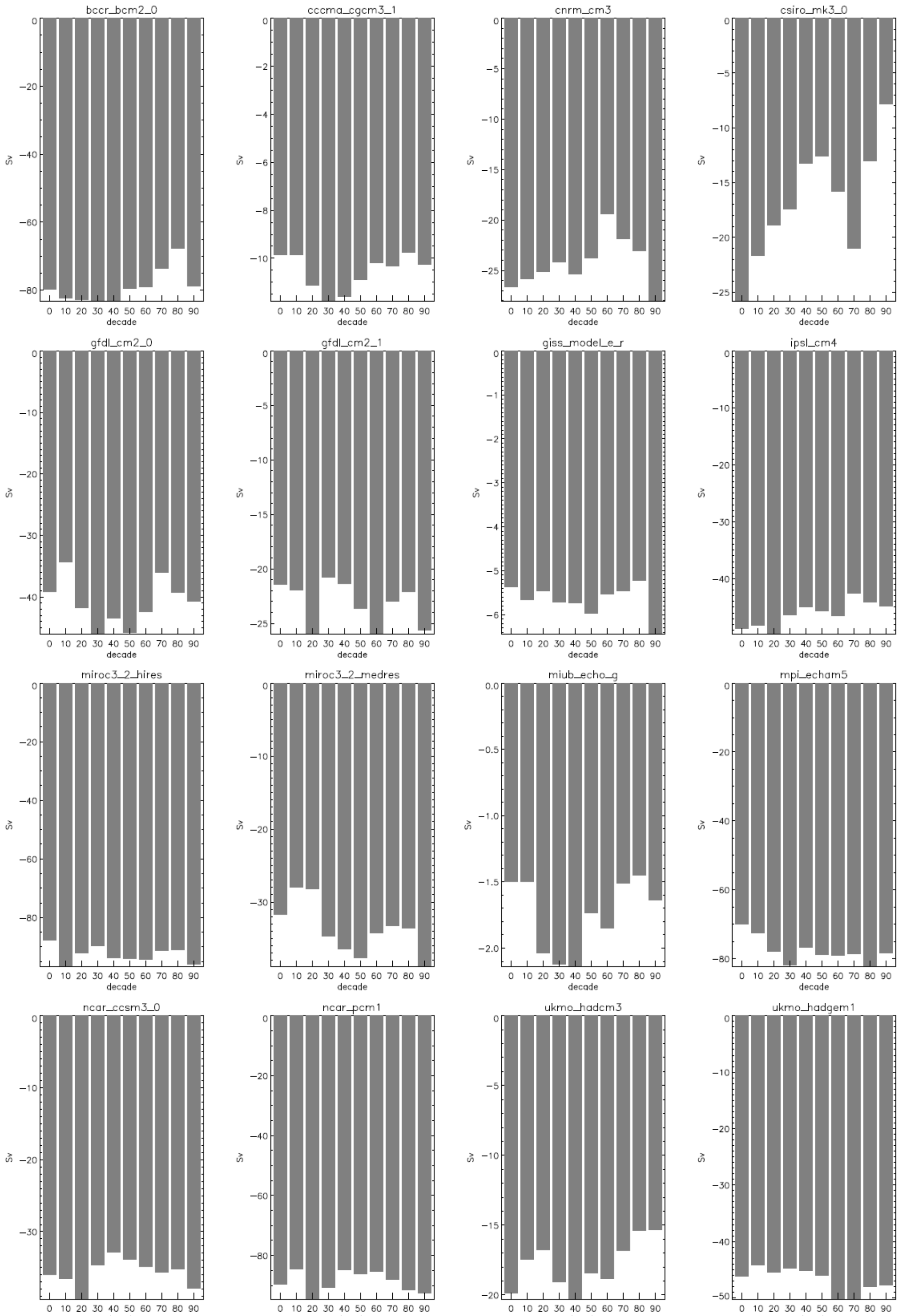


Fig. 3.19**.** Simulated decadal means of Weddell Gyre strengths by 16 IPCC AR4 coupled climate models during the 20th century. The Weddell Gyre strength is defined as the absolute transport of the southern limb, whose observed value is 56 ± 8 Sv (Klatt et al. 2005).

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.

More observational work needs to be done. The IPY 2007-2008 will provide an opportunity for climate modellers to further validate and improve their models and finally provide important insights into the 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 date to assess the model performance. In the mean time, more model intercomparison work needs to be carried out to find out the model ensemble means and the model differences.

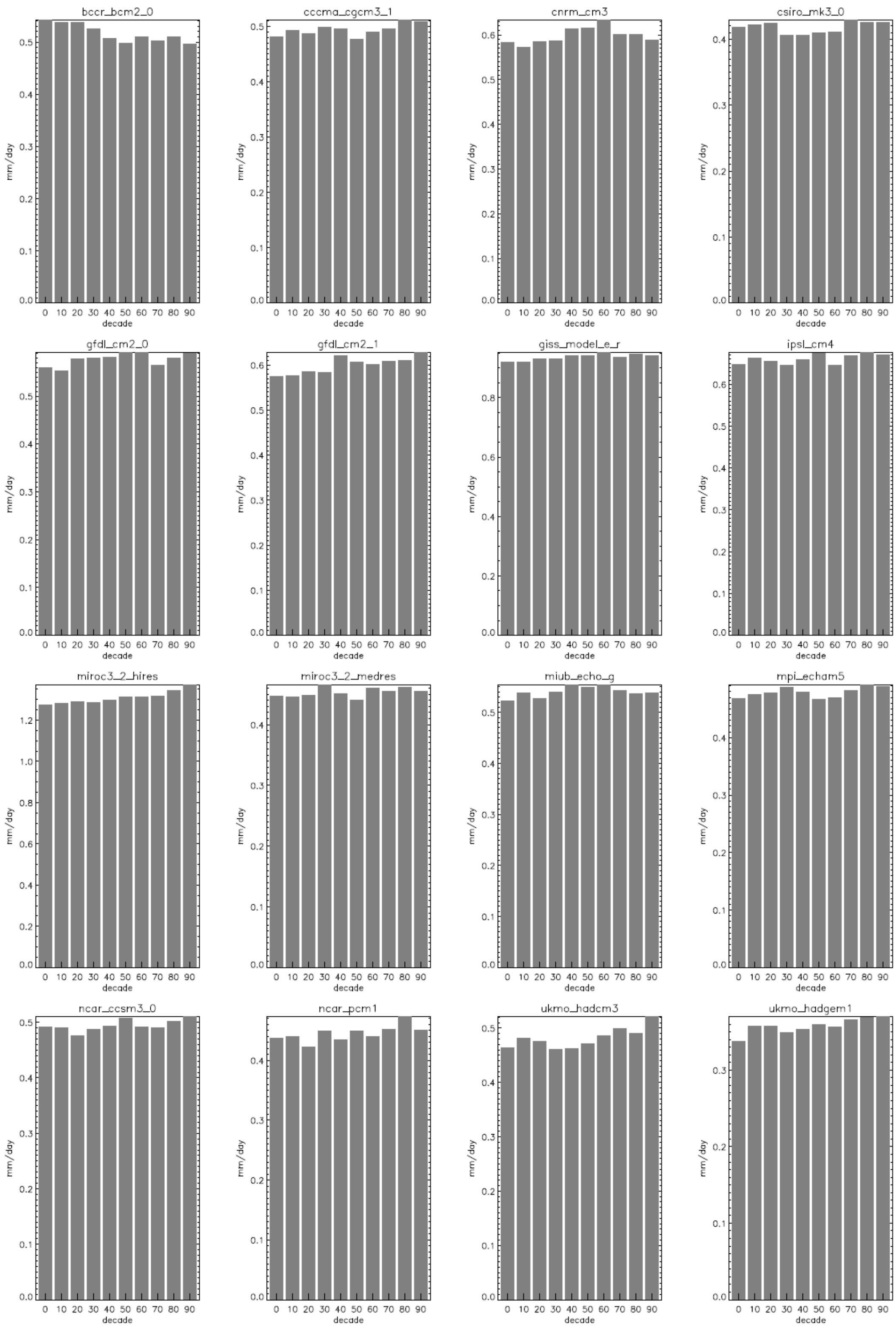


Fig. 3.20. Simulated decadal means of precipitation averaged over the Antarctic continent by 16 IPCC AR4 coupled climate models during the 20th century.

Regional Climate models

Global Circulation Models (GCMs) have increased in resolution and complexity since the first models were developed in the middle of the 20th Century. However, until more 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. However, it remains that GCMs, due to their need by definition to be Global, lack the grid resolution to sufficiently or parameterize process that occur on a subgrid scale. While the spatial resolution of GCMs will increase, it 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.

Atmosphere-only regional climate models

These 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.

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. However, they will become of increasing importance in the future.

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, typically a thousand 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. Horizontal scales in the ocean are usually determined by the Rossby radius - a measure of how far a 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 quantitative judgements. However we have a good theoretical understanding of the problem involved and long model runs usually show up gross errors. We also have satellites which provide good data on the surface fields and the ARGO program is starting to provide a good data set for the near surface layers. Detailed studies of density currents, the sea-ice field and other key regions provide further local checks on model performance.

The major model choices

Most ocean models represent the spacial 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 spacial 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 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 sale ocean models such as MOM (Griffies at al. 2004) and OCCAM (Coward & de Cuevas, 2005; Webb & 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 the flows are not geostrophic. It is also used for some large scale models (Penduff, 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, three standard schemes are used (Grifies et al. 2000). Many, following the original Bryan and Cox model use fixed horizontal layers (Bryan & 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

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

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 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 model 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, 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 & Haidvogel, 1994, Blumberg & 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 introduces small errors in calculating the horizontal pressure gradient (Shchepetkin & McWilliams, 2003). This does not matter in turbulent shallow water but in the deep ocean this 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 (Griffies, 2005) 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 develop, 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 for the coastline, the ice usually lasts for only a single year so complex multi-year ice models are not essential. Secondly the primary westerly winds means that the flow is generally divergent, so modelling of ice rheology is not so essential. As a result Southern Ocean models which 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 & Dukowicz (1997) elastic–viscous–plastic scheme is often chosen. For further details of this and more detailed schemes see section 1.8.

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 et al. 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 (Doscher & Beckman 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). However 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 which 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 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. However the velocity of the layer, which affects sea ice, does depend on near surface mixed layer processes and these do vary from model to model. Thus for any research involving sea-ice a good mixed-layer model is essential.

The mixed layer models that are available include those of Pacanowski and Philander (1981), Mellor and Yamanda (1982), Price et al. (1986), Large et al. (1994) and Gaspar et al. (1990) extend from simple bulk models to detailed turbulent closure models. The effectiveness of the different models 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 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 & 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.

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 & 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).

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 & 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 solutions is to specify the boundary conditions using climatology.

At the largest scale there have been three major models which 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, 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 upped 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.

Dinniman et al.

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 Williams, 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 strength 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 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 which 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 is becoming available from bore holes and 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.

Under Ice Shelf Models

It is important to have realistic models of the ocean flow under the 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, however, this will be necessary in the future and the current generation of ocean/shelf models provide a step in this direction.

Isolation from direct wind forcing means that the main drivers of the ocean circulation beneath ice shelves are tidal and thermohaline forces. Tidal modelling indicates that residual mean flows are small, so the primary role of the tides is as a source of energy for mixing (MacAyeal, 1984; Makinson and Nicholls, 1999; Makinson, 2002). As a consequence, most modelling studies have focussed on the thermohaline circulation beneath ice shelves (Williams et al, 1998), at the expense of neglecting both the possibly unimportant residual tidal flows and the potentially very important tidal mixing.

Melting at the ice shelf base cools and freshens the top of the water column, producing a stable stratification. The cold water must be removed and replaced by the upwelling of warmer, saltier water if melting is to be sustained. Because the ice shelf base slopes upwards from grounding line to ice front, the vertical stratification induced by melting gives rise to horizontal pressure gradients that drive meltwater towards the ice front and draw warm, salty water towards the grounding line. This motion causes the mixing and upwelling that sustain melting and maintain the forcing on the overturning circulation. A number of one- and two-dimensional models (MacAyeal, 1984; Hellmer and Olbers, 1989; Jenkins, 1991; Holland and Feltham, 2006) have focussed on this process, which has been found to respond sensitively to changes in the temperature of the inflowing water. Higher temperatures give rise to more melting, increasing the thermohaline forcing, leading to a stronger circulation and a further increase in the heat delivered to the ice shelf base. However, the models assume an essentially infinite supply of warm inflowing water having defined properties, and while they have proved valuable in advancing our understanding of fundamental processes, they have little predictive skill.

In Antarctica the warm inflows are derived from two sources. Around much of the coastline the continental shelf seas are dominated by High Salinity Shelf Water (HSSW), formed by haline convection beneath growing sea ice. This surface-freezing-point water is the warmest water on the shelf. It can melt ice at depth because the freezing point is depressed by the elevated pressure. As the product water (Ice Shelf Water - ISW) rises on its way out of the sub-ice cavity some refreezing occurs at the ice shelf base. The continental shelves of the Amundsen and Bellingshausen seas are different in that HSSW is absent and Circumpolar Deep Water (CDW) replaces it as the densest water on the shelf. With a temperature about 3°C above the surface freezing point, CDW drives much higher melt rates and the temperatures in the outflow remain high enough to prevent refreezing.

The application of three-dimensional ocean general circulation models, based on either terrain-following or isopycnic vertical coordinates, to the study of the waters beneath and in front of ice shelves (Grosfeld et al., 1997; Beckmann et al., 1999; Holland and Jenkins, 2001; Williams et al., 2002) has enabled the inflows to be modelled. Both approaches to the problem of vertical discretisation have advantages and disadvantages. Terrain-following coordinates allow a natural incorporation of the physical processes at the gently-sloping ice shelf base, but struggle with the vertical wall of the ice front. The physical topography is essentially invisible in an isopycnic coordinate system, but a non-isopycnic "mixed" layer is required at the base of the ice shelf to accept the buoyancy forcing associated with melting and freezing. Most studies have focussed on the large ice shelves of the Ross and Weddell seas, where HSSW provides the inflow. Forcing on the open ocean has been derived from restoring the surface to prescribed values of temperature and salinity (Jenkins et al., 2004), or using a sea ice model forced by idealised (Grosfeld and Gerdes, 1998) or realistic winds (Timmermann et al., 2002). The result has been a seasonal supply of inflowing HSSW. The inflows are generated in winter, and the cavity cools as the supply of HSSW dries up in the summer (Nicholls, 1997; Jenkins et al., 2004). Inter-annual changes in the sea ice distribution, resulting from calving of giant bergs from the ice front or variability in the wind fields, were found to alter the circulation pattern and the resulting net melt rate (Nøst and Østerhus, 1998; Grosfeld et al., 2001; Timmermann et al., 2002). Nicholls (1997) suggested that moderate climatic warming, leading to decreased sea ice production could reduce the supply of HSSW and hence reduce the net melting at the base of these large ice shelves.

Lack of knowledge of the sub-ice bathymetry has hampered the use of these three-dimensional models to simulate the circulation beneath the thinning ice shelves of the Amundsen, Bellingshausen and north-western Weddell seas. Thus the role of basal melting in driving the changes observed in these regions remains speculative. The first two regions are dominated by CDW (Jacobs et al., 1996), so changes in melting would be driven by changes in its temperature or flow. The earlier model of Hellmer and Olbers (1989) has been applied to Pine Island Glacier (Hellmer et al., 1998) and used to explore the sensitivity of melting to changes in water temperature. A warming of 1°C was found to increase the net melt rate by about 10 m yr^{-1}. Payne et al. (2007) applied the two-dimensional model of Holland and Feltham (2006) to the same problem and found a similar sensitivity (16 m yr^{-1} $^{o}C^{-1}$) for moderate increases in temperature. The north-western Weddell Sea represents something of a transition between the two regimes discussed above. True HSSW is absent but the on-shelf version of CDW is so modified by wintertime convection that its temperature is very close to the freezing point (Nicholls et al., 2004). It is possible that reduced sea ice formation is this area could lead to an increase in the heat carried beneath the ice shelf, in contrast to the reduction postulated for the southern Weddell Sea, although to date no model has focussed on this region.

Holland et al. (2007) used the model of Holland and Jenkins (2001) to study the sensitivity of idealised ice shelves to a series of ocean warming scenarios. They found that the net melt rate increased as a quadratic function of the ocean temperature offshore of the ice front, and demonstrated that this was consistent with most earlier findings. However, the response was also found to be dependent on the geometry of the ice shelf, making it impossible to formulate a universal relationship between ocean temperature and net melt rate applicable to all ice shelves. Even if the effect of geometry could be parameterised, trends in inflow temperatures are likely to be more influenced by changes in supply of HSSW and CDW than any warming of the surface ocean that might occur as a direct response to atmospheric warming. Overall it is difficult to assess the skill of the above models. Oceanographic observations under ice shelves are sparse and those offshore of the ice fronts are mostly restricted to the summer months. Most glaciological estimates of ice shelf mass balance yield only the basal melt rate required for equilibrium and are subject to spatial and temporal averaging that complicates direct comparison with model results.

Ice sheet models

The first numerical modeling involving the entire Antarctic Ice Sheet is the diagnostic study of Budd and others (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, sect. 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.

Realistic model simulations of the behavior of the Antarctica Ice Sheet over time require adequate surface and lower boundary conditions. Neither of these are well constrained, due to paucity of data especially in much of East Antarctica. The surface mass balance has been measured directly at a limited number of stations, most of which are located in the coastal regions so that the mass input through snowfall is poorly constrained (c.f. Bentley, 2004). While some airborne radar sounding has been conducted, resulting maps of basal topography (Drewry, 19xx; Lythe and others, 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, however, 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 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 (1998) 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 and others, 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 and others, 2004). 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 and others (2003) linked ice-shelf collapse to southward migration of the -9 ^{o}C 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 and others, 2002; De Angelis and Skvarca, 2003; Rignot and others, 2004). These observations are suggestive 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.

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 and others, 2007). The exact nature of this transfer of stress is not well known but may ultimately determine inward or outward migration (Raymond, 1996; Raymond and others, 2001; Jacobson and Raymond, 1998). Moreover, Van der Veen and others (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 and others, 1999) and over sedimentary basins (Anandakrishnan and others, 1998; Bell and others, 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, believed to be 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 and others, 2003). Velocity measurements on grounded glaciers formerly draining into these ice shelves indicate 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 towards 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 and others, 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 process 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 behaviors 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 assumed depth-independent. It may be expected 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 meso-scale 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.

3.3 Atmospheric Circulation

3.3.1 Modes of variability/SAM

Variability

Within the atmospheric circulation of the high southern latitudes there are several so-called modes of variability that can be discriminated. These are circulation patterns that appear more frequently then would be expected in a random sample and can ‘describe’ a significant proportion of the total circulation variability.

Modes of variability – The Southern Hemisphere Annular Mode

The Southern Hemisphere Annular Mode (SAM) is the principal mode of the SH 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 3.21) and *vice versa*.

The SAM is equivalent barotropic (the spatial pattern varies negligibly with height in the atmosphere) and is revealed as the leading EOF in many atmospheric fields (see Thompson and Wallace (2000) and references therein). Model experiments demonstrate that the structure and variability of the SAM result from the internal dynamics of the atmosphere (e.g. Hartmann and Lo, 1998; Limpasuvan and Hartmann, 2000). Poleward eddy momentum fluxes — synoptic-scale weather systems — interact with the zonal mean flow to sustain latitudinal displacements of the mid-latitude westerlies. The SAM contributes a significant proportion of SH climate variability (typically ~35%) from high-frequency (Baldwin, 2001) to very low-frequency timescales (Kidson, 1999), with this variability displaying a red noise distribution.

Gridded reanalysis datasets have been utilised to derive time series of the SAM (e.g. Thompson et al., 2000; Renwick, 2004). However, 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) means that long-term SAM time-series cannot be derived this way. 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. Unfortunately, this index itself is limited to the period following the International Geophysical Year (IGY) of 1957/58. However, attempts have been made to reconstruct century-scale records based on proxies of the SAM. Goodwin *et al*. (2004) use Na concentrations from Law Dome while Jones and Widmann (2003) employ tree-ring width chronologies: both these studies stress the decadal variability in their derived SAM time series.

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 trend was especially pronounced from the mid-1960s until the end of the 20th Century, since when there have been a similar frequency of seasons with positive and negative SAM values (Figure 3.22). 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 precipitation variability in 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 flow through the Drake Passage (Meredith *et al.*, in press). Modelling studies indicate 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 (Oke and England, 2004) and may explain recent patterns of observed temperature change at Southern Hemisphere high latitudes described by Gille (2002). Although the SAM is essentially zonal, a wave-number 3 pattern is superimposed (cf. Figure 3.21), 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 feedback into a more positive SAM. However, modelling work by Marshall and Connolley (2006) indicated that increased SSTs at high southern latitudes will warm the atmosphere and, through thermodynamics, cause the atmospheric centre-of-mass to rises and geopotential height increases thus producing a more negative SAM.

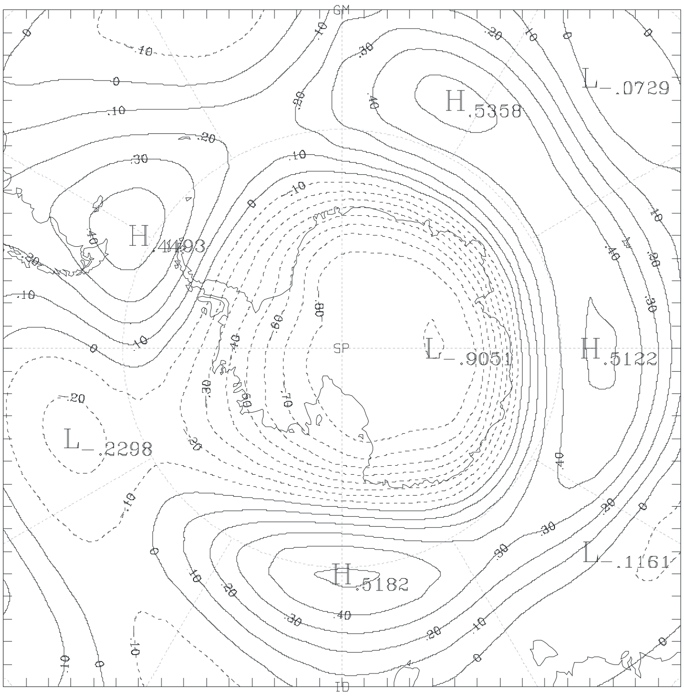


Figure 3.21. The SAM, defined using 500 hPa geopotential height monthly anomaly data for 1979-2001. Here the SAM is in its positive phase.

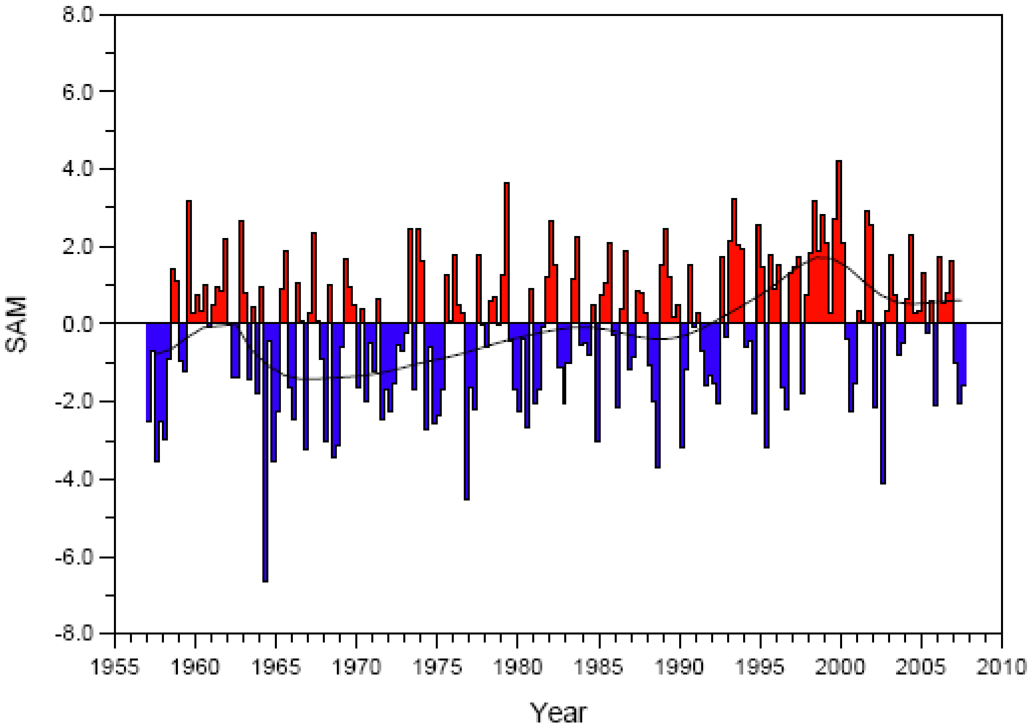


Figure 3.22. Seasonal values of the SAM index calculated from station data (Marshall, 2003). The smooth balck curve shows decadal variations. This is an updated version of Figure 3.32 from IPCC (2007).

The Pacific-South American (PSA) pattern

The PSA pattern (Mo and Ghil, 1987) is considered 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. It is generally described as a wave-like anomaly patterns emanating from the subtropical western Pacific (Figure 3.23). The PSA pattern comprises low pressure east of New Zealand, high pressure in the SE Pacific and low pressure over South America/South Atlantic and its evolution shows a coherent eastward propagation. The latter two ‘centres of action’ influence Antarctic climate and are discussed in more detail later in this section. 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) and previously seen in the work of Chen et al. (1996). 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. In the extra-tropics, the atmospheric circulation is influenced directly by storm track changes that alter the meridional eddy heat flux divergence and convergence and a shift in the latent heat release zone. 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, teleconnection correlations tend to be small. A modelling study by Lachlan-Cope and Connolley (2006) indicates 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 SH 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.

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 (a La Niña) event is associated with significantly more (less) blocking events in the SE Pacific (Renwick and Revell, 1999). This pressure anomaly in the Amundsen-Bellingshausen Sea (ABS) is primarily responsible for the Antarctic Dipole (ADP) in the interannual variance structure in the sea ice edge and SST fields of the SO, which is characterised by an out-of-phase relationship between anomalies in the central/eastern Pacific (Amundsen-Bellingshausen Sea) and the Atlantic (Weddell Sea) sectors (Yuan and Martinson, 2001: see Figure 3.24). In addition the ADP sea ice anomalies are reinforced by ENSO-related storm track variability, which influences the Ferrel cell by changing 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. However, the strongest sea ice correlations associated with ENSO occur at 120-132W (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.

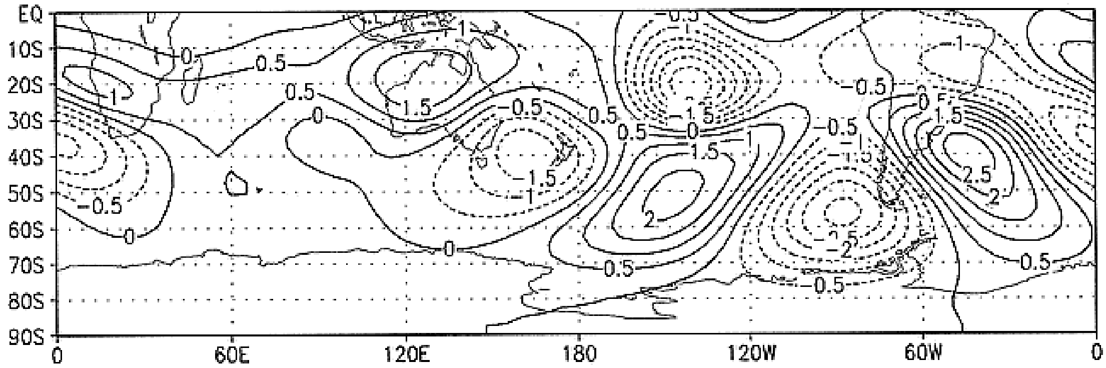


Figure 3.23. The PSA teleconnection pattern of positive and negative pressure anomalies (after Mo and Higgins, 1998). Notice the ‘centre of action’ west of the Antarctic Peninsula.

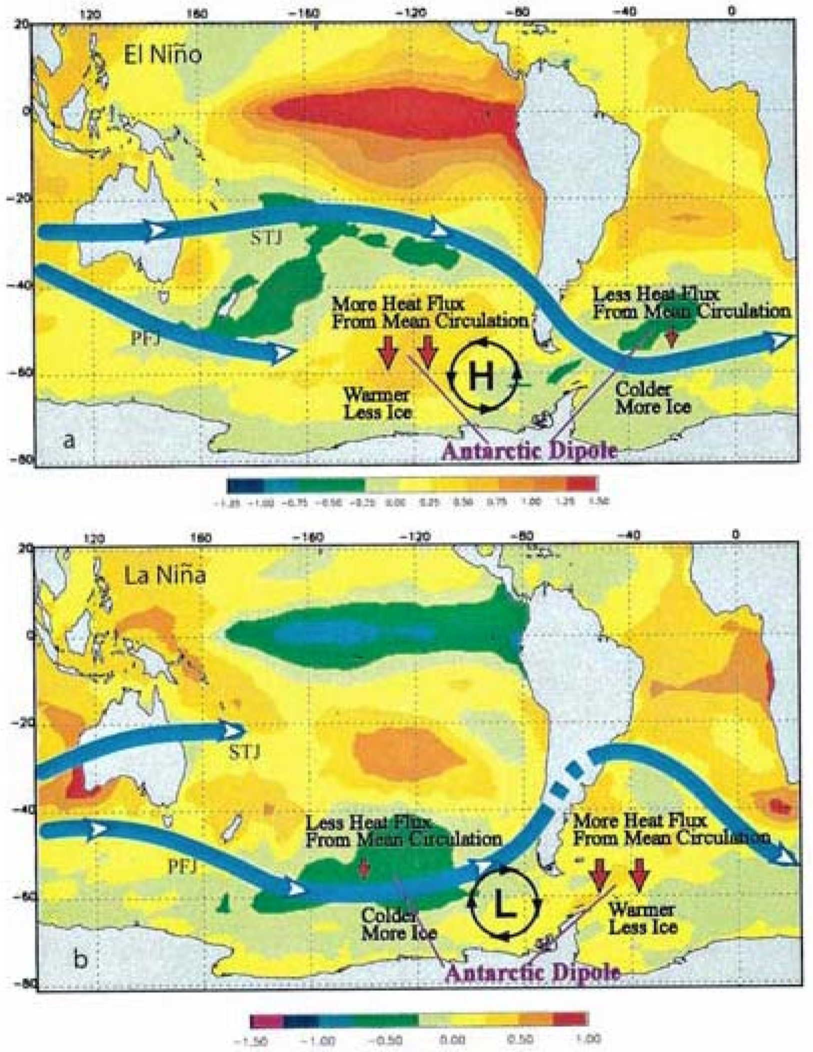


Figure 3.24. SST anomaly composites (°C) for a) El Niño conditions and b) La Niña conditions.Schematic jet streams (STJ is the subtropical jet and PFJ is the polar front jet), persistent anomalous high and low pressure centres, and heat fluxes are also marked. (Yuan, 2004, Antarctic Science).

4.4.3 The Antarctic Circumpolar Wave (ACW)

The Antarctic Circumpolar Wave (ACW), as first defined by White and Peterson (1996), is an apparent easterly progression of phase-locked anomalies in Southern Ocean (SO) surface pressure, winds, SSTs and sea-ice extent (Figure 3.25). As such, it thus represents a coupled mode of the ocean-atmosphere system. The ACW has a zonal wavenumber of 2 (a wavelength of 180°) and the anomalies propagate at a speed (6-8 cm s^{-1}) 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 & 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.

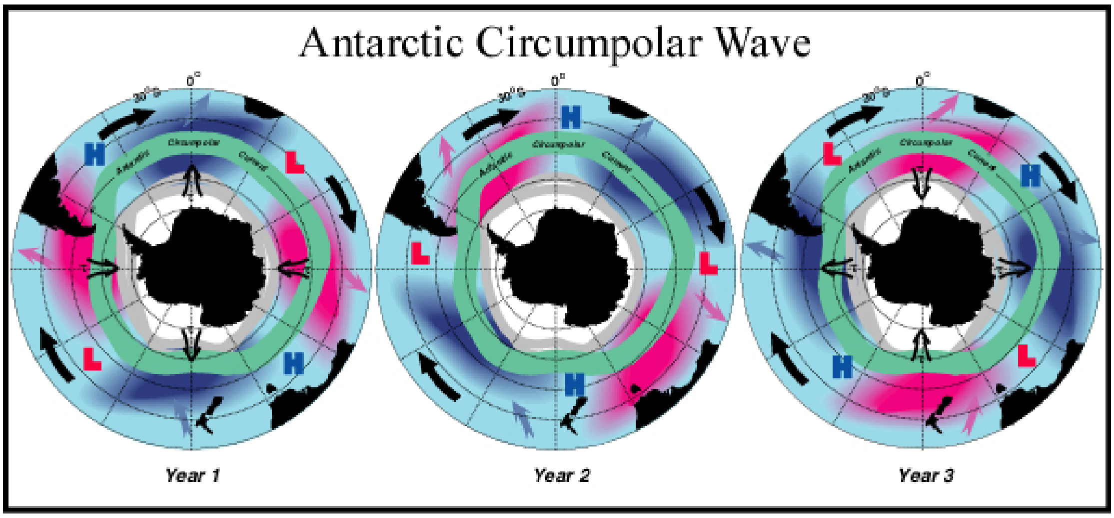


Figure 3.25. Simplified schematic summary of interannual variations in sea surface temperature (red, warm; blue, cold), atmospheric sea level pressure (bold H (high) and L (low)), and sea ice extent (grey line), together with the mean course of the Antarctic Circumpolar Current (green). Heavy black arrows depict the general eastward motion of the anomalies, and other arrows indicate communications between the circumpolar current and more northerly tropical gyres. (White & Peterson, 1996).

While 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), since the initial discovery of the ACW others have questioned its persistence: a number of observational and modelling studies have indicated that the ACW is not apparent in recent data before 1985 and after 1994 (e.g. Connolley, 2003) 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. Many studies (e.g. Christoph et al., 1998; Cai et al. 1999; Weisse 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 (Christoph et al., 1998). 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 or not. 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 advection by the Antarctic Circumpolar Current (ACC): both these points have been refuted in the literature. The Antarctic dipole (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 SO interannual variability, centred in the South Pacific and linked to the PSA pattern (Christoph et al., 1998; Cai et al., 1999; Venegas, 2003; Park 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) demonstrated that the ocean acts as an integrator of short-term atmospheric fluctuations (white noise) and turns them into a red response signal (lower frequency signal), and 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). However, the results of the study of Park et al. (2004) indicated that any such propagating anomalies comprised only ~25% of interannual SO SST variability and are often rapidly dissipated in the Indian Ocean, intermittent in phase and frequently do not complete a circumpolar journey (Figure 3.26). Hence, they questioned the very existence of the ACW, as originally described by White and Peterson (1996).

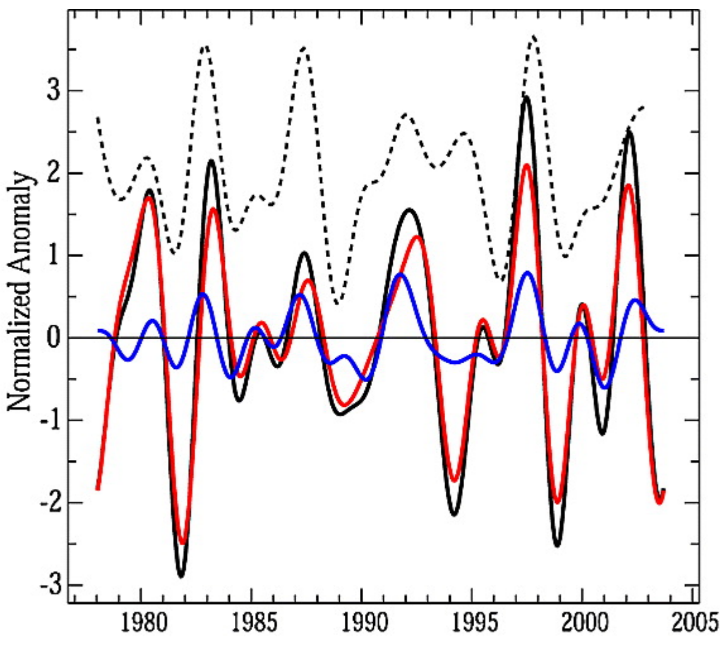


Figure 3.26: Temporal variations at 140°W of the total (black line), stationary (red line), and eastward (blue line) SST. Values are normalized by a standard deviation of the total SST. The Southern Oscillation Index is shown by a micro-dashed line, but with its sign being reversed and its zero axis being displaced by +2 to better compare with peaks of SST (Park et al. 2004).

3.3.2 **Depression Tracks**

The path 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 then their subsequent path determined from further multitemporal imagery. The polar regions are well-suited to this 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 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 3.27.a. A final methodology is to utilise Eulerian storm-track diagnostics by identifying the variance of vorticity at synoptic-timescales (2-6 days): Figure 3.27.b is an equivalent figure to 3.27.a using this method: note the significant difference in results in the CPT with the system-centred approach finding a maximum while the Eulerian method indicates relatively little cyclone activity here. The satellite-imagery study of Turner *et al.* (1998) reveals that there is indeed a maximum in cyclone activity here.

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 3.27.a) the storm track is more asymmetric with a spiral from the Atlantic and Indian Oceans in towards Antarctica and a pronounced STJ-related storm track at ~35°S over the Pacific, spiralling towards southern South America (e.g. Hoskins and Hodges, 2005). A study by Simmonds and Keay (2000b) found that the mean track length of winter systems (2315 km) was slightly longer than in summer (1946 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.

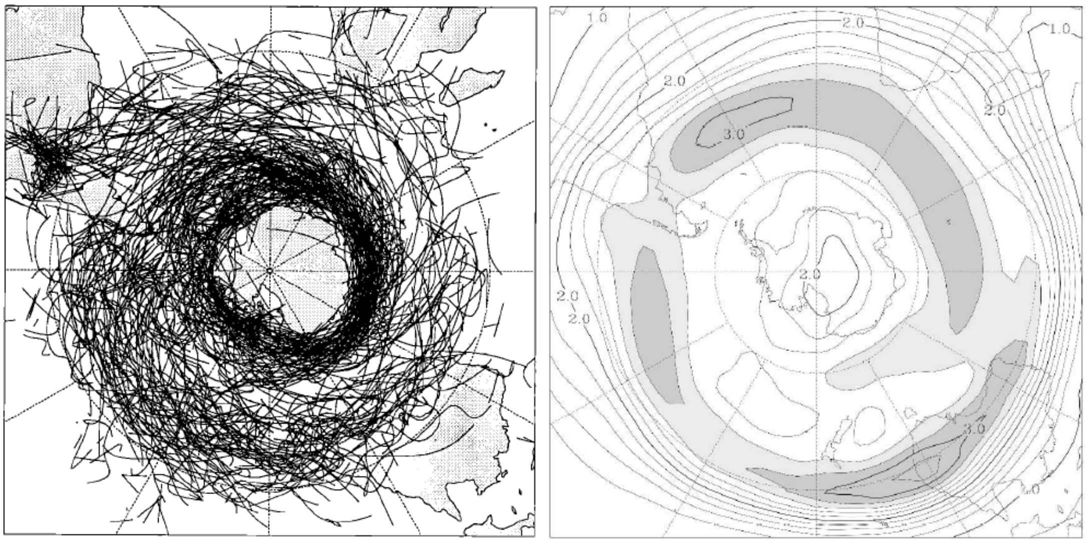


Figure 3.27. (a) Depression tracks for winter 1985-89. From Jones & Simmonds (1993). (b) Bandpass-filtered (2-6 day) variance converted to standard deviation for ξ_{250} for winter 1958-2002. From Hoskins & Hodges (2005).

Using gridded data, Simmonds and Keay (2000a) demonstrated 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 Southern Annular Mode. 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 (Simmonds *et al.*, 2003)

Depression formation and decline

An additional type of information derived from depression tracking studies is the location of cyclogenesis and cyclolysis, or when a weather system is first and last identified, respectively. By collating data from all cyclones examined regions of preferred cyclogenesis and cyclolysis may be determined. Using data from the NCEP-NCAR reanalysis, Simmonds & Keay (2000b) showed there to be a net creation of cyclones (i.e. cyclogenesis > cyclolysis) north of 50°S and net destruction south of this. However, most SH cyclogenesis actually occurs at very high latitudes (~60°S) in the CPT but rates of cyclolysis here are even higher. Turner *et al.* (1998) found that approximately half the systems in the Antarctic Peninsula region formed within the CPT and half spiralled in from further north and work by 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'.

The recent detailed study of Hoskins & Hodges (2005) revealed that the main regional depression tracks around Antarctica can be envisaged as a series of overlapping plates, each composed of cyclone lifecycles. Cyclolysis in one plate appeared to be important for cyclogenesis in the next plate further east through downstream development in the upper-troposphere spiral storm track (Fig 3.28). So for example, the cyclogenesis upstream of the Ross Sea feeds the track towards the Antarctic Peninsula, which, in combination with two more northerly cyclogenesis regions of South America feeds the storm track across the Atlantic sector.

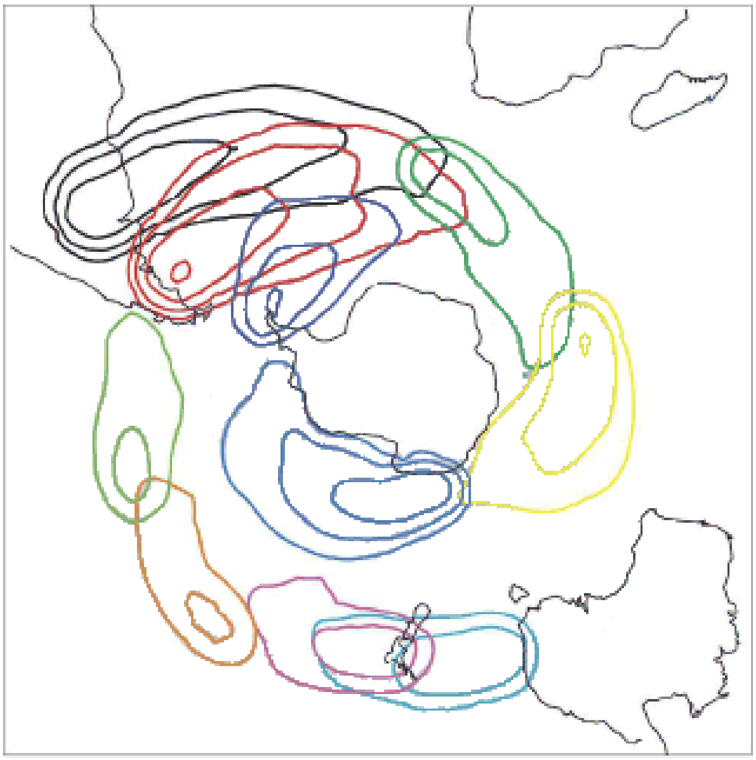


Figure 3.28. Track density contours for the main cyclogenesis regions: contours are 0.5, 1.0, 2.0 and 4.0 tracks per 5° radius per month (after Hoskins and Hodge, 2005).

3.3.3 **Teleconnections**

Atmospheric linkages

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 involved 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. However, during the El Niño phase the storm activity moves close to the date line with the 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.

In the Southern Hemisphere, the most robust signals of ENSO are found across the South Pacific during winter. At this time of year, the Rossby wave train can be identified as anomalously high mean sea level pressure (MSLP) and upper level heights (heights of constant pressure surfaces) across the Amundsen-Bellingshausen

Sea (ABS), with anomalously low values to the east of New Zealand (Fig. 3.29) (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. This atmospheric pattern of high and low pressure anomalies is known as the Pacific South American association (PSA). A comparable series of anomalies is also found in the Northern Hemisphere during El Niño events called the Pacific North American association (PNA).

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).

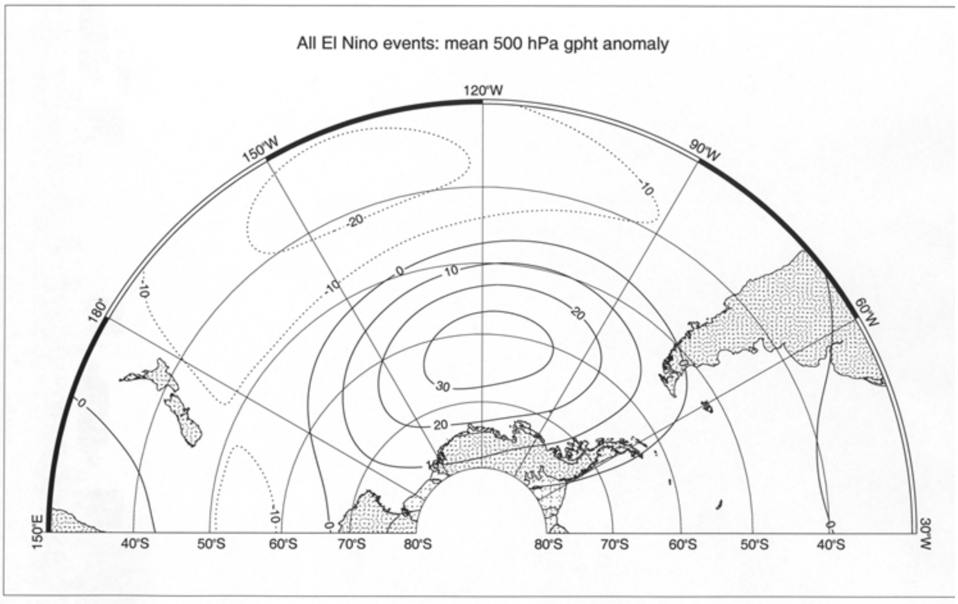


Fig. 3.29. The 500 hPa height anomalies during the Winter-season (June–August) for eight El Niño events over the period 1968–99. The figures were derived from NCEP–NCAR reanalysis data

A major difficulty in studying this teleconnection is that while many of the El Niño and La Niña events give the atmospheric anomaly patterns described above, some marked events have resulted in very different conditions. For example, the 1982/83 El Niño event was one of the most significant 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 PNA making it much more difficult to anticipate how the Antarctic atmosphere will respond as an El Niño event starts to become established. Recently, Lachlan-Cope and Connolley (2006) showed with ensemble runs of a climate model that changes due to the natural variation of the zonal flow in the Southern Hemisphere can swamp the signal resulting from changes in the Rossby wave source region during El Niño events.

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. 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. On the other hand, the Antarctic Peninsula, far from cooling, has experienced the largest increase of temperature anywhere in the Southern Hemisphere (Turner *et al.*, 2005).

In summary, 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. 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.

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 sea surface temperature (SST) 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 (Fig. 3.30).

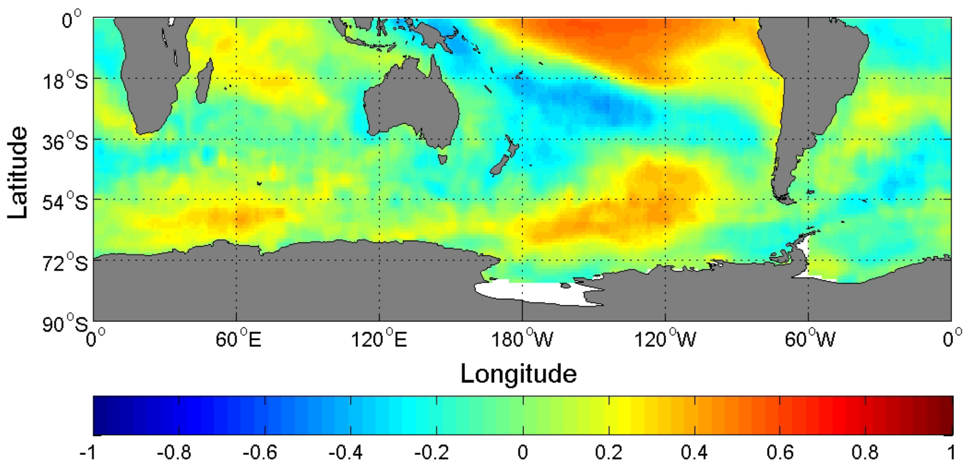


Fig. 3.30. Correlation of ENSO with sea surface temperature (SST) across the Southern Ocean. Note the positive SST in the South Pacific that coincides with an El Nino event; this warm water is subsequently advected eastward within the Antarctic Circumpolar Current (ACC). From Meredith et al. (2007)

The SST anomalies so created lie largely within the domain of the Antarctic Circumpolar Current (ACC), and hence are subsequently advected eastward. A typical timescale for advection across the South Pacific and into the South Atlantic is approximately 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, 2002).

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).

Although the most widely-known, 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. These winds are the driving force for the ACC, and it was demonstrated that changes in the ACC were induced in response. The total timescale for MJO to influence ACC transport was of order 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.

3.4 **Temperature**

3.4.1 Surface temperature

The in-situ observational record of Antarctic surface temperatures is rather sparse and sporadic before the International Geophysical Year (IGY) (see Appendix A in King and Turner, (1997)), although the Orcadas series from Laurie Island, South Orkney Islands begins in 1903 and the Faraday Station/Argentine Islands record begins in 1947. However, we are fortunate in have 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). However, 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 now online and can be accessed 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. (2005), 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. Table 3.1 gives the annual and seasonal surface temperature trends for 19 stations that have long, reliable records.

Table 3.1 Annual and seasonal surface temperature trends ($^oC (10 yr)^{-1}$) at selected Antarctic stations. [a]= Significant at the 1% level. [b]= Significant at the 5% level. [c]= Significant at the 10% level. Trends included if 90% of the observations are available. S = synoptic data, C = CLIMAT, G = GTS. ** Indicates some years missing.

Station	Annual	Spring	Summer	Autumn	Winter	Period	Data used
Novolazarevskya	+0.25 ± 0.27[c]	+0.25 ± 0.41	+0.19 ± 0.34	+0.08 ± 0.46	+0.44 ± 0.66	1962-2000	S & C
Syowa	+0.01 ± 0.35	+0.01 ± 0.48	–0.01 ± 0.27	–0.12 ± 0.67	+0.14 ± 0.58	1960-61 1967-2000	S
Molodezhnaya	–0.06 ± 0.29	–0.18 ± 0.50	–0.14 ± 0.32	–0.21 ± 0.43	+0.30 ± 0.77	1964-95, 1997-98	C & G
Mawson	–0.11 ± 0.23	–0.04 ± 0.33	–0.09 ± 0.26	–0.30 ± 0.40	+0.03 ± 0.58	1955-2000	S & G
Davis	+0.03 ± 0.35	+0.05 ± 0.50	+0.05 ± 0.30	–0.26 ± 0.60	+0.15 ± 0.67	1958-63, 1970-2000	S & G
Mirny	–0.01 ± 0.26	+0.09 ± 0.46	–0.14 ± 0.30	–0.28 ± 0.45	+0.31 ± 0.56	1956-2000	S & G
Vostok	–0.02 ± 0.34	-0.11 ± 0.51	+0.13 ± 0.42	–0.32 ± 0.63	+0.14 ± 0.85	1958-2000	C & G
Casey	+0.01 ± 0.40	+0.13 ± 0.50	–0.09 ± 0.30	–0.14 ± 0.90	+0.22 ± 0.83	1962-2000	S & G
Dumont d'Urville	+0.02 ± 0.27	+0.23 ± 0.45	0.00 ± 0.31	–0.34 ± 0.35[c]	+0.17 ± 0.60	1956-2000	S & C
Scott Base	+0.29 ± 0.36	+0.34 ± 0.68	+0.05 ± 0.38	+0.18 ± 0.65	+0.43 ± 0.71	1958-2000	C
Rothera	+1.01 ± 1.42	+1.06 ± 1.53	+0.36 ± 0.57	+1.37 ± 1.46	+1.73 ± 2.79	1978-2000	S
Faraday/Vernadsky	+0.53 ± 0.31[b]	+0.28 ± 0.29	+0.23 ± 0.10[a]	+0.55 ± 0.35[b]	+1.03 ± 0.61[b]	1951-2006	S
Bellingshausen	+0.35 ± 0.46	–0.10 ± 0.47	+0.30 ± 0.20[a]	+0.51 ± 1.05	+0.58 ± 0.97	1969-2000	S
Esperanza	+0.34 ± 0.16[c]	+0.22 ± 0.18	+0.41 ± 0.11[b]	+0.43 ± 0.37	+0.29 ± 0.31	1946-2006**	C
Marambio	<90%	-0.8 ± 10.5	<90%	<90%	+0.81 ± 1.53	1971-2000	C

Orcadas	+0.20 ± 0.10[a]	+0.15 ± 0.14[b]	+0.15 ± 0.06[a]	+0.21 ± 0.16[a]	+0.27 ± 0.24[b]	1904-2000	C
Halley	−0.11 ± 0.47	0.00 ± 0.53	+0.12 ± 0.28	−0.56 ± 0.81	+0.02 ± 0.76	1957-2000	S
Neumayer	–0.13 ± 1.03	–0.01 ± 1.69	–0.02 ± 1.25	–1.37 ± 1.32[b]	+0.30 ± 2.23	1982-2000	S
Amundsen-Scott	–0.17 ± 0.21[c]	–0.12 ± 0.63	–0.21 ± 0.49	–0.19 ± 0.45	–0.20 ± 0.50	1958-2000	S

Surface temperature trends across the Antarctic 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 (Fig. 3.31). 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 (10 yr^{-1}) 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, with the long record from Orcadas on Laurie Island, South Orkney Islands only having experienced a warming of +0.20 ^{o}C (10 yr^{-1}). However, it should be noted that 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 (10 yr^{-1}).

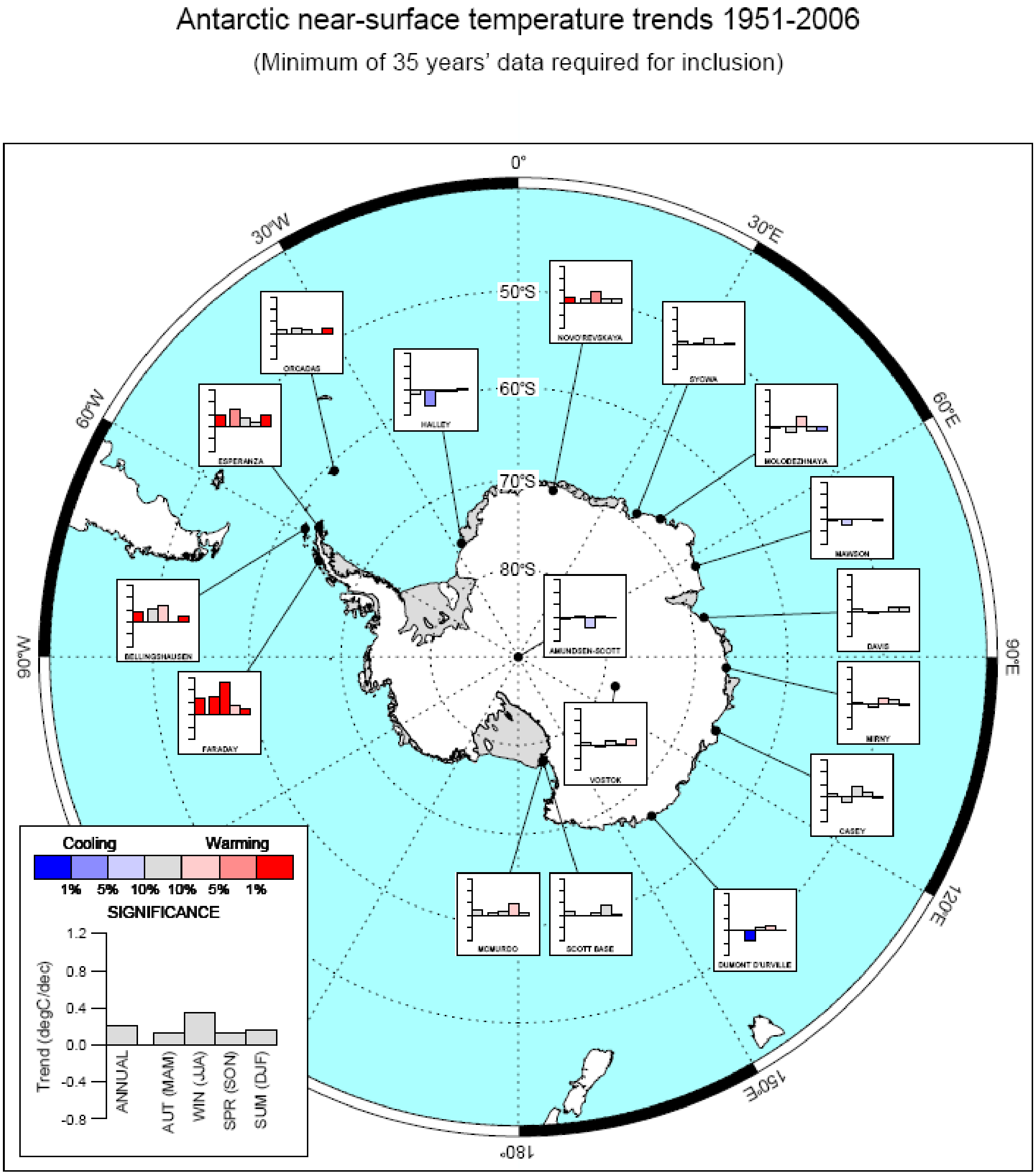


Fig. 3.31. Near-surface temperature trends for 1951-2006.

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). They 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 ^{o} C (10 yr^{-1}) 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 experiences a summer increase of +0.41 ^{o} C (10 yr^{-1}) over 1946-2006. This temperature rise has been linked to a strengthening of the westerlies that has taken place as the Southern Hemisphere Annular Mode 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 (Table 3.1). The largest warming outside the peninsula region is at Scott Base, where temperatures have risen at a rate of +0.29 °C $decade^{-1}$, although this is not statistically significant. The high spatial variability of the changes is apparent from the data for Novolazarevskya and Syowa, which are 1000 km apart. The former station has warmed at a rate of +0.25 °C $decade^{-1}$ over 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 is the McMurdo Dry Valleys. Here automatic weather station data for 1986-2000 shows that there has been a cooling of 0.7^{o} C per decade, with the most pronounced cooling being in the summer (December-February = 1.2^{o} C per decade, P = 0.02) and autumn (March-May = 2.0^{o} C per decade, P = 0.11). Winter (June-August) and spring (September-November) show small temperature *increases* of (0.6^{o} C and 0.1^{o} C per decade, P = 0.62 and 0.95, respectively) (Doran *et al.*, 2002).

On the interior plateau, Amundsen-Scott Station at the South Pole has shown a statistically significant cooling in recent decades that is thought to be a result of fewer maritime air masses penetrating into the interior of the continent. 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.W.L. 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 regions, 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 $decade^{-1}$ while summer warming rates were only +0.045°C $decade^{-1}$. For the 45 year period the temperature trend in the annual means was +0.082°C $decade^{-1}$. Interestingly the trends computed were very sensitivity 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.

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 Southern Annular Mode (SAM) into its positive phase (see Section xxx). 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 (Arblaster and Meehl, 2006). 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 (Fig. 3.32). They only considered the months of December to May, which was when the largest change in the SAM has taken place. They attribute 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, is found 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.

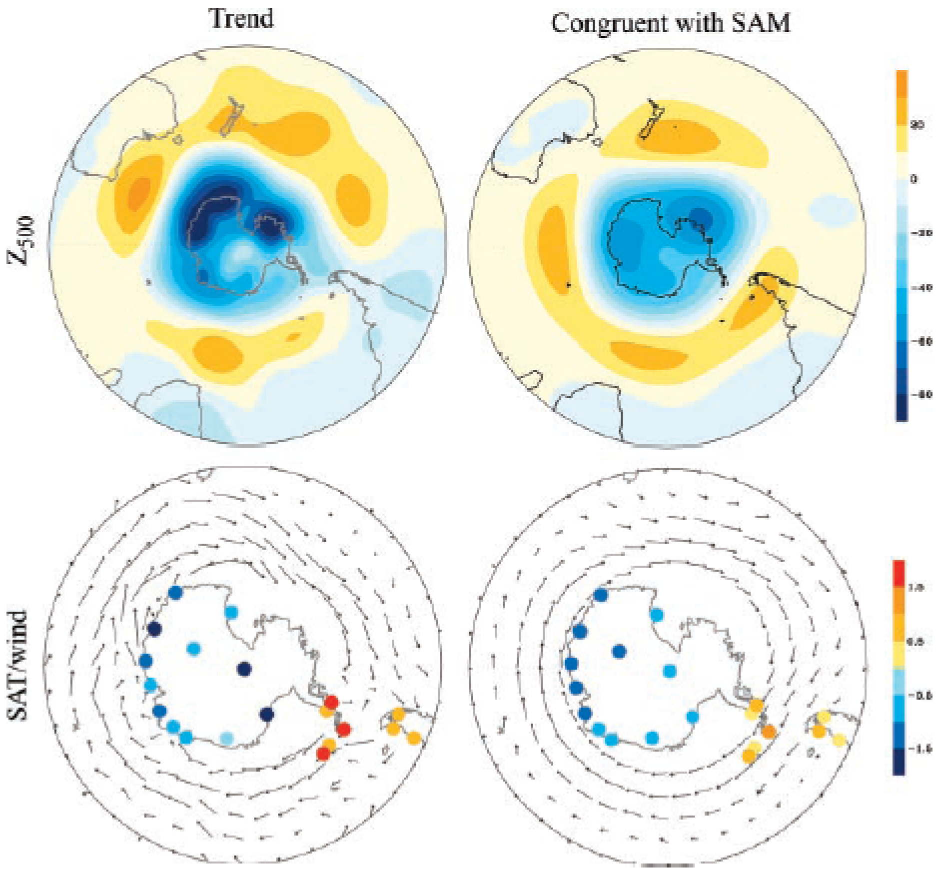


Fig 3.32. December-May trends (left) and the contribution of the SAM to the trends (right). Top, 22-year (1979-2000) linear trends in 500-hPa geopotential height. Bottom: 32-year (1969-2000) linear trends in surface temperature and 22-year (1979-2000) linear trends in 925-hPa winds. Shading is drawn 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).

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 automatic weather stations (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 (Fig. 3.33). 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, with the former being consistent with the trends derived by Thompson and Solomon, although the trends in Fig. 3.33 are for annual data and the Thompson and Solomon study is concerned only with the summer and autumn. However, the cooling across the Antarctic Peninsula in Fig. 3.33 is surprising since this area has experienced a marked warming in recent decades, although the in-situ observations are all from low-level coastal stations.

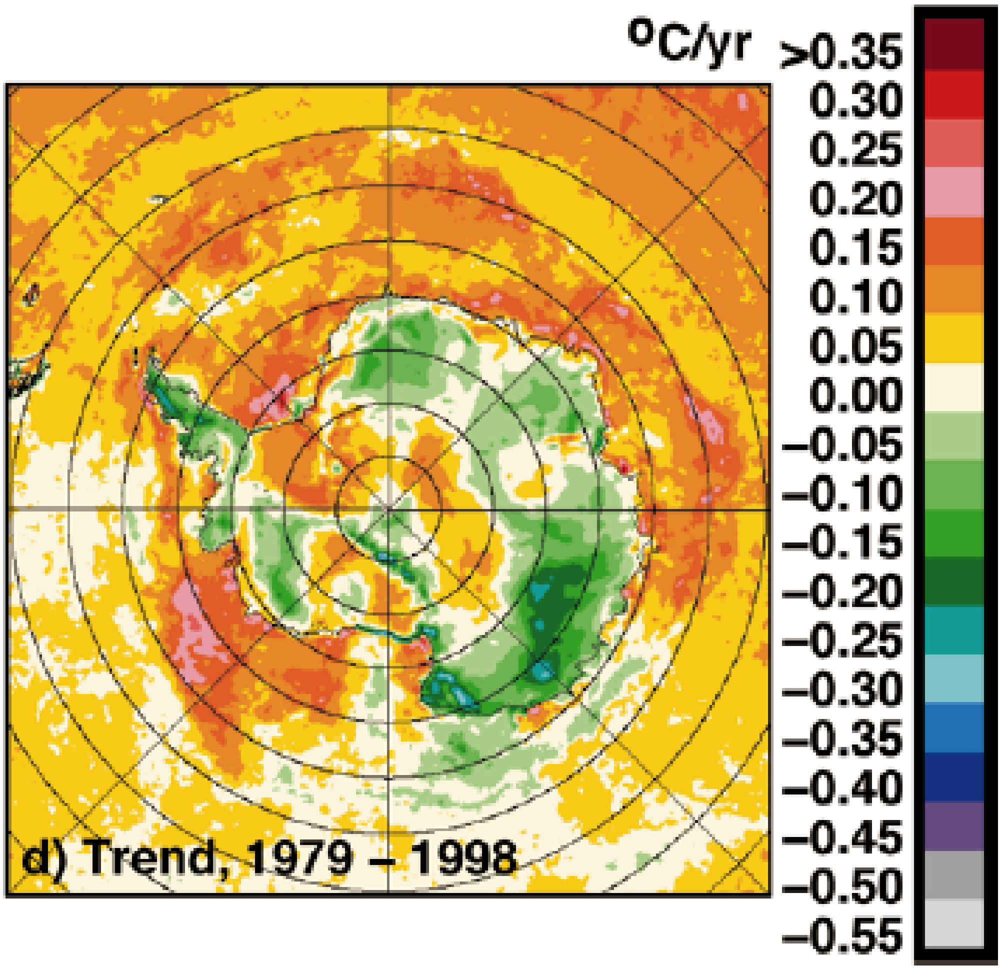


Fig. 3.33. Temperature trends determined from satellite imagery. The trends were derived using the January and July monthly average data from 1979 to 1998. From Comiso (2000).

In recent decades many relatively short ice cores have been drilled across the Antarctic by initiatives such as the International Trans Antarctic Science Expedition (ITASE) (Mayewski and 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.

3.4.2 Upper air temperatures

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 – 0.7° C dec^{-1} (Fig. 3.34) (Turner *et al.*, 2006). Fig. 3.34 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.

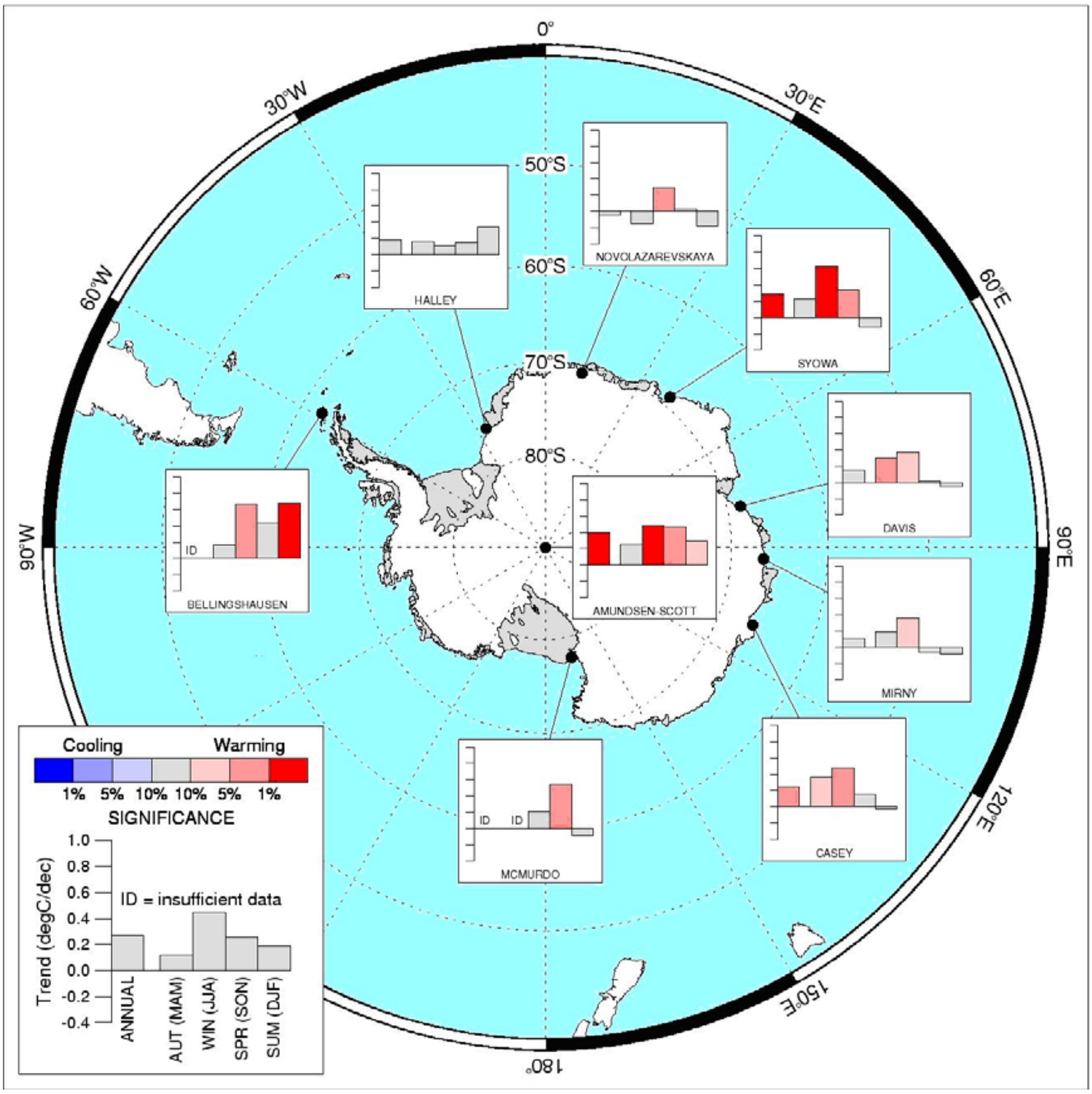


Fig 3.34. Annual and seasonal 500 hPa temperature trends for 1971-2003.

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. PSCs are a feature of the cold Antarctic winter, forming at temperatures below about 195K. However, the Antarctic stratosphere has cooled in recent decades because of the Antarctic ozone hole and rising levels of greenhouse gases. Analysis of stratospheric temperatures in the reanalysis data sets have suggests 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.

3.4.3 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. However, we still do not know the reasons for the large winter season warming on the western side.

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 greenhouse gas increases, along with a side effect of the ozone hole. However, more research is needed to confirm this.

3.5 Changes in Antarctic Snowfall Over the Past 50 Years

3.5.1 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/vapor deposition, drifting snow processes, and melt. Precipitation, which primarily occurs as snowfall, is dominant among these components (Bromwich 1988) and establishing its spatial and temporal variability is necessary to assess ice sheet surface mass balance. Comprehensive studies of snowfall characteristics over Antarctica are given by Bromwich (1988), Turner et al. (1999), Genthon and Krinner (2001), van Lipzig et al. (2002), Bromwich et al. (2004a), van den 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

(Fig. 3.35) 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 et al. 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 (Schwerdtfegar 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 El Nino-Southern Oscillation (ENSO) (Cullather et al. 1996) and the Southern Hemisphere Annular Mode (SAM) (Genthon et al. 2003, van den Broeke and van Lipzig 2004). ENSO has an intermittent teleconnection with Antarctica 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).

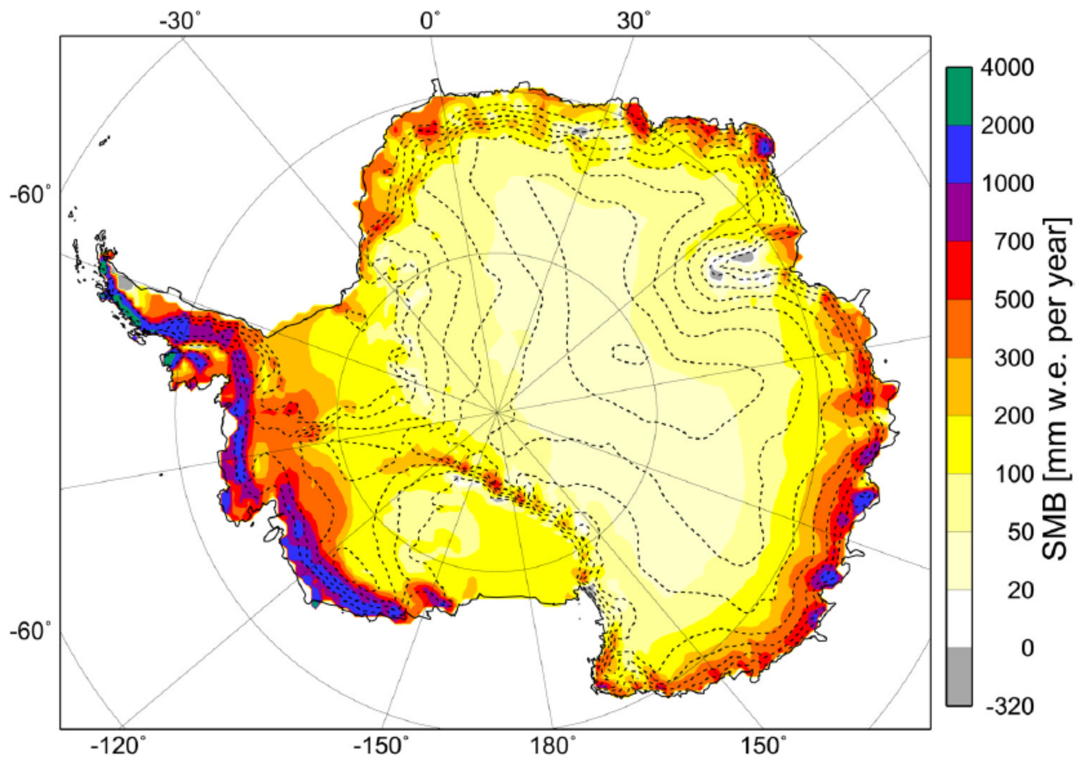


Fig. 3.35. Annual Antarctic surface mass balance (mm y^{-1}), from van de Berg et al. (2006).

3.5.2 **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 y^{-1} (van de Berg et al. 2005) to +197 mm y^{-1} (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 endeavor 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=149 mm yr^{-1} 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^{-1}. Considering the large spread between estimates, it is not surprising that calculated temporal trends vary widely (Monaghan et al. 2006a).

3.5.3 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 indicate that no statistically significant increase has occurred since ~1980 over the entire grounded ice sheet, WAIS, or the East Antarctic Ice Sheet (EAIS) (Monaghan et al. 2006a, van de Berg et al. 2005, van den Broeke et al. 2006a). A validation of the modeled-versus-observed changes (Monaghan et al. 2006a) suggests that the recent model records are more reliable than the earlier global model records that inferred an upward trend in Antarctic 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 y^{-1} water equivalent (WEQ), i.e., +/-0.69 mm y^{-1} GSL equivalent, are common (Monaghan et al. 2006a), and might easily mask underlying trends over the short record.

In contrast to modeling studies, satellite altimetry measurements by Davis et al. (2005) suggest that increased snowfall has recently caused the East Antarctic Ice Sheet (EAIS) to thicken, mitigating sea level rise by about 0.12 mm y^{-1} 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 3 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) use 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 y^{-1}), and is characterized by upward trends through the mid-1990s and downward trends thereafter (Fig. 3.36). The shape of the time series in Fig. 3.36 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 (Fig. 3.37). 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. 2005). The SMB increases, among other climate change factors, have been linked to observed decreases in Adélie penguin populations on the western side of the Antarctic Peninsula, by means of decreasing the availability of snow-free nesting habitat required by the birds (Fraser and Patterson 1997).

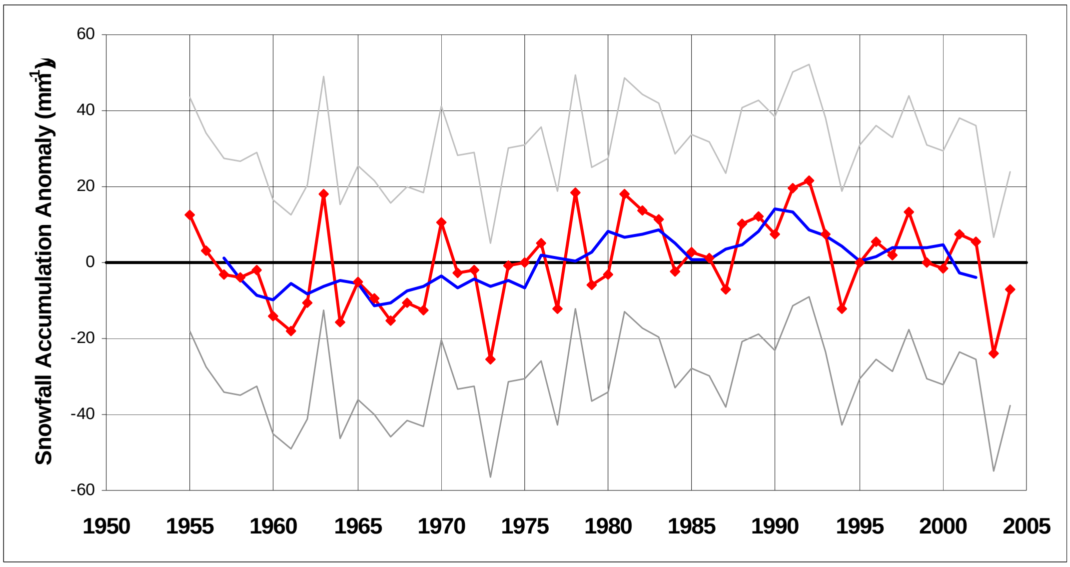


Fig. 3.36. Annual Antarctic snowfall accumulation anomalies (mm y^{-1}, in red) during 1955-2004, adapted from Monaghan et al. (2006b) and Monaghan and Bromwich (2008). Anomalies are with respect to the 1955-2004 mean. The five year running mean is shown in blue. The ±95% confidence intervals for the annual anomalies are shown in gray. The statistical uncertainty accounts for the variance, as well as that due to the methodology and measurement error.

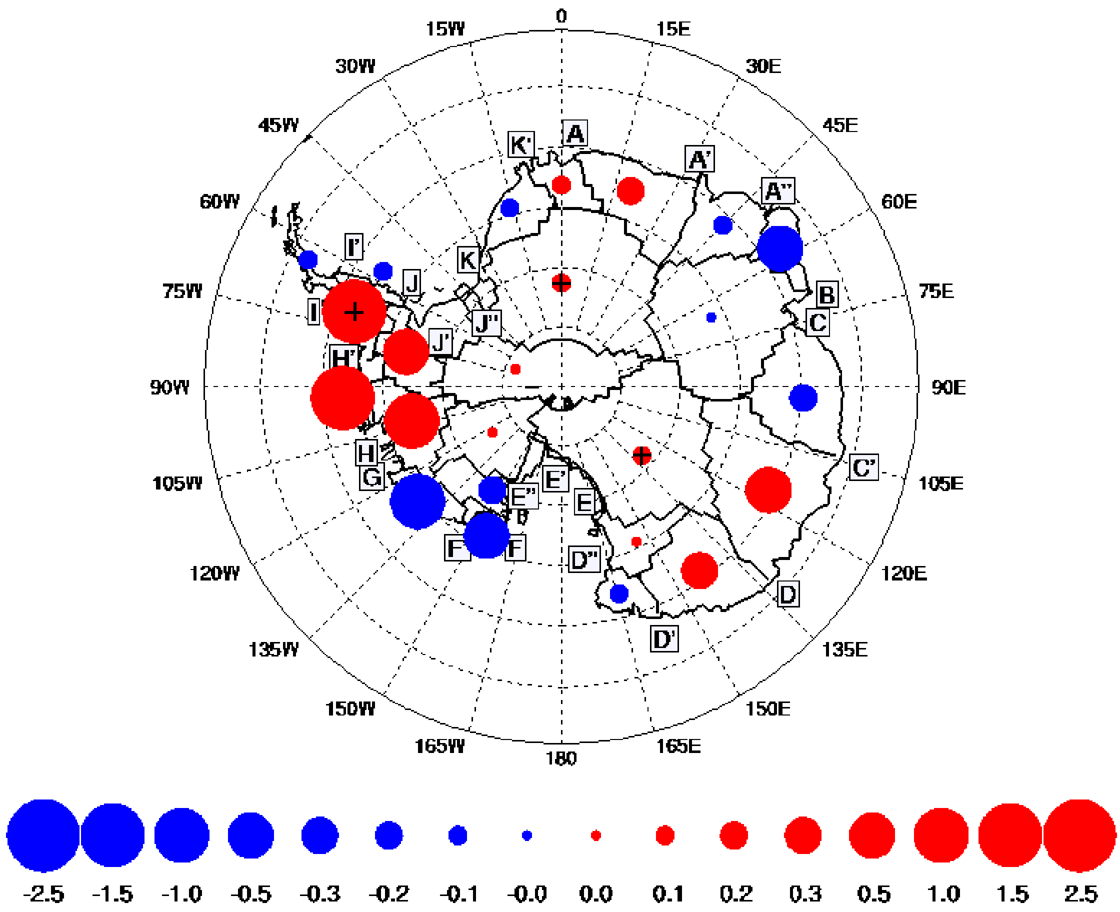


Fig. 3.37**.** Linear trends of annual Antarctic snowfall accumulation (mm y^{-2}) for 1955-2004 in each of the Antarctic glacial drainage basins, adapted from Monaghan and Bromwich (2008). Statistical significance is represented by symbols: "+" is $p<0.10$; "*" is $p<0.05$; and "Δ" is $p<0.01$. The statistical uncertainty accounts for the variance, as well as that due to the methodology and measurement error.

3.5.4 The changing nature of Antarctic precipitation

As surface temperatures have risen across the Antarctic Peninsula in recent decades so the percentage of precipitation falling has done so in the form of rain.

3.6 Atmospheric Chemistry

3.6.1. Antarctic stratospheric ozone in the instrumental period

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-stirred 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 km lower in altitude because of the reduced convective stirring 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:

$O_2 + h\nu \rightarrow 2O$	wavelength < 240 nm
$O_2 + O + M \rightarrow O_3 + M$	where a third molecule, M, is required to take away 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 prisms and a series of slits. A well-calibrated instrument can measure ozone amount 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 be just 3 mm thick.

Vertical profiles of ozone are measured 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, decreasing above.

Fig. 3.38. 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].

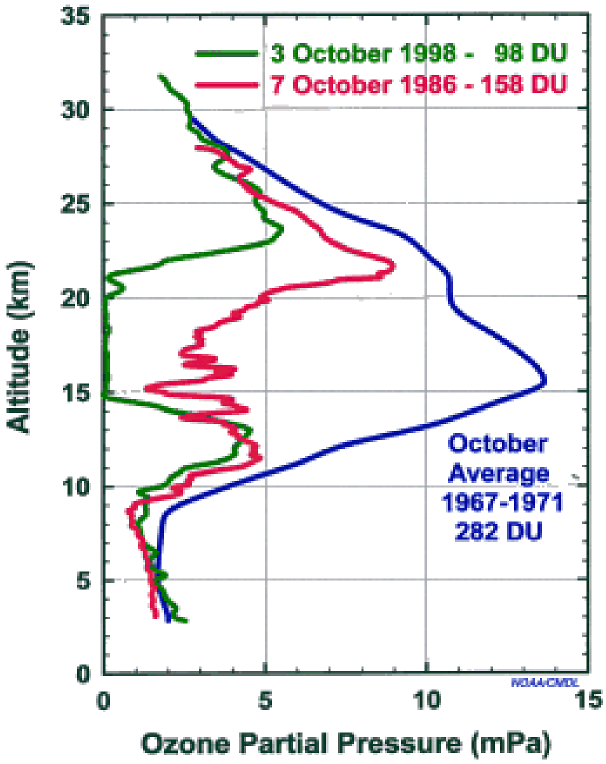


Fig. 3.39. South pole ozone profiles, showing the progressive thinning of the ozone layer in late spring as the ozone hole developed during the 1980s and 1990s [courtesy NOAA/CMDL].

In common with most Antarctic data sets, that of total ozone in Antarctica starts in 1956 when stations began operation in the International Geophysical Year of 1957 to 1958, and some later 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 (Fig. 3.38) 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).

Historically ozone values were around 300 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. The majority of this loss occurs between 14 and 22 km altitude within the polar vortex, where virtually all ozone is now destroyed (Figure 3.39). Ozone values substantially recover with the warming in late spring, when the vortex dissipates and air from outside is mixed (Figure 3.40).

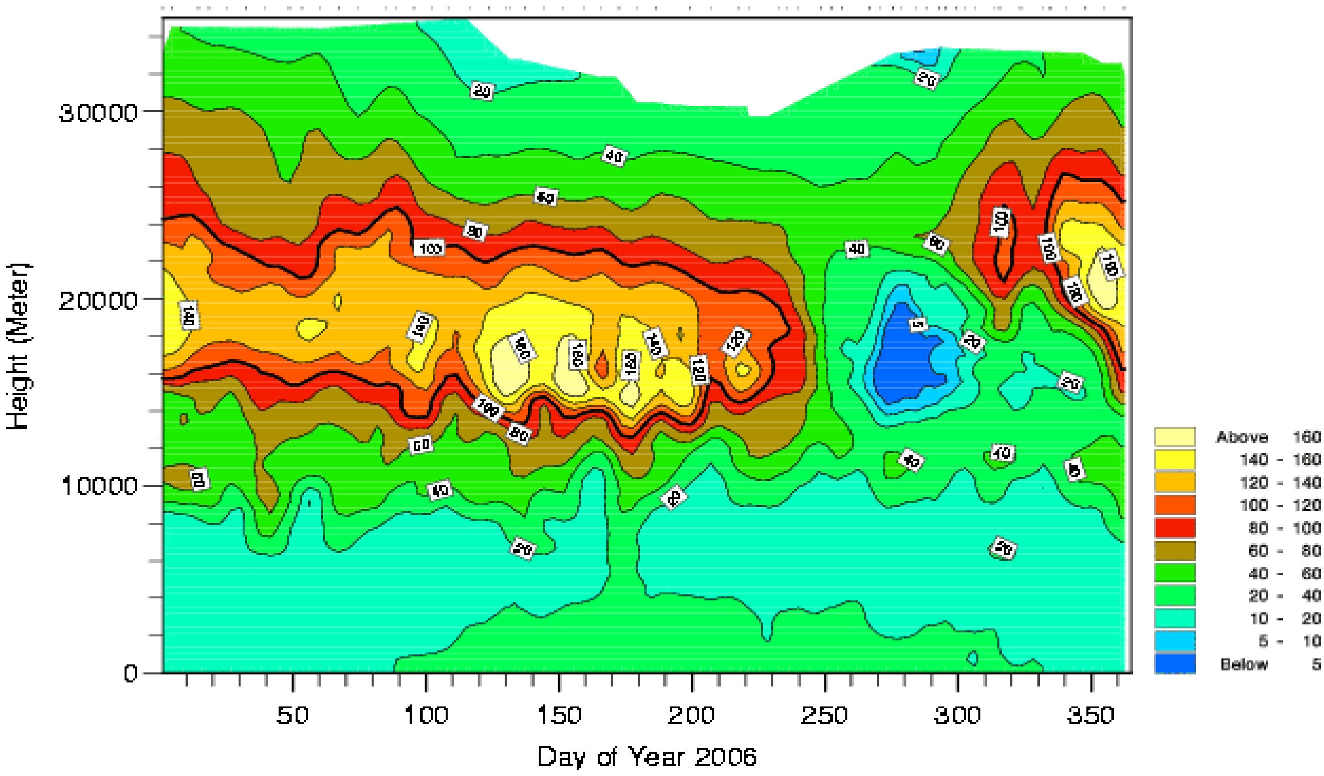


Fig. 3.40. The annual cycle of ozone (nbar) in 2006 at Neumayer, 71°S [courtesy Alfred Wegener Institute].

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 3.41a). 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° colder. On the cloud surfaces, the degradation products of the CFCs react to form chlorine gas:

$$HCl + ClNO_3 \rightarrow HNO_3 + Cl_2$$

followed by 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 that catalyse

$$2O_3 \rightarrow 3O_2$$

in which they are 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 320 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.

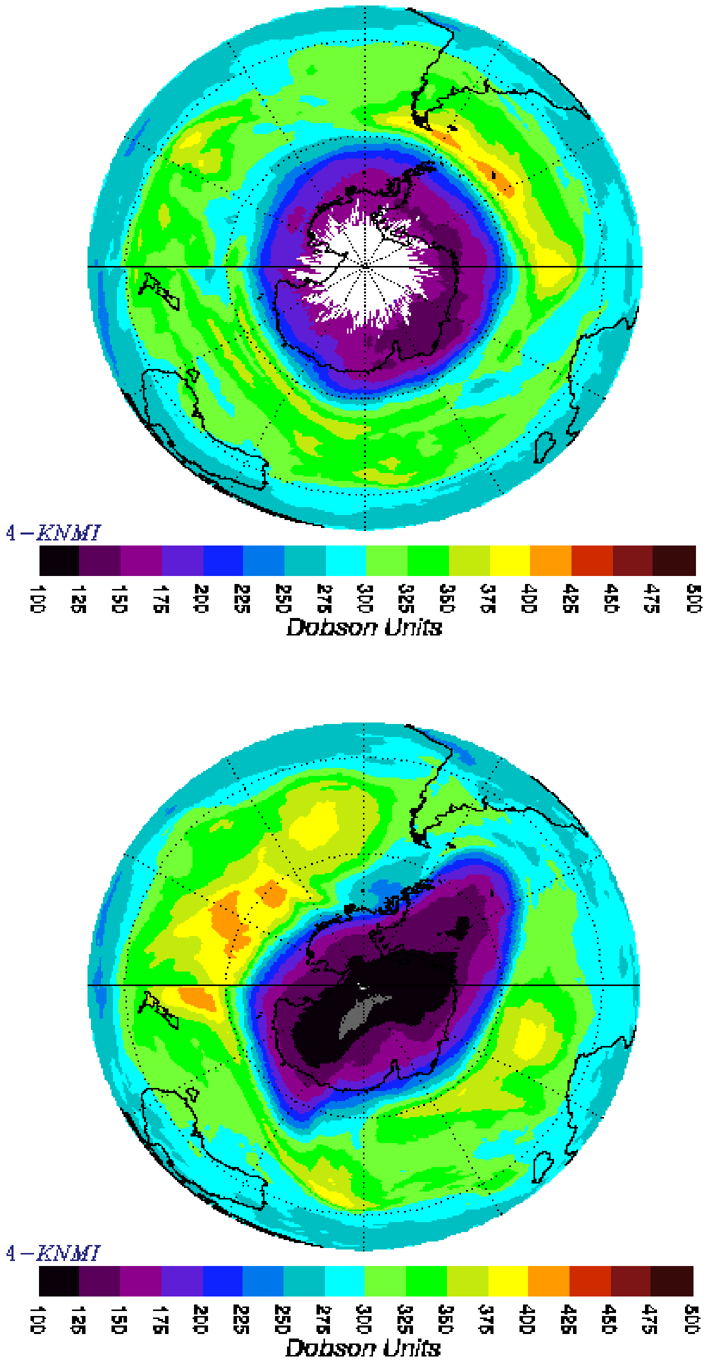
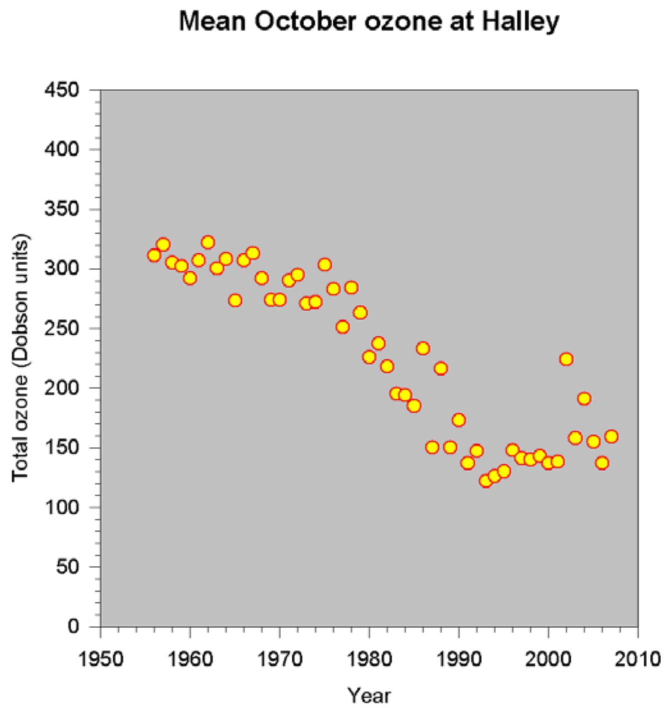


Fig. 3.41. Measurements by the Ozone Monitoring Instrument on the Aura satellite, with a scale in DU. Top – on 14 September 2006, showing an almost symmetric ozone hole covering all of Antarctica. Bottom – on 10 October 2006, showing the ozone hole extending towards South America, where it puts people at risk of damage from the sudden increase in UV-B, which can cause skin cancer and cataracts [courtesy NASA/GSFC].

Another effect of the ozone hole is on the temperatures within the vortex. In the sunlit stratosphere, temperature is strongly correlated with ozone because of ozone's strong absorption of sunlight. The ozone hole has therefore resulted in significant reductions in temperature in spring (Figure 3.42), in November reaching a difference of up to 15°C since 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 little 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 (ozone mass deficit, 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.



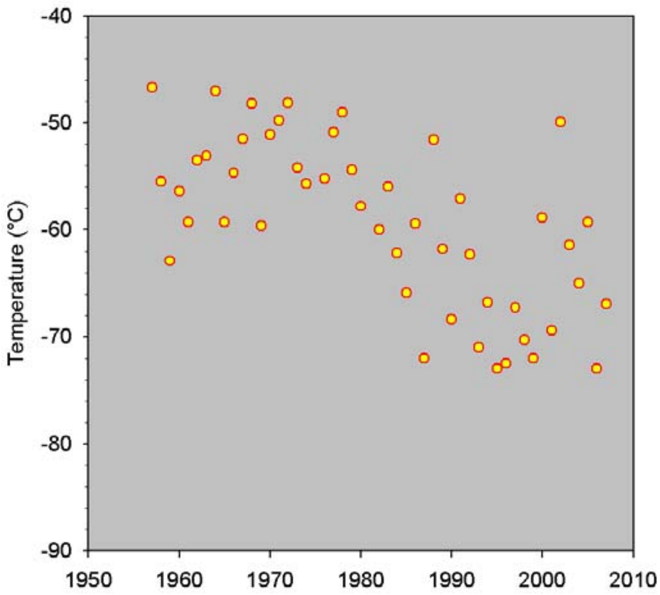


Fig. 3.42. Time series of measurements at Halley (76°S) of (top) total ozone averaged over October of each year, and (bottom) temperature at 100 hPa averaged over November of each year [courtesy British Antarctic Survey]. Although there is a lot of variability in the temperatures, they are now about 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.

3.6.2 **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, in general, 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 altitude over the polar regions. The troposphere itself is nominally subdivided into layers; the lowest is the "boundary layer", a part of the troposphere which 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 meters. 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 industrialised 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. By contrast, Antarctica supports no major population centres and lies at considerable distance from anthropogenic emission sources. 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 perceived idea that the chemical composition would be relatively uninteresting, with low concentrations of reactive trace gases like the hydroxyl or hydroperoxy radical (OH or HO_2) 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.

However, 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. Furthermore, it would be cavalier to believe we could correctly reconstruct a past atmosphere without properly understanding that of the present day.

The earlier 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. They have shown that the majority of sea salt is generated within the zone of newly-forming sea ice. This suggests that sea salt measured in ice cores can 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.

But 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, and its potential response to climate change is discussed in section 5.3.2.

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 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 3.43). At nearly all stations, surface ozone reaches its maximum concentration during the winter months and is at its minimum during the summer.

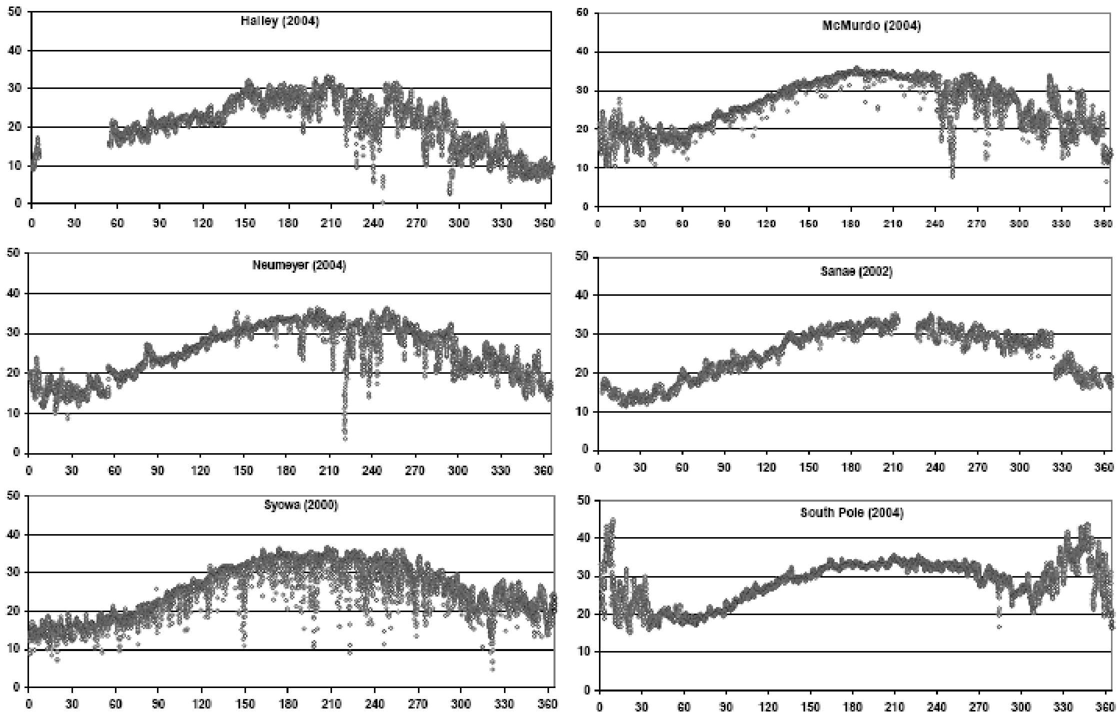


Figure 3.43. Annual records (against year calendar day) of surface ozone (in parts per billion by volume (ppbv)) for the Southern Hemisphere stations Halley, MacMurdo, Neumayer, Sanae, Syowa and South Pole, reproduced from Helmig et al., (2007)

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.

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.

It is now known that this behaviour is a natural phenomenon 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 HO2 can also be involved in BrO recycling but these are generally less important than the BrO self-reaction. 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 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 Fig. 3.44), are the source.

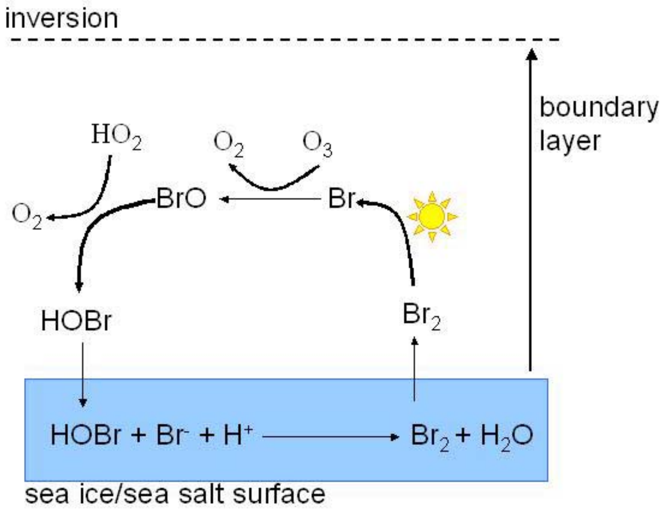


Fig. 3.44. A "Bromine Explosion".

Bromide (Br^-) is a ubiquitous ion found in sea water, albeit at concentrations ~280 times less 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 the 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 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 is it only evident at South Pole.

The answers lie within a new area of atmospheric science; that of Snow Photochemistry (Grannas et al., 2007). 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, laboratory and through 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) which 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 snowcovered 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. How they may behave in a future warmer world, is explored in section 5.3.2.

3.7 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 & Chown 1999). While the majority 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 2007 CEP workshop report). 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 RiSCC LH). 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 Antarct Sci) 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 RiSCC). 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 (Bale ref). How these patterns change in the future will be an area of major importance.

The final suite of life history traits are those 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. 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, 2007). Regional warming has also been predicted to rapidly increase the impact of certain indigenous predators (Arnold & 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 & 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.

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 Ant Res Ser, 2007 RiSCC). 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 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 – *check this paper*) 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 PPRST).

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*) [insert illustration of these] in the maritime Antarctic (Fowbert and Smith 1994, Smith 1994, Grobe et al. 1997, Gerighausen et al. 2003). At some sites numbers of plants have increased by two orders of magnitude in as little as 30 years, 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 1996c), and stimulate growth of seeds that have remained dormant in soil propagule banks (McGraw and Day 1997).

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 RiSCC, Lyons et al. 2006 RiSCC, 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 RiSCC).

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 2007 CEP). 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 program 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.

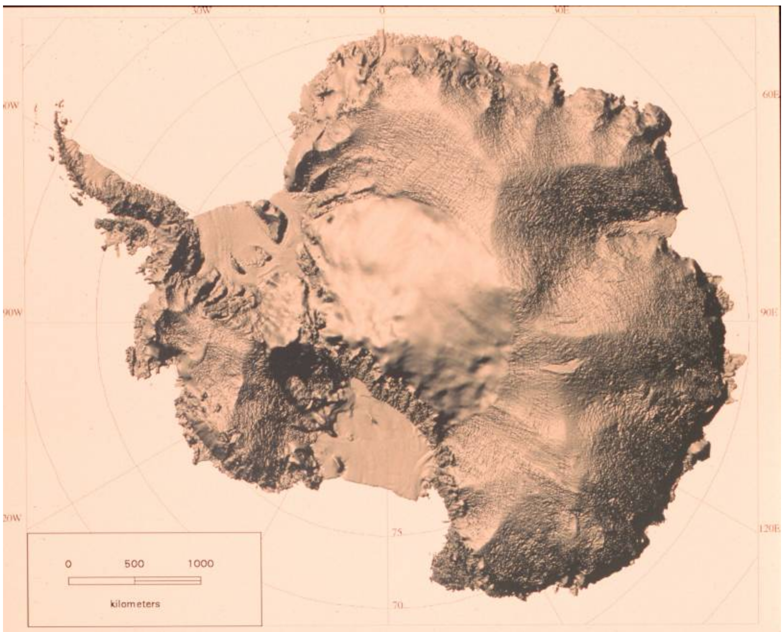
3.8 **The Antarctic Terrestrial Cryosphere**

Ice dominates the Antarctic continent, covering 99.6% of the surface area [Fox and Cooper, 1994]. The vast majority of the ice forms a continuous ice sheet thousands of meters thick in the interior and drained by numerous outlet glaciers and ice streams that usually terminate in floating ice shelves hundreds of meters thick. Isolated glaciers and small ice caps populate the coastal islands and archipeligoes. Since measurements were possible, 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 that include zero, however this belies the enormous regional variability of mass changes that express different dynamic characteristics and responses to different climate forcings. To distinguish dissimilar styles of ice sheet behaviour and to characterize differences in the ice sheet's responses to changes it is useful to separate the ice sheet into three geographic provinces.

3.8.1 **Provinces**

East Antarctica

The East Antarctic ice sheet (EAIS) comprises by the far the largest part of the ice sheet. Lying primarily in the Eastern Hemisphere, it is the part of the ice sheet bounded by Filchner Ice Shelf and the Transantarctic Mountains (roughly the slice from 30° W clockwise through the Greenwich meridian, and on to around 165° E). The EAIS contains the coldest ice and is frozen to the bed over most of its area, which restricts its rate of flow. It lies on a largely contiguous landmass lying predominantly above sea level with a few broad basins depressed below sea level largely due to the weight of the ice sheet (Fig. 3.45).



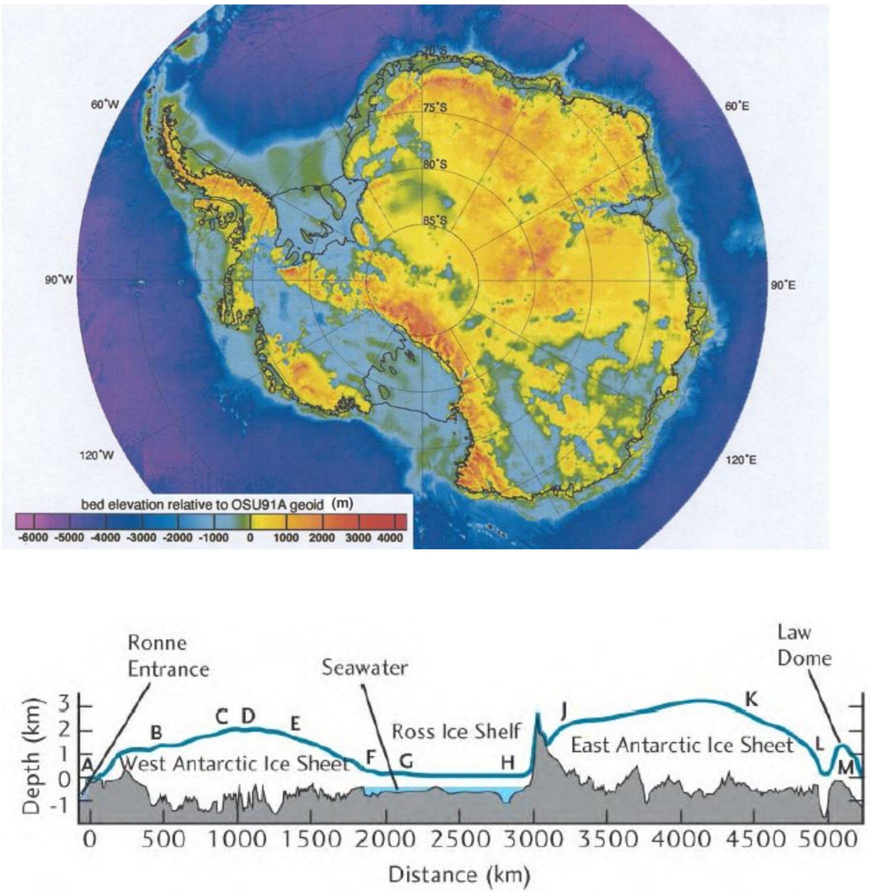


Figure 3.45. The (top) surface and (middle) bed topography of Antarctica, along with a (bottom) cross-section.

West Antarctica

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 30^{o} W (but not including the Antarctic Peninsula). In contrast to EAIS, most of this part of the ice sheet rests on a bed that is substantially below sea level, and would remain so even if the ice sheet were removed. For this reason the West Antarctic ice sheet is described as a "marine ice sheet" and as such is suspected to be capable of some unique behaviour. In many areas, the bed beneath the WAIS is over one thousand metres below sea level – far deeper than most of the world's other continental shelves. A few archipelagoes (Ellsworth Mountains, Executive Committee and Flood Ranges, Whitmore Mountains and the Ellsworth Land coast) are sufficiently tall to rise above the ice sheet. The WAIS is generally warmer, both at the surface and the bed, than its East Antarctic neighbour with basal ice close to the melting point in most areas.

The Antarctic Peninsula

The ice sheet that covers the Antarctic Peninsula is quite different from either EAIS or WAIS, so is usefully discussed separately. It consists of much smaller and much thinner ice caps that cover the central mountainous spine of the peninsula, 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 WAIS and 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 ice budget of the ice sheet in this area.

3.8.2 **Changes**

Ice sheets, like other glaciers, are to some degree self-regulating systems. Increasing snowfall over the ice sheet, will act to increase the ice thickness, but this increase in thickness will also act to 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, however it is rarely realised; the driving environmental parameters of climate are themselves constantly changing, which continually modify the equilibrium state sought by the ice sheet. At any given time, the changes in the ice sheet are a superposition of responses to recent and long-term changes in climate, and for this reason, areal extent, magnitude and duration of changes, as well as time scale all must be carefully considered in any discussion of ice-sheet change.

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: 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 in the Antarctic ice sheet come not from accounting methods, but rather 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 (Fig. 3.46). A number of research groups have evaluated data beginning in the early 1980's; spanning data from multiple satellite altimeters, they have produced broadly consistent results (e.g., Wingham *et al.*, 1998; Davis *et al.*, 2005; Zwally *et al.*, 2005; Wingham *et al.*, 2006). These results illustrate that separate catchment basins within the ice sheet behave somewhat independently. What altimetry often fail to capture are the largest changes at the ice sheet margins, due to increasing errors in the method for steeper surface slopes occurring at the ice-sheet margins, and large thickness changes on the floating ice shelves at the perimeter, because elevation changes of floating ice express only 1/8 of the full thickness change.

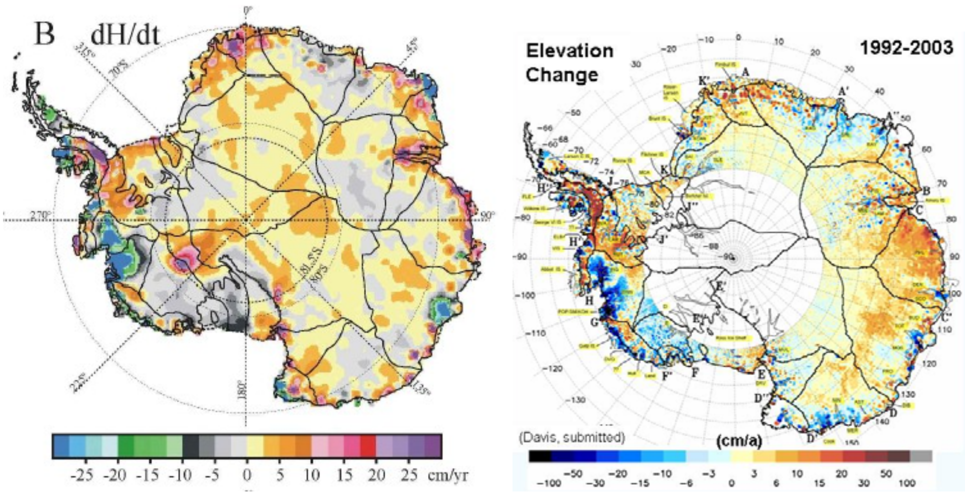


Fig. 3.46. Elevation change Zwally (left) and Davis (right).

Antarctica's ice shelves and ice tongues ('ice tongues' are narrowly-confined ice shelves bounded by fjord walls) are particularly sensitive to climate because both the upper and lower surface of the ice plate are exposed to different systems, each with a potential to cause rapid change: 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), 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 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.

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., 2004]*. 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 ºC 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 a^{-1} (which can be compared to an estimated total Antarctic Peninsula ice volume of 95,200 km^3, equivalent to 242 mm of sea-level) (P*ritchard 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^{-1} (*Morris and Mulvaney, 2004).*

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 shelves) have an area of 95 000 km^2 and a mean net annual accumulation of 143 ± 29 Gt yr^{-1} (*after van Lipzig, et al., 2004).* Changes in the ice margin around the Antarctic Peninsula based on data from 1940 to 2001 have been compiled (*Cook et al. 2005; Ferrigno et al., 2002; Ferrigno et al., 2006; ADD – see ref. 33).* 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 (Fig. 3.47).

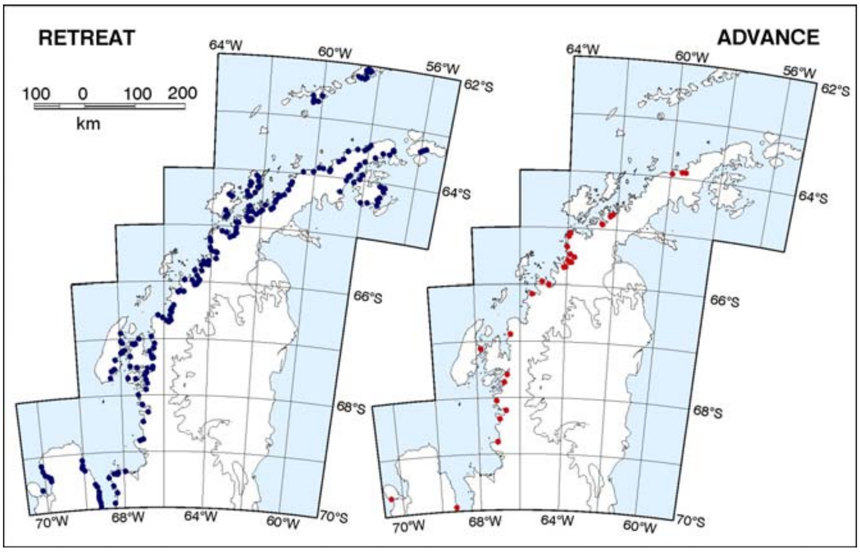


Figure 3.47. Overall change observed in glacier fronts since earliest records. *(From Cook et al., 2005).*

The glaciers that have advanced are not clustered in any pattern, but are evenly scattered down the coast. 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 (Fig. 3.48). However, 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.

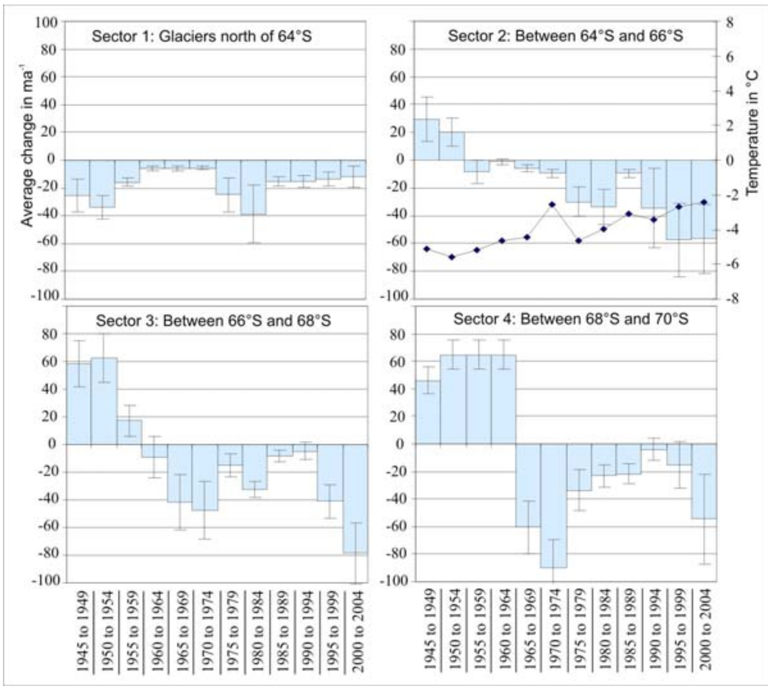


Figure 3.48. Change in Antarctic Peninsula glaciers over time and by latitude Prior to 1945 the limit of glacier retreat was north of 64°S; in 1955 it was in the interval 64-66°S; in 1960 between 66-68°S and in 1965 between 68-70°S. *(From Cook et al., 2005).*

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. It 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%.

Although these studies have assessed changes in loss in terms of glacier flow rates and frontal retreat, the underlying changes in ice volume are unknown. In assessing the 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, but 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*].

The loss of ice shelves (next section) has caused acceleration of the glaciers that fed them [*Rignot, et al., 2004; Rignot, et al., 2005; Rott, et al., 1996; Scambos, et al., 2004*] creating locally high imbalances in ice mass. Glaciers flowing into the now-collapsed sections of the Larsen Ice Shelf accelerated immediately after break-up, to speeds of 2 to 8 times the pre-disintegration flow rate [*Rignot, et al., 2004; Scambos, et al., 2004*] and the glaciers flowing into the Wordie Ice Shelf accelerated following ice shelf loss, and have been thinning and losing mass to the ocean over the last decade [*Rignot et al., 2005]*. Removal of other areas of floating ice could further increase this imbalance, and thus make a significant contribution to sea-level rise. There is still relatively little known about the mass balance of the glaciers on the Antarctic Peninsula and they are generally overlooked in global projections of sea-level rise *[e.g. Meehl, et al., 2007]*.

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; Rott, et al., 2002; Scambos, et al., 2000; Scambos, et al., 2004; Skvarca, 1994; Ward, 1995*]. Wordie Ice Shelf, the northernmost large (>1000 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 (Fig. 3.49). 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 14,000 km^2 of ice. 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).

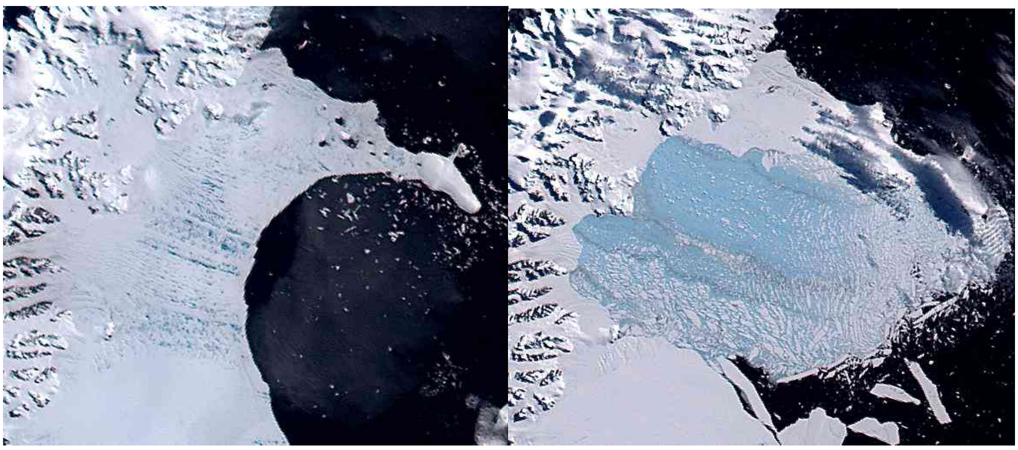


Figure 3.49. Rapid disintegration of Larsen-B ice shelf. Image on left collected on January 31, 2202 and on right collected on March 7, 2002.

Mass loss from the Larsen A and B catchments was estimated to be 24±4.5 Gta^{-1}. A re-assessment of Fleming Glacier feeding the Wordie Ice Shelf (Rignot et al., 2005) showed that it also had accelerated by about 50% during the period 1974-2003, and the region was losing mass at 18±5.4 Gta^{-1}. The ice flux increase may be partially offset by increased precipitation in the western Peninsula (Turner et al., 2005), 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 Gta^{-1}, but a more recent assessment of the region suggests this rate has slowed (28±45 Gta^{-1}, Rignot et al., 2008). This series of papers (and 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 (Table 3.2). 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., 2005; Scambos et al., 2008 in press).

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, C. G. (1995)*
Wordie	1966	1989	2000	700	-1300	35	*Doake & Vaughan (1991)*
Northern George VI	1974	1995	~ 26000	~ 25000	-993	96	*Luchitta & Rosanova (1998)*
Northern Wilkins	1990	1995	~ 17400	~ 16000	-1360	92	*Luchitta & Rosanova (1998)*
	1995	1998			-1098	85	*Scambos et al. (2000)*
Jones	1947	2003	25	0	-25	0	*Fox & Vaughan (2005)*
Prince Gustav	1945	1995	2100	~ 100	-2000	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	2488	320	-2168	13	*Rott et al. (1996)*
Larsen B	1986	2000	11500	6831	-4669	59	*Rott et al. (2002)*
	2000	2002		3631	-3200	32	*Scambos et al. (2004)*
Larsen C	1976	1986	~ 60000	~ 50000	-9200	82	*Skvarca (1994) and Vaughan & Doake (1996)*

Table 3.2. 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., 2003) 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; Kazendehar 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*] (Fig. 3.50). 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 per century [*Morris and Vaughan, 2003*]. All of the ice shelves that exist on the warmer side of the –9° C isotherm position have retreated, and none of the ice shelves on the colder side have been reported to be retreating.

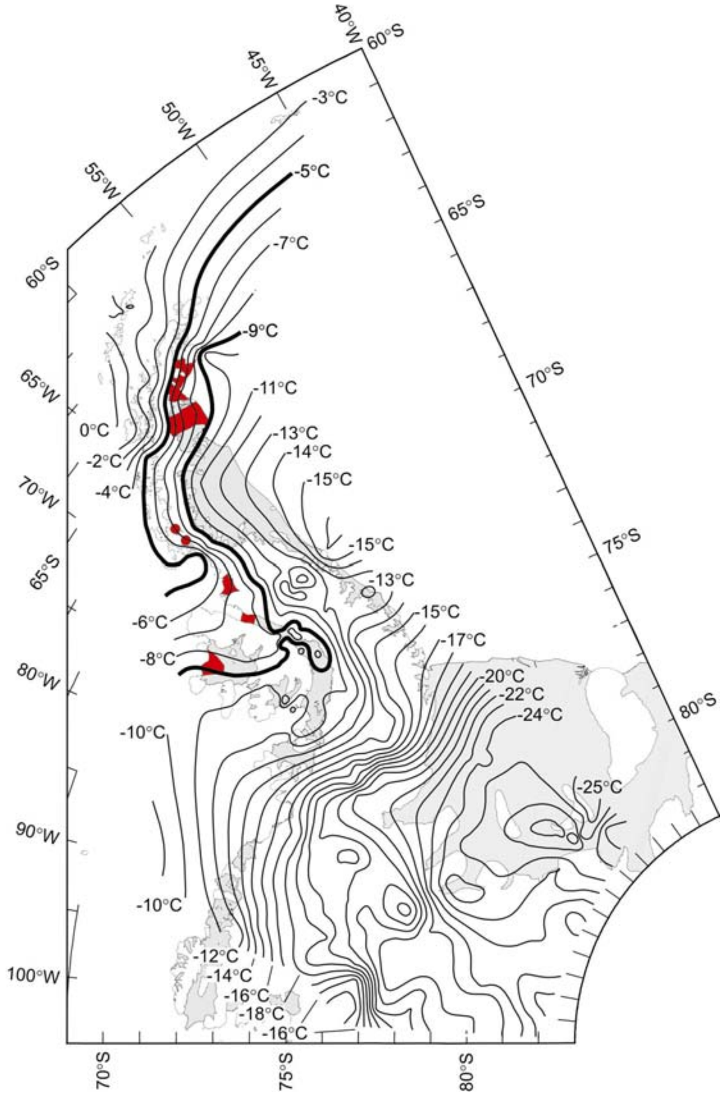


Figure 3.50. Contours of interpolated mean annual temperature. Currently existing ice shelves are shown in grey. Portions of ice shelves that have been lost through climate-driven retreat are shown in red *(From Morris and Vaughan, 2003).*

Sub-Antarctic Islands

Some formerly snow-covered islands are now increasingly snow-free during the summer. (refs?) Rates of glacial retreat have been measured on the Sub-Antarctic islands of South Georgia (54°30'S, 36°30'W) and Heard Island (53°S, 73°30'E). Since the late 1940s, the total area covered by glaciers on Heard Island has reduced by approximately 11%, and several coastal lagoons have been formed as a result (Australian Antarctic Division 2005: http://www.heardisland.aq/). Although there are twelve major glaciers and several minor glaciers, research is focussed on Brown Glacier on the northeastern coast of the island. Results so far have shown that the 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%). The glacier was also found to have decreased in thickness, with a total loss of around 7 Gt of ice each year between 2001 and 2004.

On the island of South Georgia, there are ca.160 glaciers [*Gordon et al., 2008*] and of these, 36 have been mapped and analysed for changes over the past century. The results showed that 2 of the glaciers are currently advancing, 28 are retreating and 6 are stable or show a complex, ambiguous response. Although most glaciers on the northeast coast of the island were at more advanced positions during the late 19th century, since then the smaller glaciers have progressively retreated. Most of the larger tidewater and sea-calving valley and outlet glaciers only began to retreat in the 1980s, although they have shown more variable behaviour. Since then, most glaciers have retreated and a number of small mountain glaciers will soon disappear. The response of these glaciers is related to the direct effects on glacier mass balance of sustained climate warming that began in the 1950s with glacier recession on the windward southwest coast, where precipitation is significantly higher, less widespread. The smaller, lower elevation glaciers have retreated while the higher elevation glaciers have either stabilised or advanced slightly. Individual long-profile geometry may also be a significant influence on the response and sensitivity characteristics of these glaciers [*Gordon et al., 2008*].

West Antarctica

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. However, long before these observations were available the vulnerability and potential significance of retreat in this area was highlighted in a prescient paper by Hughes (1978).

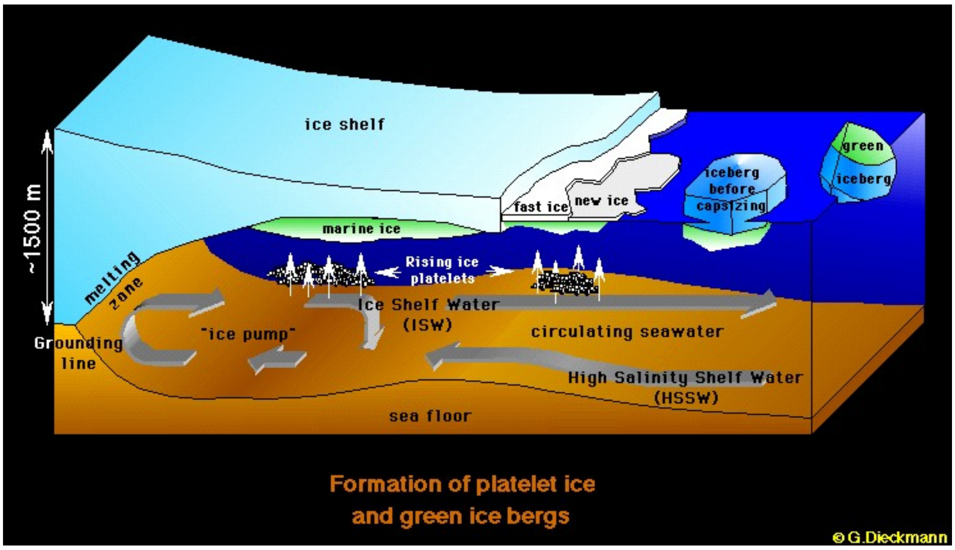


Figure 3.51. Water masses in the Antarctic coastal zone.

Acceleration of flow due to basal melting of its ice shelf and subsequent grounding line retreat of Pine Island Glacier, one of the two largest WAIS outlet glaciers draining into the Amundsen Sea, was first reported by Rignot et al. (1998). This discovery of a 10% increase 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 revealed rates as high as 55 ma^{-1}, implying basal melting rates nearly 10 times the previously calculated value (Shepherd et al., 2004). As basal melt increased, the grounding line retreated, possibly in two stages--during1980s and 1994-96—each leading to a separate increase in speed (Rignot, 1998; Joughin et al., 2003). Most recently Rignot (2008, in press) has observed the grounding line at Pine Island has retreated still further with a simultaneous increase in both speed and rate of speed increase. Pine Island Glacier is now moving at speeds 40% higher than in the 1970s.

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., 2004). Where flow rates have been observed, they too show accelerations, e.g., Smith Glacier has increased flow speed 83% since 1992.

No data is 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 as large as 134 meters, occurred in 28 years with significant shifts to the lateral margins including a major flow shift beginning perhaps as early as 1957 (Bindschadler, 2001).

Calculations of the current rate of mass loss from the Amundsen Sea embayment range from 50 to 137 Gt/a with the largest number accounting for the most recent faster glacier speeds (Lemke, 2007 #2408; 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 and suggestion of 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 Gta^{-1} of ice, while they receive only 177 ± 25 Gta^{-1} of new snowfall (Rignot et al., 2008; see also Thomas et al, 2004). 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 consider that atmospheric temperatures have had any strong role to play in the changes that have occurred there. Similarly, the pattern of thinning, which are very clearly concentrated on the most dynamic parts of the glaciers, indicates that the changes are not the result of anomalous snowfall. The most favoured explanation for the changes is a change (e.g. Payne et al., 2004) is a changes in the conditions in the sea into which this portion of WAIS flows (Fig. 3.51). 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.

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/a over an area tens of kilometers across (Price and Conway refs). 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; Joughin et al. 2005).

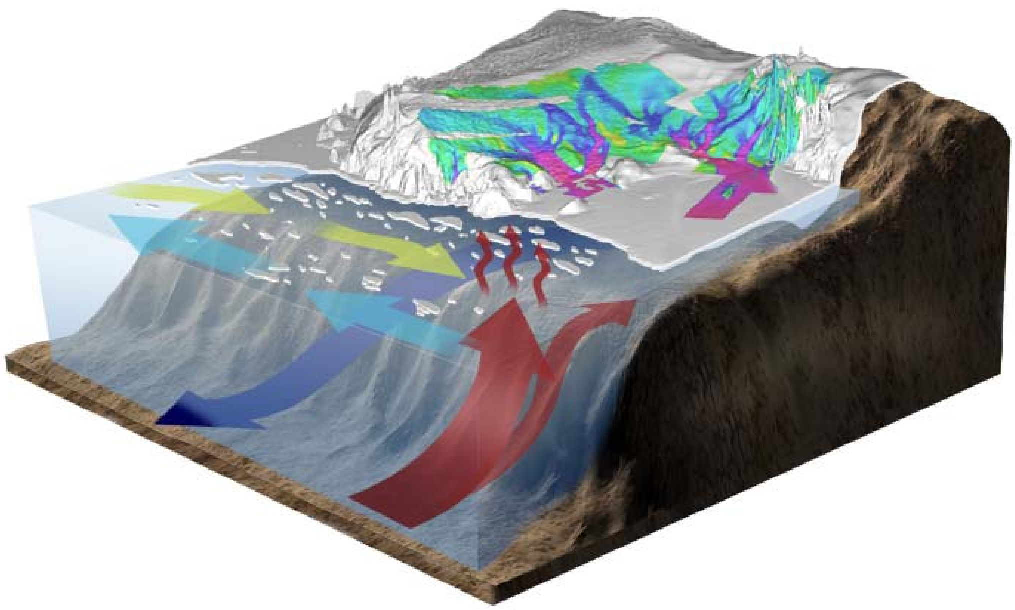


Figure 3.52. Pictoral view of the ocean-ice system of the Ross Sea (illustration from National Geographic-permission probably required).

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/a by Joughin and Tulaczyk (2002) has only been slightly modified to 34 $\pm$ 8 Gt/a recently by Rignot et al. (2008), but the errors overlap, indicating consistency. The ocean-ice system of the Ross Sea is shown in Fig. 3.52.

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 in 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/a (Rignot et al., 2008) varies insignificantly from an earlier calculation of +9 $\pm$ 8 Gt/a 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, 2007).

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 (Davis, 20XX), but ice core data do not show recent accumulation changes are significantly higher than during the past 50 years (Monaghan et al., 2006). A resolution of this apparently conflicting evidence, may be that there is a long-term imbalance in this area, this could possibly be a response to much older 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 et al., 20XX). The absence of significant atmospheric warming inland, distinct from the global trend of warming atmospheric tempertures, may have forestalled an anticipated increase in snowfall associated with the global trend.

The only significant exceptions to this broad-scale quiescence of the EAIS occur on the Cook Ice Shelf and in the mouth of the Totten Glacier where thinning rates in excess of 25 cm/a have been measured (Zwally et al., 2005). 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. It is, however, significant to note, that both these area are the outlets of the ice sheet occupying the two major marine basins lying beneath the EAIS (Wingham et al, 2007).

The mass balance of the East Antarctic ice has been calculated by many research teams with various sensors and methodologies: +22 $\pm$ 23 Gt/a (Rignot and Thomas, 2002); -4 ± 61 (Rignot et al., 2008); 0 ± 56 (Velicogna and Wahr, 2006); +15.1 ± 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 represent changes in low-density snow and firn, or whether there are dynamic in origin and represent a change in ice which has a much higher density.

Other Changes

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 calving patterns continue to follow quasi-repetitive patterns extending back to the 19th 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, which is, in the long-term, close to equilibrium.

Subglacial 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., 2007; Fricker et al., 2007). These observations highlight a subglacial hydrologic system transporting the primary lubricant permitting faster ice flow that is more active than previously thought (Fig. 3.53). However, lacking simultaneous measurements of ice flow to accompany these likely shifts in water mass, their role in ice sheet dynamics remains undetermined.

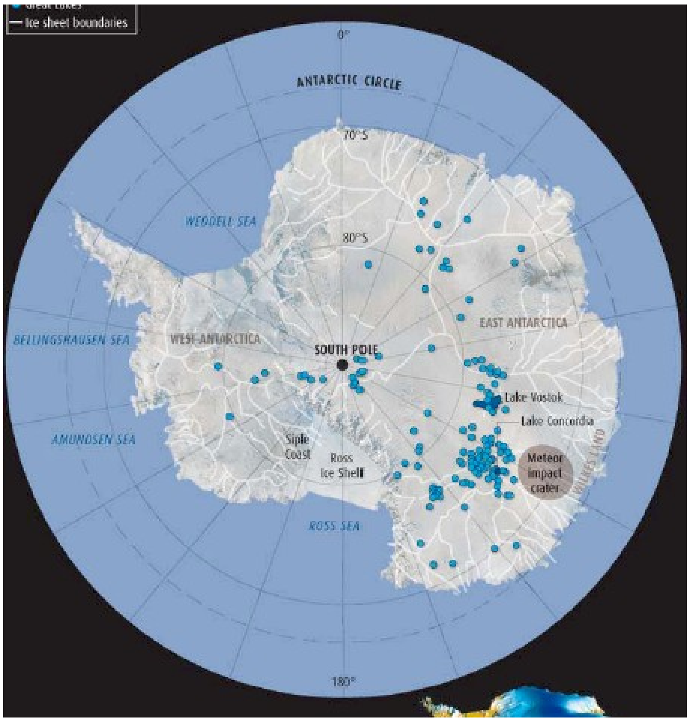


Figure 3.53. Antarctic sub-glacial lakes.

Attribution

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 causes major difficulty in attempts 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 the superposition of multiple responses on multiple time scales to multiple environmental changes.

Further complicating the situation is the lack of a firm understanding of the changes that were occurring in the ice sheet prior to the period of satellite observations (which 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. As an example, it has often been surmised that areas of thickening in the EAIS are a response to recent changes in snowfall rates, a consequence suggested by many GCM scenarios. However, Monaghan et al. (2006) have recently found no evidence in ice cores for changes in accumulation rates over the EAIS during the past 50 years that could explain the thickening and Thompson et al. (20XX) suggest their measurement of a short-lived period of increased accumulation at South Pole during the period of the altimeter study may explain the altimetry results. Thus, a more probable inference is that any sustained thickening of the EAIS areas is more likely a long-lived, and ongoing responses to late-glacial change.

Both the areas of most rapid change, the Antarctic Peninsula ice shelves and the Amundsen Sea sector outlet glaciers, are thought to be linked to greenhouse-forced changes in global circulation, specifically to changes in the Southern Annular Mode circulation around Antarctica, and the likely-related loss of sea ice cover in the Amundsen and Bellingshausen seas (Jacobs and Comiso, 1997). The increased SAM index (indicating stronger westerly winds) are consistently associated with increased greenhouse gas (GHG) loadings and stratospheric ozone depletion in global climate models (Arblaster and Meehl, 2006). Augmentation of the polar vortex flow by stratospheric circulation changes results from a net cooling of the stratosphere during the spring loss of the Antarctic ozone layer (Thompson and Solomon, 2002). To date, the SAM index has increased primarily in the summer and autumn season, but an increasing influence of GHGs, and decreasing intensity of ozone loss, is expected to promote a more general SAM increase in the course of this century (Arblaster and Meehl, 2006). Stronger 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, foen winds across the ice shelves, promoting warmer winter temperatures and longer melt seasons in summer (e.g., Van den Broeke, 2005).

A stronger (more positive) SAM climatology, with increased westerly winds, also drives the Antarctic Circumpolar Current, pushing Circumpolar Deep Water (CDW) against the western Peninsula coast and the northern coast of West Antarctica (Dinniman and Klinck, 2004; Walker et al., 2007). It has been suggested that the northerly Ekman drift of surface water, driven by the Coriolis effect, and sea ice associated with the increased westerlies, could draw sufficient volumes of underlying CDW-derived waters up onto the continental shelf and under ice shelves. CDW water occupies a deep layer, below the continental shelf break (700 to 1100 m), but can be induced to flow up onto the shelf surface and become available for basal shelf melting (Walker et al., 2007) between the upper halocline and the continental shelf, i.e. between about 100 to 500 meters. This water layer is observed to be warming significantly (Gille, 2002) and there is evidence from salinity and other measurements over the past 40 years that increased melting is resulting from interaction of warmer CDW with northern West Antarctic ice shelves (Jacobs et al., 2002).

3.8.3 **Conclusions**

The Antarctic ice sheet is not behaving in a uniform manner – this is in general, not unsurprising 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 subject 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, however slight increase 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 East Antarctic ice sheet, the current state of the ocean may be having influence in two regions of rapid thinning. This is certainly the case in the Amundsen Sea sector of West Antarctica, one of two very rapidly changing regions. 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. The other sectors of the West Antarctic 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 radically 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 lies as the only barrier to the east-west atmospheric and oceanic circulation at these latitudes. And here, high rates of snow accumulation and melting drive a more vigorous glaciological regime. Recent observations have captured the sudden, and very likely rare, succession of sudden 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 of the near-future impact of much larger ice reservoirs on global sea level.

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 portions of East Antarctica, is an expected consequence of climate change, but it is not yet possible make a satisfactory attribution of the observed changes here to any observed climate change.

3.9 Antarctic Sea Levels

3.9.1 Long Term Sea Level Change

The first three assessment reports of the Intergovernmental Panel on Climate Change (IPCC) arrived at similar conclusions with regard to global sea level change during the 20th 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/year. 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 20th century rise in global sea level was closer to 2 than 1 mm/year, with values around 1.7 mm/year 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 the 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 20th century sea level rise are believed to have originated from ocean thermal expansion and melting of glaciers and ice caps. Antarctica’s contribution appears to have been of the order of 0.1-0.2 mm/year 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/year or more (e.g. Holgate and Woodworth, 2004; Beckley et al., 2007). This is at a higher rate than one might expect from IPCC projections, which has led to concern over possibly large ice sheet contributions during the 21st century, especially due to their dynamic instabilities (Rahmstorf et al., 2007; Church et al., 2008). However, decadal variability in the rate of global sea level change makes it difficult to be confident that the apparent higher rates of the 1990s will be sustained, since high decadal rates have been observed at other times during the past century (Chambers et al., 2002; Church and White, 2006; Holgate, 2007). The Fourth Assessment determined that global sea level might rise between 18 and 59 cm during the 21st century, taking into account the full range of emission scenarios and climate models, but not including a contribution from the dynamic instability of Greenland (Church et al., 2008). However, sea levels might be expected to rise around Antarctica itself at rates lower than in other parts of the world, owing to the important role of the ocean adjustment to climate forcings (Figure 10.31 of Meehl et al., 2007) and, if the rise stems from Antarctic melt, then also due to the elastic response of the solid earth (cf. Mitrovica et al., 2001).

3.9.2 Sea Level Data Sets

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 which satisfy the quality criteria required by the Permanent Service for Mean Sea Level (Woodworth and Player, 2003) and are long enough to be of interest for long term change studies. The outstanding record is that from Vernadsky in the Argentine Islands on the western side of the Antarctic Peninsula (Figure 3.54). It is 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 at around the time of 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. Figure 3.55 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 20th 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).

Figure 3.54. Main Antarctic tide gauges described in the text.

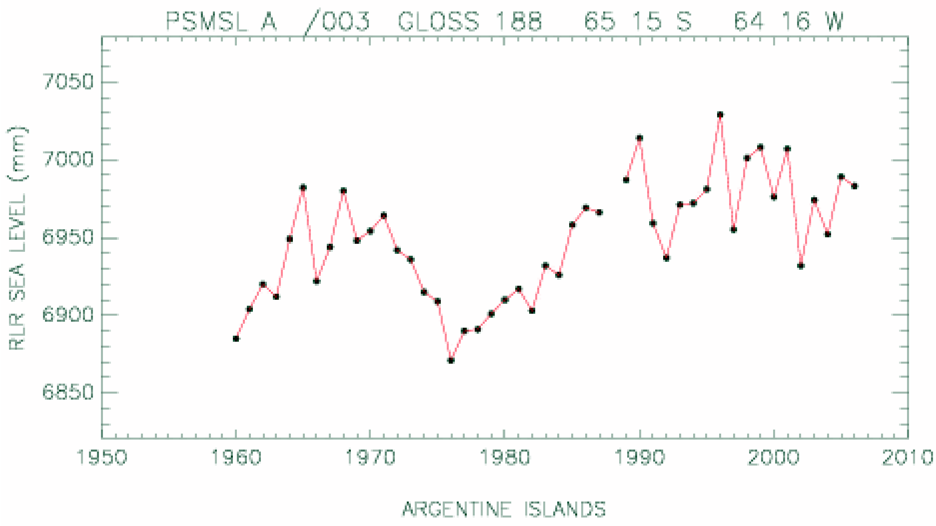


Figure 3.55. PSMSL Revised Local Reference (RLR) annual mean sea level time series for Vernadsky/Faraday (called Argentine Islands in the PSMSL data set)

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 which 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 has 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.

3.9.3 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) which 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/bprs.

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 Southern Annular Mode (SAM; the leading mode of extratropical Southern Hemisphere climate variability, and largely responsible for changes in the circumpolar westerly winds) 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 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 become available in 2008. 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. 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).

In summary, one sees that 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 GLOSS, 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 BPR 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.

3.10 **The Southern Ocean**

3.10.1 **The Circumpolar Southern Ocean Regime**

The Southern Ocean plays a critical role in driving, modifying, and regulating global change (Fig. 3.56). It is the only ocean that circles the globe without being blocked by land and is home to the largest of the world's ocean currents: the Antarctic Circumpolar Current (ACC). The Southern Ocean controls climate in a number of ways:

- The flow of the ACC from west to east around Antarctica connects the Pacific, Indian and Atlantic ocean basins. The resulting global circulation redistributes heat, salt, freshwater and other climatically and ecologically important properties. It has a global impact on patterns of temperature, rainfall and ecosystems functioning.

- The Southern Ocean is a key region in the oceanic meridional overturning circulation (MOC; often referred to as the thermohaline conveyor belt), which transports heat and salt around the world. Within the Southern Ocean, the products of deep convection in the North Atlantic are upwelled and mixed upwards into shallower layers, where they can be converted into shallow and deep return flows that complete the overturning circulation. This upwelling brings CO_2 and nutrient-rich waters to the surface, acting as a source for atmospheric CO_2 and promoting biological production.

- The lower limb of the MOC comprises the cold, dense Antarctic Bottom Water (AABW) that forms in the Southern Ocean. Close to the coast, the cooling of the ocean and the formation of sea-ice during winter increases the density of the water, which sinks from the sea surface, spills off the continental shelf and travels northwards hugging the sea floor beneath other water masses, travelling as far as the North Atlantic and North Pacific. This cold water also absorbs large amounts of atmospheric gases (oxygen and carbon dioxide), which enables it both to aerate the bed of the global ocean and to act as a sink for CO_2.

- The upper limb of the MOC is sourced toward the northern flank of the ACC. Here, the water that is upwelled within the ACC is converted into mode waters and intermediate waters that permeate much of the global ocean basin south of the equator with nutrient rich water. Both AABW and the mode waters and intermediate waters show variability in properties on a range of timescales (seasonal to decadal and longer), reflection global and regional climate variability in their source regions.

- Closer to the continent, ocean processes are strongly controlled by sea ice processes, the formation of which is the largest single seasonal phenomenon on

Earth. The size of Antarctica doubles each year with the freezing of the sea around the continent and the resulting sea ice has a profound effect on climate. Because of its high albedo, it reflects the sun's heat back into space, cooling the planet. However, the sea ice also limits heat loss from the ocean to the atmosphere. Its yearly formation injects salt into the ocean, making the water denser and causing it to sink as part of the deep circulation. The sea ice is also home to large algal populations, as well as sheltering the larvae of plankton such as krill.

Because of its upwelling nutrients, the Southern Ocean is the world's most biologically productive ocean. It is not as productive as it could be, however, because its productivity is limited by the low availability of micronutrients such as iron, except in regions such as the isolated islands that are scattered within the ACC. 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.

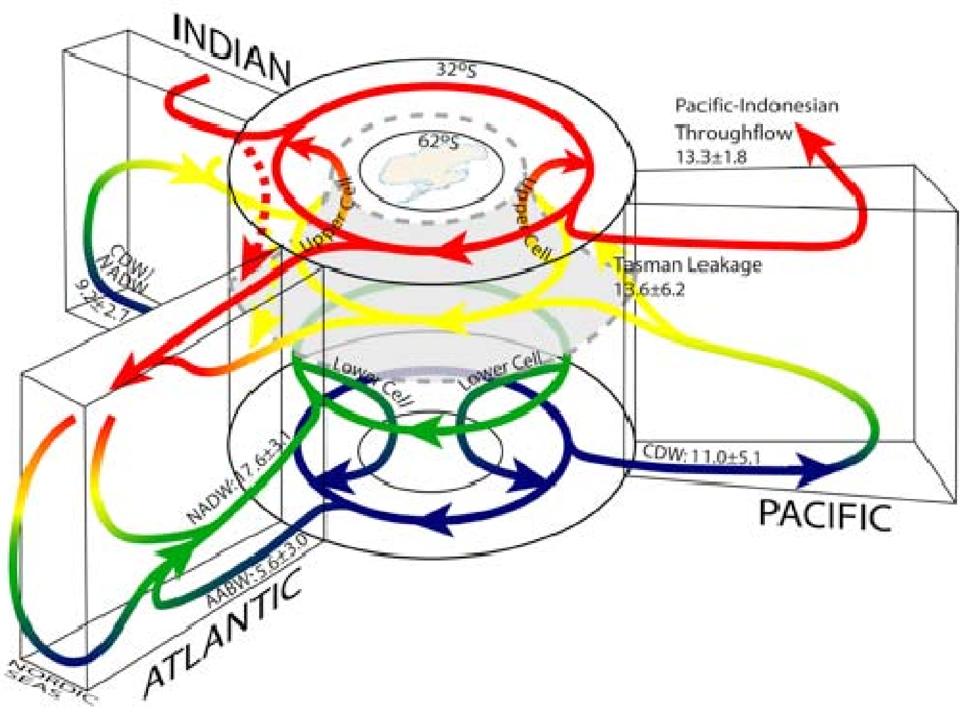


Fig. 3.56. Schematic of the Southern Ocean circulation showing the links to the World Ocean (Lumpkin and Speer, 2007). The two primary overturning cells are the Upper Cell (red and yellow), and the Lower Cell (blue, green, yellow). The bottom water of the Lower Cell upwells and joins with the southward flowing deep water to form the upper branch. This combination illustrates the global link between Southern Ocean convection and bottom water formation and convective processes in the Northern Hemisphere.

The Water Masses and circulation of the Southern Ocean

The major water masses of the Southern Ocean can be defined in various ways, but often too much emphasis is placed on these definitions, as though there were real barriers between the water masses. The overall density gradient from north to south is controlled by temperature, though salinity exerts a stronger control in the coldest waters. Of greatest importance is the effect of salinity in the North Atlantic Deep Water (Fig. 3.57, near 2000m depth), itself due ultimately to evaporation over the Mediterranean Sea. This water mass is brought in to the Southern Ocean to replace the dense water escaping along the bottom and brings with it anomalous heat. Above the deep water, lower salinities are maintained by ice melt and precipitation, and spread north in the lower salinity Antarctic Intermediate Water.

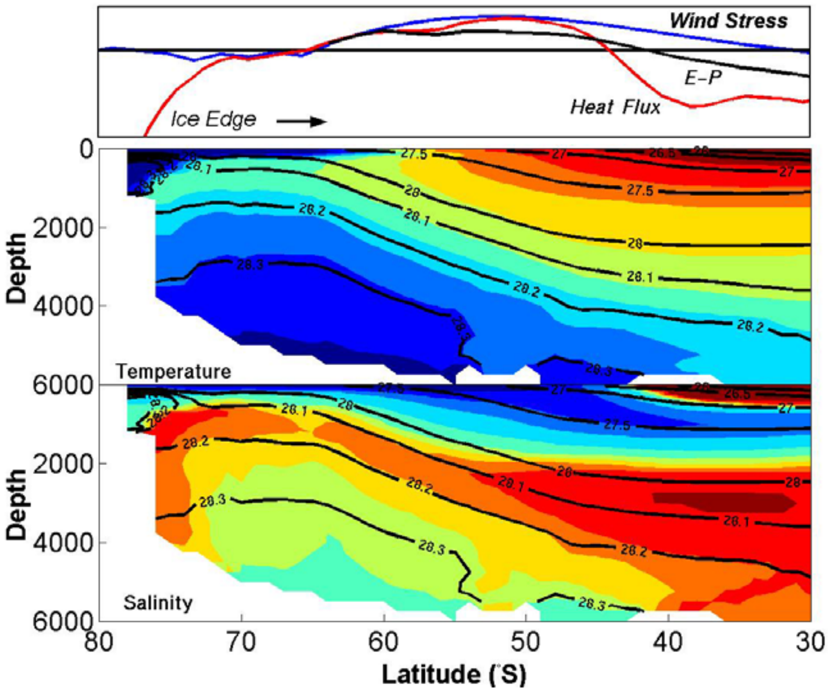


Fig. 3.57. Zonally averaged temperature, salinity, and air-sea fluxes (showing heat, freshwater, and momentum gain by the ocean from about 40S-65S) in the Southern Ocean (from Wacongne et al., 2007). Neutral density contours (black) show isopycnal spreading except at boundaries and convection regions. The seasonal ice-edge extends to roughly 65 S. An overturning cell is generated by dense water formation on the continental margins. Brine rejection results in elevated salinities on the slope, especially due to Ross Sea sources.

On top of this meridional cell are the regional gyres, principally the Weddell Gyre and Ross Gyre, whose southern boundary is the Slope Current, and whose outer limits reach to the ACC. These gyres constitute the detailed pathway for transport to, from, and along the continental margin. They are not strongly present in traditional hydrographic estimates but are clearly revealed in models of the Southern Ocean (Fig. 3.56).

Climate change affects each of the components differently, and with different time scales. The overturning cell is slowest to react to changes in the surface fluxes, with time scales of a year or greater, while the regional gyres can respond more quickly – for example days in the case of spin-up or spin-down of the gyres by anomalous wind fields. The other important time scale is the time for climate change signals like fresh anomalies to advect around each of the principal circulation elements, with the global overturning having a time scale of decades to centuries, and the gyres years to decades. Thus the imprint of climate change on the ocean, and the feedback between the components of the climate system as the anomalies reinforce or cancel one another, are regulated by the large-scale circumpolar circulation.

The Antarctic Circumpolar Current

In the Southern Ocean a meridional gradient in the air-sea buoyancy flux acts together with the wind stress to create the ACC. This current flows basically in thermal wind balance with lateral density gradient set by the freezing temperatures near Antarctica and the warm subtropical gyres. Two principal fronts typically carry a large fraction of the transport of the ACC, the Subantarctic Front (SAF) and the Polar Front (PF), (Fig. 3.58; and Cunningham *et al*., 2003), but other fronts can have comparable transports in 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 is a new paradigm for considering the response of the ACC to climate change.

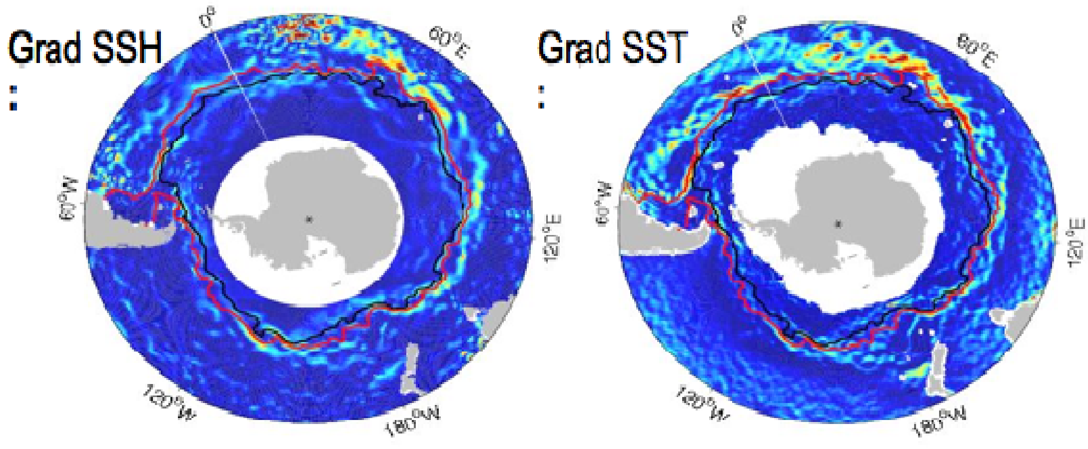


Fig 3.58. Major fronts in the ACC: SAF (red), and PF (black). Background mean gradients of sea-surface height (left) and sea-surface temperature (right) from satellite observations show the great complexity of the ACC system.

What controls the strength of the ACC? In general terms, the north-south density gradient sets the overall baroclinic (thermal wind) transport of the ACC, together with the wind, which drives the barotropic (depth-integrated) transport. Analysis of drifting buoy trajectories in the Southern Ocean by Patterson (1985) and Geosat altimetry data by Chelton, *et al*., (1990), Morrow, *et al*., (1992) and Stewart, *et al*., (1996) shows the distribution of mesoscale variability or eddies due to hydrodynamic instability of the ACC. The potential energy stored in the fronts is released through baroclinic instability and the resulting eddies are thought to play a fundamental role in the dynamical and thermodynamical balance of the ACC - see, e.g., Marshall, *et al*., 1993, 2002; Johnson and Bryden (1989), Rintoul, *et al*., (2001). These studies generally point to a dependence of the strength of the ACC on the wind stress, but as the wind becomes stronger this dependence can be less than linear, reflecting compensation by eddy transport (Hallberg and Gnanadesikan, 2006). Observationally, this means that our views of zonal flow and transport, Ekman transports, and upwelling conditioned by a zonally averaged perspective have to be modified. Changing winds, brought about by changes in the Southern Annular Mode for example, may rearrange the mean and eddy field in unknown ways. Moreover, eddies are not confined to the ACC itself, but exist throughout the Antarctic Zone: the role of eddies on the transport of heat anomalies in the ice zone or freshwater anomalies away from the ice zone is not well known; their control on sea-ice volume and their effect on glacial ice melt is a new direction for Southern Ocean climate research.

Australian Sector

Knowledge of the circulation in the Australian sector of the Southern Ocean has increased significantly in the last decade. Repeat hydrographic sections (Fig. 3.59), 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).

The ACC belt in the Australian sector has warmed in recent decades, as found elsewhere in the Southern Ocean (Gille, 2002; Levitus et al., 2000, 2005; Willis et al., 2004; Böning et al., 2008). The changes are consistent with a southward shift of the ACC. Some climate models suggest the ACC will shift south in response to a southward shift of the westerly winds driven by enhanced greenhouse forcing (Fyfe and Saenko, 2006; Bi et al., 2002). 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; Hogg et al., 2008)

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).

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). 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 140E. 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 per decade, which can be compared to a mean freshening rate of 12 ppm per 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 Adelie Land coast (144E) 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 account 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 Adelie 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 Adelie Land and Ross Sea Bottom Water.

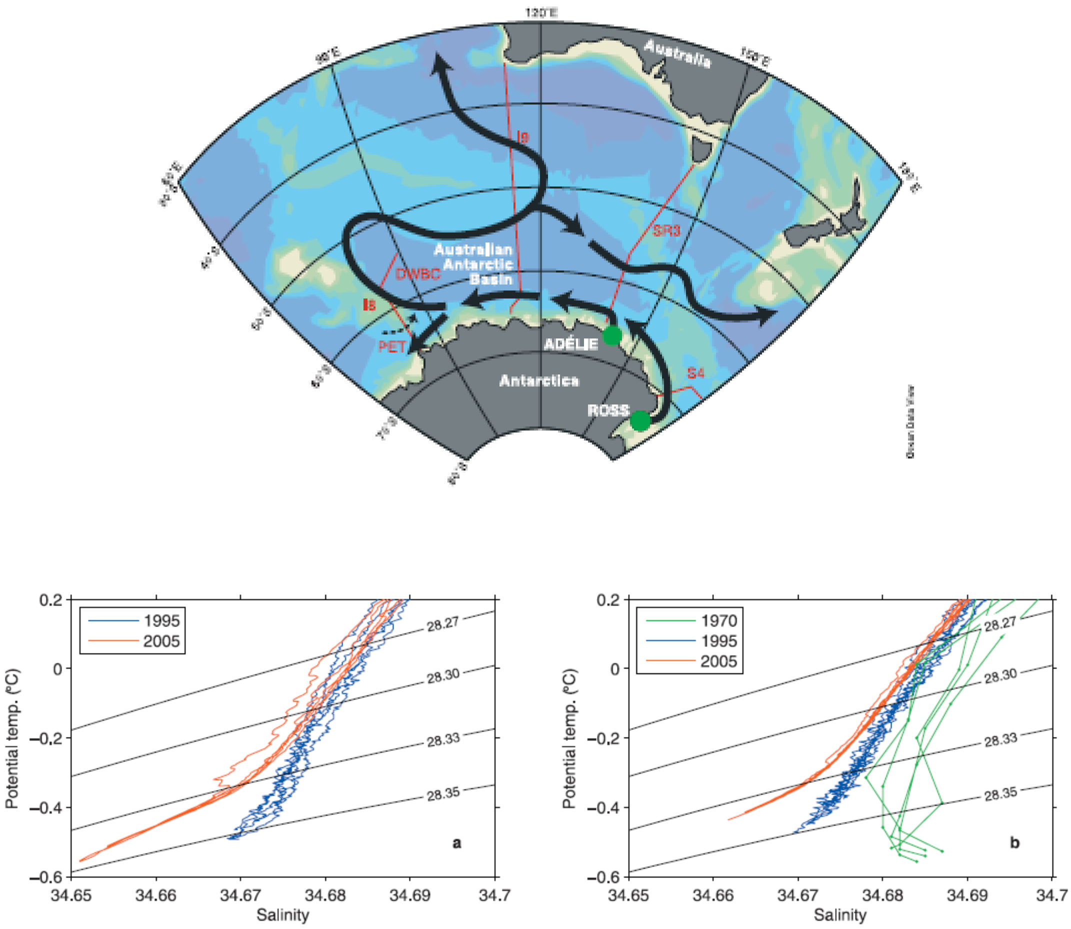


Figure 3.59. a) Map indicating location of repeat sections along which changes in bottom water properties have been assessed. The Australian Antarctic Basin is supplied by two sources of Antarctic Bottom Water: the Ross Sea and Adelie Land. b) Changes in the deep potential temperature – salinity curves along 115°E, over the continental rise (61-63.3°S, left) and further offshore (56.5 - 61°S, right). From Rintoul (2007).

The Amundsen Bellingshausen Sea

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) which can be linked to changes in the Bellingshausen Sea like sea ice retreat (Jacobs & Comiso 1993), increased ocean surface summer temperatures of more than 1 °C, enhanced upper-layer salinification (Meredith & King 2005), the disintegration of smaller ice shelves (Doake & Vaughan 1991), and accelerated retreat of glaciers (Cook et al. 2005). The changes can be related to atmospheric variability such as the Antarctic Circumpolar Wave (White & Peterson 1996) or the Southern Annular Mode (Hall & 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 (BS) 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 (AS) yet needs to be conducted. Connected via the westward flowing, in this part, weak coastal current, alterations in BS also influence the AS (100 ^{o}W – 150 ^{o}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). The southernmost position of the Antarctic Circumpolar Current's (ACC) southern front (Orsi et al. 1995) together with a relative narrow continental shelf crisscrossed by numerous channels (Fig. 3.60) allow Upper Circumpolar Deep Water (UCDW) to reach the ice shelf edges in Amundsen and Bellingshausen Seas with temperatures near 1 ^{o}C (Fig. 3.61). This ocean heat fuels melting at deep ice shelf bases of up to tens of meters per year with a linear relation of excess melting of 1m/yr per 0.1 ^{o}C ocean warming (Rignot & Jacobs 2002). Numerical modeling of ice-ocean interaction (Payne et al. 2007) concludes that the observed thinning of Pine Island Glacier at a rate of 3.9 $\pm$ 0.5 m/yr between 1992 and 2001 (Shephard et al. 2002) would correspond to a ~ 0.25 ^{o}C warming of the UCDW underneath Pine Island Glacier. Such warming has not been observed on the AS 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 ^{o}C (Jacobs et al. 2002). In addition, numerical studies with a circumpolar coupled ice shelf - ocean model show that Weddell Sea anomalies trigger on decadal time scales the intensity of the eastern recirculation of the Ross Gyre and thus influence the propagation of temperature pulses of 0.32 ^{o}C from the deep ocean onto the AS continental shelf (Hellmer et al. 2008). This temperature variability is correlated with changing freshwater fluxes due to basal melting at the fringe of the West Antarctic Ice Sheet. Upper Southern Ocean temperatures including the UCDW increased since the 1950's (Gille 2002), but the few oceanographic shnapshots (e.g., Hofmann & Klinck 1998, Walker et al. 2007) from the ABS continental margin are insufficient to identify the time scales and strength of variability or any trends. Therefore, participants of an international workshop entitled "West Antarctic Links to Sea-Level Estimation (WALSE)" formulated immediate action, among others, to (Vaughan et al. 2007):

"Significantly better oceanographic data are required from the Amundsen Sea continental shelf to understand the nature of ice-ocean interactions there and the potential for future changes. Delays in the collection of baseline data will greatly affect our ability to predict future changes in the ice sheet and future sea level rise."

"Increased effort is needed to distinguish variablity and trends in climatic variables that will influence the evolution of the ice sheet. To ensure predictability, the root causes of atmospheric and oceanic forcing on the ice sheet from drivers must be understood"

The statements above are evidence for the present and future global importance of the Amundsen and Bellingshausen Seas and the urgent need for continues field measurements in combination with high-resolution numerical modeling in the Southeast Pacific sector of the Southern Ocean.

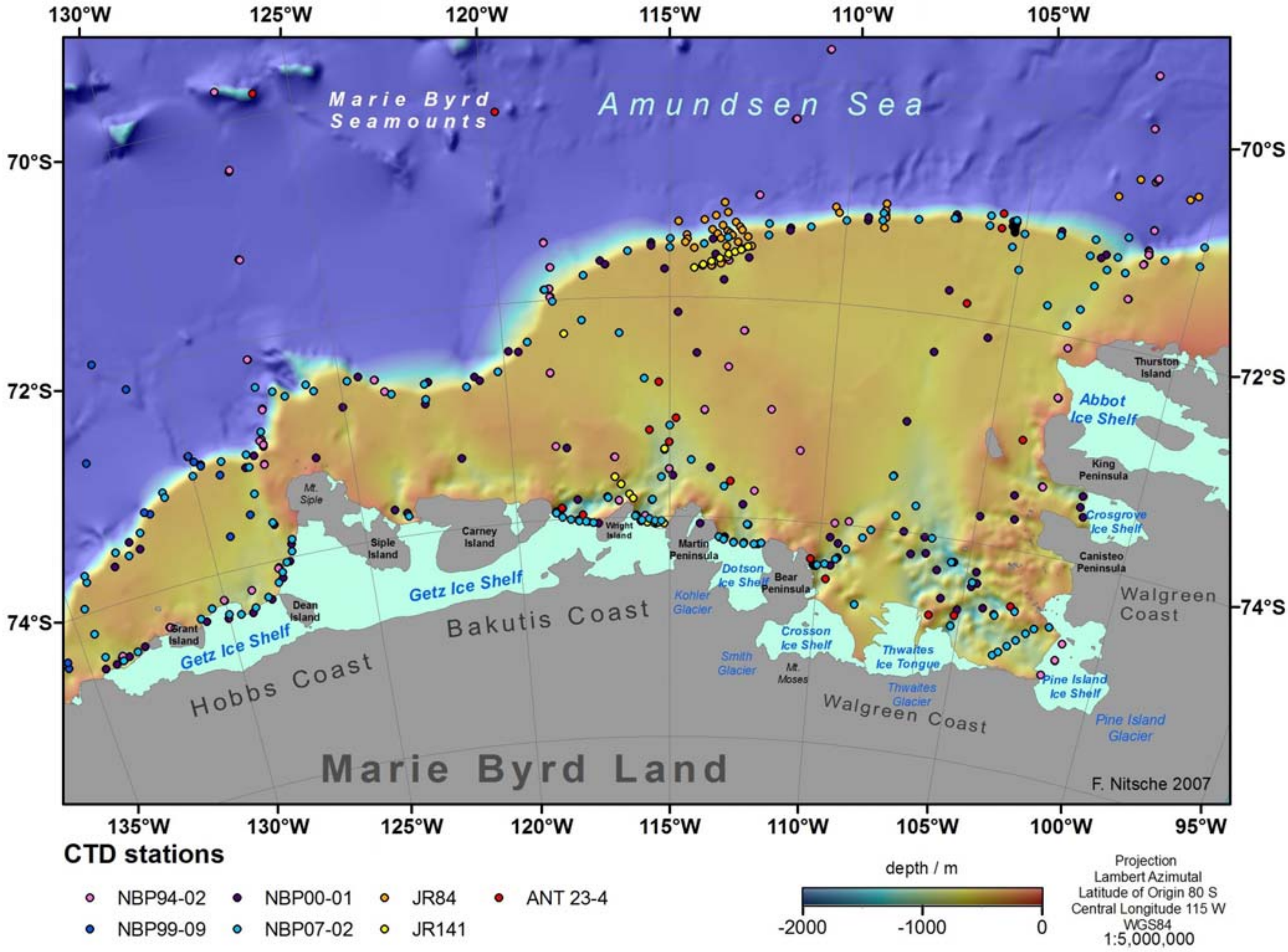


Figure 3.60. Bathymetric chart of the Amundsen Sea continental shelf and adjacent deep ocean spotted with the distribution of hydrographic stations of different cruises (color coded). Fringing ice shelves and glaciers draining the West Antarctic Ice Sheet in light blue (Nitsche et al. 2007).

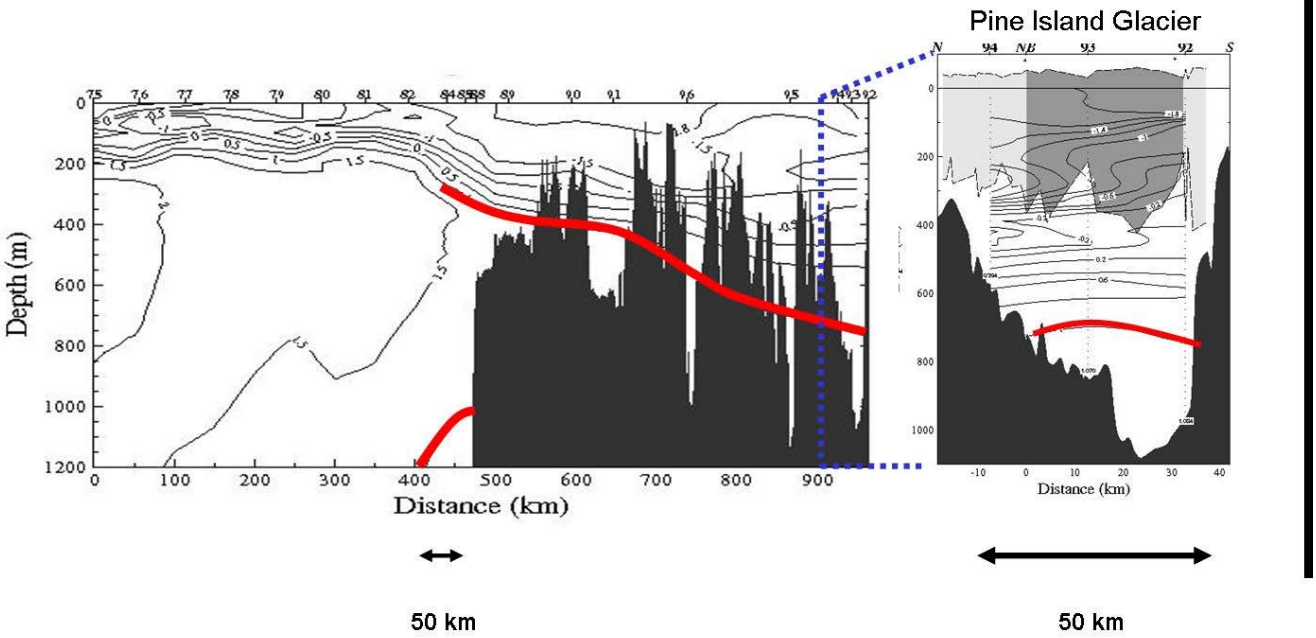


Figure 3.61. Potential temperature of the upper 1200 m along 100°W from the open ocean (left) to Pine Island Bay (righty) measured during NBP9402 (see Fig.3.60 for station locations). The sea floor depth was extracted from the ship 3.5 kHz echosounder data. Right figure depicts the temperature field in front of Pine Island Glacier with its draft shaded in gray. The 1°C isotherm on continental shelf and slope is marked in red (modified from Hellmer et al. 1998).

Variability and change in Ross Sea shelf waters

During recent decades many reports have described seasonal to decadal changes in the atmosphere, sea ice, ice shelves, icebergs, water properties, circulation and ecosystems of the Ross Sea. Here we extend the record of summer shelf water thermohaline characteristics, noting possible causes and effects of observed variability. 'Shelf waters' include the several low-temperature, ice-modified, high- and low-salinity water masses found below the surface layers on the Antarctic continental shelf (Whitworth et al 1998). Integrating a variety of forcing and mixing processes, shelf water salinity has changed substantially over the IGY to IPY time frame.

Jacobs (1985) reported seven 1963-84 summer salinity profiles from a small region of High Salinity Shelf Water (HSSW) near Ross Island. Individual records showed gradual salinity increases from 200 - 900m and interannual water column shifts of ~0.1, about 10 times the measurement accuracy and half the annual cycle at depth in McMurdo Sound (Tressler & Ommundsen 1963). Atmospheric forcing, sea ice production, HSSW residence time, ice shelf melting and intrusion of Modified Circumpolar Deep Water onto the shelf were considered as possible agents of change. Hellmer & Jacobs (1994) noted that salinity decreases could also result from fresher upstream source waters.

In a more detailed 1998 update, we tried to link observed HSSW changes to multiyear variability in regional sea ice extent, winds and local air temperature, but noted the need for longer time series. In addition, the 15-km radius reoccupation site lies above a frequent HSSW dome, which could add spatial variability, and the only database is still comprised of sporadic summer measurements. Nonetheless, the deep HSSW trend in that area has closely tracked changes at depth along the Ross Ice Shelf and near 500m throughout the western Ross Sea (Jacobs & Giulivi 1998; Smethie & Jacobs 2005).

Record low salinities at the site in Feb 2000 led to analyses that implicated changes in ice-ocean interactions upstream in the Amundsen and Bellingshausen Seas (Jacobs et al 2002). Assmann & Timmermann (2005) successfully modeled the 2000 and earlier averaged HSSW salinity profiles, but inferred that the freshening resulted from a Bellingshausen thermal anomaly. That periodic signal upwelled in the Amundsen, reducing brine drainage near the sea ice edge and inducing a subsurface salinity decrease that was advected into the Ross Sea.

Here only the 08 Feb 2007 profile in Fig 3.62 falls within the study circle, but we have added two regional profiles and the salinity ranges for summer 1960-61 casts in McMurdo Sound. Except for the latter, the additions are relatively low in salinity, in accord with modeled values prior to the mid-1960s, and with a 50-year trend of ~-0.03/decade. Temperature averages ≥300m range from ~-1.88 to -1.94, generally consistent with a primary origin near the sea surface in winter. While temperature variability will impact ice shelf melting, temperature change in shelf water is damped by freezing and melting processes (MacAyeal 1984).

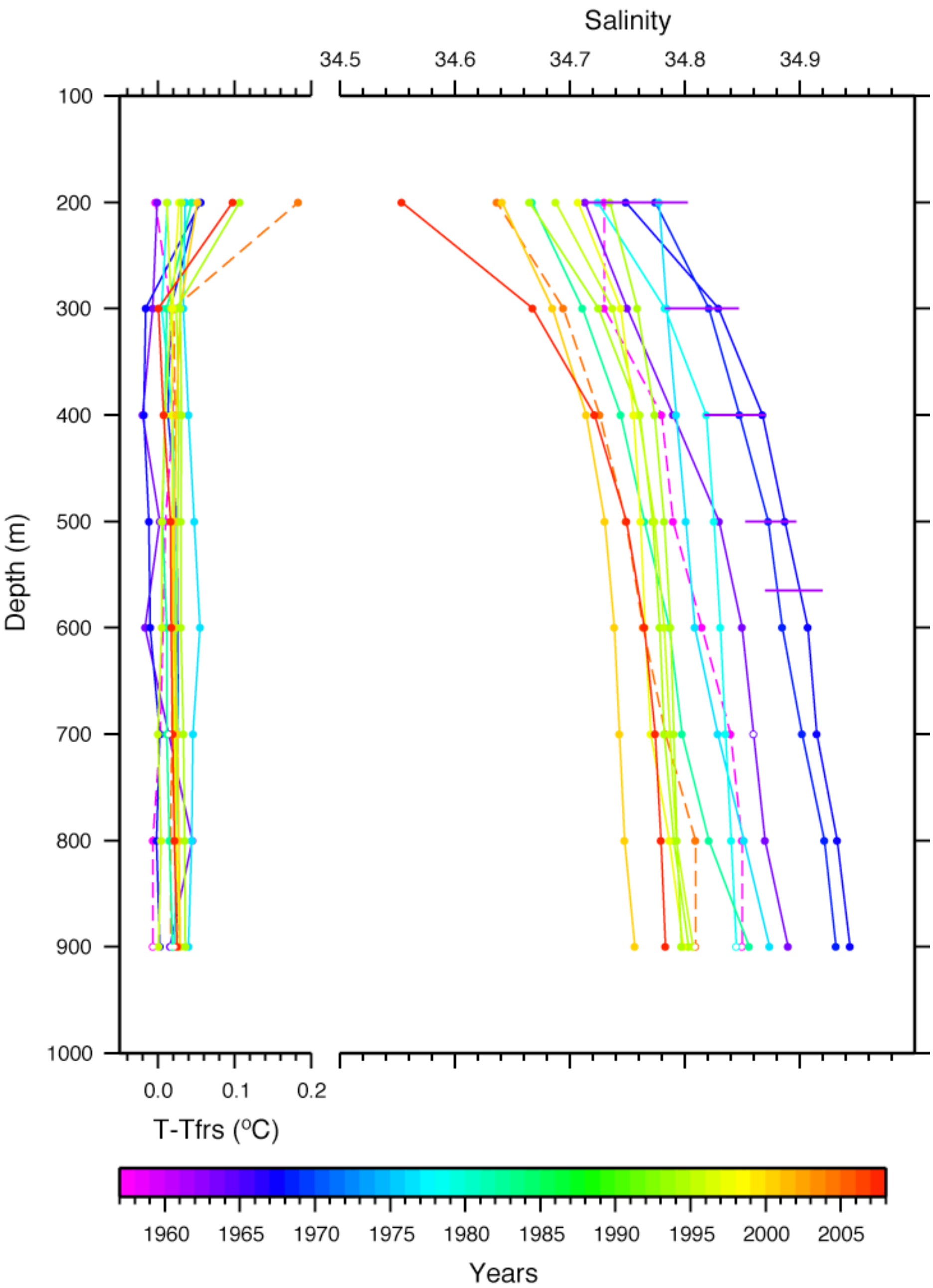


Figure 3.62. Summer temperature and salinity profiles over a 50-year period, most within 15 km of 168.333E, 77.167S, near Ross Island. Plotted values ≥200 m are 100 m averages/interpolations of CTD/bottle data, edited or extrapolated where shown by open circles. Four of the solid lines are averages of adjacent profiles, and the dashed lines are 1957 and 2004 profiles ~30 km east and 60 km west of the centerpoint. Horizontal lines depict the salinity ranges at six 23 Dec–09 Feb 1960-61 bottle casts in southern McMurdo Sound, evaluated by Jacobs & Giulivi (1999). Temperatures are referenced to the surface freezing point, ~-1.91°C at a salinity of 34.8.

Both modelers and observers have noted the possibility of aliasing in this shelf water record due to undersampling of a variable inflow. Interannual salinity variability is high, but the overall trend is statistically significant and qualitatively consistent with freshening over a much wider area (Jacobs 2006). HSSW near Ross Island thus serves as an index site to monitor change in Ross Sea shelf waters. The salinity declines appear to derive mainly from increasing continental ice meltwater, and will subsequently change the properties if not the volume of deep and bottom waters. Over regional areas the lower salinity has raised sea level via the halosteric component of seawater density. There remains a need to more continuously monitor shelf water properties at appropriate sites.

The Weddell Sea Sector

The Weddell Sea hosts a subpolar gyre, the Weddell Gyre, 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.

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 ~60S. The eastern boundary is less well defined but is generally agreed to extend as far east as ~30E (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 is difficult to measure. Integrating the wind field in a Sverdrup calculation revealed 76 Sv (Gordon et al. 1981) while the first attempts to reference the geostrophic shear to current meters yielded 97 Sv (Carmack and Foster 1975). For a section across the Weddell Gyre from the tip of the Antarctic Peninsula to Kapp Norwegia, 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 Greenwich meridian, larger transports of order 60 Sv were obtained from referencing shear to shipboard ADCP (Schroder 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 Norgwegia section. However a recent high resolution section at the Antarctic Peninsula also yields ~46 Sv (Thompson and Heywood 2007, subm. DSR I), 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 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 highly saline shelf water (ref. Nicholls review paper?). The WSBW leaves descends the continental slope and entrains ambient water as it goes (Baines & 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 +/- 4 Sv (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.

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 indicate a number of decadal-scale changes in water mass properties. The WDW warmed by some 0.04C 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, 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 0.2C accompanied by a small increase in salinity, whereas there was no discernible change in the northern limb of the gyre (Heywood & King 2002). There has been debate 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, 2006).

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.

Because much of the Weddell Sea is covered in sea ice for much of the year, there is a lack of observations on the continental shelf and slope, especially in winter. This is a priority area during IPY and moored arrays are being deployed together with hydrographic sections. Measurements beneath the sea ice in the Weddell Sea are being obtained for the first time by acoustically tracked floats and by instruments carried by marine mammals such as elephant seals. These promise to lead to new insights into the physical processes in the Weddell Sea.

Small-scale processes in the Southern Ocean

Physical processes occurring on length scales smaller than the first-baroclinic Rossby radius of deformation (5-20 km south of 40ºS, see Chelton *et al.* [1998]) play an exceptionally important role in shaping and driving the circulation of the Southern Ocean.

Thus, formation of the Antarctic Bottom Water (AABW) filling the deepest layers of much of the global ocean abyss is crucially dependent on the convective 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) (Fig. 3.63). The properties of Antarctic shelf waters are modified further by turbulent mixing associated with the globally significant dissipation of barotropic 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 underlie the conversion of barotropic tidal energy into turbulence, amongst which 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) are the most prominent. The importance of these processes is accentuated by the proximity of the Antarctic continental shelves to the critical latitude 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 sub-mesoscale processes, such as double-diffusive lateral interleaving across the shelf break zone (Foster and Carmack, 1976; Foster, 1987), viscous drainage in the bottom Ekman layer, 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 hydraulically controlled flows over small sills in confined passages (Bryden and Nurser, 2002; see also Fig. 3.64) and mid-ocean ridge-flank canyons (Thurnherr and Speer, 2003), although the breaking of internal tides (Simmons *et al.*, 2004) and internal lee waves generated by the interaction of subinertial flows with ocean-floor topography (Naveira Garabato *et al.*, 2004) must contribute significantly too.

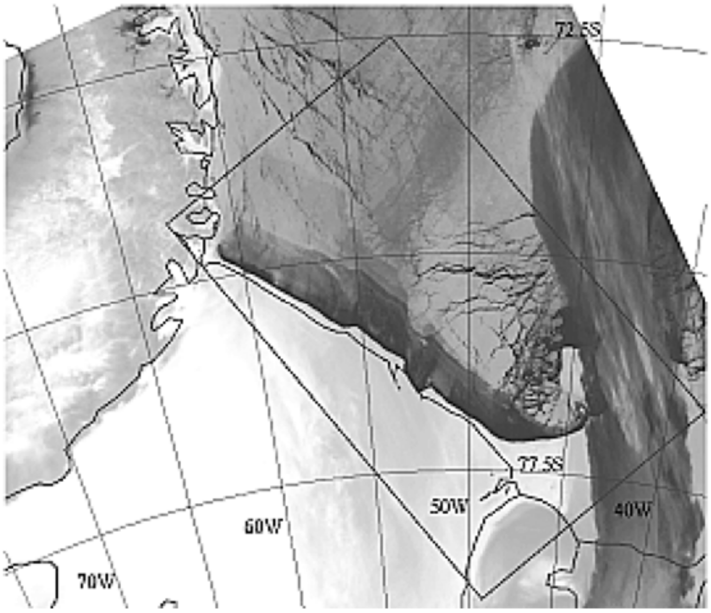


Figure 3.63. Infrared AVHRR satellite image of the southern Weddell Sea and the Ronne Ice Shelf at 19.30 UTC on 21 July 1998, processed so that white is cold and black is warm. The image shows a coastal polynya opening episode, with sea ice being blown off the ice shelf, exposing a belt of warmer open water underneath. The dark band on the right is a due to a relatively warm area of cloud. Reproduced from Renfrew *et al.* (2002).

This interaction underlies, in fact, the most outstanding illustration of the extent to which small-scale processes may influence the circulation of the Southern Ocean. 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 Antarctic Circumpolar Current (ACC), and that the bulk of that energy is transferred to the mesoscale eddy field via the action of baroclinic instability. Current conceptual and numerical models of the Southern Ocean overturning circulation (see Rintoul *et al.* [2001] and Olbers *et al.* [2004] for two recent reviews) unanimously highlight the eddies' crucial role in transporting water masses and tracers along the sloping isopycnals of the ACC — particularly in the upper overturning cell — as well as in fluxing the momentum input by the wind to the level of topographic obstacles, where it may be transferred to the solid Earth. The essential energetics implicit in these models consists of the viscous dissipation (most likely in a bottom boundary layer) of the energy contained in subinertial flows, with little or no impact on the ocean's interior stratification. While considerations based on geostrophic turbulence theory (Marshall and Naveira Garabato, 2007) and altimetric observations of an inverse energy cascade in the Southern Ocean (Scott and Wang, 2005) endorse the view that much of the eddy field's energy is ultimately fluxed toward the ocean floor, measurements of oceanic finestructure (Naveira Garabato *et al.*, 2004; Sloyan, 2005; Kunze *et al.*, 2006) suggest that a substantial fraction of the subinertial energy is dissipated via the generation of internal lee waves rather than exclusively by viscous bottom drag. These waves induce intense turbulent mixing upon breaking (Fig. 3.64) and, in so doing, contribute to driving the lower cell of the Southern Ocean overturning. It thus becomes evident that the upper and lower overturning cells, often treated as largely independent entities in descriptions of the ocean circulation, may be strongly coupled by sub-mesoscale physical processes in the Southern Ocean. This proposition is consistent with the concurrent intensification of eddy-driven isopycnal upwelling and turbulent diapycnal mixing in ACC regions of complex topography, which suggests that the eddy dampening that is required to sustain a vigorous isopycnal 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, other mechanisms are likely to enhance this transfer. In particular, theory suggests that the elevated eddy and internal wave energy levels in the ACC may provide an optimum setting for a nonlinear adiabatic interaction between internal waves and mesoscale eddies (Polzin, 2007) and the generation of internal waves by loss of balance in the mesoscale (Molemaker *et al.*, 2005) to prosper in the ocean interior. The occurrence of these processes is indicated by altimetric evidence of a significant forward eddy kinetic energy cascade at sub-deformation scales in the Pacific sector of the ACC (Scott and Wang, 2005), and has been hypothesized to play an important role in the Southern Ocean overturning by facilitating the meridional and vertical translation of water masses across (predominantly zonal) large-scale isopycnal gradients of potential vorticity (Polzin, 2007). In the upper kilometre of the ACC, this translation manifests itself in a richness of quasi-isopycnal finescale interleaving layers that are most evident in frontal regions and that may be strongly influenced by double diffusion (Joyce *et al.*, 1978; Toole, 1981). 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 rates of diapycnal mixing and turbulent energy dissipation associated with near-inertial wave breaking are likely lower than those characterizing topographically induced turbulence in the deep ocean (Fig. 3.64), and have been shown to vary seasonally and peak in winter, when wind forcing is strongest (Thompson *et al.*, 2007). The relative importance of double-diffusive processes in the driving of upper-ocean diapycnal mixing may be boosted in sea ice-covered regions (Robertson *et al.*, 1995).

The key point emerging from the preceding review is that sub-mesoscale physical processes in the Southern Ocean and their two-way interactions with the oceanic mesoscale likely 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 an alarming 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). Faced with the present period of rapid climate change, it is probable that our attempts to understand and predict future variability will be dangerously misguided until the role of small-scale processes in shaping and driving the Southern Ocean circulation is adequately acknowledged in the models.

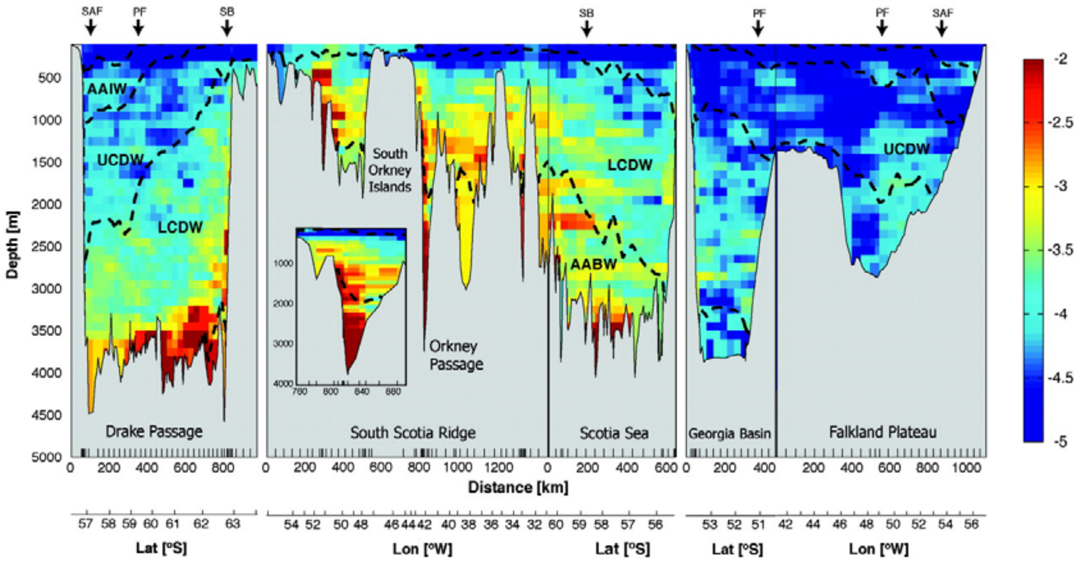


Figure 3.64. Vertical distribution of $\log_{10}$ of the turbulent diapycnal diffusivity (in $m^2 s^{-1}$) along a section following the rim of the Scotia Sea anticlockwise. Density surfaces separating water masses (AAIW, Antarctic Intermediate Water; UCDW and LCDW, Upper and Lower Circumpolar Deep Water; 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. Lat, latitude; Lon, longitude. Reproduced from Naveira Garabato *et al.* (2004).

Circulation and water masses of the ACC and the polar gyres from model results

A number of modelling studies, including both general circulation models and simplified theoretical models, have been carried out in an effort to improve our understanding of the basic circulation of the ACC. The studies have examined the controls on its transport strength and the mechanisms for import and export of various water masses for the ACC. The general circulation modelling studies have involved either regional Southern Ocean only or fully global domains, and in most cases also included the polar gyres, with the two main gyres of interest being the Ross Sea Gyre and the Weddell Sea Gyre, each of which circulates in a cyclonic fashion. These gyres are intricately linked to the ACC as they share 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 thereby brings its relatively warm Circumpolar Deep Water (CDW) masses in close proximity to the periphery of the Antarctic Ice Sheet.

A focus of the majority of modelling studies has been to identify the forces that constitute the dynamical balance for the ACC. At the latitudes of the Drake Passage where the ACC is unbounded, the current receives a zonally-continuous momentum input from a 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 of the bottom topography, with models with a smoothed representation of topography producing less form stress and thus higher transport values for the ACC than those with rougher topography (Best et al., 2004). It has also been found that 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. The radius of deformation for oceanic flows at high latitudes is relatively small, being of the order of several kilometres. So, in general, real ocean eddies are relatively small compared to a typical grid cell in an OGCM. Most modelling studies have been carried out at a relatively coarse resolution as compared to the radius of deformation, but some others at higher resolution (e.g., Maltrud and McClean, 2005), the latter being so-called eddy-permitting simulations. It has been found that 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. The influence in some regions of the ACC is for eddies to cause an upgradient transfer of kinetic energy into the mean flow, but for the major part of the ACC the transfer is downgradient (Best et al, 1999). Because coarse-resolution models have an inadequate representation of such eddy processes, and of topography, they produce an unrealistic simulation of the ACC transport, and the overall Southern Ocean circulation.

Under some scenarios of modelled climate change, there are significant changes in the wind field that drives the ACC. It has been found that for the ACC that the response to a change in the wind field occurs at the remarkably short period of two days at the latitude of the Drake Passage (Webb and de Cuevas, 2006). Furthermore, the response is largely barotropic 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 ACC to have a fast response 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 leads 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).

The numerical modelling of the ACC and the adjacent polar gyres has shed some light on the behaviour of this expansive current and its interaction with the polar gyres. There are, however, at least two pressing questions that remain poorly addressed. First, from a classical physical oceanography viewpoint, the transport volume of the ACC remains poorly constrained in different OGCMs, particularly so in models typically used in IPCC class simulations, which show a wide discrepancy in transport values even if the external forcing is similar (Ivchenko et al., 2004). Russell et al (2006) analyzed the ACC transport in 18 coupled atmosphere-ocean models and found that compared to the observational transport estimate of 135 Sv, the coupled models produced a spread of transport 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 exists in part because of a lack of grid resolution in many model simulations, a situation that is likely to alleviate itself in the coming years as Southern Ocean eddy resolving ocean models become increasingly prevalent. A second pressing question is the modelling of the interaction of the ACC and polar gyres with the periphery of the Antarctic Ice sheet. To properly tackle this problem an OGCM needs to be coupled interactively to an ice sheet model. While some regional modelling studies in which OGCMs have been coupled to static ice shelf models have been performed (see Sec. 1.16.4), no fully interactive modelling studies have been performed. This current gap in modelling ability leaves the climate modelling community without the ability to predict sea level changes that might arise from interactions of the waters of the Southern Ocean with the periphery of the Antarctic Ice Sheet.

3.11 **Biogeochemistry - Southern Ocean Carbon Cycle Response to Historical Climate Change**

3.11.1 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 CO2 [Rintoul, et al., 2001; Sarmiento, et al., 2004]. The Southern Ocean is dominated by the eastward flowing Antarctic Circumpolar Current (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 and in the north the downwelling provides a strong connection to water masses that resurface at lower latitudes e.g. [Sarmiento, et al., 2004]. These connections make the Southern Ocean extremely important in controlling the storage of carbon in the ocean and a key driver in setting atmospheric CO2 levels [Caldeira and Duffy, 2000].

3.11.2 The Marine Carbon Cycle

The marine carbon cycle can be described in terms of the anthropogenic and natural carbon cycles, but from an oceanic perspective they are treated and almost exactly the same. The anthropogenic carbon describes the emissions of CO2 into the atmosphere that has continued at an increasing rate since the start of the industrial revolution and is currently (in 2006) approaching 10 PgC/yr [*Canadell, et al.*, 2007]. The natural carbon describes behaviour of the carbon in the ocean prior to the industrial revolution. It is estimated that 30% of the total anthropogenic emissions annually is taken and sequestered by the ocean.

Air-sea CO_2 Fluxes

CO2 is exchanged between the ocean and the atmospheric primarily through air-sea fluxes and these 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]. Air-sea fluxes are a function of the differences in partial pressures between the ocean and atmosphere across this boundary (ΔpCO2), and the gas exchange coefficient (K). :

$$CO_2FLUX_{AIR-SEA} = K(pCO_{2AIR} - pCO_{2SEA}) = K(\Delta pCO_{2AIR-SEA}) \qquad (1)$$

The gas exchange coefficient (K) generally contains a power law dependence on windspeed and a weak dependence also on solubility and Schmidt number. As the time taken for CO2 for air and ocean to come to equilibrium is about 1 year, K takes this into account. To directly measure gas transfer velocity directly is extremely difficult hence many different relationships for gas transfer velocity e.g.[Caldeira and Duffy, 2000; Liss and Merlivat, 1986; Nightingale, et al., 2000; Wanninkhof, 1992] etc, have been proposed. As the Southern Ocean experiences the highest annual-averaged winds the selection of K may have an important role in determining fluxes.

3.11.3 Biological and Physical Pumps

Surface ocean CO2 levels (pCO2) and in turn atmospheric CO2 levels via air-sea fluxes can described as being controlled by the combination of physical and biological processes that move CO2 from the upper ocean into the deep ocean. These processed can be separated into the "physical pump" and the "biological pump" (Figure 3.65)

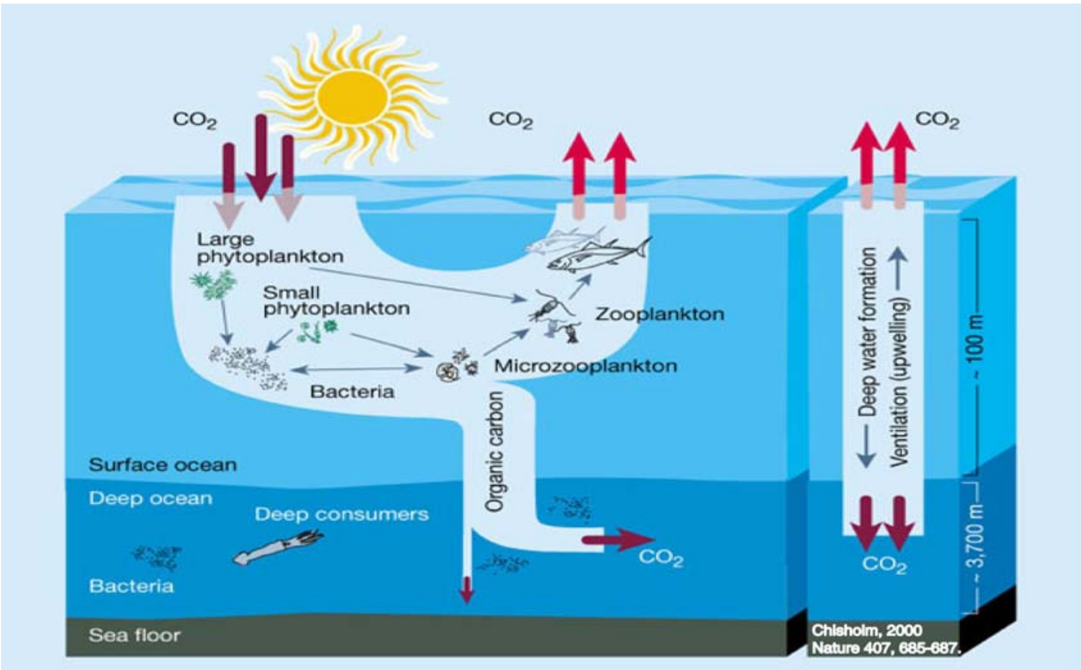


Figure 3.65. A simple schematic representation of the biological (left) and physical pumps (rights) [*Chisholm*, 2000]

The biological pump refers to the biological cycling of ocean carbon into the ocean interior. The biological pump 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 proportional to the nutrients available. This DOC is then processed, consumed and recycled within the marine ecosystem in the euphotic zone. A proportion of the organic material in the euphotic zone is exported, sinking under gravity to be either remineralised (converted back into DIC) or buried in the seafloor as sediment.

Excepted near the sea-ice, Antarctic continent and Islands, the Southern Ocean can be described as a high nutrient low chlorophyll (HNLC) region, this means that despite the region being replete in macronutrients (nitrogen, phosphate and silicate) needed by phytoplankton to grow, the phytoplankton abundance remains low. This has been hypothesised to be due to either low light levels at high latitudes, required for photosynthesis or a lack of micronutrients, in particular iron [*Martin and Fitzwater*, 1988] or both. 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 to change the distribution dissolved carbon between the upper ocean and into the deep ocean. Dynamical processes that can impact of 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 driving mixing. The thermodynamic response of the carbonate system is such that at a fixed atmosphere pCO_2, the upper ocean (in contact) can hold more dissolved inorganic carbon at cooler ocean temperatures than the same water mass at warmer temperatures. The physical pump can act in both directions to both increase and decrease surface ocean pCO2 levels at the same time. As an example in the Southern Ocean the upwelling of cold deep waters along the Antarctic Divergence can reduce surface pCO_2 levels through the thermodynamic effect, but at the same time increase surface pCO_2 by importing the total carbon at the surface.

3.11. 4 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 CO2 fluxes are well known but the magnitude remained poorly constrained [*Metzl, et al.*, 2006 and references herein). 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 strong insights into Southern Ocean behaviour at the seasonal timescales {Takahashi, 2008 #320]. During the Austral Summer biological production reduces surface oceanic pCO_2 through photosynthetic activity and then exports part of this to the deep ocean, meanwhile is offset by the changes in the physical pump reducing the capacity of the surface water to store CO_2 through changes in solubility, increasing the surface CO_2 levels. The net result of this competition between the biological and physical pumps such that the Southern Ocean acts a sink of atmospheric CO_2 in the summer in the sub-Antarctic zone and south of the Polar front (Figure 3.66). During the Austral Winter, a picture of two distinct zones separated by the Polar Front (PF ~50°S) 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 deep ocean leading to increase pCO_2 surface levels and a cooling hence increasing 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 Polar Front during the Austral Winter the same cooling and mixing processes that occur further south exist but with the addition of biological activity occurring during the period, albeit at a reduced rate, reducing the pCO_2 in surface waters. The combination of the responses of the biological and physical pumps mean that this region acts as weak sink of atmospheric CO_2 during the winter.

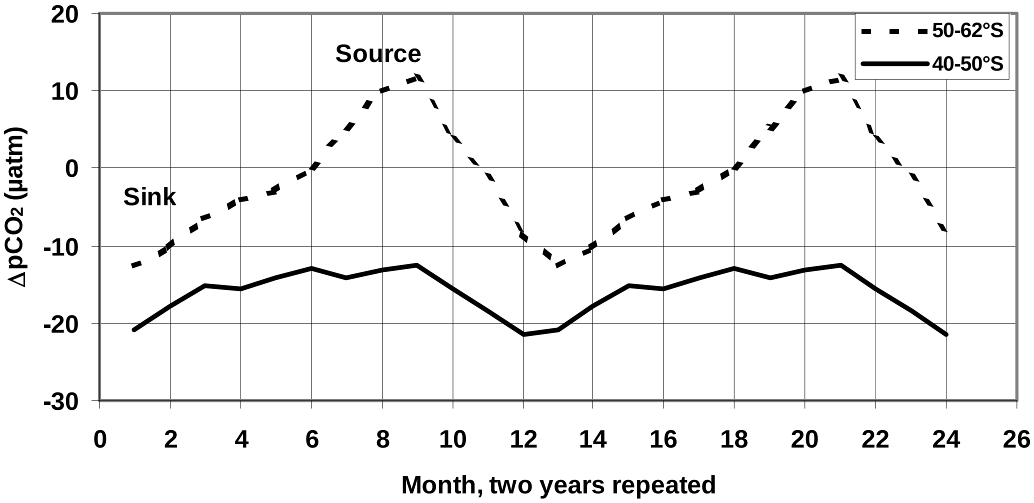


Figure 3.66. Annual cycle of ΔpCO_2 (pCO_2atm-pCO_2ocean) in the Southern Ocean for the regions 40-50°S (black line) and 50-62°S (dashed line). The monthly ΔpCO_2 have been averaged from the climatology of year 2000 [*Takahashi and al.*, 2008]. When ΔpCO_2 is negative (resp. positive) the ocean is a CO_2 sink (resp. source) with respect to the atmosphere. The Sub-Antarctic zone (40-50°S) is a permanent CO_2 sink but at higher latitudes the ocean is a sink during summer and a source during winter (Metzl et al., 2006).

When the winter and summer fluxes are integrated, the annual mean uptake is small south of 50°S (about -0.08 PgC/yr) but the sub-Antarctic zone behaves as a strong sink (-0.74 PgC/yr). Other estimates also indicate that the SAZ is a strong CO2 sink approaching -1 PgC/yr [*McNeil, et al.*, 2007; *Metzl, et al.*, 1999] which is important particularly as the SAZ is 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 that 40% of the total uptake. Note in these calculations we have used the gas transfer coefficent of [*Wanninkhof*, 1992] with the dataset of [*Takahashi and al.*, 2008].

Significant progress has occurred in recent years to simulate the annual mean uptake of CO_2 by the Southern Ocean as precursor to understanding interannual to decadal variability. Figure 3.67 shows that the annual mean uptake is well represented in comparison with that simulated from current ocean biogeochemical models (e.g. OPA/PISCES model [*Aumont and Bopp*, 2006]). Several models do contain a different representation of the annual cycle but the phase between each model and observations shows good agreement, suggesting that the underlying processes that drive this variability are well captured. We are in a good position to explore the simulated changes at interannual and longer timescales.

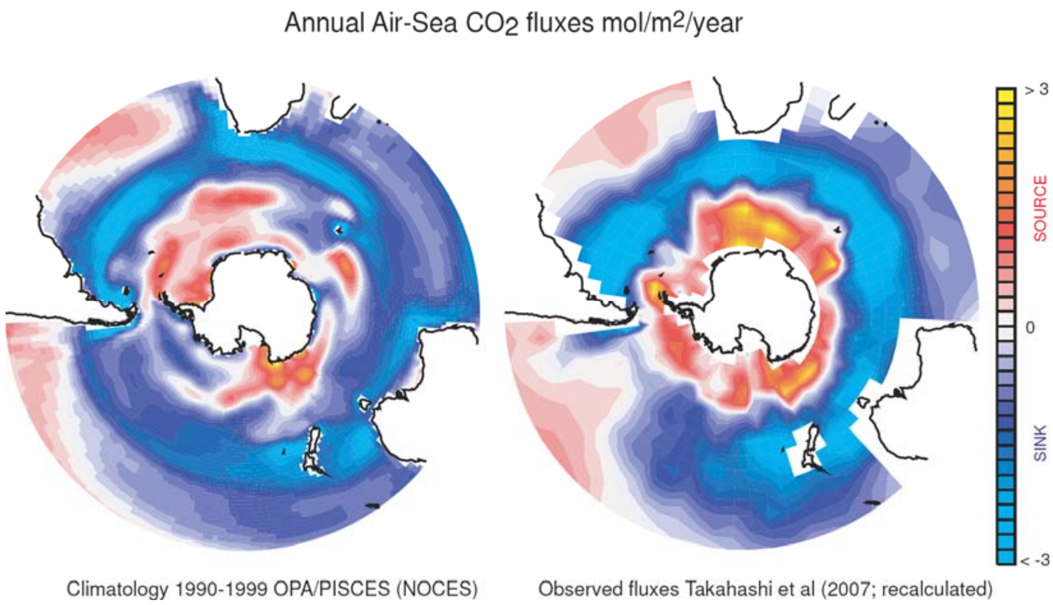


Figure 3.67. Annual mean uptake of air-sea CO2 fluxes as calculated from OPA/PISCES 1990-1999 [*Lenton, et al.*, 2006] and that from the new climatology of Takahashi et al (2008). The sub-Antarctic region (40-50°S) represents a strong sink (blue colors), whereas south of 50°S, large regions act as a CO2 source for the atmosphere (red).

3.11.5 Historical Change - Observed Response

The Southern Ocean has experienced significant changes in response to climate change; such as net increases 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 Southern Annular Mode (SAM) associated with changes in the strength of Antarctic Vortex, in response to primarily depletion of stratospheric O3 and increasing atmospheric greenhouse gas concentrations [*Arblaster and Meehl*, 2006]. The SAM is not the entire story and this has linear and non-linear interactions with other climatic modes such as ENSO and IOD to drive different responses in different regions. These physical changes impact directly on the physical pump and to a lesser extent on the biological pumps and hence the concentration of CO2 in surface waters the magnitude of uptake and export of both natural and anthropogenic CO2 from the atmosphere to the deep ocean.

In the Southern Ocean interannual changes in biological production, ocean dynamics and thermodynamics that drive pCO_2 and air-sea CO_2 exchanges remain poorly understood and very undersampled. Interannual variations have been observed at high latitudes (in the POOZ, 50-57°S) but at very few locations and during austral summer [*Jabaud-Jan, et al.*, 2004; *Brévière, et al.*, 2006] when the ocean experienced significant anomalies and maybe be related to large-scale climatic index such as ENSO and/or SAM [*Borges, et al.*, 2008]. Although these analysis gives important information regarding the response of the ocean to climate variability in the southern hemisphere, there is no clear detection of the decadal trends of oceanic CO2 and associated air-sea CO2 fluxes as it was inferred from long-term time series observations conducted in the north Atlantic or the north and equatorial pacific (e.g. [*Bates*, 2001; *Feely, et al.*, 2002]). *Takahashi et al.* [2008] have recently construction of pCO2 data synthesis, from this analysis they have been able to detect a significant increase of oceanic pCO2 during winter, about +2.1 µatm/yr, which is faster than in the atmosphere over the period 1986-2006.

Repeat underway measurements of surface pCO2 have been undertaken 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 resupply of Antarctic and subAntarctic bases they represent a valuable timeseries to explore the evolution of the surface ocean. A recent study by [*Metzl*, 2008] in the South Western Indian Ocean calculated surface trends of pCO2 between 1991-2007. These results show that in the Southern Indian Ocean, the oceanic pCO2 increased at all latitudes south of 20°S (1.5 to 2.4 µatm/yr depending the location and season). More specifically, at latitude > 40°S, it was found that oceanic pCO2 increased faster than in the atmosphere since 1991, suggesting the oceanic sink decreased. In addition, when pCO2 data are normalized to temperature, the analysis shows that the system is increasing much faster in the winter in the summer (Figure 3.68). These results suggest that the increase maybe due to ocean dynamics given that the largest response occurs in the Austral Winter. In the recent period (since the 1980s) the increase of pCO2 appeared to be faster compared to the trends based on historical observations 1969-2002 [*Inoue and Ishii*, 2005], suggesting that the Southern Ocean CO2 sink has continued to evolve in response to historical climate change.

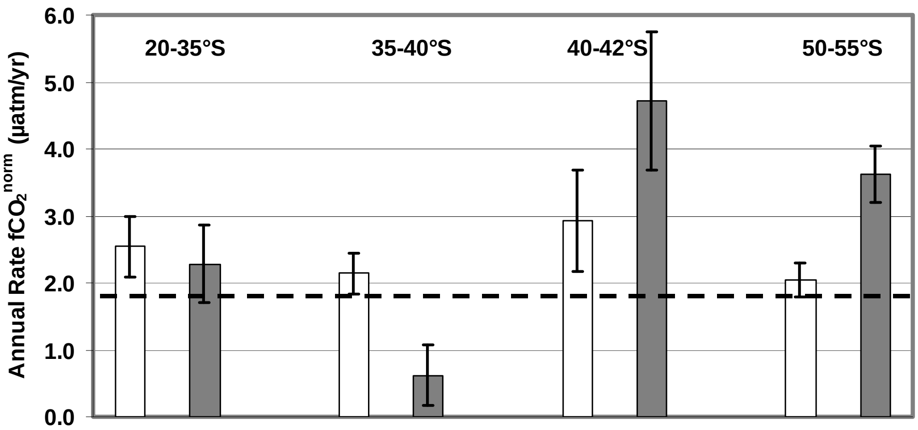


Figure 3.68. Annual mean trends of temperature normalized *f*CO_2 in 4 regions of the South-Western Indian Ocean (based on summer and winter observations in 1991-2007). The open bars indicate the growth rates estimated for summer and black bars for winter. Standard errors associated to each trend are also indicated. The dashed line indicates the atmospheric CO_2 annual growth rate (figure reproduced from Metzl, 2008).

Oceanic pCO_2 has been observed to be increasing faster than in the atmosphere in recent years consequently it might be possible to detect the signature of these changes in atmospheric CO2 data as it has been well observed in the Equatorial Pacific during ENSO event (e.g. [*Peylin, et al.*, 2005]). At latitudes north 40°S the ocean represents 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 for example is clearly seen in CO_2 record at La Nouvelle Amsterdam Island (in the South-Indian Ocean) where the seasonality of atmospheric CO_2 is very low. A recent study by [*Le Quere, et al.*, 2007] using a combination of atmospheric observations and inverse methods reported that in the period 1981-2004, that strength of the Southern Ocean CO_2 sink (south of 45°S) was reduced (Figure 3.69). Although this result remains controversial (see [*Law, et al.*, 2008] and [*Le Quéré, et al.*, 2008]), it does suggest that the oceanic CO2 levels are increasing implying that strength of the Southern Ocean CO_2 sink in response this has been reduced. The increase in upper ocean CO_2 values were attributed to changes in the natural carbon cycle through wind speed increases dominating the total response of the ocean. One of the most important aspects of this results is that historically it was believed that the anthropogenic concentration of CO_2 was driving Southern Ocean CO_2 uptake. These results is also consistent with a number of recent modeling studies, that have also suggest a decrease in uptake in the last decades [*Lenton and Matear*, 2007; *Verdy, et al.*, 2007].

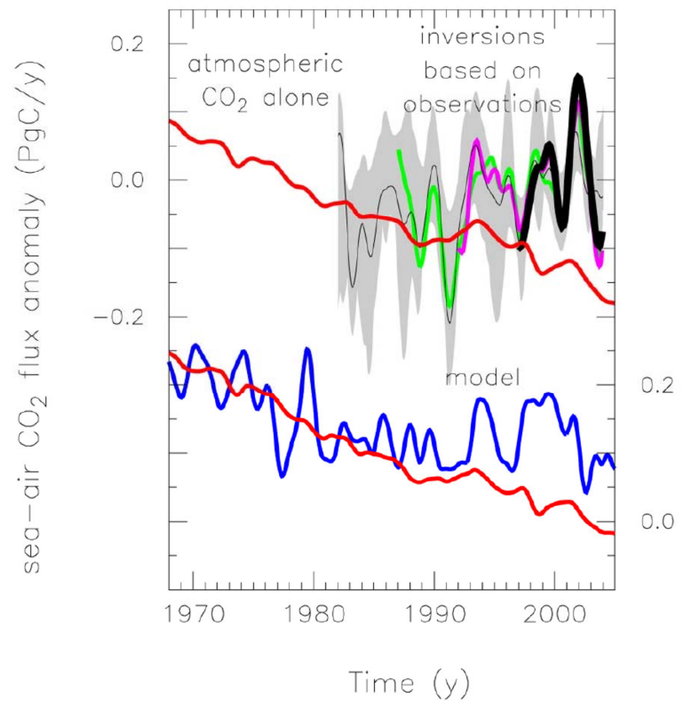


Figure 3.69. Sea-air CO_2 flux anomalies in the Southern Ocean (PgC/y, >45°S) based on atmospheric CO_2 data and inversed transport model (on top) and a global biogeochemical ocean model (bottom). Compared to experiments that do not take into account the climate variability (in red), both approaches suggest a stabilization or reduction of the ocean CO_2 sink since the 1980s (*LeQuéré et al.*, 2007).

3..1.6 Historical Changes – Simulated View

To explore changes how Southern Ocean air-sea CO2 fluxes have responsed to historical climate change between 1948-2003 [*Matear and Lenton*, 2008] used a biogeochemical ocean only model driven with NCEP R1 products[*Kalnay*, 1996]. They explored how the total carbon cycle as well as the natural and anthropogenic responded to the observed increases in windstress, heat and freshwater fluxes (Figure 3.70). The results from this study show a complex response, when only either heat or only freshwater fluxes varied the uptake the total uptake was slightly greater than the total CO2 flux response. 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 dominates the total response. The natural carbon sink dominates the total response because 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 over this period strong enough to counter the increased CO2 in surface waters due to winds bringing up water rich in dissolved inorganic carbon from the deep ocean in the Southern Ocean. Although there is some question of the validity of the changes in the presatellite era (1948-1979) [*Marshall*, 2003], it is important to note that the largest changes do occur in the period 1979-2003 (as seen in Figure 3.69, [*Le Quere, et al.*, 2007]).

As reanalysis products used in [*Matear and Lenton*, 2008] are an assimilated product all climate modes are represented. Other studies using different ocean models and experiments to explore the response of the Southern Ocean air-sea CO2 flux to the Southern Annular Mode (SAM) alone [*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 these studies suggest that more than 40% of the total variance in CO2 flux can be explained in the recent period by the SAM

Over same the period only a small increase in primary production or export production was 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 SO. Over the rest of the Southern Ocean, very little response was seen, because it is both light and iron (Fe) limited and the nitrate:iron ratio of the upwelled water still has excess nitrate relative to Fe (Bowie personal communication). Therefore even if extra Fe is brought into the system it will not be sufficient to drawdown the additional nitrate and carbon supply. In addition although upwelling of deepwater would supply more Fe to increase export production, the additional export of carbon from the upper ocean would be small given the observed range in SO deep-water Fe concentrations 0.4nM-2.8nM [*Löscher, et al.*, 1997] and the observed carbon:iron (C:F) ratio of organic matter [*Buesseler, et al.*, 2004]. Therefore only a weak link between atmospheric forcing and export production exist and hence small-simulated interannual variability in export production.

The increased ventilation of the Southern Ocean (SO) from simulations does not only alter the air-sea CO2 fluxes, and hence the concentration of CO_2 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 CO2, through changes the Revelle factor [*Revelle and Suess*, 1957] and through seawater changing pH and become more acidic. This pH change or ocean acidification means that the ability of organisms that use calcite to build shells is reduced, which impacts biological feedbacks now and in the future [*Feely, et al.*, 2004]. In response to ocean acidification a key carbon parameter is the aragonite saturation state (Ω_A), which does influence the rate of calcification of marine organisms [*Riebesell, et al.*, 2000; *Langdon and Atkinson*, 2005]. Simulations show that the increases and variability in heat, freshwater fluxes and in particular wind stress observed in the last 50 years, have moved this saturation closer to the surface, potentially having already impacting on the ecosystems in the Southern Ocean

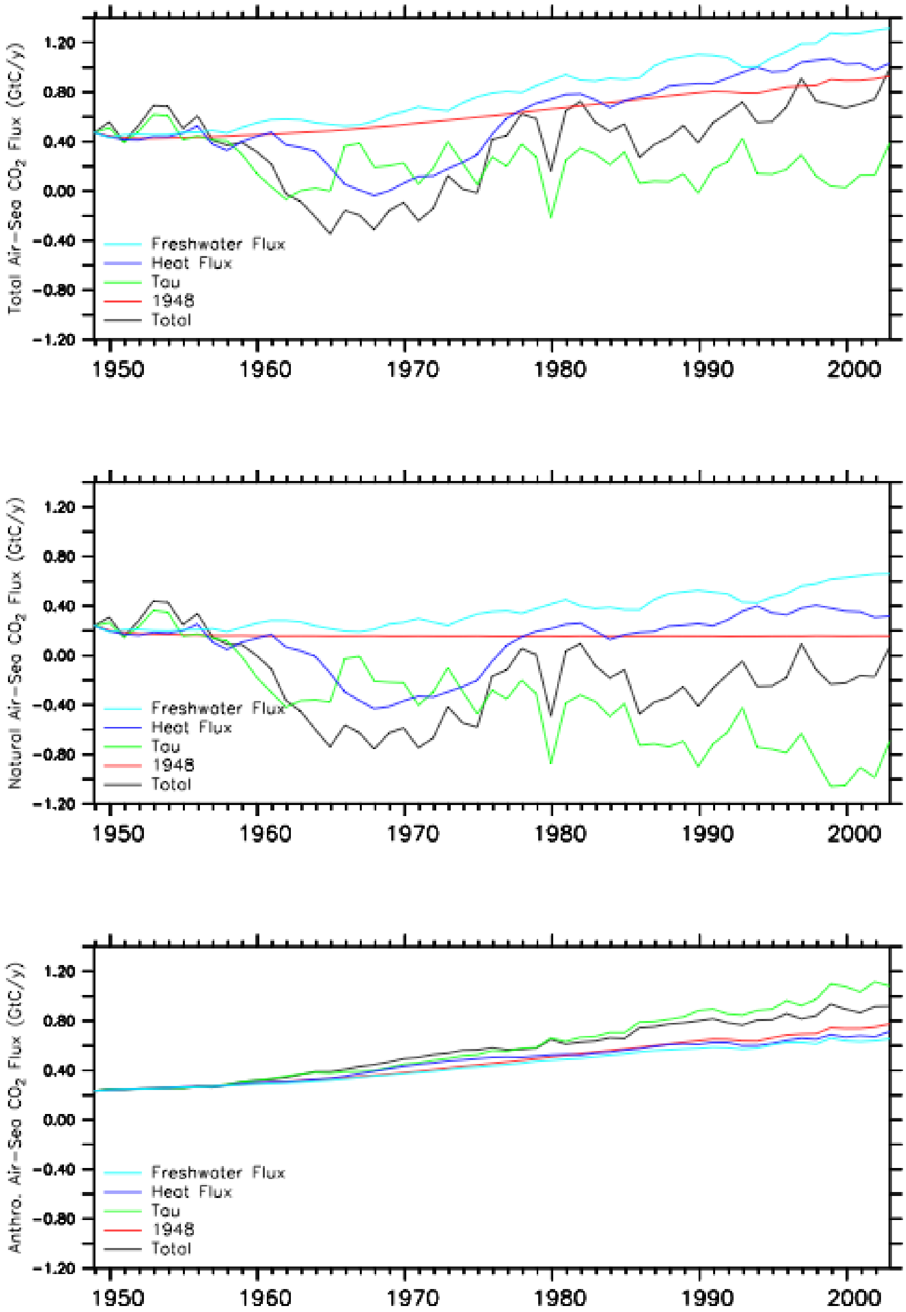


Figure 3.70. Annual-averaged Southern Ocean uptake of: a) total carbon; b) natural carbon; c) anthropogenic carbon. The different experiments use the following colour coding, total experiment (black line), 1948 (red line), tau (green line), hflx (blue line) and fflx (cyan line).

3.11.7 Changes in CO_2 inventories

In the previous sections, we focused on the changes of the air-sea CO2 fluxes as observed in recent decades and simulated over the last 50 years. Regarding the capacity of the ocean to reduce climate changes, another important evaluation is the anthropogenic CO2 changes in the water column since the preindustrial era. This is important not only to estimate the global ocean capacity to absorb anthropogenic CO2 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]. The anthropogenic CO2 in the ocean (Cant) cannot be directly measured but under several assumptions, Cant 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]). A recent comparison of data-based methods ([*Vázquez-Rodríguez, et al.*, 2008]) clearly show that all methods converge to estimate large inventories associated with mode and intermediate waters (Figure 3.71). Its has been estimated that in recent decades that 30% of the total uptake since the preindustrial period have been taken up by Southern Ocean mode waters [*Sabine, et al.*, 2004]. Uncertainties still exist in Southern Ocean Cant uptake south of 50°S, but these are differences are based methods that have been recognized to underestimates anthropogenic carbon in bottom waters along the Antarctic coast (*Lo Monaco, personal communication*). The spatial pattern and magnitude of estimates of Cant derived from observations are consistent with those simulated by ocean models [*Orr, et al.*, 2001]

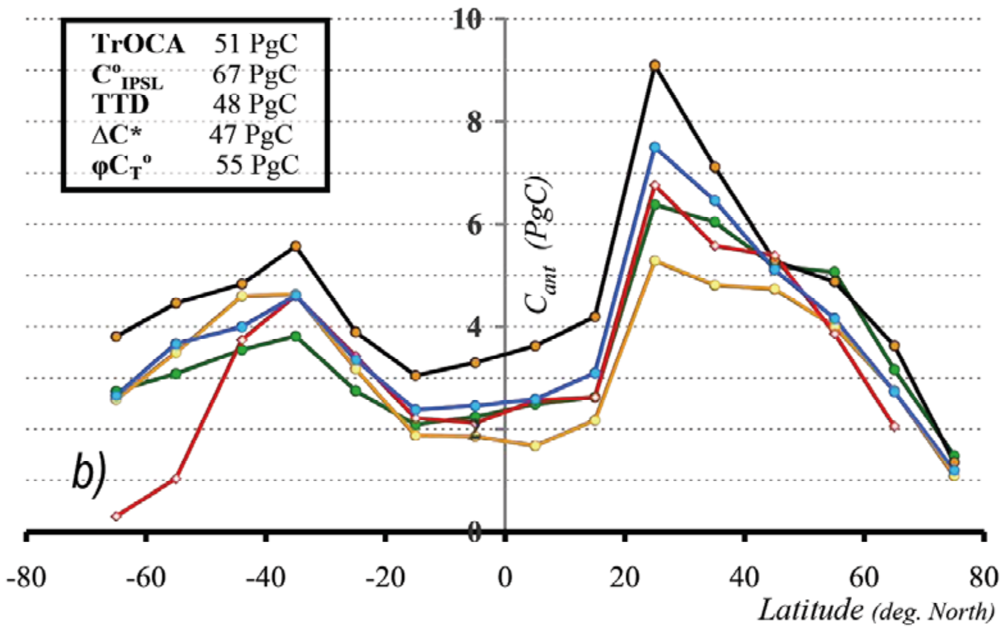


Figure 3.71. Anthropogenic carbon inventories (Pg C) in the Atlantic Ocean (from the Artic to the Antarctic) derived from different data-based methods. The inlayed box gives the integrals of the total inventories (Pg C) for each method (after [*Vázquez-Rodríguez, et al.*, 2008]). This analysis illustrates the uncertainty of the Cant estimates at latitude > 40°S in the Southern Ocean.

3.11.8 Concluding Remarks

The Southern Ocean plays a critical role in the uptake of anthropogenic CO2, more that 40% of the annual mean uptake. Modelling and observational studies have shown that the Southern Ocean has undergone significant changes in the last 50 years, and are converging towards a coherent view. The largest change has been a reduction in the total uptake in recent decades in response to the observed changes in climatic forcing particularly changes in wind speed. These changes in wind speed drive strong changes in the physical pump specifically ocean dynamics, rather than solubility or changes in the Revelle factor; and the lack of a strong response in the biological pump reinforces this view. In the future, in response to climate change, both the biological and physical pumps are expected to significantly.

In terms of simulating the Southern Ocean response to climate change the largest uncertainties in our results lie in the hydrological cycle, the most poorly observed of all the forcing fields. In order to better understand and quantify the changes due to historical and future climate change a much better estimate of freshwater inputs and outputs are required.

The Southern Ocean continues to be under sampled with respect to carbon. It has been recognised by the international CO2 community this is region that the construction and would give valuable delivery of a CO2 ocean data products e.g. ([*Metzl, et al.*, 2007]). Such data sets would provide valuable observational evidence for understanding historical climate change and would provide valuable insights, through interannual changes, into how the Southern Ocean may respond in the future [*Matear and Lenton*, 2008]. The changes in the last 50 years highlights the need also for well considered and resourced sampling plans to capture predicted future changes e.g. [*Lenton, et al.*, 2008] .

3.12 **Antarctic Sea Ice Cover during the Instrumental Period**

3.12.1 Variability and Trends using Satellite Data since 1979

The large scale seasonal variation of the sea ice cover in the Southern Hemisphere is illustrated by the color-coded multiyear monthly averages of the ice cover as presented in Figure 3.72. We use the AMSR-E ice concentration data 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 series of images starts in January which is a mid-summer month and usually the warmest and the time of highest melt rate. The minimum extent usually occurs in February although sometimes it happens in early March. It is apparent that 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 varies slightly in magnitude and location from one year to another. The period of most rapid growth is from April to June and by July it has reach close to its maximum extent which is normally reached in September or October. The period from November to January is then a period of rapid decay. Around the continent, sea ice advances the least from the coastline mainly in the Western Pacific (90^{o}E to 150^{o}E) and the tip of the Antarctic Peninsula where you would expect it to happen (i.e., coastal area are 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 rectangle corner at about 217 ^{o}E, which is in part influenced by the shape of the bathymetry in the same region.

Among the most distinctive figures in the inner zone of the ice cover are the reduced ice concentrations adjacent to the Ross Ice Shelf, at or near the Maud Rise, and near the Cosmonaut Sea. These are regions where short term (or transient) polynyas usually occur in mid-winter, as reported in Zwally et al. (1986), Comiso and Gordon (1998), and de Veaux et al. (1999), and which have been associated with the initiation of deep ocean convection and the formation of 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 are vulnerable to divergence and melt because of relatively thinner ice and warmer water in the underside of the ice. Such regions have also been the scene of high productivity (Arrigo et al., 2003?) reflecting possible influence of sea ice as have been suggested previously (Smith and Nelson, 1986).

The parameters we use for quantifying the variability and trends in the ice cover are ice extent and ice area both of which are derived from the passive microwave data. Ice extent in our case is defined as the sum of the area of 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. A time series of monthly values of the ice area, ice extent and also ice concentration as derived from SMMR and SSM/I data from November 1978 to December 2006 are presented in Figure 3.73. Over the 28 year period of consistently processed passive microwave data, the seasonality of the sea ice extent and areas appears to be very similar. The peaks and dips varied only slightly despite relatively larger interannual variabilitions 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. 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 for ice extents are coherent with those of ice area with the latter being lower in values, as expected.

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. It is thus apparent that 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 apparently counter-intuitive phenomenon is actually more apparent especially 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 [Comiso and Gordon, 1998; Zwally et al., 1983].

To assess interannual changes and trends in the sea ice extent and ice area, monthly anomalies are generated by subtracting climatological averages of each month from each monthly average. The climatological averages are averages of data from November 1979 to December 2006 using SMMR and SSM/I data. The trend results for extent in the Southern Hemisphere shows a positive trend of 0.98 ± 0.23 % per decade for the entire Southern Hemisphere. This is consistent with the 1.0 ± 0.4 % per decade reported by Zwally et al. [2002] for the 1979 to 1998 period. The trend in ice area is slightly more positive at 1.75 ± 0.25 %/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 associated with different calibration and resolution of the SMMR and SSM/I sensors are not reflected by the error. Assuming that the latter is small as inferred from the overlap 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. (Comiso, 2000, Kwok and Comiso, 2002).

Trend analysis of the ice extent in different sectors (as defined in Zwally et al., 2002) of the Antarctic region (see Figure 3.74) yielded positive trends of varying magnitude in all except in the Bellingshausen Sea sector. The trend is least positive at the Weddell Sea sector at 0.8 ± 0.6 followed by Western Pacific and Indian Oceans at 1.1 ± 0.8 and 2.0 ± 0.6, respectively. The most positive is the Ross Sea sector at 4.5 ± 0.7 %/decade. The trend in the Bellingshausen/Amundsen Seas Sector is -6.0 ± 1.0 which serves to balance the relatively high trend at the Ross Sea. Since these two sectors are adjacent to each other, the opposite trends in the two sectors are in part caused by the advection of ice from one sector to the other. However, the Antarctic Penninsula that is adjacent to the B/A sector has been an area of climate anomaly as described previously by King [1994] and Jacobs and Comiso [1997]. Also, the Ross Sea region has been associated with influences 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; Duran et al., 2001]. 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.

For completeness, we used a 5 day running average to determine the day of ice maximum and ice minimum for each year and presented in Figure 3.75. These two parameters are associated with the thermal state of the ice covered ocean and provide a means to assess if the Southern Ocean has been changing. Trend analysis of the data yielded results very similar to those of the overall trend with the trends of maximum ice values being 1.1 ± 0.4 % per decade and 1.6 ± 0.4 % per decade for ice extent and ice area, respectively. The trends of minimum ice values are 0.8 ± 2.8 and 1.6 ± 3.2 % per decade for ice extent and ice area, respectively, indicating similar trends as well but higher statistical error.

3.12.2 Sea Ice Cover during the ESMR and pre Satellite Era

While not used in the time series analysis, the sea ice cover during the 1973 to 1976 period was well documented (Zwally et al., 1983). As indicated earlier, combining this data with SMMR and SSM/I data is difficult because of no overlap data to ensure that the ESMR data represents the same ice edge as the other data sets, too many gaps including more than 3 months in succession in 1975; and inaccurate ice concentration because of only one channel for the sensor and therefore inability to correct for spatial variations in emissivity and temperature. Qualitative analysis, however can be done. For example, we can take a four-year average of the ESMR data and compare it 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.

Going backward in time into the pre-satellite data is even more challenging. Ship observations have been compiled especially during the whaling period (McKintosh) A comparison of satellite observations with the McKintosh compilations are presented in Figure 3.76. If we can assume that the McKintosh data provides a meaningful representation of average ice edge locations during the period, it is apparent that the data from the previous several decades (as represented in the McKintosh data) show relatively more extensive ice cover than the average from satellite data. However, they are normally to the south of the farthest north the ice went during the satellite era.

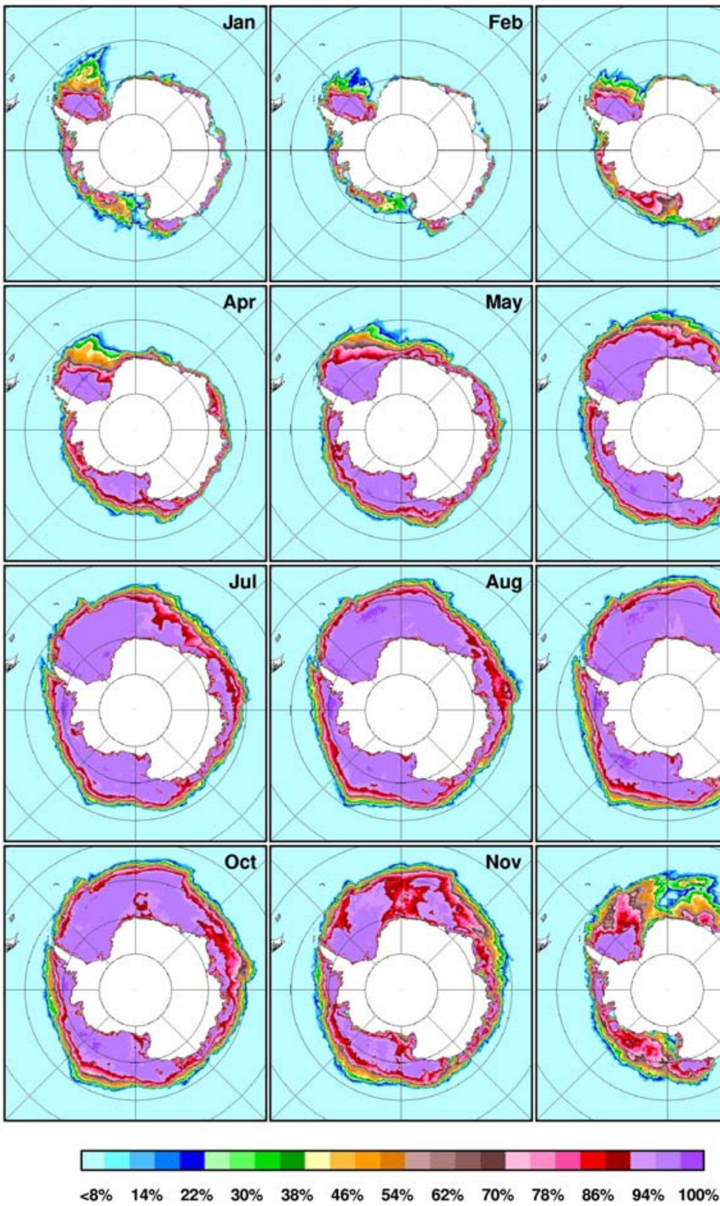


Figure 3.72. Monthly climatology of the sea ice cover as derived form AMSR-E data (2002 to 2007).

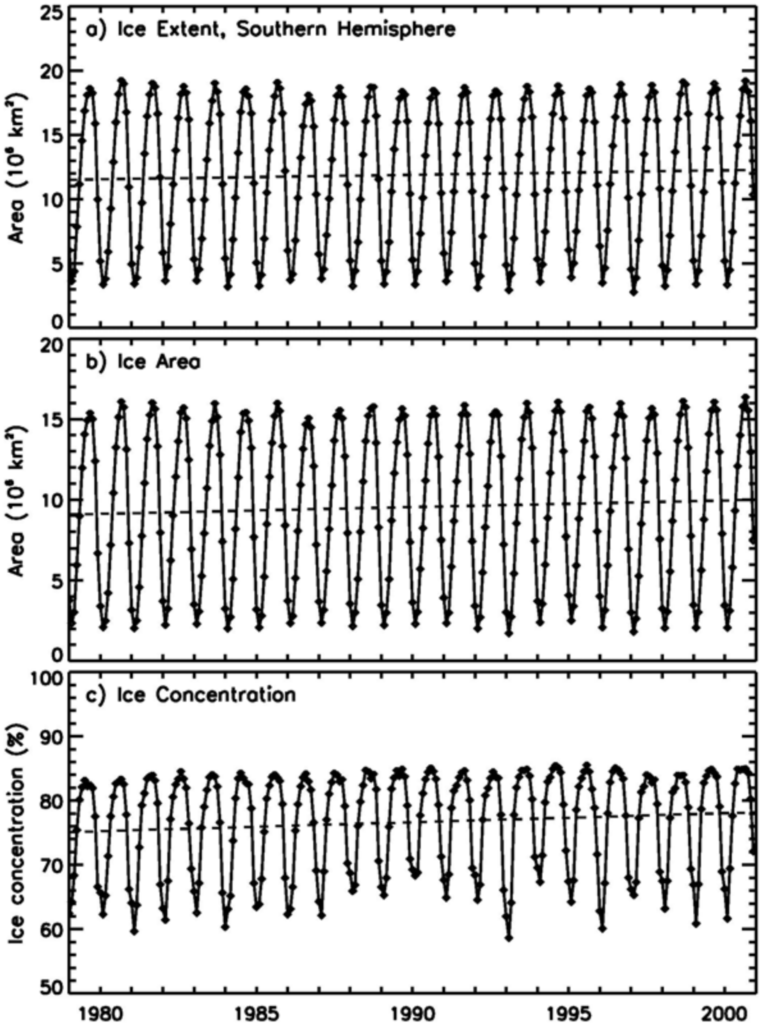


Figure 3.73. Monthly extent, area, and average ice concentration in the SH

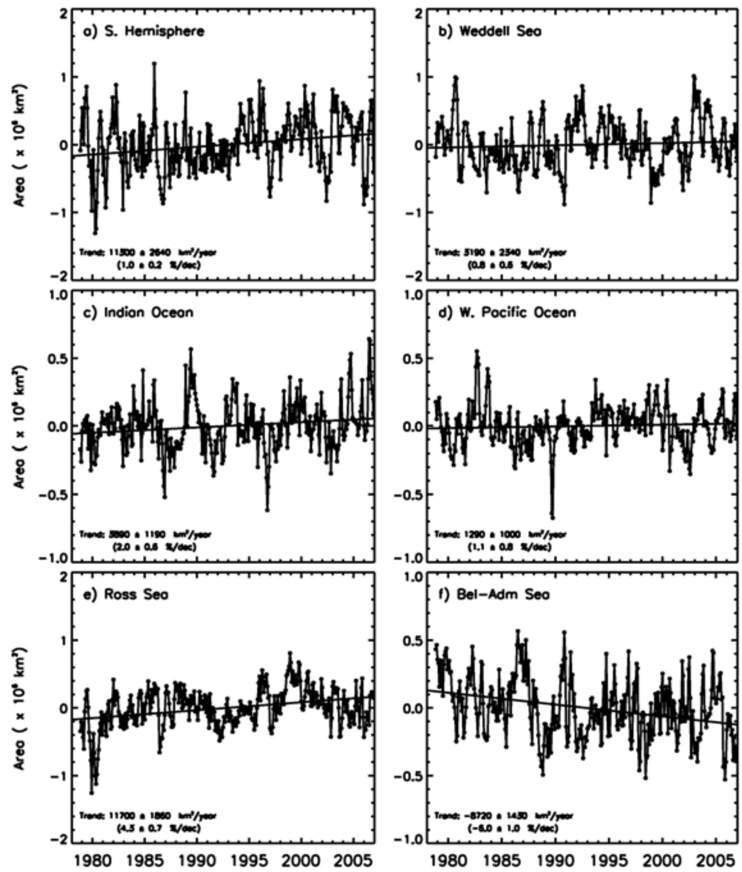


Figure 3.74. Ice Extent monthly anomalies and trend lines for (a) the Southern Hemisphere; (b) Weddell Sea; (c) Indian Ocean; (d) W. Pacific Ocean; (e) Ross Sea; and (f) Bellingshausen/Amundsen Seas Sectors

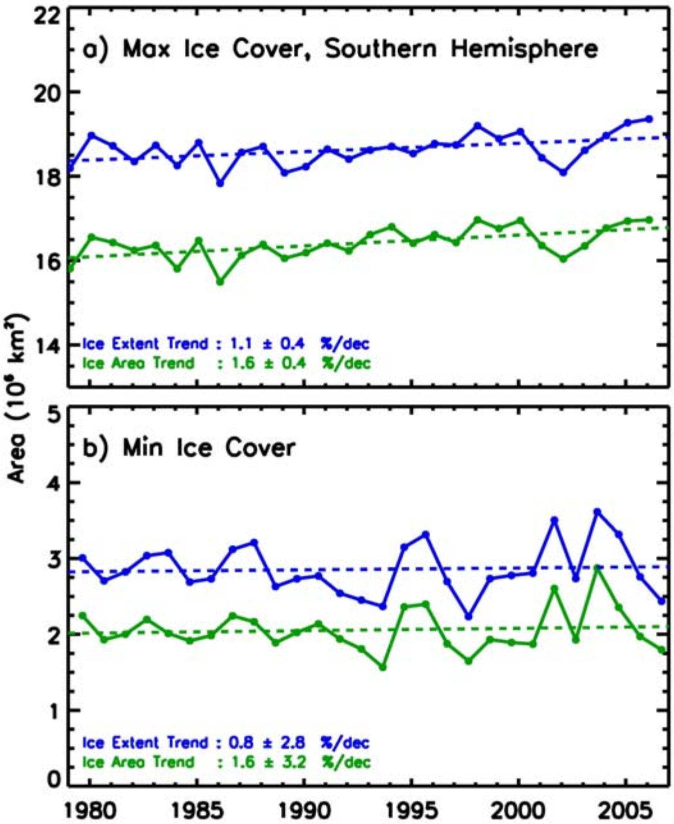
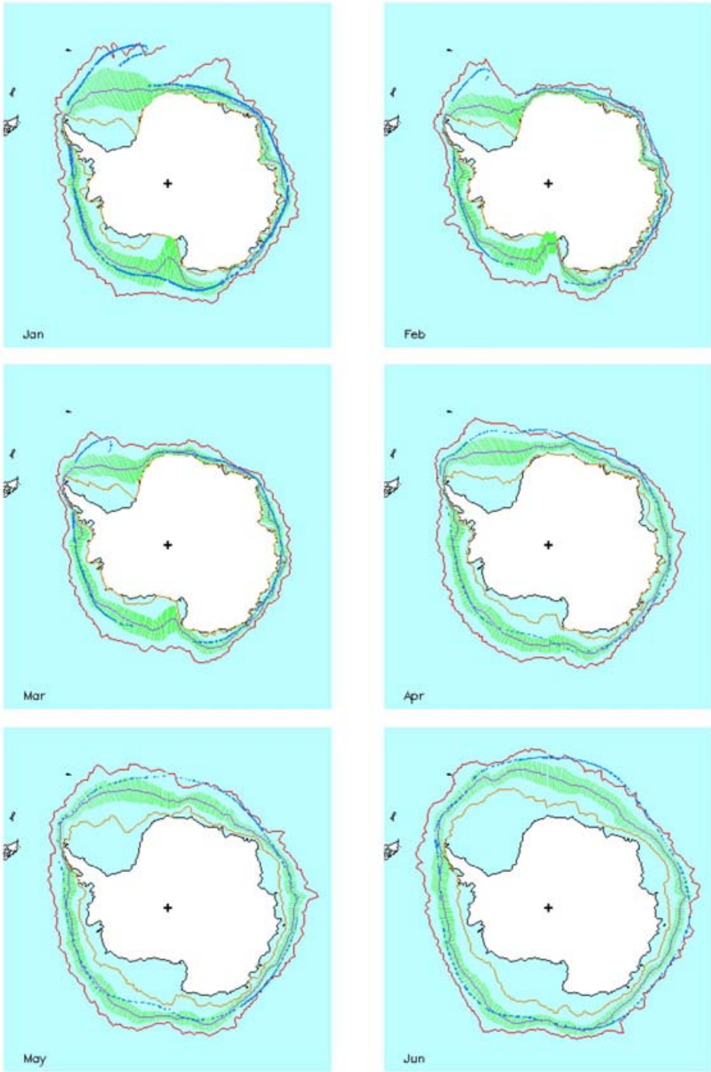


Figure 3.75. Yearly maximum and minimum extent and area of the ice cover in the Southern Hemisphere and trend results.

(a)



(b)

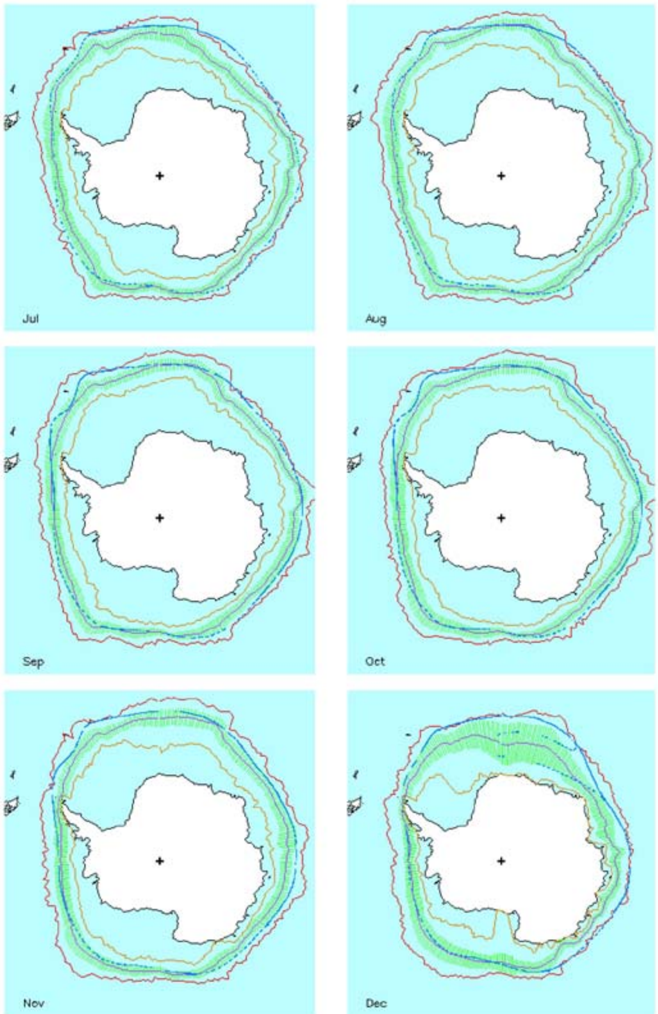


Figure 3.76. Average location of the ice edge during the satellite era (black), its standard deviation (in green), its most northern location (red) and most southern locations (orange). Data compiled by McKintosh during the whaling period is shown in blue. (Note: McKintosh digital data was provided by P. Wadham).

3.13 Changes in Antarctic permafrost and active layer over the last 50 years

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 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 considering depth progressively greater. The active layer thickness responds seasonally to the climate input. Therefore different methods are involved to monitor the permafrost thermal regime and the active layer thickness. Permafrost thermal regime is monitored 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 100x100 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 0°C isotherm, the temperature of the active layer is recorded at different depths, at least during the summer season.
In Antarctica the probing method has been tested for some years (Guglielmin, 2006) with not good results because the coarse size of the main part of the terrains.
For these reasons, a practical and basic method to monitor the active layer thickness has been proposed for the Continental Antarctica and it consists in measuring in each grid points the ground temperature at different depths (2, 5, 10, 20 and, where possible, 30 cm) with a needle thermistor only one time during the period of maximum thawing. The active layer thickness is so calculated by interpolating the two deeper temperature measurements (Guglielmin, 2006).

The Antarctic scientific community recognised the importance of these indicators and supported the creation of a SCAR Geosciences Standing Scientific Group (GSSG) on Antarctic permafrost and periglacial environments. GSSG prepared a research project called ANTPAS to fill the gap, improving the monitoring protocols and sites around the Antarctica (Bockheim, 2004), and to promote the creation of an Antarctic network, as already proposed by Guglielmin *et al.* (2003) for Victoria Land.
Active-layer thickness depends primarily on GST and the thermal properties of the ground, especially by its ice/water content. Not all the sites are really sensitive to the climate change because the heat convection, the lateral heat advection and the thermal properties of the ground produce a "thermal offset" according to Romanovsky and Osterkamp (1995, as the difference between mean annual GST and mean annual permafrost temperature) which can change in time getting harder the modelling of the relationships between climate factors and active layer thickness. In order to have the better climatic significativity, small thermal offset are required.

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

Vegetation coverage can influence deeply the ground thermal regime both because can change 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). Vegetation effects were recognized at microscale and sometimes, also at local scale, both in Maritime Antarctica (Cannone et al, 2006) and in Continental Antarctica (Guglielmin et al., 2007).

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

Since from the 1960s in many occasion, especially in Maritime Antarctica, ground temperature has been measured for different purposes, through different protocols, in general for short periods (up2-3 years) and 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 Balls, 1974; Robertson and Macdonald, 1962; Sawagaki, 1995; Thomson et al., 1971; Walton, 1977; Wójcik, 1989;).

Actually permafrost and active layer monitoring sites are located both in Continental Antarctica (Victoria Land and Queen Maud Land), as well as in different locations of Maritime Antarctica (Table 3.3). In addition there also the 5 deep boreholes (up to 282 m) drilled during the 70’ during the Dry Valley Drilling Project (DVDP) possibly available to measure permafrost temperature (Decker, 1974).

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	77º51’S 160º36’E	1273	19	P,R	4
Mt. Fleming	77º33’S 160º17’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	78º30’S 166º45’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	77º20’S 161º37’E	412	1.1	P,C (1999)	4
Victoria Valley	77° 20'S 161°37' E	380	11	P,R (2003)	3
Granite Harbour	77º00’S 162º31’E	4,6	1,2	P,C (2003)	1
		Northern Victoria Land			
Adelie Cove	74º47’S 163º58’E	35	6,1	P,R (2003)	3
Boulder Clay	74º45’S 164º01’E	205	3,6	P,C (1996)	3
Oasi	74º42’S 164º08’E	84	15,5; 30,3	P,C (2001;2008)	3

Mt. Keinath	74º33’S 163º59’E	1100	1	P,C (1998-2004)	3
Simpson Crags	74º26’S 162º53’E	830	7,8	P,C (1998-2002)	3
		Maritime Antarctica			
Livingston Island 1	62º39’S 60º21’W	35	2.4	AL,C (2000)	5
Livingston Island 2	62º39’S 60º21’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	74º34’S 11º13’W	1286	1,2	P,C (2003)	10
Wasa	73º02’S 13º26’W	450	1,2	P,C (2003)	10
Fossilbryggen	73º24’S 13º02’W	550	1,2	P,C (2003)	10

Table 3.3 – 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 the parenthesis 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 (1993); 7) Guglielmin M., Strelin J., Sone T., Mori J.; 8) Sone et al., (2001); 9) Guglielmin M. and Worland R.; 10) Boelhouwers J.

Despite the relatively small extent of ice-free areas, the actual distribution of the monitoring sites is still largely insufficient to characterise properly these areas also in consideration of the large spatial variability of the active layer, as documented by the example of Fig. 3.77a. In Antarctica the active layer monitoring is generally more difficult that in the traditional arctic areas because its spatial variability is very large due to the heterogeneity of the coarse ground sediments as well as the general high surface roughness.

The monitoring grid (100 x 100 m) located at Boulder Clay, a flat area close to the MZS station in Northern Victoria Land (Continental Antarctica)(Guglielmin, 2006), provides a striking example of the large areal variations of the active layer (Figure 3.77a), which is mainly driven by the different snow accumulation (Fig. 3.77b) influenced by the surface roughness. Actually only other two grids are installed

(Simpson Crags and Signy Island) but in both cases logistical constraints did not allow the monitoring to be repeated every year.

Future sites proposed in the framework of the IPY project "ANTPAS" both for active layer and permafrost monitoring will improve the network, although probably they will be still not enough to complete an optimal distribution.

3.13.2 Changes in the last 50 years and recomendations for the future.

Permafrost monitoring in Antarctica is a relatively new issue, although already in the 1960s Chambers (1963) at Signy Island in the South Orkneys started to monitor the active layer thermal regime and its implications on the patterned ground.

More recently (since 2001) new data on thermal active layer (Cannone et al., 2006; Guglielmin et al., 2007) were obtained. A comparison between the new active layer thickness data and the data collected at the same location four decades earlier was carried out by Cannone et al., (2006). Using a regression equation for air temperature versus ground surface temperatures the patterns of changing air temperature over time suggest that the active layer thickness changed. The analysis indicated 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 endured a series of particularly cold winters. The site of Boulder Clay (McMurdo Sound) represents the longest and almost continuous data series of permafrost and active layer temperature (Guglielmin, 2004; Guglielmin, 2006). Figure 3.78 shows the temperature recorded in the proximity of the permafrost table (at the depth of 30 cm) and at the end of the borehole (360 cm deeep).

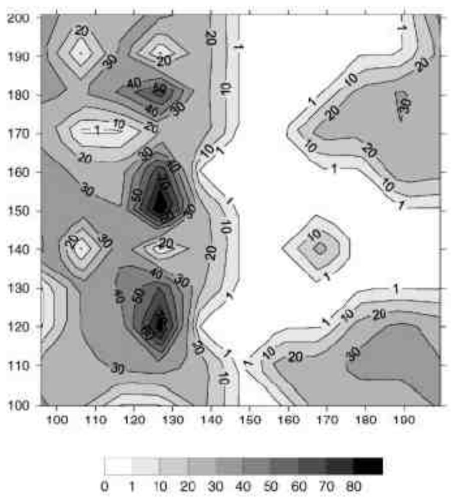
There is a substantial stability of the permafrost temperature at 360 cm, while at the permafrost table the temperature shows a sligth decrease of 0.1°C per year (Figure 3.78). This slight decrease (Guglielmin, 2007, in prep) is mainly related to the decrease of the air temperature and the decrease of the snow cover in the winter, at least in this site.

The decrease of ground surface temperature in relation to the decrease of surface air temperature confirms the pattern enhanced for the Dry Valley by Doran et al (2002) for the period 1986 to 1999 at Lake Hoare.

Not very far from Boulder Clay (only 5 km northward), at the MZS station a previous 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 meters far, was drilled as a template for the new IPY-ANTPAS monitoring network. The thermal profile obtained in the 30 m borehole (Figure 3.79) suggests at least two periods of cooling below the ZAA (around 14-15 m), following a previous period of warming. Guglielmin (2007, in prep) describes in detail these quite short fluctuations of the ground surface temperature at MZS in the last 30 years.

The paucity of the permafrost and active layer data and, above all, the lack of long-term monitoring periods avoid the use of these climatic indicators to draw any definitive conclusions on the impact of climate change on this component of the Antarctic cryosphere.

On the other hand, the proposed and the funded new boreholes and grids and a more strict coordination and standardization of the protocools as suggested by the IPY-ANTPAS project would provide for a next future a more stregthfull database.



100 110 120 130 140 150 160 170 180 190
0 1 10 20 30 40 50 60 70 80

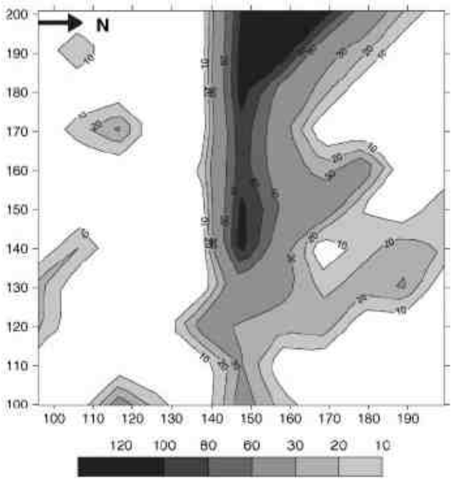


Fig. 3.77 Example of active layer spatial variability. The Boulder Clay CALM grid (100 x 100 m) is located on a flat area close to the italian antarctic station (Mario Zucchelli Station) at around 200 m of elevation. a) active layer thickness (as maximum depth of 0°C isotherm, in cm) on 7[th] January 2002; b) snow thickness recorded on the same date of a). Note the influence of snow accumulation on the active layer thickness distribution. (from Guglielmin, 2006).

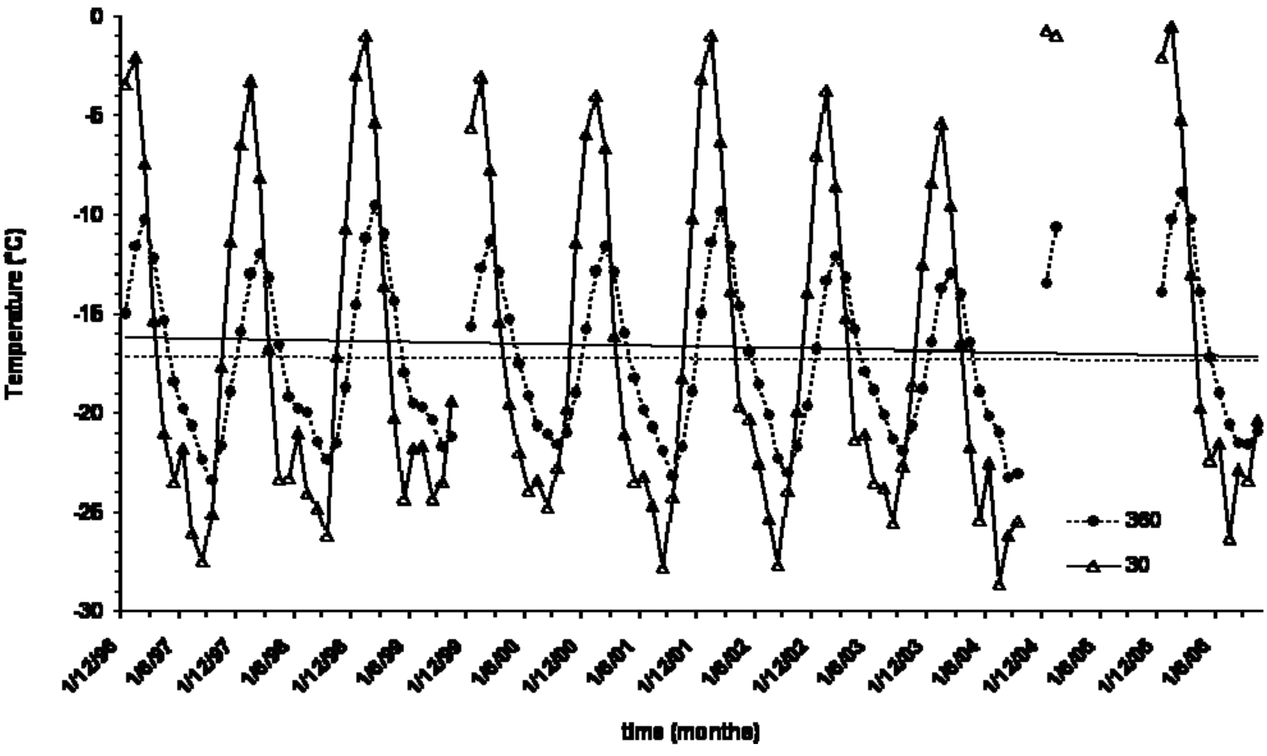


Fig. 3.78. Monthly mean temperatures at the depth of 30 and 360 cm at Boulder Clay since 1996. Note that the depth of 30 cm is very close to the permafrost table. The linear regression line for 30 cm depth (solid line) and for 360 cm (dashed line) are also reported.

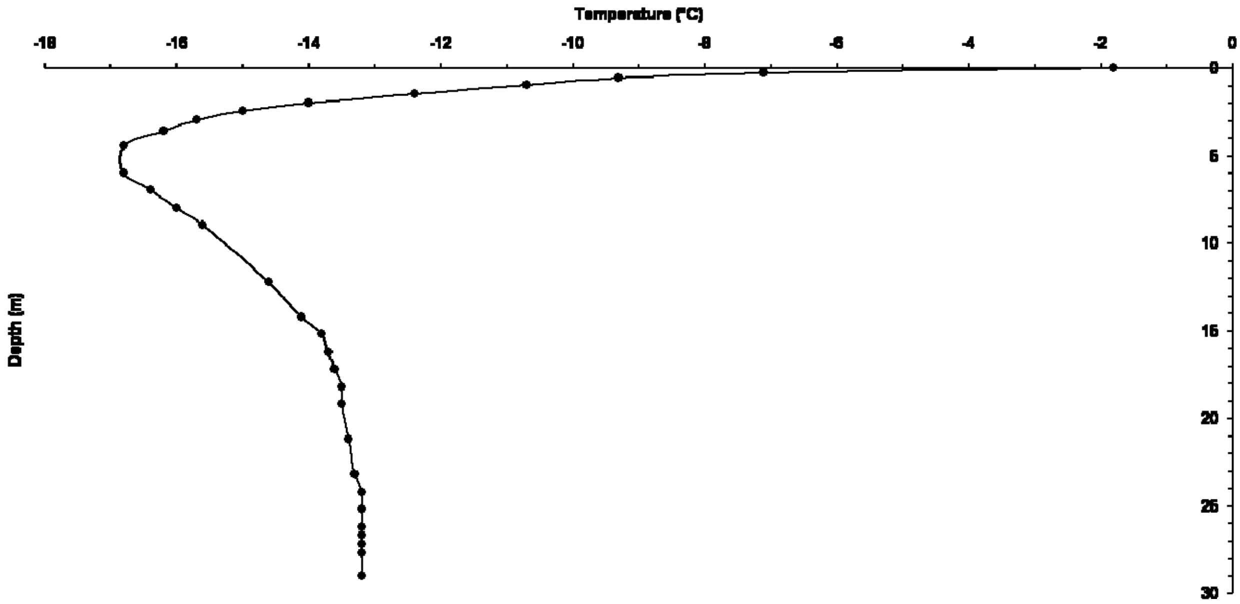


Fig. 3.79. Permafrost profile recorded in the new borehole at Oasi on 7th November 2006. The borehole was drilled through an homogenous granite outcrops on a very gentle slope at 80 m a.s.l.

3.14 **Marine Biology**

3.14.1 Adaptation and evolution

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 in the Arctic, where the sea ice cover is declining at an alarming rate. This trend is affecting the structure and functioning of Arctic marine ecosystems, particularly mammal and bird populations, in a profound manner. In the Antarctic, comparable shrinking of the winter ice cover has occurred only along the western Peninsula tip and the adjoining Scotia Sea. This is a relatively small region but home to the now proverbial whale-krill-diatom food chain where more than 50 % of the Antarctic krill stock was concentrated. This stock now appears to be in serious decline since the seventies accompanied by an increase in salp abundance. However, the extent of the krill decline and the underlying factors are under vigorous debate (Ainley et al 2007 Nicol et al. 2007 Antarctic Science) because of difficulties in unravelling the effects of industrial whaling from those of sea-ice retreat. Nevertheless, if the trend continues, recovery of the great whale populations will be jeopardised. Conversely, suppose the whale stocks had recovered to their former size, would they now be faced with starvation? This thought experiment conveys how little is known about the interplay of physical, chemical and biological processes that once enabled this region to support an extraordinarily large animal biomass.

Investigations carried out in the field and by remote sensing in the past decades indicate that the krill habitat is not as exceptionally productive as initially assumed. Further, productivity over most of the Southern Ocean was found to be even lower despite high nutrient concentrations. This condition, also found in the Equatorial and subarctic Pacific, came to be known as HNLC (high nutrient, low chlorophyll) and three, mutually inclusive factors were considered to be responsible for it: light limitation by deep mixed layers, iron limitation in the open ocean and heavy grazing pressure on phytoplankton stocks. Recently, five mesoscale, *in situ* iron fertilization experiments, carried out in the Pacific and SE 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 SW Atlantic, is maintained by input of iron from land masses and from the sediments by deep mixing and upwelling along the continental margin. Nevertheless, the presence of excess nutrients also in these regions indicates 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 has yet to be adequately explored, particularly because ongoing global change will affect coastal hydrography.

It is now acknowledged that large herbivores (megafauna) condition the terrestrial environment by promoting a vegetation cover conducive to their demands, e.g. grassland instead of forest by elephants. Their removal 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. Whereas 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. The belief that marine phytoplankton productivity is determined primarily by bottom-up driving forces is entrenched in the community, but why marine plankton should be fundamentally different from their 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.

Acquiring a mechanistic understanding of the structure and functioning of the ecosystems surrounding Antarctica is a prerequisite to 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.

In the following I deal with some of the issues raised above in the light of global change, both directly by warming and indirectly by the possible application of mitigation measures – large-scale iron fertilization – to combat global warming. The ongoing changes in the Arctic clearly indicate that current CO_2 levels have already initiated a positive feed-back process that will inexorably lead to a summer ice-free Arctic Ocean within a few decades. In the light of this alarming trend, the current debate on whether to allow large-scale iron fertilization as a mitigation measure is likely to shift its focus to exploring how to maximise its efficiency and minimise harmful side effects. It goes without saying that cutting emissions of greenhouse gases is the first order of the day but we also need to do our utmost to develop techniques to reduce atmospheric CO2 levels which, without human intervention, are expected to persist for thousands of years.

Sea ice ecosystems

The impact of ongoing global warming on the winter extent of Antarctic sea ice has not yet been as dramatic as in the case of the Arctic because of the fundamental differences in topography and hydrography between the two polar regions. Whereas the Antarctic is a closed system fed by cold water upwelling from the deep ocean along the continental margin, and shielded from lower latitudes by net northward transport of surface water along the powerful Antarctic Circumpolar Current (ACC), the Arctic is open to influx of warm surface water from the Atlantic and, to a lesser extent, the Pacific. Sea ice retreat in the Arctic has accordingly been dramatic, proceeding from the Barents and Bering seas to the central Arctic Ocean. In the Antarctic, significant winter sea ice retreat has so far occurred mainly along the northernmost extent of the continent – the Antarctic Peninsula and its oceanward plume in the Scotia Sea – known to be one of the fastest warming regions of the globe. The ice deficit in this region has been compensated by increases in other sectors implying that sea-ice production there has actually increased. This is not surprising given the fact that ice is produced mainly around the continent where winter temperatures are far below freezing and will continue to be so for some time to come. The pack ice is pushed northward from its production sites by katabatic winds from the continent, the strength of which determines how much ice is formed, hence how far north the winter sea ice extends. The air above the continent is not likely to warm for some time because reflected light (albedo) is not retained by CO_2. Hence warming of the air over open water by absorption of infrared radiation will increase the strength of these winds because south to north temperature gradients will be steeper, so any decline in the extent of maximum sea ice will be due to more rapid local melting along its northern boundary than overall ice production.

The Peninsula region is exceptional because it is influenced more by the warming ACC than by the cold continental ice shield. Part of the sea ice feeding the southwest Atlantic sector is formed here, so one can expect that decreasing sea-ice cover in the SW 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 on the underlying water column and benthos has several effects. 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. On the other hand, upward mixing of water that has contacted the sediment surface will bring with it 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 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, as we shall see next, it cannot be the only factor because the krill decline commenced earlier.

Krill stocks in the SW Atlantic sector

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, their total biomass remained only a few percent of that of the former whale population. About 300,000 blue whales were killed within the span of a few decades equivalent to more than 30 million tonnes biomass. The majority of these whales were killed on their feeding grounds in the SW Atlantic in an area of maximum 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 actually 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 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} of only blue whales and their annual food intake. This number is equivalent to biomass of an average phytoplankton bloom or, to take a visual example of more familiar grazers, to 240 cows of 500 kg each 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. The magnitude of primary production required to fulfil their food demands at the trophic transfer rule of thumb (10:1) will be around 300 g C m^{-2} yr^{-1} which is about that estimated for the North Sea, hence does not leave much scope for other grazers such as protozoa and copepods. What percentage of the production was exported from the surface through the mesopelagial and ultimately to the deep-sea benthos, hence also sequestered as carbon, is an interesting but open question.

The above calculations indicate that krill stocks in the whale feeding grounds were close to the carrying capacity of the ecosystem, implying that lowering mortality would lead to starvation of accumulating animals. This would explain why a krill surplus, at least equivalent to the amount annually eaten by whales, was not recorded. On the contrary, there is considerable evidence that krill stocks have declined significantly in subsequent decades. Thus, at their former stock sizes, and given the tendency of krill schools to appear at the very surface and discolour the water, they would be frequently observed from ship deck and this indeed is the impression left behind by the scientists of the Discovery cruises (Hardy 1967). However, despite the increase in the numbers of observers, from cruise ships to research vessels, krill swarms are now rarely seen from ship-deck. Scientific support for this visual estimate comes from a thorough statistical assessment of all net catches carried out for different sectors of the Southern Ocean. The analysis indicates an 80% decline in krill stocks of the SW Atlantic accompanied by an increase in salp populations (Atkinson et al. 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 this. Unfortunately, a comparison with phytoplankton stocks recorded during the Discovery era is not possible because the methods of that time soon became obsolete. Nevertheless, the impression gained by the Discovery scientists is one of large diatom stocks: "... the extremely rich production, which will probably be found to exceed that of any other large area in the world …" (Hart 1934). It should be mentioned that these scientists were familiar with North Sea phytoplankton which today has much higher biomass levels than those recorded for the Scotia Sea since the past decades. So it is likely that phytoplankton production has indeed decreased with that of the krill stocks, a conclusion supported by the increase in the salp population. In contrast to krill, that are equipped to deal with the characteristically spiny, heavily silicified diatoms of the Southern Ocean, salp feeding is adapted to the lower concentrations typical of the iron-limited microbial food webs. Therefore, their encroachment into the former krill habitat is an indication of declining phytoplankton, in particular diatom, stocks.

A decline in phytoplankton concentrations can only be explained by a corresponding decline in the iron supply. As mentioned above, there is reason to believe that the reduction in ice formation has resulted in a decrease in iron input from the shelf slopes of the Western Peninsula. On the other hand, the retreat of the glaciers should increase run-off from the land and hence iron input along the coasts of the convoluted Peninsula and adjoining islands. This has not been quantified but comparisons of the chlorophyll concentrations recorded by the CZCS satellite of the eighties 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 (Greg and Conkright). In contrast, a 50% increase was found 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. Whatever the mechanism, a reduction in phytoplankton biomass can only be explained by a corresponding reduction in iron supply.

An alternative but not mutually exclusive explanation for the phytoplankton decline can be the result of a decrease in the rate of recycling of 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 nanophytoplankton which are in turn eaten by their 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 this microbial community is characteristic of the iron-limited HNLC area where chlorophyll concentrations remain below 0.5 mg Chl m^{-2} throughout the year. These regions support surprisingly high zooplankton biomass, comprising slow growing copepods and fast-growing salps, throughout the year suggesting that they are an integral part of the recycling system which also regenerates iron in addition to ammonia (Barbeau et al).

Local increases above this level are invariably due to accumulation of diatoms and Phaeocystis colonies and will be due to input of new iron, whether from above or from below. The fate of these diatom blooms is under debate: whether they are consumed and their nutrients recycled in the surface or sub-surface layer or whether a significant portion sinks out to greater depths or the sea floor. The latter fate is of particular interest in the light of 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 equally effective as in the microbial food web but at a much grander scale. Predation by the whales will have contributed significantly to the recycling pool for the following reasons. With the exception of pregnant and lactating females, whales convert krill protein into blubber (hydrocarbons) implying that energy is retained but the nutrients, including iron, are returned to the system. Whale faeces are liquid and will rise to the surface after defecation. If iron released by whales is bound by ligands in the faeces it will be retained in the surface layer and increase the efficacy of recycling. Krill also have a high (50%) lipid content, and large amounts of iron released by krill faeces have been measured (ref). It follows that the exceptionally productive ecosystem characterised by the food chain of the giants was maintained by recycling of iron by krill and whale feeding. An alternative, but mutually inclusive hypothesis in which large whale stocks promoted large stocks of their prey (krill) by dispersing them over a larger area has been recently suggested.

New generation of iron fertilization experiments

The hypothetical "iron-manuring", top-down control mechanism is commensurate with those described for terrestrial and lake ecosystems. This hypothesis can 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 alarming rates of krill decline and southward encroaching global warming there is a pressing need to develop an integrated understanding of how this ecosystem functioned in the recent past but also in the glacial ocean in order to predict future changes to the pelagic and underlying benthic ecosystems all around Antarctica. In situ iron fertilization experiments represent a powerful new methodology to test this series of hypotheses because they enable the study of interactions within ecosystems with their 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 to this ecosystem. Preliminary surveys of the deep-sea benthos of the Peninsula region have shown the presence of communities with high biomass and species diversity (Brandt et al. Nature 2007) but their areal extent is not known. Since fertilization will be carried out offshore, shelf and coastal benthos would not be affected.

An added incentive to carrying out more such experiments is the ideal training ground they offer for the kind of international, interdisciplinary research demanded by earth system science 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 but also more reliable data to parameterise current and new, coupled ecological-biogeochemical, ocean-circulation models in order to assess the feasibility of using the Southern Ocean as a sink for anthropogenic CO_2. This is one of the measures currently under debate to combat the alarming rate of global warming. There is legitimate 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). International scientific and legal bodies will have to decide how best to deal with such issues, but they will require reliable models based on real-ecosystem experiments to ascertain the long-term effects of climate management schemes.

3.14.2 **ENSO links and teleconnections to vertebrate life histories and population – Forcada and trathan dynamics**

Introduction

ENSO (El Niño-Southern Oscillation) is perhaps the most important large-scale climatic impact on the world's oceans and marine ecosystems. Its impacts are local, through direct atmospheric and oceanographic effects, or remote, through changes in the atmosphere, ocean and ice systems that happen elsewhere. Remote impacts often show coherent patterns of spatial and temporal variability in the repercussions on marine ecosystems.

The biological impacts of ENSO are difficult to distinguish and quantify owing to the complexity of marine ecosystems and the effects of concurrent or past anthropic impacts. The different scales of interaction between physical and biological processes, and within biological systems, generate complex feedback effects. For populations of, and individual, Antarctic seabirds and marine mammals (predators) most responses are evident as changes in phenology, which is the annual timing of life history events in populations (migration, arrival to and departure from breeding grounds, and reproductive events). However, these organisms also dependent upon an extensive set of trophic links within the marine food web and indirect effects occur through prey, predator and competitive interactions, which have repercussions for populations and communities. Effects on individuals are often specific to sex and life history stage and may affect subsequent changes in population frequency and compensation-depensation, species interactions, and feedbacks in foodwebs (Stenseth et al. 2002). These effects also depend upon where animals breed and spend the winter, and the level of regional climate change impact. Ultimately, understanding how marine predators respond to their highly seasonal environment requires that we also understand how the wider food web responds, which trophic interactions are key, and how ecosystems fluctuate with physical processes.

Propagation of ENSO variability

ENSO effects and variability in the tropical Pacific propagate to southern high latitudes through atmospheric teleconnection and oceanic processes (Kwok and Comiso 2002, Liu et al. 2002, White et al. 2002, Turner 2004, Meredith et al. 2004). The Southern Ocean, a major component within the global ocean and climate system, plays an important role in the propagation of the ENSO signals. It unites the Atlantic, Indian and Pacific Oceans, connecting low tropical latitudes with high polar latitudes. Southern Ocean atmosphere, ocean and sea ice systems are strongly coupled showing marked variation, but also covariation, between regions and on time scales that range from years to decades. Variation is largely driven by ENSO impacts on the atmospheric circulation patterns across the southwest Pacific, and generates sea surface temperature (SST) anomalies in the Pacific sector of the Southern Ocean (Li 2000, White et al. 2002, Turner 2004). Anomalies in atmospheric pressure, SST, surface tempearture and sea ice slowly propagate westward across the South Pacific sector into the Atlantic at a rate consistent with oceanographic transport, and in association with the Antarctic Circumpolar Current (ACC) (White et al 2002, Venegas 2003). This is a phenomenon originally known as Antarctic Circumpolar Wave (ACW) (White and Peterson 1996), which has long term variation (Connolley 2002).

ENSO-related variation in the South Pacific and the Atlantic has been quasi-cyclic in the last decades, showing a periodicity of approximately between 4 and 6 years, and high variation in anomaly intensity and duration on decadal and longer time scales (White and Peterson 1996, Trathan and Murphy 2002, Carleton 2003, Turner 2004). Changes in the ENSO patterns and in particular in the development of El Niños since the late 1970s have been emphasized by Trenberth and Hoar (1996). They have related the more frequent occurrence of El Niños and the less frequent occurrence of La Niñas to decadal changes in climate throughout the Pacific. These have created persistently warm background conditions (Fedorov and Philander 2000), which in turn have been associated with oceanic regime shifts causing major changes in the biological environment (McGowan et al. 1998, Stenseth et al. 2002, Chávez et al. 2003).

SST anomalies that propagate across the South Pacific and into the South Atlantic sector of the Southern Ocean are correlated with the changing phases of ENSO in the equatorial region of the Pacific. Sea surface temperatures and sea-ice extent in both the Atlantic and the Indian Oceans are related to the Pacific Ocean, but coherent correlations decrease between the Indian Ocean and the western Pacific Ocean (Trathan et al. 2007). After the southeastern Pacific, the Southwest Atlantic sector is the region where ecosystem fluctuation associated with ENSO is stronger and best known. Ecosystem-wide effects have been observed in the Scotia Sea, between the Antarctic Peninsula and adjacent islands, South Georgia, and the other islands of the Scotia Arc. Positive anomalies in the ENSO region occur approximately 2-3 years prior to positive anomalies close to Southwest Atlantic sector (Trathan and Murphy 2002, Forcada et al 2005, Trathan et al. 2006, Murphy et al. 2007). This is consistent with the ACW (White and Peterson 1996), but may also result from spatial differences in the geographical relationships of SST variation across the southern Pacific fluctuating in phase with ENSO (Park et al. 2004), or some combination of these processes.

Southern Ocean ecosystems fluctuation and consequences for predators

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 this 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 a commercial fishery. The relative density of krill across the region has shown a significant negative trend over the last

30 years, correlated positively with sea ice contraction and negatively with sea ice season duration (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 characterised 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. Whereas 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, Murphy et al. 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 congeneric gentoo penguins, but their population trends are positive. Similar results on population trends at the South Orkneys suggest the loss of buffering against the changing sea ice environment of the more abundant and ice-dependent chinstrap and Adélie penguins, and positive population consequences through habitat improvement for the les ice-related gentoo penguin (Forcada et al. 2006).

Other krill dependent predators have shown comparable responses to the propagation of ENSO variability. Gentoo penguins and Antarctic fur seals breeding 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) have shown a negatively correlated breeding output with SST anomalies. Low chick and pup production and survival in penguins and fur seals are a consequence of reduce food supply, which comprises krill and krill-dependent fish species, such as mackerel icefish. Reductions in food supply associated to ENSO variability have been recently explained by Murphy et al. (2007).

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 (SOI, and an opposite trend in ENSO) (Barbraud and Weimerskirsch 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 Weimerskirsch 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 (Weimerskirsch 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 periodicity of 3-5 years, consistent with the periodicity in sea-ice anomalies and the Southern Oscillation Index (and hence ENSO). The observed cyclical patterns also indicated significant change since 1980, consistent with a regime shift. Although these populations have remained stable to date.

In the Pacific Ocean and in particular the Ross Sea sector, trends in surface temperatures are less apparent in relation with ENSO, although there has been a decrease in SIO. 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, 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 as 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 polynia, providing improved penguin habitat. Similar to the emperor penguins in Terre Adélie, Indian Ocean, population trends of Weddell seals at McMurdo Sound, Ross Sea, declined in the early 1970s, to become later on stable (Ainley et al. 2005). This suggests that ENSO impacts have not been so obvious as in the Atlantic Ocean and Indian Ocean sectors of the Southern Ocean.

3.15 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 became 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 green-house 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 determined a remarkable 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 sewages, are inevitable. Another serious anthropogenic impact, especially in the subantarctic islands (where climatic and environmental conditions are less extreme) is the introduction of alien organisms. 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 o 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 major 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).

Local-scale pollution

The International Geophysical Year (1958-1959) 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 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 meters of scientific stations (UNEP, 2002).

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 startling 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 costal marine environment through wastewater, 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 program 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, mollusks, 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 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 became one of the most important places for research on bacteria that thrive on ice or in subglacial lakes. One of the main challenge 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 which 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 (3400 and 3600 m) ice borehole at Vostok.

Global-scale contaminants

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 generally cause local disturbance of the Antarctic environment. As pesticides have neither been produced nor applied in the continent, the discovery of Dichlorodiphenyltrichloroethane (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, Hexachlobenzene (HCB), Hexachlorocycloexanes (HCHs), aldrin, dieldrin, chlordane, endrin, heptachlor and other POPs have 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). Although long-range transported contaminants in the Antarctic environment and biota have lower concentrations then in the rest of the Southern Hemisphere, they show similar patterns (UNEP 1996). 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 gasoline, 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 system 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 benthos invertebrates (Goerke et al., 2004).

Concern about the future impact of human activity

New classes of global pollutants are emerging, such as perfluorinated compounds (PFCs) which 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 yet no significant trend in meteorological temperature, a loss of ice sheets such as that in the Antarctic Peninsula (seven ice sheets have disappeared in the last 50 years) could have dramatic effects on atmospheric precipitation (i.e., the deposition of contaminants) and the biogeochemical cycle of 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 which 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. Our field evidence of enhanced Hg accumulation in soil and cryptogamic organisms from terrestrial ecosystems facing the Terra Nova Bay costal 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.

The Protocol on Environmental Protection to the Antarctic Treaty provides strict guidelines for the protection of the Antarctic environment and its value to scientific research. Although the rigorous application of the Protocol will help minimize the local impact of tourists and scientists, it is inadequate to reduce or prevent large-scale environmental contamination by metals and POPs from other continents. Conservation measures should focus on the following:

- 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;
- developing continental-scale monitoring programs in order to asses the responses of terrestrial and marine ecosystems to climate changes and anthropogenic activity (local and remote);
- promoting international agreements and the transfer of financial aid and technologies from rich countries to developing countries in the Southern Hemisphere in order to address global environmental threats.

Chapter 4

The next 100 years

Coordinating Lead Author:
John Turner

Lead Authors:
Tom Bracegirdle, John Turner, Pete Convey, Ad Huiskes, Frenot, Dana Bergstrom, Steve Chown, Dominic Hodgson, Anthony Lenton, Bockheim, Amanda Lynch, Howard Roscoe, Greg Bodeker, Robert Bindscadler, Victor Smetacek, Guido di Prisco, C Verde, Lloyd Peck, David Barnes, Julian Gutt, Siobhan O'Farrell, Hughes Goosse, Daniel Smale, James Bockheim, Anna Jones

Contributing Authors:
N Metzl, David Vaughan, Hans Pörtner, Jaume Forcada

Review Editors:

4.1 Introduction

Determining how the environment of the Antarctic will evolve over the next century presents many challenges, yet it is a crucial question that has implications for many areas of science, as well as for 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 only spatially and temporally well separated observations of marine biota.

The evolution of the Antarctic climate over the next 100 years can only be predicted using coupled atmosphere-ocean-ice models. These have many problems in correctly simulating the observed changes that have taken place, so there is still a degree of uncertainty about the projections, particularly on the regional scale. 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, since it is very sensitive to changes in atmospheric and oceanic conditions. The IPCC report took the mean of the 20 models that it employed, without regard to how individual models performed globally or regionally. The predictions for temperature and precipitation were used to estimate how much sea level would rise under various greenhouse gas emission scenarios. With a quantity such as near-surface temperature it is possible to use the predictions from the various models to derive various estimates of how the quantity may change over land areas, the ocean and for various regions of the Earth.

But an extremely important question is how will the Antarctic ice sheet change over the coming decades? The models can give estimates of how the precipitation onto the continent might change, but ice sheet models currently cannot help us regarding possible dynamical changes that might occur in the flow of the ice streams.

Predicting how Antarctic terrestrial and marine biota might change over the coming decades is also a major challenge. Some experiments have been carried out into the survival of biota when subject to thermal stress, but conditions in the Antarctic involve many complex feedbacks and interactions that cannot be replicated.

In this chapter we consider how the physical environment of the Antarctic might change over the next century. The key factor in driving future change is how concentrations of greenhouse gases will increase in the future, and here have assumed a doubling of greenhouse gas concentrations by 2100. This is a frequently used scenario, and is in line with the increases seen in the early years of the Twenty first Century.

4.2 Climate Change

4.2.1 IPCC scenarios

4.2.1.1 Introduction

The degree to which the climate of the Earth will change over the next century is 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 centre of population, but greenhouse gases are well mixed and fairly uniformly mixed across the Earth. Whatever happens it will take a long time for the levels of GHGs to drop. For instance, even if anthropogenic emissions of CO2 were halted now it may take thousands of years for CO2 concentrations to naturally return to pre-industrial amounts (*Archer*, 2005).

Future levels of greenhouse gas emissions will be determined by many complex factors, such as technological changes and regionally varying social and economic development. The future evolution of GHGs and aerosols is therefore very uncertain. Nevertheless, in order to have emission levels for runs of climate models through the Twenty First Century it is necessary to have a range of possible scenarios that can be prescribed. The IPCC developed emission scenarios in 1990 and 1992 that were used for the model runs that contributed towards the first two assessment reports. Following a review in 1995, it was decided to develop a new series of scenarios in light of the greater knowledge that was available on the contribution of different forms of energy to gas emissions and the rapid industrialisation that was taking place in some developing countries. A new set of scenarios was therefore developed and published in 2000 (Nakicenovic *et al*., 2000).

4.2.1.2 The IPCC greenhouse gas and aerosol emission scenarios

The scenarios developed in 2000 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 described in detail in Nakicenovic et al. (2000) (*Nakicenovic, N. et al (2000). Special Report on Emissions Scenarios: A Special Report of Working Group III of the Intergovernmental Panel on*

Climate Change, Cambridge University Press, Cambridge, U.K., 599 pp. Available online at: http://www.grida.no/climate/ipcc/emission/index.htm) are:

- A1: a future 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 very 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 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. This resulted in a total of 40 scenarios, each of which was considered equally valid with no assigned probability. 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). Scenario A1B (often referred to as SRESA1B) is the most frequently used scenario since climate model projections of the 21st century using the A1B 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.

Figure 4.1 illustrates how total global annual CO2 emissions from all sources would change during the Twenty First Century for the four emission families (A1, A2, B1 and B2) and the six scenario groups (A1F1, A1T, A1B in Figure 4.1a, and A2, B1 and B2). The figure shows that the greatest emission would be with the A1F1 scenario group, which includes high coal and high oil and gas scenarios. With this group the CO2 emissions by the end of the end of the century would have risen from the 1990 value of about 8 GT/year to be in the range 28-37 GT/year. The frequently used A1B scenario has emissions increasing to a peak around the middle of the century and then decreasing to about 14 GT/year by 2100.

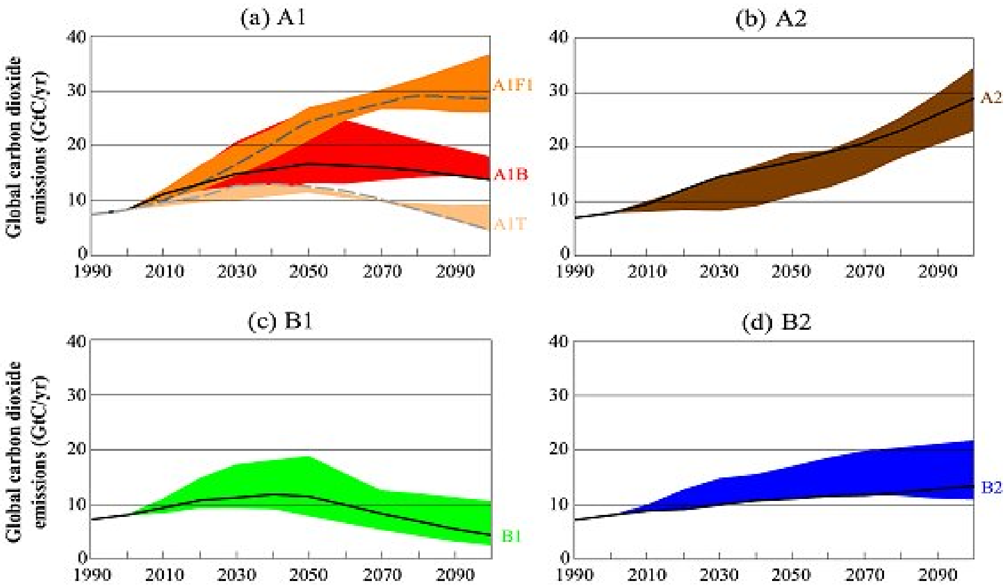


Fig. 4.1. Total global annual CO_2 emissions from all sources (energy, industry, and land-use change) from 1990 to 2100 (in gigatonnes of carbon (GtC/yr) for the families and six scenario groups. The 40 SRES scenarios are presented by the four families (A1, A2, B1, and B2) and six scenario groups: the fossil-intensive A1FI (comprising the high-coal and high-oil-and-gas scenarios), the predominantly non-fossil fuel A1T, the balanced A1B in Figure 4.1a; A2 in Figure 4.1b; B1 in Figure 4.1c, and B2 in Figure 4.1d. Each colored emission band shows the range of harmonized and non-harmonized scenarios within each group. For each of the six scenario groups an illustrative scenario is provided, including the four illustrative marker scenarios (A1, A2, B1, B2, solid lines) and two illustrative scenarios for A1FI and A1T (dashed lines).

With the other three scenario groups, A2 anticipates a steady increase of CO2 emissions during the century, with values peaking around 30 GT/year in 2100. B1 suggests increasing emissions until around 2050, and then a decrease to about 4 GT/year, which is less than the 1990 level. B2 has a modest increase throughout the century to about 12 GT/year by 2100.

The Arctic Climate Impact Assessment (ACIA) made extensive use of model projections from the IPCC Third Assessment Report, with the focus being on output from the runs with the A2 and B2 emission scenarios.

4.2.1.3 **The emission scenarios in the AR4 models**

The AR4 climate model simulations were run using prescribed concentrations of GHGs (CO_2, CH_4 and N_2O) according to the scenarios of Nakicenovic et al. (2000). As well the scenarios for the future two other emissions experiments were conducted as part of the AR4. A "climate of the 20th century" (20C3M) experiment, which was forced by observed GHG concentrations from the mid 19th century to the end of the 20th century, and a "pre-industrial control" (PICNTRL) experiment for which GHG and aerosol concentrations were kept at constant pre-industrial levels. For the SRES A1B future scenario, concentrations of CO_2 are projected to rise from the current concentration of around 380 ppm (Figure 4.2) to reach 720 ppm in the year 2100. For both the 20C3M and scenario runs the sulfate aerosol concentrations were calculated from prescribed sulfur dioxide (SO_2) emissions, but vary between models due to different methods of calculation. Both sets of runs also include inter-model differences of forcing by stratospheric ozone and volcanic aerosols, which are important for the Southern Hemisphere circulation due their role in long-term changes of the southern annular mode (SAM) (*Shindell and Schmidt*, 2004; *Miller, et al.*, 2006). For the SRES A1B scenario, most models were run with a gradual recovery of stratospheric ozone to pre-industrial levels over the 21st century, but some did not include any stratospheric ozone forcing (see *Miller, et al.*, 2006).

In this volume we will present results from models run through the Twenty First Century with the SRESA1B emission scenario, in other words assuming a doubling of CO2 and other gases by 2100. In terms of temperature increase, the response to rises in greenhouse gases in fairly linear, so that a quadrupling of gases would give temperature increase that were double those presented here.

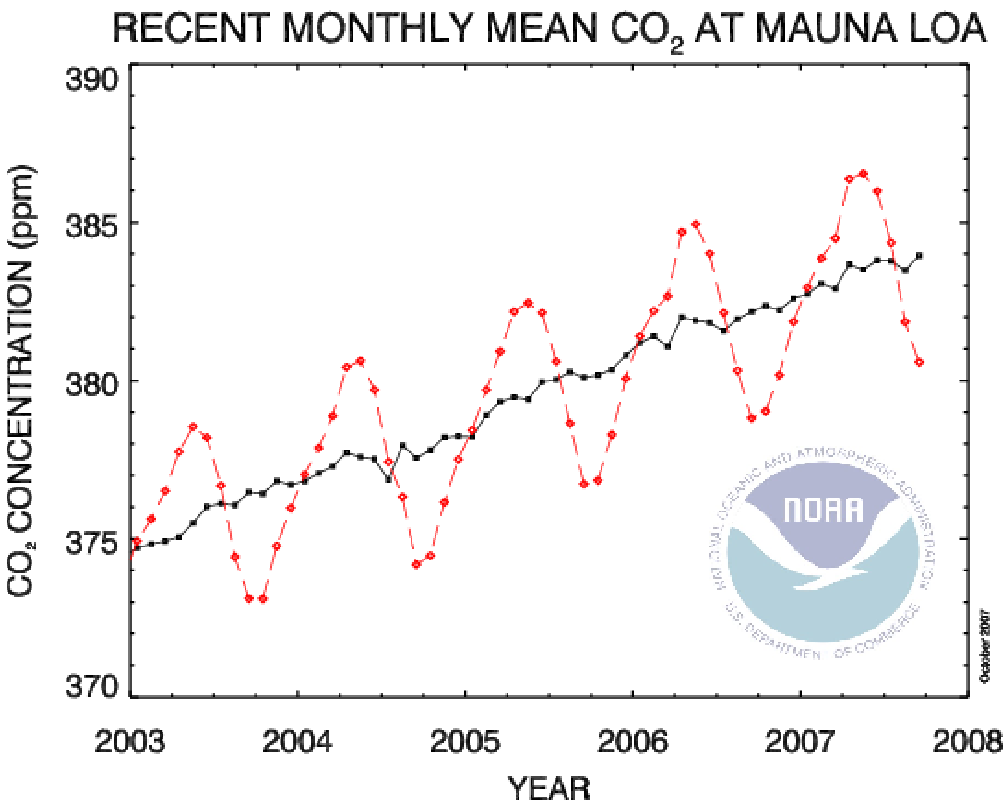


Figure 4.2. "The graph shows recent monthly mean carbon dioxide measured at Mauna Loa Observatory, Hawaii. The last four complete years of the Mauna Loa CO2 record plus the current year are shown. Data are reported as a dry mole fraction defined as the number of molecules of carbon dioxide divided by the number of molecules of dry air multiplied by one million (ppm). Click for a graph of the full Mauna Loa record. The last year of data is still preliminary, pending recalibrations of reference gases and other quality control checks. The dashed red line with diamond

symbols represents the monthly mean values, centered on the middle of each month. The black line with the square symbols represents the same, after correction for the average seasonal cycle. The latter is determined as a moving average of ten adjacent seasonal cycles centered on the month to be corrected, except for the first and last five years of the record, where the seasonal cycle has been averaged over the first and last ten years, respectively. The Mauna Loa data are being obtained at an altitude of 3400 m in the northern subtropics, and may not be the same as the globally averaged CO2 concentration at the surface."

4.2.2 Climate models

Antarctica has historically been rather peripheral to global climate modelling, since few people live there and computing resources were more limited than now. However, with the current concern about the future impacts of climate change on humans, issues such as sea level rise have initiated a greater effort in modelling high latitude climate processes accurately. Currently effort is moving towards developing climate models 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.

With the growing concern over human-induced climate change, many countries are working to contribute to our understanding of climate change. This is done through various types of research, of which the development and interpretation of climate models is a major component. The results from 24 climate models contributed by many modelling centres around the world were compiled into a central database for the Intergovernmental Panel on Climate Change (IPCC) Assessment Report Four (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 the strengths and weaknesses.

The CMIP3 climate models are coupled ocean/atmosphere/sea-ice/land-surface models. In other words each model is made up of sub models that simulate the atmosphere, ocean, sea ice and 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.

Atmosphere models

Models of the atmosphere are founded on the fundamental laws of motion, thermodynamics and chemistry. These equations are used to step forward in time from some initial state. The result is a representation of the evolution of the atmosphere discretized 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 some synoptic-scale phenomena, such as extra-tropical cyclones. 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 and not explicity included). Many of the parameterizations are developed and optimised for mid and low latitudes and therefore are not always appropriate for the polar regions.

One example is parameterizations of the atmospheric boundary layer, which perform poorly in the polar regions where a very stable boundary layer is often present over snow/ice covered surfaces. Parameterizations that are based on observations of more weakly stable boundary layers at lower latitudes are often not suitable for such conditions. As a result fluxes of momentum, heat and water vapour are often too small in the very stable stratified conditions.

Another parameterization challenge in the polar regions is clear air precipitation. This is not explicitly included in CMIP3 climate models.

Climate models do not include parameterizations for polar stratospheric cloud (PSC), which are the thin, tenuous clouds found in the startosphere. PSCs are important for the radiation balance and temperature of the mid/upper troposphere to the lower stratosphere.

Not all of the CMIP3 models include the effects of stratospheric ozone. This has a significant impact on their representation of Southern Hemisphere wind patterns. Stratospheric ozone has been found to affect the strength of the circumpolar westerlies in climate models. These circumpolar westerly winds blow around the mid-latitude regions just north of the Antarctic continent. At least part of an increase of the circumpolar westerlies that has been observed in the late 20th century has been attributed to concurrent decreases of stratospheric ozone. Miller *et al.* (2006) found that the evolution of the circumpolar westerlies as simulated by the CMIP3 models depends significantly on whether or not the models include stratospheric ozone changes.

Ocean models

As with the atmosphere, the ocean model uses a set of equations to represent the evolution in discrete time and space intervals. Baroclinic eddies in the ocean are smaller than in the atmosphere, therefore the resolution of the ocean models is higher. The grid spacing for the CMIP3 ocean models is around 100 km in the horizontal. The steep ocean ridges also require a high resolution to be represented realistically.

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. 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 retrieving observations in such conditions.

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 CMIP3 sea ice models is in their treatment of rheology (deformation and flow). The rheology used ranges from simple ocean drift models to more advanced Elastic Viscus Plastic (EVP) schemes. For more details see p606 of the AR4 WG1 report.

Terrestrial ice and snow

Most global climate models (including the CMIP3 models) represent the large ice sheets that cover much of Antarctica non-interactively. Their presence is implicit in the land surface orography and surface albedo. There is also not generally an explicit representation of glaciers.

Some studies have 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 moving ice sheets and glaciers. The dynamical discharge into the ocean due to rapid changes in the ice sheet properties is a large uncertainty at present, but an important problem to address due to its potential impact on sea level rise (Rignot *et al.*, 2005).

Regional climate models

Regional climate models are used because they can 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. This can give improved regional detail of precipitation ({berg}, 2005) and winds (van Lipzig *et al.*, 2004). Due to the sparse observation network over Antarctica it is difficult to determine whether or not regional climate models improve Antarctic-wide precipitation totals.

Model evaluation

As yet there is no widely accepted way to evaluate the performance of a climate model. The root of the problem is that long-term projections cannot be checked against observations. Probably the most common way to assess the performance of climate models is to compare their output with observations (refs). The main issue with this approach is that 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. However, at high latitudes it has been found that biases in surface temperature are significantly correlated with projected temperature change under future scenarios (Räisänen, 2007). This may be caused by, at least in part, sea ice biases, which have a strong impact on projected regional changes in the sea ice zone.

Sea ice in the CMIP3 models

All models produce a seasonal cycle 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 measure based on the pointwise RMS difference from climatology produces more usable rankings, in which MRI, CSIRO, HADGEM and MIROC_hires 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. Of the models, most use a VP or EVP rheology, CSIRO uses cavitating fluid, and HADCM3 and MRI implement "ocean drift". Only INM has no ice advection. It is perhaps notable that the best performing model is then MRI, with the most primitive "rheology". However, the MRI model is flux-corrected globally, and this is likely to strongly affect the seaice simulation. The next best, CSIRO, uses the relatively simple cavitating fluid rheology. This illustrates the fact (Connolley, 2006) that many aspects of the model simulation besides sea ice model quality goes into making up the simulation of the sea ice. It provides no support for the need for a sophisticated sea ice dynamics scheme, although clearly if all else is equal a more sophisticated and physically plausible scheme will be preferable. Parkinson et al. 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 other; we have only examined the SH in this paper.

The model average displays significantly higher skill (0.42) than any of the individual models, presumably due to cancellation of errors; Parkinson et al. Also noted the qualitative virtues of the model average.

Temperature

Station observations (Turner et al, 2005) show a maximum temperature trend since 1958 on the west side of the Antarctic peninsula, with smaller and generally non-significant changes around East Antarctica. 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 to evaluate the temperature trends. The model average, for JJA of 1960-1999, qualitatively reproduces this observed pattern (figure 4.3).

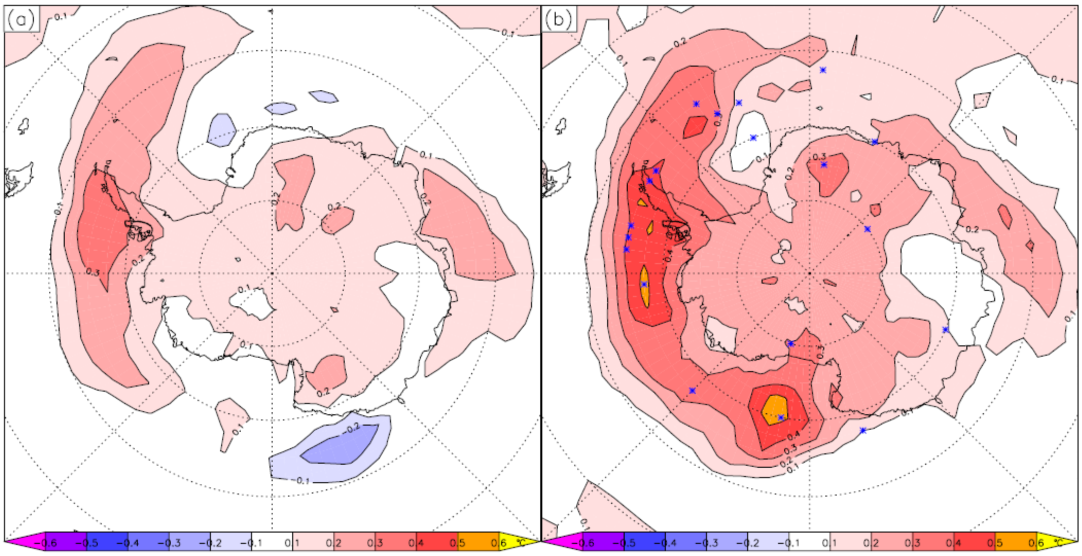


Figure 4.3. Temperature trends in ^{o} C/decade from 1960–2000 for winter (JJA). (a) Unweighted average of 19 models. (b) Weighted average. Also plotted on Figure 4.3b are the locations of the maximum trends from the individual models.

The maximum trend is, correctly, in the region of the west side of the Antarctic Peninsula. Figure 4.3 (a) is the simple average; (b) is the skill-weighted average using the weights. The trend at (65S, 70E) is 0.38 °C/decade (unweighted) or 0.45 (weighted), both of which are smaller than the observed value at Faraday of approximately 1 °C/decade, though the weighted average performs somewhat better. However the observed winter temperature at Faraday is highly variable and the trend depends strongly on the exact start date chosen. In the weighted average 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.

Individual models show great scatter in their trends; Figure 4.4. Only 4 of the individual models have their maximum trends in the "correct" place, and the average maximum trend is 1.1 °C/decade, three times higher than the average, but in rather better accord with observations. Some models (e.g. giss_e_h) show small trends almost everywhere; however most models have large trends somewhere; 9 of the 20 models have a greatest absolute trend that is negative. All the large model trends are over the seaice rather than the continent, and are closely related to seaice changes. In this they are behaving realistically, in that the observed winter trends around the Peninsula are believed to be reinforced by seaice feedbacks (King, 1994).

Fig 4.4 shows the locations of the maximum trends from the models. Four models position it west of the Peninsula; a further two to the east; four in the Weddell Sea; three in the seas around East Antartica; three over the continent itself (although the absolute magnitudes of these trends are small); one on the Ross ice shelf and two in the Ross sea.

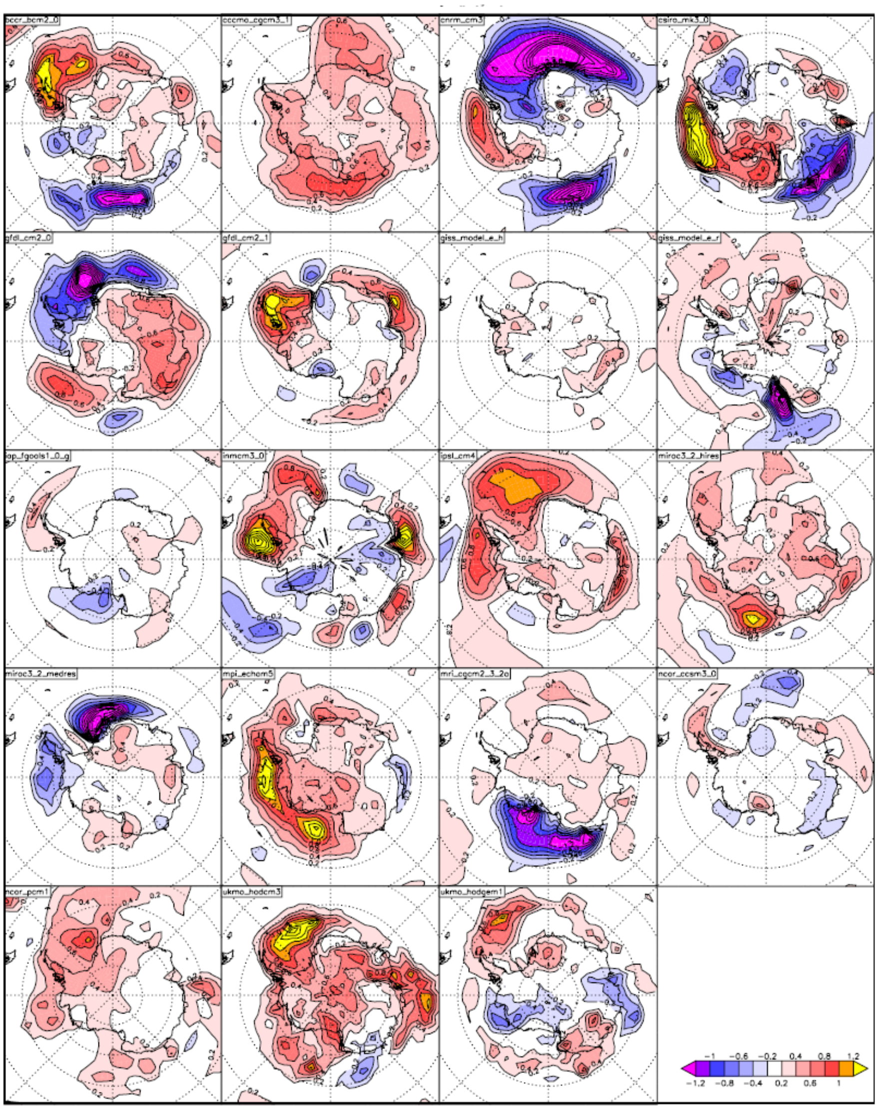


Figure 4.4. Temperature trends in ^{o} C/decade from 1960–2000 for winter (JJA) for individual models.

Surface Mass Balance

Estimates of Antarctic surface mass balance (SMB) vary (Uotila et al); we shall use a central value of 167 mm/yr from Vaughan et al with a spread of 30 mm/yr which recognises the considerable spread in estimates from observations and models, and also allows for inter annual variation. Typically, models indicate that this SMB is made up of 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); the GISS models (despite having a large value for sublimation) and MRI overestimate by about 100 mm/yr, but for different reasons: GISS have a "central desert" area but it is too small; whereas MRI does not simulate the very low values of SMB in the interior. Only MIROC_medres (116) and HADGEM (131) substantially underestimate SMB. Of those models that do well on overall totals, two (BCCR and CNRM) nonetheless produce SMB simulations that, on inspection of their maps, are implausible: they fail to produce large (> 500 mm/yr) SMB on and around the coast of East Antarctica.

4.2.2 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 the 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.

The role of modes of circulation variability

Climate model projections of circulation change in high southern latitudes must be evaluated in the context of simulation correspondence to observed changes in the 20th century. Mean sea level pressure changes associated with the long term variability of the Southern Hemisphere circulation 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 20th and 21st centuries.

Given the large inter-annual variability of the high southern latitudes, secular trends must be put in the context of manifestations to changes in modes of variability that are dominant in the region: El Niño Southern Oscillation (ENSO) and the Southern Annular Mode (SAM) (Turner, 2004; Sen Gupta and England 2007). The climate response to increases in concentrations of greenhouse gases has been found to project strongly onto the leading modes of present-day variability, although perhaps with changed spatial characteristics (Brandefelt and Källén, 2004). Of less significance, due largely to its contested nature (Park 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 demonstrates an association with negative sea ice cover anomalies in the Ross and Amundsen Seas and positive sea ice anomalies in the Bellingshausen and Weddell Seas (Kwok and Comiso, 2002). The SAM index has demonstrated a positive trends over recent decades which is a reflection of the trends in the zonally averaged mid-latitude westerlies. These increases in the SAM index are associated with strong warming over 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 Peninsula warming and reduced sea ice in the region (Liu et al., 2004b). 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).

The AR4 reports on progress in simulating ENSO variability has led to significant improvements in the representation of 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 the structure of the thermocline (Cai et al., 2003; Davey et al., 2002). 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 subject to substantial uncertainty, with no coupled model consensus on the likelihood of a relationship between more frequent El Niño conditions and increasing greenhouse gas concentrations (van Oldenborgh et al. 2005; Collins et al. 2005, Wang, 2007).

Realizations of climate system models submitted to the AR4 simulate the SAM with a high degree of verisimilitude (e.g., Miller et al., 2006), generally with spatial correlations greater than 0.95 (Randall et al. 2007). The SAM signature in the surface warm anomaly over the Antarctic Peninsula is also captured by some models (e.g., Delworth et al., 2006) reflecting advection of the climatological temperature distribution. Other features, including the zonal structure and the temporal signal, exhibit large variance across realizations even in a single model (Miller et al., 2006; Raphael and Holland, 2006). Hence, the true 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).

Projected changes in modes of circulation variability

In the range of studies reported by the AR4, it has been demonstrated that the ENSO response of climate system models in the 21st century is highly model dependent (see figure 4.5 (Meehl et al. 2007). This outcome has changed little in the most recent analyses (e.g. Yeh et al. 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 21st century, although some of the models that simulated 20th century ENSO variability well do indicate 21st 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).

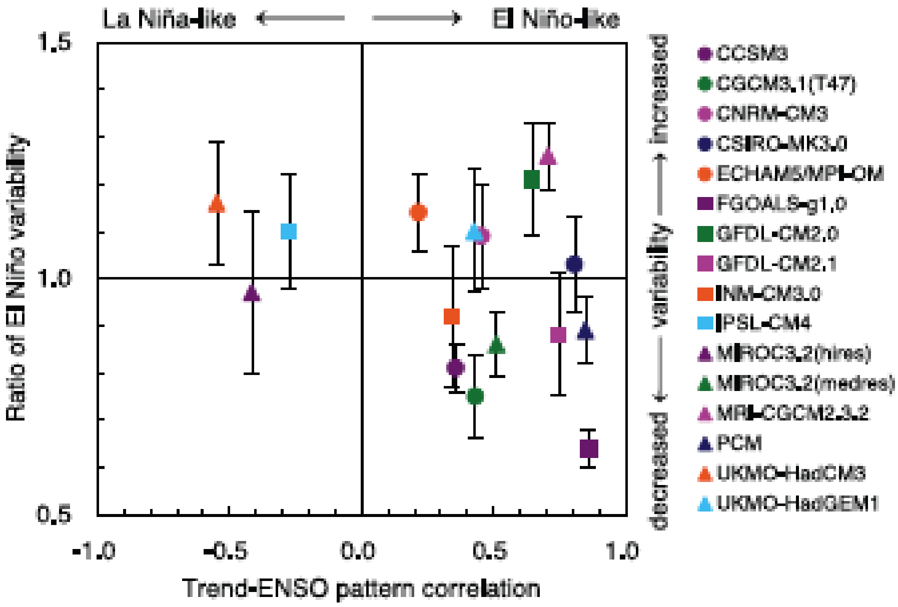


Figure 4.5. Base state change in average tropical Pacific SSTs and change in El Niño variability simulated by AOGCMs. The base state change (horizontal axis) is denoted by the spatial anomaly pattern correlation coefficient between the linear trend of SST in the 1% yr^{-1} CO_2 increase climate change experiment and the first Empirical Orthogonal Function (EOF) of SST in the control experiment over the area 10°S to 10°N, 120°E to 80°W (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 SRES A2 experiments (2051–2100), in the region 30°S to 30°N, 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 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 is projected during the late 20th century by models that include stratospheric ozone changes (e.g. Cai and Cowan, 2007; see Figure 4.6). 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 21st 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, precipitated 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 20th century, the confounding element of future trajectories in stratospheric ozone concentration makes projection of the SAM more problematic than other elements of Antarctic climate.

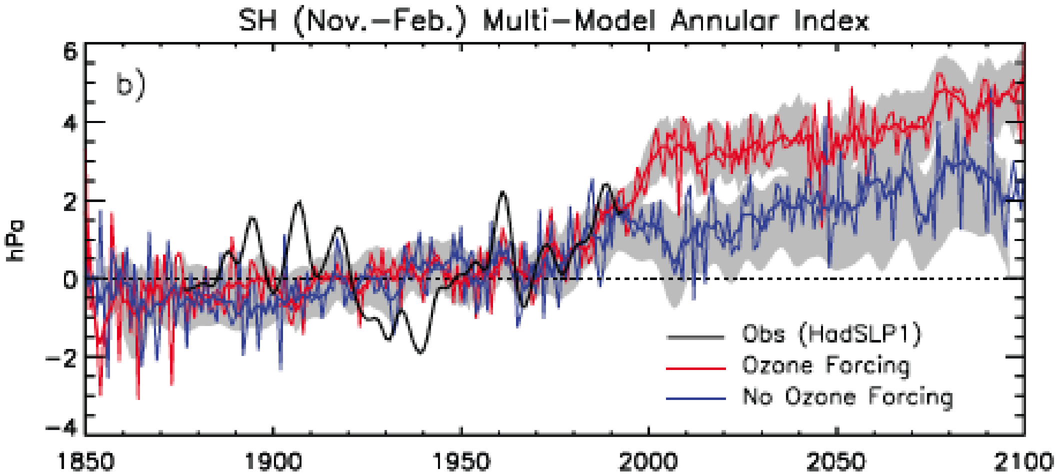


Figure 4.6. Multi-model mean of the regression of the leading EOF of ensemble mean Southern Hemisphere sea level pressure. The time series of regression coefficients has zero mean between year 1900 and 1970. The thick red (blue) line is a 10-year low-pass filtered version of the mean with (without) ozone 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).

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 re-analysis 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 Antarctic Ocean (e.g. Fyfe, 2003).

A consistent result that has emerged recently from 21st 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) demonstrate increasing cyclonicity and stronger westerlies in high southern latitudes in a 10-member multi-model ensemble simulation of the 21st 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).

Impacts on accumulation

The Antarctic ice sheet constitutes the largest reservoir of freshwater on earth, representing tens of meters of sea level rise if it was 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 represents a large fraction of net precipitation. Synoptic activity intensification over the Antarctic Ocean would suggest a potential for increase in accumulation along the coasts (Sinclair et al., 1997). However, significant correlations with the SO index 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).

4.2.4 **Temperature change over the 21st century**

A significant surface warming over Antarctica is projected over the 21st century. The average of the SRESA1B scenario runs of the CMIP3 models shows an increase of the annual average surface temperature of 0.34°C over land and grounded ice sheets (Bracegirdle *et al.*, 2008) (Fig. 4.7). All the CMIP3 models show a warming, but with a large range from 0.14 to 0.5°C dec^{-1} under the SRESA1B scenario. The difference between the ensemble average projections of each scenario is smaller then the inter-model spread for any given scenario.

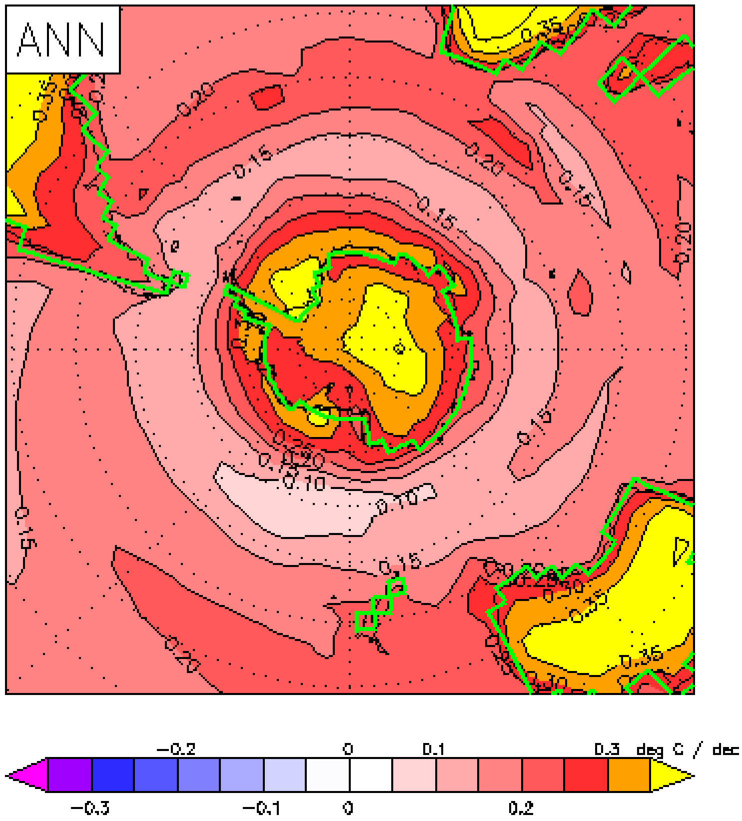


Figure 4.7. Skin temperature trend over 21st century in °C/decade.

Due to the retreat of the sea-ice edge, the largest warming occurs during the winter when the sea ice extent approaches its maximum, e.g. 0.51 ± 0.26°C dec^{-1} off East Antarctica (Bracegirdle *et al.*, submitted). 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 (Fig. 4.8). 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 from simulations and observations of temperature change for the latter part of the 20th century. The most notable difference is that the observed and simulated maximum of warming over the Antarctic Peninsula for the latter part of the 20th century is not present in projections of change over the 21st century.

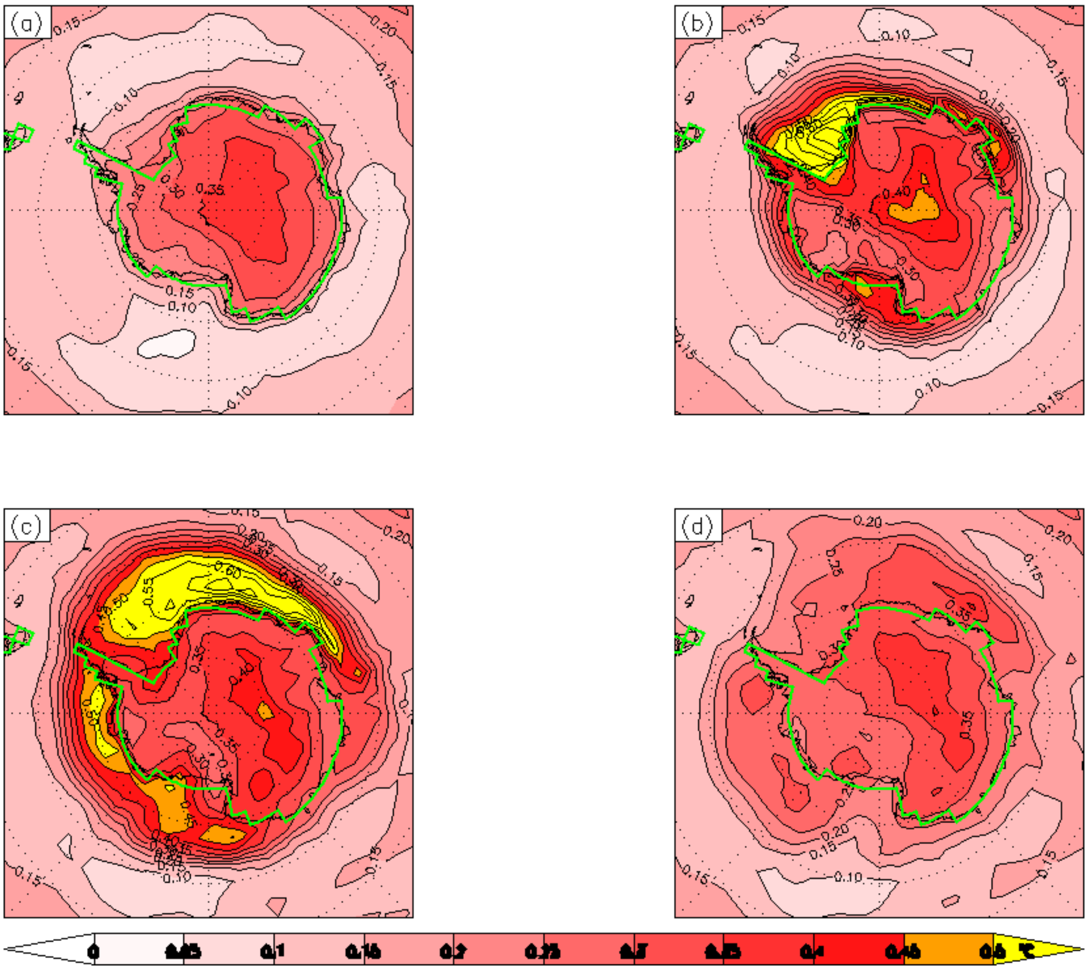


Figure 4.8. Skin temperature change over 21st century in °C/decade. (a) DJF, (b) MAM, (c) JJA and (d) SON.

The model consensus for warming is strong for Antarctica as a whole. However, 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 4.9 it can be seen that at most grid points the projected increases of temperature are significant. However, there is less confidence of the large warming trends around the coast than the smaller changes over the high interior. This is due to the large uncertainty over the sea ice and ocean projections.

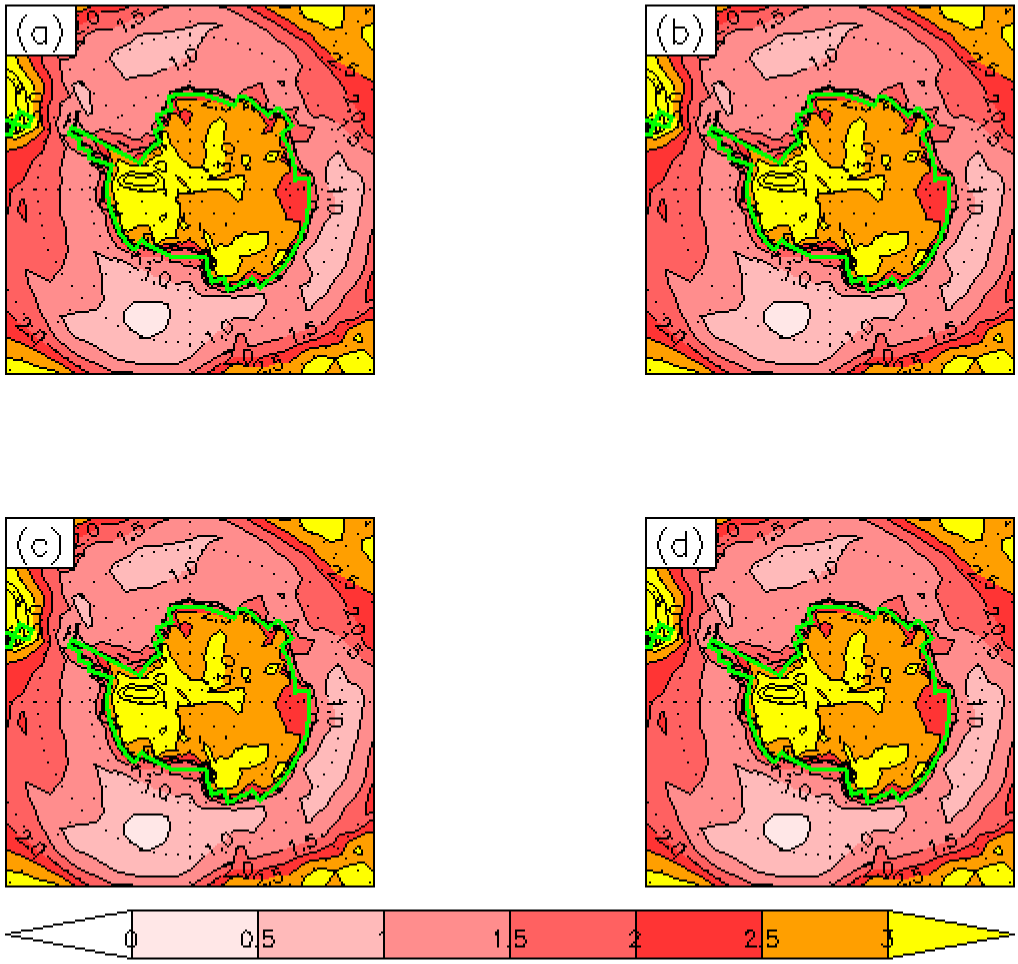


Figure 4.9. Inter-model standard deviation for individual grid points. (a) DJF, (b) MAM, (c) JJA and (d) SON.

Extensive projected sea ice retreat over the Arctic Ocean leads to much larger annual-mean surface temperature increases over the Arctic than over the Antarctic. According to the IPCC report the warming over the Antarctic continent is also 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 warming is much smaller than the global average due to the large heat uptake.

The annual ensemble mean warming rate at 500 hPa of 0.28 °C dec^{-1} is slightly smaller than the surface warming, with no evidence of a mid-tropospheric maximum, which 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).

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.

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 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 20^{th} century over the Antarctic Peninsula (Hughes *et al.*, 2006).

4.2.5 **Precipitation**

The net precipitation (precipitation less evaporation and ablation) over Antarctica is an important factor in the mass balance of the continental ice sheet, as has been noted. 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). Hence, quantification of precipitation changes in high southern latitudes is paramount resolving the uncertainty surrounding the future of the Antarctic ice sheet.

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. (2006) 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 20^{th} century. This range falls approximately between 150 and 190 mm $yearr^{-1}$, although recent analyses suggest the likely figure is at the high end of this range (Monaghan et al. 2006), who quoted 182 mm $year^{-1}$. Monaghan et al. (2006) 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. On the other hand, a significant increase in the net precipitation has been reported in the interior of the continent based on in situ measurements (Mosley-Thompson et al., 1999) and on satellite altimetry (Davis et al. 2005).

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 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

[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.

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 program to provide the necessary data support for such an endeavour. The precipitation simulations in high southern latitudes by global models result 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 a sensitivity can only be resolved by measurements that are currently not available.

An important driving mechanism for precipitation in this region is the 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. However, global models exhibit significant biases in this regard also. Placed in the context of the available re-analyses, as discussed in section 5.2.3, 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 AMIP-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) note that the reanalysis products, and therefore probably global models in general, underestimate the Antarctic precipitation in the 20th 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).

Projected changes in precipitation

The lack of a 20th century trend in net precipitation in the some recent comprehensive analyses is particularly problematic in the context of 21st century model projections. Almost all climate models simulate a continuing robust precipitation increase over Antarctica in the coming century (see figure 4.10.) 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^{-2}. 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. They 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 and hence an increase of precipitation. 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

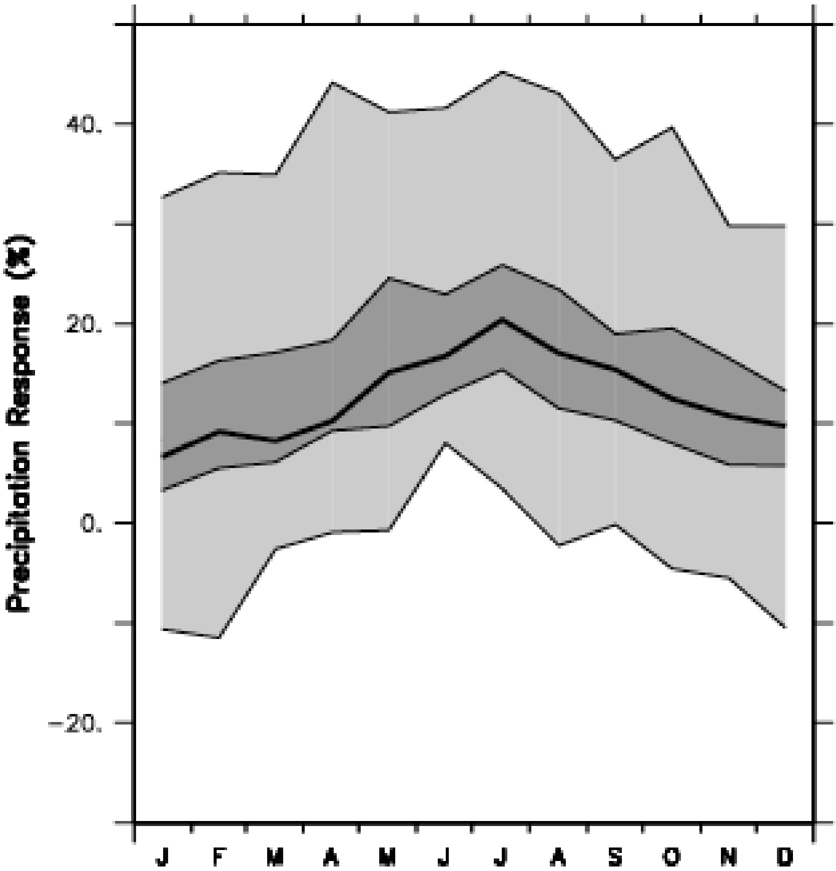


Figure 4.10. Annual cycle of Antarctic continent area mean percentage precipitation changes (averaged over the Antarctic continent) for 2080-2099 minus 1980-1999, under the A1B scenario. Thick lines represent the ensemble median of the 21 MMD models. The dark grey area represents the 25% and 75% quartile values among the 21 models, while the light grey area shows the total range of the models.

after the middle of the 21st 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., (2006).

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 4.11.) 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 lower future cyclone frequency.

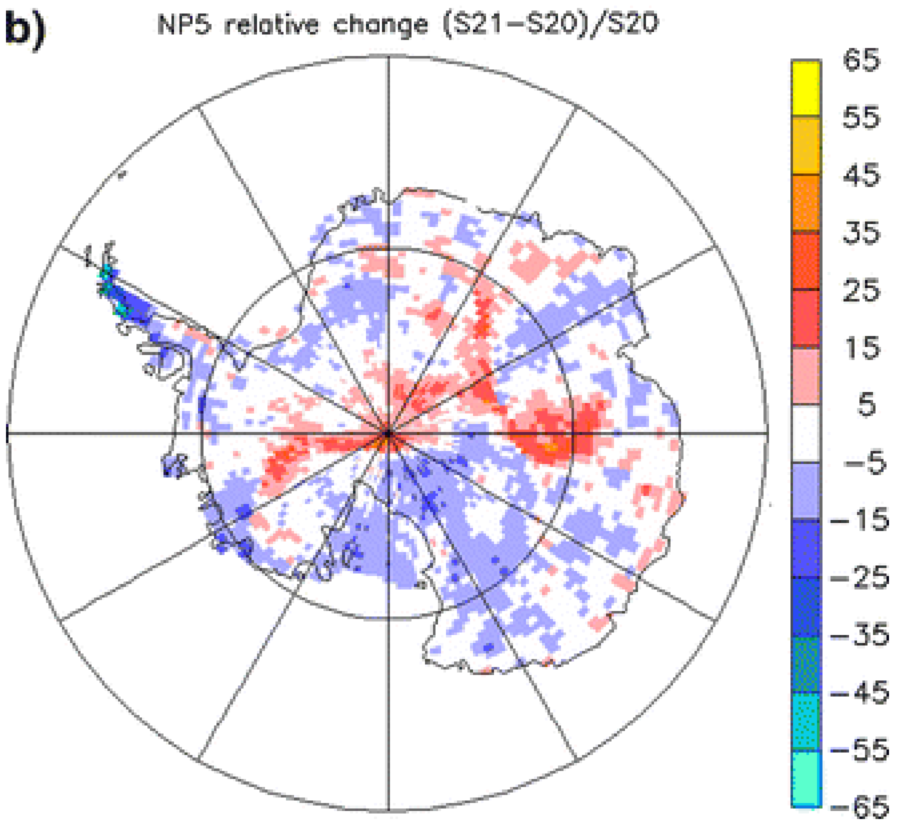


Figure 4.11 from Krinner et al. 2007: Number of days with precipitation exceeding five times the annual daily mean. b relative change from the end of the twentieth to the end of the twenty-first century (in percent).

The scatter among the individual models reported on in the AR4 is considerable. During the first half of the 21st century, models using the A1B scenario predict anything from maximum upward trend of 0.71 mm $year^{-2}$ a the maximum downward trend of 0.13 mm $year^{-2}$. Some models project stronger net precipitation increases in the first half of the 21st 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. estimated that by the end of the century the snowfall rate over the continent was simulated to 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 5cm.

4.2.6 **Atmospheric chemistry**

4.2.6.1 **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 fieldsfrom 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.

Table 4.1. Models whose results appear in Figures 4.12, 4.13 and 4.14.

Model	Institute	Resolution	Number 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)
MAECHAM4CHEM*	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 & Chipperfield (2005)
WACCM	NCAR (USA)	4.0° x5°	66	4.5×10^{-6}	Park et al. (2004)

* 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 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

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 ClNO3, as well as the reactive Cl, ClO, OClO, Cl2O2 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, 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, changes in temperature due to increased greenhouse gases may lead to amounts of ozone quite different to those of pre-ozone hole days.

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, and this cooling will slow gas-phase ozone loss reactions, which will tend to increased ozone.
2. The same cooling acts to increase amounts of Polar Stratospheric Clouds (PSCs), on which the heterogeneous reactions leading to ozone loss occur, so we can expect reduced ozone given the same Cly - opposite to the effect of gas-phase chemistry. This is particularly likely in the edge region of the vortex (Lee et al. 2001) because PSCs are not ubiquitous there. It is ozone loss in this edge region that defines the ozone hole area diagnosed in Figure 4.13.
3. Increased greenhouse gases 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) EESC. Each of these helps remove ozone, so this process also acts to reduce ozone in the short term. In the long term, the increased production of EESC would more rapidly reduce ozone-depleting substances in the whole atmosphere, so that the effect via EESC would oppose those via CH4 and N2O.

Despite these difficulties, the general characteristics of future Antarctic ozone in Figure 4.12 are similar in all models, and similar to projections in earlier models in WMO (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 to detect the onset of ozone hole recovery (Hofmann et al. 1997).

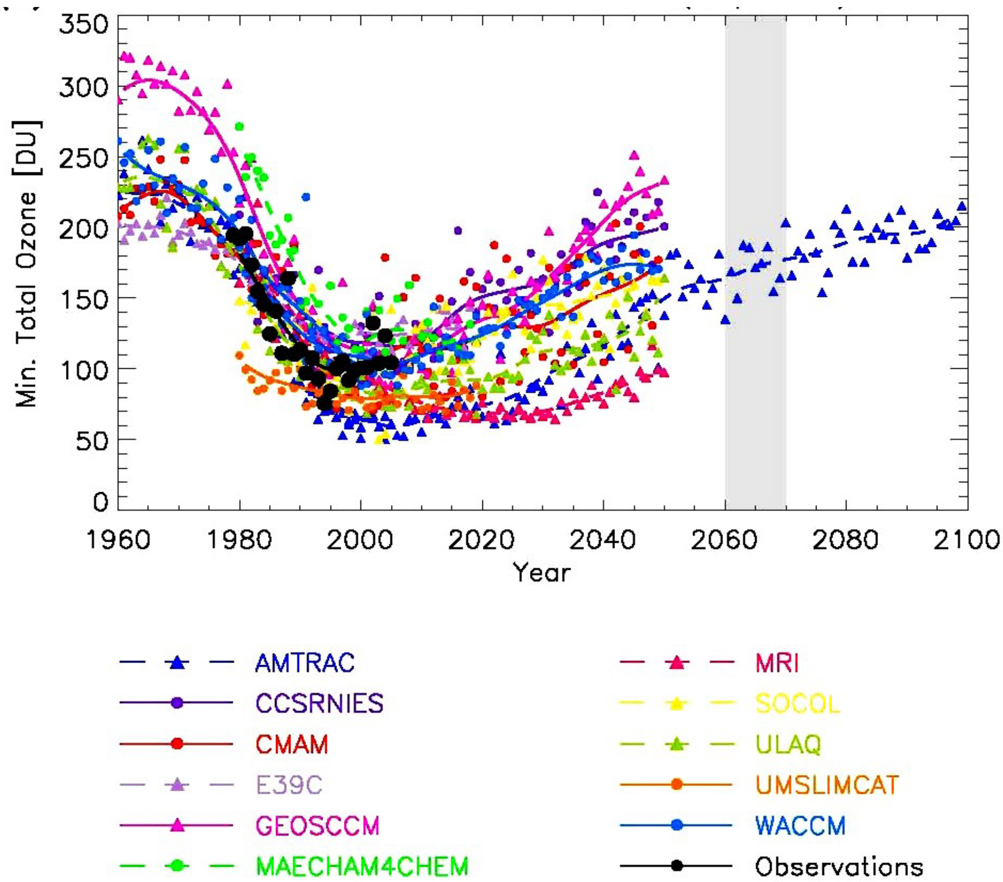


Fig. 4.12. Minimum total column ozone in September to October predicted by various models, plus observations from the NIWA total ozone database (Bodeker et al., 2005). Solid and dashed curves show smoothed values. Light gray shading shows when EESC is expected to return to 1980 values. Adapted from WMO (2006).

The minimum amount of ozone in Figure 4.12 differs widely between models, ranging from 60 DU to over 120 DU compared to the observed 80 DU, highlighting the difficulty of predictions in 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.

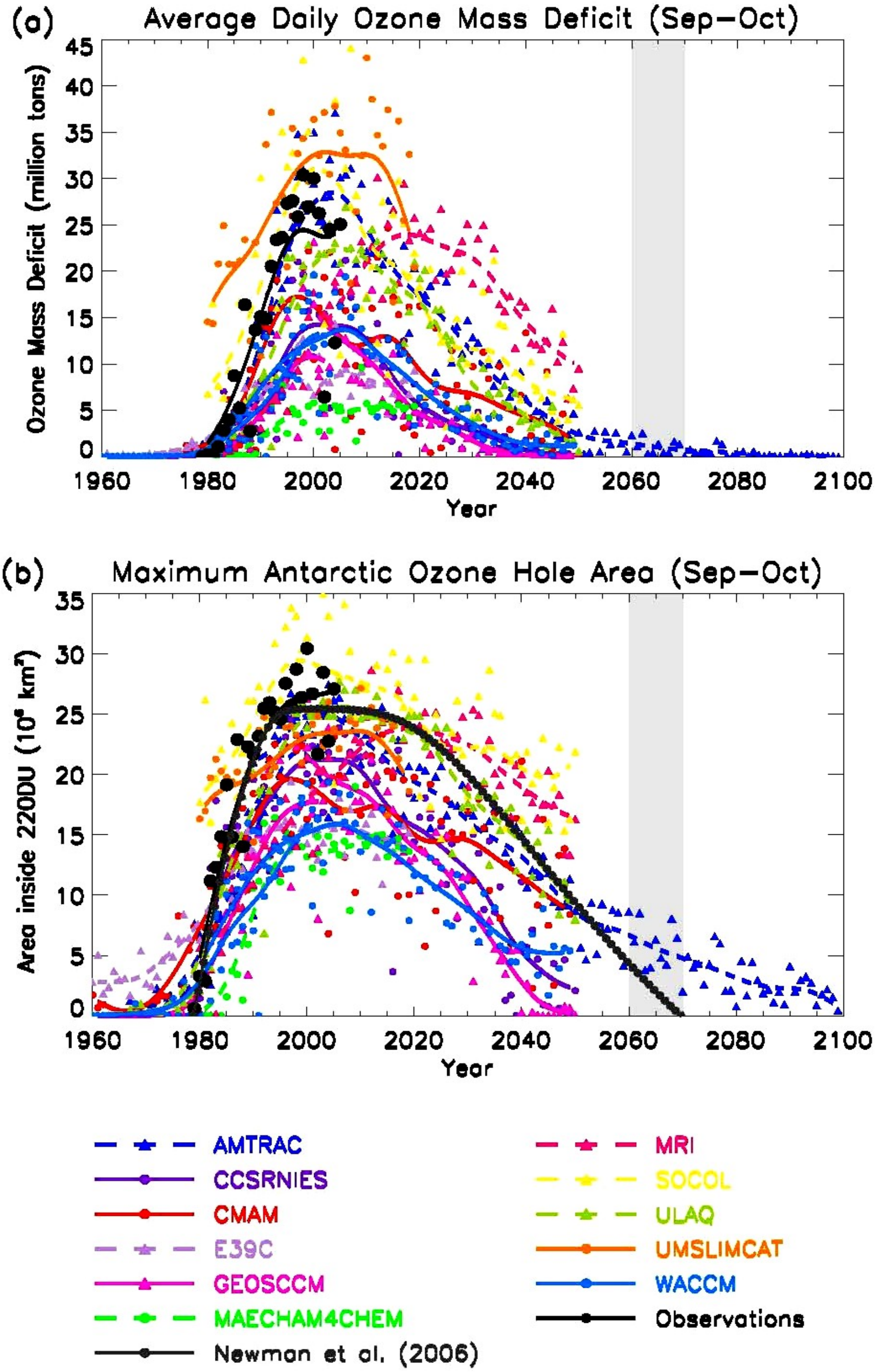


Fig. 4.13. (a) September to October average daily ozone mass deficit, and (b) the maximum ozone hole area, for each year from each model. Curves, shading and source of observations are as in Figure 4.12. Adapted from WMO (2006).

Similarly, the predicted values of maximum ozone mass deficit in Figure 4.13 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 amounts 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).

Some insight into the model differences can be obtained from comparisons of Cly in the models. As shown in Figure 4.14, 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 EESC. AMTRAC matches the observations of EESC best, and predicts the latest return to 1980 values.

AMTRAC also predicts the latest recovery of ozone in Figures 4.12 and 4.13. This is almost consistent with the study of Newman et al. (2006), who used a parametric model of spring ozone amounts that includes EESC amounts and stratospheric temperatures. Figure 4.13 shows that they predicted a return to 1980 ozone amounts would not occur until about 2070.

Despite the differences in models, extrapolating their results suggests that at the end of the next 100 years, 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.

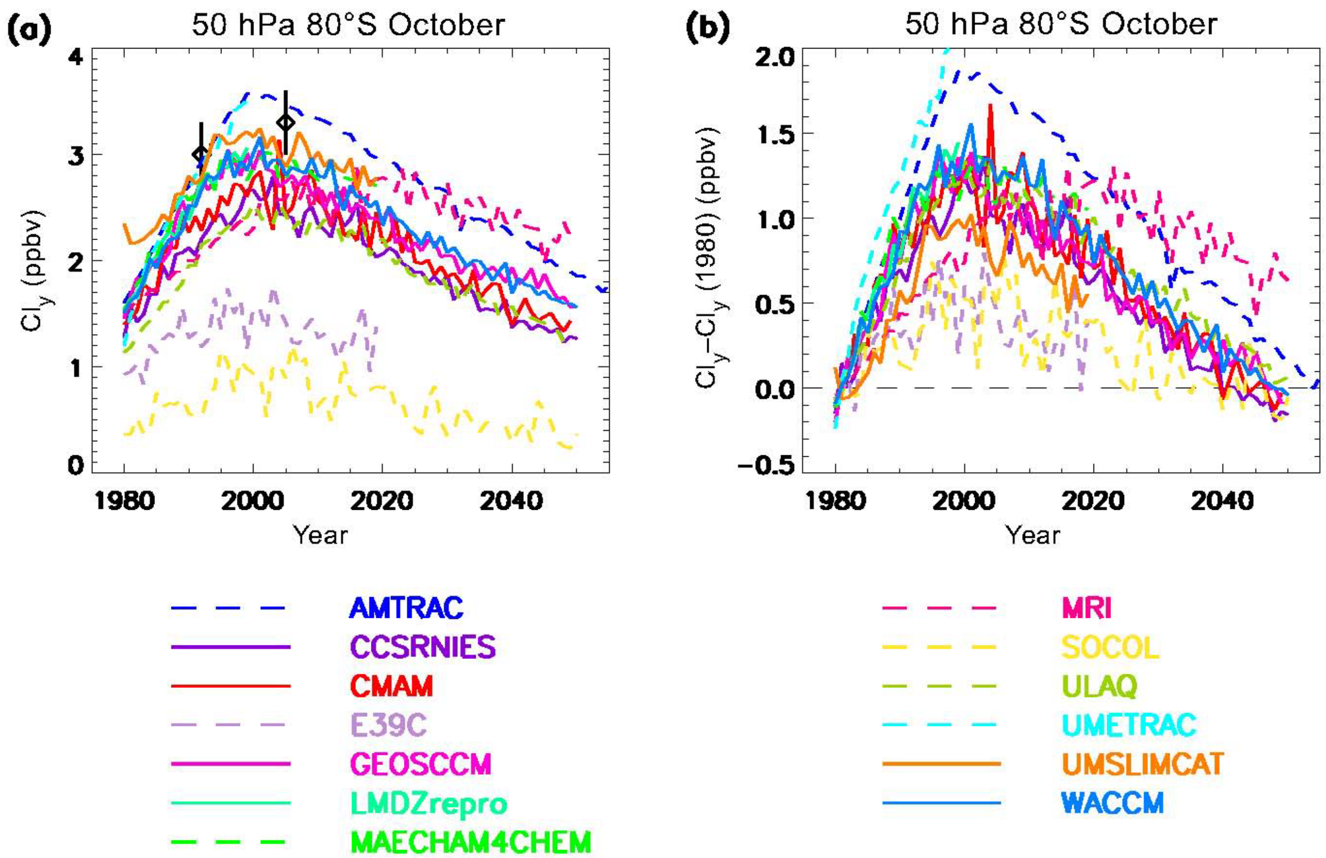


Fig. 4.14. Zonal mean values of total inorganic chlorine (ppbv) in October at 50 hPa

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.

4.2.6.2 **Antarctic Tropospheric Chemistry**

The chemistry of the Antarctic troposphere, its composition and reaction pathways, are effectively dominated by the presence of the cryosphere. The influences are various and act in both complimentary and competitive ways, some establishing feedbacks. In this section, we discuss the influences and how they would be likely to change in a warmer world with a reduced cryosphere.

At the most fundamental level, the cryosphere provides a bright surface which reflects incoming radiation from the sun. For tropospheric chemistry, solar radiation provides energy to the system; trace gases are photolysed, ie. split apart by 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. The albedo over snow is generally taken to be around 0.98, such that the presence of snow on both land and sea ice surfaces, will effectively double the pathway of solar radiation, and therefore double the likelihood of trace gases in the troposphere being photolysed. The albedo of snow is a powerful influence that maintains a reactive system. Without it, the lifetime of trace gases would be increased and the chemistry would in general move to a less reactive state.

A further influence of the cryosphere is as a cap to emissions from the underlying land or ocean. In Antarctica, should temperatures rise to the point where the cryosphere disappeared, it is likely that microbes would become active in the continental soil. In present day soils, microbes are recognised sources of nitrous oxide (N_2O) which is a long-lived greenhouse gas, and additional emissions would be likely to contribute to some extent to greenhouse warming.

With regards the ocean, a loss of the sea ice would enable emissions of trace gases with an oceanic origin. Various gases that are measured in the troposphere are known to be released from the oceans around Antarctica. Examples 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 of these are generated by phytoplankton. Precisely which phytoplankton species release which trace gases is not known and, consequently, neither is the geographic distribution of these emissions. For the discussion here, we assume that all gases have the same, or similar, origins. The seasonal cycle in the atmosphere of these trace gases is closely linked to the extent of sea ice but also to the sun. In a warmer world with a reduced sea ice extent, emissions from the ocean would likely increase (all other things e.g. nutrient availability, being equal). The dominant control would then be the Sun, so that a seasonality with a wintertime minimum but an extended summer maximum could be expected. DMS plays a critical role as a source of cloud condensation nuclei (CCN) via its oxidation to sulphate (see e.g. Finlayson-Pitts and Pitts, 1999). Changing the number of CCN alters cloud properties and albedo with a consequent influence on the Earth's radiation budget, surface temperature and climate. Indeed, it was proposed some years ago that feedback loops could 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). Oxidation of DMS is predominantly driven by reaction with 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, and at temperatures below 285K the addition channel leading to dimethylsulphoxide (DMSO) is believed to dominate (Arsene et al., 1999). However, it is the abstraction channel that ultimately results in production of SO_2, and hence CCN. In a warmer world, with a warmer troposphere, the proportion of DMS oxidising with OH via the abstraction channel may therefore increase, leading to the expectation of additional CCN formation. However, given anticipated increases in other oceanic trace gases, this may not be the case. For example, BrO can also react with DMS, oxidising it to DMSO. Indeed a recent modelling study (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., in prep). 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. Thus, even with enhanced levels of DMS in the future it is not clear that greater numbers of CCN would result. The competing reaction with BrO would oxidise DMS to DMSO rather than SO_2 and consequent CCN. Whether or not any climatic influence resulted from these reactions would depend on the balance between increases in the oceanic sources as well as ambient temperature.

Sea ice also has a direct influence on tropospheric chemistry. As described in section 4.5.2, 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. However, 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 0.05 K, i.e. a small but positive feedback.

The potential for feedbacks also exists elsewhere. Recent measurements from Halley station, in coastal Antarctica, found significant concentrations of iodine monoxide (IO) in the boundary layer (Saiz-Lopez et al., 2007). The seasonal maximum occurred in spring, but even during the summer months concentrations were sufficiently high to influence tropospheric chemistry processes. The IO was postulated to originate from diatoms that live underneath the sea ice. IO in the remote atmosphere is known to be a major source of new particles (marine aerosols and CCN) from which clouds originate. As discussed above, marine aerosols and clouds scatter incoming solar radiation and so contribute a cooling to the Earth's radiation budget. If the source of IO is linked to the presence of sea ice, then in a warmer world with a reduced sea ice coverage, the production of IO is likely to decrease. The consequent effect would be fewer aerosol and CCN particles, less radiative scattering, and hence an additional warming at the Earth's surface. If on the other hand, IO emanates directly from the ocean and is not dependent upon the presence of sea ice, the opposite effect might be expected; reduced sea ice could increase IO emissions and thereby increase CCN and aerosol production. The consequent cooling would then be a restorative influence rather than a positive feedback. In order to clarify whether the feedback would be positive or negative, it will be necessary to determine the source of boundary layer IO.

The final issue to be discussed here is the role of the snowpack itself. Section 4.5.2 outlined the influence that the snowpack has on the present day boundary layer by acting as a source of highly reactive trace gases. Over parts of Antarctica, such as the polar plateau, these dramatically increase the oxidising capacity of the boundary layer above what had been expected. In coastal regions, the effect is less pronounced, but nonetheless significant. The differences occur partly as a result of the fetch of snow and also on the stability of the boundary layer. Clearly, if all the Antarctic snow disappeared, this source of trace gases would also be removed, and the atmosphere would most likely move to a more sluggish state with longer-lived and less reactive chemical species. However, assuming that at least the East Antarctic Ice Sheet is still in existence in 100 years time, the likelihood is that emissions from snow will still be an important driver of local tropospheric chemistry.

Clearly the cryosphere over and around Antarctica has an enormous influence on tropospheric composition and chemistry. The discussion presented here attempts to focus on specific processes that would be influenced by changes in the cryosphere. Although there are significant uncertainties, as outlined in the discussion, it is nonetheless clear that a change in the Antarctic cryosphere would greatly alter Antarctic tropospheric chemistry from its present day state. Furthermore, within these changes there are potential feedbacks that might themselves influence the future climate system.

4.3 Terrestrial Biology

Antarctica is not specifically different from other continents, just extremely isolated and at the end of the spectrum of planetary conditions. Climate change will impose a complex web of threats and interactions on the plants and animals 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. Antarctica contains some of only places on Earth where natural biological phenomena can be studied in their pristine state, but human visitation risk breaking Antarctica's isolation, and seriously threatens Antarctica's unique legacy.

The consequences even of direct environmental changes might not always be straightforward to ascertain. For example, many sub-Antarctic islands are showing 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 very 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 2005b, 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 soluble carbohydrate loss (Tearle 1987, Melick and Seppelt 1992). Thus, one of the major consequences of climate change might paradoxically not be an increase in upper lethal 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 (Chown 2001).

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 RiSCC). 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 species more strongly 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 it is salient to note that 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 defence and mitigation strategies, reducing that available for other life history components (Convey 2006 RiSCC, Hennion et al. 2006 RiSCC, Robinson et al. 2005). Changes in water availability will also impact on both terrestrial and the more the stable limnetic environments. Local reduction in water availability in terrestrial habitats can lead to desiccation stress (Convey 2006 RiSCC) 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 in stress tolerance, plasticity and variation in life histories (Convey 1996). However, and critically, these adaptations have been at the expense of reduced competitive ability, leaving Antarctic ecosystems vulnerable to the impact of colonisation by better competitors, that may be at more advantage under changed climatic conditions (Bergstrom and Chown 1999, Convey and Chown 2006 RiSCC, Convey et al. 2006a RiSCC). 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, 2007 CEP, Whinam et al. 2005, Convey et al. 2006 RiSCC, Convey 2007 CEP). 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 establishment.

Changes in temperature, precipitation and windspeed, 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 RiSCC, Lyons et al. 2006 RiSCC, Quayle et al. 2002, 2003). Indeed, Quayle et al. (2002, 2003) highlight 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, has resulted in rapid increases in lake salinity in the last few decades (Hodgson et al. 2006).

Increased inputs of meltwater into the mixolimna 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 RiSCC). 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 RiSCC) 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.

4.4 The terrestrial cryosphere

4.4.1 Introduction

Recent observations of ice sheet behavior in Greenland and Antarctica have forced experts to radically revise their view of ice sheet sensitivity to climate change. Existing ice-sheet models, do not properly reproduce this observed behavior, casting doubt on the value of these models at predicting future changes. Predictions of the future state of ice sheets must, therefore, be based upon a combination of inference from past behavior, extension of current behavior, and interpretation of proxy data and analogs from the geological record. The broad range of response time-scale exhibited by ice sheets, however, guarantees that future behavior 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 (see 4.7.1), growth in some areas, loss in others, some related directly to atmospheric climate, some related to changes in oceans, possibly driven by climate change. No single external driver, no single aspect of climate change (changing atmospheric temperature, changing snowfall rate, or change 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 likelihood that in a warmer world there will be less ice—and higher sea level. One million years of paleoclimate data support this expectation. However, these 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 behavior, will be episodic to some degree. Translated into a picture of future ice sheet behavior, this suggests there will be a combination of coherent activity and disjoint activity among individual glacier basins and groups of glacier basins.

Rates of sea level rise at least twenty times the current rate (~3.1 mm/a) sustained over more than a century have been measured during the transition to the current warm period following the termination of the last ice age. Although, this occurred at a time when there was far more ice on Earth, it occurred during a period when climate change was probably not as fast as is projected for 2000-2100 under many emission scenarios. Thus, until an improved predictive capability for ice sheets is achieved, this might be regarded as the upper bound of Antarctica's potential contribution to global sea level. The present contribution is far less than this value and it is not likely such a maximum rate can be realized quickly; even an exponential increase in rates to 2100, while capable of producing 1 to 2 meters of sea level rise, contributions in excess of this are unlikely this century. Nevertheless, even these smaller contributions would come at great human and environmental cost, underscoring the need for a better understanding of ice-sheet sensitivity to climate change.

The most likely regions of near-future change are those that have been shown to be changing today. However, most models agree that, although warming was not observed everywhere in Antarctica in the last 50 years, it will be strong in the coming century, and so it is likely both snowfall and melt will increase during this century. Even the relatively small increases in the rate of snowfall that are expected to parallel this warming, would cause significant volumetric growth due to the vast area of the ice sheet—average accumulation across all of Antarctica is only 15 cm/a (water equivalent) and much lower in the cold and high elevation, deep interior of East Antarctica. But warming 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., 2006). However, models disagree over many of the other details of atmospheric climate change in Antarctica, and generally share no agreement in predictions of future ocean changes, which might be the most significant influence of all on the ice sheets.

4.4.2 **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 (Zwally et al., 2006).

As discussed in Section 3.8, the cause of the interior thickening is probably a long-term dynamic response to a distant change in climate and not recent increased snowfall. This effect is likely to continue and change only slowly. Eventual atmospheric warming over the EAIS 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 ºC 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, 2007 #2409}. 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^{o}C.

There are marine basins beneath the EAIS and the potential for these to promote 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.

4.4.3 **West Antarctic ice sheet**

The current loss of mass from the Amundsen Sea embayment of the West Antarctic ice sheet (WAIS) is equivalent to that from the entire Greenland ice sheet (~50 Gt per year) {Lemke, 2007 #2408}. This sector of the WAIS is by far the most active, substantially outweighing the Ross Sea sector where the stagnation of Kamb Ice Stream has caused a positive mass balance contribution. Little net change is measured from the remaining Weddell Sea sector.

The future of WAIS is always cast against the backdrop of the "marine 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). And 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).

The process believed responsible for this dramatic behavior is a progressive thinning of the fringing ice shelves seaward of the Amundsen Sea outlet glaciers—in effect, a slower version of the glacier acceleration that was observed to follow ice-shelf disintegrations in the Antarctic Peninsula. The likely chain of events leading to this thinning may begin 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 meters 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 0.75 meters to global sea level if the area of ice lost is limited to that where the bed slopes downward toward the interior of West Antarctica (Holt et al, 2006; Vaughan et al., 2006), so a contribution from this sector alone of some tens of centimeters by this century's end cannot be discounted.

To look at this another way, the loss of the Larsen – B ice shelf caused a 2 – 8-fold increase in the glaciers feeding it. If a similar acceleration (e.g., due to loss of its ice shelf), were to occur on either Pine Island or Thwaites glacier, the sea-level rise contribution would be 0.5 and 2 mm per year in global sea-level rise. Which could clearly mount up to a substantial additional contribution in just a few decades.

The other sectors of the WAIS are doing little to offset this contribution. 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 scientist's attention and is being monitored by a host of sensors.

Measurements made of 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 in 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 potential for counterintuitive behavior of the Antarctic ice sheet does exist and must not be dismissed too easily given our limited understanding of recent ice-sheet behavior. 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, that reduce sea-ice production, could decrease ocean overturning and actually 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 could actually cause thickening upstream and a local advance of the ice-sheet.

4.4.4 **Antarctic Peninsula and sub-Antarctic islands**

Within the Antarctic Peninsula, recent climate changes have had a dominant impact on the behavior of resident glaciers and ice shelves. Thus, 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, and thus poor confidence in current projections.

Nevertheless, some aspects of the mechanisms causing recent warming are becoming understood. Warming rates vary seasonally. Winter warming, maybe primarily 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 $0^{o}C$) is an important parameter tied to glacier growth/shrinkage and these are projected to increase as well, offsetting some portion of the increased snowfall.

Applying the concept to 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, and/or progressive retreat of the calving front. Prediction of the timing of ice shelf disintegration is not yet possible, but field programs 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 prescient signs of disintegration. Data collected during a disintegration would improve the understanding of the causal mechanisms and lead toward a better predictive capability. Studies of tidewater glaciers, a possible useful analog to glaciers on the Antarctic Peninsula, supported by years of observations in some cases, could add insight into the controlling mechanics and sensitivities.

Although at present most of the effects leading to loss of ice are confined to the more northerly section of the Antarctic Peninsula (which contains a few centimetres of global sea-level rise), the total volume contained on the Antarctic Peninsula is 95,200 km^3 (equivalent to 242 mm of sea-level; Pritchard and Vaughan, 2007). This is roughly half that of all glaciers and ice caps outside of either Greenland or Antarctica. The mechanisms that have led to recent acceleration of these glaciers as an immediate consequence of ice shelf disintegration, and glacier retreat, which could progress further south in the coming century, suggests that this ice could impact global sea level relatively rapidly, and perhaps be comparable to the more gradual melting of glaciers and ice caps in other mountainous environments across the Earth. Once GCMs are able to properly reproduce the regional warming and the seasonality of this warming, they may become a useful tool of predicting ice shelf futures through the concept of their thermal viability.

4.4.5 **Conclusions**

Efforts to measure the mass balance of ice sheets have borne fruit in recent years with multiple emerging methodologies that have provided a troubling picture of increasing ice loss in the past one or two decades. The short periods for which established records exist compared to the rapid rate of change make it difficult to extrapolate these observations into the future. In addition, 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, and led to the understandable caution expressed by the authors of the recent IPCC review, to express the uppermost bound on sea-level rise.

Nonetheless, a conservative expectation of changes no greater than those already observed leads to an expectation of increasing loss of ice from the Antarctic Peninsula and from 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 by the IPCC, have aimed at the predictable components, shying away from the less predictable components, 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.
In summary, in an assessment of expert glaciological opinion made in 2000 (Vaughan and Spouge, 2002), it was determined that a group of leading glaciologists believed that within the next 200 years there remained a 30% probability that loss of ice from the West Antarctic ice sheet could cause sea level rise at a rate of 2 mm/year, and a

5% probability it would cause rates of 1 cm/year. Since that opinion was gathered there has been enormous scientific progress made in understanding the ice sheet. However, little has been discovered that would cause a reduction of the levels of risks expressed in that expert judgment. Conversely, the recent observations that inland ice sheets can be impacted by the loss of floating ice shelves and the continued acceleration of ice-sheet thinning and glacier-flow in the Amundsen Sea embayment are now firmly understood to be the result of glacier-acceleration, and can no longer be argued to result from a few years of unusually low-snowfall rates (as might have been interpreted at the time that the opinion was gathered {Wingham, 1998 #944}), but can be argued to support the hypothesis that the ASE could already be entering a phase of collapse. Similarly, recent improvements in numerical analysis of the stability of marine ice sheets {Schoof, 2007 #2283; Schoof, 2007 #2287}, which are supported by many other ice-sheet modellers, appear to reinforce earlier concerns that a potential instability does exist in marine ice sheets that could lead to deglaciation of parts of WAIS.

4.5 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 from 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 scenario, a range of climate models and an additional uncertainty for land-ice changes. For the IPCC's Fourth Assessment Report (AR4, 2007), the range of sea-level projections, using a larger range of models, is 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. In these projections, the great ice sheets of Greenland and Antarctica together contribute little to sea-level rise during the 21st century. Increased snow fall on Antarctica partially offsets positive sea-level contributions from other components. However, it is recognized that our ice sheet models are incomplete and do not allow for a rapid dynamic response of the ice sheets, as indicated in some recent observations. To allow for ice sheet uncertainties, IPCC AR4 increased the upper limit by 10 to 20 cm and stated 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 (Figure 4.15) are similar.

There is increasing concern about the stability of ice sheets. For Greenland, this concern is based on measurements indicating an increasing contribution from the ice sheet and melt water possibly finding its way to the base of the ice sheet facilitating rapid sliding of glaciers and thus contributing to a more rapid sea-level rise. Much of the West Antarctic Ice Sheet is grounded below sea level and the penetration of warmer water beneath the ice shelves to the base of the ice sheet and the subsequent dynamic response could also lead to a more rapid rate of sea-level rise. However, the current suite of ice sheet models does not adequately represent many of these processes and thus projections of ice sheet contributions to both 21st century and longer term sea-level rise may be underestimated. Concern that the sea-level projections may be biased low has been reinforced by a comparison showing that observed sea level has been rising more rapidly than the central range of the IPCC projections, since 1990, and is at the very upper end of the IPCC projections (Rahmstorf et al, 2007) suggesting that one or more of the model contributions to sea-level rise is underestimated.

Since 1993, there are high-quality satellite-altimeter observations of sea level over nearly all 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 20th century, with an increase in the rate of rise over this period (Figure 4.15). Jevrejeva et al. (200?) and Holgate and Woordworth (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 about 3.2 ±0.4 mm yr^{-1}. Note that these rate of increase are an order of magnitude faster than the average rate of rise over the previous several thousand years.

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.

Sea-level rise over the early 21st century has largely been determined by past emissions of greenhouse gases. However, sea-level projections closer 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 increasing concern about the longer-term contributions of the ice sheets. For example, for the Greenland Ice Sheet, global average temperature increase relative to pre-industrial values of greater than 3.1 °C (with a range of 1.9 °C to 4.6 °C) leads to surface melting exceeding precipitation, resulting in an ongoing wastage of the Greenland Ice Sheet for centuries and millennia (Gregory and Huybrechts 2006), consistent with sea levels in the last interglacial being several metres higher than today's value. This value could potentially be crossed late in the 21st century if effective mitigation measures are not adopted. A second cause for concern is associated with the poorly understood dynamic responses of the Greenland and West Antarctic Ice Sheets that could lead to a significantly more rapid rate of sea-level rise than from surface melting alone.

Regional Projections of mean sea-level rise

Overlying the global sea-level rise is a large regional variability and sea-level rise during the 21st century is not expected to be uniform around the globe. This is a result of changing atmospheric conditions, particularly surface winds (Lowe and Gregory, 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 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 and Gregory, 2005).

Sixteen of the models used for the IPCC assessment are shown in Figure 4.16 using the mid-range SRESA1b scenario (stabilization of CO2 equivalent at 720ppm by 2100). The models are showing 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 have reduced sea level increases compared to the global average of 15-20cm. 4 of the models show however in the coastal region close to Antarctica there is small increase in sea level above the average of 5-10cm. 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.

Mid latitude studies (McInnes 200, Lowe, 200) 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, and Hansom, 1981). The storm surges in these areas are driven by the inverse barometer effect. The long record from the Falklands has an upward sea level secular trend of 0.70+- 0.18mm year since 1964, about half the observed global figure.

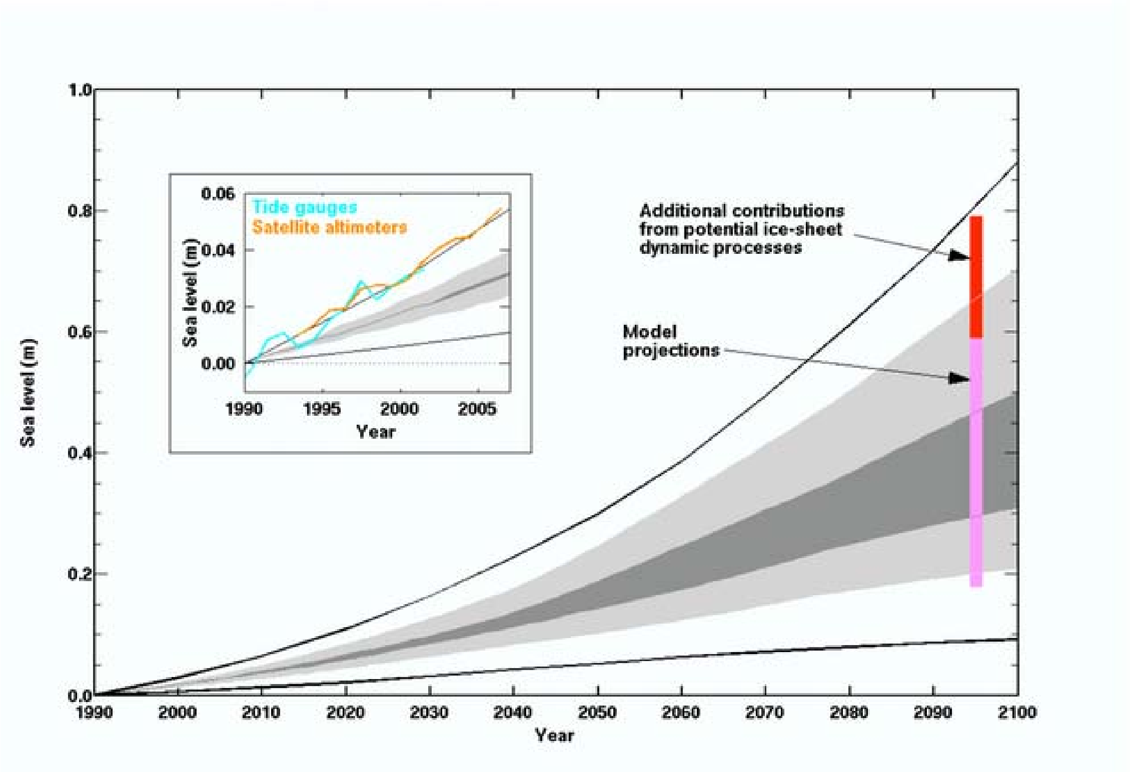


Figure 4.15. Projected sea-level rise for the 21st century. The projected range of global averaged sea-level rise from the IPCC 2001 Assessment Report for the period 1990 to 2100 is shown by the lines and shading. (The dark shading is the model average envelope for all SRES greenhouse gas scenarios, the light shading is the envelope for all models and all SRES 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).

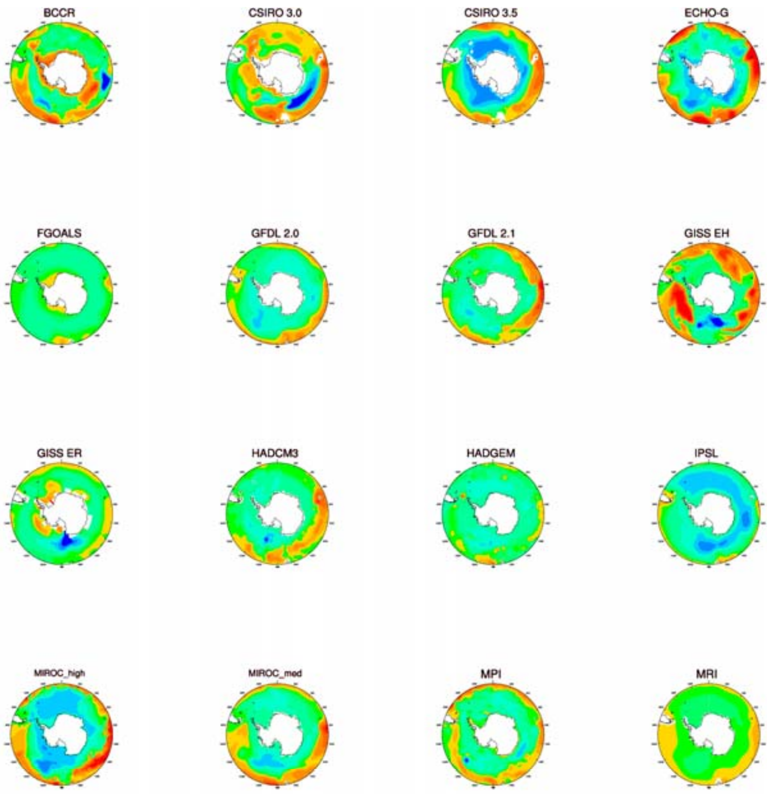


Fig. 4.16. Sea level rise projected by sixteen of the models used for the IPCC assessment.

4.6 Biogeochemistry – response of the Souhern Ocean carbon cycle to future climate change

Introduction

Atmospheric CO2 emissions continue to increase at unprecedented rates, already anthropogenic emissions have equaled or exceed the worst cast emission scenarios for the IPCC [Raupach, et al., 2007], Figure 4.17. Currently it is estimated that about 30 % of the CO2 emitted annually is taken up by the ocean. The Southern Ocean plays a critical role taking up this atmospheric CO2, with more than 40 % of the annual mean uptake of atmospheric CO2 being taken up in the region south of 40°S [Takahashi and al., 2008]. The important role of the Southern Ocean in taking up atmospheric CO2 emissions suggests that how this region will respond to climate change will directly impact atmospheric CO2 levels and hence the rate at of the earth's warming. In addition this must be a priority in order to start to formulate effective global policy for stabilizing atmospheric CO2 levels.

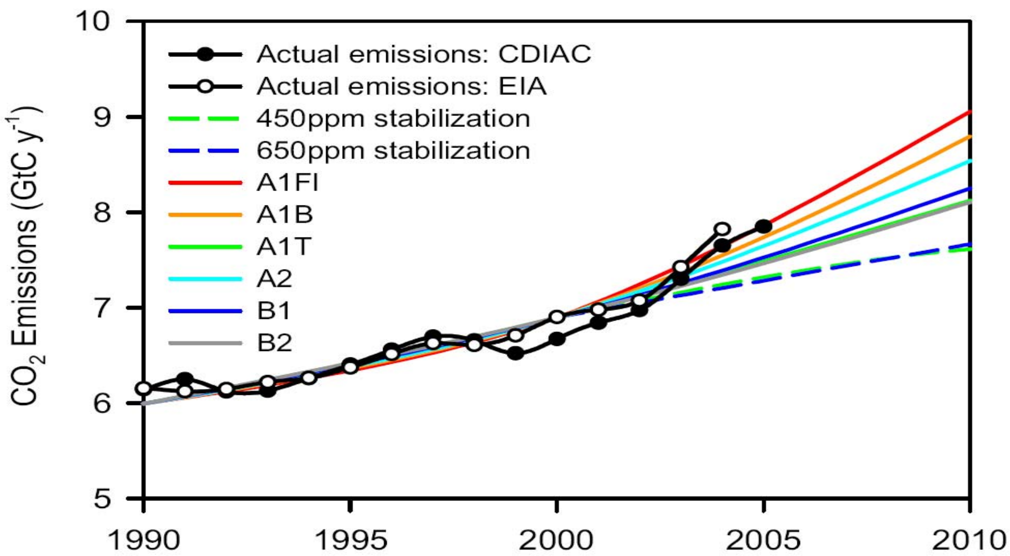


Fig 4.17. Observed CO2 emissions over the last 25 years as compared with the different IPCC emission scenarios [*Raupach, et al.*, 2007].

As discussed in chapter 3.11 the Southern Ocean in recent decades has undergone significant change; in particular there has been a reduction in CO2 uptake at interannual and longer timescales [Le Quere, et al., 2007]. This reduced uptake is consistent with a number of ocean modeling studies e.g. [Lenton and Matear, 2007] and is driven by changes in wind speed bringing carbon rich water from the deep ocean to surface. This carbon rich water reduces the CO2 gradient between the ocean and the atmosphere, thereby reducing uptake.

As outlined in earlier chapters the Southern Ocean in the next 100 years is predicted to become warmer, fresher and windier. The major mechanisms by which climate change will drive changes are through ocean warming and stratification, oceanic chemistry changes and circulation changes. Simulations from coupled climate carbon models suggest that the Southern Ocean will be a region of increased CO2 uptake but the range in predicted magnitude of these increases, even from models driven with the same emission scenarios, can be large e.g. [Friedlingstein, et al., 2006]. The goal of this chapter is not to try and make a definitive prediction of how the Southern Ocean will respond to climate change but to discuss the potential impact of each of these mechanisms on CO2 uptake and sequestration and to conclude by looking at feedback of the marine carbon cycle changes on the climate system.

Future Southern Ocean carbon response

Response to increased winds

The predicted increase in winds over the Southern Ocean is expected to both reduce, in the short term, the uptake of CO2 through the changes in the upwelling and increase uptake through increased southward eddy flow in response to this increased upwelling [Zickfeld, et al., 2007]. As the Southern Ocean becomes winder in response to climate there will be a continued upwelling of carbon rich water from the deep ocean. As this carbon reaches the surface it increases the CO2 content of surface water thereby reducing the gradient between the atmosphere and the ocean, reducing CO2 uptake. This mechanism is responsible for the observed changes in Southern Ocean uptake in recent decades [Le Quere, et al., 2007]. As the wind speed continues to increase more deep water will be upwelled to the surface further reducing the gradient, however at the same time atmospheric CO2 concentrations will continue to rise. Once the atmospheric concentration of CO2 exceeds the maximum deep water pCO2 value, estimated to be 430 matm [McNeil, et al., 2007] the Southern Ocean CO2 sink will change character to become an increasing sink instead of a decreasing one. This cross-over point is estimated by coupled from recent studies to be between 2020-2030 under SRES A2 emission scenario [Matear and Lenton, 2008; Zickfeld, et al., 2008].

The predicted increases in wind speed are expected to drive an increase in northward Ekman transport and with this an associated compensating southward increasing mesoscale eddy return flow not currently resolved in coupled climate carbon models e.g. [Zickfeld, et al., 2007]. This eddy field acts to reduce the depth of pynocline, thereby increasing the volume of northward transport light actively ventilated water above the level [Mignone, et al., 2006]. The result of increased northward transport is a larger uptake of anthropogenic CO2, but this is simply a redistribution of carbon from mid-latitude and equatorial oceans and therefore although large increases maybe seen in the Southern Ocean, this does not significantly impact global uptake by the end of 2100 [Zickfeld, et al., 2007]. In addition it has also been suggested that return flow will reduce the strength of Southern Ocean upwelling, thereby bringing less carbon stored in the deep ocean to the surface. In this way reducing the pCO2 of the upwelled surface water, this return flow would strengthen the gradient between the atmosphere and ocean and hence improve the efficiency of the CO2 uptake.

CO2 response to ocean warming

The impact of increased warming and/or freshening under climate change over the ocean is increased stratification. The Southern Ocean can be described as a high nutrient low chlorophyll (HNLC) region, which means that despite the region being replete in macronutrients (nitrogen, phosphate and silicate) needed by phytoplankton to grow, the phytoplankton abundance remains low. This has been hypothesised to be due to either low levels of the light required for photosynthesis at high latitudes, or a lack of micronutrients, in particular iron, or both. As the upper ocean starts to stratify, more light is available for photosynthesis and this combined with warmer temperatures increases biological production; hence CO2 levels in surface waters decreases (Figure 4.18). Conversely as the ocean stratifies this has the effect of reducing nutrients and also reducing the efficiency of export of organic matter out of the mixed layer, reducing biological efficiency and hence CO2 uptake.

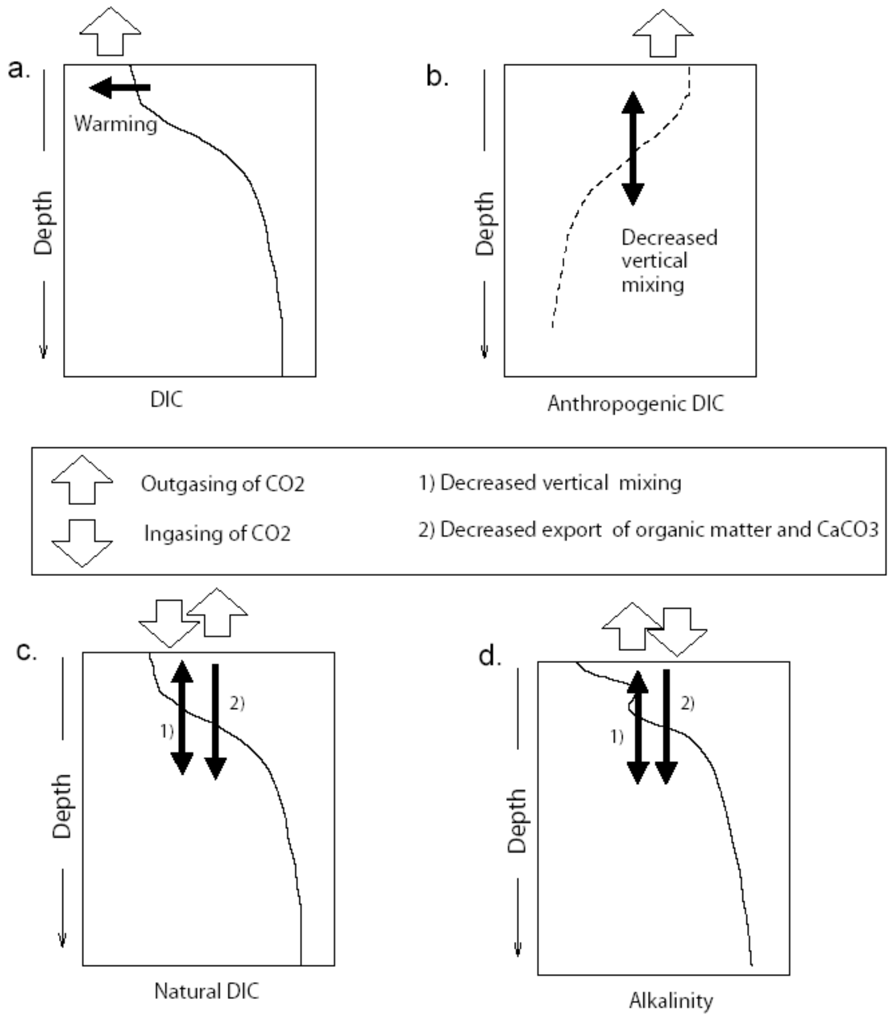


Fig 4.18. Schematic of how climate change might affect air-sea carbon fluxes. a). Sea surface warming decreases CO2 solubility and drives outgasing. b). Decreased vertical mixing prevents the penetration of anthropogenic carbon. c). Decreased vertical mixing and decreased export of organic matter have counter-effects on CO2 fluxes. d). Decreased vertical mixing and decreased export of CaCO3 also impact CO2 fluxes.

Bopp, et al. [2001] explored the relationship between the different competing process in a coupled climate carbon model and showed that in the Southern Ocean, because it is already nutrient limited, the large effect was from increased light leading to a longer, more efficient growing season and hence a 30% increase in marine productivity and export production. Consistent with this Sarmiento, et al., [2004] using six different coupled climate carbon models found that each model showed increased primary production in the Southern Ocean (Figure 4.19). In this study the magnitude of the response was high correlated to the strength of stratification; this in turn was related to changes in sea ice extent in response to climate change.

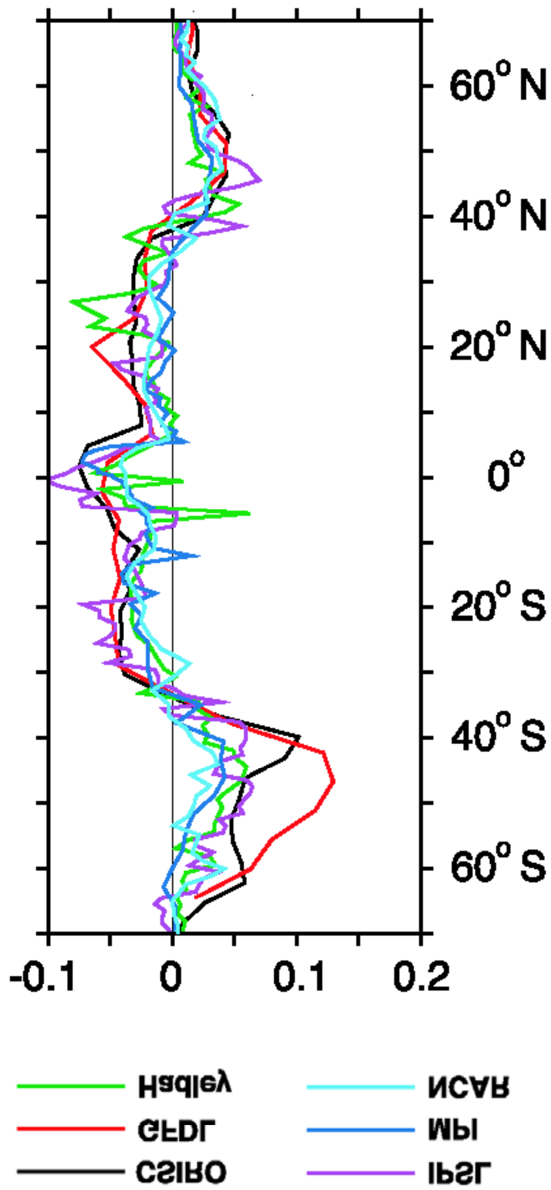


Fig 4.19 Primary Productivity (PP) changes PgC/deg calculated for the period averaged 2040-2060 using six different coupled climate carbon models [*Sarmiento, et al.*, 2004]. Note that PP changes are calculated using [*Behrenfield and Falkowski*, 1997], and changes assessed against control simulations

Increased ocean stratification has the potential to reduce uptake of anthropogenic CO2 through Southern Ocean density changes. The results of Mignone, et al., [2006], discussed above, show that the depth of the pynocline is highly correlated with anthropogenic CO2 uptake. They showed that as the pynocline shallowed (deepened) the volume of northward flowing, light, well-ventilated water reduces (increases) and hence anthropogenic uptake reduces (increases) in response to this. Therefore as density associated with stratification causes a net shallowing of the pyncocline the future anthropogenic CO2 uptake is expected to reduce also [Sarmiento, et al., 1998].

Warming of the ocean in response to climate change also reduces the solubility of CO2 in seawater. The effect of reduced solubility is a reduction of ability to take and store CO2: a warming of 1% increases oceanic pCO2 by 4% [Takahashi, et al., 1993]. These changes manifest in CO2 uptake by reducing the gradient between the atmosphere and the ocean, thus reducing the efficiency of the CO2 uptake. Various studies of future uptake all identify that the solubility effect will be very significant [Matear and Hirst, 1999; Plattner, et al., 2001; Sarmiento, et al., 1998], but they give a large range of values, reflecting the uncertainty that exists in predicting future Southern Ocean warming.

Response to increased CO2 uptake

Changes in the Revelle factor and acidification

The predicted increase in uptake of CO2 by Southern Ocean will increase the concentration of CO2 in the ocean and in turn alter the carbonate chemistry of the upper ocean through changes in the Revelle factor and ocean acidification. These changes in carbonate chemistry affect the ability of the ocean to take up more atmospheric CO2, through changes in the Revelle factor and ocean acidification.

The Revelle factor [Revelle and Suess, 1957] describes how much CO2 in the ocean will change for an increase in the partial pressure (pCO2), often termed the buffer factor. The higher the Revelle Factor, the less able the ocean is to take up atmospheric CO2. Currently in the Southern Ocean it lies between 10-15, in the next 100 years it is predicted to increase and hence reduce the efficiency of the ocean to take up anthropogenic CO2 will be reduced, Figure 4.20 The change in buffer capacity is expected to impact significantly on the ability of the Southern ocean to take up atmospheric CO2 in the future.

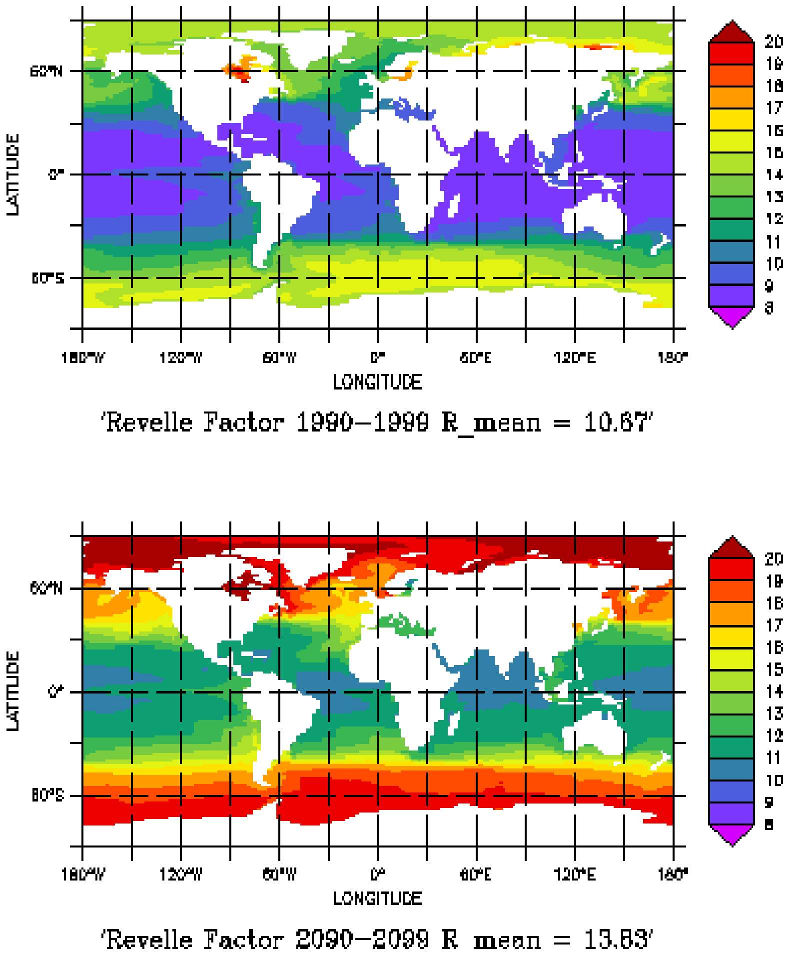


Fig 4.20. Surface values of the Revelle factor computed with PISCES model forced by the SRESA2 scenario and for the recent past and future (1990-1999 and 2090-2099).

As the ocean takes up more CO2 the ocean changes in carbonate chemistry will bring about a lowering of pH, commonly refereed to as ocean acidification. Currently the upper ocean is supersaturated with respect to aragonite and calcite, both used by some marine organisms to build calcium carbonate shells e.g. pteropods, although aragonite is a more soluble form than calcite and has a lower super saturation. As the ocean becomes more acidic the saturation states of both aragonite and calcite will be reduced. When the state drops below 1 the ocean becomes under saturated in either argonite or calcite it is no longer possible for marine organisms to use these to build calcium carbonate shells [Feely, et al., 2004]. [Orr, et al., 2005] using a suite of ocean models showed that by 2100 the saturation horizon - the point where the saturation state changes from super to under saturated - would shallow significantly in the next 100 years, with an additional pH drop of 0.3, and that by 2100 much of the Southern Ocean will be under saturated with respect to aragonite (Figure 4.21).

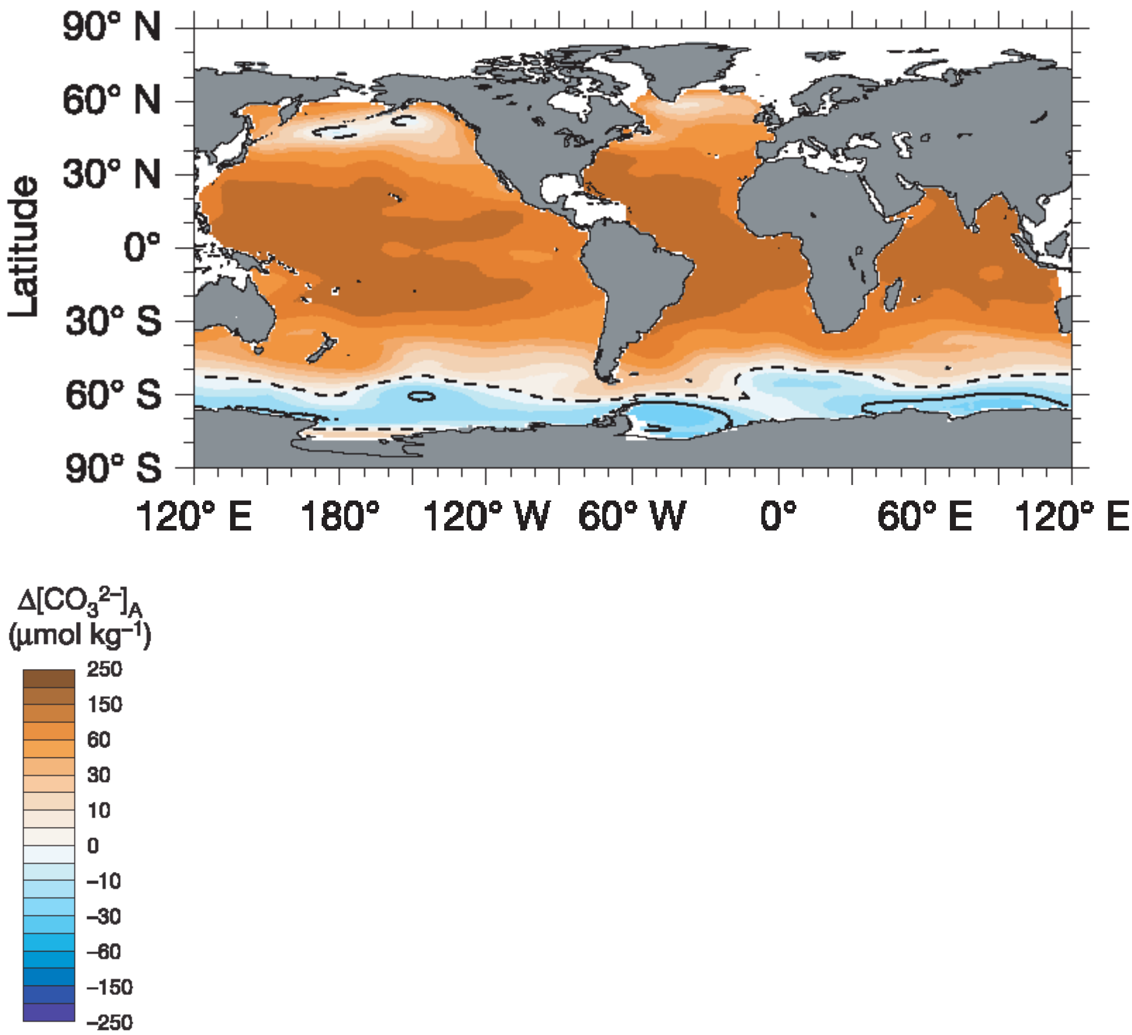


Fig 4.21. Predicted aragonite surface state in the year 2100 [*Orr, et al.*, 2005] The dashed line represents where the ocean will undersaturated with respect to aragonite

Ocean 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] this effect maybe be offset by increased CO2 uptake due to increases in availability of alkalinity [Heinze, 2004]. It is important to note that the changes in acidification (and Revelle factor) although undergoing significant longer-term changes are subject to large interannual variability. [Matear and Lenton, 2008] showed that ocean saturation horizons show large interannual variability. This large variability has the potential particularly in the case of ocean acidification to perturb the system sufficiently that far reaching changes in ecosystem can occur well in advance of that driven by climate change alone.

Studies of the future global uptake in coupled and uncoupled simulations show that the global marine ocean carbon cycle acts as a positive feedback on climate change i.e. amplifying climate change. This highlights the importance of the Southern Ocean as it is expected to act as a negative feedback in the 100 years. We note also that Southern Ocean uptake changes are vulnerable to changes in the terrestrial biosphere. Changes in uptake particularly at mid-latitudes can be very large and therefore can significantly impact response of the Southern Ocean in the future by impacting the gradient between the atmosphere and the ocean.

Conclusions

The response of the Southern Ocean to climate change remains highly uncertain and simulations from coupled climate carbon models show a large range of responses e.g. [Friedlingstein, et al., 2006]. Patterns have emerged from future scenarios that suggest that the Southern Ocean will be an increased sink of atmospheric CO2 in the future and that the recent decrease in uptake will not continue into the future. The magnitude of the total uptake is very dependent on how the ocean responds to predicted increases in ocean warming and stratification, which can drive both increases in CO2 uptake through biological and export changes and decreases through solubility changes and density changes. The increased absorption of atmospheric CO2 and upwelling of deep-water impacts the ability of the ocean to store CO2 through changes in the Revelle Factor and potentially through ecosystem changes in response to acidification.

Studies of the future global uptake in coupled and uncoupled simulations show that the global marine ocean carbon cycle acts as a positive feedback on climate change i.e. amplifying climate change. This highlights the importance of the Southern Ocean as it is expected to act as a negative feedback in the 100 years. We note also that Southern Ocean uptake changes are vulnerable to changes in the terrestrial biosphere. Changes in uptake, particularly at mid-latitudes, can be very large and therefore can significantly impact the response of the Southern Ocean in the future by impacting the gradient between the atmosphere and the ocean.

Currently our predictions of the future response are based primarily on coupled climate carbon simulations but these need to be validated. One very important goal must be therefore the development of strategies to observe and detect climate changes in the future e.g. [Lenton and Matear, 2007] to augment the ongoing and planned physical measurements e.g. ARGO [Gould, et al., 2004].

4.7 Ocean circulation and water masses

4.7.1 Simulation of present-day conditions in the Southern Ocean.

Coupled General Circulation Models (CGCMs) used in the framework of the IPCC AR4 have made numerous progresses in their representation of high latitude processes compared to earlier model versions (Randall et al. 2007). However, the Southern Ocean remains one of the region were the largest differences between different models (and between models and observations) are found. In particular, the transport of the Antarctic Circumpolar Current (ACC) through Drake Passage simulated by those models ranges from –6Sv (i.e. a westward transport) to more than 300 Sv. Only two models among the nineteen analysed by Russell et al. (2006a) are able to obtain transports that are within 20% of the value estimated from observation (135 Sv) while most of the simulated values are within 50 Sv of this estimate. Those strong biases have been attributed to different factors (Russell et al. 2006). One of the most important appears to be the representation of the zonal surface wind stress in models. Indeed, many models tend to simulate too low zonal wind stresses in the Southern Ocean or to have a maximum in zonal wind stress that is located too northward 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 (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 the export of North Atlantic Deep Water (NADW) towards the Southern Ocean, could thus play a strong role in the difference in transport between model and observations (Russell et al. 2006a).

When the temperature and salinity averaged over all the models is compared to observations (figure 4.22), the zonal mean differences are relatively small. One could however notice a tendency to have too warm and too salty water masses around 30-40°S in the depth range 500-1000m (Randall et al. 2007). This could be related to a too 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 South or by excessive exchanges with the surface. Overall, the representation of the vertical stratification for the ensemble mean is reasonable. However, this relatively good average behaviour should not mask the fact that the majority on the models either significantly overestimates the vertical density gradient or underestimates it, only a few of them being about right (Russell et al. 2006a).

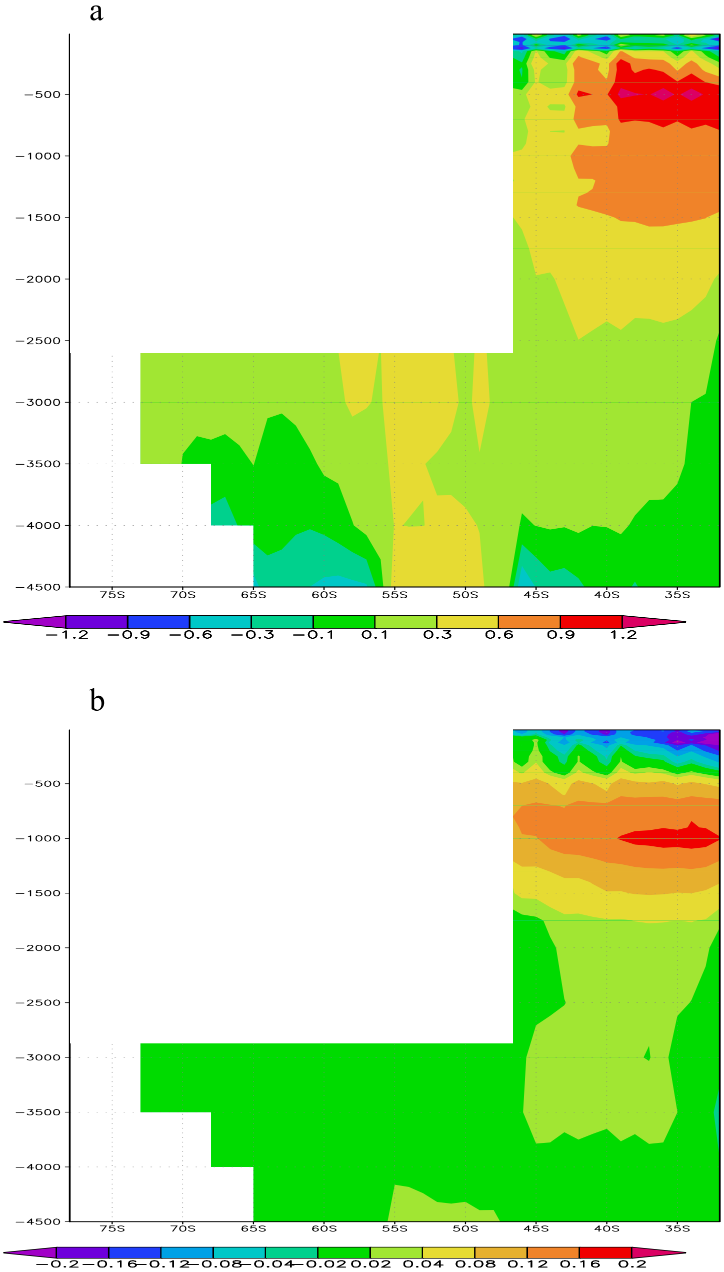


Figure 4.22. The zonal mean difference between observed (a) temperature (°C) and (b) salinity (psu) and the average of 20 CGCMS simulations for the period 1981-2000 (20C3M simulation). The observations are taken from Levitus and al. (1994) and Levitus and Boyer (1994).The 20 models for which sufficient data for the 20C3M-runs was available are CGCM3.1 (T63), BCCR-BCM2.0, CCSM3, CNRM-CM3, CGCM3.1 (T47), INGV-ECHAM4, PCM, GISS-AOM, ECHO-G, MRI-CGCM2.3.2, GISS-EH, GISS-ER, UKMO-HadCM3, IPSL-CM4, MIROC3.2 (hires), MIROC3.2 (medres), FGOALS-g1.0, GFDL-CM2.0, GFDL-CM2.1

4.7.2 **Projections for the 21st century**

An intensification of the ACC is generally simulated in response to the southward shift and intensification of the westerly winds over the Southern Ocean simulated over the 21st century. Averaged over all the CGCMs simulations performed in the framework of the IPCC AR4, the increase in transport projected for the end of the 21st

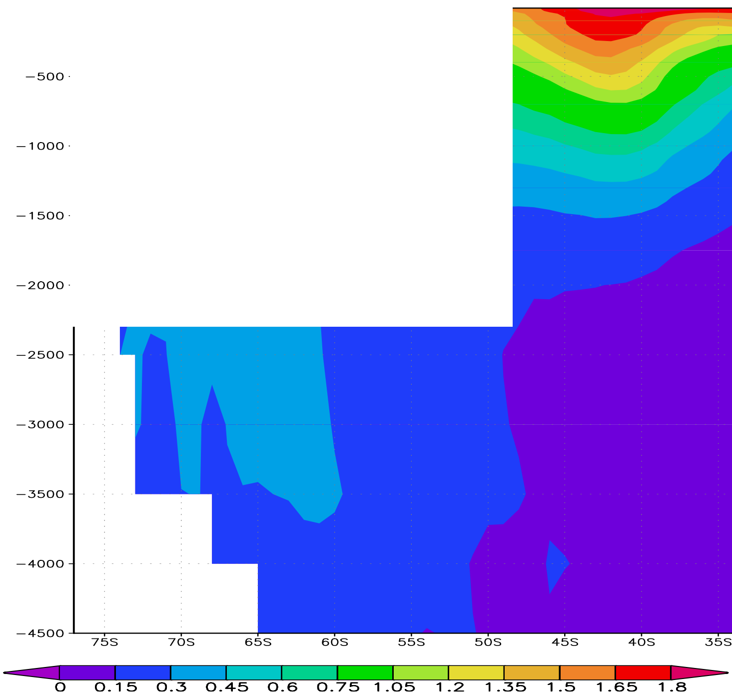
century reaches a few sverdrups at Drake passage. The enhanced winds induce in addition a small (less than1° in latitude on average) but robust southward displacement of the core of the ACC as well as a stronger northward Ekman transport close to the surface and a return southward flow around 2000 m in the models (e.g., Saenko et al. 2005, Fyfe and Saenko,2005, 2006, Lefebvre and Goosse 2007). Furthermore, nearly all the models simulate regional changes in the horizontal oceanic circulation but the patterns of the projected changes vary significantly between the models. One exception is probably the Ross Sea where a large number of models simulate a cyclonic anomaly at the end of the 21st century compared to the late 20th century (Lefebvre and Goosse, 2007).

The CGCMs reproduce quite well the mid-depth warming observed during the second half of the 20th century, in particular if the effect of volcanic eruptions is taken into account in models (Fyfe, 2006). During the 21st century, this warming is projected to continue, reaching nearly all depths when averaged over the ensemble of models (Fig. 4.23). However, close to the surface, the warming of the Southern ocean during the 21st century is weaker than in other regions. This is partly related to the large heat storage by the ocean which remove 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 can be found. Furthermore, the vertical stratification increases in the majority of the models during the 21st century because of the surface warming as well as of the freshening southward of 45°S (Figure 4.23). This surface salinity decrease, which has a dominant impact on the density changes at high latitudes, is due on the one hand to the increase in precipitation at high latitude. On the other hand, 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 the 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 thus to the moderate surface temperature increase simulated in the Southern Ocean during the 21st century. In addition, changes in the stratification alter the isopycnal diffusion in models by modifying the slope of the isopycals, resulting in a reduction of the heat transport towards the surface by this mechanism (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. More generally, as a result of the surface density decrease, the ocean ventilation and in particular the formation of Antarctic Bottom Water decreases 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 the uptake of carbon dioxide, more carbon remaining in the atmosphere during the 21st century compared to a case with constant ventilation of the Southern Ocean (e.g., Sarmieto et al. 1998, Matear and Hirst 1999).

However, one should pay attention to the fact that vertical mixing does certainly not decrease over the 21st century at all locations of the Southern Ocean in all models. Increased wind stresses could result in deeper mixed layer in some areas. The decrease in ice extent could induce stronger cooling in winter that leads to stronger mixing at the new ice edge. Shift 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 Menéendez 2006).

Furthermore, the ocean ventilation could be enhanced because of the surface divergence induced by the increase in the wind stress projected during the 21st century. Russel et al. (2006b) argue that this effect could be strongly underestimated in some models because of westerlies located too northerly 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 a much lower heat (and carbon dioxide) storage in the Southern Ocean during the next centuries than 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.

a



b

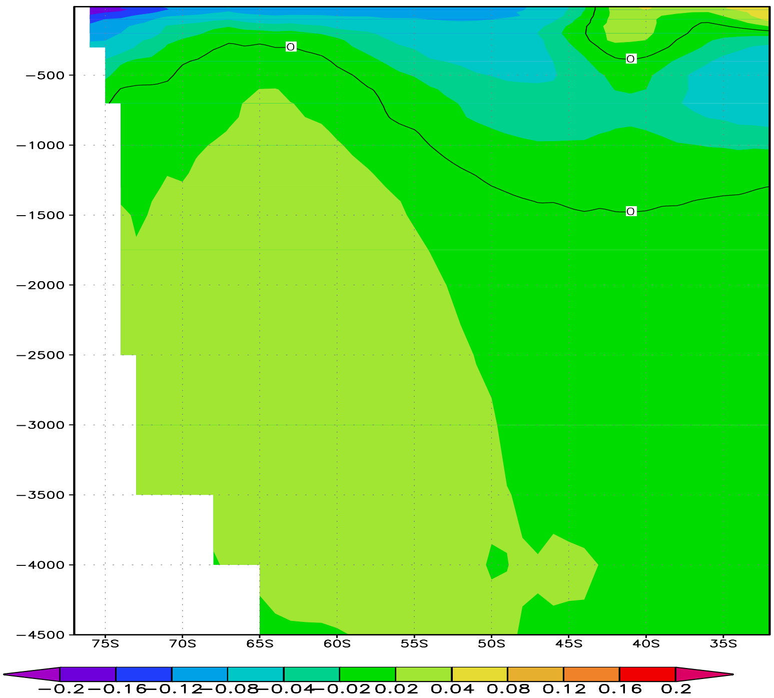


Figure 4.23. Difference in (a) zonal mean temperature (°C) and (b) salinity (psu) between 2081-2100 (scenario SRES A1B) and 1981-2000 (simulation 20C3M) averaged over 17CGCMs simulations. The 17 models for which sufficient data for the 20C3M-runs and the SRESA1B were available are CGCM3.1 (T63), BCCR-BCM2.0, CCSM3, CNRM-CM3, CGCM3.1 (T47), INGV-ECHAM4, PCM, GISS-AOM, ECHO-G, MRI-CGCM2.3.2, GISS-EH, GISS-ER, UKMO-HadCM3, IPSL-CM4, MIROC3.2 (hires), MIROC3.2 (medres).

4.7.3 **Long-term evolution**

Only a few studies have addressed the evolution of 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 in a robust manner 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) during the simulation. 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 21st are less operand during the 3rd 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 the ocean currents. In a first step, the general decrease found in AABW formation over the 21st 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 as in Hirst et al. 1999 and 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 allow deep-water formation to start again or to be enhanced (Bi et al. 2001). The disappearance of sea ice could also plays a role because very strong atmosphere-ocean interactions takes 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 strong 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 this 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 form the melting of the Antarctic ice sheet, which could strongly influence stratification and deep water formation in the Southern Ocean at this time-scale, is included in none of those simulations, increasing our uncertainties in the long-term evolution of the system.

4.7 Sea ice change over the 21st century

The average of the CMIP3 models sea ice extent compares well with observations. Although there is a large inter-model spread (Arzel *et al.*, 2006; Parkinson *et al.*, 2006).

According to the SRESA1B scenario, over the 21st century show the annual average total sea-ice area is projected to decrease by 2.6×10^6 km^2, which is a 33% decrease compared to current simulated values (Bracegirdle *et al.*, 2007). There is strong consensus for an Antarctica-wide decrease amongst the models, with an inter-model standard deviation of 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 SRESA1B projections and found 21st century deceases of 34% for sea-ice volume and 24% for sea-ice extent. The reductions of sea ice area are larger because of reductions of the projected sea ice concentration.

Most of the ice retreat occurs for winter and spring when the sea ice extent is largest (Figure 1.24). The amplitude of the seasonal cycle of sea ice area will therefore decrease. The smaller amount of seasonal ice melting/freezing will affect ocean due to changes in processes such as brine rejection.

In contrast to the Southern Hemisphere, the amplitude of the seasonal cycle of sea ice extent in the Northern Hemisphere is projected to increase, with the largest reductions of sea ice extent in late summer. Another contrast with the Northern Hemisphere is in the projections of sea ice volume. In the Northern Hemisphere the projected decrease of sea ice volume is more than twice that of sea ice extent, whereas in the Southern Hemisphere the projected decrease of sea ice volume is only 40% larger than sea ice extent. The difference might be due to the greater amount of thicker perennial sea ice over the Arctic (Arzel *et al.*, 2006).

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.

There is strong confidence in the Antarctic-wide decreases of sea ice extent. However, at a more regional level decreases of sea ice are less significant. 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 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 4.22). However, at the scale of one grid point the confidence of a reduction of sea ice over the 21st 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).

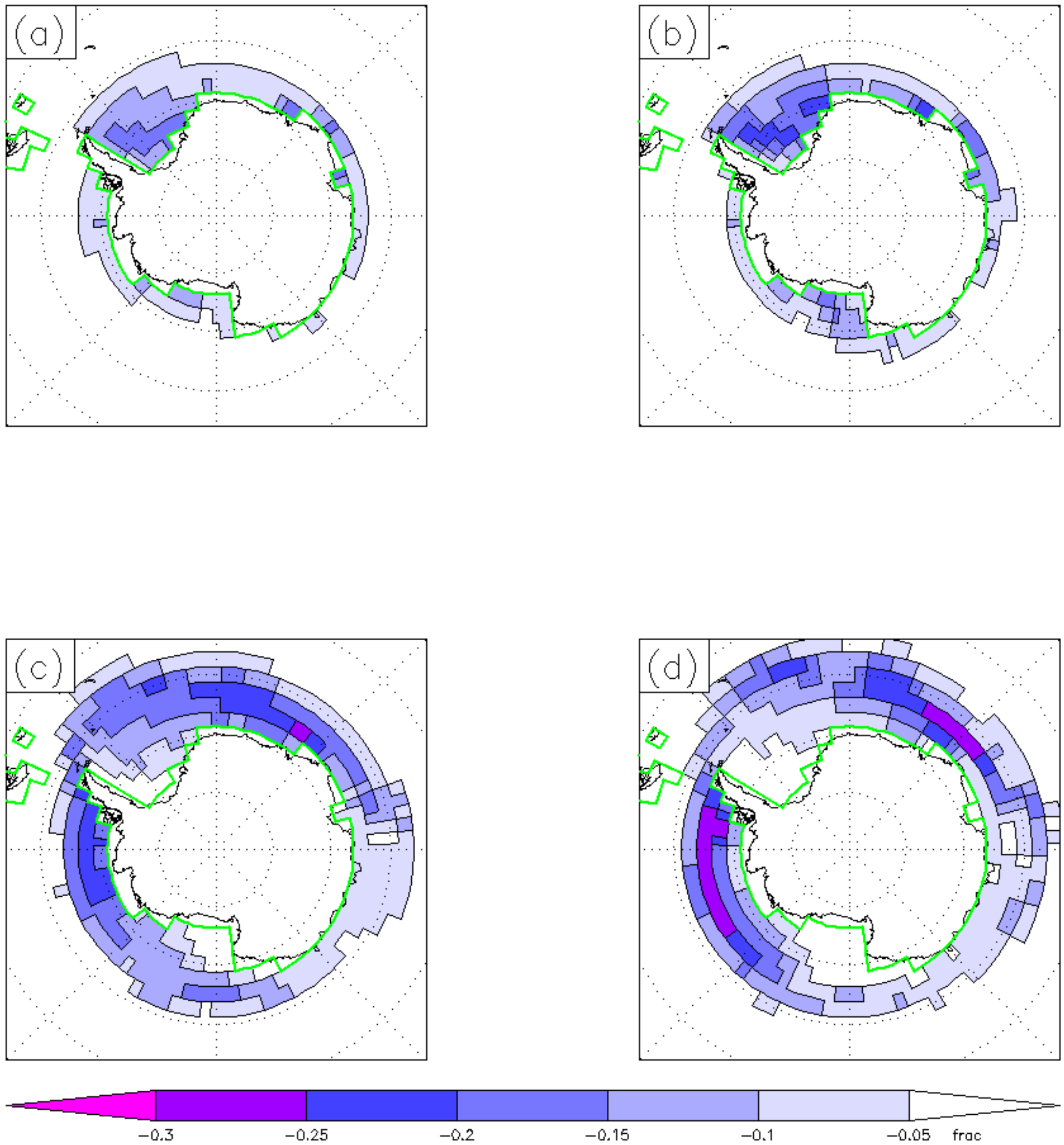


Figure 1.24. 21st century sea ice concentration change for (a) DJF, (b) MAM, (c) JJA and (d) SON. Difference between 2080-2099 mean and 2004-2023 mean.

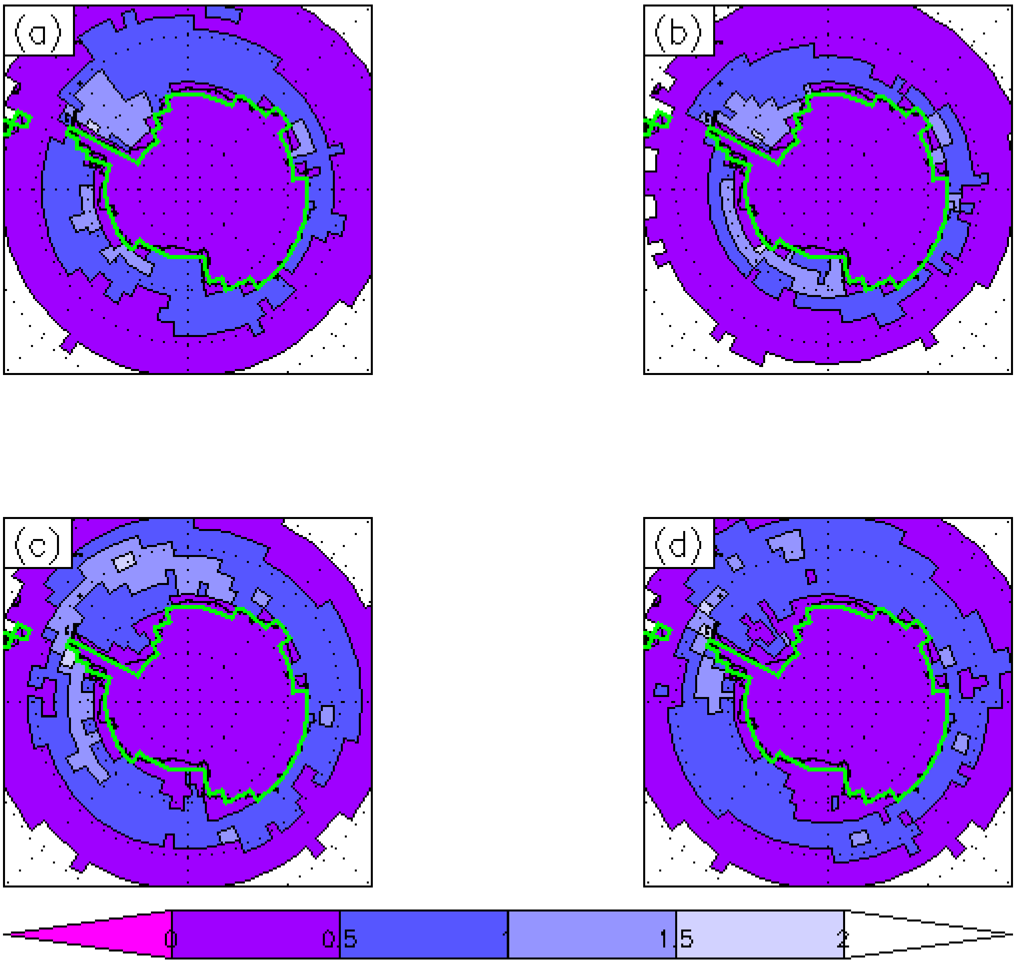


Figure 4.22. Signal to noise ratio of projections of sea ice reduction.

4.9 Predictions of how Antarctic permafrost may evolve over the next century

Given the possible warming scenarios in Antarctica (Sec. 5.2), the following changes in permafrost and associated land features and processes could occur over the next 100 years:

- Reduction in permafrost area
- Subsidence of ground surface
- Changes in hydrology and mass movements
- Risks to infrastructure
- Feedbacks and interactions

Reduction in permafrost area

Permafrost in Antarctica is restricted to ice-free areas, which account for only 0.32% (49,500 km^2) (Sec. 1.10). The areas most susceptible to permafrost degradation lie within the zones of discontinuous and sporadic permafrost, which includes the northern Antarctic Peninsula and the South Shetland and South Orkney Islands (ca. 60-69ºS, 43-70ºW) and possibly coastal areas in East Antarctica. However, the total ice-free area of these regions is probably less than 5,000 km^2.

Subsidence of ground surface
Although we do not envision a major reduction in permafrost area over the next 100 years, melting of ice wedges in polygons may lead to thermokarst, which refers to subsidence from melting of ground ice. Areas that are particularly susceptible to thermokarst occur in coastal areas and feature extensive ice-wedge polygons, including Casey Bay near Molodezhnaya Station (70.5ºS, 12ºE), the Pennell-Borchgrevink Coasts in North Victoria land (70.5-73ºS, 165-171ºE), the Scott Coast in the McMurdo Sound area (74-78ºS, 165ºE), and throughout the Antarctic Peninsula and its offshore islands (55.5-72ºS, 45-70 ºW). The total ice-free area that may be impacted is approximately 15,000 km^2.

Impact on hydrologic and geomorphic processes
Although climate warming has not been reported in continental Antarctica (Doran et al., 2002), there have been dramatic changes in hydrologic and geomorphologic processes as a consequence of unusual summer warming events during the 2000s. These events could represent changes should the climate warm in interior Antarctica. The mean summer (December and January) temperature during the period 1994 to 2003 in the MDV was -0.19º C, and there were 30 days per year in which the mean daily temperature exceeded 0ºC. In 2001-2002 the summer temperature was 1.5ºC and there were 43 days in which the mean daily temperature exceeded 0ºC, which 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 level of the preceding 9 years for a 4-year period (Barrett, MCM-LTER, unpublished). In addition, observational evidence suggests that extreme events in the MDV may alter subsurface flow 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.

Risks to infrastructure
There are approximately 80 year round and summer bases in Antarctica, of which about 90% are in areas that are sensitive to thermokarst and mass wasting. Therefore, the effect of climate warming on melting of ice wedges and ground ice should be of concern to the Council of Managers of National Antarctic Programs (COMNAP). Examples of unusual warming at McMurdo Station on road failure and undercutting of pilings for utilidors can be found.

Feedbacks and interactions
Unlike the arctic and central Asian regions, there is minimal concern about feedback to climate change through the release of trace gases such as carbon dioxide and methane. The reason for this is that Antarctic soils, particularly those in interior regions, generally contain very low concentrations of organic C (<0.05%).

Recommended future research
- Expand GTN-P network in Antarctica to include monitored boreholes in the Sør Rondane Mountains (72ºS, 25ºE), Enderby Land (66.50ºS, 52ºE), the Vestfold Hills (70ºS, 73ºE), Wilkes Land (66.30ºS, 110ºE), and additional sites along the Antarctica Peninsula and its offshore islands.

- Expand CALM-S network to included sites along the Palmer Archipelago, the Schirmacher Hills (Drønning Maud Land), the Bunger Hills (Enderby Land), and the Vestfold Hills.
- Apply the PERMAMODEL to predict changes in permafrost distribution under different climate change scenarios, particularly along the Antarctic Peninsula and in maritime East Antarctica.
- Establish a central location for management of permafrost, active-layer, and ground ice data.

4.10 Marine biology

4.10.1 Biological response

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 are the level at which natural selection works and therefore other responses ultimately depend on the 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 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 rate of environmental change and of phenological response. A match (mismatch) between rates will be likely to occur with a stable (decreasing) optimum mean population 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 become essential to unequivocally identify and characterise successful responses of organisms to change.

Organisms will depend on phenotypic plasticity to cope with environmental change, or will adapt through changes in gene frequencies between generations which, at the population level, are known as microevolution. This irreversible mechanism 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 icebers (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 vital rates. 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 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 guild studied, the fitness of three populations -emperor penguin, snow petrel and southern fulmar- has also been recently evaluated (Jenouvrier et al. 2005). These species show a stable population growth to date, and only southern fulmars show a significant phenological change, which is the delay in date of arrival to the nesting grounds. In southern fulmars, however, approximately 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. 2005). 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 affect sometimes 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 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 season duration variability (Jenouvrier et al. 2005).

The majority of 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 impact of climate is likely to be the change in abundance, quality or predictability of the food resources, which result from food web modifications related to changes in the physical environment. The increase in climate-related food shortages has been the cause of 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 20th century. The recent rapid increase in ENSO-related variability, with more frequent extreme events causing frequent food shortages, have lead to a local decline in female fur seal fitness, and a loss of buffering of survival and propensity to breed in adult breeders to the current 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 type of vital rates increases the long-term fitness tends to decrease, causing loss of buffering against the environment (Morris and Doak 2004).

In the fur seal case, however, changes in phenology were not apparent, which suggests that the loss of predictability of the food supply and its shortage, 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 start near one of the Antarctic regions with most rapid warming, the Antarctic Peninsula. Therefore, it is likely that the loss of buffering against the environment occurs in species most sensitive to loss of their critical habitats, but also those most affected by changes in predictability 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. The 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 which case, phenotypic plasticity may not be enough to ensure persistence, and it is likely that most successful species replace those more sensitive to loss in critical habitats (e.g. Forcada et al. 2006). Alternatively, microevolution may help long-term adaptation, but there is little hard evidence that it is occurring already.

All organisms, both terrestrial and marine, are susceptible to environmental changes. Many polar species are vulnerable, because they are specialised to cope with harsh conditions; in addition, they must face human-induced stresses. As it is common in evolution, if a species fails to adapt to climate-change requirements exceeding the internal flexibility limits or fails to migrate, its fate will be extinction. The vulnerability of Antarctic benthos (Peck 2005) strongly suggests that, compared to temperate environments, long-term survival of these organisms under rapid climate change involves far higher risk.

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 have poorer physiological capacities to deal with change than species elsewhere (Peck 2005, Peck et al. 2007, Pörtner et al. 2007). They die when temperatures are raised by only 5–10°C above the annual average, and 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 is often narrow, thus the efficiency of this ability is poor.

(ii) *Adapt to the new conditions and alter the range of biological capacity*. This strategy depends on magnitude and rate of change, and aquatic habitats change temperature at a far slower rate than terrestrial ones, possibly creating fewer adaptation problems to most fish species. The ability to adapt, or evolve new characters to changing conditions depends on many characters 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, Brey & Clarke 1993, Peck 2002) and develop at rates often x5–x10 slower than similar temperate species (Peck 2002, Peck et al. 2006a). They also live to great age, and exhibit deferred maturity (Peck et al. 2006b). 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 2 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 in the 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 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 locomote 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 (Bosch et al. 1987, Stanwell-Smith & Peck 1998, 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 mollusks. 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. Thus in a warming environment there are fewer places to migrate to. On all three major criteria Antarctic benthic species appear less capable than species elsewhere of responding to change in ways that can enhance survival.

Although the absence of wide latitudinal gradients in the Antarctic continent minimises the advantage of migration, it highlights the importance of the sub-Antarctic as a critical research area, largely populated by eurythermal fish. These 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.

Adaptations and physiological limits

Specific adaptations to the Antarctic marine environment

Most invertebrate species have body fluids isosmotic with seawater and do not, therefore, freeze unless the water around them freezes. This 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 antifreeze compounds year-round to avoid freezing (DeVries, 1971, 1982; Clarke 1998; Cheng & 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), and that the trait has evolved on several separate occasions (Chen et al. 1997a,b; Cheng &

Chen 1999). 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 & 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 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). However, other factors may also be limiting. These include the possibility that protein synthesis at low temperature may be more difficult, and costly, and this 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. Recently, Hofman 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 with some invertebrates expressing heat shock genes when temperatures are elevated to over 10°C but not below (Clark et al. 2008a), whereas other invertebrates appear to be similar to the fish data reported by Hofman et al. (2000) in having no HSP response to elevated temperature (Clark et al. 2008b). Surprisingly in this context of laboratory studies, the limpet *Nacella concinna* has a heat shock response in wild intertidal populations when emersed, even though foot temperatures do not exceed 3°C (Clark et al. 2008c). After 22 days of acclimation to 4°C there was little or no production of heat shock proteins, and no stress-induced chaperone molecules could be detected on short-term warming. It is not known whether the lack of a heat shock response 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 the terrestrial condition.

Physiological limits

Antarctic marine species are highly stenothermal, with the vast majority having experimental upper lethal temperatures between 5°C and 10°C (Somero & DeVries 1967; Peck & 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 & Davison 2005; Seebacher et al. 2005; Podrabsky & Somero 2006; Jin & 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. obs.) 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. Recently 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).

4.10.2 **Adaptive Evolution - marine**

In this section, much of the information has been drawn from the 2007 IPCC report (Anisimov et al. 2007).

Although the impacts of climate change in the 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. A summary of the changes observed over the 20th century was supplied by Anisimov et al. (2001).

The recognition of the important role of the polar habitats in global climate changes has recently awakened great interest in the evolutionary biology of the organisms that live there, as well as in the increasing threat of loss of biological diversity and depletion of marine fisheries. Recent evidence has indicated that global change is already affecting the physiology and ecology of some species (Hughes 2000; Walther et al. 2002). Also natural variations might be responsible for some of the observed trends, regional and short-term climatic variations seem to be more frequent and intense in recent years, and human-induced climate change is very often the most likely cause.

The impact of Global Climate Change in polar environments

Given the differences in topography and glaciation history, the Antarctic and Arctic Oceans may respond differently to climate change, but both habitats appear sensitive. Small temperature differences 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. At present, despite climate change, the polar regions offer an important opportunity to study species biodiversity in relatively undisturbed environments. In absolute terms, this rare advantage mostly refers to the Antarctic; national territorial claims are still not accepted, 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. Thus, the main direct influence on the Antarctic marine ecosystem comes from global climate sources.

The Arctic lies between North America, Greenland, Europe and Asia beyond the Arctic Circle and its geography is very complex. Unlike in the Antarctic, other human-induced impacts add to those due to climate change; the sea is almost completely enclosed and influenced by large human populations in extensive colonised land areas, and by industrial activities. The marine ecosystem is strongly influenced also by local sources, with far greater anthropogenic impact. Pollutants generated in Europe, Asia and North America are efficiently carried into the Arctic by atmospheric and oceanic circulation processes, and into the ocean via the huge rivers of northern Europe; some sea areas contain significant amounts of radionuclide pollution. National sovereignties bring about further difficulties on political grounds.

Impacts and concerns

Climate change is already having significant impacts on marine and terrestrial systems (Hughes 2000; Walther et al. 2002), and will continue to influence biological diversity. Many species are susceptible to this environmental change, and those of the marine environment are particularly vulnerable, even though warming is more evident in the air than in the sea. The northern and southern regions are undergoing rapid environmental changes, in many instances due to the combined effects of natural climate change and human activity.

Present patterns of biodiversity and distribution are a consequence of processes working on both evolutionary and ecological timescales. Among the ecological factors controlling distribution and biodiversity of the modern polar ichthyofaunas, the most important are temperature, ice cover, oxygen, light, UVB and wind. Besides being largely interconnected, 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 UVB levels. Although more limited, illegal fishing and introduction of alien species are further threats. In contrast to these widespread phenomena, pollution and visitor pressure are causing only local effects on 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 impact on primary producers may affect consumers and higher levels in the food web.

The following is a pertinent example. The western side of the Antarctic Peninsula is currently subject to one of the fastest rates of climate change on the planet (Cook et al. 2005), leading to reduction of annual mean sea-ice extent (reviewed in Clarke et al. 2006). There are indications that *Pleuragramma antarcticum*, a key fish species of the trophic web and whose reproduction is closely associated to sea ice, has disappeared, undergoing replacement by myctophids, a new food item for predators (M Vacchi, personal communication; WR Fraser, Regional loss of Antarctic Silverfish from the western Antarctic Peninsula food web, in preparation). This event is suspected to be caused by seasonal changes in sea-ice dynamics, upsetting reproduction processes.

However, except in the Peninsula, temperature trends in Antarctica are not straightforward. In some regions, in fact, cooling has been recorded, at times accompanied by clear local impacts, such as in the lakes and soil of the Dry Valleys. Moreover, there is no evidence of a continent-wide "polar amplification" similar to that predicted in the Arctic. Sea-ice duration has increased in the Ross Sea, and decreased in the Bellingshausen and Amundsen Sea, reflecting trends in atmospheric temperature. The extent of sea ice has seen no overall change over 1973-1996, but a substantial loss has been projected. Acidification is occurring in the ocean, changing the chemistry, affecting some organisms (thereby altering the availability of nutrients) and reducing the uptake of carbon dioxide from the atmosphere. For many species, uncertainty in climate predictions leads to uncertainty in projecting impacts; however, warming and winter sea-ice decrease will affect reproduction and growth of fish and krill, leading to population reduction and distribution changes.

In areas which experienced warming, increases in shallow-water sponges and their predators, and decreases in krill, Adélie and Emperor penguins and Weddell seals have been recorded (Ainley et al. 2005). Major changes have also occurred in terrestrial ecosystems. The reduction in krill biomass and the increase in abundance of salps (pelagic organisms) may also be linked to sea-ice regional changes, which in turn may underlie the recent changes in the demography of krill predators, e.g. mammals and birds (Fraser & Hofmann 2003). Negative impacts on the local biota have occurred in some sub-Antarctic islands. Signy Island has witnessed the explosion of the fur-seal population, due to decreased ice cover and increased areas available for resting and moulting, with deleterious impacts on vegetation.

The ecological and evolutionary consequences of global climate change are a great concern also for the Arctic. All sub-Arctic regions are highly ecologically sensitive, implying that anthropogenic warming will affect habitat and species resilience and may potentially induce dramatic changes in community dynamic and structure, with compelling social and economic implications. A very partial overview of recent reports is summarised below.

Although the temperature record across the Arctic regions is not complete, warming appears to be concentrated in the last century (Moritz et al. 2002). During the 20th century, air temperatures increased by up to 5°C, warming being about 1°C per decade since 1980 (McBean et al. 2005). Warming is projected to be in the range 2-9°C by 2100. Together with the Antarctic Peninsula, Arctic regions appear to be the most rapidly warming areas of the world (Turner et al. 2007). Climate change has completely changed the life style, entirely dependent on marine food, of the human population living in the Bering Sea area, raising both scientific and political concerns

(Smetacek & Nicol 2005). Fish stocks in the Bering and Barents Seas (among the richest in the world) have fluctuated significantly in the last decades as a result of changes in fishing pressure in response to climate conditions, such as storm frequency. It is realistic to expect that some of the variations documented in the Arctic may show similar trends in the Antarctic. Climate models have recently indicated that the retreat of Arctic sea ice (it has declined ca. 7% since 1978) is unrelated to natural climate variability and is caused by anthropogenic impact (Johannessen & Miles 2000; Hasselmann et al. 2003). According to current models, by the end of the century the Arctic Ocean might be essentially ice-free during the summer (Zhang & Walsh 2006). Similar to the Arctic, the reductions in the extent of cover and thickness of sea ice in the western side of the Antarctic Peninsula are dramatic, and potentially devastating to some species. The warming of sea surface is accompanied by increase in phytoplankton in cooler regions (which might actually be beneficial to most of the commercial fish stocks) and decrease in warmer regions (Richardson & Schoeman 2004); similar variations occur on seasonal cycles of micro-organisms and invertebrates. If the sea-ice cover continues to decrease, such differences in response to changes will impair predation processes and will impact on community composition and levels of primary and secondary producers. For instance, marine ice algae would disappear due to loss of habitat, and this may cause a cascade effect to higher trophic levels in the food web: zooplankton feed on algae; many fish (e.g. cod) feed on zooplankton, and sea birds and mammals in turn feed on fish.

A "*polar amplification*" of the anthropogenic warming, predicted and strongly supported by recent accelerations of glacier retreat (Dyurgerov & Meier 2000), sea-ice thinning and permafrost degradation (Oechel et al. 2000; Romanovsky et al. 2002; Lemke et al. 2007) in the Arctic, is thus already evident in the Peninsula.

Observations on polar lakes have indicated that high-latitude Arctic lakes are extremely sensitive to climate change, because even slight warming results in decreased ice cover, and only the coldest sites may retain ice cover throughout the summer. Smol et al. (2005) analysed 55 palaeolimnological records from Arctic lakes and showed that many Arctic freshwater ecosystems have experienced dramatic and unidirectional regime shifts within the last 150 years. Similar changes have been documented in Antarctic maritime lakes (Quayle et al. 2002; Hodgson et al. 2004).

The responsiveness of species to recent and past climate change does raise the possibility that human influence may cause a major extinction event in the near future for some vulnerable species. Thomas et al. (2004) have performed a complete analysis of the extinction risk from climate warming. The alarming conclusion, namely that many of the extant species could be driven to extinction by climate change over the next 50 years, is compelling and provides important arguments urging new policies aimed at reducing the impact of warming due to human activity. Indeed, at present, the awareness of global warming has prompted scientists and governments to consider whether climate change in combination with its strong acceleration due to human influence might cause extinction of species inhabiting environments covered by sea ice. Although the available models and data sets as yet do not justify such a prediction for Antarctica, such an event may be averted only by resorting to concerted, multidisciplinary, international efforts aimed at protecting the polar marine life.

Cold-adapted organisms and climate change: pathways of research

The development and growth of research on adaptations to polar environmental conditions are relatively recent events, and have originated from the increasing number of nations which have engaged in polar science. Adaptations of the polar ichthyofauna in response to environmental change are commanding attention as biodiversity and climate change are increasingly considered in a global context. There is ample evidence that recent climate changes already cause physiological problems to a broad range of species, drive evolutionary responses (Thomas et al. 2001, 2004; Walther et al. 2002), and produce micro-evolutionary changes in some species (Rodriguez-Trelles & Rodriguez 1998). But species do not live in isolation and it is necessary to evaluate responses at community and ecosystem levels. Ecologists and physiologists are thus faced with the difficult challenge to predict the effect of warming, not only on individual species, but on the community as a whole. For instance, ice-shelf collapse increases the number of icebergs, increasing the impact by scouring on benthos biodiversity as well as on the food web. Although the changes in sea temperature are as yet small, increasing 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. 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? Another key question in Antarctic biology is whether climate change will result in either relaxation of selection pressure on genomes, or tighter constraints and ultimately extinction of species and populations.

Only recently, essentially due to commercial implications, has the Arctic habitat been receiving extensive attention. The urge to study the molecular mechanisms underlying fish thermal adaptations and biodiversity also at the North Pole is becoming stronger as efforts to understand cold adaptation widen and shift to a more analytical phase. The Arctic offers a remarkable opportunity to develop comparative studies on evolutionary differences among 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 the evolutionary pressures and provide insight into gene selection. Climate change will affect every aspect of an organism's biology, from cellular physiology and biochemistry to food web and habitat. Organisms must alter their physiology/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). In spite of their fundamental physical and biological differences, there is the need to establish links between the Northern and Southern regions. Understanding the adaptation-response mechanisms in species living in both polar habitats will offer an ideal background for extrapolation to lower latitudes.

In addition to adaptation, other key themes include life cycles (tactics and strategies to respond to environment features), micro-evolutionary processes driven by anthropogenic impact, interactions between changing abiotic conditions (e.g. temperature, UVB) and biotic responses, modelling of interactions between environmental change and organism responses (to facilitate predictions of change), development of conservation policies.

The urgent and challenging agenda for the next decade will be to incorporate thinking along the physiological/biochemical viewpoint into a field which is currently labelled as evolutionary biology. Such an integrative approach can provide answers to the question of how Antarctic and Arctic fish will respond, and whether they will be able to adjust, to ongoing Global Warming, already in full action in the polar regions. A future challenge pertains to analysing the ability of polar fish to develop repair mechanisms to changes induced by a wide variety of natural and anthropogenic processes, in the general framework of species and ecosystem responses, as well as the ways in which responses feed back to influence these processes.

The synergy of disciplines is essential. Studying the response of (micro-)evolutionary processes to changes in selection pressures needs collaboration with physical sciences and modelling. Analyses of adaptive evolution across the biological organisation from molecules to species must integrate physiology, biochemistry/molecular biology, morphology, taxonomy, biogeography, ecology, ethology. Investigating changes in the physical environment that have driven evolution over geological time requires collaboration with palaeoscience, geophysics, glaciology and oceanography. Statistical and molecular genetic approaches are needed to monitor the biodiversity. This multidisciplinary approach will allow to establish links between tectonics, climate evolution, 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.

Concluding remarks

The largest challenge facing humankind is the management of the Earth System to ensure a sustainable future. To this end, understanding of the functioning of the Earth System in the context of both natural and anthropogenic change is essential. The polar habitats and their biota are an instrumental part of the Earth System, not only influencing the pace and nature of environmental change, but also responding to it in an integrated system of biologically modulated connections, which project into temperate areas.

The Antarctic offers an immensely valuable, regionally focussed approach. Its ecosystems offer examples of how both structure and function have evolved to cope with an extreme environment, and the likely responses of species and ecosystems to change induced by a wide variety of natural and anthropogenic processes, as well as the ways in which their responses feed back to influence these processes.

In 2004 the Scientific Committee on Antarctic Research (SCAR), fully aware of the problems inherent to climate change, launched the 8-year international programme "Evolution and Biodiversity in the Antarctic: the Response of Life to Change" (EBA). It 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 major intention is to provide a platform for interactions amongst disciplines and researchers that are essential to understand the role of biodiversity in the Earth System and its responses to change, by offering the Antarctic context, and establishing crosslinks with the Arctic, enhancing our ability 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.

Together with another international programme that highlights the importance of the sub-Antarctic, named "International Collaborative Expedition to collect and study Fish Indigenous to Sub-Antarctic Habitats" (ICEFISH), EBA has been selected by ICSU/WMO (International Council for Science of UNESCO/World Meteorological Organisation) as "Lead Project" for the International Polar Year (IPY 2007-2008). IPY has taken place half a century after the International Geophysical Year (IGY, 1957-8), to which we owe countless outstanding milestones of Polar Science. Such timely event meets the increasing concerns expressed by the Antarctic Treaty System regarding the responses of Antarctic environments to natural and anthropogenic disturbances, and the 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.

New information, including the choice of suitable target species, long-term data sets and the concerted efforts from international multidisciplinary programmes, will help us 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 which are essentially driven by anthropogenic contributions.

4.10.3 **Invertebrates**

It has been well documented that decadal-scale variations in the coupled ocean-atmosphere system make 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, for example, associated with shifts in the geographical distribution of ectothermic animals along a latitudinal cline or with pole-ward or high-altitude extensions of geographic species ranges (Walther et al. 2002, Parmesan & Yohe 2003, Root et al. 2003). Temperature means and variability as associated with the climate regime can be interpreted as major driving forces setting the large scale biogeography of marine water breathing animals. These relationships lead us to expect significant effects of climate warming on polar including Antarctic ecosystems.

The marine Antarctic is currently characterized by cold temperatures close to freezing in several areas with lowest temperature variability at highest latitudes (Clarke 1998, Peck 2005). Accordingly, Antarctic marine ectotherms live at the low end of the aquatic temperature continuum and are considered highly stenothermal, i.e. these animals are specialized on narrow thermal windows (Somero & De Vries 1967, Peck & Conway 2000; Somero et al. 1996, 1998; Pörtner et al. 1999, 2000; Peck et al. 2002). Changes in temperature means and variability in relation to the climate regime should thus exert key influences in shaping survival and functional adaptation to temperature. However, the question currently arises to what extent stenothermy in Antarctic species and phyla has been overestimated and how species may differ with respect to the level of stenothermy and their capacity to thermally acclimate. Recent data show the capacity of Antarctic fish to undergo thermal acclimation and shift physiological characters accordingly, as verified in a zoarcid (Lannig et al. 2005) and a notothenioid (Seebacher et al. 2005).

A wider comparison of the mechanisms characterizing thermal intolerance between and within species of marine invertebrates and fish led to the development of a unifying physiological concept of thermal limitation and adaptation. The capacity limitation of oxygen supply mechanisms characterizes the first line of thermal intolerance in animals and restricts performance in behaviours, growth and reproduction (Pörtner 2001, 2002, Pörtner et al. 2001). The data base supporting these conclusions comprises findings in several Antarctic invertebrates and fish. Thermally induced reduction in oxygen supply capacity begins at low and high pejus temperatures (pejus = getting worse), before any biochemical stress indicators are affected. These pejus temperatures border the temperature window of maximum scope for aerobic activity and associated performance, suitable to support successful survival in the natural environment. Beyond pejus limits, at more extreme low and high temperature thresholds transition to anaerobic mitochondrial metabolism occurs. These thresholds were defined as critical temperatures (Tc) (review by Pörtner 2001). Similar to the respective findings in crustaceans (Frederich & Pörtner 2000) and other invertebrate phyla (for review see Pörtner 2001, 2002), thermal limitation in temperate and Antarctic fish, seen among zoarcids (temperate *Zoarces viviparus*, Antarctic *Pachycara brachycephalum*) and sub-Arctic fish like Atlantic cod (*G. morhua*), also begins with a limitation in oxygen supply capacity at pejus limits (Mark et al. 2002, Zakhartsev et al. 2003, Lannig et al. 2004) which finally causes transition to anaerobic metabolism at critical temperatures (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 & Knust 2007). Heat stress in the field causes reduced performance and enhanced mortality even before critical temperatures are reached. These findings emphasize the early effect and crucial role of limitations in oxygen supply in compromising fitness.

The data available for Antarctic marine invertebrates indicate that they may, on average, be more thermally sensitive than fish (Pörtner et al. 2007). The critical temperature in the infaunal bivalve *Laternula elliptica* was found at about 6°C (Pörtner et al. 1999, Peck et al. 2002). Before critical temperatures are reached, this species develops systemic hypoxia (hypoxemia) within the pejus range which reduces whole organism performance. The early reductions in aerobic scope were demonstrated as a complete loss of ability to burrow in *L. elliptica* or to right in the limpet *Nacella concinna* at 5°C, 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 at 2°C. The early loss of performance and a progressive reduction in haemolymph oxygenation indicates pejus thresholds close to 0°C in *L. elliptica* (Peck et al. 2004; Pörtner et al. 2006). However, 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). Pejus temperatures will be felt even earlier in this species. Finally, 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*), 50% mortality of specimens was obtained in 19 days at 4°C (D. Bailey, pers. commun.). In infaunal clams (*L. elliptica*), 50% mortality has been obtained in 2 months at 3°C, and in the brittle star *Ophionotus victoriae* 50% mortality occurred in less than 1 month at 3°C (L. Peck, pers. obs.). As a corollary, Antarctic stenotherms, especially among the invertebrates, live close to their thermal optimum while others, e.g. those able to acclimate as among Antarctic fish, may live permanently below their optimum. For example, the Antarctic eelpout (*P. brachycephalum*) grows at an optimum of 5°C, above ambient temperatures (Brodte et al. 2006), in accordance with its ability to acclimate to the warmth (Lannig et al. 2005).

Several Antarctic marine invertebrate species also dwell in the intertidal zone and may experience elevated temperatures in summer. Their more limited acclimation capacity suggests that they sustain higher temperatures beyond the pejus range passively. They likely use metabolic depression strategies and anaerobic metabolism 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 has been known for longer in temperate zone invertebrates (Grieshaber et al. 1994).

All of these findings suggest limited thermal tolerance according the restriction of oxygen supply capacity to within a limited thermal window in marine, including Antarctic ectotherms (Pörtner 2001). Molecular mechanisms of thermal acclimation, protection and limitation are adjusted to the same temperature range, but for individual mechanisms windows are wider and reach beyond the limits set beyond those of aerobic scope (Pörtner 2002, Pörtner & Knust 2007).

In conclusion, long term climate sensitivity is ultimately set by pejus limits at the borders of acclimation capacity of a species. Here the species has no capacity left to shift upper thermal limits to higher temperatures. In temperate eelpout, the loss of aerobic performance at acclimation limits was mirrored in the progressive reduction of growth rates beyond temperature dependent growth optima, paralleled by the onset of a decrease in abundance in the field (Pörtner & Knust 2007). Under field conditions the loss of aerobic performance capacity beyond pejus rather than critical temperatures thus appear crucial in limiting the survival of warmer summers. We have no field data yet to demonstrate the respective phenomena in Antarctic marine ectotherms but the information available from physiological and other lab studies (see above) clearly indicates that further warming of the marine environment by as little as 1°C will reach the pejus limits of some, firstly marine invertebrate species. From a global perspective, polar ectotherms clearly do not have the opportunity to retreat to cooler, i.e. higher latitude waters. Depending on the ubiquity of warming trends in the Antarctic, this 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 (Jacob 2006) may at least temporarily buffer such changes, this cannot prevent the potential loss of typical Antarctic fauna.

Seasonality on the high antarctic benthic shelf communities

The Antarctic spring is considered one of the planet's principal episodes of oceanic primary production (Hence et al. 2003), reaching values in excess of 1 mg $Chl \cdot l^{-1}$ in just a few weeks. More than 10^7 km^2 of sea ice containing a huge trapped biomass melt (Thomas & Dieckmann 2002), whereas the sunlight period considerably increases, 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. Most of the large predators abandon the high Antarctic at the start of the long Austral winter, when the continental shelf and large areas of the open ocean pass through a seasonal coverage of ice more than a metre thick. This general model set conditions for one of the long-lasting paradoxes in Antarctic science: pronounced marine seasonality (Clarke 1988). Concomitantly, a series of misconceptions arose, chief among them the notion that the high Antarctic fauna undergoes a period of low activity in winter as a consequence of reduced food availability, given that the sea water temperature remains 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 (Arntz et al. 2004). 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 & 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 the currents. Due to low temperatures, a large number of species present slow metabolic rates, associated to 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 that force 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 quickly occupy areas scraped clean by icebergs, presumably showing higher growth rates than expected (Teixidó et al. 2004). A series of experimental observations have furnished solid evidence supporting the suspicion 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 pattern 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.

2. Seabed sediment with high nutritive quality and grain sizes suitable for the anatomic structures of benthic suspension feeders. In average, measured concentrations of protein (3 mg g^{-1}) and lipids (2 mg g^{-1}) were higher than on other continental shelves and similar to the contents found in settling particles (Smith et al. 2006).

3. Tides acting as the incessant mechanism that resuspends the “food banks” and supplies 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 at the high Antarctic shelf seabed makes it necessary to reconsider a paradox that has served as a cornerstone for the understanding of polar ecosystems. Resuspension by tidal currents and the high nutritional quality of the seabed sediment after the summer allowing benthic trophic conditions to remain almost constant throughout the year set the basis for a new model of Antarctic seasonality. A model including these characteristics could help explain the diversity and high biomass of benthic communities in the high

Antarctic, even when food input from the euphotic zone becomes scarce during the long winter.

Further, the high diversity levels found on the Antarctic shelf will be easier to explain if the general process of speciation that has been taking place for millions of years has not been periodically interrupted by temporary shortages of food resources. These new findings must be taken into account when planning future research to be carried out on the Antarctic bottom-dwelling fauna. They obviously invite to place special emphasis on carrying out studies during winter and open dew doors to polar research. Processes occurring near the seabed during austral winter (Fig 4.26) could be a key factor for to understand both, the high productivity of the system at the early spring and high biodiversity in one of the more oldest benthic ecosystems in the world oceans (Gili et al. 2006).

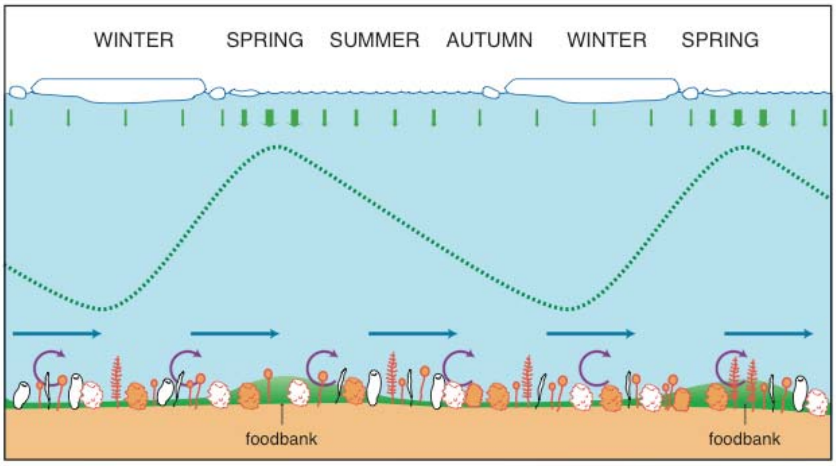


Figure 4.26. Synoptic view of the processes described in the text showing the seasonal vertical flux of new organic matter originated mainly at the beginning of spring (green line), the seasonal variation of food banks and the lateral and resuspension transport just above the seabed (arrows close the bottom).

4.10.4 Nearshore marine disturbances over the next 100 years

The Southern Ocean has the highest wind speeds, wave heights and most icebergs of any water body. Since the formation of a massive ice-cap over the southern polar region, the biodiversity on the shelf of the Southern Ocean has been exposed to high frequencies and intensities of disturbance, mainly from ice scour. These huge forces, enough to sculpt the sea bed, take place in ecological time by icebergs grounding and in evolutionary time through iceshelves advancing. In contrast the region also includes amongst the least disturbed places in the abyss and seasonally the shelf when under fast-ice. Both in the shallows (Smale et al. 2007) and on the shelf at hundreds of meters depth (Gutt & Piepenburg 2003) the seabed resembles a patchwork quilt of recovery from the last chance iceberg scour. The top few hundred meters of the Southern Ocean shelf is a very disturbed place, but has regionally high biodiversity as a result (Gutt & Piepenburg 2003). Newly scoured areas are dominated by rapid spreading and growing pioneers, whereas areas impacted less recently are occupied by slower growing, more dominant species. Polar benthic communities can be highly hierarchical so without mechanical disturbance, the major space competitors could monopolise space virtually as a monoculture. Within the Southern Ocean any change in disturbance is likely to be highly variable regionally as warming signals have been so far. For example we should expect much change in disturbance the Antarctic Peninsula region, the area where the greatest warming and ice retreat has been detected so far (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. Models suggest a 2ºC ±2ºC rise in the Southern Ocean over the next 80 years but like elsewhere on the planet surface temperature change is very geographically patchy. Warming has been detected both in Weddell Sea deep water and Bellingshausen surface water (see Meredith & King 2005) but to date only the latter is biologically meaningful. Even this warming has only strongly influenced the top few tens of meters seasonally. However given recent intensification of atmospheric CO_2 levels and air and land temperature rises it seems inevitable that warming of surface waters will not only continue, but accelerate and widen in geographic extent. 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. 2002), glacial retreat (Cook et al. 2005), rising acidification and desaturation of aragonite levels (Key et al. 2004).

Many conditions will and altering drastically with rising atmospheric CO_2 and regional warming, e.g. ice shelf disappearance releasing large areas of new habitat and increased area for primary productivity, but which are not considered disturbance so not discussed. Furthermore not all aspects of disturbance will change, for example volcanism and meteorite impacts, and thus are not considered here. The implications and the actual nature of disturbance from other factors such as weather, wind and anthropogenic impacts such as pollution and trawling are likely to change over the next decades and centuries but may not link or link only complexly to regional warming and are also not discussed in this article. Changes in disturbance with regional warming can be categorised into two distinct categories, acute and chronic, the former based on rapid change on short ecological time scales and the latter changing on time scales over hundreds of years. Major elements of disturbance envisaged to alter with current and future regional warming include:

Acute

1. Rapid ice-loading increases and decreases
2. Coastal sedimentation
3. Freshening events
4. Thermal 'events'
5. Arrival of invasive non-indigenous species (covered in a separate section)

Chronic

6. Long term ice scour decreases
7. Warming

8. Acidification
9. Deoxygenation

Here the way in which each of these disturbance events will alter is considered as 'best guesses'. 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 maybe more impact.

Acute disturbance

Rapid ice-loading increases and decreases

Currently we are in an interglacial and as such expect that some of the glacier and ice shelve retreat would be an expected response following a glacial maximum. There is evidence that a number of ice shelves today were not present during the last interglacial, when it was warmer than now. Thus to evaluate changes in ocean ice-loading, the major form of disturbance to the shelf, we must consider current rapid change in that context. However, recent glacier retreat and ice shelf collapse in some regions, such as the Antarctic Peninsula, has almost certainly occurred at a greater rate 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, generate floating icebergs. An increased 'population' of Southern Ocean icebergs, leading to an increased frequency of ice scouring, could have severe implications for shelf benthos. When icebergs collide with the seabed, large areas of benthos may be completely destroyed as the iceberg scours through the sediment or scrapes across the bedrock. As modern Antarctic icebergs can be very large, with draughts of up to 600 m depth, most of the Antarctic shelf system is exposed to this disturbance. As a result, the shelf system is effectively a patchwork of discrete disturbance events of different ages, and these relict scours support benthic communities at various stages of recovery (Gutt & Piepenburg 2003).

At small scales (tens to hundreds of metres) iceberg groundings are catastrophic and cause high mortality to benthic assemblages. For example, in shallow waters Peck et al. (1999) observed a 99.5% reduction in macrofaunal abundance following iceberg scouring in the South Orkney Islands. Similar decreases in abundance have been found further south at an Antarctic Peninsula site. Similarly, considerable alterations to community structure following ice scouring have been recorded in deeper waters along the shelf. At highly disturbed sites that experience chronic ice scouring, benthic communities may be held at early successional stages and are characterised by a high abundance of pioneer species and often low (α and β) species diversity. Few studies have directly linked the intensity of iceberg scouring with community structure but it seems increased ice disturbance can be strongly related to decreased biodiversity, biomass and space coverage (Fig 4.27). Also, the iceberg groundings damage benthos and promote feeding opportunities for scavenging species, so that at chronically disturbed locations the relative abundance of scavengers may be considerably elevated. If ice scouring to nearshore benthos intensifies due to greater ice loading into coastal waters, it seems likely that communities at some (or even many) locations will become dominated by pioneers (r strategists) and scavengers, whilst the species richness and the abundance of large and old sessile animals will decline.

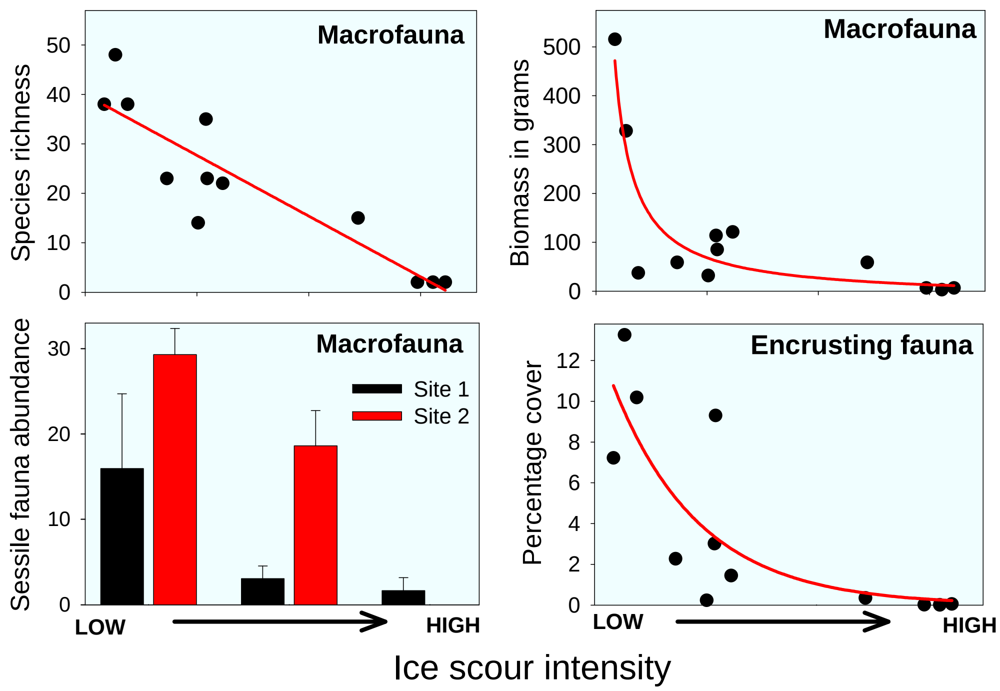


Figure 4.27. Influence of ice scour on fauna in the shallows at Adelaide Is., West Antarctic Peninsula. Patterns of faunal decrease with increasing ice scour frequencies in terms of macrofaunal richness (top left), biomass (top right) and abundance (bottom left) and cryptofauna (encrusting lithophiles) (bottom right). Author's unpublished data.

At greater spatial scales (i.e. regional) it has been suggested that iceberg scouring promotes and maintains biodiversity by increasing habitat heterogeneity and preventing monopolisation by dominant competitors. That is, within a region there exists a number of disturbance patches at various stages of recovery, which are characterised by a different suite of species. The physical structure of the iceberg scours also changes with time and therefore the region represents a large number of different microhabitats, which are inhabited by a diverse group of animals. Whether or not the shelf ecosystem is robust to regional changes in disturbance frequency remains unclear, but Johst et al (2006), using the Weddell Sea as a model system, found that a decrease in disturbance frequency is likely to more detrimental to regional biodiversity than an increase.

There are a number of uncertainties involved in predicting future disturbance frequencies. Whilst it is likely that ice loading will increase over the next hundred years or so, the influence of floating ice as an agent of disturbance to the benthos is less easy to predict. For example, if surface ocean waters continue to warm then the life expectancy of floating ice would intuitively decrease. Perhaps most importantly, the frequency of ice scouring in coastal waters is strongly linked with the persistence of seasonal fast ice (Smale et al. 2007). During the winter months icebergs are 'locked in' by fast ice that has formed around them, which restricts their movements and therefore the potential to cause disturbance. The extent and duration of fast ice has decreased in some regions during recent decades (Zwally et al. 2002), and further reductions in sea ice duration would mean that icebergs are free to move around coastal waters for longer periods of the year. Thus, the net result is that ice scouring frequencies in coastal waters will almost certainly increase with diminishing sea ice extent and duration.

Coastal sedimentation

Recent studies have shown that a number of maritime glaciers have dramatically retreated in the last few decades, particularly along the Antarctic Peninsula, and that the flow of ice into coastal waters is currently accelerating in some regions. 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. It should first be noted that whilst fjord systems are excellent 'natural laboratories' for studying the response of benthic assemblages to sedimentation, they are complex systems strongly influenced by other physical factors, such as ice scour, post-disturbance recovery times, freshening and sediment load. Hence, although fjords do represent gradients of sedimentation, this process cannot be considered in isolation and conclusions should be drawn tentatively.

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 subsequently supported by studies elsewhere and it is likely that Antarctic communities inhabiting the shelf adjacent to retreating glacier fronts will undergo similar change over ecological timescales.

Furthermore, it is likely that increasing sedimentation at ice fronts will disturb a much greater area of shelf by destabilising the substrata. On steep marine slopes ‘slumping’ of unstable sediments may smother considerable areas of benthos. Although localised, slumping events can result in high mortality of sessile species, reduced richness and significant community restructuring (see e.g. Gambi & Bussotti 1999). Increased sediment loading in the shallows of steep sections of shelf, caused by glacial retreat, could therefore destabilise the substrata and promote slumping events. Such events could disturb assemblages over a considerable area, perhaps to depths in the order of hundreds of metres. Finally, it should be noted that retreating ice fronts also deposit material larger than fine glacial sediments, and an increased deposition of boulders or ‘drop stones’ would potentially have both adverse and beneficial effects on assemblage richness. On one hand, small scale smothering by drop stones damages sessile benthos, whilst conversely drop stones represent an important source of hard substratum for colonisation.

Freshening events

Compared with the Arctic (and lower latitude coastal environments) the influence of freshwater on marine benthos in Antarctica has been and is minimal. Even so, during Antarctic summers melt water dilutes surface seawater (to depths of ~10 m), and is undoubtedly an important stressor acting on shallow water benthos. The coastal waters of some regions are likely to experience a considerable increase in freshwater input over the next hundred years as a result of glacial retreat and melting ice sheets. Although such freshening will mainly influence benthos in very shallow waters, this fauna is not, historically, adapted to low salinity stress. The Southern Ocean is the only large marine water body with no rivers flowing into it, and thus no estuaries or ‘estuarine’ communities. Intuitively then freshening events could more strongly affect Antarctic benthos than elsewhere, and perhaps be synergistic with other changing disturbance pressures in the shallows. Stockton (1984) observed mass mortality in an epifaunal bivalve population following the summer formation of a hyposaline lens of seawater at McMurdo Sound, but this is the only field report concerning this process. Conversely, in the laboratory some Antarctic algae are unaffected by hyposaline conditions but are severely impaired by increased salinity (Wiencke 1996). Both the intensity and frequency of biologically important freshening events are highly likely to increase in some regions as a result of climate change. For example, Jacobs et al. (2002) showed that the salinity of the surface waters of the Ross Sea decreased during the late 20th Century, probably as a result of increased precipitation, reduced sea ice formation and continued melting of the West Antarctic ice sheet. 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.

Short-term thermal 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. Such events and responses are not obvious in the Southern Ocean for a number of reasons, such as winter sea temperatures already reach minima possible and the strong wind and wave induced mixing plus the sinking of water makes temporary rises of sea temperatures in the shallows minor. The exception has been adjacent to active volcanism, such as at Deception Island, but these are accompanied by covarying water chemistry. However research into potential responses of biodiversity to climate change, in Antarctica as elsewhere, have been dominated by acute warming. This is for a number of reasons but perhaps most important experimental practicality and scientific funding for simulating long-term, gradual rises is difficult. Experimental manipulation of sea water temperatures in aquaria have shown that in very rapid rises (≥0.5°C.day^{-1}) or rapid rises (e.g.1-3°C per week) most Antarctic metazoans tested appear highly stenothermal and die before 10°C. Long before such model animals reach lethal temperatures they lose the ability to perform critical activities (Fig 4.28). This has now been demonstrated in a number of species, particularly molluscs, such as 'righting' in limpets (Fig 4.28) and reburying or swimming in certain bivalves. Both the inability to perform work at slightly raised temperatures and death at higher temperatures seems to be largely due to oxygen limitation and the loss of aerobic scope (Peck 2005). Experiments on a range of invertebrate types and species have failed to acclimate animals to just 3°C. However, some studies have reported acclimation and functioning of several species of Antarctic fish for weeks to months at ~4°C (Gonzalez-Cabrera et al. 1995, and more recent studies). To date 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 but certainly they would suggest the fauna is highly sensitive to acute warming.

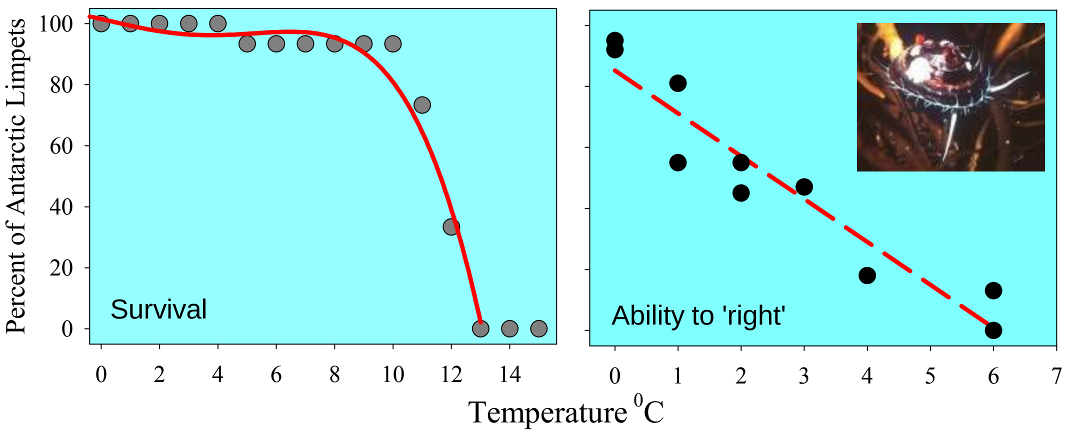


Figure 4.28. Acute temperature influences on Antarctic invertebrates. Survival and ability to 'right' (turn back over) of the Antarctic limpet *Nacella concinna*. Data from Peck (2005) and unpublished.

Many of these impacts have not only acute events but also a longer term component and are now discussed as chronic influences on Southern Ocean biodiversity.

Chronic disturbance

Long term ice scour decreases

Whilst over short timescales (i.e. decades) it seems likely that ice loading into coastal waters will increase in many regions, over longer timescales (i.e. centuries) it is more likely that the frequency of ice scouring will be less than at the present time. Currently, about 55% of the seaward margin of the Antarctic Ice Sheet is floating, but this proportion will be considerably reduced as more ice shelves collapse. Once maritime glaciers and ice 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 & 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, which could have severe implications for habitat heterogeneity and regional biodiversity (Johst et al. 2006).

It may, however, be in the shallowest waters where disturbance regimes (and therefore community structure) 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, encapsulates 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 be diminished 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

Across much of the planet sea surface temperatures are warming but increases are much greater in some areas than others (Hansen et al. 2006). Whilst much of the Southern Ocean surface shows no warming signal, the surface of the Bellingshausen Sea has warmed rapidly over the last 50 years, but this signal rapidly decreases across the top hundred meters depth (Meredith & King 2005). Temperatures at these depths are by far the best studied because measurement by remotely sensed (satellite) imagery is possible and scientists can deploy CTD probes to take direct readings. The best known, upper layer of water is Antarctic Surface Water (AASW) seasonally varies by just a few degrees C. Underlying this is Winter Water extending from about 50-200 m, but most Antarctic shelf benthos live at depths of several hundred meters. The water masses they live in, their temperature means and variabilities differ but along most of the WAP it is Circum Polar Deep Water (CDW). This is typically about ~1.5˚C but can vary by even as much as 3˚C locally (BAS unpublished data). At these deeper shelf depths in the Weddell Sea, warming over ~50 years has been detected in Weddell Deep Water (WDW) but only by biologically insignificant amounts and it is not clear that it is associated with regional warming. So substantial Southern Ocean warming detected to date is limited geographically and bathymetrically to the shallows of the Bellingshausen Sea. Even this restricted warming is rapid though in the context of past changes in the Southern Ocean and the unprecedented and accelerating levels of aerial warming suggest that oceanic warming will become much greater.

How the rich biodiversity of the Southern Ocean will respond to medium/long term warming is unclear. Certain changes over decades have been found in pelagic and benthic populations of some Southern Ocean species but none definitively linked to climate change. Most of the evidence currently being discussed for how benthos might cope with temperature rises is experimental. Indeed being typically 'stenothermal' is considered a key trait characterising Antarctic marine animals and if this is the case they would be highly sensitive to predicted climate change (Peck 2005). For a number of decades it has been clear that, compared with temperate animals, polar ectotherms are highly stenothermal. This was discovered using laboratory conditions to expose Antarctic fish to raised temperatures (Somero & DeVries 1967). Such experiments showed that most species, including a wide range of invertebrates, had upper lethal temperatures below or near to 10°C (Peck 2005). Some, e.g. the bivalve *Limopsis marionensis*, can survive just a 7°C temperature window (Pörtner et al. 1999). Even in the most severe warming scenarios a rise of this magnitude in the Southern Ocean is very unlikely by 2100 but organisms can be critically affected at lower temperatures long before lethal levels. For example the bivalve *Laternula elliptica* dies at ~9°C, metabolises anaerobically at more than 5°C and can not rebury if dug out of mud at this temperature and even half can not rebury at 2-3°C (Peck 2005). So whether populations or species will survive future temperature rises may be more dictated by their ability to do critical activities e.g. feeding. It seems that mitochondria may be key in explaining reduced aerobic scope and poor temperature tolerance in Antarctic marine ectotherms. In summary experimental physiological evidence would suggest that future warming will pose major problems to marine invertebrates, especially in the WAP region.

Such findings contrast with ecological context of experiments and observations on current species distributions. Experiments (laboratory acclimations) necessarily can only be performed on ecologically short-term timescales. A crucial question is whether the level of sensitivity exhibited in short term experiments represents 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. There is question too as to how representative the 'critical activities' measured to date are of truly important activities such as feeding and indeed 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 representatives of 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 maximum summer temperatures being 3°C warmer than some localities of the WAP (Barnes et al. 2006). Although Antarctic species level endemism is high, 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 a contrast between physiological and ecological evaluations of vulnerability but both are missing valuable context. To gain consensus and more meaningful estimates of true vulnerability will require much work such as by investigations of populations across latitudes and depths, and at sites with differing temperature regimes. Assessment of community level responses is one of the most critical pieces of evidence missing. Thus response to predicted temperature rises in the Southern Ocean over the next 100 years is unclear despite experimental evidence of sensitivity.

Acidification

Burning fossil fuels and other anthropogenic activities, such as cement production, have drastically increased CO_2 emissions over the last two centuries – such that both levels and rates of increase of atmospheric concentrations are unprecedented for millions of years. Yet Key et al. (2004) estimated that half of anthropogenically emitted CO_2 has been absorbed by the global ocean. When dissolved in water this gas dissociates into ions, raising the concentration of H^+ ions resulting in raised acidity. Absorbing raised CO_2 over two centuries has shifted oceanic pH by 0.1, this seems like a small rise, but it is equivalent to a 30% increase in hydrogen ions because the pH scale is logarithmic. On longer ecologically and certainly on evolutionary time scales this constitutes a major disturbance – the total pH change in just two centuries is almost double that from the previous 20 million years. Orr et al. (2005) project that the rate of acidification will increase over the next few centuries, the depth this will penetrate will increase and both will keep increasing through the global ocean centuries after fossil fuels have been used up. Thus the magnitude of acidification will drastically increase as an agent of disturbance. Levels and rates of acidification are greatest at the surface, so organisms in the shallows and shelf are being influenced first and more seriously – this is a major problem for animals that secrete $CaCO_3$ for skeletons.

The Southern Ocean has low saturation levels of $CaCO_3$ and the Calcite Compensation Depth (CCD) is much shallower than elsewhere. Thus animals that use $CaCO_3$ in the Southern Ocean are more quickly and immediately seriously influenced by any increase in acidity, particularly those which use the Aragonite form of $CaCO_3$ as the saturation levels of this are even lower than Calcite (the other form). Model projections of changing oceanic pH show that Southern Ocean levels could become critical for organisms using aragonite even within the next 100 years (Orr et al. 2005). Many animals in the Southern Ocean, notably some molluscs and brachiopods have very thin shells leading to debate about whether this is due to a lack of durophagus predators, physiological difficulty and cost in synthesising $CaCO_3$ in the cold, both or other reasons. As with corals in the tropics and subtropics, increased difficulty in synthesising skeletal material, could pose major problems for the survival of many Southern Ocean species. There is no evidence, known to the authors, that animals are able to adapt to rapidly and substantially changing pH levels, but even if it is possible Antarctic benthos is hindered by long generation times.

Deoxygenation

Antarctic Bottom Water (AABW) is the principal carrier of oxygen (O_2) 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 current systems were disrupted or surface water is warmed the flow of O_2 to the deep is potentially reduced. The significance of reducing oxygenation to deep waters is considerable – it is the planets' largest habitat type and covers 50% of its' area. The deep sea is so poorly known that it is not easy to gauge the significance of deoxygenation in terms of losses of biomass, abundance, species richness or other metrics. A strong indicator that the Southern Ocean deep sea is very rich and contains species not found elsewhere is the huge number of species new to science caught by recent cruises in the Weddell Sea (Brandt et al. 2007). Despite only comprising ~2% of Southern Ocean samples to date, deep sea richness comprises far more than 2% of total Antarctic species currently known.

Signals of warming have already been detected in the shallows and mid-water masses of some areas of the Southern Ocean. Warmer water holds less gas and therefore O_2, but to date the levels of warming detected in the Southern Ocean are too low to be of significance. However, given the acceleration of atmospheric CO_2 and warming in just the last decade, substantial warming may be measured well before the end of the Century. However even much greater warming than the 2°C predicted would only involve a relatively small drop in the amount of gas held in the water. What warming of upper layers could do is stratify the water column more and reduce the amounts and rates of sinking, thereby transporting less O_2 to the seabed. Disruption of current directions or velocities could similarly influence and reduce the transport of O_2 to the global deep sea but is difficult to model and predict.

Conclusions

The upper shelf environments of both the Arctic and Antarctic are amongst the most disturbed places on the planet both in frequency and intensity, especially considering the slow tempo of polar ectotherms and so their ability to recolonise. Over the next 100 years there will be many influences of climate change on polar disturbance, acute and chronic, major and minor. In parallel there will also be influences that increase with time alongside, but not driven by, climate change. However non-direct influences, such as pollution and trawling may be made more extreme by aspects of climate change and some factors may be driven by climate change as well as other causes, such as introduction of non-indigenous species. Teasing apart potential causes and drivers will be complex and effects as well are likely to be complicated by secondary and cascade influences. Amongst the acute agents of disturbance we principally highlight three of note, 1) rapid increases in both ice-loading and coastal concentrations of large icebergs from iceshelf collapses; 2) coastal sedimentation associated with ice melt, smothering benthos and hindering feeding; 3) freshening of surface waters leading to amongst other changes, stratification. The impact of the arrival and establishment of invasive non-indigenous species are covered in a separate section. Unlike at lower latitudes sea level rise is unlikely to have major acute effects, due to steep topography, lack of a specialised intertidal and subtidal fauna and the shallows being in a state of near perpetual recolonisation naturally. The chronic impacts of climate change we consider important are 1) long term decreases in ice scour probably leading to increased local scale, but decreased regional biodiversity; 2) the physiological effect of direct warming leading to reduced performance at critical activities and thus geographic and bathymetric migrations; 3) increased acidification leading to skeletal synthesis and maintenance problems, particularly for animals using aragonite, and; 4) slight deoxygenation of surface waters but disruption of currents and downwelling could ultimately lead to more serious deoxygenation for deeper layers.

To date physical and biological, rates and impacts of climate change have, for a number of reasons, been underestimated in many parts of the planet. The patterns established so far and the models built have yet to enable us to build strong predictions or understanding of what is likely to happen in the coming decades. However, there seems to be strong consensus that acidification, warming and ice melting will continue, intensify and accelerate in their effect. Many of the most heavily influenced areas have been in the polar regions and the biota their may indeed prove to be the most sensitive indicators of change, that many scientists think.

4.10.5 Diversity of the Future Fauna

Almost 15% (4.59 km^2 x 10^6) of global continental shelf areas is situated around Antarctica (Clarke & Johnston 2003). The continental shelf is very deep in the Southern Ocean, reaching 800 m in places, depressed by the massive ice sheet which formed over the continent. More than 95% of it is at depths outside the reaches of sea-ice keels scouring the seafloor, wave action, PAR (sunlight i.e. outside the euphotic zone) and scuba divers. All ecosystems on earth have been subjected to constant change but with rates differing in time and space. Life on the Antarctic shelf in preceding times has altered in response to varying degrees of cooling, ice shelf dynamics, isolation and oceanography and will to acidification and warming in the present and future. In the past, environmental changes during the transition from glacial to interglacial periods and vice versa have been very significant. Finding evidence of past changes in benthos is difficult because of poor preservation conditions and destruction of evidence by successive advances towards glacial maxima. However, within the recent interglacial physical conditions have been comparatively predictable or stable. During the past glacial approx. 90% of the continental shelf was characterised by grounded ice shelves (Harris & O'Brien 1996), though perhaps not all simultaneously, therefore unavailable to benthic species. Wherever the benthos lived during these periods, its physical and biological environment differed significantly from today, especially with regard to food conditions (see Bonn et al. 1998). Therefore, only during interglacials was the sea-floor and the overlying water column a suitable habitat for rich benthic and pelagic communities (Gutt 2007; Clarke et al. 2004). Due to intensive efforts of off-shore research activities, e.g. within the international initiatives EPOS, EASIZ, EVOLANTA, CAML, LGP, FOODBANK and single national projects we know that the macrobenthic fauna is highly endemic, eurybathic and possibly stenothermal but it can also be relatively euryoec, it can be dominated by a high number of suspension feeders, by deposit feeders or a mixture of both together with their predators and scavengers, that it can be extremely rich but also poor in biomass, and processes are generally slow (Gutt 2007).

In assessing likely future trends we need background information on the future of the physical settings (such as temperature, and ice dynamics), and the ranges of tolerance, and individual response to environmental changes of a broad variety of ecological key species. We also need to understand interspecific interactions, the variety of different niches especially among similar species or their overlap, and the consequences for ecosystem functioning and synergistic effects. Of these, changes in the physical setting can most reliably be predicted (Clarke et al. 2007). We have a good working knowledge of how some individual organisms respond to acute changes in temperature. Whether the response of these model species in the shallows reflect the species of deeper shelf depths or short term laboratory experiments reflect field reactions is unknown. At the moment species have only been investigated in isolation and we have almost no understanding of how species interactions may change. Some predictions we can make, however are (see also Fig. 4.29):

a. Grounding icebergs devastate the sea-floor down to almost 600 m depth with the highest frequencies between 30 and 250 m (Gutt 1996, 2001; Gray 2001). Shallower than 30 m smaller growlers may scour much more of the seafloor in a single summer season with increasing force with increasing depth (Smale et al. 2007). Populations of most species are completely destroyed during a scour but about 1-5% of individuals survive of some smaller species (Smale et al., in press). Early stages of recolonisation are generally poor in species (Gutt & Piepenburg 2003). We consider increased iceberg disturbance due to increasing calving events would lead to a local expansion of low diversity areas.

After disintegration of entire ice shelves, e.g. Larsen A/B locally, iceberg calving and therefore iceberg disturbance ceases. According to field studies it is unlikely that this will cause an additional decrease in local diversity due to increased competitive displacement (Gutt & Piepenburg 2003). However, such a scenario cannot be excluded for highly dynamic especially shallower benthic assemblages, which reach a final stage of succession faster than deeper ones (Gutt et al. 2007).

b. Spatially explicit models show that increased disturbance will increase regional benthic biodiversity due to habitat fragmentation and the resulting coexistence of different stages of faunistic succession (Johst et al. 2000) between 50 and 250 m water depths. According to the Intermediate-Disturbance-Hypothesis (Huston 1979), beyond a certain magnitude of disturbance diversity of the many sessile species will decrease since a large amount of a mature assemblage will be removed (Fig. 4.30). However, a few mobile and sessile pioneer species would benefit from such a development.

c. Iceberg scouring also shapes bottom topography. In combination with winnowing and sedimentation processes, the substratum can get more heterogeneous and could provide more ecological niches for more species. However, since sediments are generally poorly sorted this effect is supposed to be of significant relevance only where extreme sediment characteristics are generated such as almost pure mud, gravel, or bedrock.

d. Locally dense concentrations of suspension are presently washed into coastal waters along the Antarctic Peninsula due to local deglaciation. Especially species-rich suspension feeder communities are expected not to tolerate such conditions but some opportunistic deposit feeders and infaunal species could benefit from the formation of softer sediments (Wlodarska-Kowalczuk et al. 2005) and the fertilisation of the pelagic system (Smetacek & Nicol 2005).

e. Principally 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, both can be predicted, an increase due to an extended period for algal blooms in the open water and a reduction (see Thrush et al. 2006), at least at the regional level, due to a less melting sea-ice and, as a consequence, a less developed halocline. In addition it has to be considered that this food source is modified on its way from the euphotic zone to the sea-floor, which is, compared to other continents, vertically longer due to the great water depth on the shelf and also horizontally for under-ice shelf habitats.

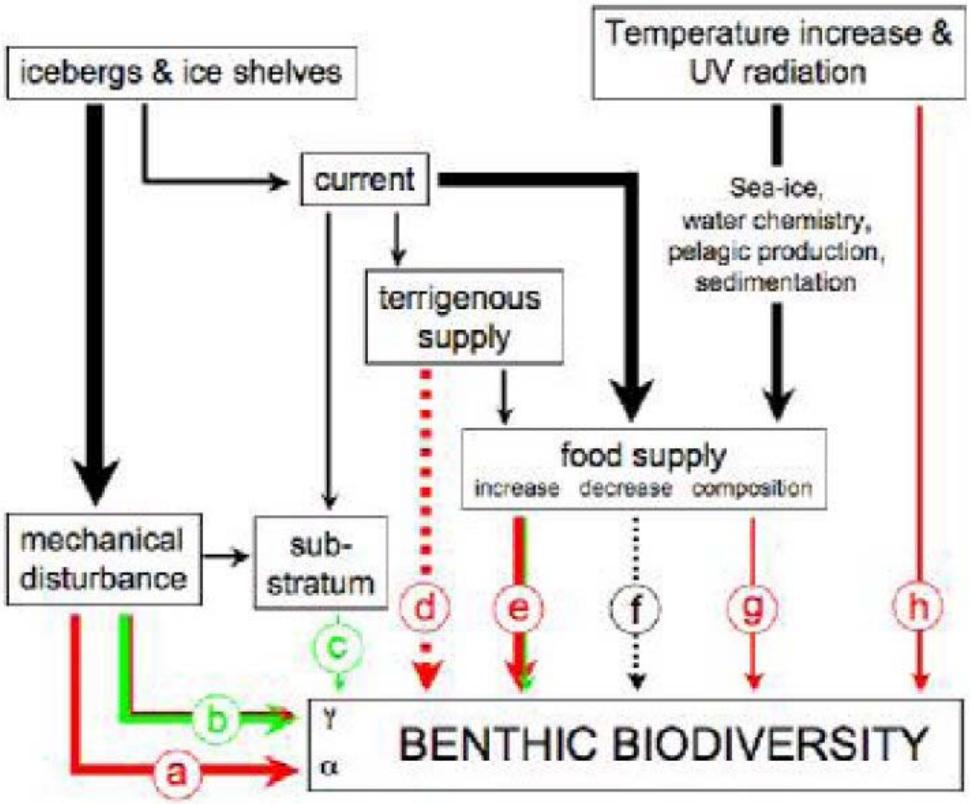


Fig. 4.29. Different pathways within the Antarctic ecosystem indicating possible responses of the macrobenthos to Global Change impacts. Red denotes a possible decrease and green an increase in biodiversity. Broken lines indicate unlikely scenarios. Line thickness indicates severity of impact.

In terms of quantity, most suspension feeders might not only be adapted to today's assumed summer surplus food conditions but also to limited food availability during glaciation periods (Gutt 2000). Some benthic species are already able to consume different kinds of food and to extend their feeding period beyond the summer bloom when they feed primarily on fresh phytodetritus. As a consequence, an increase in food could be a problem rather than a decrease. Single species could suffer from a clogging of their feeding apparatus. At the community level, in contrast to ecological theory, enriched systems can be characterised by the dominance of a few species only (Schewe & Soltwedel 2003). The demise of suspension feeders would also have potential consequences for the rich associated fauna since their microhabitat or food source is lost. Only benthic systems living under strongly limiting food conditions (see Post et al. 2007; Riddle et al. 2007) might get richer in diversity and biomass, when food conditions improve, e.g. in areas where ice shelves disintegrate. On the other hand an assemblage being well adapted to former extremely poor food conditions might suffer rather than increase in diversity.

In case of a reduced food supply 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 & Gutt 2007) or communities living under seriously limiting food conditions beneath ice shelves will regionally disappear if these conditions further deteriorate.

In terms of food quality a shift in the pelagic realm from a retention-system (only few sinking microorganisms reach the bottom) to a loss-system (much organic material e.g. in form of faecal pellets sinks rapidly to the bottom, (see Peinert et al. 1989)) caused by possible geographic shifts of big krill populations (Atkinson et al. 2004) reaching the shelves could also affect the benthos. What would be the response of the benthos? With few exceptions only little is known about the qualitative food preference of benthic key species (Orejas et al. 2003). Consequently, this question can only be answered on a general basis. Suspension feeding species that prefer small food particles will eventually suffer whilst those preferring larger particles and opportunists will benefit. Deposit feeders might be more flexible and benefit more from an increase in the amount of food but less from a change in food quality. However, dense concentrations of deep-sea holothurians being obviously adapted to low-food conditions are found on the Antarctic shelf (Gutt & Piepenburg 1991). It is still unclear whether increased UV radiation will only alter the composition of the plankton or also have a negative effect to quantitative sedimentation rates. In general, however, similar consequences as for a warming of the ocean's surface layer can be expected.

Many of the predators seem to be specialized on very specific food items. As a consequence they are assumed to be very sensitive to changes in food availability if they cannot adapt rapidly to other food sources.

At present an increase in sea-water temperature is known from near shore and off shore areas, along the Antarctic Peninsula only down to water depths of less than 100 m. These are affected by the well known marked atmospheric warming in the area (Meredith & King 2005; Barnes et al. 2006). Direct effects of increasing temperature to shallow water benthic species are described in detail by Barnes & Peck (this volume). Species whose representatives are endangered by this warming have a good chance to survive if they can migrate south along both sides of the Antarctic Peninsula into colder waters or if populations of such species already exist on the high latitude shelves. Data on temperature anomalies in waters around East Antarctica show a slight decrease in temperature rather than an increase and the sea-ice development is inconsistent (this volume). On the one hand, if unexpectedly the increase of temperature will extend, most of the assumed minimum of 17,000 species on the shelf do not have a chance to escape to colder areas and survive there if we assume that the fauna of the deeper shelves is not as tolerant to temperature increase as those adapted to the naturally more variable shallow conditions. If they cannot adapt fast enough to increasing temperatures a mass extinction would be the consequence. On the other hand, the fact that many species inhabiting the deeper shelf are highly eurybath and consequently can tolerate temperatures of the Warm Deep Water can lead to the assumption that several species have the capacity to survive a modest warming. A replacement of the typical Antarctic species composition by invaders from northerly adjacent continents and oceans will only be successful if the hydrodynamic isolation of the Southern Ocean will not act as one of the most efficient barriers of species dispersal on Earth and if the invaders are superior in competition.

Finally, it can be concluded that a continued and extended Global Warming could lead to both, a decrease and an increase in benthic diversity due to changes in ecosystem functioning. Under realistic scenarios diversity will be changed considerably only in terms of the numeric composition of the species; based on our present knowledge and on projections for the environment a significant extinction of species or invasion of many new species is rather unlikely. However, if genetic techniques show that many cryptic species exist, which have a spatially very restricted distribution, in contrast to the still assumed circumpolar distribution of species, such species are generally more endangered by any kind of environmental changes than the presently known species.

References

Abram NJ, McConnell J, Mulvaney R, Wolff E (2007a) Ice core records of regional sea ice changes around Antarctica during the 20th century. Geophysical Research Abstracts 01599

Abram NJ, Mulvaney R, Wolff E, Mudelsee M (2007b) Ice core records as sea ice proxies: An evaluation from the Weddell Sea region of Antarctica. Journal of Geophysical Research 112:D15101

ACIA 2005 – Chapter 2

Adams et al. 2006 – Chapter 3

Adamson DA, Colhoun EA (1992) Late Quaternary Glaciation and Deglaciation of the Bunger Hills, Antarctica. Antarctic Science 4:435-446

Adamson DA, Pickard J (1986) Physiography and geomorphology of the Vestfold Hills. In: Pickard J (ed) Antarctic Oasis: terrestrial environments and history of the Vestfold Hills. Academic Press, Sydney, p 99-139

AGCS (2008) State of the Antarctic and Southern Ocean Climate System (SASOCS). Reviews in Geophysics

Ainley 2005 – chapter 3

Ainley D G, Clarke E D, Arrigo K, Fraser W R, Kato A, Barton K J and Wilson P R (2005) Decadal-scale changes in the climate and biota of the Pacific sector of the Southern Ocean, 1950s to the 1990s. Antarctic Science 17, 171-182.

Ainley DG, Hobson KA, Crosta X, Rau GH, Wassenaar LI, Augustinus PC (2006) Holocene variation in the Antarctic coastal food web: linking δD and δ13Cin snow petrel diet and marine sediments. Marine Ecology Progress Series 306:31-40

Ainley et al 2007 – chapter 3 – page 172

Aislabie JM, Balks MR, Foght JM, Waterhous EJ (2004) Hydrocarbon spills on Antarctic soils: effects and management. Environ Sci Technol 38:1265-1274

Akiyoshi, H., T. Sugita, H. Kanzawa, N. Kawamoto, "Ozone perturbations in the Arctic summer lower stratosphere as a reflection of NOx chemistry and planetary scale wave activity", *J. Geophys. Res.*, **109**, D03304, (2004).

Alekhina IA, Marie D, Petit JR, Lukin VV, Zubkov VM, Bulat SA (2007) Molecular analysis of bacterial diversity in kerosene-based drilling fluid from the deep ice borehole at Vostok, East Antarctica. FEMS Microbiol Ecol 59:289-299

Alford, M. H. (2003). Improved global maps and 54-year history of wind-work on ocean inertial motions. *Geophys. Res. Lett.* **30**, doi:10.1029/2002GL016614.

Allegrucci G, Carchini G, Todiscso V, Convey P, Sbordoni V (2006) A molecular phylogeny of Antarctic Chironomidae and its implications for biogeographical history. Polar Biology 29:320-326

Alley, R.B., Anandakrishnan, S., Bentley, C.R., Lord, N., 1994, A water-piracy hypothesis for the stagnation of Ice Stream C, Antarctica. *Ann. Glaciol.* **20**, 187–194.

Alley, R.B., Anandakrishnan, S., Dupont, T.K., Parizek, B.R. and Pollard, D., 2007, Effect of sedimentation on ice-sheet grounding-line stability. *Science* **315**, 1838-1841.

Alley, R.B., and R.A. Bindschadler (2001), *The West Antarctic Ice Sheet: Behavior and environment.* Washington DC: American Geophysical Union, Antarctic Research Series vol. 77, 296 pp.

Alley, R.B., D.D. Blankenship, C.R. Bentley, and S.T. Rooney, 1987, Till deformation beneath ice stream B. 3. Till deformation: Evidence and implications. *Journal of Geophysical Research* , **92**(B9), 8921-8929.

Anandakrishnan S. and R.B. Alley, 1994, Ice stream C, Antarctica, sticky spots detected by microearthquake monitoring, *Ann. Glaciol.* **20**, 183-186.

Anandakrishnan, S., Catania, G.A., Alley, R.B. and Horgan, H.J. 2007. Discovery of till deposition at the grounding line of Whillans Ice Stream, *Science* **315**, 1835-1838.

Anandakrishnan, S., D.D. Blankenship, R.B. Alley, and P.L. Stoffa (1998), Influence of subglacial geology on the position of a West Antarctic ice stream from seismic observations. *Nature 394*, 62-65.

Anderson J.B. (1999) Antarctic marine geology, Vol. Cambridge University Press, Cambridge, UK

Anderson J.B., Andrews JT (1999) Radiocarbon constraints on ice sheet advance and retreat in the Weddell Sea, Antarctica. Geology 27:179-182

Anderson JB, Shipp SS, Lowe AL, Wellner JS, Mosola AB (2002) The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history: a review. Quaternary Science Reviews 21:49-70

Anderson, T. R., and I. J. Totterdell (2004), Modelling the response to the biological pump to climate change., in *The Ocean Carbon Cycle and Climate*, edited by M. Folllows and T. Oguz, pp. 65-96, Kluwer Academic Press, Netherlands.

Andrews JE, Tandon SK, Dennis PF (1995) Concentration of Carbon Dioxide in the Late Cretaceous Atmosphere. J. Geol. Soc. (Lond.) 152:1-3

Andrews JT (1992) Glaciation - a Case of Missing Water. Nature 358:281-281

Andrews PL, Bell RS. 1998. Optimizing the United Kingdom Meteorological Office data assimilation for ERS-1 scatterometer winds. *Monthly Weather Review* **126**: 736–746.

Andriashev AP, Chernova NV (1995) Annotated list of fishlike vertebrates and fish of the Arctic seas and adjacent waters. J Ichthyol 35: 81–123

Anisimov O, Fitzharris BB, Hagen JO, Jefferies B, Marchant H, Nelson F, Prowse T, Vaughan D (2001) Polar regions (Arctic and Antarctic). In: Climate Change 2001: Impacts, Adaptation, and Vulnerability. Contribution of Working Group II to the Third Assessment Report of the Intergovernmental Panel on Climate Change, pp 801-841 (McCarthy JJ, Canziani OF, Leary NA, Dokken DJ & White KS, Eds), Cambridge University Press, Cambridge, UK

Anisimov OA, Vaughan DG, Callaghan TV, Furgal C, Marchant H, Prowse TD, Vilhjálmsson H, Walsh JE (2007). In: Climate Change 2007: impacts, adaptation and vulnerability. Contribution of working group II to the fourth assessment. Report of the Intergovernmental Panel on Climate Change (IPCC), pp. 653-685 (Parry ML, Canziani OF, Palutikof JP, van der Linden PJ, Hanson CE, Eds). Cambridge University Press, Cambridge, UK

Antarctica as a Global Indicator p161-175

Aoki, S. 2002. Coherent sea level response to the Antarctic Oscillation. Geophysical Research Letters, 29(20), 1950, doi:10.1029/2002GL015733.

Aoki, S., N. L. Bindoff, and J. A. Church, Interdecadal water mass changes in the Southern Ocean between 30°E and 160°E. *Geophys. Res. Letters*, **32**, L07607, doi:10.1029/2004GL022220 (2005).

Aoki, S., S.R. Rintoul, S. Ushio, S. Watanabe, and N.L. Bindoff, 2005. Freshening of the Adelie Land Bottom Water near 140°E, Geophys. Res. Lett., Vol. 32, L23601, doi10.1029/2005GL024246, 2005.

Appleby PG, Jones VJ, Ellis-Evans JC (1995) Radiometric dating of lake-sediments from Signy Island (Maritime Antarctic) - evidence of recent climatic-change. Journal of Paleolimnology 13:179-191

Arakawa, A., 1966: Computational design for long-term numerical integration of the equations of fluid motion: Two-dimensional incompressible flow, Part 1. Journal of Computational Physics, 1, 119-143.

Arblaster, J. M., and G. A. Meehl, 2006. Contributions of external forcings to Southern Annual Mode trends. Journal of Climate, 19, 2896-2905.

Archer, D. (2005), Fate of fossil fuel CO2 in geologic time, *J. Geophys. Res.*, DOI: 10.1029/2004JC002625.

Arctic Climate Impact Assessment (ACIA), 2005 Ch. 6, Cryosphere and hydrology.

Armand LK, Crosta X, Romero O, Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean sediments: 1. Sea ice related species. Palaeogeography, Palaeoclimatology, Palaeoecology 223:93-126

Arnold & Convey 1998 – chapter 3

Arntz WE, Brey T, Gallardo VA (1994) Antarctic zoobenthos. Oceanogr Mar Biol Ann Rev 32: 241-304

Arrigo et al., 2003 – chapter 3

Arsene,C., I.Barnes, and K.H.Becker, FT-IR product study of the photo-oxidation of dimethyl sulfide: Temperature and O_2 partial pressure dependence. Phys. Chem. Chem. Phys 1:5463-5470, 1999.

Arthern R. J. and R. C. A. Hindmarsh, 2006, Determining the contribution of Antarctica to sea-level rise using data assimilation methods, *Phil. Trans. R. Soc.* A, **364**, 1841-1865. doi:10:1098/rsta.2006.1801.

Arthern, R. J. and Wingham, D. J., 1998, The Natural Fluctuations of Firn Densification and Their Effect on the Geodetic Determination of Ice Sheet Mass Balance, *Climatic Change,* **40**, 605–624.

Arthern, R. J., D. J. Wingham, and A. L. Ridout, 2001, Controls on ERS altimeter measurements over ice sheets: Footprint-scale topography, backscatter fluctuations, and the dependence of microwave penetration depth on satellite orientation, *J. Geophys. Res.*, **106**(D24), 33,471–33,484.

Arthern, R. J., D. P. Winebrenner, and D. G. Vaughan, 2006, Antarctic snow accumulation mapped using polarization of 4.3-cm wavelength microwave emission, *J. Geophys. Res.*, 111, D06107, doi:10.1029/2004JD005667.

Arzel, O., T. Fichifet, and H. Goosse, 2006: Sea ice evolution over the 20th and 21st centuries as simulated by current AOGCMs. *Ocean Modelling*, **12,** 401-415.

Ashley GM, Smith ND (2000) Marine sedimentation at a calving glacier margin. Geol Soc Amer 112: 657–667

Assmann, K., Hellmer, H.H., and Beckmann, A., 2003: Seasonal variation in circulation and water mass distribution on the Ross Sea continental shelf. Antarctic Science, 15 (1), 3–11 (doi: 10.1017/S0954102003001007).

Assmann, K.M. and R. Timmermann (2005) Variability of dense water formation in the Ross Sea. *Ocean Dynamics* **55**, 68-87.

Associate Committee on Geotechnical Research (ACGR). 1988. Glossary of permafrost and related ground-ice terms. *Permafrost Subcommittee, National Research Council of Canada, Technical Memorandum 142*, Ottawa, 156pp

Atkinson A, Pakhomov E, Rothery P, Siegel V (2004) Long-term decline in krill stocks and increase in salps within the Southern Ocean. Nature 432 (7013): 100-103

Atkinson A, Siegel V, Pakhomov E and Rothery P (2004) Long-term decline in krill stock and increase in salps within the Southern Ocean. Nature, 432, 100-103.

Atmos. Chem. Phys., 7, 4171-4185, 2007.

Aumont, O., and L. Bopp (2006), Globalizing results from ocean in situ iron fertilization studies, *Global Biogeochemical Cycles*, *20*, doi:10,029/2005GB002519.

Austin, J., J. Wilson, F. Li, H. Vömel, "Evolution of water vapor and age of air in coupled chemistry climate model simulations of the stratosphere", *J. Atmos. Sci.*, in press (2006).

AUTHOR? dominant Antarctic contribution to meltwater pulse IA. Quaternary Science Reviews 26:2113-2127

Bacci E, Calamari D, Gaggi C, Fanelli R, Focardi S, Morosini M (1986) Chlorinated hydrocarbons in lichen and moss samples from the Antarctic Peninsula. Chemosphere, 15:747-754

Bailey D. pers. Commun - Chapter 4, page 73

Bailey DM, Johnston IA, Peck LS (2005) Invertebrate muscle performance at high latitude: swimming activity in the Antarctic scallop, *Adamussium colbecki.* Polar Biol 28: 464-469

Baines, P. G. and Condie, S. (1998). Observations and modelling of Antarctic downslope flows: a review. In *Ocean, Ice and Atmosphere: Interactions at the Antarctic Continental Margin*, AGU Antarctic Research Series Vol. 75, S.S. Jacobs & R. Weiss editors, 29-49.

Bakun A (1996) Patterns in the Ocean-Ocean Processes and Marine Population Dynamics. California Sea Grant College System, La Jolla

Baldwin, 2001 – chapter 3

Bale ref – chapter 3 – page 101

Bamber, J., Vaughan, D.G. and Joughin, I., 2000, Widespread complex flow in the interior of the Antarctic ice sheet. *Science* **287**, 1248–1250, (doi:10.1126/science.287.5456.1248).

Banks, H. T. and N. L. Bindoff, Comparison of observed temperature and salinity changes in the Indo-Pacific with results from the coupled climate model HadCM3: processes and mechanisms. *J. Climate*, **16**, 156-166 (2003).

Barbante C, Cozzi G, Capodaglio G, van de Velde K, Ferrari C, Veyssery A, Boutron CF, Scarponi G, Cescon P (1999) Determination of Rh, Pd, and Pt in polar and

Alpine snow and ice by double-focusing ICP-MS with microcentric nebulization. Anal Chem 71:4125-4133

Barbante C, Turetta C, Gambaro A, Capodaglio G, Scarponi G (1998) Sources and origins of aerosols reaching Antarctica as revealed by lead concentrations profiles in shallow snow. Ann Glaciol 27:674-678

Barbeau et al – chapter 3, page 177

Barber DC, Dyke A, Hillaire-Marcel C, Jennings AE, Andrews JT, Kerwin MW, Bilodeau G, McNeely R, Southon J, Morehead MD, Gagnon JM (1999) Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes. Nature 400:344-348

Barbraud C, Weimerskirsch H (2006) Antarctic birds breed later in response to climate change. Proceedings of the National Academy of Sciences, 103, 6248-6251.

Bard E, Hamelin B, Arnold M, Montaggioni L, Cabioch G, Faure G, Rougerie F (1996) Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. Nature 382:241-244

Bargagli R (2001) Trace metals in Antarctic organisms and the development of circumpolar biomonitoring networks. Rev Environ Contam Toxicol 171:53-110

Bargagli R, Agnorelli C, Borghini F, Monaci F (2005) Enhanced deposition and bioaccumulation of mercury in Antarctic terrestrial ecosystems facing a coastal polynya. Environ Sci Technol 39:8150-8155

Bargagli, 2005 – chapter 3 – Bargagli et al 2005?

Bargelloni et al. 1994 – Chapter 2

Bargelloni et al. 1998 – Chapter 2

Bargelloni L, Marcato S, Zane LG & Patarnello T (2000) Mitochondrial phylogeny of Notothenioids: a molecular approach to Antarctic fish evolution and biogeography. Syst Biol 49: 114-129

Baringer et al. 1997 – chapter 3 – Baringer and Price 1997?

Baringer, M.O.,and Price, J.F. 1997:Momentum and Energy Balance of the Mediterranean Outflow. Journal of Physical Oceanography, 27 (8) 1678–1692.

Barker 2001 – Chapter 2

Barnes DKA, Fuentes V, Clarke A, Schloss IR, Wallace MI (2006) Spatial and temporal variation in shallow seawater temperature around Antarctica. Deep-Sea Res II 53: 853-865

Barnes DKA, Hodgson DA, Convey P, Allen CS, Clarke AC (2006) Incursion and excursion of Antarctic biota: past, present and future. Global Ecol Biogeogr 15: 121-142

Barnes DKA, Webb KE, Linse K (2007). Growth rate and its variability in erect Antarctic bryozoans . Polar Biol 30: 1069-1081

Baroni C, Orombelli G (1994) Abandoned pengiun rookeries as Holocene paleoclimatic indicators in Antarctica. Geology 22:23-26

Barrett PJ (1996) Antarctic palaeoenvironment through Cenzoic times - a review. *Terra Antartica* 3:103-119

Barrett PJ (2006) The second great climate shift in the last 65 million years. In: Chapman R, Boston, J., and Schwass, M., (ed) *Confronting Climate Change: Critical Issues for New Zealand*. Victoria University Press., Wellington, p 6pp

Barrett PJ (2007) Cenozoic climate and sea-level history of glacimarine strata off the Victoria Land coast, Cape Roberts Project, Antarctica. In: Hambrey, al. e (eds) Glacial Sedimentary Processes and Products. International Association of Sedimentologists, Special Publication

Barrett, MCM-LTER, unpublished - chapter 4 Page 60

Barrie, L.A., Bottenheim, J.W., Schnell, R.C., Crutzen, P.J., and Rasmussen, R.A., Ozone destruction and photochemical reactions at polar sunrise in the lower Arctic atmosphere, Nature, 334, 138-141, 1988.

Bassett SE, Milne GA, Bentley MJ, Huybrechts P (2007a) Modelling Antarctic sea-level data to explore the possibility of a dominant Antarctic contribution to meltwater pulse IA. Quaternary Science Reviews 26:2113-2127

Bates M.L., M.H. England and W.P. Sijp (2005). On the multi-century Southern Hemisphere response to changes in atmospheric CO2-concentration in a global climate model. Meteorol. Atmos. Phys. 89, 17-36.

Bates, N. (2001), Interannual variability of oceanic CO_2 and biogeochemical properties in the Western North Atlantic subtropical gyre, *Deep-Sea Research II*, *48*, 1507-1528.

Beagley, S.R., J. de Grandpré, J.N. Koshyk, N.A. McFarlane, T.G. Shepherd, “Radiative-dynamical climatology of the first-generation Canadian Middle Atmosphere Model”, *Atmos. Ocean*, **35**, 293-331 (1997).

Beamish RJ (1995) Climate Change and Northern Fish Populations. Canadian Special Publication of Fisheries and Aquatic Science, 121, 739 pp

Beckley, B.D., Lemoine, F.G., Lutchke, S.B., Ray, R.D. and Zelensky, N.P. 2007. A reassessment of global and regional mean sea level trends from TOPEX and Jason-1 altimtery based on revised reference frame and orbits. Geophysical Research Letters, 34, L14608, doi:10.1029/2007GL030002.

Beckmann, A., H.H. Hellmer and R. Timmermann, 1999: A numerical model of the Weddell Sea: large-scale circulation and water mass distribution. *J. Geophys. Res*. **104**(C10), 23375-23,391.

Behrenfeld M. J. et al, Nature 444, 752-755; 2006

Behrenfield, M. J., and P. G. Falkowski (1997), A consumers guide to phytoplankton primary production models *Limnol. Oceanogr.*, *47*, 1479-1491.

Beine, H.J., Dominé, F., Simpson, W.R., Honrath, R.E., Sparapani, R., Zhou, X., and King, M.: Snow-pile and chamber experiments during the Polar Sunrise Experiment 'Alert 2000': Exploration of nitrogen chemistry, Atmos. Environment, 36(15-16), 2707-2719, 2002a.

Beine, H.J., Honrath, R.E., Dominé, F., Simpson, W.R., and Fuentes, J.D.: NO_x during background and ozone depletion periods at Alert: Fluxes above the snow surface, J. Geophys. Res., 107(D21), 4584, doi:10.1029/2002JD002082, 2002b.

Bell, R.E., D.D. Blankenship, C.A. Finn, D.L. Morse, T.A. Scambos, J.M. Brozena, and S.M. Hodge (1998), Influence of subglacial geology on the onset of a West Antarctic ice stream from aerogeophysical observations. *Nature 394*, 58-62.

Belt S, Masse G, Rowland SJ, Poulin M, Michel C, LeBlanc B (2007) A novel chemical fossil of palaeo sea ice: IP25. Organic Geochemistry 38:16-27

Bengtsson, L., K.I. Hodges, and E. Roeckner, 2006: Storm tracks and climate change. *J. Climate*, **19**, 3518–3543.

Bentley MJ (1999) Volume of Antarctic Ice at the Last Glacial Maximum, and its impact on global sea level change. Quaternary Science Reviews 18:1569-1595

Bentley MJ, Anderson JB (1998) Glacial and marine geological evidence for the ice sheet configuration in the Weddell Sea-Antarctic Peninsula region during the last glacial maximum. Antarctic Science 10:309-325

Bentley MJ, Fogwill CJ, Kubik PW, Sugden DE (2006) Geomorphological evidence and cosmogenic 10Be/26Al exposure ages for the Last Glacial Maximum and deglaciation of the Antarctic Peninsula Ice Sheet. GSA Bulletin 118:1149–1159

Bentley MJ, Hodgson DA, Smith JA, Cox NJ (2005a) Relative sea level curves for the South Shetland Islands and Marguerite Bay, Antarctic Peninsula. Quaternary Science Reviews 24:1203-1216

Bentley MJ, Hodgson DA, Smith JA, Ó Cofaigh C, Domack EW, Larter RD, Roberts SJ, Brachfeld S, Leventer A, Hjort C, Hillenbrand C-D, Evans J (0000) Mechanisms of Holocene palaeoenvironmental change in the Antarctic Peninsula region. The Holocene 00:00-00

Bentley MJ, Hodgson DA, Sugden DE, Roberts SJ, Smith JA, Leng MJ, Bryant C (2005b) Early Holocene retreat of the George VI Ice Shelf, Antarctic Peninsula. Geology 33:173-176

Bentley, C.R. (2004), Mass balance of the Antarctic Ice Sheet: observational aspects. In *Mass balance of the Cryosphere: Observations and modeling of contemporary and future changes (J.L. Bamber and A.J. Payne, eds.)*. Cambridge: Cambridge University Press, 459-489.

Berg, 2005 – Chapter 4

Berger & Loutre 1991b – Chapter 2 – *1991a but not 1991b*

Berger A, Loutre MF (1991a) Insolation values for the climate of the last 10 million years. Quaternary Sciences Reviews 10:297-317

Bergstrom DM, Chown SL (1999) Life at the front: history, ecology and changes on southern ocean islands. Trends in Ecology and Evolution 14:427-477

Berkman et al. 1998 – Chapter 2

Berner RA (1990) Atmospheric Carbon Dioxide Levels Over Phanerozoic Time. Science 249:1382-1386

Berteaux D, Réale D, McAdam AG, Boutin S (2004) Keeping pace with fast climate change: can Arctic life count on evolution? Integrative Comp Biol 44: 140-151

Bertler, N. A. N., Barrett, P. J., Mayewski, P. A., Fogt, R. L., Kreutz, K. J. and Shulmeister, J. 2004. El Niño suppresses Antarctic warming. *Geophysics Research Letters* **31**: L15207, doi:10.1029/2004GL020749.

Best et al., 2004 – chapter 3

Best, S.E., V.O. Ivchenko, K.J. Richards, R.D. Smith, and R.C. Malone, 1999: Eddies in numerical models of the Antarctic Circumpolar Current and their influence on the mean flow. J. Phys. Oceanogr., 29(3), 328-350.

Beyer A, Matthies M (2001) Long-range transport potential of semivolatile organic chemicals in coupled air-water systems. Environ Sci Pollut Res 8:173-179

Bi D., W.F. Budd, A.C. Hirst and X. Wu (2002). Response of the Antarctic circumpolar current transport to global warming in a coupled model. Geophys. Res. Lett. 29 (24) 2173, doi:10.1029/2002GL015919.

Bi D.H., W.F. Budd, A.C. Hirst and X.R. Wu (2001). Collapse and reorganisation of the Southern Ocean overturning under global warming in a coupled model. Geophys. Res. Let. 28 (20), 3927-3930

Bindoff, N., Willebrand, J., Artale, V., Cazenave, A., Gregory, J., Gulev, S., Hanawa, K., Le Quéré, C., Levitus, S., Nojiri, Y., Shum, C., Talley, L. and Unnikrishnan, A. 2007. Observations: oceanic climate change and sea level. In Climate Change 2007: The Physical Science Basis. Contribution of Working

Group 1 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (eds. S. Solomon, D. Qin, and M. Manning). Cambridge University Press. United Kingdom and New York, NY, USA.

Bindschadler R, 2006, The environment and evolution of the West Antarctic ice sheet: setting the stage. Phil. Trans. R. Soc. A **364**, 1583–1605 doi:10.1098/rsta.2006.1790.

Bindschadler R. A., M.A. King, R.B. Alley, S. Anandakrishnan and L. Padman, 2003, Tidally controlled stick-slip discharge of a West Antarctic ice stream, *Science* **301**, 1087–1089.

Bindschadler, R., 2001. Flow History of Pine Island Glacier from Landsat Imagery, *Eos* Transactions (Fall 2001).

Bindschadler, R.A. and P. L. Vornberger, 1998. Changes in West Antarctic ice sheet since 1963 from Declassified Satellite Photography, Science, 279, 689-692.

Bindschadler, R.A., 2002, History of lower Pine Island Glacier, West Antarctica, from Landsat. imagery, *J. Glaciol.*, **48** (163), 536-544.

Bitz, C. M., P.R. Gent, R.A. Woodgate, M.M. Holland and R. Lindsay (2006). The influence of sea ice on ocean heat uptake in response to increasing CO2. J. Clim. 19, 2437-2450

Biuw, M., L. Boehme, C. Guinet, M. Hindell, D. Costa, J.-B. Charrassin, F. Roquet, F. Bailleul, M. Meredith, S. Thorpe, Y. Tremblay, B. McDonald, Y.-H. Park, S.R. Rintoul, N. Bindoff, M. Goebel, D. Crocker, P. Lovell, J. Nicholson, F. Monks and M.A. Fedak, Variations in behaviour and condition of a Southern Ocean top predator in relation to in situ oceanographic conditions. *Proceedings of the National Academy of Sciences*,104, 34, 13705-13710. 2007

Björck S, Håkansson H, Olsson S, Barnekow L, Janssens J (1993) Palaeoclimatic studies in South Shetland Islands, Antarctica, based on numerous stratigraphic variables in lake sediments. Journal of Paleolimnology 8:233-272

Björck S, Håkansson H, Zale R, Karlen W, Jonsson BL (1991a) A late Holocene sediment sequence from Livingston Island, South Shetland Islands, with palaeoclimatic implications. Antarctic Science 3:61-72

Björck S, Malmer N, Hjort C, Sandgren P, Ingólfsson O, Wallen B, Smith RIL, Jonsson BL (1991b) Stratigraphic and paleoclimatic studies of a 5500-year-old moss bank on Elephant Island, Antarctica. Arctic and Alpine Research 23:361-374

Björck S, Olsson S, Ellis-Evans JC, Häkansson H, Humlum O, de Lirio JM (1996) Late Holocene palaeoclimatic records from lake sediments on James Ross Island, Antarctica. Palaeogeography Palaeoclimatology Palaeoecology 121:195-220

Björck S., Hjort C., Ingólfsson Ó., Zale R., J. I (1996) Holocene deglaciation chronology from lake sediments. Geomorphological map of Byers Peninsula LIBGSSS-A120. British Antarctic Survey, Cambridge

Blaker, A. T., B. Sinha, V. O. Ivchenko, N. C. Wells, and V. B. Zalesny, 2006: Identifying the roles of the ocean and atmosphere in creating a rapid equatorial response to a Southern Ocean anomaly, Geophys. Res. Lett., 33, L06720, doi:10.1029/2005GL025474.

Blankenship, D.D., Bentley, C.R., Rooney, S.T. and Alley, R.B., 1986, Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. *Nature* **322**, 54–57.

Bleck, R., Rooth, C., Hu, D., Smith, L.T., 1992. Ventilation patterns and mode water formation in a wind- and thermodynamically driven isopycnic coordinate model of the North Atlantic. Journal of Physical Oceanography 22, 1486-1505.

Block 1984 – chapter 3

Bloom, S., A. da Silva, D. Dee, M. Bosilovich, J.-D. Chern, S. Pawson, S. Schubert, M. Sienkiewicz, I. Stajner, W.-W. Tan, and M.-L. Wu, "Documentation and Validation of the Goddard Earth Observing System (GEOS) Data Assimilation System, Version 4", *NASA/TM-2005-104606*, *26*, 187 pp., NASA Goddard Space Flight Center, Greenbelt, Md. (2005).

Bloss, W. J., Lee, J. D., Heard, D. E., Salmon, R. A., Bauguitte, S. J.-B., Roscoe, H. K. and Jones, A. E. , Observations of OH and HO_2 radicals in coastal Antarctica,

Blumberg, A. F. and G. L. Mellor, 1987: A description of a three-dimensional coastal ocean circulation model, In Three-Dimensional Coastal Ocean Models, N. S. Heaps (Ed.), 1-16, American Geophysical Union, Washington, DC.

Blunier T, Schwander J, Stauffer B, Stocker T, Dallebach A, Indermühle A, Tschumi J, Chappellaz J, Raynaud D, Barnola JM (1997) Timing of the Antarctic Cold Revearsal and the atmospheric CO_2 increase with respect to the Younger Dryas event. Geophys.Res.Letters 24:2683 - 2686

Blunier T, Spahni R, Barnola JM, Chappellaz J, Loulergue L, Schwander J (2007) Synchronisation of ice core records via atmospheric gases. Climate of the Past 3:325-330

Bockheim, J.G. 1995. Permafrost distribution in the Southern Circumpolar Region and its relation to the environment: a review and recommendations for further research. Permafrost & Periglacial Processes 6:27-45.

Bockheim, J.G. and K.J. Hall. 2002. Permafrost, active layer dynamics and periglacial environments of continental Antarctica. South Afr. J. Sci. 98:82-90.

Bockheim, J.G., I.B. Campbell, and M. McLeod. 2007. Permafrost distribution and active-layer depths in the McMurdo Dry Valleys, Antarctica. Permafrost & Periglacial Processes:in press.

Bockheim, JG. 2004. Workshop goals and approach of Workshop on Antarctic permafrost and soils, Madison, 14-18 November 2004, oral presentation.

Bodeker, G.E., H. Shiona, H. Eskes, "Indicators of Antarctic ozone depletion", *Atmos. Chem. Phys.*, **5**, 2603-2615, (2005).

Böning, C. W., A. Dispert, M. Visbeck, S. R. Rintoul, and F. Schwarzkopf, 2008. Observed multi-decadal ocean warming and density trends across the Antarctic Circumpolar Current, submitted.

Bonn WJ, Gingele FX, Grobe H, Mackensen A, Fütterer DK (1998) Palaeoproductivity at the Antarctic continental margin: opal and barium records for the last 400 ka. Palaeogeography, Palaeoclimatology, Palaeoecology 139: 195-211

Bonstra R (2004) Coping with changing northern environments: the role of the stress axis in birds and mammals. Integr Comp Biol 4:95-108

Bopp, L., et al. (2001), Potential impact of climate change on marine export production, *Global Biogeochemical Cycles*, *15*, 81-99.

Borges, A. V., et al. (2008), Inter-annual variability of the carbon dioxide oceanic sink south of Tasmania, *Biogeosciences Discuss.*, *5*, 141-145.

Borzatta, E. and D. Trombotto. 2003. Analysis of permafrost thickness in Antarctica using VES and MTS data. In: Haeberli, W. and D. Brandova (eds.) Permafrost; extended abstracts reporting current research and new information. 8th Internat. Conf.on Permafrost, Zurich, Switzerland.

Bosch et al. 1987 - chapter 4

Bracegirdle T.J., Connolley W.M., Turner J. (2008) Antarctic climate change over the Twenty First Century. Journal of Geophysical Research - Atmospheres 113:D03103, doi:03110.01029/02007JD008933

Bracegirdle, T. J., W. Connolley, and J. Turner, 2007: Antarctica climate change over the 21st century. *Journal of Geophysical Research - Atmospheres*, **In Press**.

Brachfeld S, Domack E, Kissel C, Laj C, Leventer A, Ishman S, Gilbert R, Camerlenghi A, Eglinton LB (2003) Holocene history of the Larsen-A Ice Shelf constrained by geomagnetic paleointensity dating. Geology 31:749-752

Bradley RS, Alverson K, Pedersen TF (2002) Challenges of a Changing Earth: Past perspectives, Future Concerns. In: Alverson K, Bradley R, Pedersen T (eds) Paleoclimate, Global Change and the Future. Springer Verlag, p 163-168

Brandefelt, J., and E. Källén, 2004: The response of the Southern Hemisphere atmospheric circulation to an enhanced greenhouse gas forcing. *J. Climate*, **17**, 4425–4442.

Brandt A, Gooday AJ et al (2007) First insights into the biodiversity and biogeography of the Southern Ocean deep sea. Nature 447(7142): 307–311

Brévière, E., et al. (2006), Changes of the oceanic CO2 sink in the Eastern Indian sector of the Southern Ocean, *Tellus B*, *52*, 438-446.

Brewer, P. G. (1978), Direct observation of oceanic CO2 increase, *Geophys. Res. Let.*, *5*, 997-1000.

Brey & Clarke 1993 - chapter 4

Brierley, A.S., P.G. Fernandes, M.A. Brandon, F. Armstrong, N.W. Millard, S.D. McPhail, P. Stevenson, M. Pebody, J. Perrett, M. Squires, D.G. Bone, and G. Griffiths, Antarctic Krill under sea ice: elevated abundance in a narrow band just south of ice edge. *Science* 295, pp 1890–1892, 2002

Brodte E, Knust R, Pörtner HO (2006) Temperature dependent energy allocation to growth in Antarctic and boreal eelpout (Zoarcidae). Pol Biol 30: 95–107

Bromwich 1988 – chapter 3

Bromwich D.H., Fogt R.L., Hodges K.I., Walsh J.E., 2007: A tropospheric assessment of the ERA-40, NCEP, and JRA-25 global reanalyses in the polar regions, *J. Geophys. Res.*, **112,** Art. No. D10111 MAY 22 2007

Bromwich, D. H., A. N. Rogers, P. Kållberg, R. I. Cullather, J. W. C. White, and K. J. Kreutz, 2000: ECMWF analyses and reanalyses depiction of ENSO signal in Antarctic precipitation. *J. Climate*, **13,** 1406-1420.

Bromwich, D. H., F. M. Robasky, R. I. Cullather, and M. L. Van Woert, 1995: The atmospheric hydrologic cycle over the Southern Ocean and Antarctica from operational numerical analyses. *Mon. Wea. Rev.*, **123**, 3518-3538.

Bromwich, D. H., J. J. Cassano, T. Klein, G. Heinemann, K. M. Hines, K. Steffen and J. E. Box, Mesoscale modeling of katabatic winds over Greenland with the Polar MM5, *Mon. Wea. Rev., 129*, 2290-2309, 2001.

Bromwich, D. H., Z. Guo, L. Bai, and Q.-S. Chen, 2004b, Modeled Antarctic precipitation. Part I: Spatial and temporal variability, *J. Climate*, **17**, 427–447.

Bromwich, D.H., A.J. Monaghan, and Z. Guo, 2004a: Modeling the ENSO modulation of Antarctic climate in the late 1990s with the Polar MM5. *J. Clim.*, **17**, 109–132.

Bromwich, D.H., A.N. Rogers, P. Kallberg, R.I. Cullather, J.W.C. White, and K.J. Kreutz, 2000: ECMWF analyses and reanalyses depiction of ENSO signal in Antarctic precipitation. *J. Climate,* **13**, 1406-1420.

Bromwich, D.H., Z. Guo, L. Bai, and Q-S. Chen, 2004a: Modeled Antarctic precipitation. Part I: spatial and temporal variability. *J. Climate,* **17**, 427-447.

Bronge C (1992) Holocene climatic record from lacustrine sediments in a freshwater lake in the Vestfold Hills, Antarctica. Geografiska Annaler 74A:47-58

Brook EJ, White JWC, Schilla ASM, Bender ML, Barnett B, Severinghaus JP, K.C.Taylor, Alley RB, Steig EJ (2005) Timing of millenial-scale climate change at Siple Dome, West Antarctica, during the last glacial period. Quarternary Science Reviews 24:1333-1343

Brown, R.J.E. and T.L. Péwé. 1973. Distribution of permafrost in North America and its relationship to the environment: a review, 1963-1973. In: Permafrost: North American Contribution to the 2nd International Permafrost Conference, Yakutsk,USSR. Natl. Acad. Sci., Washington, D.C., pp. 71-100.

Bryan F.O., G. Danabasoglu, P.R. Gent and K. Lindsay (2006). Changes in ocean ventilation during the 21st century in the CCSM3. Ocean Model. 15, 141-156.

Bryan, K. & Cox, M.D., 1968: A non-linear model of an ocean drivem by wind and differential heating: Part 1. Description of the Three-dimensional velocity and density fields. Journal of the Atmospheric Sciences, 25 (6), 945-967.

Bryden and Nurser, 2002 – chapter 3 – 2003?

Bryden, H. L. and Nurser, A. J. G. (2003). Effects of strait mixing on ocean stratification. *J. Phys. Oceanogr.* **33,** 1870-1872.

Budd, W., 1966. The dynamics of the Amery Ice Shelf. Journal of Glaciology 6(XX), 335-358.

Budd, W.F., and I. Simmonds, 1991: The impact of global warming on the Antarctic mass balance and global sea level. *In Proceedings of the International Conference on the Role of the Polar Regions in Global Change, Vol. 1*, University of Alaska Fairbanks, Alaska, 11-15 June 1990, 489-494.

Budd, W.F., and I.N. Smith (1982), Large-scale numerical modeling of the Antarctic Ice Sheet. *Annals of Glaciology 3*, 42-49.

Budd, W.F., D. Jenssen, and U. Radok (1971), *Derived Physical Characteristics of the Antarctic Ice Sheet.* ANARE Interim Report, Series A(IV), Glaciology Publication no. 18, University of Melbourne.

Buesseler, K. O., et al. (2004), The effects of iron fertilization of carbon sequestration in the Southern Ocean, *Science*, *304*.

Burgess JS, Spate AP, Shevlin J (1994) The onset of deglaciation in the Larsemann Hills, eastern Antarctica. Antarctic Science 6:491 - 495

Bustnes JO, Tveraa T, Varpe Ø, Henden JA, Skaare JU (2007) Reproductive performance and organochlorine pollutants in an Antarctic marine top predators: the south polar skua. Environ Int 33:911-918

Cai et al. 1999 – chapter 3

Cai, W. (2006), Antarctic ozone depletion causes in intensification of the Southern Ocean super-gyre circulation, *Geophys. Res. Lett.*, 33, doi:10.1029/2005GL024911.

Cai, W., P.H. Whetton, and D.J. Karoly, 2003: The response of the Antarctic Oscillation to increasing and stabilized atmospheric CO2. *J. Climate*, **16**, 1525–1538.

Cai, W.J. and T. Cowan, 2007: Trends in Southern Hemisphere circulation in IPCC AR4 models over 1950-99: Ozone depletion versus greenhouse forcing. *J. Climate*, **20**, 681-693.

Caldeira, K., and P. B. Duffy (2000), The role of the Southern Ocean in uptake and storage of anthropogenic carbon dioxide, *Science*, *287*, 620-622.

Cameron RE (1972) Pollution and conservation of the Antarctic terrestrial ecosystems. In: Parker BC (ed) Proc Coll Conservation Problems in Antarctica. Allen Press, Lawrence, Kansas, pp 267-305

Campbell, I.B. and G.G.C. Claridge. 2006. Permafrost properties, patterns, and processes in the Transantarctic Mountains region of Antarctica. Permafrost & Periglacial Processes 17:215-232.

Campbell, I.B., G.G.C. Claridge, M.R. Balks, and D.I. Campbell. 1997. Moisture content in soils of the McMurdo Sound and Dry Valley region of Antarctica, pp. 61-76. In:Lyons, W.B., C. Howard-Williams, and I. Hawes (eds.) Ecosystem processes in Antarctic ice-free landscapes, Balkema, Rotterdam.

Canadell, J. G., et al. (2007), Contributions to accelerating atmospheric CO2 growth from economic activity, carbon intensity, and efficiency of natural sinks, *PNAS*, doi:10.1073/pnas.0700609104.

Cannone, N., Ellis Evans, J. C., Strachan, R., Guglielmin, M. 2006. Interactions between climate, vegetation and active layer in Maritime Antarctica. *Antarctic Science*, 18(3): 323-333.

Carleton A M (2003) Atmospheric teleconnections involving the Southern Ocean. Journal of Geophysical Research, Oceans, 108, 8080.

Carmack, E.C. and T.D. Foster (1975) On the flow of water out of the Weddell Sea, Deep-Sea Res., 22, 711-724.

Carsey, F., Behar, A., Lane, A.L., Realmuto, V., Engelhardt, H., 2002. A borehole camera system for imaging the deep interior of ice sheets. *J. Glaciol.* **48** (163), 622–628.

Catania, G. A., H. Conway, C. F. Raymond, and T. A. Scambos, 2006, Evidence for floatation or near floatation in the mouth of Kamb Ice Stream, West Antarctica, prior to stagnation, *J. Geophys. Res.*, **111**, F01005, doi:10.1029/2005JF000355.

Chambers (1963) – chapter 3

Chambers, D.P, Mehltaff, C.A., Urban, T.J., Fujii, D. and Nerem, R.S., 2002. Low frequency variations in global mean sea level:1950-2000. Journal of Geophysical Research, 107, 10.1029/2001JC001089.

Chambers, M. J. G. 1967. Investigations of patterned ground at Signy Island, South Orkney Islands: II, Temperature regimes in the active layer. *British Antarctic Survey Bulletin*, 10, 71-83.

Chapman.W.L. and Walsh, J. E. 2007. **A Synthesis of Antarctic Temperatures**. *Journal of Climate* **20**: 4096-4117.

Chappellaz J, Bluniert T, Raynaud D, Barnola JM, Schwander J, Stauffert B (1993) Synchronous changes in atmospheric CH4 and Greenland climate between 40 and 8 kyr BP. Nature 366:443-445

Chapter 6

Charlson, R.J., Lovelock, J.E., Andreae, M.O., and Warren, S.G., Oceanic phytoplankton, atmospheric sulphur, cloud albedo and climate, Nature (323), 655-661, 1987.

Chassignet, E.P. & Verron, J.: 2006: Ocean Weather Forecasting: An Integrated View of Oceanography. (Eds.). Springer, 577pp.

Chassignet, E.P., H.E. Hurlburt, O.M. Smedstad, G.R. Halliwell, A.J. Wallcraft, E.J. Metzger, B.O. Blanton, C. Lozano, D.B. Rao, P.J. Hogan, and A. Srinivasan, 2006: Generalized vertical coordinates for eddy-resolving global and coastal ocean forecasts. Oceanography, 19, 20-31.

Chassignet, E.P., Hurlburt, H.E., Smedstad, O.M., Halliwell, G.R., Hogan, P.J., Wallcraft, A.J., Baraille, R., & Bleck, R., 2007. The HYCOM (HYbrid Coordinate Ocean Model) data assimilative system. Journal of Marine Systems, 65, 60-83.

Chavez F P, Ryan J, Lluch-Cota S E and Niquen M (2003) From Anchovies to Sardines and Back: Multidecadal Change in the Pacific Ocean. Science, 299, 217-221

Chelton D. B., M. G. Schlax, D. L. Witter, and J. G. Richman, 1990. Geosat altimeter observations of the surface circulation of the Southern ocean. *J. Geophys. Res*., **95**, 17877–17903.

Chelton, D. B., R. A. deSzoeke, M. G. Schlax, K. El Naggar and N. Siwertz (1998). Geographical variability of the first-baroclinic Rossby radius of deformation. *J. Phys. Oceanogr.* **28**, 433-460.

Chen et al. (1996) – chapter 3

Chen et al. (1998) – Chapter 2

Chen et al. 1997 – Chapter 2 (1997a and 1997b but not 1997)

Chen L B, DeVries AL, Cheng C-HC (1997a) Convergent evolution of antifreeze glycoproteins in Antarctic notothenioid fish and Arctic cod. Proc Natl Acad Sci USA 94: 3817–3822

Chen L B, DeVries AL, Cheng C-HC (1997b) Evolution of antifreeze glycoprotein gene from trypsinogen gene in Antarctic notothenioid fish. Proc Natl Acad Sci USA 94: 3811–3816

Chen, C. T., and F. J. Millero (1979), Gradual increase of oceanic CO2, *Nature*, *277*, 205-206.

Chen, J. L., C. R. Wilson, D. D. Blankenship, and B. D. Tapley, 2006, Antarctic mass rates from GRACE, *Geophys. Res. Lett.*, **33**, L11502, doi:10.1029/2006GL026369.

Chen, X. 1993. Permafrost around CGWS, Antarctica. In: *International Conference on Permafrost, Proceedings*, 6(1), 84-88.

Cheng C-HC (1998) Origin and mechanism of evolution of antifreeze glycoproteins in polar fishes. In: di Prisco G, Pisano E, Clarke A (Eds) Fishes of Antarctica. A Biological Overview (pp 311-328). Springer, Milano

Cheng C-HC, Chen L B (1999) Evolution of an antifreeze glycoprotein. Nature 401: 443–444

Cheng C-HC, Chen L, Near TJ, Jin Y (2003) Functional antifreeze glycoprotein genes in temperate-water New Zealand nototheniid fish infer an Antarctic evolutionary origin. Mol Biol Evol 20: 1897-1908

Chisholm *et al*. 2001 – chapter 3

Chisholm, S. W. (2000), Stirring times in the Southern Ocean, *Nature*, *407*, 685-687.

Chown & Convey 2007 – chapter 3

Chown 2001 – chapter 4

Chown and Convey 2006, 2007 – Chapter 3

Chown SL, Gremmen NJM, Gaston KJ (1998) Ecological biogeography of southern ocean islands: species-area relationships, human impacts and conservation. The American Naturalist 152:562-575

Christoffersen, P., and S. Tulaczyk, 2003, Response of subglacial sediments to basal freeze-on, 1. Theory and comparison to observations from beneath the West Antarctic Ice Sheet, J. Geophys. Res., **108**(B4), 2222, doi:10.1029/2002JB001935.

Christoph et al., 1998 – chapter 3

Church, J., Wilson, S., Woodworth, P. and Aarup, T. 2007. Understanding sea level rise and variability. Meeting report. EOS, Transactions of the American Geophysical Union, 88(4), 23 January 2007, p43.

Church, J.A. and White, N.J. 2006 A 20th century acceleration in global sea-level rise. Geophysical Research Letters, 33, L01602, 10.1029/2005GL024826.

Church, J.A., J.M. Gregory, P. Huybrechts, M. Kuhn, K. Lambeck, M.T. Nhuan, D. Qin, and P.L. Woodworth, 2001. Changes in sea level. In Climate change 2001: the scientific basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change, J.T. Houghton, Y. Ding, D.J. Griggs, M. Noguer, P.J. van der Linden, X. Dai, K. Maskell and C.A. Johnson, eds. Cambridge University Press, Cambridge, 881pp.

Church, J.A., White, N.J., Aarup, T., Wilson, S.L., Woodworth, P.L., Domingues, C., Hunter, J. and Lambeck, K. 2008. Global sea level: past, present and future. Sustainability Science (submitted).

Church, J.A., White, N.J., Coleman, R., Lambeck, K. and Mitrovica, J.X. 2004. Estimates of the regional distribution of sea-level rise over the 1950 to 2000 period. Journal of Climate, 17, 2609–262.

Claridge GGC, Campbell IB, Powell HKJ, Amin ZH, Balks MR (1995) Heavy metal contamination in some soils of the McMurdo region, Antarctica. Antarct Sci 7:9-14

Clark et al 2008b – *Paraceradocus* and *Odontaster validus* paper.

Clark et al. 2008a - chapter 4 - missing page 65

Clark M. pers. Obs – chapter 4, page 66

Clark MS, Fraser KPP, Peck LS (2008b) The HSP70 heat shock response in Antarctic marine molluscs. Cell Stress & Chaperones

Clark MS, Fraser KPP, Peck LS (2008c) The environmental HSP70 heat shock response in the Antarctic limpet: *Nacella concinna*. Cell Stress & Chaperones

Clark PU, Mitrovica JX, Milne GA, Tamisiea ME (2002) Sea-level fingerprinting as a direct test for the source of global meltwater pulse IA. Science 295:2438-2441

Clarke A (1988) Seasonality in the Antarctic marine environment. Comp Biochem Physiol 90(B): 461-473

Clarke A (1991a) Cold adaptation. J Zool 225: 691–699

Clarke A (1991b) What is cold adaptation and how should we measure it ? Amer Zool 31: 81–92

Clarke A (1992) Reproduction in the cold: Thorson revisited. Invert Reprod & Dev 22: 175-184

Clarke A (1998). Temperature and energetics: a review of cold ocean physiology. In Cold Ocean Physiology (eds HO Pörtner, RC Playle), Soc Exp Biol Seminar Ser 66, pp. 3–30

Clarke A, Aronson RB, Crame JA, Gili J-M, Blake DB (2004) Evolution and diversity of the benthic fauna of the Southern Ocean continental shelf. Antarctic Sci 16: 559-568

Clarke A, Johnston NM (2003) Antarctic marine benthic diversity. Oceanogr Mar Biol Ann Rev 41: 47-114

Clarke A, Murphy EJ, Meredith MP, King JC, Peck LS, Barnes DKA, Smith RC (2006) Climate change and the marine ecosystem of the western Antarctic Peninsula. Phil Trans R Soc B 362: 149-166

Claussen, M. et al. (2002), Earth system models of intermediate complexity: closing the gap in the spectrum of climate system models, Clim. Dyn., 18, 579-586.

CLIMAP (1981) Seasonal Reconstructions of the Earth's Surface at the Last Glacial Maximum Map Chart Series MC-36, Geological Society of America, Boulder, CO

Cocca E, Ratnayake-Lecamwasam M, Parker SK, Camardella L, Ciaramella M, di Prisco G, Detrich HW III (1995) Genomic remnants of α-globin genes in the hemoglobinless Antarctic icefishes. Proc Natl Acad Sci USA 92: 1817-1821

Colbecki A. - Chapter 4 page 73

Colhoun EC (1991) Geological evidence for changes in the East Antarctica ice sheet (60°-120°E) during the last glaciation. Polar Record 27:345-355

Collins, M. and the CMIP Modelling Groups, 2005: El Niño- or La Niña-like climate change? *Climate Dynamics*, 24, 89-104.

Collins, W.D. and Coauthors, 2006: The Community Climate System Model Version 3 (CCSM3). Journal of Climate, 19, 2122-2143.

Comiso 1995 – Chapter 3

Comiso and Gordon (1998) – chapter 3

Comiso, J. C. 2000. Variability and trends in Antarctic surface temperatures from in situ and satellite infrared measurements. *Journal of Climate* **13**: 1674-1696.

COMNAP-AEON (2001) Summary and environmental monitoring activities in Antarctica. Council of Managers of National Antarctic Programs, Hobart

Conil S. And C.G. Menéndez (2006). Climate fluctuations of the Weddell Sea and its surroundings in a transient climate change scenario. Clim. Dyn 27 (1): 83-99

Conlan KE, Rau GH, McFeters GA, Kvitek RG (2000) Influence of McMurdo Station sewage on Antarctic marine benthos: evidence from stable isotopes, bacteria, and sewage on Antarctic marine benthos: evidence from stable isotopes, bacteria, and biotic indices. In: Davison W, Howard-Williams C, Broady P (eds) Antarctic Ecosystems: Models for Wider Ecological Understanding. Caxton Press, Christchurch, pp 315-318

Connolley W M (2002) Long term variability of the Antarctic Circumpolar Wave. Journal of Geophysical Research, Oceans, 108, 8076.

Connolley, 2003 – chapter 3

Connolley, 2006 – Chapter 4

Convey 1996 Antarct Sci – chapter 3

Convey 1996c – chapter 3

Convey 2001 – chapter 3

Convey 2003 Ant Res Ser, 2007 RiSCC – chapter 3

Convey 2006 RiSCC – chapter 4

Convey 2007 – Chapter 3

Convey 2007 CEP workshop report – chapter 3

Convey 2007 PPRST – chapter 3

Convey 2007b – chap 1

Convey and Chown 2006 RiSCC - chapter 4

Convey and McInnes (2005) – chap 1

Convey et al in press – chapter 3

Convey et al. 2006 RiSCC LH – chapter 3

Convey *et al.* 2006a – chap 1 & chap 4

Convey *et al.* 2006b – chap 1

Convey P (1996) The influence of environmental characteristics on life history attributes of Antarctic terrestrial biota. Biological Reviews of the Cambridge Philosophical Society 71:191-225

Convey P, Gibson JAE, Hillenbrand C-D, Hodgson DA, Pugh PJA, Smellie JL, Stevens MI (2008) Antarctic terrestrial life - challenging the history of the frozen continent? Biological Reviews in press:00-00

Convey P, Stevens MI (2007) Antarctic Biodiversity. Science 317:1877-1878

Conway et al. 1999b – Chapter 2 – *1999 but not 1999b*

Conway H, Hall BL, Denton GH, Gades AM, Waddington ED (1999a) Past and future grounding-line retreat of the West Antarctic Ice Sheet. Science 286:280-283

Conway, H., B. Hall, G. Denton, A. Gades, and E. Waddington, 1999, Past and future grounding-line retreat of the West Antarctic Ice Sheet, *Science*, 286, 280–283.

Cook, A., A. Fox, D. Vaughan, and J. Ferrigno, 2005, Retreating glacier fronts on the Antarctic Peninsula over the past half-century. *Science* **308**, 541-544.

Cooper, A. P. R., 1997. Historical observations of Prince Gustav Ice Shelf. Polar Record 33(187), 285-294.

Corlett, G.K., Barton, I.J., Donlan, C.J., Edwards, M.C., Good, S.A., Horrocks, L.A., Llewellyn-Jones, D.T., Merchant, C.J., Minnett, P.J., Nightingale, T.J., Noyes, E.J., O'Carroll, A.G., Remedios, J.J., Robinson, I.S., Saunders, R.W. and Watts, J.G. (2006) The accuracy of SST retrievals from AATSR: An initial assessment through geophysical validation against in situ radiometers, buoys and other SST data sets. *Advances in Space Research*, 37, (4), 764-769.

Corr H. F. J., A. Jenkins, K. W. Nicholls, and C. S. M. Doake, 2002 Precise measurement of changes in ice-shelf thickness by phase-sensitive radar to determine basal melt rates. *Geophys. Res. Lett.* **29**(8): 10.1029/2001GL014618.

Corsolini S, Covaci A, Ademollo N, Focardi S, Schepens P (2006) Occurrence of organochlorine pesticides (OCPs) and their enantiometric signature, and concentrations of polybrominated diphenyl ethers (PBDEs) in the Adélie penguin food web, Antarctica. Environ Pollut 140:371-382

Coulson in press – from chap 1

Covey, C., , K. M. AchutaRao, U. Cubasch, P. Jones, S. J. Lambert, M. E. Mann, T. J. Phillips, and K. E. Taylor, 2003: An overview of results from the Coupled Model Intercomparison Project (CMIP). *Global Planet. Change*, **37**, 103–133, doi:10.1016/S0921-8181(02)00193-5.

Cowan EA, Brachfeld, SA, Powell RD, Schoolfield SC (2006) Terrane-specific rock magnetic characteristics preserved in glacimarine sediment from southern coastal Alaska. Canadian J Earth Sci 43(9): 1269–1282

Coward, A. & de Cuevas, B., 2005: The OCCAM 66 level model: model description, physics, initial conditions and external forcing, Tech. rep., Southampton Oceanography Centre, Internal Document No. 99, 83 pp. (http://eprints.soton.ac.uk/18792)

Cox, M.D., 1984. A primitive equation 3-dimensional model of the ocean. Tech. Rep. 1, Geophys. Fluid Dyn. Lab, Natl. Oceanic and Atmos. Admin., Princeton Uni. Press, Princeton, N.J.

Craven, M., I. Allison, R. Brand, A. Elcheikh, J. Hunter, M. Hemer, S. Donoghue, 2004, Initial borehole results from the Amery Ice Shelf hot-water drilling project. *Ann. Glaciol.* **39**, 531-539.

Crawford, J. H., Davis, D. D., Chen, G., Buhr, M., Oltmans, S., Weller, R., Mauldin, L., Eisele, F., Shetter, R., Lefer, B., Arimoto, R., and Hogan, A.: Evidence for photochemical production of ozone at the South Pole surface, Geophys. Res. Lett., 28, 3641-3644, 2001.

Cremer H, Gore D, Melles M, Roberts D (2003) Palaeoclimatic significance of late Quaternary diatom assemblages from southern Windmill Islands, East Antarctica. Palaeogeography Palaeoclimatology Palaeoecology 195:261-280

Cremer H, Heiri O, Wagner B, Wagner-Cremer F (2007) Abrupt climate warming in East Antarctica during the early Holocene. Quaternary Science Reviews 26:2012-2018

Cromer L, Gibson JAE, Swadling K, Ritz DA (2005) Faunal microfossils: Indicators of Holocene ecological change in a saline Antarctic lake. Palaeogeography, Palaeoclimatology, Palaeoecology 221:83-97

Cromer L, Gibson JAE, Swadling KM, Hodgson DA (2006) Evidence for a lacustrine faunal refuge in the Larsemann Hills, East Antarctica, during the Last Glacial Maximum. Journal of Biogeography 33 1314-1323

Crosta X, Armand LK, Romero O, Pichon J-J (*2005*) The biogeography of major diatom taxa in Southern Ocean sediments: 2. Open ocean related species. Palaeogeography, Palaeoclimatology, Palaeoecology 223:66-92

Crosta X, Debret M, Denis D, Courty M-A, Ther O (2007) Holocene long- and short-term climate changes off Adelie Land, East Antarctica. Geochemistry Geophysics Geosystems 8:Q11009, doi:11010.11029/12007GC001718

Crosta X, Denis D, Ther O (2008) Sea ice seasonality during the Holocene, Adelie Land, East Antarctica. Marine Micropaleontology 66:222-232

Crosta X, Pichon JJ, Burckle LH (1998) Reappraisal of Antarctic seasonal sea-ice at the Last Glacial Maximum. Geophysical Research Letters 25:2703-2706

Crosta X, Sturm A, Armand L, Pichon JJ (2004) Late Qaternary sea ice history in the Indian sector of the Southern Ocean as record by diatom assemblages. Marine Micropaleontology 50:209-223

Crowley TJ, Kim K (1995) Comparison of long-term greenhouse projections with the Geologic Record. Geophysical Research Letters 22:933-936

Croxall J P, McCann T S, Prince P A and Rothery P (1988) Reproductive performance of seabirds and seals at South Georgia and Signy Island, South Orkney Islands, 1976-1987: implications for Southerrn Ocean monitoring studies. In Antarctic Ocean resources variability (ed. D Sahrhage), pp. 261–285. Berlin, Germany: Springer.

Csatho, B., Y. Ahn, T. Yoon, C. J. van der Veen, S. Vogel, G. Hamilton, D. Morse, B. Smith, and V. B. Spikes, 2005, ICESat measurements reveal complex pattern of elevation changes on Siple Coast ice streams, Antarctica, *Geophys. Res. Lett.*, **32**, L23S04, doi:10.1029/2005GL024289.

Cullather, R. I., Bromwich, D. H. and Van Woert, M. L. 1996. Interannual variations in Antarctic precipitation related to El Nino southern oscillation. *Journal of Geophysical Research* **101**: 19109-19118.

Cullather, R. I., D. H. Bromwich, and M. L. van Woert, 1998: Spatial and temporal variability of Antarctic precipitation from atmospheric methods, *J. Climate*, **11**, 334–367.

Cunningham L, Raymond B, Snape I, Riddle MJ (2005) Benthic diatom communities as indicators of anthropogenic metal contamination at Casey Station, Antarctica. J Paleolimnol 33:499-513

Cunningham S. A., S. G. Alderson, B. A. King, M. A. Brandon, Transport and variability of the Antarctic Circumpolar Current in Drake Passage, 2003. J. Geophys. Res., **108** (C5), 8084, doi:10.1029/2001JC001147.

Curran MAJ, van Ommen TD, Morgan VI, Phillips KL, Palmer AS (2003) Ice Core Evidence for Antarctic Sea Ice Decline Since the 1950s. Science 302:1203-1206

Cushing DH (1982) Climate and Fisheries. Academic Press, London.

Dameris, M., V. Grewe, M. Ponater, R. Deckert, V. Eyring, F. Mager, S. Matthes, C. Schnadt, A. Stenke, B. Steil, C. Brühl, M.A. Giorgetta, “Long-term changes and variability in a transient simulation with a chemistry-climate model employing realistic forcing”, *Atmos. Chem. Phys.*, **5**, 2121-2145 (2005).

Dartnall (2005) – chap 1

Dartnall et al. (2005) – chap 1

Davey 1997 - chapter 4

Davey, M.K., M. Huddleston, K. Sperber, P. Braconnot, F. Bryan, D. Chen, R. Colman, C. Cooper, U. Cubasch, P. Delecluse, D. DeWitt, L. Fairhead, G. Flato, C. Gordon, T. Hogan, M. Ji, M. Kimoto, A. Kitoh, T. Knutson, M. Latif, H. Le Treut, T. Li, S. Manabe, C. Mechoso, G. Meehl, S. Power, E. Roeckner, L. Terray, A. Vintzileos, R. Voss, B. Wang, W. Washington, I. Yoshikawa, J. Yu, S. Yukimoto and S. Zebiak, 2002: STOIC: A study of coupled model climatology and variability in tropical ocean regions. *Clim. Dyn.*, **18**, 403–420.

Davis C.H., Li Y., McConnell J.R., Frey M.M., Hanna E., 2005: Snowfall-driven growth in East Antarctic Ice Sheet mitigates recent sea-level rise. *Science* **308**,1898–1901.

Davis, 20XX – chapter 3 – page 117

Davis, D., Chen, G., Buhr, M., Crawford, J., Lenschow, D., Lefer, B., Shetter, R., Eisele, F., Mauldin, L., and Hogan, A.: South Pole NO_x chemistry: an assessment of factors controlling variability and absolute levels, Atmos. Environ., 38, 5375-5388, 2004.

Dayton PK (1990) Polar benthos. In: Smith WO (Ed), Polar Oceanography. Academic Press, London, pp 631-685

Dayton PK, Mordida BJ, Bacon F (1994) Polar marine communities. Am Zool 34: 90–99

De Angelis, H., and P. Skvarca (2003), Glacier surge after ice shelf collapse. *Science 299*, 1560-1562.

de la Mare WK (1997) Abrupt mid-twentieth-century decline in Antarctic sea-ice extent from whaling records. Nature 389:57-60

de Veaux et al. (1999) – chapter 3

De Vries AL (1982) Biological antifreeze agents in coldwater fishes. Comp Biochem Physiol 73A(4): 627–640

Decker, E.R. 1974. Preliminary geothermal studies of the Dry Valley Drilling Project holes at McMurdo Station, Lake Vanda, Lake Vida, and New Harbor, Antarctica. *Bulletin - Dry Valley Drilling Project (DVDP)*, 4, 22-23.

Decker, E.R. and G.J. Bucher. 1977. Geothermal studies in Antarctica. Antarctic Journal of the United States 12(4):102-104.

DeConto RM, Pollard D (2003) Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO2. Nature 421:245-249

Delworth, T.L., A.J. Broccoli, A. Rosati, R.J. Stouffer, V. Balaji, J.A. Beesley, W.F. Cooke, K.W. Dixon, J. Dunne, K.A. Dunne, J.W. Durachta, K.L. Findell, P. Ginoux, A. Gnanadesikan, C.T. Gordon, S.M. Griffies, R. Gudgel, M.J. Harrison, I.M. Held, R.S. Hemler, L.W. Horowitz, S.A. Klein, T.R. Knutson, P.J. Kushner, A.R. Langenhorst, H.C. Lee, S.J. Lin, J. Lu, S.L. Malyshev, P.C.D. Milly, V. Ramaswamy, J. Russell, M.D. Schwarzkopf, E. Shevliakova, J.J. Sirutis, M.J. Spelman, W.F. Stern, M. Winton, A.T. Wittenberg, B. Wyman, F. Zeng, and R. Zhang, 2006: GFDL's CM2 Global Coupled Climate Models. Part I: Formulation and Simulation Characteristics. *J. Climate*, **19**, 643–674.

Denton GH, Karlén W (1973) Holocene climatic variations: Their pattern and possible cause. Quaternary Research 3:155-205

Denton GH, Marchant DR (2000) The geologic basis for a reconstruction of a grounded ice sheet in McMurdo Sound, Antarctica, at the last glacial maximum. Geografiska Annaler Series a-Physical Geography 82A:167-211

Denton GH, Prentice ML, Burckle LH (1991) Cainozoic history of the Antarctic ice sheet. In: Tingey (ed) Geology of Antarctica. Oxford University Press, New York, p 365-433

Deprez PP, Arens M, Locher E (1999) Identification and assessment of contaminated sites at Casey Station, Wilkes Land, Antarctica. Polar Record 35:299-316

Detrich HW III, Johnson KA, Marchese-Ragona SP (1989) Polymerization of Antarctic fish tubulins at low temperature: energetic aspects. Biochemistry 28: 10085–93

Dettaï & Lecointre 2004, 2005 – Chapter 2

DeVries AL (1971) Glycoproteins as biological antifreeze agents in Antarctic fishes. Science 172: 1152–1155

DeVries AL, Cheng C-HC (2005) Antifreeze proteins and organismal freezing avoidance in polar fishes. In The Physiology of Polar Fishes (Farrell AP, Steffensen JF, Eds), Vol. 22, Fish Physiology, pp. 155-201. Elsevier Academic Press, San Diego

DeVries, 1971, 1982; 1982 - chapter 4 - missing page 64

di Prisco et al. 2007 - Chapter 2

di Prisco G (1998) Molecular adaptations in Antarctic fish hemoglobins. In: di Prisco G, Pisano E & Clarke A (Eds) Fishes of Antarctica. A Biological Overview (pp 339-353). Springer, Milano

di Prisco G (2000) Life style and biochemical adaptation in Antarctic fishes. J Mar Syst 27: 253-265

di Prisco G, Cocca E, Parker SK, Detrich HW (2002) Tracking the evolutionary loss of hemoglobin expression by the white-blooded Antarctic icefishes. Gene 295: 185-191

di Prisco G, D'Avino R, Caruso C, Tamburrini M, Camardella L, Rutigliano B, Carratore V, Romano M (1991) The biochemistry of oxygen transport in red-blooded Antarctic fishes. In: di Prisco G, Maresca B & Tota B (Eds) Biology of Antarctic Fish (pp 263-281). Springer-Verlag, Berlin, Heidelberg, New York

Dibb, J. E., Arsenault, M., Peterson, M. C., and Honrath, R. E.: Fast nitrogen oxide photochemistry in Summit, Greenland snow, Atmos. Environ., 26, 2501-2511, 2002.

Dickson, B., I. Yashayaev, J. Meincke, B. Turrell, S. Dye and J. Holfort (2002), Rapid freshening of the deep North Atlantic Ocean over the past four decades. *Nature*, 416, 832-837.

Diekmann B, Kuhn G, Rachold V, Abelmann A, Brathauer U, Fütterer DK, Gersonde R, Grobe H (2000) Terrigenous sediment supply in the Scotia Sea (Southern

Ocean): response to Late Quaternary ice dynamics in Patagonia and on the Antarctic Peninsula. Palaeogeography Palaeoclimatology Palaeoecology 162:357-387

Dinniman and Klinck, 2004 – chapter 3

Dinniman, M.S., John M. Klinck,J.M. & Smith Jr., W.O., 2007: Influence of sea ice cover and icebergs on circulation and water mass formation in a numerical circulation model of the Ross Sea, Antarctica. Journal of Geophysical Research, 112, C11013, (doi:10.1029/2006JC004036).

DiVenere VJ, Kent DV, Dalziel IWD (1994) Mid-Cretaceous paleomagnetic results from Marie Byrd Land, WestAntarctica: a test of post-100 Ma relative motion between East and West Antarctica. Journal of Geophysical Research 99:15115-15139

Doake C S M & Vaughan D G (1991) Rapid disintegration of the Wordie Ice shelf in response to atmospheric warming, Nature, 350, 328-330.

Domack E (2002) A synthesis for site 1098: Palmer Deep. In: Barker PF, Camerlenghi A, Acton GD, Ramsay ATS (eds) Proceedings of the Ocean Drilling Program, Scientific Results, Vol 178. Ocean Drilling Program, Texas A&M University, College Station TX 77843-9547, USA

Domack E, Duran D, Leventer A, Ishman S, Doane S, McCallum S, Amblas D, Ring J, Gilbert R, Prentice M (2005) Stability of the Larsen B ice shelf on the Antarctic Peninsula during the Holocene epoch. Nature 436:681-685

Domack E, Leventer A, Burnett A, Bindschadler R, Convey P, Kirby M (2003) Antarctic Peninsula Climate Variability: Historical and Paleoenvironmental Perspectives, Vol Antarctic Research Series, Vol. 79. American Geophysical Union

Domack E, Leventer A, Dunbar RB, Taylor F, Brachfeld S, Sjunneskog C, Party OLS (2001) Chronology of the Palmer Deep site, Antarctic Peninsula: a Holocene palaeoenvironmental reference for the circum-Antarctic. The Holocene 11:1-9

Domack E, O'Brien P, Harris P, Taylor F, Quilty PG, De Santis L, Raker B (1998) Late Quaternary sediment facies in Prydz Bay, East Antarctica and their relationship to glacial advance onto the continental shelf. Antarctic Science 10:234-244

Domack EW, Ishman SE, Stein AB, McClennen CE, Jull AJT (1995) Late Holocene Advance of the Muller Ice Shelf, Antarctic Peninsula - Sedimentological, Geochemical and Paleontological Evidence. Antarctic Science 7:159-170

Domack EW, Jull AJT, Anderson JB, Linick TW (1991) Mid-Holocene ice sheet recession from the Wilkes Land continental shelf, East Antarctica. In: Thomson MRA, Crame JA, Thomson JW (eds) Geological Evolution of Antarctica. Cambridge University Press, Cambridge, p 693-698

Domack EW, Jull AJT, Anderson JB, Linick TW, Williams CR (1989) Application of tandem accelerator mass-spectrometer dating to late Pleistocene-Holocene Sediments of the East Antarctic continental shelf. Quaternary Research 31:277-287

Domack EW, O'Brien P, Harris P, Taylor F, Quilty PG, De Santis L, Raker B (1999) Late Quaternary sediment facies in Prydz Bay, East Antarctica and their relationship to glacial advance onto the continental shelf. Antarctic Science 10:236-246

Doran, P. T., Priscu, J. C., Lyons, W. B., Walsh, J. E., Fountain, A. G., McKnight, D. M., Moorhead, D. L., Virginia, R. A., Wall, D. H., Clow, G. D., Fritsen, C. H., McKay, C. P. and Parsons, A. N. 2002. Antarctic climate cooling and terrestrial ecosystem response. *Nature* **415**: 517-520.

Doscher, R. & Beckman, A., 2000: Effects of a Bottom Boundary Layer Parameterization in a Coarse-Resolution Model of the North Atlantic Ocean. Journal of Atmospheric and Oceanic Technology, 17(5), 698-707

Dowdeswell JA, Bamber JL (2007) Keel depths of modern Antarctic icebergs and implications for sea-floor scouring in the geological record. Mar Geol doi: 10.1016/j.margeo.2007.04.008

Dowdeswell JA, Ó Cofaigh C, Pudsey CJ (2004) Continental slope morphology and sedimentary processes at the mouth of an Antarctic palaeo-ice stream. Marine Geology 204:203-214

Drewry, D.J. (1983), *Antarctica: Glaciological and Geophysical Folio.* Cambridge, Scott Polar Research Institute, University of Cambridge.

Duran et al., 2001 – chapter 3

Dyurgerov MB, Meier MF (2000) Twentieth century climate change: evidence from small glaciers. Proc Natl Acad Sci USA 97: 1406-1411

Eastman & Clarke 1998 – Chapter 2

Eastman & McCune 2000 – Chapter 2

Eastman JT (1993) Antarctic Fish Biology. Evolution in a Unique Environment. Academic Press, San Diego, USA, 322 pp

Eastman JT (1997) Comparison of the Antarctic and Arctic fish faunas. Cybium 12: 276-287

Eastman JT (2000) Antarctic notothenioid fishes as subjects for research in evolutionary biology. Antarctic Sci 12: 276-287

Eastman JT (2005) The nature of the diversity of Antarctic fishes. Polar Biol. 28: 93-107.

Ebinghaus R, Kock HH, Temme C, Einax JW, Richter A, Burrows JP, Schroeder WH (2002) Antarctic springtime depletion of atmospheric mercury. Environ Sci Technol 36:1238-1244

Egbert, G. D. and Ray, R. D. (2003). Semi-diurnal and diurnal tidal dissipation from TOPEX/Poseidon altimetry. *Geophys. Res. Lett.* **30**, doi:10.1029/2003GL017676.

Egginton et al. 2002 – Chapter 2

Elliptica, L, Chapter 4 – page 73

Emslie SD, Coats L, Licht K (2007) A 45,000 Yr Record of Adelie Penguins and Climate Change in the Ross Sea, Antarctica. Geology 35:61-64

Emslie SD, Woehler EJ (2005) A 9000-year record of Adélie penguin occupation and diet in the Windmill Islands, East Antarctica. Antarctic Science 17:57-66

Engelhardt, H, 2005, Thermal regime and dynamics of the West Antarctic ice sheet. *Ann. Glaciol*. **39**, 85-92.

EPICA (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. Nature 444

EPICA-community-members (2004) Eight glacial cycles from an Antarctic ice core. Nature 429:623-628

Ernsting *et al.* 1995 – chapter 3

Etheridge et al., 1996 – Chapter 2

Evans J, Dowdswell JA, O'Cofaigh C, Benham TJ, Anderson JB (2006) Extent and dynamics of the West Antarctic Ice Sheet on the outer continental shelf of Pine Island Bay during the last glaciation. Marine Geology 230:53-72

Evans J, Pudsey CJ, Ocofaigh C, Morris P, Domack E (2005) Late Quaternary Glacial History, Flow Dynamics and Sedimentation Along the Eastern Margin of the Antarctic Peninsula Ice Sheet. Quaternary Science Reviews 24:741-774

Fahnestock, M. A., T. A. Scambos, C. A. Shuman, R. J. Arthern, D. P. Winebrenner and R. Kwok, 2000, Snow Megadune Fields on the East Antarctic Plateau: Extreme Atmosphere-Ice Interaction, *Geophys. Res. Lett.*, ***27****(22)*, 3719 - 3722.

Fahnestock, M., W. Abdalati, and C. Shuman, 2002. Long melt seasons on ice shelves of the Antarctic Peninsula: an analysis using satellite-based microwave emission measurements. Annals of Glaciology 34, 127-133.

Fahrbach, E., G. Rohardt, M. Schroder and V. Strass (1994) Transport and structure of the Weddell Gyre, Ann. Geophysicae, 12, 840-855.

Fahrbach, E., M. Hoppema, G. Rohardt, M. Schroder and A. Wisotzki (2006) Causes of deep-water variation: Comment in the paper by L.H. Smedsrud, 2005, "Warming of the deep water in the Weddell Sea along the Greenwich Meridian: 1997-2001", Deep-Sea Research I, 53, 574-577.

Fahrbach, E., M. Hoppema, G. Rohardt, M. Schröder, A. Wisotzki, Decadal-scale variations of water mass properties in the deep Weddell Sea, *Ocean Dynamics*, 54, 77-91, 2004

Fahrbach, E., M. Knoche and G. Rohardt (1991) An estimate of water mass transformation in the southern Weddell Sea, *Mar. Chem.,* 35, 25-44.

Farman, J.C., B.G. Gardiner, J.D. Shanklin, "Large losses of total ozone in Antarctica reveal seasonal ClO_X/NO_X interaction", *Nature*, **315**, 207-210 (1985).

Fedorov V and Philander S G (2000) Is El Niño changing. Science, 288, 1997–2002.

Feely, R. A., et al. (2002), Seasonal and interannual variability of CO2 in the equatorial Pacific, *Deep-Sea Research Part Ii-Topical Studies in Oceanography*, *49*, 2443-2469.

Feely, R. A., et al. (2004), Impact of Anthropogenic CO_2 on the $CaCO_3$ System in the Oceans, *Science*, *305*, 362-366.

Fell 2002 – chap 1

Ferrigno, J. G., Cook, A. J., Foley, K. M., Williams, R.S. Jr., Swithinbank, C., Fox, A. J., Thomson, J. W. and Sievers, J. (2006), Coastal-Change and Glaciological Maps of the Trinity Peninsula area, Antarctica: 1843-2002 *(USGS map number I-2600-A).*

Ferrigno, J. G., R. S. Jr. Williams and J. W. Thomson (2002) Joint U.S. Geological Survey – British Antarctic Survey Fact Sheet FS-017-02, 2p.

Ferrigno, J.G., Williams, R.S., Jr., Rosanova, C.E., Lucchitta, B.K., and Swithinbank, Charles, 1998, Analysis of coastal change in Marie Byrd Land and Ellsworth Land, West Antarctica, using Landsat imagery: *Ann. Glaciol.*, **27**, 33–40.

Fink D, Mckelvey B, Hambrey MJ, Fabel D, Brown R (2006) Pleistocene Deglaciation Chronology of the Amery Oasis and Radok Lake, Northern Prince Charles Mountains, Antarctica. Earth and Planetary Science Letters 243:229-243

Finlayson-Pitts and Pitts, 1999 - chapter 4 - Year query 2000? – page 28

Finlayson-Pitts, B.J. and Pitts, J.N., "Chemistry of the Upper and Lower Atmosphere", Academic Press, 2000.

Finney BP, Gregory-Eaves I, Douglas MSV, Smol JP (2002) Fisheries productivity in the northeastern Pacific Ocean over the past 2,200 years. Nature 416: 729-733

Fischer et al., 2004 – chapter 3

Fischer H, Fundel F, Ruth U, Twarloh B, Wegener A, Udisti R, Becagli s, Castellano E, Morganti A, Severi M, Wolff EW, Littot GC, Röthlisberger R, Mulvaney R, Hutterli MA, Kaufmann P, Federer U, Lambert F, Bigler M, Hansson M, Jonsell U, De Angelis M, Boutron C, Siggaard-Andersen ML, Steffensen JP, Barbante C, Gaspari V, Gabrielli P, Wagenbach D (*2007*) Reconstruction of millennial changes in dust emission, transport and regional sea ice coverage using the deep EPICA ice cores from the Atlantic and Indian Ocean sector of Antarctica. Earth and Planetary Science Letters 260:340-354

Fischer H. (from AGCS, in press (2008)

Fischer, *AH*, Rott, H and Björnsson, H, 2003, Observation of recent surges of Vatnajökull, Iceland, by means of ERS SAR interferometry, Annals of Glaciology, 37, 69-76.

Fischer-Bruns, I., H. Von Storch, J.F. Gonzalez-Rouco, and E. Zorita, 2005: Modelling the variability of midlatitude storm activity on decadal to century time scales. *Clim. Dyn.*, **25**, 461–476.

Fisher DA, Wake CP, Kreutz KJ, Yalcin K, Steig E, Mayewski PA, Anderson L, Zheng J, Rupper S, Zdanowicz C, Demuth M, Waszkiewicz M, Dahl-Jensen D, Goto-Azuma K, Bourgeois JB, Koerner RM, Sekerka J, Osterberg E, Abbot MB, Finney BP, Burns SJ (2004) Stable isotope records from Mount Logan, Eclipse ice cores and nearby Jellybean Lake. Water cycle of the North Pacific over 2000 years and over five vertical kilometers: sudden shifts and tropical connections. Geographie Physique et Quaternaire 58:9033-9048

Flower BP, Kennett JP (1994) The middle Miocene climatic transition: East Antarctic ice sheet development, deep ocean circulation and global carbon cycling. Palaeogeogr., palaeoclimatol., palaeoecol. 108:537-555

Fogt, R.L., and D.H. Bromwich, 2006: Decadal Variability of the ENSO Teleconnection to the High-Latitude South Pacific Governed by Coupling with the Southern Annular Mode. *J. Climate*, **19**, 979–997.

Foldvik et al. 2004

Forcada J, Trathan P N, Reid K and Murphy E J (2005) The effects of global climate variability in pup production of Antarctic fur seals. Ecology, 86, 2408–2417

Forcada J, Trathan P N, Reid K, Murphy E J and Croxall J P (2006) Contrasting population changes in sympatric penguin species with climate warming. Global Change Biology, 12, 411–423.

Forcada J, Trathan PN, Murphy EJ (submitted) Life history buffering in Antarctic mammals and birds against changing patterns of climate and environmental variation. Global Change Biol

Foreman, C.M., C.F. Wolf, and J.C. Priscu. 2004. Impact of episodic warming events on the physical, chemical and biological relationships of lakes in the McMurdo Dry Valleys, Antarctica. Aquatic Geochemistry 10:239-268.

Foster, T. D. (1987). Mixing and bottom water formation in the shelf break region of the southern Weddell Sea. *Deep-Sea Res.* **34**, 1771–1794.

Foster, T. D. and Carmack, E. C. (1976). Temperature and salinity structure in the Weddell Sea. *J. Phys. Oceanogr.* **6**, 36–44.

Fowbert and Smith 1994 – chapter 3

Fox, A. J. and A. Cziferszky (in press) Unlocking the time-capsule of historic aerial photography to measure volume changes in Antarctic Peninsula glaciers over a fifty-year period. *Photogram. Record.*

Fox, A.J., and A.P.R. Cooper, 1994: Measured properties of the Antarctic Ice Sheet derived from the SCAR Antarctic Digital Database. Pol. Rec., 30(174), pp. 201-206.

Fox, A.J., and D.G. Vaughan, 2005: The retreat of Jones Ice Shelf, Antarctic Peninsula. J. Glaciol., 51(175), pp. 555-560.

FRAM Group (Webb, D.J. et al) 1991. An Eddy-Resolving Model of the Southern Ocean. EOS, Transactions, American Geophysical Union, 72(15), 169-174.

Francis JA. 1994. Improvements to TOVS retrievals over sea ice and applications to estimating Arctic energy fluxes. *Journal of Geophysical Research*, **99**: 10395–10408.

Francis JE, Ashworth A, Cantrill DJ, Crame JA, Howe J, Stephens R, Tosolini A-M, Thorn V (2008) 100 million years of Antarctic climate evolution: evidence from fossil plants. U.S. Geological Survey and The National Academies; USGS OFR-2007.

Francis JE, Poole I (2002) Cretaceous and early Tertiary climates of Antarctica : evidence from fossil wood. Palaeogeography, Palaeoclimatology, Palaeoecology 182:47-64

Fraser KPP, Clarke A, Peck LS (2002) Feast and famine in Antarctica: seasonal physiology in the limpet *Nacella concinna*. Mar Ecol Prog Ser 242: 169–177

Fraser W R and Hofmann E E (2003) A predator's perspective on causal links between climate change, physical forcing and ecosystem response. Marine Ecology Progress Series, 265, 1-15.

Fraser, W.R. and D.L. Patterson, 1997: Human disturbance and long-term changes in Adelie penguin populations: a natural experiment at Palmer Station, Antarctic Peninsula. *Antarctic communities: species, structure and survival*. B. Battaglia, J. Valencia, and D.W.H. Walton, Eds., Cambridge, UK: Cambridge University Press, 445-452.

Frederich M, Pörtner HO (2000) Oxygen limitation of thermal tolerance defined by cardiac and ventilatory performance in the spider crab *Maja squinado*. Amer J Physiol 279: R1531-R1538

Frenot *et al.* 2005, 2007 – from chap 1, chap 3 and chap 4

Frezzotti M, S. Gandolfi, S. Urbini, Snow megadune in Antarctica: Sedimentary structure and genesis, J. Geophys. Res. **107** (D18) (2002) 4344, doi:10.1029/2001JD000673.

Frezzotti, M. *and 13 others*, 2005, Spatial and temporal variability of snow accumulation in East Antarctica from traverse data. *J. Glaciol.,* **51**(172), 113–124.

Frezzotti, M., and M. Polizzi, 2002. 50 years of ice-front changes between the Adélie and Banzare Coasts, East Antarctica. Annals of Glaciology, 34, 235-240.

Fricker H., R. Warner, I. Allison, 2000, Mass balance of the Lambert Glacier–Amery Ice shelf system, East Antarctica: a comparison of computed balance fluxes and measured fluxes, J. Glaciol. **46** (155), 561–570.

Fricker, H. A., N. W. Young, R. Coleman, J. N. Bassis, and J.-B. Minster, 2005. Multi-year monitoring of rift propagation on the Amery Ice Shelf, East Antarcitca. Geophysical Research Letters, 32, L02502, doi:10.1029/2004GL021036.

Fricker, H. A., T. A. Scambos, R. Bindschadler and L. Padman, 2007. An Active Subglacial Water System in West Antarctica Mapped from Space, *Science*, Vol. 315, p. 1544 (doi: 10.1126/science.1136897.

Fricker, HA, Young, NW, Allison, I and Coleman, R (2002). Iceberg calving of the Amery Ice Shelf, East Antarctica: New field and remote sensing data. Annals of Glaciology, 34, 81-88.

Friedlingstein, P., et al. (2006), Climate-carbon cycle feedback analysis: Results from the C4MIP model intercomparison, *Journal of Climate*, *19*, 3337-3353.

Frieβ, U., J. Hollwedel, G. Konig-Langlo, T. Wagner, and U. Platt, Dynamics and chemistry of tropospheric bromine explosion events in the Antarctic coastal region, J. Geophys. Res., 109, (D6): D06305, 2004.

Fulford-Smith SP, Sikes EL (1996) The evolution of Ace Lake, Antarctica, determined from sedimentary diatom assemblages. Palaeogeography Palaeoclimatology Palaeoecology 124:73-86

Fundel F, Fischer H, Weller R, Traufetter F, Oerter H, Miller H (2006) Influence of large-scale teleconnection patterns on methane sulfonate ice core records in Dronning Maud Land. Journal of Geophysical Research 111:D04103

Futuyma D J (1998) Evolutionary Biology. Sinauer, Sunderland, USA

Fyfe and Lorenz, 2005 – chapter 3

Fyfe J. C. and O. A. Saenko, Simulated changes in extratropical Southern Hemisphere winds and currents. *Geophys. Res. Letters*, **33**, L06701, doi:10.1029/2005GL025332 (2006).

Fyfe J.C. (2006) Southern Ocean warming due to human influence. Geophys. Res. Let. 33 L19701, doi:10.1029/2006GL027247

Fyfe J.C. and O. Saenko (2005). Human induced change in the Antarctic Circumpolar Current. J. Clim. 18, 3068-3073.

Fyfe, J. C., G. J. Boer and G. M. Flato (1999). The Arctic and Antarctic oscillations and their projected changes under global warming. *Geophys. Res. Lett*. 26, 1601-1604.

Fyfe, J., and O. Saenko, 2005: Human-induced change in the Antarctic Circumpolar Current. J. Clim., 18, 3068-3073.

Fyfe, J.C., 2003: Extratropical southern hemisphere cyclones: Harbingers of climate change? *J. Climate*, **16**, 2802–2805.

Gambi MC, Bussotti S (1999) Composition, abundance and stratification of soft-bottom macrobenthos from selected areas of the Ross Sea shelf (Antarctica). Polar Biol 21: 347-354

Gaspar, P., Gregoris, Y., and Lefevre, J.-M., 1990: A simple eddy kinetic energy model for stimulations of the oceanic vertical mixing: tests at station Papa and Long-Term Upper Ocean Study Site. Journal of Geophysical Research, 95, 16179-16193.

Gent, P. R, J. Willebrand, T. J. McDougall, and J. C. McWilliams, 1995: Parameterizing eddy-induced tracer transports in ocean circulation models. Journal of Physical Oceanography, 25, 463–474.

Gent, P. R., and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. Journal of Physical Oceanography, 20, 150–155.

Genthon, C., 2004: Space-time Antarctic surface mass-balance variability from climate models, *Ann. Glaciol.*, **39**, 271–275.

Genthon, C., and E. Cosme, 2003: Intermittent signature of ENSO in west-Antarctic precipitation. *Geophys. Res. Lett.*, **30**, 2081, doi:10.1029/2003GL018280.

Genthon, C., and G. Krinner, 1998: Convergence and disposal of energy and moisture on the Antarctic polar cap from ECMWF reanalyses and forecasts. *J. Climate*, **11**, 1703-1716.

Genthon, C., and G. Krinner, 2001: Antarctic surface mass balance and systematic biases in general circulation models. *J. Geophys. Res*. **106**, 20653-20664.

Genthon, C., G. Krinner, and M. Sacchettini, 2003: Interannual Antarctic tropospheric circulation and precipitation variability. *Clim. Dyn.,* **21**, 289-307.

Gerighausen et al. 2003 – chapter 3

Gersonde R, Crosta X, Abelmann A, Armand L (2005) Sea-surface temperature and sea ice distribution of the Southern Ocean at the EPILOG Last Glacial

Maximum-a circum-Antarctic view based on siliceous microfossil records. Quaternary Science Reviews 24:869-896

Gibson JAE, Bayly IAE (2007) New insights into the origins of crustaceans of Antarctic lakes. Antarctic Science 19:157–164

Gibson JAE, Zale R (2006) Holocene development of the fauna of Lake Boeckella, northern Antarctic Peninsula, p 625-634

Gilbert R, Nielsen N, Moller H, Desloges JR, Rasch M (2002) Glacimarine sedimentation in Kangerdluk (Disko Fjord) West Greenland, in response to a surging glacier. Mar Geol 191: 1-18

Giles, K. A.; Laxon, S. W.; Worby, T., Validating Satellite Radar Altimetry Estimates of Antarctic sea ice Thickness Using the ASPeCt Data set. Proceedings of 2006 American Geophysical Union Fall Meeting.

Gili J-M, Arntz WE, Palanques A, Orejas C, Clarke A, Dayton P, Isla E, Teixidó, Rossi S, López-Gonzales PJ (2006) A unique assemblage of epibenthic sessile suspension feeders with archaic features in the high-Antarctic. Deep-Sea Res II 53: 1029-1052

Gilichinsky, D.A., G.S. Wilson (21 other authors). 2007. Microbial populations in Antarctic permafrost: biodiversity, state, age, and implications for astrobiology.Astrobiology 7:275-311.

Gille S.T., M.M. Yale, D.T. Sandwell, 2000: Correlation of mesoscale ocean variability with seafloor roughness from satellite altimetry. Geophys. Res. Lett., 27, 1251-1254.

Gille, S. T., 2002. Warming of the Southern Ocean since the 1950s. Science 295(5558), 1275-1277, doi:10.1126/science.1065863.

Gille, S.T., D.P. Stevens, R.T. Tokmakina, and K.J. Heywood, 2001: Antarctic Circumpolar Current response to zonally averaged winds. J. Geophys. Res., 106(C2), 2743-2760.

Giordano et al. 2007 – Chapter 2

Giovinetto, M. B., and H. J. Zwally, 2000, Spatial distribution of net surface accumulation on the Antarctic ice sheet, *Ann. Glaciol.* **31**, 171–178.

Giovinetto, M.B., and C.R. Bentley, 1985: Surface balance in ice drainage systems of Antarctica. *Ant. J. U. S.*, **20**, 6-13.

Glasser, N., and T. A. Scambos, 2008. A structural glaciological analysis of the 2002 Larsen B ice shelf collapse. Journal of Glaciology 54(184), 3-16.

Gloersen (1995) – chapter 3

Gloersen and White (2001) – chapter 3

Gloersen et al., 1993 – Chapter 3

Gnanadesikan 2006 – chapter 3

Gnanadesikan, A., Griffies, S.M. & Samuels, B.L., 2007: Effects in a climate model of slope tapering in neutral physics schemes. Ocean Modelling, 16, 1-16.

Gnandadesikan, A., and R.W. Hallberg, 2000: On the relationship of the circumpolar current to southern hemisphere winds in coarse-resolution ocean models. J. Phys. Oceanogr., ???, 2013-????.

Goerke H., Weber K, Bornemann H, Ramdohr S, Plötz J (2004) Increasing levels and biomagnification of persistent pollutants (POPs) in Antarctic biota. Mar Pollut Bull 48:295-302

Goldstein, M., H. Engelhardt, B. Kamb and R. M. Frolich, 1993, Satellite radar interferometry for monitoring ice sheet motion: application to an Antarctic ice stream. *Science* **262**: 1525-1530.

Gong and Wang (1999) – chapter 3

Gonzalez-Cabrera JJ, Dowd F, Pedibhotla VK, Rosario R, Stanley-Samuelson D, Petzel D (1995) Enhanced hypo-osmoregulation induced bt warm acclimation in Antarctic fish is mediated by increased gill and kidneyNa^+/K^+-ATPase activities. J Exp Biol 198: 2279-2291

Goodwin *et al*. (2004) – chapter 3

Goodwin ID (1993) Holocene deglaciation, sea-level change, and the emergence of the Windmill Islands, Budd Coast, Antarctica. Quaternary Research 40:70-80

Goodwin ID (1996) A mid to late Holocene re advance of the Law Dome ice margin, Budd Coast, East Antarctica. Antarctic Science 8:395-406

Goosse H. and H. Renssen, 2001. A two-phase response of Southern Ocean to an increase in greenhouse gas concentrations. Geophysical Research Letters 28, 3469-3473

Gordon et al. 2007

Gordon, A.L. (1982) Weddell deep water variability, Journal of Marine Research, 40, 199-217.

Gordon, A.L., D.G. Martinson and H.W. Taylor (1981) The wind-driven circulation in the Weddell-Enderby Basin, Deep-Sea Res., 28, 151-163.

Gordon, C., C. Cooper, C. A. Senior, H. Banks, J. M. Gregory, T. C. Johns, J. F. B. Mitchell, and R. A. Wood (2000), The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments, Clim. Dyn., 16, 147-168.

Gordon, J., Haynes, V. and Hubbard, A. (2008), Recent glacier changes and climate trends on South Georgia, *Global and Planetary Change, 60.*

Gordon, A. L., Zambianchi, E., Orsi, A., Visbeck, M., Giulivi, C. F., Whitworth III, T. and Spezie, G (2004). Energetic plumes over the western Ross Sea continental slope. *Geophys. Res. Lett*. **31**, doi: 10.1029/2004GL020785.

Gore DB (1997) Blanketing snow and ice; constraints on radiocarbon dating deglaciation in East Antarctic oases. Antarctic Science 9:336-346

Gore DB, Rhodes EJ, Augustinus PC, Leishman MR, Colhoun EA, Rees-Jones J (2001) Bunger Hills, East Antarctica: Ice free at the Last Glacial Maximum. Geology 29:1103-1106

Goring and Pyne, 2003 - chapter 4

Gould, J., et al. (2004), Argo Profiling Floats Bring New Era of In Situ Ocean Observations, *EOS*, *85*.

Gouretski and Danilov 1993

Grannas, A. M., A. E. Jones, J. Dibb, M. Ammann, C. Anastasio, H. J. Beine, M. Bergin, J. Bottenheim, C. S. Boxe, G. Carver, G. Chen, J. H. Crawford, F. Dominé, M. M. Frey, M. I. Guzmán, D. E. Heard, D. Helmig, M. R. Hoffmann, R. E. Honrath, L. G. Huey, M. Hutterli, H. W. Jacobi, P. Klán, B. Lefer, J. McConnell, J. Plane, R. Sander, J. Savarino, P. B. Shepson, W. R. Simpson, J. R. Sodeau, R. von Glasow, R. Weller, E. W. Wolff, and T. Zhu, An overview of snow photochemistry: evidence, mechanisms and impacts, Atmos. Chem. Phys., 7, 4329-4373, 2007

Gray JS (2001) Antarctic marine benthic diversity in a world-wide context. Polar Biol 24: 473-485

Gray, L., I. Joughin, S. Tulazcyk, V.B. Spikes, R. Bindschadler and K. Jezek, 2005. Evidence for subglacial water transport in the West Antarctica Ice Sheet through three-dimensional satellite radar interferometry, *Geophysical Research Letters*, VOL. 32, L03501, doi:10.1029/2004GL021387,.

Gray, L.; Short, N.; Bindschadler, R.; Joughin, I.; Padman, L.; Vornberger, P.; Khananian, A. Radarsat interferometry for Antarctic grounding-zone mapping. Annals of Glaciology, 34, 269-276.

Greenslade (in press) – chap 1

Greenslade 2006 – chapter 3

Greg and Conkright – chapter 3, page 176

Gregory J.M., 2000. Vertical heat transport in the ocean and their effect on time-dependent climate change. Clim. Dyn. 16, 501-515.

Gregory, J. and P. Huybrechts, 2006: Ice-sheet contributions to future sea-level change. *Philosophical Transactions of the Royal Society A,* 1709-1731

Gressitt 1970 – chap 1

Grezio, A., N.C. Wells, V.O. Ivchenko, and B.A. de Cuevas, 2005: Dynamical budgets of the Antarctic Circumpolar Current using ocean general-circulation models. Quart. J. Roy. Met. Soc., 131(607), 833-860.

Grieshaber MK, Hardewig I, Kreutzer U, Pörtner HO (1994) Physiological and metabolic responses to hypoxia in invertebrates. Rev Physiol Biochem Pharmacol 125: 43-147

Griffies, 2005 – chapter 3

Griffies, S. M., 1998: The Gent–McWilliams Skew Flux. Journal of Physical Oceanography, 28, 831-841.

Griffies, S. M., 2006: Some ocean model fundamentals. Pp 19-74 in: Ocean Weather Forecasting: an integrated view of Oceanography. E. P. Chassignet & J. Verron, eds., Berlin, Germany: Springer, 574 pp.

Griffies, S.M., Boning, C., Bryan, F.O., Chassignet, E.P., Gerdes, R., Hasumi, H., Hurst, A., Treguier, A-M, & Webb, D.J., 2000: Developments in ocean climate modelling. Ocean Modelling, 2, 123-192.

Griffies, S.M., Gnanadesikan, A., Dixon, K.W., Dunne, J.P., Gerdes, R., Harrison, M.J., Rosati, A., Russell, J.L., Samuels, B. L., Spelman, M.J., Winton, M. & Zhang, R., 2005: Formulation of an ocean model for global climate simulations. Ocean Science, 1, 45-79.

Griffies, S.M., Harrison, M.J., Pacanowski, R.C., and Rosati, A., 2004: A technical guide to MOM4. GFDL ocean group technical report No. 5. NOAA/Geophysical Fluid Dynamics Laboratory, Princeton, USA.

Grobe et al. 1997 – chapter 3

Grootes PM, Stuiver M (1997) Oxygen 18/16 variability in Greenland snow and ice with 10^3 to 10^5-year time resolution. Journal of Geophysical Research 102:26455-26470

Grosfeld, K. and R. Gerdes, 1998: Circulation beneath the Filchner Ice Shelf, Antarctica, and its sensitivity to changes in the oceanic environment: a case-study. *Ann. Glaciol*. **27**, 99-104.

Grosfeld, K., M. Schröder, E. Fahrbach, R. Gerdes and A Mackensen, 2001: How iceberg calving and grounding change the circulation and hydrography in the Filchner Ice Shelf-ocean system. *J. Geophys. Res*. **106**(C5), 9039-9055.

Grosfeld, K., R. Gerdes and J. Determann, 1997: Thermohaline circulation and interaction between ice shelf cavities and the adjacent open ocean. *J. Geophys. Res*. **102**(C7), 15,595-15,610.

Gudmundsson, G. H., 2006, Fortnightly variations in the flow velocity of Rutford Ice Stream, West Antarctica, *Nature* **444**, 1063-1064 doi:10.1038/nature05430.

Guglielmin, M. (2007, in prep.) Permafrost cooling in Continental Antarctica.

Guglielmin, M. 2004. Observations on permafrost ground thermal regimes from Antarctica and the Italian Alps, and their relevance to global climate change. *Global and Planetary Change* 40: 159-167.

Guglielmin, M. 2006. Ground surface temperature (GST), active layer, and permafrost monitoring in continental Antarctica. Permafrost & Periglacial Processes:17:133-143.

Guglielmin, M. and H.M. French. 2004. Ground ice in the Northern Foothills, northern Victoria Land, Antarctica. Annals of Glaciology 39:495-500.

Guglielmin, M., Ellis Evans, C.J., Cannone, N. 2007 Active layer thermal regime under different vegetation conditions in permafrost areas. A case study at Signy Island (Maritime Antarctica). Geoderma (accepted paper)

Guglielmin, M., M. Balks, and R. Paetzold. 2003. Towards an Antarctic active layer and permafrost monitoring network. In: Phillips, M., S.M. Springman, and L. Arenson (eds.) Proceedings of the 8th International Conference on Permaforst, Zurich, Switzerland, 21-25 July 2003. Balkema Publishers, Lisse, pp. 337-341.

Guo, Z., D.H. Bromwich, and J.J. Cassano, 2003: Evaluation of Polar MM5 simulations of Antarctic atmospheric circulation. *Mon. Wea. Rev.,* **131**, 384-411.

Guo, Z., D.H. Bromwich, and K.M. Hines, 2004: Modeled Antarctic precipitation. Part II: ENSO signal over West Antarctica. *J. Climate* **17**, 448-465.

Gutt J (2000) Some "driving forces" structuring communities of the sublittoral Antarctic macrobenthos. Antarctic Sci 12 (3): 297-313

Gutt J (2007) Antarctic macro-zpoobenthic communities: a review and an ecological classification. Antarctic Sci 109 (2): 165-182

Gutt J, Koubbi P, Eléaume M (2007) Mega-epibenthic diversity off Terre Adélie (Antarctica) in relation to disturbance. Polar Biol 30: 1323-1329

Gutt J, Piepenburg D (1991) Dense aggregations of three deep-sea holothurians in the southern Weddell Sea, Antarctica. Mar Ecol Prog Ser 68: 277-285

Gutt J, Piepenburg D (2003) Scale-dependent impact on diversity of Antarctic benthos caused by grounding of icebergs. Mar Ecol Prog Ser 253: 77-83

Haidvogel, 1991 – chapter 3 – Haidvogel et al 1991?

Haidvogel, D.B., Wilkin, J.L. & Young, R.E., 1991: A semi-spectral primitive equation ocean circulation model using vertical sigma and orthogonal curvilinear horizontal coordinates. Journal of Computational Physics, 94, 151–185.

Hales and Takahashi (2004) – Chapter 3 – only 2007 in list

Hales, B., and T. Takahashi, High-resolution biogeochemical investigation of the

Ross Sea, Antarctica, during the AESOPS (U.S. JGOFS) Program. *Global Biogeochemical Cycles* 18, GB3006, doi:10.1029/2003GB002165, 2007

Hall A, Visbeck M (2002) Synchronous variability in the Southern Hemisphere atmosphere, sea ice, and ocean resulting from the annular mode, J. Clim., 15, 3043-3057.

Hall BL, Baroni C, Denton GH (2003) Holocene relative sea-level history of the southern Victoria Land coast, Antarctica:. Global and Planetary Change

Hall BL, Denton GH (1995) Late Quaternary lake levels in the Dry Valleys, Antarctica. Antarctic Journal of the United States 30:52-53

Hall BL, Denton GH (2000) Extent and Chronology of the Ross Sea Ice Sheet and the Wilson Piedmont Glacier Along the Scott Coast at and Since the Last Glacial Maximum. Geografiska Annaler Series a-Physical Geography 82A:337-363

Hall BL, Denton GH (2002) Holocene history of the Wilson Piedmont Glacier along the southern Scott Coast, Antarctica, p 619-627

Hall BL, Denton GH, Hendy CH (2000) Evidence from Taylor Valley for a Grounded Ice Sheet in the Ross Sea, Antarctica, p 275-303

Hall BL, Denton GH, Overturf B (2001) Glacial Lake Wright, a high-level Antarctic lake during the LGM and early Holocene. Antarctic Science 13:53-60

Hall BL, Hoelzel AR, Baroni C, Denton GH, Le Boeuf BJ, Overturf B, Topf AL (2006) Holocene Elephant Seal Distribution Implies Warmer-Than-Present Climate in the Ross Sea. Proceedings of the National Academy of Sciences of the United States of America 103: 10213-10217

Hallberg, R. and A. Gnanadesikan, A., The role of eddies in determining the structure and response of the wind-driven Southern Hemisphere overturning: Initial results from the Modelling Eddies in the Southern Ocean project. *J. Phys. Oceanogr.*, **36**, 3312-3330 (2006).

Hallberg, R.W., 1997: Stable split time stepping schemes for large-scale ocean modelling. Journal of Computational Physics 135, 54-6

Hambrey MJ, Glasser NF, Mckelvey BC, Sugden DE, Fink D (2007) Cenozoic Landscape Evolution of an East Antarctic Oasis (Radok Lake Area, Northern Prince Charles Mountains), and Its Implications for the Glacial and Climatic History of Antarctica. Quaternary Science Reviews 26:598-626

Hamilton G.S., 2005, Spatial patterns in mass balance of the Siple Coast and Amundsen Sea sectors, West Antarctica, *Ann. Glaciol.,* **41**, 105-106.

Hammer CU, Clausen HB, Langway CCJ (1994) Electrical conductivity method (ECM) stratigraphic dating of the Byrd Station ice core, Antarctica. Annals of Glaciology 20:115-120

Hansen J, Sato M, Ruedy R, Lo K, Lea DW, Medina-Elizade M (2006) Global temperature change. Proc Natl Acad Sci 103: 14288-14293

Hansen, J.E. (2005), A slippery slope: how much global warming constitutes "dangerous anthropogenic interference?" *Climatic Change 68*, 269-279.

Hansom JD, Flint CP (2004) Holocene ice fluctuations on Brabant Island, Antarctic Peninsula. Cambridge Journals Online, p 165-166

Hansom, 1981 - chapter 4

Hardy 1967 – chapter 3

Harris PT, O'Brien PE (1996) Geomorphology and sedimentology of the continental shelf adjacent to Mac. Robertson Land, East Antarctica: a scalped shelf. Geo-Marine Lett 16: 287-296

Hart 1934 – chapter 3, page 176

Hartmann and Lo, 1998 – chapter 3

Hartmann, D.L., J.M. Wallace, V. Limpasuvan, D.W.J. Thompson, and J.R. Holton, 2000: Can ozone depletion and global warming interact to produce rapid climate change? *Proc. Natl. Acad. Sci. U.S.A.*, **97**, 1412–1417.

Harwood DM, Mcminn A, Quilty PG (2000) Diatom biostratigraphy and age of the Pliocene Sorsdal Formation, Vestfold Hills, East Antarctica. Antarctic Science 12:443-462

Hasselmann K, Latif M, Hooss G, Azar C, Edenhofer O, Jaeger CC, Johannessen OM, Kemfert C Welp M, Wokaun A (2003) The challenge of long-term climate change. Science 302: 1923-1925

Hauck, C., G. Vieira, S. Gruber, J.J. Blanco, and M. Ramos. 2007. Geophysical identification of permafrost in Livingston island, maritime Antarctica. J. Geophys. Res., doi:10.1029/2006JF0005444, in press.

Haug GH, Hughen KA, Sigman DM, Peterson LC, Röhl U (2001) Southward migration of the Intertropical Convergence Zone through the Holocene. Science 293:1304-1308

Hawley, R. L., E. M. Morris, R. Cullen, U. Nixdorf, A. P. Shepherd, and D. J. Wingham, 2006, ASIRAS airborne radar resolves internal annual layers in the dry-snow zone of Greenland, *Geophys. Res. Lett.*, **33**, L04502, doi:10.1029/2005GL025147.

Haywood AM, Valdes PJ, Sellwood BW (2000) Gloal scale paleoclimate reconstruction of the middle Pliocene climare using the UKMO GCM: initial results. Global Planet. Change 25:239-256

Heinze, C. (2004), Simulating oceanic CaCO3 export production in the greenhouse, *Geophys. Res. Let.*, *31*.

Hellmer H H, Jacobs S S, Jenkins A (1998) Oceanic erosion of a floating Antarctic glacier in the Amundsen Sea, in: Ocean, Ice, and Atmosphere: Interactions at the Antarctic continental margin, S S Jacobs & R F Weiss (eds), Antarc. Res. Ser., 75, 83-99.

Hellmer H H, Kauker F, Timmermann R (2008) Western impacts of Weddell Sea anomalies, Science, submitted.

Hellmer, H. H. (2004). Impact of Antarctic shelf basal melting on sea ice and deep ocean properties. *Geophys. Res. Lett.* **31**, doi:10.1029/2004GL019506.

Hellmer, H.H. and D.J. Olbers, 1989: A two-dimensional model for the thermohaline circulation under an ice shelf. *Antarct. Sci.* **1**(4), 325-336.

Hellmer, H.H. and S.S. Jacobs (1994) Temporal changes in shelf water of the southern Ross Sea. *Antarct. J. of the U.S.* **29**(5), 123-124.

Hellmer, H.H., S.S. Jacobs and A. Jenkins, 1998: Oceanic erosion of a floating Antarctic glacier in the Amundsen Sea, *Antarct. Res. Ser*. **75**, 83-99.

Helmig, D., Oltmans, S., Carlson, D., Lamarque, J.-F., Jones, A.E., Labuschagne, C., Anlauf, K. and Hayden, K., A review of surface ozone in the polar regions", Atmos. Env., 41, 5138-5161, doi:10.1016/j.atmosenv.2006.09.053 (2007)

Hendy CH (2000) Late Quaternary lakes in the McMurdo Sound region of Antarctica. Geografiska Annaler 82 A:411-432

Hennion et al. 2006 RiSCC – chapter 3 & 4

Hense I, Timmermann R, Beckmann A, Bathmann UV (2003) Regional and interannual variability of ecosystem dynamics in the Southern Ocean. Ocean Dynamics 53: 1-10

Heroy DC, Anderson JB (2005) Ice-sheet extent of the Antarctic Peninsula region during the Last Glacial maximum (LGM) - Insights from glacial geomorphology. GSA Bulletin 117:1497-1512

Heroy DC, Anderson JB (in press) Ice Sheet Configuration and Paleo-Ice streams of the Antarctic Peninsula during the Last Glacial Maximum (LGM).

Herterich, K. 1988. A three-dimensional model of the Antarctic ice sheet. Annals of Glaciology 11:32-35.

Heywood et al. 1999

Heywood, K. J., Naveira Garabato, A. C. and Stevens, D. P. (2002). High mixing rates in the abyssal Southern Ocean. *Nature* **415**, 1011-1014.

Heywood, K.J. and B.A. King (2002) Water masses and baroclinic transports in the South Atlantic and Southern Oceans, J. Mar. Res., 60(5), 639-676.

Hibler III, W.D., 1979: A dynamic-thermodynamic sea ice model. Journal of Physical Oceanography, 9, 815–846.

Hiller A, Wand U, Kampf H, Stackebrandt W (1988) Occupation of the Antarctic continent by petrels during the past 35 000 years: inferences from a ^{14}C study of stomach oil deposits. Polar Biology 9:69-77

Hines K. M., D. H. Bromwich, and G. J. Marshall, 2000: Artificial surface pressure trends in the NCEP–NCAR reanalysis over the Southern Ocean and Antarctica. J. Climate, 13, 3940–3952.

Hines, K.M., D.H. Bromwich, P.J. Rasch, and M.J. Iacono, 2004: Antarctic Clouds and Radiation within the NCAR Climate Models. *J. Climate*, **17**, 1198–1212.

Hinke J T, Salwicka K, Trivelpiece S G, Watters G M, Trivelpiece W Z (2007) Divergent responses of *Pygoscelis* penguins reveal a common environmental driver. Oecologia, 153, 845-855.

Hirst et al. 1999 - chapter 4

Hirst, A.C. (1999). The Southern Ocean response to a global warming in the CSIRO coupled ocean-atmosphere model. Environmental Modeling and Software, 227-241.

Hjört C, Björck S, Ingólfsson O, Möller P (1998) Holocene deglaciation and climate history of the northern Antarctic Peninsula region: a discussion of correlations between the Southern and Northern Hemispheres. Annals of Glaciology 27:110-112

Hjört C, Ingólfsson O, Bentley MJ, Björck S (2003) The Late Pleistocene and Holocene glacial and climate history of the Antarctic Peninsula region as documented by the land and lake sediment records - a review. Antarctic Research Series 79:95-102

Hjört C, Ingólfsson O, Möller P, Lirio JM (1997) Holocene glacial history and sea-level changes on James Ross Island, Antarctic Peninsula. Journal of Quaternary Science 12:259-271

Hodell DA, Kanfoush SL, Shemesh A, Crosta X, Charles CD, Guilderson TP (2001) Abrupt cooling of Antarctic surface waters and sea ice expansion in the South Atlantic sector of the Southern Ocean at 5000 cal yr B.P. Quaternary Research 56:191-198

Hodgson DA, Bentley MJ, Roberts SJ, Smith JA, Sugden DE, Domack EW (2006a) Examining Holocene stability of Antarctic Peninsula Ice Shelves. Eos Transactions, American Geophysical Union 87:305-312

Hodgson DA, Convey P (2005) A 7000-year record of oribatid mite communities on a maritime-Antarctic island: responses to climate change. Arctic Antarctic and Alpine Research 37:239-245

Hodgson DA, Doran PT, Roberts D, McMinn A (2004) Paleolimnological studies from the Antarctic and subantarctic islands. In: Long-term Environmental Change in Arctic and Antarctic Lakes. Developments in Paleoenvironmental Research, Vol. 8, pp 419-474 (Pienitz R, Douglas MSV, Smol JP, Eds), Springer, Dordrecht

Hodgson DA, McMinn A, Kirkup H, Cremer H, Gore D, Melles M, Roberts D, Montiel P (2003) Colonization, succession, and extinction of marine floras during a glacial cycle: A case study from the Windmill Islands (east Antarctica) using biomarkers. Paleoceanography 18(3):1067, doi:1010.1029/2002PA000775

Hodgson DA, Noon PE, Vyverman W, Bryant CL, Gore DB, Appleby P, Gilmour M, Verleyen E, Sabbe K, Jones VJ, Ellis-Evans JC, Wood PB (2001) Were the Larsemann Hills ice-free through the Last Glacial Maximum? Antarctic Science 13:440-454

Hodgson DA, Roberts D, McMinn A, Verleyen E, Terry B, Corbett C, Vyverman W (2006b) Recent rapid salinity rise in three East Antarctic lakes. Journal of Paleolimnology 36:385-406

Hodgson DA, Verleyen E, Sabbe K, Squier AH, Keely BJ, Leng MJ, Saunders KM, Vyverman W (2005) Late Quaternary climate-driven environmental change in the Larsemann Hills, East Antarctica, multi-proxy evidence from a lake sediment core. Quaternary Research 64:83-99

Hodgson DA, Verleyen E, Squier AH, Sabbe K, Keely BJ, Saunders KM, Vyverman W (2006c) Interglacial environments of coastal east Antarctica: comparison of MIS 1 (Holocene) and MIS 5e (Last Interglacial) lake-sediment records. Quaternary Science Reviews 25:179-197

Hodgson, D. A., and Smol, J. P. (2008). High latitude palaeolimnology. *In* "Polar Lakes and Rivers - Arctic and Antarctic Aquatic Ecosystems." (W. F. Vincent, and J. Laybourn-Parry, Eds.). Oxford University Press, Oxford, UK.

Hodgson, D. A., Roberts, D., McMinn, A., Verleyen, E., Terry, B., Corbett, C., and Vyverman , W. (2006). Recent rapid salinity rise in three East Antarctic lakes. *Journal of Paleolimnology* **36,** 385-406.

Hofman GE, Buckley BA, AIraksinen S, Keen JE, Somero GN (2000) Heat-shock protein expression is absent in the Antarctic fish *Trematomus bernacchii* (family Nototheniidae). J Exp Biol 203: 2331–2339

Hofmann E E & Klinck J M (1998) Thermohaline variability of the waters overlying the West Antarctic Peninsula continental shelf, in: Ocean, Ice, and Atmosphere: Interactions at the Antarctic continental margin, S S Jacobs & R F Weiss (eds), Antarc. Res. Ser., 75, 67-81.

Hofmann GE, Lund SG, Place SP, Whitmer AC (2005) Some like it hot, some like it cold: the heat shock response is found in New Zealand but not Antarctic notothenioid fishes. J Exp Mar Biol Ecol 316: 79-89

Hofmann, D.J., S.J. Oltmans, J.M. Harris, B.J. Johnson, J.A. Lathrop, "Ten years of ozonesonde measurements at the south pole: Implications for recovery of springtime Antarctic ozone", *J. Geophys. Res.*, **102**, 8931-8943, (1997).

Hogg ID, Cary SC, Convey P, Newsham K, O'Donnell T, Adams BJ, Aislabie J, Frati FF, Stevens MI, Wall DH (2006) Biotic interactions in Antarctic terrestrial ecosystems: are they a factor? Soil Biology and Biochemistry 38:3035-3040

Hogg, A. McC., M. P. Meredith, J. R. Blundell and C. Wilson, 2008. Eddy Heat Flux in the Southern Ocean: Response to Variable Wind Forcing. Journal of Climate, 4, 608-620.

Holgate, S.J. 2007. On the decadal rates of sea level change during the twentieth century. Geophysical Research Letters, 34, L01602, 10.1029/2006GL028492.

Holgate, S.J. and Woodworth, P.L. 2004. Evidence for enhanced coastal sea level rise during the 1990s. Geophysical Research Letters, 31, L07305, doi:10.1029/2004GL019626.

Holland, D.M. and A. Jenkins, 2001: Adaptation of an isopycnic coordinate ocean model for the study of circulation beneath ice shelves. *Mon. Wea. Rev*. **129**(8), 1905-1927.

Holland, D.M., Jacobs, S.S. & Jenkins, A., 2003: Modelling the ocean circulation beneath the Ross Ice Shelf. Antarctic Science, 15 (1), 13–23 (doi: 10.1017/S0954102003001019).

Holland, P.R. and D.L. Feltham, 2006: The effects of rotation and ice shelf topography on frazil-laden Ice Shelf Water plumes. *J. Phys. Oceanogr.*, **36**(12), 2312-2327, doi: 10.1175/JPO2970.1.

Holland, P.R., A. Jenkins and D.M. Holland, 2007: The response of ice shelf basal melting to variation in ocean temperature, *J. Clim.*, in press.

Holm-Hansen, M. Kahru, C.D. Hewes, S. Kawaguchi, T. Kameda, V.A. Sushin, I. Krasovski, J. Priddle, R. Korb, R.P. Hewitt and B.G. Mitchell, Temporal and spatial distribution of chlorophyll-*a* in surface waters of the Scotia Sea as determined by both shipboard measurements and satellite data, *Deep-Sea Research Part II—Topical Studies in Oceanography* **51** (2004) (12–13), pp. 1323–1331.

Holt, J. W., D. D. Blankenship, D. L. Morse, D. A. Young, M. E. Peters, S. D. Kempf, T. G. Richter, D. G. Vaughan, and H. F. J. Corr, 2006, New boundary conditions for the West Antarctic Ice Sheet: Subglacial topography of the Thwaites and Smith glacier catchments, *Geophys. Res. Lett.*, **33**, L09502, doi:10.1029/2005GL025561.

Honrath, R.E., Lu, Y., Peterson, M.C., Dibb, J.E., Arsenault, M.A., Cullen, N.J., and Steffen, K.: Vertical fluxes of NO_x, HONO and HNO_3 above the snowpack at Summit, Greenland, Atmos. Environ., 36, 2629-2640, 2002.

Honrath, R.E., Peterson, M.C., Guo, S., Dibb, J.E., Shepson, P.B., and Campbell, B.: Evidence of NO_x production within or upon ice particles in the Greenland snowpack, Geophys. Res. Lett., 26, 695-698, 1999.

Hormes A, Müller BJ, Schlüchter C (2001) The Alps with little ice: evidence for eight Holocene phases of reduced glacier extent in central Swiss Alps. Holocene 11:255-265

Hoskins and Hodge, 2005 – chapter 3 – Hoskins and Hodges?

Hoskins and Hodges, 2005 – chapter 3

Houghton JT (2001) Climate Change 2001: The Scientific Basis. Intergovernmental Panel of Climate Change, 3rd Assessment Report, Vol. Cambridge University Press

Houghton JT, Meir A, Filho LG, Callander BA, Harris N, Kattenberg A, Markell KJ (1996) Climate Change 1995: The Science of Climate Change. Intergovernmental Panel of Climate Change Report, Vol. Cambridge University Press

Huck, P.E., S. Tilmes, G.E. Bodeker, W.J. Randel, A.J. McDonald, H. Nakajima, "An improved measure of ozone depletion in the Antarctic stratosphere", *J. Geophys. Res.* **112**, D11104 (2007).

Hughen KA, Baillie MGL, Bard E, Beck JW, Bertrand CJH, Blackwell PG, Buck CE, Burr GS, Cutler KB, Damon PE, Edwards RL, Fairbanks RG, Friedrich M, Guilderson TP, Kromer B, Mccormac G, Manning S, Ramsey CB, Reimer PJ, Reimer RW, Remmele S, Southon JR, Stuiver M, Talamo S, Taylor FW, Van Der Plicht J, CE W (2004) Marine04 Marine Radiocarbon Age Calibration 0-26 Cal Kyr Bp. Radiocarbon 46:1059-1108

Hughes et al., 2001 – Chapter 3 - Hughes & Ash 2001?

Hughes L (2000) Biological consequences of global warming: is the signal already apparent? Trends Ecol Evol 15: 56-61

Hughes, C.W. and E.R. Ash, 2001: Eddy forcing of the mean flow in the Southern Ocean. Journal of Geophysical Research. 106, 2713-2722.

Hughes, C.W., Meredith, M.P. and Heywood, K. 1999. Wind-driven transport fluctuations through Drake Passage: a Southern mode. Journal of Physical Oceanography, 29, 1971 – 1992.

Hughes, C.W., Woodworth, P.L., Meredith, M.P., Stepanov, V., Whitworth, T. and Pyne A.R. 2003. Coherence of Antarctic sea levels, Southern Hemisphere Annular Mode, and flow through Drake Passage. Geophysical Research Letters, 30(9), 1464, doi:10.1029/2003GL017240.

Hughes, G. L., S. S. Rao, and T. S. Rao, 2007: Statistical analysis and time-series models for minimum/maximum termperatures in the Antarctic Peninsula. *Proceedings of the Royal Society A*, **463** 241-259.

Hughes, G. O. and Griffiths, R. W. (2006). A simple convective model of the global overturning circulation, including effects of entrainment into sinking regions. *Ocean Model.* **12**, 46-79.

Hughes, T., 1981. The weak underbelly of the West Antarctic ice sheet. Journal of Glaciology 27(97), 518-525.

Hughes, T.J., 1973, Is the West Antarctic ice sheet disintegrating? *J. Geophys. Res.* **78**(33), 7884-7910.

Huiskes et al. 2006 – Chapter 3

Hulbe, C.L. and M.A. Fahnestock, 2007, Century-scale discharge stagnation and reactivation of the Ross ice streams, West Antarctica, *J. Geophys. Res. Earth Surface*, **112**, doi:10.1029/2006JF000603.

Hunke, E.C. & Dukowicz, J.K., 1997: An Elastic–Viscous–Plastic Model for Sea Ice Dynamics. Journal of Physical Oceanography, 27 (9), 1849-1867.

Hunt B P V, Pakhomov E A and Mcquaid C D (2001) Community structure of mesozooplankton in the Antarctic Polar Frontal Zone in the vicinity of the Prince Edward Islands (Southern Ocean): Small-scale distribution patterns in relation to physical parameters. Deep-Sea Research Part II Topical Studies Oceanography, 49, 3307–3325.

Hunt, B.P.V, and G W. Hosie The Continuous Plankton Recorder in the Southern Ocean: a comparative analysis of zooplankton communities sampled by the CPR and vertical net hauls along 140 E. *Journal of Plankton Research*, Volume 25,12 , pp 1561–1579, 2003

Hunter, J., Coleman, R. and Pugh, D. 2003 The sea level at Port Arthur, Tasmania, from 1841 to the present. Geophysical Research Letters, 30, 1401, doi:10.1029/2002GL016813.

Hurrell, J.W. and H. van Loon, 1994: A Modulation of the Atmospheric Annual Cycle in the Southern Hemisphere. *Tellus:* Vol. 46A, pp.325-338.

Huston M (1979) A general hypothesis of species diversity. Amer Naturalist 113 (1): 81-101

Hutter, K. (1983), *Theoretical Glaciology.* Dordrecht: Reidel Publ. Co., 510 pp.

Huybrechts P (2002) Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles. Quaternary Science Reviews 21:203-231

Huybrechts, P. (2004), Antarctica: modeling. In *Mass balance of the Cryosphere: Observations and modeling of contemporary and future changes (J.L. Bamber and A.J. Payne, eds.).* Cambridge: Cambridge University Press, 491-523.

Huybrechts, P., J. Gregory, I. Janssens, and M. Wild, 2004: Modelling Antarctic and Greenland volume changes during the 20th and 21st centuries forced by GCM time slice integrations, *Glob. Planet. Change*, **42**, 83–105.

Igarashi A, Harada N, Moriwaki Kv, p. (1995) Marine fossils of 30-40 ka in raised beach deposits, and late Pleistocene glacial history around Lutzow-Holm Bay, East Antarctica. Proceedings of the NIPR Symposium on Antarctic Geosciences 8:219-229

IGY, 1957-8 - Chapter 4, Page 72

IHB, 2002. Status of natutical charting (part A) and hydrographic surveying (part B) in Antarctica. International Hydrographic Bureau publication S-59. (see also previous editions of this publication with the same number.)

Impacts of a warming Arctic: Arctic Climate Impact Assessment. (2005) Cambridge University Press, Cambridge, UK, 1042 pp

Indermühle A, Monnin E, Stauffer B, Stocker TF, Wahlen M (2000) Atmospheric CO2 concentration from 60 to 20 kyr BP from the Taylor dome ice core, Antarctica. Geophysical Research Letters 27:735-738

Indermuhle A, Stocker TF, Joos F, Fischer H, Smith HJ, Wahlen M, Deck B, Mastroianni D, Tschumi J, Blunier T, Meyer R, Stauffer B (1999) Holocene carbon-cycle dynamics based on CO2 trapped in ice at Taylor Dome, Antarctica. Nature 398:121-126

Ingólfsson Ó (2004) Quaternary glacial and climate history of Antarctica. In: Ehlers J, Gibbard PL (eds) Quaternary Glaciations - Extent and Chronology, Part III. Elsevier, p 3-43

Ingolfsson O, Hjort C (2002) Glacial history of the Antarctic Peninsula since the Last Glacial Maximum-a synthesis, p 227-234

Ingólfsson Ó, Hjort C, Berkman PA, Björck S, Colhoun EA, Goodwin ID, Hall BL, Hirakawa K, Melles M, Möller P, Prentice ML (1998) Antarctic glacial history since the last glacial maximum: an overview of the record on land. Antarctic Science 10:326-344

Ingólfsson Ó, Hjort C, Humlum O (2003) Glacial and Climate History of the Antarctic Peninsula since the Last Glacial Maximum. Arctic, Antarctic and Alpine Research 35:175-186

Inoue, H. Y., and M. Ishii (2005), Variations and trends of CO2 in the surface seawater in the Southern Ocean south of Australia between 1969 and 2002, *Tellus B*, *57*, 58-69.

Intergovernmental Panel on Climate Change. 2001. Climate change 2001: impacts, adaption and vulnerability. Ch. 16 Polar regions (Arctic and Antarctic). United Nations Environmental Programme and World Health Association (http://www.grida.no/climate/ipcc_tar/wg2/index.htm).

Intergovernmental Panel on Climate change. 2007. IPCC Fourth Assessment Report: the Polar Perspective (http://www.sciencepoles.org/index.php?articles/ipcc_fourth_assessment_report_the polar_perspective&s=2&rs=home&uid=897&lg=en)

IOC, 2006. Manual on sea-level measurement and interpretation. Volume 4 - An update to 2006. (eds. Aarup, T., Merrifield, M., Perez, B., Vassie, I and Woodworth, P.). Intergovernmental Oceanographic Commission Manuals and Guides No. 14. IOC, Paris, 80pp.

IOC. 1997. Global Sea Level Observing System (GLOSS) implementation plan-1997. Intergovernmental Oceanographic Commission, Technical Series, No. 50, 91pp. & Annexes.

IPCC (2007) Climate Change 2007: The Physical Science Basis. Contribution of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge

IPCC 2000 – Chapter 2

IPY 2007-2008 - Chapter 4 - Page 72

Iskandarani, M., Haidvogel, D.B., Levin, J.C., 2003: A three-dimensional spectral element model for the solution of the hydrostatic primitive equations. Journal of Computational Physics 186, 397–425

Isla E, Rossi S, Palanques A, Gili JM, Gerdes D, Arntz WE (2006) Biochemical composition of marine sediments from the eastern Weddell Sea (Antarctica): High nutritive value in a high benthic-biomass environment. J Mar Sist 60: 255-267

Ivchenko, V., N. Wells, and A. Grezio, 2004: What controls the transport of the Antarctic Circumpolar Current in ocean general circulation models? Geophys. Res. Abs, 6, 01847.

Ivchenko, V.O., V. B. Zalesny, and M. R. Drinkwater, 2004: Can the Equatorial Ocean quickly responds to Antarctic sea ice-salinity anomalies? Geophys. Res. Lett., 31(15), doi: 10.1029/2004GL020472.

Ivins ER, James TS (2005) Antarctic glacial isostatic adjustment: a new assessment. Antarctic Science 17:541-553

Jabaud-Jan, A., et al. (2004), Interannual varaibility of the carbon dioxide system in the southern Indian Ocean (20˚-60˚S): the impact of a warm anomaly in the austral summer 1998, *Global Biogeochemical Cycles*, *18*, GB1042,1010.1029/2002GB002017.

Jacka, T. H. and Budd, W. F. 1991. Detection of temperature and sea ice extent changes in the Antarctic and Southern Ocean. Weller, G., Wilson, C. L., and Severin, B. A. Proceedings of the International Conference on the Role of the Polar Regions in Global Change. June 11-15, 1990, University of Alaska Fairbanks. 63-70. Fairbanks, AK, University of Alaska, Geophysical Institute.

Jacka, T. H. and Budd, W. F. 1998. Detection of temperature and sea-ice-extent changes in the Antarctic and Southern Ocean, 1949-96. *Annals of Glaciology* **27**: 553-559.

Jacob U (2006) Trophic dynamics of Antarctic shelf ecosystems – food webs and energy flow budgets. PhD thesis, AWI Bremen University

Jacobs & Mitchell, 1996 – chapter 3

Jacobs S S & Comiso J C (1993) A recent sea-ice retreat west of the Antarctic Peninsula, Geophys. Res. Lett., 20, 1171-1174.

Jacobs, S. S., and J. C. Comiso, 1997. Climate variability in the Amundsen and Bellingshausen Seas. Journal of Climate, 10(4), 697-709.

Jacobs, S. S., C. F. Giulivi, and P. A. Mele, 2002. Freshening of the Ross Sea during the late 20th century. Science 297(5580), 386-389, doi:10.1126/science.1069574.

Jacobs, S. S., D. R. MacAyeal, and J. L. Ardai, Jr., 1986. The recent advance of the Ross Ice Shelf, Antarctica. Journal of Glaciology 32(112), 464-474.

Jacobs, S. S., H. H. Hellmer, and A. Jenkins, 1996. Antarctic ice sheet melting in the southeast Pacific. Geophysical Research Letters, 23(9), 23(9), 957-960.

Jacobs, S.S. (1985) Oceanographic evidence for land ice/ocean interactions in the Southern ocean. In: Glaciers, Ice Sheets and Sea Level: Effect of a CO2-Induced Climatic Change, Report of a Workshop, Seattle, 13-15 Sep 1984, DOE/ER/60235-1, 116-128.

Jacobs, S.S. (2006) Observations of change in the Southern Ocean. *Phil. Trans. Roy. Soc. A,* **364**, 1657-1681.

Jacobs, S.S. and C.F. Giulivi (1998) Interannual ocean and sea ice variability in the Ross Sea. *Antarct. Res. Ser.* **75**, 135-150.

Jacobs, S.S. and C.F. Giulivi (1999) Thermohaline Data and Ocean Circulation on the Ross Sea Continental Shelf. In: *Oceanography of the Ross Sea, Antarctica*, G. Spezie & G. Manzella (eds), Springer, Milan, 3-16.

Jacobs, S.S., H.H. Hellmer and A. Jenkins, 1996: Antarctic ice sheet melting in the Southeast Pacific. *Geophys. Res. Lett.*, **23**(9), 957–960.

Jacobson, H.P., and C.F. Raymond (1998), Thermal effects on the location of ice stream shear margins. *Journal of Geophysical Research 103*, 12,111-12,122.

Jansen E, J., Overpeck KR, Briffa J-C, Duplessy F, Joos V, Masson-Delmotte D, Olago B, Otto-Bliesner WR, Peltier S, Rahmstorf R, Ramesh D, Raynaud D, Rind O, Solomina R, Villalba, Zhang D (2007) Palaeoclimate. In: Solomon S, D.,, Qin M, Manning Z, Chen M, Marquis KB, Averyt M, Tignor, Miller HL (eds) *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental*

Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, p 433-497

Jenkins, A., 1991: A one-dimensional model of ice shelf-ocean interaction. *J. Geophys. Res.*, **96**(C11), 20,671-20,677.

Jenkins, A., D. G. Vaughan, S. S. Jacobs, H. H. Hellmer, and J. R. Keys, 1997. Glaciological and oceanographic evidence of high melt rates beneath Pine Island Glacier, West Antarctica. Journal of Glaciology 43(143), 114-121.

Jenkins, A., D.M. Holland, K.W. Nicholls, M. Schröder and S. Østerhus, 2004: Seasonal ventilation of the cavity beneath Filchner-Ronne Ice Shelf simulated with an isopycnic coordinate ocean model. *J. Geophys. Res*. **109**, C01024, doi:10.1029/2001JC001086.

Jenkins, A., H. F. J. Corr, K. W. Nicholls, C. L. Stewart, C. S. M. Doake, Christopher S.M., 2006, Interactions between ice and ocean observed with phase-sensitive radar near an Antarctic ice-shelf grounding line. *J. Glaciol.* **52** (178), 325-346.

Jenouvrier S, Barbraud C, Weimerskirch H (2003) Effects of climate variability on the temporal population dynamics of southern fulmars. J Animal Ecol 72: 576–587

Jenouvrier S, Barbraud C, Weimerskirch H (2005). Long-term contrasted responses to climate of two antarctic seabird species. Ecol 86: 2889–2903

Jenouvrier S,Weimerskirch H, Barbraud C, Park Y H and Cazelles B (2005a) Evidence of a shift in the cyclicity of Antarctic seabird dynamics linked to climate. Proceeding of the Royal Society B, 272, 887-895.

Jenssen, D. (1977), A three-dimensional polar ice sheet model. *Journal of Glaciology 18*, 373-389.

Jevrejeva et al. (200?) - chapter 4

Jin Y, DeVries A (2006) Antifreeze glycoprotein levels in Antarctic notothenioid fishes inhabiting different environments and the effect of warm acclimation. Comp Biochem Physiol B 144: 290-300

Johannessen OM, Miles MW (2000) Arctic sea ice and climate change - will the ice disappear in this century? Sci Prog 83: 209-222

Johnsen K-P, Miao J, Kidder SQ. 2004. Comparison of atmospheric water vapour overarctica derived from CHAMP/GPS and AMSU-B data. *Physics and Chemistry of the Earth*, **29**: 251–255.

Johnson, G. C., and H. L. Bryden, 1989. On the size of the Antarctic Circumpolar Current. *Deep Sea Res.*, **36**, 39–53.

Johnston IA (2003) Muscle metabolism and growth in Antarctic fishes (suborder Notothenioidei):

Johnston IA, Calvo J, Guderley H, Fernandez D, Palmer L (1998) Latitudinal variation in the abundance and oxidative capacities of muscle mitochondria in perciform fishes. J Exp Biol 201: 1-12

Johst K, Gutt J, Wissel C, Grimm V (2006) Diversity and disturbances in the Antarctic megabenthos: feasible versus theoretical disturbance ranges. Ecosystems 9: 1145-1155

Jones and Simmonds, 1993 – chapter 3

Jones and Widmann (2003) – chapter 3

Jones VJ, Hodgson DA, Chepstow-Lusty A (2000) Palaeolimnological evidence for marked Holocene environmental changes on Signy Island, Antarctica. The Holocene 10:43-60

Jones, A.E., Anderson, P.S., Wolff, E.W., Turner, J., Rankin, A.M. and Colwell, S. R. ‚A role for newly-forming sea ice in springtime polar tropospheric ozone loss? Observational evidence from Halley station, Antarctica" J. Geophys. Res., 111, D08306, doi:10.1029/2005JD006566 (2006)

Jones, A.E., Weller, R., Anderson, P.S., Jacobi, H.-W., Wolff, E.W., Schrems, O., and Miller, H.: Measurements of NO_x emissions from the Antarctic snowpack, Geophys. Res. Lett., (28), 1499-1502, 2001.

Jones, A.E., Weller, R., Wolff, E.W., and Jacobi, H.-W.: Speciation and rate of photochemical NO and NO_2 production in Antarctic snow, Geophys. Res. Lett., (27), 345-348, 2000.

Jones, P. D. 1995. Recent variations in mean temperature and the diurnal temperature range in the Antarctic. *Geophysics Research Letters* **22** [11]: 1345-1348.

Jorgensen, M.T., Y.L. Shur, and E.R. Pullman. 2006. Abrupt increase in permafrost degradation in arctic Alaska. Geophysical Research Letters 33, L020503, doi:10.1029/2005GL024960.

Joseph, R. and S. Nigam, 2006: ENSO evolution and teleconnections in IPCC's twentieth-century climate simulations: Realistic representation? *J. Climate*, **19**, 4360-4377.

Joughin et al. 2005 – chapter 3

Joughin, I. And S. Tulaczyk, 2002. Positive mass balance of the Ross Ice Streams, West Antarctica, *Science*, Vol 295 (Jan 18), No. 5554, p. 476-480.

Joughin, I., 2002, Ice-sheet velocity mapping: a combined interferometric and speckle-tracking approach. *Ann. Glaciol.* **34**, 195–201.

Joughin, I., E. Rignot, C. Rosanova, B. Lucchitta, and J. Bohlander, 2003. Timing of recent accelerations of Pine Island Glacier, Antarctica. Geophysical Research Letters, 30(13), doi:10.1029/2003GL017609.

Joughin, I., Gray, L., Bindschadler, R., Price, S., Morse, D., Hulbe, C., Mattar, K. & Werner, C., 1999, Tributaries of West Antarctic ice streams revealed by Radarsat interferometry. *Science* **286**, 283–286, (doi:10.1126/science.286.5438.283).

Joughin, I., S. Tulaczyk, R.A. Bindschadler and S.F. Price, 2002. Changes in west Antarctic ice stream velocities, *Journal of Geophysical Research*, Vol. 107, No. B11, p. 2289, doi:10.1029/2001/JB001029.

Jouzel J, Masson V, Cattani O, Falourd S, Stievenard M, Stenni B, Longinelli A, Johnsen SJ, Steffenssen JP, Petit JR, Schwander J, Souchez R, Barkov NI (2001) A new 27 ky high resolution East Antarctic climate record. Geophysical Research Letters 28:3199-3202

Jouzel J, Masson-Delmotte V, Cattani O, Dreyfus G, Falourd S, Hoffmann G, Minster B, Nouet J, Barnola JM, Chappellaz J, Fischer H, Gallet JC, Johnsen S, Leuenberger M, Loulergue L, Luethi D, Oerter H, Parrenin F, Raisbeck G, Raynaud D, Schilt A, Schwander J, Selmo E, Souchez R, Spahni R, Stauffer B, Steffensen JP, Stenni BS, T.F., Tison JL, Werner M, Wolff E (2007) Orbital and millenial Antarctic climate variability over the past 800,000 years. Science 317:793-796

Jouzel J, Vaikmae R, Petit JR, Martin M, Duclos Y, Stiévenard M, Lorius C, Toots M, A. MM, Burckle LH, Barkov NI, Kotlyakov VM (1995) The two-step shape and timing of the last deglaciation in Antarctica. Climate Dynamics 11:151-161

Jouzel J, Vimeux F, Caillon N, Delaygue G, Hoffmann G, Masson-Delmotte V, Parrenin F (2003) Magnitude of the Isotope/Temperature scaling for interpretation of central Antarctic ice cores. J.Geophys.Res. 108:1029-1046

Joyce, T. M., Zenk, W. and Toole, J. M. (1978). The anatomy of the Antarctic Polar Front in the Drake Passage. *J. Geophys. Res.* **83**, 6093-6113.

Kalnay, E. (1996), The NCEP/NCAR 40-year reanalysis project, *Bulletin of the American Meteorological Society*, *77*, 437-470.

Kamb, B., Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion, in The West Antarctic Ice Sheet: Behavior and Environment, AGU Antarct. Res. Ser., vol. 77, edited by R. B. Alley and R. A. Bindschadler, pp. 157– 201, AGU, Washington, D.C., 2001a.

Karoly, D. J. 1989. Southern Hemisphere circulation features associated with El Niño-Southern Oscillation events. *Journal of Climate* **2**: 1239-1252.

Kazendehar et al., 2007 – chapter 3

Kejna, M. and Laska, K. 1999. Spatial differentiation of ground temperature in the region of Arctowski Station, King George Island, Antarctica in 1996. *Polish Polar Research*, 20(3), 221-241.

Kennett (1978) – Chapter 2

Kennett JP (1977) Cenozoic evolution of the Antarctic glaciation, the circum-Antarctic ocean and their impact on global paleoceanography. J Geophys Res 82: 3843-3876

Kennett JP, Hodell DA (1993) Evidence for Relative Climatic Stability of Antarctica during the Early Pliocene: A Marine Perspective. Geografiska Annaler. Series A, Physical Geography 75:205-220

Kennicutt MC, McDonald SJ, Sericano JL, Boothe P, Oliver J, Safe F, Presley BJ, Liu H, Wolfe D, Wade TL, Crockett A, Bockus D (1995) Human contamination of the marine environment: Arthur Harbor and McMurdo Sound, Antarctica. Environ Sci Technol 29:1279-1287

Key R M, Kozyr A, SabineC L, Lee K, Wanninkhof R, Bullister J, Feely R A, Millero F, Mordy C, Peng T-H (2004) A global ocean carbon climatology: results from GLODAP. Global Biogeochemical Cycles 18, GB4031

Keys, H., S. Jacobs, and D. Barnett, 1990. The calving and drift of iceberg B-9 in the Ross Sea, Antarctica. Antarctic Science 2(3), 243-257.

Khazendar, A., E. Rignot, and E. Larour, 2007. Larsen B Ice Shelf rheology preceeding its disintegration inferred by a control method. Geophysical Research Letters 34 L19503, doi:10.1029/2007GL030980.

Kidson, 1999 – chapter 3

Kim, K., K. C. Jezek, and H.Liu, 2007. Orthorectifield image mosaic of Antarctica from 1963 Argon satellite photography: image processing and glaciological applications. International Journal of Remote Sensing, 28(23), 5357-5373, doi: 10.1080/01431160601105850.

King J C (1994) Recent climate variability in the vicinity of the Antarctic Peninsula, Int. J. Climat., 14, 357-369.

King, 1994 - chapter 4 - Which one – 2 referring to 1994 (page 10)

King, J. C. (2003), Antarctic Peninsula climate variability and its causes as revealed by analysis of instrumental records, *Antarctic Peninsula Climate Variability: Historical and Paleoenvironmental Perspectives. Antarctic Research Series, 79*, edited by E. Domack, et al., pp. 17-30, AGU, Washington, DC.

King, J. C. and Comiso, J. C. 2003. The spatial coherence of interannual temperature variations in the Antarctic Peninsula. *Geophysics Research Letters* **30** [2 10.1029/2002GL015580].

King, J. C. and Harangozo, S. A. 1998. Climate change in the western Antarctic Peninsula since 1945: observations and possible causes. *Annals of Glaciology* **27**: 571-575.

King, J. C. and Turner, J. 1997. Antarctic Meteorology and Climatology. 1-409. Cambridge, Cambridge University Press.

King, J. C., 1994. Recent variability in the Antarctic Peninsula. International Journal of Climatology 14(4), 357-369.

King, M. A., 2004, Rigorous GPS data-processing strategies for glaciological. applications. *J. Glaciol.* **50**, 601–607.

King, M.A., R. Coleman, P.J. Morgan, and R.S. Hurd, 2007, Velocity change of the Amery Ice Shelf, East Antarctica, during the period 1968-1999, *J. Geophys. Res.*, **112**, F01013, doi:10.1029/2006JF000609.

Kirkup H, Melles M, Gore DB (2002) Late Quaternary environment of Southern Windmill Islands, East Antarctica. Antarctic Science 14:385-394

Klaas, C., and D. E. Archer (2002), Association of sinking organic matter with various types of mineral ballast in the deep sea: Implications for the rain ratio, *Global Biogeochemical Cycles*, *16*.

Klatt, O., E. Fahrbach, M. Hoppema, and G. Rohardt (2005), The transport of the Weddell Gyre across the Prime Meridian, Deep-Sea Research II, 52, 513-528.

Klatt, O., O. Boebel, and E. Fahrbach, A profiling float's sense of ice. *J. Atmos. Oceanic Technol*., 24, 1301-1308, 2007

Knorr G, Lohmann G (2003) Southern Ocean origin for the resumption of Atlantic thermohaline circulation during deglaciation. Nature 424:532-536

Köhler P, Fischer H (2006) Simulating low frequency changes in atmospheric CO_2 during the last 740 000 years. Climate of the Past 2:57-78

Kohler P, Fischer H, Munhoven G, Zeebe RE (2005) Quantitative interpretation of atmospheric carbon records over the last glacial termination. Glob. Biogeochem. Cycles 19:Art. No. GB4020

Koide M, Goldberg ED, Herron MM, Langway CC (1979) Deposition history of artificial radionuclides in the Ross Ice Shelf, Antarctica. Earth Planet Sci Lett 44:205-223

Korb R.E, M.J. Whitehouse and P. Ward, SeaWiFS in the Southern Ocean: spatial and temporal variability in phytoplankton biomass around South Georgia, *Deep-Sea Research Part II—Topical Studies in Oceanography* **51** (2004) (1–3), pp. 99–116.

Kreher, K., P. V. Johnston, S. W. Wood, B. Nardi, and U. Platt. Ground-based measurements of tropospheric and stratospheric BrO at Arrival Heights, Antarctica. Geophys. Res. Lett., 24(23), 3021 – 3024, 1997

Kreutz KJ, Mayewski PA, Meeker LD, Twickler MS, Whitlow SI (2000) The effect of spatial and temporal accumulation rate variability in West Antarctica on soluble ion deposition. Geophysical Research Letters 27:2517

Krinner G., Magand O., Simmonds I., Genthon C., Dufresne J.L., 2007: Simulated Antarctic precipitation and surface mass balance at the end of the twentieth and twenty-first centuries. *Clim. Dyn.*, **28**, 215-230.

Kulbe T, Melles M, Verkulich SR, Pushina ZV (2001) East Antarctic climate and environmental variability over the last 9400 years inferred from marine sediments of the Bunger Oasis. Arctic Antarctic and Alpine Research 33:223-230

Kunze, E., Firing, E., Hummon, J. M., Chereskin, T. K. and Thurnherr, A. M. (2006). Global abyssal mixing inferred from lowered ADCP shear and CTD strain profiles. *J. Phys. Oceanogr.* **36**, 1553-1576.

Kurbatov AV, Zielinski GA, Dunbar NW, Mayewski PA, Meyerson EA, Sneed SB, Taylor KC (2006) A 12,000 year record of explosive volcanism in the Siple Dome Ice Core, West Antarctica. Journal of Geophysical Research 111, D12307, doi:10.1029/2005JD006072.

Kwok, R., and J.C. Comiso, 2002: Southern ocean climate and sea ice anomalies associated with the Southern Oscillation. *J. Climate*, **15**, 487–501.

Kwok, R., G. F. Cunningham, H. J. Zwally, and D. Yi, "ICESat over Arctic Sea Ice: Retrieval of Freeboard", *J. Geophys. Res.*, 2006.

L'Heureux, M.L., and D.W.J. Thompson, 2006: Observed relationships between the El-Niño/Southern Oscillation and the extratropical zonal mean circulation. *J. Climate*, **19**, 276–287.

Lachlan-Cope TA. 1992. The use of a simultaneous physical retrieval scheme for satellite derived atmospheric temperaturesL Weddell Sea, Antarctica. *International Journal of Remote Sensing*, **13**: 141–154.

Lachlan-Cope, T. A. and Connolley, W. M. 2006. Teleconnections between the tropical Pacific and the Amundsen-Bellingshausen Sea: role of the El Niño/Southern Oscillation. *Journal of Geophysical Research* **111** [D23]: doi:10.1029/2005JD006386.

Lacis, A.A., Weubbles, D.J., Logan, J.A., Radiative forcing of climate change by changes in the vertical distribution of ozone, J. Geophys. Res., 95, 9971-9981, 1990.

Lambeck K, Chappell J (2001) Sea level change through the last glacial cycle. Science 292:679-686

Langdon, C., and M. J. Atkinson (2005), Effect of elevated pCO(2) on photosynthesis and calcification of corals and interactions with seasonal change in temperature/irradiance and nutrient enrichment, *Journal of Geophysical Research-Oceans*, *110*, -.

Lannig G, Bock C, Sartoris FJ, Pörtner HO (2004) Oxygen limitation of thermal tolerance in cod, *Gadus morhua* L. studied by non-invasive NMR techniques and on-line venous oxygen monitoring. Amer J Physiol 287: R902–R910

Lannig G, Storch D, Pörtner HO (2005) Aerobic mitochondrial capacities in Antarctic and temperate eelpout (Zoarcidae) subjected to warm versus cold acclimation. Polar Biol 28: 575-584. DOI: 10.1007/s00300-005-0730-9.

Large, W.G. and Nurser, A.J.G., 1998: Ocean Surface Water Mass Transformations. Pp 317-336 in: Eds: Siedler, G., Church, J. and Gould, J.: Ocean Circulation and Climate. Academic Press, San Diego, 715pp.

Large, W.G., Williams, J.C. and Doney, S.C., 1994: Oceanic vertical mixing: a review and a model with nonlocal boundary layer parameterization. Reviews of Geophysics, 32, 363-403.

Law, R. M., et al. (2008), Comment on "Saturation of the Southern Ocean CO2 Sink Due to Recent Climate Change", *Science*, *in press*.

Laws 1984 – chap 1

Lazzara, M. A., K. C. Jezek, T. A. Scambos, D. R. MacAyeal, and C. J. Van der Veen, 1999. On the recent calving of icebergs from the Ross Ice Shelf. Polar Geography, 23(3), 201-212.

Le Quere, C., et al. (2007), Saturation of the Southern Ocean CO2 Sink Due to Recent Climate Change, *Science %R 10.1126/science.1136188*, *316*, 1735-1738.

Le Quéré, C., et al. (2008), Response to Comments on 3Saturatio of the Southern Ocean CO2 Sink Due to Recent Climate Change", *Science*, *319*, 570c.

Leaper R, Cooke J, Trathan P N, Reid K, Rowntree V and Payne R (2006) Global climate drives southern right whale (*Eubalaena australis*) population dynamics. Biology Letters, 2, 289–292.

Ledley and Huang, 1997 – chapter 3

Lee, A.M., H.K. Roscoe, A.E. Jones, P.H. Haynes, E.F. Shuckburgh, M.J. Morrey, H.C. Pumphrey, "The impact of the mixing properties within the Antarctic stratospheric vortex on ozone loss in spring", *J. Geophys. Res.* **106**, 3203-3211 (2001).

Lefebvre W. and H. Goosse, 2007. Analysis of the projected regional sea-ice changes in the Southern Ocean during the 21st century. Climate Dynamics DOI 10.1007/s00382-007-0273-6.

Lefebvre, W., H. Goosse, R. Timmermann, T. Fichefet, 2004: Influence of the Southern Annular Mode on the sea ice-ocean system. *J. Geophys. Res.*, **109**, C09005, doi:10.1029/2004JC002403.

Legg, S., Hallberg, R.R., Girton, J.B., 2006: Comparison of entrainment in overflows simulated by z-coordinate, isopycnal and non-hydrostatic models. Ocean Modelling, 11 (1-2), 69-79.

Legresy, B., and F. Remy, 1997, Altimetric observations of surface characteristics of the Antarctic ice sheet, *J. Glaciol.*, ***43***(144), 265–275.

Lemke P, Ren J, Alley R, Allison I, Carrasco J, Flato G, Fujii Y, Kaser G, Mote P, Thomas R, Zhang T (2007) Observations: change in snow, ice and frozen ground. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, pp 337-384 (Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL, Eds), Cambridge University Press, Cambridge, UK

Lemke, 2007 #2408 – chapter 3 & 4 – Lemke et al 2007?

Lenihan HS, Oliver JS (1995) Anthropogenic and natural disturbance to main benthic communities in Antarctica. Ecol Appl 5:311-326.

Lenihan HS, Oliver JS, Oakden JM, Stephenson MD (1990) Intense and localized benthic marine pollution around McMurdo Station, Antarctica. Mar Pollut Bull 21:422-430

Lenton, A., and R. J. Matear (2007), The role of the Southern Annular Mode (SAM) in Southern Ocean CO2 uptake, *Global Biogeochemical Cycles*, *21*, doi: 10:1029/2006GB002714.

Lenton, A., et al. (2006), Simulating the Ocean Carbon Cycle: A focus on the Southern Ocean, paper presented at CARBOOCEAN Annual Meeting, Spain.

Lenton, A., et al. (2008), Strategies for High Latitude CO2 Sampling Now and in the Future, *Deep Sea Research II*, *SOCOVV Special Volume*.

Leventer A, E. Domack, A. Barkoukis, B. McAndrews, Murray J (2002) Laminations from the Palmer Deep: A diatom-based interpretation. *Paleoceanography 17*:8002

Levitus, S. and T. Boyer. 1994. World Ocean Atlas. Vol. 4: Temperature, 117 pp., NOAA, Washington, D.C.

Levitus, S., J. Antonov, and T. Boyer, Warming of the world ocean. *Geophysical Res. Letters*, **32**, L02604, doi:10.1029/2004GL021592 (2005).

Levitus, S., J.I. Antanov, T.P. Buyer and C. Stephens (2000) Warming of the world ocean, Science, 287 (5461), 2225-2229.

Levitus, S., R. Burgett, and T. Boyer. 1994. World Ocean Atlas. Vol. 3: Salinity, 100 pp., NOAA, Washington, D.C.

Li J., H. J. Zwally, J. C. Comiso, 2007, Ice-sheet elevation changes caused by variations of the firn compaction rate induced by satellite-observed temperature variations (1982–2003). *Ann. Glaciol.* **46**, 8-13.

Li, Z.X., 2000. Influence of Tropical Pacific El Nino on the SST of the Southern Ocean through atmospheric bridge. Geophysical Research Letters 27 (21), 3505-3508.

Licht KJ, Dunbar NW, Andrews JT, Jennings AE (1999) Distinguishing subglacial till and glacial marine diamictons in the western Ross Sea, Antarctica:

implications for last glacial maximum grounding line. Geological Society of America Bulletin 111:91-103

Licht KJ, Jennings AE, Andrews JT, Williams KM (1996) Chronology of late Wisconsin ice retreat from the western Ross Sea, Antarctica. Geology 24:223-226

Limpasuvan and Hartmann, 2000 – chapter 3

Lisiecki LE, Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic d18O records. Paleoceanography 20:PA1003, doi:1010.1029/2004PA001071

Liss, P. S., and L. Merlivat (1986), Air-sea gas exchange rates: Introduction and synthesis, in *The Role of Air-Sea Exchange in Geochemical Cycling*, edited by P. Buat-Menard, pp. 113-129, Reidel, Boston.

Liu J P, Yuan X J, Rind D and Martinson D G (2002) Mechanism study of the ENSO and southern high latitude climate teleconnections. Geophysical Research Letters 29, 1679.

Liu X, Sun L, Xie Z, Yin X, Zhu R, Wang Y (2007) A preliminary record of the historical seabird population in the Larsemann Hills, East Antarctica, from geochemical analyses of Mochou Lake sediments. Boreas 36:182-197

Liu X-D, Li H-C, Sun L-G, Yin X-B, Zhao S-P, Wang Y-H (2006) δ13C and δ15N in the ornithogenic sediments from the Antarctic maritime as palaeoecological proxies during the past 2000 yr. Earth and Planetary Science Letters 243:424-438

Liu XD, Sun L, Xie Z, Yin X, Wang Y (2005) A 1300-Year Record of Penguin Populations at Ardley Island in the Antarctic, as Deduced From the Geochemical Data in the Ornithogenic Lake Sediments. Arctic Antarctic and Alpine Research 37:490-498

Liu, H., L. Wang and K.C. Jezek. 2006. Spatiotemporal variations of snowmelt in Antarctica derived from satellite scanning multichannel microwave radiometer and Special Sensor Microwave Imager data (1978–2004). *J. Geophys. Res.,* 111(F1), F01003. (10.1029/2005JF000318.)

Liu, J., J.A. Curry, and D.G. Martinson, 2004: Interpretation of recent Antarctic sea ice variability. *Geophys. Res. Lett.*, **31**, L02205, doi:10.1029/2003GL018732.

Lo Monaco, C., et al. (2005), Anthropogenic CO_2 in the Southern Ocean: Distribution and inventory at the Indian-Atlantic boundary (WOCE line I6), *Journal of Geophysical Research*, *110*, 18.

Lo Monaco, personal communication – chapter 3 – page 156

Long DG, Drinkwater MR, Holt B, Saatchi S, Bertoia C. 2001. Global ice and land climate studies using scatterometer image data. *EOS, Transactions of the*

American Geophysical Union **82:** 503 (see also *http://www.agu.org/eos_elec/010126e.html*).

Long, D.G. and M.R. Drinkwater. 1999. Cryosphere applications of NSCAT data. *IEEE Trans. Geosci. Remote Sens.*, **37**(3), 1671–1684.

Longton 1988 – Chapter 3

Lorius C, Merlivat L, Jouzel J, Pourchet M (1979) A 30,000-yr isotope climatic record from Antarctic ice. Nature 280:644-648

Löscher, B. M., et al. (1997), The distribution of Fe in the Antarctic Circumpolar Current, *Deep-Sea Research II*, *44*, 143-187.

Loulergue L, Schilt A, Spahni R, Masson-Delmotte V, Blunier T, Lemieux B, Barnola JM, Raynaud D, Stocker T, Chappelaz J (submitted) Orbital and millenial-scale features of atmospheric CH_4 over the last 800 000 years. Nature

Lovenduski, N., et al. (2007), Enhanced CO2 outagssing in the Southern Ocean from a positive phase of the Southern Annular Mode, *Global Biogeochemical Cycles*, *in press*.

Lowe AL, Anderson JB (2002) Late Quaternary advance and retreat of the West Antarctic Ice Sheet in Pine Island Bay, Antarctica. Quaternary Science Reviews 21:1879-1897

Lowe and Gregory, 2005 - chapter 4

Lowe and Gregory, 2006 - chapter 4

Lowe CJ, Davison W (2005) Plasma osmolarity, glucose concentration and erythrocyte responses of two Antarctic nototheniid fishes to acute and chronic thermal change. J Fish Biol 67(3): 752-766

Lowe, 200 - chapter 4 - is this a reference?

Luchitta, B. K. and C. E. Rosanova (1998), Retreat of northern margins of George VI and Wilkins ice shelves, Antarctic Peninsula, *Ann. Glaciol*. 27, 41.

Lynch, A., P. Uotila, and J.J. Cassano, 2006: Changes in synoptic weather patterns in the polar regions in the twentieth and twenty-first centuries, Part 2: Antarctic. *Int. J. Climatol.*, **26**, 1181–1199.

Lyons et al. 2006 RiSCC – chapter 3 & 4

Lyons WB, Tyler SW, Wharton RA, McKnight DM, Vaughan BH (1998) A late Holocene desiccation of Lake Hoare and Lake Fryxell, McMurdo Dry Valleys, Antarctica. Antarctic Science 10:247-256

Lythe, M.B., Vaughan, D.G. and the BEDMAP Consortium. (2000), BEDMAP - bed topography of the Antarctic. 1:10,000,000 scale map. BAS (Misc) 9. Cambridge, British Antarctic Survey.

Lythe, M.B., D.G. Vaughan, The Bedmap consortium, 2001, Bedmap: a new ice thickness and subglacial topographic model of Antarctica, J. Geophys. Res. **106** 11335–11351.

MacAyeal, 2003 – Chapter 3 – MayAyeal et al 2003?

MacAyeal, D. R., T. A. Scambos, C. L. Hulbe and M. A. Fahnestock, 2003, Catastrophic Ice-Shelf Break-Up by an Ice-Shelf-Fragment-Capsize Mechanism, *J. Glaciol.*, **49**(164) 22-36.

MacAyeal, D.R. (1984) Thermohaline circulation below the Ross Ice Shelf: A consequence of tidally induced vertical mixing and basal melting. *J. Geophys. Res.* **89**(C1), 597-606.

MacAyeal, D.R. (1989), Large-scale ice flow over a viscous basal sediment: theory and application to Ice Stream B, Antarctica. *Journal of Geophysical Research 94*, 4071-4087.

Mackintosh [1972] – chapter 3

Mackintosh and Herdman [1940] – chapter 3

Madden, R.A. and P.R. Julian. Observations of the 40-50 day tropical oscillation - A review. *Monthly Weather Review, 122,* 814-837, 1994.

Mahadevan, A., et al. (2004), Mesoscale variability of sea surface pCO_2: What does it respond to?, *Global Biogeochemical Cycles*, *18*.

Mahaffy, M.W. (1976), A three-dimensional numerical model of ice sheets: tests on the Barnes Ice Cap, Northwest Territories. *Journal of Geophysical Research 81*, 1059-1066.

Makinson K. and K.W. Nicholls, 1999: Modeling tidal currents beneath Filchner-Ronne Ice Shelf and on the adjacent continental shelf: their effects on mixing and transport. *J. Geophys. Res.*, **104**(C6), 13,449-13,465.

Makinson, K. (2002). Modeling tidal current profiles and vertical mixing beneath Filchner-Ronne Ice Shelf, Antarctica. *J. Phys. Oceanogr.* **32**, 202-215.

Makinson, K., Schröder, M. and Osterhus, S. (2006). Effect of critical latitude and seasonal stratification on tidal current profiles along Ronne Ice Front, Antarctica. *J. Geophys. Res.* **111**, doi:10.1029/2005JC003062.

Maltrud, M.E., and McClean, J.L., 2005: An eddy resolving global 1/10 degrees ocean simulation. Ocean Modelling, 8(1-2), 31-54.

Manabe S., R.J. Stouffer, M.J. Spelman and K. Bryan (1991). Transient responses of a coupled atmosphere-ocean model to gradual changes of atmospheric CO_2. I. Annual mean response. J. Clim. 4, 785-818.

Manabe, S and R.J. Stouffer (1993). Century-scale effects of increased atmospheric CO2 on the ocean-atmosphere system. Nature 364, 215-218.

Mann ME, Jones PD (2003) Global surface temperatures over the past two millennia. Geophysical Research Letters 30, doi: 10.129/2003GL017814

Manzini, E., B. Steil, C. Brühl, M.A. Giorgetta, K. Krüger, "A new interactive chemistry-climate model: 2. Sensitivity of the middle atmosphere to ozone depletion and increase in greenhouse gases and implications for recent stratospheric cooling", *J. Geophys. Res.*, **108**, 4429 (2003).

Marchant D.R., A. Lewis, W.M. Phillips, R. Souchez, G.H. Denton, D.E. Sugden, and P. Landis. 2002. Formation of patterned ground and sublimation till over Miocene glacier ice in Beacon Valley, southern Victoria Land. Geological Society of America Bulletin 114:718-730.

Mark FC, Bock C, Pörtner HO (2002) Oxygen limited thermal tolerance in Antarctic fish investigated by magnetic resonance imaging (MRI) and spectroscopy (^{31}P-MRS). Amer J Physiol 283: R1254-R1262

Marshall 2002 – chapter 3

Marshall and Connolley (2006) – chapter 3

Marshall DJ, Coetzee L (2000) Historical biogeography and ecology of a continental Antarctic mite genus, Maudheimia (Acari, Oribatida): evidence for a Gondwanan origin and Pliocene-Pleistocene speciation. Zoological Journal of the Linnean Society 129:111-128

Marshall DJ, Convey P (2004) Latitudinal variation in habitat specificity of ameronothroid mites. Experimental and Applied Acarology 34:21-35

Marshall DM, Pugh PJA (1996) Origin of the inland Acari of Continental Antarctica, with particular reference to Dronning Maud Land. Journal of the Linnean Society of Zoology 118

Marshall GJ, Turner J. 1997a. Surface wind fields of Antarctic mesocyclones derived from ERS-1 scatterometer data. *Journal of Geophysical Research* **102:** 13907–13921.

Marshall GJ, Turner J. 1997b. Katabatic wind propagation over the western Ross Sea observed using ERS-1 scatterometer data. *Antarctic Science* **9:** 221–226.

Marshall GJ, Turner J. 1999. Synoptic-scale weather systems observed during the FROST project via scatterometer winds. *Weather and Forecasting* **14:** 867–877.

Marshall, A., A.H. Lynch, and K. Görgen, 2007: A model intercomparison study of the Australian summer monsoon. Part II: Variability at intraseasonal and interannual timescales Aust. Met. Mag. (in review).

Marshall, D. P. and Naveira Garabato A. C. (2007). A conjecture on the role of bottom-enhanced diapycnal mixing in the parameterization of geostrophic eddies. *J. Phys. Oceanogr.*, in press.

Marshall, G. J. (2003), Trends in the Southern Annular Mode from Observations and Reanalyses, *Journal of Climate*, *16*, 4134-4143.

Marshall, G. J., Orr, A., van Lipzig, N. P. M. and King, J. C. 2006. The impact of a changing Southern Hemisphere Annular Mode on Antarctic Peninsula summer temperatures. *Journal of Climate* **19** [20]: 5388-5404.

Marshall, G.J., 2007: Half-century seasonal relationships between the Southern Annular Mode and Antarctic temperatures. *Int. J. Climatol*., **27**, 373-383.

Marshall, J., D. Olbers, H. Ross, and D. Wolf-Gladrow, 1993. Potential Vorticity Constraints on the dynamics and Hydrography of the Southern Ocean. *J. Phys. Oc*eanogr., **23**, 465–487.

Marshall, J., H. Jones, R. Karsten, and R. Wardle, 2002. Can eddies set the ocean stratification? *J. Phys. Oceanogr.*, **32**:26-38**.**

Martín, C., R. C. A. Hindmarsh, F. J. Navarro, 2006, Dating ice flow change near the flow divide at Roosevelt Island, Antarctica, by using a thermomechanical model to predict radar stratigraphy, *J. Geophys. Res.*, **111**, No. F1, F01011, doi: 10.1029/2005JF000326

Martin, J. H., and S. E. Fitzwater (1988), Iron-deficiency limits phytoplankton growth in the Northeast Pacific Subarctic, *Nature*, *331*, 341-343.

Maslen and Convey, in press – chap 1

Maslin M, Seidov D, Lowe J (2001) Synthesis of the nature and causes of rapid climate transitions during the Quaternary. Geophysical monograph 126:9-52

Massé G, Rowland S, Sicre M-A, Jacob J, Jansen E, Belt S (2008) Abrupt climate changes for Iceland during the last millennium: Evidence from high resolution sea ice reconstructions. Earth and Planetary Science Letters In press

Massom R.A., Pook M.J., Comiso J.C., Adams N., Turner J., Lachlan-Cope T., Gibson T.T., 2004: Precipitation over the interior East Antarctic Ice Sheet related to midlatitude blocking-high activity. *J. Climate*, **17**, 1914–1928

Masson V, Vimeux F, Jouzel J, Morgan V, Delmotte M, Ciais P, Hammer C, Johnsen S, Lipenkov VY, Mosley-Thompson E, Petit JR, Steig EJ, Stievenard M, Vaikmae R (2000) Holocene climate variability in Antarctica based on 11 ice-core isotope records. Quaternary Research 54:348-358

Masson-Delmotte V, Jouzel J, Landais A, Stievenard M, Johnsen SJ, White JWC, Werner M, Sveinbjornsdottir A, Fuhrer K (2005) GRIP Deuterium Excess Reveals Rapid and Orbital-Scale Changes in Greenland Moisture Origin. Science 309:118 - 121

Masson-Delmotte V, Kageyama M, Braconnot P, Charbit S, Krinner G, Ritz C, Guilyardi E, Jouzel J, Abe-Ouchi A, Crucifix M, Gladstone RM, Hewitt CD, Kitoh A, Legrande A, Marti O, Merkel U, Motoi T, Ohgaito R, Otto-Bliesner B, Peltier WR, Ross I, Valdes PJ, Vettoretti G, Weber SL, Wolk F (2006) Past and future polar amplification of climate change: climate model intercomparisons and ice-core constraints. Climate Dynamics

Masson-Delmotte V, Stenni B, Jouzel J (2004) Common millennial-scale variability of Antarctic and Southern Ocean temperatures during the past 5000 years reconstructed from the EPICA Dome C ice core. The Holocene 14:145-151

Matear R.J. and A.C. Hirst (1999). Climate change feedback on the future oceanic CO_2 uptake. Tellus, 51B, 722-733.

Matear, R. J., and A. Lenton (2008), Impact of Historical Climate Change on the Southern Ocean Carbon Cycle, *Journal of Climate*, *accepted*.

Matsumoto GI, Komori K, Enomoto A, Imura S, Takemura T, Ohyama Y, Kanda H (2006) Environmental changes in Syowa Station area of Antarctica during the last 2300 years inferred from organic components in lake sediment cores. Polar Biosciences 19:52-61

Matsuoka, N., Moriwaki, K., Iwata, S., Hirakawa, K. 1990. Ground temperature regimes and their relation to periglacial processes in the S/,or Rondane Mountains, East Antarctica. In: *Proceedings of the NIPR Symposium on Antarctic Geosciences*, 4,55-66.

Matthews, A.J. and M.P. Meredith. “Variability of Antarctic circumpolar transport and the Southern Annular Mode associated with the Madden-Julian Oscillation”. Geophysical Research Letters, 31(24), L24312, 10.1029/2004GL021666, 2004.

Mäusbacher R, Müller J, Schmidt R (1989) Evolution of postglacial sedimentation in Antarctic Lakes (King-George-Island). Zeitschrift Fur Geomorphologie 33:219-234

Mayewski et al., 2004a – Chapter 2 – *2004 but not 2004a*

Mayewski PA, Denton GH, Hughes T (1981) Late Wisconsin ice margins of North America. In: Denton GH, Hughes T (eds) The Last Great Ice Sheets, p 67-178

Mayewski PA, Maasch K (2006) Recent warming inconsistent with natural association between temperature and atmospheric circulation over the last 2000 years. Climate of the Past (Discussions) http://www.copernicus.org/EGU/cp/cpd/2/327/cpd-2-327.htm

Mayewski PA, Maasch K, Yan Y, Kang S, Meyerson E, Sneed S, Kaspari S, Dixon D, Morgan V, van Ommen T, Curran M (2005) Solar forcing of the polar atmosphere. Annals of Glaciology 41:147-154

Mayewski PA, Rohling E, Stager C, Karlén K, Maasch K, Meeker LD, Meyerson E, Gasse F, van Kreveld K, Holmgren S, Lee-Thorp J, Rosqvist G, Rack F, Staubwasser M, (2004) SR (2004) Holocene climate variability. Quaternary Research 62:243-255

Mayewski, P. A. and et al. 2006. The International Trans-Antarctic Scientific Expedition (ITASE): An overview. *Annals of Glaciology* **41**: 180-185.

Mayewski, P. A., Frezzotti, M., Bertler, N., Van Ommen, T., Hamilton, G., Jacka, T. H.; Welch, B., Frey, M., Dahe, Q., Jiawen, R., Simöes, J., Fily, M., Oerter, H., Nishio, F., Isaksson, E., Mulvaney, R., Holmund, P., Lipenkov, V., Goodwin, I., 2005, The International Trans-Antarctic Scientific Expedition (ITASE): an overview. Ann. Glaciol. **41**(1), 180-185.

McBean G, Alekseev G, Chen D, Førland E, Fyfe J, Groisman PY, King R, Melling H, Vose R, Whitfield PH (2005) Arctic climate: past and present. In Arctic Climate Impacts Assessment (ACIA), pp 21-60 (Symon C, Arris L, Heal B, Eds), Cambridge University Press, Cambridge, UK

McCarron JJ, Larter RD (1998) Late Cretaceous to Early Tertiary Subduction History of the Antarctic Peninsula. J. Geol. Soc. Lond. 155:255-268

McCartney, MS and Donohue, KA, 2007. A deep cyclonic gyre in the Australian-Antarctic Basin, Progress in Oceanography, 75, 675-750.

McConnell, J.C., Henderson, G.S., Barrie, L., Bottenheim, J., Niki, H., Langford, C.H., and Templeton, E.M.J.: Photochemical bromine production implicated in Arctic boundary-layer ozone depletion, Nature, 355, 150-152, 1992.

McCormac F, Hogg A, Blackwell P, Buck C, Higham T, Reimer P (2004) Shcal04 Southern Hemisphere Calibration 0-11.0 Cal Kyr Bp. Radiocarbon 46:1087-1092

McGinnis, L.D. (ed.). 1981. Dry Valley Drilling Project. Antarctic Research Series 33, American Geophysical Union, Washington, D.C.

McGowan J A, Cayan D R and Dorman L M (1998) Climate-ocean variability and ecosystem response in the northeast Pacific. Science 281, 210–217.

McGraw and Day 1997 – chapter 3

McInnes 200 - chapter 4 – is this a reference?

McKay, C.P., Mellon, M.T., Friedmann, E.I. 1998. Soil temperatures and stability of ice-cemented ground in the McMurdo Dry Valleys, Antarctica. *Antarctic Science*, 10(1), 31-38.

McMinn A (2000) Late Holocene increase in sea ice extent in fjords of the Vestfold Hills, eastern Antarctica. Antarctic Science 12:80-88

McMullen K, Domack EW, Leventer A, Dunbar RB, Brachfeld S (2006) Glacial morphology and sediment formation in the Mertz Trough, East Antarctica. Palaeogeography, Palaeoclimatology, Palaeoecology 231:169- 180

McMurdie *et al.*, 1997 – Chapter 3

McNeil, B. I., et al. (2007), An empirical estimate of the Southern Ocean air-sea CO2 flux, *accepted by Global Biogeochemical cycles*.

Meehl, 2007 #2409 - chapter 4

Meehl, G.A., T.F. Stocker, W.D. Collins, P. Friedlingstein, A.T. Gaye, J.M. Gregory, A. Kitoh, R. Knutti, J.M. Murphy, A. Noda, S.C.B. Raper, I.G. Watterson, A.J. Weaver, and Z.-C. Zhao, 2007: Global Climate Projections. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 749-845.

Meeker LD, Mayewski PA (2002) A 1400-year high-resolution record of atmospheric circulation over the North Atlantic and Asia. The Holocene 12:257-266

Melick and Seppelt 1992 – chapter 4

Melles M, Kulbe T, Verkulich SR, Pushina ZV, Hubberten H-W (1997) Late Pleistocene and Holocene environmental history of Bunger Hills, East Antarctica, as revealed by fresh-water and epishelf lake sediments. In: Ricci CA (ed) The Antarctic Region: Geological Evolution and Processes. Proceedings of the VII International Symposium on Antarctic Earth Sciences, Siena, 1995. Terra Antarctica, Sienna, p 809-820

Mellor, G. L., 2003: Users guide for a three-dimensional, primitive equation, numerical ocean model (June 2003 version), 53 pp., Prog. in Atmos. and Ocean. Sci, Princeton University.

Mellor, G.L. and Yamada, T., 1982: Development of a turbulent closure model for geophysical fluid problems. Geophysics and Space Physics, 20, 851-875.

Mercer, J.H. (1968), Antarctic ice and Sangamon sea level. *IASH publ. 79*, 217-225.

Mercer, J.H. (1978), West Antarctic ice sheet and CO2 greenhouse effect: a threat of disaster. *Nature 271*, 321-325.

Meredith M P, Renfrew I A, Clarke A, King J C and Brandon M A (2004) Impact of the 1997/98 ENSO on upper ocean characteristics in Marguerite Bay, western Antarctic Peninsula. Journal of Geophysical Research Oceans, 109, C09013.

Meredith MP, King JC (2005) Rapid climate change in the ocean west of the Antarctic Peninsula during the second half of the 20^{th} century. Geophys Res Lett 32:L19604. (doi: 10.1029/2005GL024042)

Meredith, M. P. and A. M. Hogg, Circumpolar response of Southern Ocean eddy activity to a change in the Southern Annular Mode. *Geophys. Res. Letters*, **33**, doi: 10.1029/2006GL026499 (2006).

Meredith, M.P. and Hughes, C.W. 2005. On the sampling timescale required to reliably monitor interannual variability in the Antarctic circumpolar transport. Geophysical Research Letters, 32(3), L03609, 10.1029/2004GL022086.

Meredith, M.P., E.J. Murphy, E.J. Hawker, J.C. King and M.I. Wallce. On the interannual variability of ocean temperatures around South Georgia, Southern Ocean: forcing by El Niño/Southern Oscillation and the Southern Annular Mode. *Deep-Sea Research II,* in press, 2007.

Meredith, M.P., P.L. Woodworth, C.W. Hughes and V. Stepanov. "Changes in the ocean transport through Drake Passage during the 1980s and 1990s, forced by changes in the Southern Annular Mode". Geophysical Research Letters, 31(21), L21305, 10.1029/2004GL021169, 2004.

Meredith, M.P., R.A. Locarnini, K.A. Van Scoy, A.J. Watson, K.J. Heywood and B.A. King (2000) On the sources of Weddell Gyre Antarctic Bottom Water, J. Geophys. Res., 105, 1093-1104.

Metzl, N. (2008), Decadal Increase of ocean carbon dioxide in the Southwest Indian Ocean Surface Waters (1991-2007), *Deep Sea Research II*, *SOCOVV Special Volume*, accepted.

Metzl, N., et al. (1999), The annual fCO2 cycle and the air-sea CO2 flux in the sub-Antarctic Ocean, *Tellus*, *51B*, 849-861.

Metzl, N., et al. (2006), Summer and Winter Air-Sea CO_2 Fluxes in the Southern Ocean, *Deep-Sea Research II*, *53*.

Metzl, N., et al. (2007), Global Changes in Ocean Carbon: Variability and Vulnerability, *EOS*, *88*, 287.

Meurk *et al.* 1994a – chap 1

Meurk *et al.* 1994b – chap1

Meyers, G., R. J. Bailey and A. P. Worby, 1995. Geostrophic transport of Indonesian Throughflow. Deep-Sea Res., I, 42, 1163-1174.

Meyerson et al., 2002 – Chapter 2

Mignone, B. K., et al. (2006), Central role of the Southern Hemisphere winds and eddies in modulating the oceanic uptake of anthropogenic carbon, *Geophysical Research Letters*, *33*, doi:10.0129/2005GL024464.

Miller, R. L., G. A. Schmidt, and D. T. Shindell, 2006: Forced annular variations in the 20th century Intergovernmental Panel on Climate Change Fourth Assessment Report models. *J. Geophys. Res.*, **111,** doi:10.1029/2005JD006323.

Mincks SL, Smith CR, DeMaster DJ (2005) Persistence of labile organic matter and microbioal biomass in Antarctic shelf sediments: evidence of a sediment ‘food bank’. Mar Ecol Prog Ser 300: 3-19

Mitchell *et al.* 1999 – chap 1

Mitrovica, J. X., Tamisiea, M.E., Milne, G.A. and Davis, J.L. 2001. Recent mass balance of polar ice sheets inferred from patterns of global sea-level change. Nature, 409, 1026-1029.

Miura H, Maemoku H, Igarishi A, Moriwaki K (1998) Late Quaternary raised beach deposits and radiocarbon dates of marine fossils around Lützow-Holm Bay., Special map series of National Institute of Polar Research 6, Tokyo

Mo and Ghil, 1987 – chapter 3

Mo and Higgins (1998) – chapter 3

Molemaker, J., McWilliams, J. C. and Yavneh, I. (2005). Baroclinic instability and loss of balance. *J. Phys. Oceanogr.* **35**, 1505-1517.

Monaghan A.J., Bromwich D.H., Fogt R.L., Wang S.H., Mayewski P.A., Dixon D.A., Ekaykin A., Frezzotti M., Goodwin I., Isaksson E., Kaspari S.D., Morgan V.I., Oerter H., Van Ommen T.D., Van der Veen C.J., Wen J.H., 2006: Insignificant change in Antarctic snowfall since the International Geophysical Year. *Science,* **313**, 827-831.

Monaghan and Bromwich. 2008 – chapter 3

Monaghan et al. (2006) - chapter 4 -a or b page 19

Monaghan, A.J., D.H. Bromwich, and S-H. Wang, 2006a: Recent trends in Antarctic snow accumulation from Polar MM5. *Philosophical Trans. Royal. Soc. A*, **364**, 1683-1708.

Monaghan, A.J., D.H. Bromwich, R.L. Fogt, S.-H. Wang, P.A. Mayewski, D.A. Dixon, A.A. Ekaykin, M. Frezzotti, I. D. Goodwin, E. Isaksson, S.D. Kaspari, V.I. Morgan, H. Oerter, T.D. van Ommen, C.J. van der Veen, and J. Wen, 2006b: Insignificant change in Antarctic snowfall since the International Geophysical Year. *Science*, **313**, 827-831.

Monnin E, Indermühle A, Dällenbach A, Flückiger J, Stauffer B, Stocker T, Raynaud D, Barnola JM (2001) Atmospheric CO_2 concentrations over the last glacial termination from the Dome Concordia, Antarctica, ice core. Science 291:112-114

Morales Maqueda, M. A., Willmott, A. J. and Biggs, N. R. T. (2004). Polynya dynamics: A review of observations and modeling. *Rev. Geophys.* **42**, doi:10.1029/2002RG000116.

Morgan V, Delmotte M, van Ommen T, Jouzel J, Chappellaz J, Woon S, Masson-Delmotte V, Raynaud D (2002) Relative timing of Deglacial Climate events in Antarctica and Greenland. Science 297:1862-1864

Moritz RE, Bitz CM, Steig EJ (2002) Dynamics of recent climate change in the Arctic. Science 297: 1497–1502

Morley S., pers. Comm. – chapter 4

Morley SA, Peck LS, Miller AJ, Pörtner HO (2007) Hypoxia tolerance associated with activity reduction is a key adaptation for *Laternula elliptica* seasonal energetics. Oecologia 153: 29–36

Morris WF, Doak DF (2004) Buffering of life histories against environmental stochasticity: accounting for a spurious correlation between the variabilities of vital rates and their contributions to fitness. American Naturalist, 163: 579-590

Morris, E. M. and R. Mulvaney (2004), Recent variations in surface mass balance of the Antarctic Peninsula Ice Sheet, *J. Glaciol., 50,* 257-267.

Morris, E. M., and D. G. Vaughan (2003), Spatial and temporal variation of surface temperature on the Antarctic Peninsula and the limit of viability of ice shelves, *Antarctic Peninsula Climate Variability: Historical and Paleoenvironmental Perspectives. Antarctic Research Series, 79*, edited by E. Domack, et al., pp. 61-68, AGU, Washington, DC.

Morrow et al., 2008 – chapter 3

Morrow R., J. Church, R. Coleman, D. Chelton, and N. White, 1992. Eddy momentum flux and its contribution to the Southern Ocean momentum balance. *Nature*, **357**, 482–484.

Mosley-Thompson, E., J. F. Paskievitch, A. J. Gow, L. G. Thompson, 1999: Late 20th century increase in South Pole snow accumulation, *J. Geophys. Res.,* **104**(D4), 3877-3886, 10.1029/1998JD200092.

Mosola AB, Anderson JB (2006) Expansion and rapid retreat of the West Antarctic Ice Sheet in Eastern Ross Sea: possible consequence of over extended ice streams? Quaternary Science Reviews 25:2177-2196

Muerk *et al.* 1994 – chap 1

Murphy E J, Trathan P N, Watkins J L, Reid K, Meredith M P, Forcada J, Thorpe S E, Jonston N M and Rothery P (2007) Climatically driven fluctuations in Southern Ocean ecosystems. Proceedings of the Royal Society B, doi:10.1098/rspb.2007.1180.

Murphy E J, Watkins J L, Reid K, Trathan P N, Everson I, Croxall J P, Priddle J, Brandon M A, Brierley A S and Hofmann E (1998) Interannual variability of the South Georgia marine ecosystem: biological and physical sources of variation in the abundance of krill. Fisheries Oceanography, 7, 381-390.

Murphy E. J. et al. 2006. Spatial and temporal operation of the Scotia Sea ecosystem: a review of large-scale links in a krill centred food web. Philosophical Transactions of the Royal Society B: Biological Sciences, 362, (1477): 113 148. DOI: 10.1098/rstb.2006.1957

Naish TR, R.D. Powell, S. Henrys, G.S. Wilson, L.A. Krissek, F. Niessen, M. Pompilio, R. Scherer, F. Talarico, R.H. Levy, Pyne A (2007) New insights into Late Neogene (13-0Ma) climate history of Antarctica: Initial results from the ANDRILL McMurdo Ice Shelf (MIS) Project (AND-1B). In: A.K. Cooper, C.R. Raymond, al e (eds) *A Keystone in a Changing World - Online Proceedings of the 10th ISAES* USGS Open-File Report 2007-xxx, yyy, p 1-11

Naish TR, Woolfe KJ, Barrett PJ, Wilson GS, Atkins C, Bohaty SM, Bucker CJ, Claps M, Davey FJ, Dunbar GB, Dunn AG, Fielding CR, Florindo F, Hannah MJ, Harwood DM, Henrys SA, Krissek LA, Lavelle M, van der Meer J, McIntosh WC, Niessen F, Passchier S, Powell RD, Roberts AP, Sagnotti L, Scherer RP, Strong CP, Talarico F, Verosub KL, Villa G, Watkins DK, Webb P-N, Wonik T (2001) Orbitally induced oscillations in the East Antarctic ice sheet at the Oligocene/Miocene boundary. Nature 413:719-723

Nakada M, Kimura R, Okuno J, Moriwaki K, Miura H, Maemoku H (2000) Late Pleistocene and Holocene melting history of the Antarctic ice sheet derived from sea-level variations. Marine Geology 167:85-103

Nakada M, Lambeck K (1988) The melting history of the Late Pleistocene Antarctic ice-sheet. Nature 333:36-40

Nakicenovic et al., 2000 – Chapter 4

Naud, Catherine; Muller, Jan-Peter; de Valk, Paul. 2005. On the use of ICESAT-GLAS measurements for MODIS and SEVIRI cloud-top height accuracy assessment. Geophys. Res. Lett., Vol. 32, No. 19, L19815, 10.1029/2005GL023275, 13 October 2005

Naveira Garabato, A. C., Polzin, K. L., King, B. A., Heywood, K. J. and Visbeck, M. (2004). Widespread intense turbulent mixing in the Southern Ocean. *Science* **303**, 210-213.

Naveira Garabato, A. C., Stevens, D. P., Watson, A. J. and Roether, W. (2007). Short-circuiting of the overturning circulation in the Antarctic Circumpolar Current. *Nature* **447**, 194-197.

Naveira Garabato, A.C., E.L. McDonagh, D.P. Stevens, K.J. Heywood and R.J. Sanders (2002) On the export of Antarctic Bottom Water from the Weddell Sea, Deep-Sea Res. II 49, 4715-4742.

Near T (2004) Estimating divergence times of notothenioid fishes using a fossil-calibrated molecular clock. Antarctic Sci 16: 37-44

Nelson, FE., Outcalt, SI., Brown, J, Shiklomanov, NI., Hinkel, KM. 1998. Spatial temporal attributes of the active-layer thickness record, Barrow, Alaska, USA. In: Lewkowicz, AG., Allard M. (eds.) *Permafrost, Seventh International Conference, June 23-27, 1998, Proceedings.*, Université Laval , Nordicana 57**:** 797-802.

Nerem R.S., É. Leuliette, A. Cazenave, 2006, Present-day sea-level change: A review, *C. R. Geoscience,* **338**, 1077–1083.

Newman, P.A., E.R. Nash, S.R. Kawa, S.A. Montzka, S.M. Schauffler, “When will the Antarctic ozone hole recover?”, *Geophys. Res. Lett.*, **33**, L12814 (2006).

Nicholls K.W., C.J. Pudsey and P. Morris, 2004: Summertime water masses off the northern Larsen C Ice Shelf, Antarctica. *Geophys. Res. Lett.*, **31**, L09309, doi:10.1029/2004GL019924.

Nicholls review paper – chapter 3 – page 139

Nicholls, K. W., and A. Jenkins, 1993, Temperature and salinity beneath Ronne Ice Shelf, Antarctica. *J. Geophys. Res.,* **98,** 22,553–22,568.

Nicholls, K. W., Makinson, K. and Osterhus, S. (2004). Circulation and water masses beneath the northern Ronne Ice Shelf, Antarctica. *J. Geophys. Res.* **109**, doi:10.1029/2004JC002302.

Nicholls, K.W., 1997: Predicted reduction in basal melt rates of an Antarctic ice shelf in a warmer climate. *Nature* **388**(6641), 460-462.

Nichols and Balls, 1974 – chapter 3

Nichols, R.L. and Ball, D.G. 1964. Soil temperatures, Marble point, McMurdo Sound, Antarctica. *Journal of Glaciology*, 5, 357-359.

Nicol et al. 2007 Antarctic Science – chapter 3 – page 172

Nielsen SHH, Koc N, Crosta X (2004) Holocene climate in the Atlantic sector of the Southern Ocean: controlled by insolation or oceanic circulation? . Geology 32:317-320

Nightingale, P. D., et al. (2000), In-situ evaluation of air-sea gas exchange parameterizations using novel conservative and volatile tracers, *Global Biogeochemical Cycles*, *14*, 313-387.

Nikolaiev VI, Kotlyakov VM, Smirnov KE (1988) Isotopic studies of the ice core from the Komsomolskaia station, Antarctica. Data of Glaciological Studies of the USSR Academy of Sciences 63:97--102

Nitsche F O, Jacobs S S, Larter R D, Gohl K (2007) Bathymetry of the Amundsen Sea continental shelf: Implications for geology, oceanography, and glaciology, Geochem. Geophys. Geosyst., 8, Q10009, doi:10.1029/2007GC001694.

Noone and Simmonds (2002) - chapter 4 -page 15

Noone, D. and I. Simmonds, 2002: Annular variations in moisture transport mechanisms and the abundance of delta18O in Antarctic snow. *J. Geophys. Res.*, **107**, 4742, doi:10.1029/2002JD002262.

NorthGRIP-community-members (2004) High resolution climate record of the northern hemisphere reaching into last interglacial period. Nature 431:147-151

Nøst, O.A. and S. Østerhus, 1998: Impact of grounded icebergs on the hydrographic conditions near the Filchner Ice Front, *Antarct. Res. Ser.* **75**, 267-284.

Nye, J.F. (1957), The distribution of stress and velocity in glaciers and ice sheets. *Proceedings of the Royal Society of London Series A 275*, 87-112.

Ó Cofaigh C, Dowdeswell JA, Allen CS, Hiemstra JF, Pudsey CJ, Evans J, Evans DJA (2005) Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream. Quaternary Science Reviews 24:709-740

Ó Cofaigh C, Pudsey CJ, Dowdeswell JA, Morris P (2002) Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophysical Research Letters 29:1199

O'Brien PE (1994) Morphology and late glacial history of Prydz Bay, Antarctica, based on echo sounder data. Terra Antarctica 1:403-405

O'Brien PE, Harris PT (1996) Patterns of glacial erosion and deposition in Prydz Bay and the past behavior of the Lambert Glacier. Papers and Proceedings of the Royal Society of Tasmania 130:79-85

O'Brien PE, Leitchenkogv G (1997) Deglaciation of Prydz Bay, East Antarctica based on echo sounder and topographic features. Antarctic Research Series 71:109-126

O'Dowd, C.D., Jmenez, J.L., Bahreini, R., Flagan, R.C., Seinfeld, J.H., Hämeri, Kaarle, Pirjola, L., Kulmala, M., Jennings, S.G., and Hoffmann, T., Marine aerosol formation from biogenic iodine emissions, Nature (417), 632-636, 2002.

Oechel WC, Vourlitis GL, Hastings SJ, Zulueta RC, Hinzman L, Kane D (2000) Acclimation of ecosystem CO_2 exchange in the Alaskan Arctic in response to decadal climate warming. Nature 406: 978–981

Oerlemans, J. (1982a), A model of the Antarctic Ice Sheet. *Nature 297*, 550-553.

Oerlemans, J. (1982b), Response of the Antarctic Ice Sheet to a climatic warming: a model study. *Journal of Climatology 2*, 1-11.

Ohmura, A., M.A. Wild, and L. Bengtsson, 1996: A possible change in mass balance of Greenland and Antarctic ice sheets in the coming century. *J. Climate* **9**, 2124-2135.

Oke and England, 2004 – chapter 3

Oke, T. R., 1987. Boundary layer of climate, Mehuen and C Ed., London.

Okuno J, Miura H, Maemoku H (2007) The possibility of the rapid melting of ice sheet in Holocene around the Lutzow-Holm Bay, Antarctica. Geophysical Research Abstracts 7

Olbers, D., Borowski, D., Völker, C. and Wolff, J.-O. (2004). The dynamical balance, transport and circulation of the Antarctic Circumpolar Current. *Antarct. Sci.* **16**, 439–470.

Orejas C, Gili J-M, Arntz W (2003) Role of small-plankton communities in the diet of two Antarctic octocorals (*Primnoisis antarctica* and *Primnoella* sp.). Mar Ecol Prog Ser 250: 105-116

Orr, A., D. Cresswell, G. J. Marshall, J. C. R. Hunt, J. Sommeria, C. G. Wang, M. Light, 2004: A 'low-level' explanation for the recent large warming trend over the western Antarctic Peninsula involving blocked winds and changes in zonal circulation. *Geophys. Res. Lett.*, **31**, L06204, doi:10.1029/2003GL019160.

Orr, J. C., et al. (2001), Estimates of anthropogenic carbon uptake from four three-dimensional global ocean models, *Global Biogeochemical Cycles*, *15*, 43-60.

Orr, J. C., et al. (2005), Anthropogenic ocean acidification over the twenty-first century and its impact on calcifying organisms, *Nature*, *437*, 681-686.

Orsi, A. H., T.W. Whitworth III, W. D. Nowlin Jr., 1995. On the meridional extent and fronts of the Antarctic Circumpolar Current. *Deep-Sea Res*., 42: 641-673.

Orsi, A.H., G. C. Johnson and J. B. Bullister (1999). Circulation, mixing and production of Antarctic Bottom Water. *Prog. Oceanog*. 43, 55-109.

Pacanowski R.C. and Philander, S.G.H., 1981: Paramerization of vertical mixing in numerical models of tropical oceans. Journal of Physical Oceanography, 11, 1443-1451.

Pagani M, Zachos JC, Freeman KH, Tipple B, Bohaty S (2005) Marked Decline in Atmospheric Carbon Dioxide Concentrations During the Paleogene. Science 309:600-603

Papetti C, Liò P, Rüber L, Patarnello T & Zardoya R (2007) Antarctic fish mitochondrial genomes lack *ND6* gene. J Mol Evol 65: 519-528

Park, M., W.J. Randel, D.E. Kinnison, R.R. Garcia, W. Choi, "Seasonal variation of methane, water vapor, and nitrogen oxides near the tropopause: Satellite observations and model simulations", *J. Geophys. Res.*, **109**, D03302 (2004).

Park, Y., F. Roquet, and F. Vivier, 2004: Quasi-stationary ENSO wave signals versus the Antarctic Circumpolar Wave scenario. *Geophys. Res. Lett*., **31**, L09315, doi:10.1029/2004GL019806.

Parkinson C L (2004) Southern Ocean sea ice and its wider linkages: insights revealed from models and observations. Antarctic Science, 16, 387–400.

Parkinson et al – chapter 4 - which one 2004 or 2006

Parkinson, C. L., K. Y. Vinnikov, and D. J. Cavalieri (2006), Evaluation of the simulation of the annual cycle of Arctic and Antarctic sea ice coverages by 11 major global climate models, J. Geophys. Res., 111, doi:10.1029/2005JC003408.

Parmesan C, Yohe G (2003) A globally coherent fingerprint of climate change impacts across natural systems. Nature 421: 37-42

Patarnello et al, 2003 – Chapter 2

Patterson S. L., 1985. Surface circulation and kinetic energy distributions in the southern hemisphere oceans from FGGE drifting buoys. *J. Phys. Oceanogr.*, **15**, 865–884.

Payne A J, Holland P R, Shepherd A P, Rutt I C, Jenkins A, Joughin I (2007) Numerical modeling of ocean-ice interactions under Pine Island Bay's ice shelf, J. Geophys. Res., 112, doi:10.1029/2006JC003733.

Payne, A.J. (1998), Dynamics of the Siple Coast ice streams, West Antarctica: results from a thermomechanical ice sheet model. *Geophysical Research Letters 25*, 3173-3176.

Payne, A.J., A. Vieli, A.P. Shepherd, D.J. Wingham, and E. Rignot (2004), Recent dramatic thinning of largest West Antarctic ice stream triggered by oceans. *Geophysical Research Letters 31*, L23401, doi: 10.1029/2004GL02184.

Payne, A.J., P. Sammonds, and J.C.R. Hunt, 2006, Evolution of the Antarctic ice sheet: new understanding and new challenges, *Philosophical Transactions of the Royal Society A* 364, 1844.

Payne, A.J., P.R. Holland, A.P. Shepherd, I.C. Rutt, A. Jenkins and I. Joughin, 2007: Numerical modelling of ocean-ice interactions under Pine Island Bay's ice shelf, *J. Geophys. Res.*, in press.

Pearson PN, Palmer MR (2000) Atmospheric carbon dioxide concentrations over the past 60 million years. Nature 406:695-699

Peat et al. 2007 – chapter 3

Peck et al. 2004 - Pörtner et al. 1999a - chapter 4 - page 66

Peck et al. 2006a - chapter 4

Peck et al. 2006b - chapter 4

Peck L. pers. Obs - Chapter 4 – page 73

Peck LS (1989) Temperature and basal metabolism in two Antarctic marine herbivores. J Exp Mar Biol Ecol 127: 1-12

Peck LS (1998) Feeding, metabolism and metabolic scope in Antarctic ectotherms. Soc Exp Biol Seminar Ser 66: 365-389

Peck LS (2005) Prospects for survival in the Southern Ocean: vulnerability of benthic species to temperature change. Antarctic Sci 17: 497-507

Peck LS (2005) Prospects for surviving climate change in Antarctic aquatic species. Frontiers in Zoology 2: 9

Peck LS, Ansell AD, Webb KE, Hepburn L, Burrows M (2004) Burrowing in Antarctic bivalve molluscs. Polar Biol 27: 357-367

Peck LS, Brockington S, Vanhove S, Beghyn M (1999) Community recovery following catastrophic iceberg impacts in a soft-sediment shallow-water site at Signy Island, Antarctica. Mar Ecol Prog Ser 186: 1-8

Peck LS, Convey P, Barnes DKA (2006) Environmental constraints on life histories in Antarctic ecosystems: tempos, timings and predictability. Biol Rev 81: 75-109

Peck LS, Conway LZ (2000) The myth of metabolic cold adaptation: oxygen consumption in stenothermal Antarctic bivalves. In Harper E, Tatlor JD, Crame JA (eds). Evolutionary biology of the bivalvia, vol 177. Geol Soc Special Publications, London, pp 441-450

Peck LS, Morley SA, P**ö**rtner HO, Clark MS (2007) Thermal limits of burrowing capacity are linked to oxygen availability and size in the Antarctic clam *Laternula elliptica*. Oecologia. Published on line DOI 10.1007/s00442-007-0858-0

Peck LS, Pörtner HO, Hardewig I (2002) Metabolic demand, oxygen supply, and critical temperatures in the Antarctic bivalve *Laternula elliptica*. Physiol Biochem Zool 75:123-133

Peck LS, Webb K, Bailey D (2004) Extreme sensitivity of biological function to temperature in Antarctic marine species. Funct Ecol 18: 625-630

Peck LS, Webb KE, Miller A, Clark MS, Hill, T (in press). Temperature limits to activity, feeding and metabolism in the Antarctic starfish *Odontaster validus*. Mar Ecol Prog Ser

Peinert R, Bodungen B von, Smetacek VS (1989) Food web structure and loss rate. In Productivity in the ocean: present and past (eds: Berger WH, Smetacek VS, Wefer G) Wiley-Interscience, Chichester, pp 35-48

Peltier WR (2002) Comments on the paper of Yokoyama et al. (2000), entitled "Timing of the Last Glacial Maximum from observed sea level minima". Quaternary Science Reviews 21:409-414

Peltier WR (2005) On the hemispheric origins of meltwater pulse 1a. Quaternary Science Reviews 24:1655-1671

Penduff, 2007 – chapter 3 – Penduff et al 2007?

Penduff, T., Le Sommer, J., Barnier, B., Treguier, A.-M., Molines, J.-M., & Madec, G., 2007: Influence of numerical schemes on current-topography interactions in 1/4° global ocean simulations. Ocean Science, 3, 509-524..

Pereira, A. F., Beckmann, A. and Hellmer, H. H. (2002). Tidal mixing in the southern Weddell Sea: Results from a three-dimensional model. *J. Phys. Oceanogr*. **32**, 2151-2170.

Peterson, R.G., White, W.B., 1998. Slow oceanic teleconnections linking the Antarctic Circumpolar Wave with tropical ENSO. Journal of Geophysical Research 103, 24573-24583.

Petit et al. 1999b – Chapter 2

Petit JR, Basile I, Leruyuet A, Raynaud D, Lorius C, Jouzel J, Stievenard M, Lipenkov VY, Barkov NI, Kudryashov BB, Davis M, Saltzman E, Kotlyakov V (1997) Four climatic cycles in Vostok ice core. Nature 387:359

Petit JR, Jouzel J, Raynaud D, Barkov NI, Barnola JM, Basile I, Bender M, Chappellaz J, Davis J, Delaygue G, Delmotte M, Kotyakov VM, Legrand M, Lipenkov VY, Lorius C, Pépin L, Ritz C, Saltzman E, Stievenard M (1999a) Climate and Atmospheric History of the Past 420000 years from the Vostok Ice Core, Antarctica. Nature 399:429-436

Petoukhov V., M. Claussen, A. Berger, M. Crucifix, M. Eby, A. Eliseev, T. Fichefet, A. Ganopolski, H. Goosse, I. Kamenkovich, I. Mokhov, M. Montoya, L.A. Mysak, A. Sokolov, P. Stone, Z. Wang and A. J. Weaver (2005). EMIC Inter-comparison Project (EMIP-CO2): Comparative analysis of EMIC simulations of climate, and of equilibrium and transient responses to atmospheric CO2 doubling. Climate Dynamics 25 363-385.

Peylin, P., et al. (2005), Multiple constrains on regional CO2 flux variations over land and oceans, *Global Biogeochemical Cycles*, *19*, doi:10.1029/2003GB002214.

Phillips, H. E. and S. R. Rintoul, 2000. Eddy variability and energetics from direct current measurements in the Antarctic Circumpolar Current south of Australia. Journal of Physical Oceanography, 30, 3050-3076.

Picard, G., and M. Fily, 2006. Surface melting observations in Antarctica by microwave radiometers: correcting 26-year record from changes in acquisition times. Remote Sensing of Environment, 104(3), 325-336, doi:10.1016/j.rse.2006.05.010.

Pickard and Seppelt 1984 – Chapter 3

Pickard J, Adamson DA, Harwood DM, Miller GH, Quilty PG, Dell RK (1986) Early Pliocene marine sediments in the Vestfold Hills, East Antarctica: implications for coastline, ice sheet and climate. South African Journal of Science 82:520-521

Pisano E, Cocca E, Mazzei F, Ghigliotti L, di Prisco G, Detrich HW III & Ozouf-Costaz C (2003) Mapping of alpha- and beta-globin on Antarctic fish chromosomes by fluorescence in situ hybridisation. Chromosome Res 11: 633-640

Pisano E, Ozouf-Costaz C & Prirodina V (1998) Chromosome diversification in Antarctic fish (Notothenioidei). In: di Prisco G, Pisano E & Clarke A (Eds) Fishes of Antarctica. A Biological Overview (pp 275-285). Springer, Milano

Pitari, G., E. Mancini, V. Rizi, D.T. Shindell, "Impact of future climate change and emission changes on stratospheric aerosols and ozone", *J. Atmos. Sci.*, **59**, 414-440 (2002).

Planchon FAM, Boutron CF, Barbante C, Cozzi G, Gaspari V, Wolff EW, Ferrari CP, Cescon P (2002) Changes in heavy metals in Antarctic snow from Coats Land since mid-19th to the late-20th century. Earth Planet Sci Lett 200:207-222

Plattner, G. K., et al. (2001), Feedback Mechanisms and sensitivities of ocean carbon uptake under global warming, *Tellus B*, *53*, 564-592.

Podrabsky JE, Somero GN (2006) Inducible heat tolerance in Antarctic notothenioid fishes. Pol Biol 30(1): 39-43

Polzin, K. L. (2007). How Rossby Waves Break. Results from POLYMODE and the end of the enstrophy cascade. *J. Phys. Oceanogr.*, in press.

Popova, E.E., Coward, A.C., Nurser, G.A., de Cuevas, B., Fasham, M.J.R., and Anderson, T.R., 2006: Mechanisms controlling primary and new production in a global ecosystem model – Part I: Validation of the biological simulation. Ocean Science, 2, 249-266.

Pörtner et al. 1999a - chapter 4 missing page 65

Pörtner et al. 1999b - chapter 4

Pörtner HO (2001) Climate change and temperature-dependent biogeography: oxygen limitation of thermal tolerance in animals. Naturwiss 88: 137-146

Pörtner HO (2002) Climate variations and the physiological basis of temperature dependent biogeography: systemic to molecular hierarchy of thermal tolerance in animals. Comp Biochem Physiol *A* 132: 739-761

Pörtner HO (2006) Climate dependent evolution of Antarctic ectotherms: an integrative analysis. Deep Sea Res II 53: 1071–1104

Pörtner HO, Berdal B, Blus, R, Brix O, Colosimo A, De Wachter B, Giuliani A, Johansen T, Fischer T, Knust R, Naevdal G, Nedenes A, Nyhammer G, Sartoris FJ, Serendero I, Sirabella P, Thorkildsen S, Zakhartsev M (2001) Climate effects on growth performance, fecundity and recruitment in marine fish: developing a hypothesis for cause and effect relationships in Atlantic cod (*Gadus morhua*) and common eelpout (*Zoarces viviparus*). Continental Shelf Res 21: 1975-1997

Pörtner HO, Hardewig I, Peck LS (1999) Mitochondrial function and critical temperature in the Antarctic bivalve, *Laternula elliptica*. Comp Biochem Physiol 124A: 179-189

Pörtner HO, Hardewig I, Sartorius FJ, van Dijk PLM (1998) Energetic aspects of cold adaptation: critical temperatures in metabolic, ionic and acid-base regulation? In Pörtner HO, Playle R (eds) Cold Ocean Physiology. Cambridge University Press, Cambridge, pp 88-120

Pörtner HO, Knust R (2007) Climate change affects marine fishes through the oxygen limitation of thermal tolerance. Science 315: 95 - 97

Pörtner HO, Mark FC, Bock C (2004) Oxygen limited thermal tolerance in fish? Answers obtained by Nuclear Magnetic Resonance techniques. Resp Physiol Neurobiol 141: 243–260

Pörtner HO, Peck LS, Hirse T (2006) Hyperoxia alleviates thermal stress in the Antarctic bivalve, *Laternula elliptica*: evidence for oxygen limited thermal tolerance. Polar Biol 29: 688-693

Pörtner HO, Peck LS, Somero GN (2007) Thermal limits and adaptation: an integrative view (Antarctic Ecology: From Genes to Ecosystems). Phil Trans R Soc B 362: 2233–2258

Pörtner H-O, Peck LS, Zielinski S, Conway LZ (1999) Intracellular pH and energy metabolism in the highly stenothermal Antarctic bivalve *Limopsis marionensis* as a function of ambient temperature. Polar Biol 22: 17-30

Pörtner HO, Somero GN, Peck LS (2007) Thermal limits and adaptation: an integrated view. Phil Trans R Soc Lond B (in press)

Pörtner HO, Van Dijk PLM, Hardewig I, Sommer A (2000) Levels of metabolic cold adaptation: tradeoffs in eurythermal and stenothermal ectotherms. In: Davison W, Williams CW (Eds) Antarctic Ecosystems: Models for a Wider Understanding. Caxton Press, Christchurch, New Zealand, pp. 109-122

Post LA, Hemer MA, O'Brien PE, Roberts D, Craven M (2007) History of benthic colonization beneath the Amery Ice Shelf, East Antarctica. Mar Ecol Prog Ser 344: 29-37

Poulin E, Palma AT, Feral J-P (2002) Evolutionary versus ecological success in Antarctic benthic invertebrates. Trends Ecol Evol 17: 218-222

Powell DK, Tyler PA, Peck LS (2001). Effect of sperm concentration and sperm ageing on fertilization success in the Antarctic soft-shelled clam *Laternula elliptica* and the Antarctic limpet *Nacella concinna*. Mar Ecol Prog Ser 215: 191-200

Price and Conway refs – chapter 3 (page 116)

Price, J.F., Weller, R.A., & Oinkel, R., 1986: Duirnal cycling: Observations and models of the upper ocean response to diurnal heating, cooling and wind mixing. Journal of Geophysical Research, 91, 8411-8427.

Pritchard, H. D. and D. G. Vaughan, 2007, Widespread acceleration of tidewater glaciers on the Antarctic Peninsula, *J. Geophys. Res.*, **112**, F03S29, doi:10.1029/2006JF000597.

Pudsey CJ, Evans J (2001) First survey of Antarctic sub-ice shelf sediments reveals mid-Holocene ice shelf retreat. Geology 29:787-790

Pugh (1993) – chap 1

Pugh and Scott (2002) – chap 1

Pugh PJA, Convey P (In press) Surviving out in the cold: Antarctic endemic invertebrates and their refugia. Journal of Biogeography

Pugh PJA, Dartnall HJG, McInnes SJ (2002) The non-marine crustacea of Antarctica and the islands of the Southern Ocean: biodiversity and biogeography. Journal of Natural History 36:1047-1103

Quayle et al. 2003 – chapter 3 & 4

Quayle WC, Peck LS, Peat H, Ellis-Evans JC, Harrigan PR (2002) Extreme responses to climate change in Antarctic lakes. Science 295:645-645

Quesada et al. 2006 RiSCC – chapter 3 & 4

Quetin L B and Ross R M (2003) Episodic recruitment in Antarctic krill *Euphausia superba* in the Palmer LTER study region. Marine Ecology Progress Series, 259, 185–200.

Quinn and Neal, 1992 – Chapter 2

Quinn et al., 1987 – Chapter 2

Rahmstorf, S., Cazenave, A., Church, J.A., Hansen, J.E., Keeling, R., Parker, D.E. and Somerville, R.C.J. 2007. Recent climate observations compared to projections. Science, 316, 709, doi:10.1126/science.1136843.

Räisänen, J., 2007: How reliable are climate models? *Tellus*, **59A,** 2-29.

Ramillien G., A. Lombard, A. Cazenave, E.R. Ivins, M. Llubes, F. Remy, R. Biancale, 2006, Interannual variations of the mass balance of the Antarctica and Greenland ice sheets from GRACE, Global and Planetary Change **53,** 198–208.

Ramos, M. and G. Vieira. 2003. Active layer and permafrost monitoring in Livingston Island, Antarctic. First results from 2000 to 2001. In: Phillips, M., S.M. Springman, and L. Arenson (eds.) Proceedings of the 8th International Conference on Permaforst, Zurich, Switzerland, 21-25 July 2003. Balkema Publishers, Lisse, pp. 929-933.

Randall, D.A., R.A. Wood, S. Bony, R. Colman, T. Fichefet, J. Fyfe, V. Kattsov, A. Pitman, J. Shukla, J. Srinivasan, R.J. Stouffer, A. Sumi and K.E. Taylor, 2007: Cilmate Models and Their Evaluation. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M.Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Rankin, A.M., Wolff, E.W., and Martin, S.: Frost flowers: Implications for tropospheric chemistry and ice core interpretation, J. Geophys., Res., 107, 4683, doi:10.1029/2002JD002492, 2002.

Raper, S. C., Wigley, T. M., Jones, P. D. and Salinger, M. J. 1984. Variations in surface air temperatures: Part 3. The Antarctic, 1957-1982. *Monthly Weather Review* **112**: 1341-1353.

Raphael (2003) – chapter 3

Raphael, M.N., and M.M. Holland, 2006: Twentieth century simulation of the Southern Hemisphere climate in coupled models. Part 1: Large scale circulation variability. *Clim. Dyn.*, **26**, 217–228, doi:10.1007/s00382-005-0082-8.

Rasmussen SO, Andersen KK, Svensson AM, Steffensen JP, Vinther BM, Clausen HB, Siggaard-Andersen M-L, Johnsen SJ, Larsen LB, Dahl-Jensen D, Bigler M, Röthlisberger R, Fischer H, Goto-Azuma K, Hansson ME, Ruth U (2006) A new Greenland ice core chronology for the last glacial termination. J. Geophys. Res.:doi:10.1029/2005JD006079

Rau, F., and M. Braun (2002), The regional distribution of the dry-snow zone on the Antarctic Peninsula north of 70° S, *Annals of Glaciology, 34*, 95-100.

Raupach, M., et al. (2007), Global and regional drivers of accelerating CO2 emissions, *PNAS*.

Ravelo AC, Andreasen DH, Lyle M, Olivarez Lyle A, Wara MW (2004) Regional climate shifts caused by gradual global cooling in the Pliocene epoch. Nature 429:263-267

Raymond, C. (1996), Shear margins in glaciers and ice sheets. *Journal of Glaciology 42*, 90-102.

Raymond, C.F., K.A. Echelmeyer, I.M. Whillans, and C.S.M. Doake (2001), Ice stream shear margins. In: *The West Antarctic Ice Sheet: Behavior and environment (R.B. Alley and R.A. Bindschadler, eds.).* Washing ton DC: American Geophysical Union, Antarctic Research Series 77, 137-155.

Raynaud D, Blunier T, Ono Y, Delmas RJ (2003) The Late Quaternary History of Atmospheric Trace Gases and Aerosols: Interactions Between Climate and Biogeochemical Cycles. In: Alverson K, Bradley R, Pedersen T (eds) Paleoclimate, Global Change and the Future. Springer-Verlag, p 13-35

Read, K. A., Lewis, A.C., Salmon, R.A., Bauguitte, S., Rankin, A.M., and Wolff, E.W., DMS and MSA measurements in the Antarctic Boundary Layer: Impact of BrO on MSA production, in prep.

Reimer PJ, Baillie MGL, Bard E, Bayliss A, Beck JW, Bertrand CJH, Blackwell PG, Buck CE, Burr GS, Cutler KB, Damon PE, Edwards RL, Fairbanks RG, Friedrich M, Guilderson TP, Hogg AG, Hughen KA, Kromer B, Mccormac G, Manning S, Ramsey CB, Reimer RW, Remmele S, Southon JR, Stuiver M, Talamo S, Taylor FW, Van Der Plicht J, Weyhenmeyer CE (2004) Intcal04 Terrestrial Radiocarbon Age Calibration, 0-26 Cal Kyr Bp. Radiocarbon 46:1029-1058

Renfrew, I. A., King, J. C. and Markus, T. (2002). Coastal polynyas in the southern Weddell Sea: Variability of the surface energy budget. *J. Geophys. Res.* **107**, doi:10.1029/2000JC000720.

Renssen et al. 2005b – Chapter 2 (2005a is there but not 2005b)

Renssen H, Goosse H, Fichefet T, Masson-Delmotte V, Koc N (2005a) Holocene climate evolution in the high-latitude Southern Hemisphere simulated by a coupled atmosphere-sea ice-ocean-vegetation model. The Holocene 15:951-964

Renwick and Revell, 1999 – chapter 3

Renwick, 2004 – chapter 3

Retzlaff, R. and C.R. Bentley, 1993. Timing of stagnation of Ice Stream C, West Antarctica, from short-pulse radar studies of buried surface crevasses, *Journal of Glaciology* ,Vol. 39, No. 133, 553-561.

Revelle, R., and H. E. Suess (1957), Carbon dioxide exchange between atmosphere and ocean and the question of an increase of atmospheric CO2 during the past decades, *Tellus*, *9*, 18-27.

Rhines, P.B., 1975. Waves and turbulence on a beta-plane. *J. Fluid Mech*, **69**:417-443.

Richardson AJ, Schoeman DS (2004) Climate impact on ecosystems in the northeast Atlantic. Science 305: 1609-1612

Richardson-Naslund C., 2004, Spatial characteristics of snow accumulation in Dronning Maud Land, Antarctica *Global and Planetary Change*, **42** (1 4), 31-43.

Riddle MJ, Craven M, Goldsworthy PM, Carsey F (2007) A diverse benthic assemblage 100 km from open water under the Amery Ice Shelf, Antarctic. Palaeoceanography 22, PA1204; DIO: 10.1029/2006PA001327

Ridley, J. 1993. Surface melting on Antarctic peninsula ice shelves detected by passive microwave sensors. *Geophys. Res. Lett.,* **20**(23), 2639–2642.

Riebesell, U., et al. (2000), Reduced calcification of marine plankton in response to increased atmospheric CO2, *Nature*, *407*, 364-367.

Rignot E & Jacobs S S (2002) Rapid bottom melting widespread near Antarctic ice sheet grounding lines, Science, 296, 2020-2023.

Rignot E. and R.H. Thomas, 2002. Mass balance of polar ice sheets, *Science*, 297 (5586): 1502-1506 AUG 30 2002

Rignot E., D. G. Vaughan, M. Schmeltz, T. Dupont and D. MacAyeal, 2002, Acceleration of Pine island and Thwaites Glaciers, West Antarctica. *Ann. Glaciol.* **34**, 189–194.

Rignot et al. (1998) – chapter 3

Rignot et al., 2005 – Chapter 4

Rignot, 2008 – chapter 3 – 2008b?

Rignot, E. 1998. Radar Interferometry Detection of Hinge Line Migration on. Rutford Ice Stream and Carkon Inlet, Antarctica. Annals of Glaciology, 27, 25-32.

Rignot, E. and S. S. Jacobs, (2002), Rapid bottom melting widespread near Antarctic ice sheet grounding lines. *Science* 296, 2020-2023.

Rignot, E., 1998, Fast recession of a West Antarctic Glacier. *Science* **281**, 549–551, (doi:10.1126/science.281.5376.549).

Rignot, E., 2008b in press. Changes in West Antarctic ice stream dynamics observed with ALOS PALSAR data. Geophysical Research Letters, in press.

Rignot, E., G. Casassa, P. Gogineni, P. Kanagaratnam,W. Krabill, H. Pritchard, A. Rivera, R. Thomas, J. Turner and D. Vaughan (2005), Recent ice loss from the Fleming and other glaciers,Wordie Bay, West Antarctic Peninsula, *Geophys. Res. Let.*, *32*, doi:10.1029/2004GL021947.

Rignot, E., G. Casassa, P. Gogineni, W. Krabill, A. Rivera and R. Thomas (2004), Accelerated ice discharge from the Antarctic Peninsula following the collapse of Larsen B ice shelf, *Geophys. Res. Let.*, *31*, doi:10.1029/2004GL020679.

Rignot, E., J.L. Bamber, M.R. van den Broeke, C. Davis, L. Yonghong, W.J. van deBerg and E. van Meijgaard, 2008. Recent Antarctic ice mass loss from radar interferometry and regional climate modeling, *Nature Geoscience*, 13 January 2008; doi:10.1038/ngeo102

Rintoul, S. R. and S. Sokolov, 2001. Baroclinic transport variability of the Antarctic Circumpolar Current south of Australia (WOCE repeat section SR3). Journal of Geophysical Research, 106, 2795-2814.

Rintoul, S. R., 1998. On the origin and influence of Adelie Land Bottom Water. In: Ocean, Ice and Atmosphere: Interactions at the Antarctic Continental Margin, S. Jacobs and R. Weiss (eds.), Antarctic Research Series, 75, 151-171, American Geophysical Union, Washington.

Rintoul, S. R., Hughes, C. W. and Olbers, D. (2001). The Antarctic Circumpolar Current system. In: Eds. G. Siedler, J. Church & J. Gould, *Ocean circulation and climate; observing and modelling the global ocean*. International Geophysics Series Vol. 77, pp. 271-302, Academic Press.

Rintoul, S. R., S. Sokolov, and J. Church, 2002. A six year record of baroclinic transport variability of the Antarctic Circumpolar Current at 140°E, derived from XBT and altimeter measurements. Journal of Geophysical Research – Oceans, 107 (C10): art.no. 3155.

Rintoul., S. R., 2007. Rapid freshening of Antarctic Bottom Water formed in the Indian and Pacific Oceans. Geophys. Res. Lett., 34, L06606, doi:10.1029/2006GL028550.

Risebrough RW, De Lappe BW, Younghans-Haug C (1990) PCB and PCT contamination in Winter Quarters Bay, Antarctica. Mar Pollut Bull 21:523-529

Rist, M. A., P. R. Sammonds, H. Oerter, and C. S. M. Doake, 2002, Fracture of Antarctic shelf ice, *J. Geophys. Res.*, **107**(B1), 10.1029/2000JB000058.

Roberts D, McMinn A (1996) Palaeosalinity reconstruction from saline lake diatom assemblages in the Vestfold Hills, Antarctica. In: Lyons WB, Howard-Williams C, Hawes I (eds) Ecosystem Processes in Antarctic Ice-Free Landscapes: Polar Desert Ecosystems Conference Proceedings, Christchurch, New Zealand, p 207-220

Roberts D, McMinn A (1999a) A diatom-based palaeosalinity history of Ace Lake, Vestfold Hills, Antarctica. The Holocene 9:401-408

Roberts D, McMinn A (1999b) Palaeohydrological modelling of Ace Lake, Vestfold Hills, Antarctica. The Holocene 9:515-520

Roberts D, McMinn A, Cremer H, Gore D, Melles M (2004) The Holocene evolution and palaeosalinity history of Beall Lake, Windmill Islands (East Antarctica) using an expanded diatom-based weighted averaging model. Palaeogeography Palaeoclimatology Palaeoecology 208:121-140

Roberts D, Van Ommen TD, McMinn A, Morgan V, Roberts JL (2001) Late Holocene East Antarctic climatic trends from ice core and lake sediment proxies. The Holocene 11:117-120

Roberts SJ, Hodgson DA, Bentley MJ, Smith JA, Millar I, Olive V, Sugden DE (2008) The Holocene history of George VI Ice Shelf, Antarctic Peninsula from clast-provenance analysis of epishelf lake sediments. Palaeogeography Palaeoclimatology Palaeoecology 259:258-283

Robertson, E.I. and Macdonald, W.J.P. 1962. Electrical resistivity and ground temperature at Scott Base, Antarctica. *New Zealand Journal of Geology and Geophysics*, 5(5), 797-809.

Robertson, R. (2001). Internal tides and baroclinicity in the southern Weddell Sea: Part II: Effects of the critical latitude and stratification. *J. Geophys. Res.* **106**, 27017-27034.

Robertson, R., Beckmann, A. and Hellmer, H. (2003). M_2 tidal dynamics in the Ross Sea. *Antarct. Sci.* **15**, 41-46.

Robertson, R., M. Visbeck, A.L. Gordon and E. Fahrbach (2002) Long-term temperature trends in the Deep Waters of the Weddell Sea, Deep-Sea Research II 49(21), 4791-4806.

Robertson, R., Padman, L. and Levine, M. D. (1995). Fine structure, microstructure, and vertical mixing processes in the upper ocean in the western Weddell Sea. *J. Geophys. Res.* **100**, 18517-18535.

Robin, G. de Q., Swithinbank, C.W.M. and Smith, B.M.E., 1970. Radio echo exploration of the Antarctic ice sheet. *Int. Assoc. Sci. Hydrol. Publ.* **86**, 97-115.

Robinson et al. 2003 – chapter 3 & 4

Robinson et al. 2005 - chapter 4

Rodriguez-Trelles F, Rodriguez MA (1998) Rapid micro-evolution and loss of chromosomal diversity in *Drosophila* in response to climate warming. Evol Ecol 12: 829-838

Roemmich, D., J. Gilson, R. Davis, P. Sutton, S. Wijffels, and S. Riser, 2007: Decadal Spinup of the South Pacific Subtropical Gyre. *J. Phys. Oceanogr.*, **37**, 162–173.

Rogers and van Loon, 1982 – chapter 3

Romanovsky VE, Burgess M, Smith S, Yoshikawa K, Brown J (2002) Permafrost temperature records: indicators of climate change. EOS Transactions 83: 589-594

Romanovsky, V.E. and Osterkamp T. E. 1995. Interannual variations of the thermal regime of the active layer and near surface permafrost in northern Alaska. *Permafrost and Periglacial Processes*, 6: 313-335.

Rønning 1996 – from Chap 1

Root TL, Price JT, Hall KR, Schneider SH, Rosenzweig C, Pounds JA (2003) Fingerprints of global warming on wild animals and plants. Nature 421: 57-60

Roscoe, H.K., Kreher, K., Frieβ, U., Ozone loss episodes in the free Antarctic troposphere, suggesting a possible climate feedback, Geophys. Res. Letts. (28), 2911-2914, 2001.

Rosenheim JA, Tabashnik BE (1991) Influence of generation time on the rate of response to selection. American Naturalist 137: 527-541

Rotschky G., W. Rack, W. Dierking, and H. Oerter, 2006, Retrieving Snowpack Properties and Accumulation Estimates from a Combination of SAR and

Rott, H., P. Skvarca and T. Nagler (1996), Rapid collapse of Northern Larsen Ice Shelf, Antarctica, *Science*, *271*, 788-792.

Rott, H., W. Rack, P. Skvarca, and H. De Angelis (2002), Northern Larsen Ice Shelf, Antarctica: further retreat after collapse. *Annals of Glaciology 34*, 277-282.

Rott, H., W. Rack, T. Nagler, and P. Skvarca, 1998. Climatically induced retreat and collapse of northern Larsen Ice Shelf, Antarctic Peninsula. Annals of Glaciology 27, 86-92.

Rowley R J, Kostelnick J C, Braaten D, Li X, Meisel J (2007) Risk of rising sea level to population and land area, EOS Transactions, 88, 105, 107.

Royer DL (2006) CO2-forced climate thresholds during the Phanerozoic. *Geochim. Cosmochim. Acta 70* 5665-5675

Rozanov, E., M. Schraner, C. Schnadt, T. Egorova, M. Wild, A. Ohmura, V. Zubov, W. Schmutz, Th. Peter, "Assessment of the ozone and temperature variability during 1979-1993 with the chemistry-climate model SOCOL", *Adv. Space Res.*, **35**, 1375-1384 (2005).

Russel J.L., K.W. Dixon , A. Gnanadesikan, R.J. Stouffer and J.R. Toggweiler (2006b). The Southern hemisphere westerlies in a warming world: propping open the door to the deep ocean. J. Clim. 19 4560-4575.

Russel J.L., R.J. Stouffer and K.W. Dixon (2006a). Intercomparison of the Southern Ocean circulation in IPCC coupled model controm simulations. J. Clim. 19 4560-4575

Russell et al (2006) – chapter 3 & 4 – Russel et al 2006a and 2006b?

Ruud JT (1954) Vertebrates without erythrocytes and blood pigment. Nature 173: 848-850

Sabine, C. L., et al. (2004), The Oceanic Sink for Anthropogenic CO_2, *Science*, *305*, 367-371.

Saenko O.A., J.C. Fyfe and M.H. England (2005). On the response of the wind-driven circulation to atmospheric CO_2 increase. Clim. Dyn. 25, 415-426.

Saiz-Lopez, A., Mahajan, A. S., Salmon, R. A., Bauguitte, S. J.-B., Jones, A. E., Roscoe, H.K. and Plane, J.M.C., Boundary layer halogens in coastal Antarctica,
Science 317: 348-351, DOI: 10.1126/science.1141408, 2007

Sallee, Speer, Morrow (2007) Response of the ACC to the atmosphere, J. Climate submitted.

Sands et al – chapter 3

Sands et al, in press – chapter 3

Sandvik A., Biryulina M., Kvamsto N.G., Stamnes J.J., Stamnes K., 2007: Observed and simulated microphysical composition of arctic clouds: Data properties and model validation. *J. Geophys. Res.*, **112,** Art. No. D05205.

Sandwell D.T., W.H.F. Smith, 1997: Marine gravity anomaly from Geosat and ERS-1 satellite altimetry. J. Geophys. Res., 102, 10,039-10,054.

Sarmiento, J. L., et al. (2004), High-latitude Controls of Thermocline Nutrients and Low Latitude Biological Productivity, *Nature*, *427*, 56-60.

Sarmiento, J. L., et al. (2004), Response of Ocean Ecosystems to Climate Warming, *Global Biogeochemical Cycles*, *18*, doi:10.1029/2003GB002134.

Sarmieto J.L., T.M.C Hughes, R.J. Stouffer and S. Manabe (1998). Simulated response of the ocean carbon cycle to an anthropogenic climate warming. Nature 393, 245-249.

Sawagaki, T. 1995. Ground temperature regimes and frost heave activity in the vicinity of Syowa Station, East Antarctica. In: *Proceedings of the NIPR symposium on Antarctic geosciences*, 8, 239-249.

Scambos et al., 2005 – chapter 3

Scambos et al., 2008 in press – chapter 3

Scambos TA, Bohlander J, Shuman C, Skvarca P (2004) Glacier acceleration and thinner after ice shelf collapse in the Larsen B embayment, Antarctica. Geophys Res Lett 31:doi:10.1029/2004GL020670

Scambos, T. A., C. Hulbe, and M. Fahnestock, 2003. Climate-induced ice shelf disintegration in the Antarcitc Peninsula. In : Antarctic Peninsula Climate Variability; Historical and Paleoenvironmental Perspectives (E. Domack, A. Burnett, A. Leventer, P. Conley, M. Kirby, and R. Bindschadler, editors), Antarctic Research Series. Vol 79, 79-92.

Scambos, T., M. Huble, M. Fahnestock and J. Bohlander (2000), The link between climate warming and break-up of ice shelves in the Antarctic Peninsula, *Journal of Glaciology*, *46*, 516-530.

Scatterometer Measurements, *IEEE Trans. Geosci. Remote Sens.*, **44**(4), 943-956.

Scher HD, Martin EE (2006) Timing and Climatic Consequences of the Opening of Drake Passage. Science 312:428-430

Schewe I, Soltwedel T (2003) Benthic response to ice-edge-induced particle flux in the Arctic Ocean. Polar Biol 26: 610-620

Schmidt R, Mäusbacher R, Muller J (1990) Holocene diatom flora and stratigraphy from sediment cores of two Antarctic lakes (King George Island). Journal of Paleolimnology 3:55-74

Schneider *et al.*, 2004 – chapter 3

Schneider, D. P., Steig, E. J., van Ommen, T. D., Dixon, D. A., Mayewski, P. A., Jones, J. M. and Bitz, C. M. 2006. Antarctic temperatures over the past two centuries from ice cores. *Geophysics Research Letters* **33**: L16707, doi:10.1029/2006GL027057.

Schoof, 2007 #2283 - chapter 4

Schoof, C. 2007. Marine ice sheet dynamics. Part I: The case of rapid sliding. *J. Fluid. Mech.*, **573**, 27-55.

Schroder and Fahrbach 1999

Schroeder WH, Anlauf KG, Barrie LA, Lu JY, Steffen A, Schneeberger DR, Berg T (1998) Arctic springtime depletion of mercury. Nature 394:331-332

Schulz KG, Zeebe RE (2006) Pleistocene glacial terminations triggered by synchronous changes in Southern and Northern Hemisphere insolation : the insolation canon hypothesis. Earth and Planetary Science Letters 249:326-336

Schutz, B. E., H. J. Zwally, C. A. Shuman, D. Hancock, and J. P. DiMarzio (2005), Overview of the ICESat Mission, Geophys. Res. Lett., 32, L21S01, doi:10.1029/2005GL024009

Schwerdtfeger, W., 1970: The climate of the Antarctic. World Survey of Climatology. Chapter 4: 253-355.

Scott, R. B. and Wang, F. (2005). Direct evidence of an oceanic inverse kinetic energy cascade from satellite altimetry. *J. Phys. Oceanogr.* **35**, 1650-1666.

Seebacher F, Davison W, Lowe CJ, Franklin CE (2005) A falsification of the thermal specialization paradigm: compensation for elevated temperatures in Antarctic fish. Biol Lett 2: 151-154

Seiler J, Gutt J (2007) Can dead sponges still talk? Antarctic Sci 19 (3): 337-338

Semtner, A., 1976: A model for the thermodynamic growth of sea ice in numerical investigations of climate. Journal of Physical Oceanography, 6 (3), 379–389.

Semtner, A.J., 1974: An oceanic general circulation model with bottom topography. Technical Report No. 9, Dept. of Meteorology, UCLA, Los Angeles, CA 90095, USA.

Sen Gupta, A. and M.H. England, 2007: Coupled ocean-atmosphere feedback in the Southern Annular Mode. *J. Climate*, **20**, 3677-3692.

Severinghaus JP, Sowers T, Brook EJ, Alley RB, Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. Nature 391:141-146

Sexton, D.M.H., 2001: The effect of stratospheric ozone depletion on the phase of the Antarctic Oscillation. *Geophys. Res. Lett.*, **28**, 3697–3700.

Seybold, unpublished – Chapter 3

Shabtaie, S. and Bentley, C. R., 1987, West Antarctic ice streams draining into the Ross ice shelf: configuration and mass balance. *J. Geophys. Res.* **92**, 1311–1336.

Shackleton NJ, Backman J, Zimmerman H, Kent DV, Hall MA, Roberts DG, Schnitker D, Baldauf JG, Desprairies A, Homrighausen R, Huddlestun P, Keene JB, Kaltenback AJ, Krumsiek KAO, Morton AC, Murray JW, Westberg-Smith J (1984) Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. Nature 307:620-623

Shaw 2005 – chap 1

Shaw *et al.* 2005 – chap 1

Shaw GE (1988) Antarctic aerosol: a review. Rev Geophys 26:89-112

Shchepetkin , A.F. & McWilliams, J.C., 2003: A method for computing horizontal pressure-gradient force in an oceanic model with a nonaligned vertical coordinate. Journal of Geophysical Research, 108 (c3), 3090, doi:10.1029/2001JC001047.

Shchepetlin and Williams, 2005 – chapter 3

Shepherd A P, Wingham D J, Mansley, J A D (2002) Inland thinning of the Amundsen Sea sector, West Antarctica, Geophys. Res. Lett., 29, 1364, doi:10.1029/2001GL014183.

Shepherd A. and D. Wingham, 2007, Recent sea-level contributions of the Antarctic and Greenland ice sheets, *Science* **315** (5818), 1529-1532.

Shepherd et al., 2003 – chapter 3

Shepherd LD, Millar CD, Ballard G, Ainley DG, Wilson PR, Haynes GD, Baroni, Lambert DM (2005) Microevolution and mega-icebergs in the Antarctic. Proc Natl Acad Sci 102: 16717-16722

Shepherd, A., D. J. Wingham, and J. A. D. Mansley, 2002. Inland thinning of the Amundsen Sea sector, West Antarctica. Geophysical Research Letters, 29(10), doi:10.1029/2001GL014183.

Shepherd, A., D. Wingham and E. Rignot (2004). Warm ocean is eroding West Antarctic ice sheet. *Geophys. Res. Lett*. 31, L23402, doi:10.1029/2004GL021106.

Shepherd, A., Wingham, D.J., Mansley, J.A.D. & Corr, H.F.J., 2001, Inland thinning of Pine Island Glacier, West Antarctica. *Science* **291**, 862–864, (doi:10.1126/science.291.5505.862).

Sherjal and Fily 1995 – Chapter 3 – 1994?

Sherjal, I. and M. Fily. 1994. Temporal variations of microwave brightness temperatures over Antarctica. *Ann. Glaciol.,* **20**, 19–25.

Shevenell AE, Domack EW, Kernan G (1996) Record of Holocene paleoclimate change along the Antarctic Peninsula: Evidence from glacial marine sediments, Lallemand Fjord. Papers and Proceedings of the Royal Society of Tasmania 130:55-64

Shibata, K., M. Deushi, T.T. Sekiyama, H. Yoshimura, "Development of an MRI chemical transport model for the study of stratospheric chemistry", *Pap. Meteorol. Geophys.*, **55**, 75-119 (2005).

Shindell and Schmidt 2004 - chapter 4

Shindell, D.T., and G.A. Schmidt, 2004: Southern hemisphere climate response to ozone changes and greenhouse gas increases. *Geophys. Res. Lett.*, **31**, L18209, doi:10.1029/2004GL020724.

Shipp S, Anderson JB, Domack EW (1999) Seismic signature of the Late Pleistocene fluctuation of the West Antarctic Ice Sheet system in Ross Sea: a new perspective, Part I. . eological Society of America Bulletin 111:1486-1516

Sidell & O'Brien 2006 – Chapter 2

Siegel V (2005) Distribution and population dynamics of Euphausia superba: summary of recent findings, Pol. Biol., 29, 1-22, doi:10.1007/s00300-005-0058-5.

Siegenthaler U, E. , Monnin K, Kawamura R, Spahni J, Schwander B, Stauffer TF, Stocker J-M, Barnola H, Fischer H (2005a) Supporting evidence from the EPICA Dronning Maud Land ice core for atmospheric CO2 changes during the past millennium. Tellus B 57: 51-57

Siegenthaler U, Stocker TF, Monnin E, Luthi D, Schwander J, Stauffer B, Raynaud D, Barnola J-M, Fischer H, Masson-Delmotte V, Jouzel J (2005b) Stable Carbon Cycle-Climate Relationship During the Late Pleistocene. Science 310:1313-1317

Siegert M. J., M. Pokar, J. A. Dowdeswell and T. Benham, 2005, Radio-echo layering in West Antarctica: a spreadsheet dataset. *Earth Surf. Process. Landforms* **30**, 1583–1591.

Siegert MJ, Barrett P, DeConto R, Dunbar R, Cofaigh CO, Passchier S, Naish TR (2008) Recent Advances in Understanding Antarctic Climate Evolution. Antarctic Sci. 1: 1-13

Simmonds and Keay (2000a) – chapter 3 – 2000?

Simmonds and Keay (2000b) – chapter 3 – 2000?

Simmonds *et al.*, 2003 – chapter 3

Simmonds, I., and K. Keay, 2000: Variability of Southern Hemisphere extratropical cyclone behavior 1958–97. *J. Climate*, **13**, 550–561.

Simmons, H, Hallberg, R. and Arbic, B. (2004). Internal wave generation in a global baroclinic tide model. *Deep-Sea Res. II* **51**, 3043-3068.

Simpson W R.,, R. von Glasow, K. Riedel, P. Anderson, P. Ariya, J. Bottenheim, J. Burrows, L. J. Carpenter, U. Frieß, M. E. Goodsite, D. Heard, M. Hutterli, H.-W. Jacobi, L. Kaleschke, B. Neff, J. Plane, U. Platt, A. Richter, H. Roscoe, R. Sander, P. Shepson, J. Sodeau, A. Steffen, T. Wagner, and E. Wolff, Halogens and their role in polar boundary-layer ozone depletion, Atmos. Chem. Phys., 7, 4375-4418, 2007

Sinclair (1996) – chapter 3

Sinclair 2001 - chapter 4

Sinclair and Chown 2005b - chapter 4

Sinclair, M.R., J.A. Renwick, and J.W. Kidson, 1997: Low-frequency variability of Southern Hemisphere sea level pressure and weather system activity. *Mon. Weather Rev.*, **125**, 2531–2543.

Sinha, B. and K. J. Richards, 1999. Jet structure and scaling in Southern Ocean models. *J. Phys. Oceanogr.*, **29**:1143-1155.

Sjunneskog C, Taylor F (2002) Postglacial marine diatom record of the Palmer Deep, Antarctic Peninsula (ODP Leg 178, Site 1098) 1. Total diatom abundance. Paleoceanography 17:art. no.-8003

Skotnicki ML, Selkirk PM (2006) Plant Biodiversity in an Extreme Environment: genetic studies of origins, diversity and evolution in the Antarctic. In: Bergstrom DM, Convey P, Huiskes AHL (eds) Trends in Antarctic Terrestrial and Limnetic Ecosystems

Skvarca, P. (1994), Changes and surface features of the Larsen Ice Shelf, Antarctica, derived from Landsat and Kosmos mosaics, *Ann. Glaciol. 20*, 6.

Skvarca, P., 1993. Fast recession of the northern Larsen Ice Shelf monitored by space images. Annals of Glaciology, 17, 317-321.

Skvarca, P., W. Rack, H. Rott, T. Ibarzábal y Donángelo, 1998. Evidence of recent climatic warming on the eastern Antarctic Peninsula. Annals of Glaciology 27, 628-635.

Slabber & Chown 2002 – chapter 3

Slabber 2005 – chapter 4

Slattery M, Bockus D (1997) Sedimentation in McMurdo Sound, Antarctica: a disturbance mechanism for benthic invertebrates. Polar Biol 18: 172-179

Sloyan, B. M. (2005). Spatial variability of mixing in the Southern Ocean. *Geophys. Res. Lett.* **32**, doi:10.1029/2005GL023568.

Smale DA, Barnes DKA, Fraser KPP (2007) The influence of depth, site exposure and season on the intensity of iceberg scouring in nearshore Antarctic waters. Polar Biol 30: 769-779

Smedsrud 2006 – chapter 3

Smedsrud, L.H. (2005) Warming of the deep water in the Weddell Sea along the Greenwich Meridian: 1997-2001, Deep-Sea Research I, 52(2), 241-258.

Smetacek V, Nicol S (2005) Polar ocean ecosystems in a changing world. Nature 437: 362-368

Smethie, W.M. and S.S. Jacobs (2005) Circulation and melting under the Ross Ice Shelf: Estimates from evolving CFC, salinity and temperature fields in the Ross Sea. *Deep-Sea Res*. I **52**, 959-978.

Smith 1984 – Chapter 3

Smith 1994 – chapter 3

Smith 2002 - chapter 4

Smith A. M., T. Murray, K.W. Nicholls, K. Makinson, G. Aðalgeirsdóttir, A.E. Behar, D.G. Vaughan, 2007a, Rapid erosion, drumlin formation, and changing hydrology beneath an Antarctic ice stream. *Geology* 35, 127-130.

Smith A.M., Vaughan D.G., Doake C.S.M., et al., 1998, Surface lowering of the ice ramp at Rothera Point, Antarctic Peninsula, in response to regional climate change. *Ann. Glaciol.,* **27**, 113-118.

Smith and Nelson, 1986 – chapter 3

Smith and Steenkamp 1990, Smith 2002 – chapter 4

Smith B. E., C. F. Raymond and C. R. Bentley, 2005, Recent elevation changes on the ice streams and ridges of the Ross Embayment from ICESat crossovers. *Geophys. Res. Lett.* **32**, L21509. (doi:10.1029/2005GL024365).

Smith CR, Minks S, DeMaster DJ (2006) A síntesis of bentho-pelagic coupling on the Antarctic shelf: Food banks, ecosystem inertia and global climate change. Deep-Sea Res II 53: 875-894

Smith JA, Bentley MJ, Hodgson DA, Roberts SJ, Leng MJ, Lloyd JM, Barrett MJ, Bryant C, Sugden DE (2007) Oceanic and atmospheric forcing of early Holocene ice shelf retreat, George VI Ice Shelf, Antarctica Peninsula. Quaternary Science Reviews 26:500-516

Smith R C et al. (1999) Marine ecosystem sensitivity to climate change. BioScience, 49, 393–404.

Smith, A. M., 2007b, Subglacial Bed Properties from Normal-Incidence Seismic Reflection Data, *J. Env. and Eng. Geophys.*, **12**, 3-13.

Smith, W., V. A. Asper. New Ways to Collect Data in the Antarctic, EOS, *Transactions American Geophysical Union*, Vol. 88, no. 48, p. 525, 2007

Smol JP, Wolfe AP, Birks HJ, Douglas MS, Jones VJ, Korhola A, Pienitz R, Ruhland K, Sorvari S, Antoniades D, Brooks SJ, Fallu MA, Hughes M, Keatley BE, Laing TE, Michelutti N, Nazarova L, Nyman M, Paterson AM, Perren B, Quinlan R, Rautio M, Saulnier-Talbot E, Siitonen S, Solovieva N, Weckstrom J (2005) Climate-driven regime shifts in the biological communities of arctic lakes. Proc Natl Acad Sci USA 102: 4397-4402

Sokolov, S. and S. R. Rintoul, 2002: The structure of Southern Ocean fronts at 140E. Journal of Marine Systems, 37, 151-184.

Sokolov, S. and S. R. Rintoul, 2003: The subsurface structure of interannual temperature anomalies in the Australian sector of the Southern Ocean. Journal of Geophysical Research, 108 (C9): Art. No. 3285.

Sokolov, S. and S. R. Rintoul, 2007. Multiple jets of the Antarctic Circumpolar Current south of Australia. Journal of Physical Oceanography, 37, 1394-1412.

Somero GN (2005) Linking biogeography to physiology: evolutionary and acclamatory adjustments of thermal limits. Front Zool 2, 1, DOI: 10.1186/1742-9994-2-1

Somero GN, Dahlhoff E, Lin JJ (1996) Stenotherms and eurytherms: mechanisms establishing thermal optima and tolerance ranges. In: Johnston IA, Bennett AF (Eds) Animals and Temperature: Phenotypic and Evolutionary Adaptation. Cambridge University Press, Cambridge, pp 53-57

Somero GN, De Vries AL (1967) Temperature tolerance of some Antarctic fishes. Science 156: 257-258

Somero GN, Fields PA, Hofmann GE, Weinstein RB, Kawall H (1998) Cold adaptation and stenothermy in Antarctic notothenioid fishes: what has been gained and what has been lost. In: di Prisco G, Pisano E, Clarke A, (Eds), Fishes of Antarctica: a Biological Overview. Springer, Milano, pp 97-109

Sone, T., Strelin, J., Mori, J. 2001 Thermal characteristics of permafrost in Seymour Island and James Ross Island, Antarctic Peninsula. *In: 1st European Permafrost Conference, Rome 26-28 March, 2001– Abstracts*, 100.

Song, Y. T. & Haidvogel, D., 1994: A semi-implicit ocean circulation model using a generalized topography following coordinate system. Journal of Computational Physics, 115, 228–248

Spahni R, Chappellaz J, Stocker TF, Loulergue L, Hausammann G, Kawamura K, Fluckiger J, Schwander J, Raynaud D, Masson-Delmotte V, Jouzel J (2005) Atmospheric Methane and Nitrous Oxide of the Late Pleistocene from Antarctic Ice Cores. Science 310:1317-1321

Sparrow, M., The Southern Ocean Observing System, *CLIVAR Exchanges* Vol. 12, No.4 Oct. 2007

Speich, S., B. Blanke, and W. J. Cai, 2007. Atlantic meridional overturning circulation and the Southern Hemisphere supergyre. Geophys. Res. Let., 34, L23614.

Speich, S., B. Blanke, P. de Vries, et al., 2002. Tasman leakage: A new route in the global ocean conveyor belt. Geophys. Res. Lett., 29, 1416.

Spencer, R. and Vassie, J.M. 1997. The evolution of deep ocean pressure measurements in the U.K. Progress in Oceanography, 40, 423-435.

Spikes, V.B., G.S. Hamilton, S.A. Arcone, S. Kaspari, and P.A. Mayewski, 2004, Variability in accumulation rates from GPR profiling on the West Antarctic Plateau. *Ann. Glaciol.* **39**, 238-244.

Stam WT, Beintema JJ, D'Avino R, Tamburrini M, Cocca E & di Prisco G (1998) Evolutionary studies on teleost hemoglobin sequences. In: di Prisco G, Pisano E & Clarke A (Eds) Fishes of Antarctica. A Biological Overview (pp 355-359). Springer, Milano

Stam WT, Beintema JJ, D'Avino R, Tamburrini M & di Prisco G (1997) Molecular evolution of Antarctic fishes (Notothenioidei). J Mol Evol 45: 437-445

Stanwell-Smith & Peck 1998 - chapter 4

Stark JS, Snape I, Riddle MJ (2003) The effects of petroleum hydrocarbon and heavy metal contamination of marine sediments on recruitment on Antarctic soft-bottom assemblages: a field experimental investigation. J Exp Mar Biol Ecol 283:21-50

Stark JS, Snape I, Riddle MJ (2006) Abandoned Antarctic waste disposal sites: monitoring remediation outcomes and limitations at Casey Station. Ecol Manage Rest 7:21-31

Steig EJ, Brook EJ, White JWC, Sucher CM, Bender ML, Lehman SJ, Morse DL, Waddington ED, Clow GD (1998) Synchronous climate changes in Antarctica and the North Atlantic. Science 282:92-95

Steig EJ, Morse DL, Waddington ED, Stuiver M, Grootes PM, Mayewski PA, Twickler MS, Whitlow SI (2000) Wisconsinan and Holocene climate history from an ice core at Taylor Dome, western Ross Embayment, Antarctica. Geografiska Annaler Series a-Physical Geography 82A:213-235

Steinhage, D., Nixdorf, U., Meyer, U., Miller, H., 1999, New maps of the ice thickness and subglacial topography in Dronning Maud Land, Antarctica, determined by means of airborne radio echo sounding. *Ann. Glaciol.* **29**, 267-272.

Stenseth N C, Mysterud A, Ottersen G, Hurrell J W, Chan K S and Lima M (2002) Ecological effects of climate fluctuations. Science, 297, 1292–1296.

Stephens BB, Keeling RF (2000) The influence of Antarctic sea ice on glacial-interglacial CO2 variations. Nature 404:171-174

Stevens MI, Hogg ID (2003) Long-term isolation and recent range expansion revealed for the endemic springtail Gomphiocephalus hodgsoni from southern Victoria Land, Antarctica. Molecular Ecology 12:2357-2369

Stevens MI, Hogg ID (2006) The molecular ecology of Antarctic terrestrial and limnetic invertebrates and microbes. In: Bergstrom DM, Convey P, Huiskes AHL (eds) Trends in Antarctic Terrestrial and Limnetic Ecosystems: Antarctica as a global indicator. Springer, Dordrecht, p 177-192

Stewart, R. H., C. K. Shum, B. Tapley, and L. Ji, 1996. Statistics of geostrophic turbulence in the southern ocean from satellite altimetry and numerical models. *Physica D*, **98**, 499–513.

Stickley et al. 2004 – Chapter 2

Stocker TF, Wright DG (1991) Rapid transitions of the ocean's deep circulation induced by changes in surface water fluxes. Nature 351:729-732

Stockton WL (1984) The biology and ecology of the epifaunal scallop *Adamussium colbecki* on the west side of McMurdo Sound, Antarctica. Mar Biol 78: 171-178

Stockwell CA, Hendry A, Kinnison MT (2003) Contemporary evolution meets conservation biology. Trends Ecol Evol 18: 94-101

Stokes C.R., C. D. Clark, O. B. Lian, S. Tulaczyk, 2007, Ice stream sticky spots: A review of their identification and influence beneath contemporary and palaeo-ice streams. *Earth-Science Reviews* **81**. 217–249.

Stott L, Timmermann A, Thunell R (2007) Southern Hemisphere and Deep-Sea Warming Led Deglacial Atmospheric CO2 Rise and Tropical Warming. Science 318:435-438

Stratton, R.A., and V.D. Pope, 2004: Modelling the climatology of storm tracks - Sensitivity to resolution. In: *The Second Phase of the Atmospheric Model Intercomparison Project (AMIP2)* [Gleckler, P. (ed.)]. Proceedings of the WCRP/WGNE Workshop, Toulouse, pp. 207-210.

Struthers, H., K. Kreher, J. Austin, R. Schofield, G.E. Bodeker, P.V. Johnston, H. Shiona, A. Thomas, "Past and future simulations of NO2 from a coupled chemistry-climate model in comparison with observations", *Atmos. Chem. Phys.*, **4**, 2227-2239 (2004).

Stuiver and Braziunas, 1993 – Chapter 2

Stuiver M, Denton GH, Hughes TJ, Fastook JL (1981) History of the marine ice sheet in West Antarctica during the last glaciation: a working hypothesis. In: Denton GH, Hughes TJ (eds) The Last Great Ice Sheets. Wiley-Inter-science, New York, p 319-436

Stuiver M, P.J. Reimer, Braziunas TF (1998) High precision radiocarbon age calibration for terrestrial and marine samples. *Radiocarbon 40*:1127-1151

Summerhayes, C., B. Dickson, M. Meredith, P. Dexter and K. Alverson. Observing the Polar Oceans during the International Polar Year and Beyond, *WMO Bulletin* 56 (4), pp 270-2893, 2007

Surdyk and Fily, 1994 – Chapter 3 – 1995?

Surdyk, S. and M. Fily. 1995. Results of a stratified snow emissivity model based on the wave approach: application to the Antarctic ice sheet. *J. Geophys. Res.,* **100**(C5), 8837–8848.

Syvitski JPM, Farrow GE, Atkinson RJA, Moore PG, Andrews JT (1989) Baffin Island macrobenthos: bottom communities and environmental significance. Arctic 42: 232-247

Takahashi, T., and e. al. (2008), Climatological mean surface pCO2 and net CO2 flux over the global oceans, *Deep Sea Research*, *SOCOVV Special Volume*, accepted.

Takahashi, T., et al. (1993), Seasonal variation of CO_2 and nutrients in the high-latitude surfcae oceans: a comparative study, *Global Biogeochemical Cycles*, *7*, 843-878.

Taylor F, Sjunneskog C (2002) Postglacial marine diatom record of the Palmer Deep, Antarctic Peninsula (ODP Leg 178, Site 1098). Paleoceanography 17:10.1029/2000PA000564

Taylor F, Whitehead JM, Domack E (2001) Holocene paleoclimate change in the Antarctic Peninsula: evidence from the diatom, sedimentary and geochemical record. Marine Micropaleontology 41:25-43

Tearle 1987 - chapter 4

Tebaldi et al. (2006) - chapter 4

Tebaldi, C., R. l. Smith, D. Nychka, and L. O. Mearns, 2004: Quantifying Uncertainty in Projections of Regional Climate Change: A Bayesian Approach to the Analysis of Multimodel Ensembles. *J. Clim*, **18,** 1524-1540.

Tedesco, 2008 – chapter 3

Teixidó N, Garrabou J, Gutt J, Arntz WE (2004) Recovery in Antarctic benthos after iceberg disturbance: trends in benthic composition, abundance and growth forms. Mar Ecol Prog Ser 278: 1-16

Thomas CD, Bodsworth EJ, Wilson RJ, Simmons AD, Davies ZG, Musche M, Conradt L (2001) Ecological and evolutionary processes at expanding range margins. Nature 411: 577-581

Thomas CD, Cameron A, Green RE, Bakkenes M, Beaumont LJ, Collingham YC, Erasmus BF, De Siqueira MF, Grainger A, Hannah L, Hughes L, Huntley B, Van Jaarsveld AS, Midgley GF, Miles L, Ortega-Huerta MA, Peterson AT, Phillips OL, Williams SE (2004) Extinction risk from climate change. Nature 427: 145-148

Thomas DN, Dieckmann GS (2002) Antarctic sea ice-a habitat for extremophiles. Science 295: 641-644

Thomas et al. 2001 - Chapter 4 - page 70

Thomas, R. H., E. Rignot, P. Kanagaratnam, W. Krabill, and G. Casassa, 2004, Force perturbation analysis of Pine Island Glacier suggests cause for recent acceleration, *Ann. Glaciol.,* **39**, 133-138.

Thomas, R., and 17 others (2004), Accelerated sea-level rise from West Antarctica. *Science 306*, 255-258.

Thomas, R., E. Rignot, G. Cassasa, P. Kanagaratnam, and 14 others, 2004. Accelerated sea-level rise from West Antarctica. Science 306(5694), doi:10.1126/science.1099650.

Thompson A.F. and K. J. Heywood (subm.) Frontal structure and transport in the northwestern Weddell Sea, submitted to Deep-Sea Research, 2007.

Thompson and Wallace 2000 – chapter 3

Thompson DWJ, Wallace JM (2001) Regional climate impacts of the northern hemisphere annular mode. Science 293:85-89

Thompson et al. (20XX) – Chapter 3 – page 117 & 119

Thompson et al., 2000 – chapter 3

Thompson, A. F., Gille, S. T., MacKinnon, J. A. and Sprintall, J. (2007). Spatial and temporal patterns of small-scale mixing in Drake Passage. *J. Phys. Oceanogr.* **37**, 572-592.

Thompson, D., and S. Solomon, 2002. Interpretation of recent southern hemisphere climate change. Science, 296(5569), 895-899.

Thomson et al., 1971 – chapter 3, page 167 – 1971a?

Thomson, D.C., Bromley, A.M., Craig, R.M.F. 1971a. Ground temperatures in an Antarctic dry valley. *New Zealand Journal of Geology andGeophysics*, 14, 477-483.

Thrush S, Dayton P, Cattaneo-Vietti R, Chiantore M, Cummings V, Andrew N, Hawes I, Kim S, Kvitel R, Schwarz A-M (2006) Broad-scale factors influencing the biodiversity of coastal benthic communities of the Ross Sea. Deep-Sea Res II 53: 959-971

Thurnherr, A. M. and Speer, K. G. (2003). Boundary mixing and topographic blocking on the Mid-Atlantic Ridge in the South Atlantic. *J. Phys. Oceanogr.* **33**, 848-862.

Tian, W., and M.P. Chipperfield, “A new coupled chemistry-climate model for the stratosphere: The importance of coupling for future O3-climate predictions”, *Quart. J. Roy. Meteorol. Soc.*, **131**, 281-303 (2005).

Timmermann, R., H.H. Hellmer and A. Beckmann, 2002: Simulations of ice-ocean dynamics in the Weddell Sea 2. Interannual variability 1985-1993. *J. Geophys. Res.* **107**(C3), 3025, doi:10.1029/2000JC000742.

Toole, J. M. (1981). Intrusion characteristics in the Antarctic Polar Front. *J. Phys. Oceanogr.* **11**, 780-793.

Toon, O. B. & R.P. Turco, “Polar Stratospheric Clouds and Ozone Depletion”, *Scientific American,* **264**, 40-47 (1991).

Torinesi, O., M. Fily and C. Genthon, 2003, Variability and trends of summer melt period of Antarctic ice margins since 1980 from microwave sensors. *J. Climate*, **16**(7), 1047–1060.

Trathan P N and Murphy E J (2002) Sea surface temperature anomalies near South Georgia: relationships with the Pacific El Nino regions. Journal of Geophysical Research Oceans, 108, 8075.

Trathan P N, Forcada J and Murphy E J (2007) Environmental forcing and Southern Ocean marine predator populations: effects of climate change and variability. Philosophical Transactions of the Royal Society B, 362, 2351–2365

Trathan P N, Murphy E J, Forcada J, Croxall J P, Reid K and Thorpe S E (2006) Physical forcing in the southwest Atlantic: ecosystem control. In Top predators in marine ecosystems: their role inmonitoring andmanagement (eds I L Boyd, S Wanless and C J Camphuysen), pp. 28–45. Cambridge, UK: Cambridge University Press.

Trenberth K E and Hoar T L (1996) The 1990-1995 El Niño-Southern Oscillation event: longest on record. Geophysical Research Letters, 57, 60.

Tressler, W.L. and A.M. Ommundsen (1963) Oceanographic studies in McMurdo Sound, Antarctica. IG Bull. 67, *Trans. Am. Geophys. Un.* **44**(1), 217-225.

Truffer M., M. Fahnestock, 2007, Rethinking Ice Sheet Time Scales. *Science* 315, 1508 (2007).

Turner J (2004) The El Niño-Southern Oscillation and Antarctica. International Journal of Climatology, 24, 1-31.

Turner J, Overland JE & Walsh JE (2007) An Arctic and Antarctic perspective on recent climate change. Intern J Climatol 27: 277-293

Turner, J., A. Lachlan-Cope, S. Colwell, G.J. Marshall, and W.M. Connolley, 2006: Significant warming of the Antarctic winter troposphere, *Science*, **311**, 1914–1917.

Turner, J., Colwell, S. R., Marshall, G. J., Lachlan-Cope, T. A., Carleton, A. M., Jones, P. D., Lagun, V., Reid, P. A. and Iagovkina, S. 2005. Antarctic climate change during the last 50 years. *International Journal of Climatology* **25**: 279-294.

Turner, J., Colwell, S. R., Marshall, G. J., Lachlan-Cope, T. A., Carleton, A. M., Jones, P. D., Lagun, V., Reid, P. A. and Iagovkina, S. 2004. The SCAR READER project: Towards a high-quality database of mean Antarctic meteorological observations. *Journal of Climate* **17**: 2890-2898.

Turner, J., Lachlan-Cope, T. A., Colwell, S. R., Marshall, G. J. and Connolley, W. M. 2006. Significant warming of the Antarctic winter troposphere. *Science* **311** [31 March 2006]: 1914-1917.

Turner, J., S. Leonard, T. Lachlan-Cope, and G. J. Marshall, 1998: Understanding Antarctic Peninsula precipitation distribution and variability using a numerical weather prediction model, *Ann. Glaciol.*, **27**, 591–596.

Turner, J., T. A. Lachlan-Cope, S. R. Colwell, G. J. Marshall, and W. M. Connolley, 2006: Significant warming of the Antarctic winter troposphere. *Science*, **311,** 1914-1917.

Turner, J., T.A. Lachlan-Cope, S. Colwell, S., and G.J. Marshall, 2005: A positive trend in western Antarctic Peninsula precipitation over the last 50 years reflecting regional and Antarctic-wide atmospheric circulation changes. *Ann. Glaciol*. **41**, 85-91.

Turner, J., W.M. Connolley, S. Leonard, G.J. Marshall, and D.G. Vaughan, 1999: Spatial and temporal variability of net snow accumulation over the Antarctic from ECMWF re-analysis project data. *Int. J. Climatol*. **19**, 697-724.

UNEP (1966) Question of Antarctica. State of the Environment in Antarctica. Unite Nations Environmental Programme, Chemicals, Geneva

UNEP (2002) Regionally based assessment of persistent toxic substances – Antarctica regional report. United Nations Environmental Programmme, Chemicals, Geneva

Uotila et al. (2006) - chapter 4

Uotila P., A.H. Lynch, J.J. Cassano, R.I. Cullather, 2007: Changes in Antarctic net precipitation in the 21st century based on Intergovernmental Panel on Climate Change (IPCC) model scenarios, *J. Geophys. Res*., **112**, D10107, doi:10.1029/2006JD007482.

Urban, T. J., and B. E. Schutz (2005), ICESat sea level comparisons, Geophys. Res. Lett., 32, L23S10, doi:10.1029/2005GL024306

Van de Berg, W.J., M.R. van den Broeke, C.H. Reijmer, and E. van Meijgaard, 2006: Reassessment of the Antarctic surface mass balance using calibrated output of a regional atmospheric climate model. *J. Geophys. Res*., **111**, D11104, doi:10.10292005JD006495.

van de Berg, W.J., M.R. van den Broeke, C.H. Reijmer, and E. van Meijgaard, 2005: Characteristics of the Antarctic surface mass balance (1958-2002) using a regional atmospheric climate model. *Ann. Glaciol.*, **41**, 97–104.

van den Broeke et al. 2006a – chapter 3 – 2006?

Van den Broeke, M. 2005. Strong surface melting preceded collapse of Antarctic Peninsula ice shelf. Geophysical Research Letters, 32, L12815, doi:10.1029/2005GL023247.

Van den Broeke, M.R., and N.P.M. van Lipzig, 2004: Changes in Antarctic temperature, wind and precipitation in response to the Antarctic Oscillation, *Ann. Glaciol.,* **39**, 199-126.

Van den Broeke, M.R., W.J. van de Berg and E. van Meijgaard, 2006: Snowfall in coastal West Antarctica much greater than previously assumed. *Geophys. Res. Letts.* **33**, L02505, doi:10.1029/2005GL025239.

Van der Veen, C.J. (1999a), *Fundamentals of Glacier Dynamics.* Rotterdam: A.A. Balkema, 462 pp.

Van der Veen, C.J. (1999b), Evaluating the performance of cryospheric models. *Polar Geography 23*, 83-96.

Van der Veen, C.J., and A.J. Payne (2003), Modelling land-ice dynamics. In: *Mass Balance of the Cryosphere: observations and modelling of contemporary and future changes (eds. J.L. Bamber and A.J. Payne).* Cambridge: Cambridge University Press, 169-225.

Van der Veen, C.J., and ISMASS (2007), A need for more realistic ice-sheet models. *SCAR Report No. 30*, 27 pp.

Van der Veen, C.J., K.C. Jezek, and L. Stearns (2007), Shear measurements across the northern margin of Whillans Ice Stream. *Journal of Glaciology 53*, 17-29.

Van Dijk PLM, Tesch C, Hardewig I, Pörtner HO (1999) Physiological disturbances at critically high temperatures: a comparison between stenothermal Antarctic and eurythermal temperate eelpouts (Zoarcidae). J Exp Biol 202: 3611-3621

Van Everdingen, R.O. 1998. Multi-language glossary of permafrost and related ground ice terms. International Permafrost Association (http://www.geo.uro.no/IPA/).

van Lipzig, N. P. M., G. J. Marshall, A. Orr and J. C. King, The relation between the Southern Hemisphere Annular Mode and Antarctic Peninsula summer temperatures: Analysis of a high-resolution model climatology, *J. Clim.,* 2006.

van Lipzig, N. P. M., J. C. King, T. A. Lachlan-Cope, and M. R. van den Broeke (2004), Precipitation, Sublimation and Snow Drift in the Antarctic Peninsula Region from a Regional Atmospheric Model, *Journal of Geophysical Research, 109,* doi 10.1029/2004JD004701.

van Lipzig, N. P. M., J. Turner, S. R. Colwell, and M. R. Van den Broeke, 2004: The near-surface wind field over the Antarctic continent. *International Journal of Climatology*, **24,** 1973-1982.

Van Lipzig, N.P.M., E. van Meijgaard, and J. Oerlemans, 2002: The spatial and temporal variability of the surface mass balance in Antarctica: results from a regional atmospheric climate model. *Int. J. Climatol.* **22**, 1197-1217.

van Loon, H., 1967: The half yearly oscillations in middle and high southern latitudes and the coreless winter. *J.Atmos.Sci.*, **24**, 472-486.

Van Oldenborgh, G.J., S.Y. Philip, and M. Collins, 2005: El Nino in a changing climate: a multi-model study. *Ocean Sci.*, **1**, 81–95.

van Ommen and Morgan, 1997 – Chapter 2

Vaughan and Spouge, 2002 - chapter 4

Vaughan D G, Holt J W, Blankenship D D (2007) West Antarctic links to sea level estimation, EOS, Transactions, 88(46), 485-486.

Vaughan, D. G., D. R. Mantripp, J. Sievers, and C. S. M. Doake, 1993. A synthesis of remote sensing data on Wilkins Ice Shelf, Antarctica. Annals of Glaciology 17, 211-218.

Vaughan, D. G. (2006), Recent trends in melting conditions on the Antarctic Peninsula and their implications for ice-sheet mass balance, *Arctic, Antarctic and Alpine Research*, *38*, 147-152.

Vaughan, D. G. , and C. S. M. Doake, 1996. Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula. Nature 379, 328-330.

Vaughan, D. G., 1993. Implications of the break-up of the Wordie Ice Shelf, Antarctica for sea level rise. Antarctic Science 5(4), 403-408.

Vaughan, D. G., H. F. J. Corr, F. Ferraccioli, N. Frearson, A. O'Hare, D. Mach, J. W. Holt, D. D. Blankenship, D. L. Morse, and D. A. Young (2006), New boundary conditions for the West Antarctic ice sheet: Subglacial topography beneath Pine Island Glacier, *Geophys. Res. Lett.*, **33**, L09501, doi:10.1029/2005GL025588.

Vaughan, D.G. (1993), Implications of the break-up of Wordie Ice Shelf, Antarctica for sea level. *Antarctic Science 5*, 403-408.

Vaughan, D.G., and C.S.M. Doake (1996), Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula. *Nature 379*, 328-331.

Vaughan, D.G., G. Marshall, W.M. Connolley, C. Parkinson, R. Mulvaney, D.A. Hodgson, J.C. King, C.J. Pudsey, J. Turner, and E. Wolff (2003), Recent rapid regional climate warming on the Antarctic Peninsula. *Climatic Change 60*, 243-274.

Vaughan, D.G., J.L. Bamber, M. Giovinetto, J. Russell and A.P.R. Cooper, 1999: Reassessment of net surface mass balance in Antarctica. *J. Climate* **12**, 933-946.

Vazquez et al., "NOAA/NASA AVHRR Oceans Pathfinder Sea Surface Temperature Data Set User's Reference Manual Version 4.0", 10 April 1998, JPL Publication D-14070

Vázquez-Rodríguez, M., et al. (2008), Anthropogenic carbon distributions in the Atlantic ocean: data-based estimates from the Arctic to the Antarctic, *Biogeosciences Discussion*, *5*, 1421-1443.

Velicogna, I., and J. Wahr, 2002, A method for separating Antarctic postglacial rebound and ice mass balance using future ICESat Geoscience Laser Altimeter System, Gravity Recovery and Climate Experiment, and GPS satellite data, *J. Geophys. Res*. **107**, 10.1029/2001JB000708.

Velicogna, I., and J. Wahr, 2006, Measurements of Time-Variable Gravity Show Mass Loss in Antarctica, *Science* **311** (5768), 1754 [DOI:10.1126/science.1123785]

Venegas S A (2003) The Antarctic Circumpolar Wave: a combinationof two signals? Journal of Climate, 16, 2509-2525.

Venegas, S.A., and M.R. Drinkwater. Sea ice, atmosphere and upper ocean variability in the Weddell Sea, Antarctica. Journal of Geophysical Research, 106(C8), 16,747–16,766, 2001

Verde et al. 2002 – Chapter 2

Verde et al. 2006a – Chapter 2

Verde et al. 2006a, 2006b – Chapter 2

Verde et al. 2006b – Chapter 2

Verdy, A., et al. (2007), Carbon dioxide and oxygen fluxes in the Southern Ocean: mechanisms of interannual variability, *Global Biogeochemical Cycles*, *GB2020*.

Verkulich SR, Hiller A (1994) Holocene Deglaciation of the Bunger Hills Revealed by C-14 Measurements on Stomach Oil Deposits in Snow Petrel Colonies. Antarctic Science 6:395-399

Verkulich SR, Melles M, Pushina ZV, Hubberten H-W (2002) Holocene Environmental changes and development of Figurnoye Lake in the southern Bunger Hills, East Antarctica. Journal of Paleolimnology 28:253-267

Verleyen E, Hodgson DA, Milne GA, Sabbe K, Vyverman W (2005) Relative sea level history from the Lambert Glacier region (East Antarctica) and its relation to deglaciation and Holocene glacier re-advance. Quaternary Research 63:45-52

Verleyen E, Hodgson DA, Sabbe K, Vanhoutte K, Vyverman W (2004a) Coastal oceanographic conditions in the Prydz Bay region (East Antarctica) during the Holocene recorded in an isolation basin. The Holocene 14:246-257

Verleyen E, Hodgson DA, Sabbe K, Vyverman W (2004b) Late Quaternary deglaciation and climate history of the Larsemann Hills (East Antarctica). Journal of Quaternary Science 19:361-375

Vetter RAH, Buchholz F (1998). Kinetics of enzymes in cold stenothermal invertebrates. In Cold Ocean Physiology (eds HO Pörtner, RC Playle), pp 190–211. Soc Exp Biol Seminar Ser 66

Vieli, A., A. J. Payne, A. Shepherd, and Z. Du, 2007. Causes of pre-collapse o fthe Larsen B ice shelf: numerical modeling and assimilation of satellite observations. Earth and Planetary Science Letters 259, 297-306, doi:10.1016/j.epsl.2007.04.50.

Vieli, A., and A.J. Payne (2005), Assessing the ability of numerical ice sheet models to simulate grounding line migration. *Journal of Geophysical Research 110*, F01003, doi: 10.1029/2004JF000202.

Vieli, A., J. Jania and L. Kolondra (2002), The retreat of a tidewater glacier: observations and model calculations on Hansbreen, Spitsbergen, *J. Glaciol.*, *48*, 592-600.

Vieli, A., M. Funk, and H. Blatter, 2001, Flow dynamics of tidewater glaciers: a numerical modelling approach, *J. Glaciol.,* **47**, 595– 606.

Vimeux F, Cuffey K, Jouzel J (2002) New insights into Southern Hemisphere temperature changes from Vostok ice cores using deuterium excess correction over the last 420, 000 years. Earth Planet. Sci. Lett. 2O3:829-843

Vogel, S. W., S. Tulaczyk, B. Kamb, H. Engelhardt, F. Carsey, A. Behar, A. Lane and I. Joughin, 2005, Subglacial Conditions During And After Stoppage Of An Antarctic Ice Stream: Is Reactivation Imminent. *Geophys. Res. Lett.* **32** (14), L14502, doi:10.1029/2005GL022563.

Volk, T., and M. I. Hoffert (1985), Ocean Carbon Pumps: Analysis of relative strenghts and efficencies in ocean-driven circulation.

von Glasow, R. R. Sander, A. Bott, and P.J. Crutzen, Modelling halogen chemistry in the marine boundary layer. 2. Interactions with sulfur and cloud-covered MBL. J. Geophys. Res., 107:4323, 2002.

Wacongne, S., K. Speer, R. Lumpkin, and V. Sadouley (2007) Warm water formation in the midst of the Southern Ocean, J. Marine Res., submitted.

Waelbroeck C, Duplessy JC, Michel E, Labeyrie L, Paillard D, Duprat J (2001) The timing of the last deglaciation in North Atlantic climate records (vol 412, pg 724, 2001). Nature 414:470-470

Wagner B, Cremer H, Hultzsch N, Gore DB, Melles M (2004) Late Pleistocene and Holocene history of Lake Terrasovoje, Amery Oasis, East Antarctica, and its climatic and environmental implications. Journal of Paleolimnology 32:321-339

Wagner B, Hultzsch N, Melles M, Gore DB (2007) Indications of Holocene sea-level rise in Beaver Lake, East Antarctica. Cambridge Journals Online, p 125-128

Wagner B, Melles M, Doran PT, Kenig F, S.L. F, Pierau R, Allan P (2006) Glacial and postglacial sedimentation in the Fryxell basin, Taylor Valley, southern Victoria Land, Antarctica Palaeography, Palaeoclimatology, Palaeoecology 241:320-337

Walker D P, Brandon M A, Jenkins A, Allen J T, Dowdeswell J A, Evans J (2007) Oceanic heat transport onto the Amundsen Sea shelf through a submarine glacial trough, Geophys. Res. Lett., 34, doi:10.1029/2006GL028154.

Wallace, D. W. R. (2001), Introduction to special section: Ocean measurements and models of carbon sources and sinks, *Global Biogeochemical Cycles*, *15*, 3-10.

Walther G-R, Post E, Convey P, Menzel A, Parmesan C, Beebee TJC, Fromentin J-M, Hoegh-Guldberg O, Bairlein F (2002) Ecological responses to recent climate change. Nature 416: 389-395

Walton, D.W.H. 1977. Radiation and soil temperatures 1972-74: Signy Island Terrestrial Reference Site. *British Antarctic Survey Data*, 1, 49pp.

Wang, DeVries, Cheng, 1995

Wang, F.M., 2007: Investigating ENSO sensitivity to mean climate in an intermediate model using a novel statistical technique. *Geophys. Res. Lett.*, **34**, L07705.

Wang, X.L., V.R. Swail, and F.W. Zwiers, 2006: Climatology and changes of extratropical storm tracks and cyclone activity: Comparison of ERA-40 with NCEP/NCAR Reanalysis for 1958–2001. *J. Climate*, **19**, 3145−3166.

Wania F, Mackay D (1993) Global fractionation and cold condensation of volatile organochlorine compounds in polar regions. Ambio 22:10-18

Wanninkhof, R. (1992), Relationship between wind speed and gas exchange over the ocean, *Journal of Geophysical Research*, *97*, 7373-7382.

Ward, C. G. (1995), Mapping ice front changes of Müller Ice Shelf, Antarctic Peninsula, *Antarct. Science, 7,* 197.

Wardle 1991 – chap 1

Wasley et al. 2006 – chapter 3 & 4

Watanabe O, Jouzel. J, Johnsen S, Parrenin F, Shoji H, Yoshida N (2003) Homogeneous climate variability across East Antarctica over the past three glacial cycles. Nature 422:509-512

Watson, A. J., and J. C. Orr (2003), Carbon Dioxide Fluxes in the Global Ocean, in *Ocean Biogeochemistry: The Role of the Ocean carbon Cycle in Global Change*, edited by M. J. R. Fasham, Springer, Heidelberg.

Waugh, D. W., et al. (2006), Anthropogenic CO2 in the oceans estimated using transit-time distributions, *Tellus B*, *58*, 376–389.

Weaver AJ, Saenko OA, Clark PU, Mitrovica JX (2003) Meltwater Pulse 1A from Antarctica as a trigger of the Bølling-Allerød warm interval. Science 299:1709-1713

Webb, D.J. & de Cuevas, B. A., 2007: On the fast response of the Southern Ocean to changes in the zonal wind. Ocean Science, 3, 417-427.

Webb, D.J., and B.A.de Cuevas, 2006: On the fast response of the Southern Ocean to changes in the zonal wind. Ocean Science Discussions, 3(3), 471-501.

Webster J, Hawes I, Downes M, Timpeley M, Howard-Williams C (1996) Evidence for regional climate change in the recent evolution of a high latitude pro-glacial lake. Antarctic Science 8:49–59

Webster J, Webster K, Nelson P, Waterhouse E (2003) The behaviour of residual contaminants at a former station site, Antarctica. Environ Pollut 123:163-179

Weertman J, 1974, Stability of the junction of an ice sheet and an ice shelf. *J Glaciol* **13**, 3–11.

Weertman, J., 1973, Can a water-filled crevasse reach the bottom surface of a glacier? *International Association of Scientific Hydrology Publication* **95**, 139-145.

Wei W (2004) Opening of the Australia-Antarctica Gateway as dated by nannofossils. Marine Micropaleontology 52:133-152

Weimerskirsch H, Inchausti P, Guinet C and Barbraud C (2003) Trends in bird and seal populations as indicators of a system shift in the Southern Ocean. Antarctic Science, 15, 249–256.

Weisse et al. 1999 – chapter 3

Welch KA, Mayewski PA, Whitlow SI (1993) Methanesulphonic acid in coastal Antarctic snow related to sea-ice extent. Geophysical Research Letters 20:443-446

Wellner JS, Heroy DC, Anderson JB (2006) The death mask of the antarctic ice sheet: Comparison of glacial geomorphic features across the continental shelf. Geomorphology 75:157-171

Wells RMG, Macdonald JA, di Prisco G (1990) A rheological comparison of the haemoglobin-free icefishes, *Chionodraco kathleenae* and *Cryodraco antarcticus*, with a red-blooded notothenioid, *Pagothenia bernacchii*. J Fish Biol 36: 595-609

Wessel, S., S. Aoki, P. Winkler, R. Weller, A. Herber, H. Gernandt and O. Schrems, Tropospheric ozone depletion in polar regions: A comparison of observations in the Arctic and Antarctic. Tellus, 50B, 34-50, 1998.

Whinam et al. 2005 - chapter 4

White et al. (2004) – chapter 3

White W B & Peterson R G (1996) An Antarctic circumpolar wave in surface pressure, wind, temperature, and sea ice extent, Nature, 380, 699-702.

White W B, Chen S C, Allan R J and Stone R C (2002) Positive feedbacks between the Antarctic Circumpolar Wave and the global El Nino-Southern Oscillation Wave. Journal of Geophysical Research Oceans 107, 3165.

White, W.B., and R.G. Peterson, 1996: An Antarctic Circumpolar Wave in surface pressure, wind, temperature and sea-ice extent. *Nature*, **380**, 699–702.

Whitehead J.M., S. Wotherspoon, Bohaty S.M. (2005) Minimal Antarctic sea ice during the Pliocene. *Geology 33*:137-140

Whitworth, T. and Peterson, R.G. 1985. Volume transport of the Antarctic Circumpolar Current from bottom pressure measurements. Journal of Physical Oceanography, 15, 810-816.

Whitworth, T., A.H. Orsi, S.J. Kim, W.D. Nowlin and R.A. Locarnini (1998) Water masses and mixing near the Antarctic Slope Front. *Antarct. Res. Ser.* **75**, 1-27.

Whitworth, T., III (2002), Two modes of bottom water in the Australian-Antarctic Basin. *Geophys. Res. Lett.* 29(5), 1073, doi:10.1029/2001GL014282.

Wiencke C (1996) Recent advances in the investigation of Antarctic macroalgae. Polar Biol 16: 231-240

Wild, M., P. Calanca, S. C. Scherrer, and A. Ohmura, 2003: Effects of polar ice sheets on global sea level in high-resolution greenhouse scenarios, *J. Geophys. Res.*, **108**, 4165, doi:10.1029/2002JD002451.

Willebrand, J., Barnier, B., Boening, C., Dieterich, C., Killworth, P, Le Provost, C,. Jia. Y., Molines, J. & New, A.L., 2001: Circulation characteristics in three eddy-permitting models of the North Atlantic. Progress in Oceanography, 48, 123–161

Williams, M.J.M., A. Jenkins and J. Determann, 1998: Physical controls on ocean circulation beneath ice shelves revealed by numerical models. *Antarct. Res. Ser.* **75**, 285-299.

Williams, M.J.M., R.C. Warner and W.F. Budd, 2002: Sensitivity of the Amery Ice Shelf, Antarctica, to changes in the climate of the Southern Ocean, *J. Climate* **15**(19), 2740-2757. Bromwich, D.H., 1988: Snowfall in high southern latitudes. *Rev. Geophys.* **26**, 149-168.

Williams, R. S. Jr. and J. G. Ferrigno (1998) U.S. Geological Survey Fact Sheet 50-98, 2p.

Williams, R. S. Jr., J. G. Ferrigno, C. Swithinbank, B. K. Lucchitta, B. and A. Seekins (1995) Coastal-change and glaciological maps of Antarctica, *Ann. Glaciol. 21,* 284.

Willis, J. K., D. Roemmich, and B. Cornuelle, Interannual variability in upper ocean heat content, temperature, and thermosteric expansion on global scales. *J. Geophys. Res.*, **109**, doi:10.1029/2003JC002260 (2004).

Wilson P R, Ainley D G, Nur N, Jacobs S S, Barton K J, Ballard G and Comiso J C (2001) Adélie penguin population change in the pacific sector of Antarctica: relation to sea-ice extent and the Antarctic Circumpolar Current. Marine Ecology Progress Series, 213, 301-309.

Wingham et al, 2007 – chapter 3

Wingham, 1998 #944 - chapter 4

Wingham, D.J., A. L. Ridout, R. Scharroo, R. J. Arthern, and C. K. Schum, 1998, Antarctic elevation change from 1990 to 1996. *Science* 282, 456–458, (doi:10.1126/science.282.5388.456).

Wingham, D.J., A. Shepherd, A. Muir, and G.J. Marshall, 2006: Mass balance of the Antarctic ice sheet. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences, pp. DOI: 10.1098/rsta.2006.1792.

Wingham, D.J., A.J. Ridout, R. Scharroo, R.J. Arthern, and C.K. Schum, 1998: Antarctic elevation change from 1992 to 1996. Science, 282, pp. 456-458.

Wingham, D.J., M.J. Siegert, A. Shepherd and A.S. Muir, 2006. Rapid discharge connects Antarctic subglacial lakes, *Nature*, Vol. 440, pp. 1033-1036 , doi:10.1038/nature04660.

Wlodarska-Kowalczuk M, Pearson TH, Kendall MA (2005) Benthic response to chronic natural physical disturbance by glacial sedimentation in an Arctic fjord. Mar Ecol Prog Ser 303:31-41

WMO 2006. South Orkneys – a century of Antarctic meteorology. WMO Bulletin 55: 49-52

WMO Guide to Meteorological Instruments and Methods of Observations (1996).

WMO, "Scientific Assessment of Ozone Depletion: 2002", *World Meteorological Organisation Global Ozone Research and Monitoring Project - Report No. 47* (2002).

Wójcik, G. 1989. Temperature of active layer at Bunger Oasis in Antarctica in summer 1978-79, *In* K. Senneset (ed.): *5th International Conference onPermafrost*, 1. Tapir, Trondheim, 505-510.

Wolff E, Rankin AM, Rothlisberger R (2003) An ice core indicator of Antarctic sea ice production? . Geophysical Research Letters 30:2158

Wolff EW, Cachier H (1998) Concentrations and seasonal cycle of black carbon in aerosol at a coastal Antarctic station. J Geophys Res D 103:11033-11041

Wolff EW, Fischer H, Fundel F, Ruth U, Twarloh B, Littot GC, Mulvaney R, Röthlisberger R, de Angelis M, Boutron CF, Hansson M, Jonsell U, Hutterli MA, Lambert F, Kaufmann P, Stauffer B, Stocker TF, Steffensen JP, Bigler M, Siggaard-Andersen ML, Udisti R, Becagli S, Castellano E, Severi M, Wagenbach D, Barbante C, Gabrielli P, Gaspari V (2006) Southern Ocean sea-ice extent, productivity and iron flux over the past eight glacial cycles. Nature 440:491-496

Woodworth, P.L. and Player, R. 2003. The Permanent Service for Mean Sea Level: an update to the 21st century. Journal of Coastal Research, 19, 287-295.

Woodworth, P.L., Aarup, T., Merrifield, M., Mitchum, G.T. and Le Provost, C. 2003. Measuring progress of the Global Sea Level Observing System. EOS, Transactions of the American Geophysical Union, 84(50), 16 December 2003, 565, 10.1029/2003EO500009.

Woodworth, P.L., Gregory, J.M. and Nicholls, R.J. 2004. Long term sea level changes and their impacts. pp.715-753 (chapter 18) in The Sea, Volume 13, edited by A.R. Robinson and K.H. Brink. Harvard University Press.

Woodworth, P.L., Hughes, C.W., Blackman, D.L., Stepanov, V.N., Holgate, S.J., Foden, P.R., Pugh, J.P., Mack, S., Hargreaves, G.W., Meredith, M.P., Milinevsky, G. and Fierro Contreras, J.J. 2006. Antarctic peninsula sea levels: a real time system for monitoring Drake Passage transport. Antarctic Science, 18(3), 429-436.

Woodworth, P.L., Le Provost, C., Rickards, L.J., Mitchum, G.T. and Merrifield, M. 2002. A review of sea-level research from tide gauges during the World Ocean Circulation Experiment. Oceanography and Marine Biology: An Annual Review, 40, 1-35.

Woodworth, P.L., Pugh, D.T., Meredith, M.P. and Blackman, D.L. 2005. Sea level changes at Port Stanley, Falkland Islands. Journal of Geophysical Research, Vol. 110, C06013, doi:10.1029/2004JC002648.

Woodworth, P.L., Vassie, J.M., Hughes, C.W. and Meredith, M.P. 1996. A test of the ability of TOPEX/POSEIDON to monitor flows through the Drake Passage. Journal of Geophysical Research, 101(C5), 11935-11947.

Worby and Allison, 1999 – chapter 3

Working Group I Contribution to the Fourth Assessment Report of the IPCC Intergovernmental Panel on Climate Change, Cambridge Universiy Press, Cambridge

World Climate Research Programme workshop on Sea Level Rise and Variability – chapter 3

World Meteorological Organization, "Scientific Assessment of Ozone Depletion: 2002", *Global Ozone Research and Monitoring Project–Report No. 47*, Geneva, Switzerland (2003).

World Meteorological Organization, "Scientific Assessment of Ozone Depletion: 2006", *Global Ozone Research and Monitoring Project–Report No. 50*, Geneva, Switzerland (2006).

World Meteorological Organization, 1970 – chapter 3

Wunsch C. (1998). The work done by the wind on the oceanic general circulation. *J. Phys. Oceanogr,* **28**, 2332–2340.

Wunsch, C. and Ferrari, R. (2004). Vertical mixing, energy, and the general circulation of the oceans. *Ann. Rev. Fluid Mech.* **36**, 281-314.

Xie, P.P., and P.A. Arkin, 1998: Global monthly precipitation estimates from satellite-observed outgoing longwave radiation. *J. Climate* **11**, 137-164.

Yamaguchi and Noda, 2006 - chapter 4

Yamashita N, Kannan K, Taniyasu S, Horii Y, Petrick G, Gamo T (2005) A global survey of perfluorinated acids in oceans. Mar Pollut Bull 51:658-668

Yan Y, Mayewski PA, Kang S, Meyerson E (2005) An ice-core proxy for Antarctic circumpolar zonal wind intensity. Annals of Glaciology 41:121-130

Yang E.-S., D. M. Cunnold, R. J. Salawitch, M. P. McCormick, J. Russell III, J. M. Zawodny, S. Oltmans, M. J. Newchurch (2006), Attribution of recovery in lower-stratospheric ozone, J. Geophys. Res., 111, D17309, doi:10.1029/2005JD006371.

Yao T, Petit JR, Jouzel J, Lorius C, Duval P (1990) Climatic record from an ice margin area in East Antarctica. Annals of Glaciology 14:323-327

Yeh, S.-W. and B.P. Kirtman, 2007: ENSO amplitude changes due to climate change projections in different coupled models. *J. Climate*, **20**, 203-217.

Yin, J. H., 2005: A consistent poleward shift of the storm tracks in simulations of 21st century climate. *Geophysical Research Letters*, **32,** L18701.

Yokoyama Y, Lambeck K, De Deckker P, Johnston P, Fifield IK (2001) Timing of the Last Glacial Maximum from observed sea-level minima (vol 406, pg 713, 2000). Nature 412:99-119

Yoon HI, Park BK, Kim Y, Kim D (2000) Glaciomarine Sedimentation and Its Paleoceanographic Implications Along the Fjord Margins in the South Shetland Islands, Antarctica During the Last 6000 Years. Palaeogeography Palaeoclimatology Palaeoecology 157:189-211

Younger Dryas episode – Chapter 2

Yuan (2004) – chapter 3

Yuan and Martinson, 2000 – chapter 3

Yuan and Martinson, 2001 – chapter 3

Yuan, 2004, Antarctic Science – chapter 3

Zachos JC, J. Breza, Wise SW (1992) Early Oligocene ice sheet expansion on Antarctica, sedimentological and isotopic evidence from the Kergulen Plateau. *Geology 20*:569-573

Zachos JC, Rohl U, Schellenberg SA, Sluijs A, Hodell DA, Kelly DC, Thomas E, Nicolo M, Raffi I, Lourens LJ, McCarren H, Kroon D (2005) Rapid Acidification of the Ocean During the Paleocene-Eocene Thermal Maximum. Science 308:1611-1615

Zachos JC, Wara MW, Bohaty S, Delaney ML, Petrizzo MR, Brill A, Bralower TJ, Premoli-Silva I (2003) A Transient Rise in Tropical Sea Surface Temperature During the Paleocene-Eocene Thermal Maximum. Science 302:1551-1554

Zakhartsev MV, De Wachter B, Sartoris FJ, Pörtner HO, Blust R (2003) Thermal physiology of the common eelpout (*Zoarces viviparus*). J Comp Physiol B 173: 365-378

Zane et al. 2006 – Chapter 2

Zhang X, Walsh JE (2006) Toward a seasonally ice-covered Arctic Ocean: scenarios from the IPCC AR4 model simulations. J Climate 19: 1730-1747

Zhang, T., and Stammes, K. 1998. Impact of climatic factors on the active layer and permafrost at Barrow, Alaska. *Permafrost and Periglacial Processes*, 9: 1, 229-246.

Zhao Y, Ratnayake-Lecamwasam M, Parker SK, Cocca E, Camardella L, di Prisco G, Detrich HW III (1998) The major adult α-globin gene of Antarctic teleosts and its remnants in the hemoglobinless icefishes. Calibration of the mutational clock for nuclear genes. J Biol Chem 273: 14745-14752

Zickfeld, K., et al. (2007), Response of the global carbon cycle to human-induced changes in Southern Hemisphere winds, *Geophysical Research Letters*, *34*, L12712.

Zickfeld, K., et al. (2008), Comment on "Saturation of the Southern Ocean CO2 Sink due to Recent Climate Change", *Nature*, 570b.

Zielinski GA, Mayewski PA, Meeker LD, Whitlow S, Twickler MS, Morrison M, Meese DA, Gow AJ, Alley RB (1994) Record of Volcanism Since 7000 B.C. from the GISP2 Greenland Ice Core and Implications for the Volcano-Climate System, p 948-952

Zielinski U, Gersonde R (1997) Diatom distribution in Southern Ocean surface sediments (Atlantic sector): Implications for paleoenvironmental reconstructions. Palaeogeography, Palaeoclimatology, Palaeoecology 129:213-250

Zwally et al. (1986) – chapter 3

Zwally et al., 1983 – chapter 3

Zwally et al., 2006 - chapter 4

Zwally H.J, Comiso JC, Parkinson CL, Cavalieri DJ, Gloersen P (2002) Variability of Antarctic sea ice 1979-1998. Journal of Geophysical Research 107:3041, doi: 3010.1029/2000JC000733

Zwally, H. J., M. Giovinetto, J. Li, H. Cornejo, M. Beckley, A. Brenner, J. Saba, and D. Yi, 2005. Mass changes of the Greenland and Antarctic ice sheets and shelves and contributions to sea-level rise: 1992-2002. Journal of Glaciology 51(175), 509-527.

Zwally, H. J., W. Abdalati, T. Herring, K. Larson, J. Saba and K. Seffen (2002), Surface melt-induced acceleration of Greenland ice-sheet flow, *Science*, *297*, 218-222.

Zwally, H.J. and S. Fiegles, 1994, Extent and duration of Antarctic surface melting. *J. Glaciol.,* **40**(136), 463–476.

Zwartz D, Bird M, Stone J, Lambeck K (1998) Holocene sea-level change and ice-sheet history in the Vestfold Hills, East Antarctica. Earth and Planetary Science Letters 155:131-145