CHAPTER ONE

Impacts of the Oceans on Climate Change

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Abstract

The oceans play a key role in climate regulation especially in part buffering (neutralising) the effects of increasing levels of greenhouse gases in the atmosphere and rising global temperatures. This chapter examines how the regulatory processes performed by the oceans alter as a response to climate change and assesses the extent to which positive feedbacks from the ocean may exacerbate climate change. There is clear evidence for rapid change in the oceans. As the main heat store for the world there has been an accelerating change in sea temperatures over the last few decades, which has contributed to rising sea-level. The oceans are also the main store of carbon dioxide (CO_2), and are estimated to have taken up $\sim$40% of anthropogenic-sourced CO_2 from the atmosphere since the beginning of the industrial revolution. A proportion of the carbon uptake is exported via the four ocean 'carbon pumps' (Solubility, Biological, Continental Shelf and Carbonate Counter) to the deep ocean reservoir. Increases in sea temperature and changing planktonic systems and ocean currents may lead to a reduction in the uptake of CO_2 by the ocean; some evidence suggests a suppression of parts of the marine carbon sink is already underway. While the oceans have buffered climate change through the uptake of CO_2 produced by fossil fuel burning this has already had an impact on ocean chemistry through ocean acidification and will continue to do so. Feedbacks to climate change from acidification may result from expected impacts on marine organisms (especially corals and calcareous plankton), ecosystems and biogeochemical cycles. The polar regions of the world are showing the most rapid responses to climate change. As a result of a strong ice–ocean influence, small changes in temperature, salinity and ice cover may trigger large and sudden changes in regional climate with potential downstream feedbacks to the climate of the rest of the world. A warming Arctic Ocean may lead to further releases of the potent greenhouse gas methane from hydrates and permafrost. The Southern Ocean plays a critical role in driving, modifying and regulating global climate change via the carbon cycle and through its impact on adjacent Antarctica. The Antarctic Peninsula has shown some of the most rapid rises in atmospheric and oceanic temperature in the world, with an associated retreat of the majority of glaciers. Parts of the West Antarctic ice sheet are deflating rapidly, very likely due to a change in the flux of oceanic heat to the undersides of the floating ice shelves. The final section on modelling feedbacks from the ocean to climate change identifies limitations and priorities for model development and associated observations. Considering the importance of the oceans to climate change and our limited understanding of climate-related ocean processes, our ability to measure the changes that are taking place are conspicuously inadequate. The chapter highlights the need for a comprehensive, adequately funded and globally extensive ocean observing system to be implemented and sustained as a high priority. Unless feedbacks from the oceans to climate change are adequately included in climate change models, it is possible that the mitigation actions needed to stabilise CO_2 and limit temperature rise over the next century will be underestimated.

1. Introduction

Through many natural processes and feedback mechanisms, the oceans[1] regulate climate on a range of timescales, from geological and millennial to decadal, interannual and shorter. Over the last two centuries, because of the ability of the oceans to take up heat and absorb greenhouse gases such as carbon dioxide (CO_2), they have partially buffered (neutralised) the effects of increasing levels of human-sourced greenhouse gases in the atmosphere. There is, however, clear evidence that many of the processes that contribute to this buffering role have been changing, in some cases almost certainly as a response to climate change. These processes provide a number of feedbacks that may be positive (reinforcing) or negative (ameliorating) to climate change.

There has been insufficient attention paid in the past to the key role that the oceans play in regulating climate and particularly to the feedback mechanisms that have the potential to and, in some cases, may already be intensifying climate change. For example, the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) in 2007 included as much information as was possible at that time on the ocean carbon cycle,

[1] All the oceans are interconnected and are often referred to in the singular. In this chapter, the plural version is generally used.

but recognised that many feedback mechanisms were incompletely included. This chapter explores the role of the oceans in regulating the climate and especially those changes that can accelerate climate change and have important implications for achieving stabilisation targets to mitigate climate change. Some of the key issues that are addressed are summarised below, followed by an outline of the structure of the chapter and a brief summary of conclusions and recommendations.

1.1. Heat budget

Comprising 97% of the Earth's water and covering 71% of the surface, the oceans are the main heat store for the world. Over the last few decades there has been a rapid and accelerating change in ocean temperatures and an increase in heat storage affecting seasonal and decadal variability in climate, heat transport, ocean circulation, stratification, biology and biogeochemistry. All of these ocean factors can lead to feedbacks to climate change.

The main positive feedbacks derive from rising temperatures and changing salinities. Higher temperatures are causing a loss of Arctic sea-ice, which feeds back to warming and climate change through many processes, including the potential release of the potent greenhouse gas methane. Changes in the oceans have led to an expansion of tropical/subtropical stratified (layered) waters, changing patterns of wind and altered ocean currents. Together these changes are likely to have led to a net reduction in the drawdown of CO_2 from the air into the ocean. However, expansion of the suboxic layers in the tropics and Atlantic Ocean (but not in the Indian Ocean) may, on the contrary, increase the preservation of organic matter and thus provide a sink for CO_2. A rising sea-level has also resulted from increasing temperatures through thermal expansion of the oceans, as well as shrinking polar ice sheets and glaciers. Some of these feedbacks may be compounded by the impacts of ocean acidification from CO_2.

1.2. Ocean circulation

Marked changes in salinity have been observed, reflecting an alteration in the hydrological cycle of the world through changes in precipitation, evaporation, river runoff and ice melt, with especially clear reductions in the North Atlantic, and in deeper waters and some upper layers of the Southern Ocean. Changes in ocean temperature have also been observed, with some regions warming very rapidly. Changes in buoyancy forcing (heat and salinity) and mechanical forcing (e.g. winds and tides) have the potential to change the large-scale circulation of the global ocean, including its overturning circulation and horizontal flows [Thermohaline Circulation (THC)/Meridional Overturning Circulation (MOC), commonly known as the 'global conveyor belt']. The general consensus from modelling projections for the twenty-first century is that there is likely to be a reduction in the strength of the Atlantic MOC by up to 50% of its current strength. This will not necessarily lead to a cooling of Europe, but more likely to a slower rate of warming, because the general atmospheric warming tends to dominate over the cooling expected from a reduced MOC.

Recent increases in the poleward ocean heat flux are likely to have played a central role in the decline of Arctic sea-ice. The signal from the changes in the Arctic has, and is expected to continue to, propagate south through subarctic seas on either side of Greenland, to modulate the Atlantic thermohaline overturning.

1.3. Tropical storms

The intensity of tropical storms has increased by 75% from 1970 to 2004 in the North Atlantic and western North Pacific and a global increase in their destructiveness is documented. The possible feedback role to climate change is still unclear, but it is expected that as global temperatures rise, storm intensity and possibly their frequency may increase.

1.4. Storage and transfer of CO_2

The oceans are the main store for the greenhouse gas CO_2, each year taking in about 40% of anthropogenic CO_2 from the atmosphere and exporting carbon via physical and biological processes to the deep ocean reservoir. Emissions of CO_2 from human sources have already grown to over 7 GtC (gigatonnes carbon) per year. The sensitivity of atmosphere/ocean fluxes of the carbon cycle is particularly evident. Increases in sea surface temperature (SST) and changing biological systems and ocean currents may lead to a reduction in the uptake of CO_2 by the oceans. Measurements taken over the last few decades of atmospheric greenhouse gases and ocean observations are indicating that a reduction in the buffering capacity of the oceans is underway in some regions. A slowing down of the ocean sink and any large change to the different ocean carbon pumps could lead to an acceleration of levels of atmospheric CO_2 and thus to intensified climate change.

1.5. Acidification

Through the uptake of nearly 50% of CO_2 produced by burning fossil fuel over time, the oceans have buffered the cause and effects of climate change. This large addition of CO_2 to the oceans has also had a profound effect on ocean chemistry. As CO_2 dissolves into the ocean, it reacts with seawater, forming carbonic acid which causes a reduction in pH (lower alkalinity), a process that has been termed 'ocean acidification'. Since the beginning of the industrial revolution, pH has reduced by $\sim$0.1 units (representing a 30%

increase in H^+ ions), a substantial amount considering that the units are logarithmic. Rapid acidification is expected to continue to the extent that in 50 years time the oceans are predicted to be less alkaline than at any time in the past 20, and likely 55, million years.

Feedbacks to climate change from ocean acidification may result from expected impacts on marine organisms, ecosystems and biogeochemical cycles. Planktonic plants (phytoplankton) comprise 50% of global primary production and play a crucial role in the uptake of CO_2 from the atmosphere. There is concern that oceanic organisms will not be able to adapt to the rate and scale of change now underway. These organisms are vital to the way the oceans draw down CO_2 from the atmosphere and play a profound role in the biological pump and the way it transfers CO_2 to the deep ocean store. In addition, the effects of projected changes in the pH of the oceans on corals and plankton community structure are likely to have profound implications for biodiversity, marine living resources and again with likely feedback to the carbon cycle.

1.6. Polar regions

The polar regions are thought to be especially susceptible to planetary-scale climate change, and a number of indicators of this have been observed. For example, there have been considerable reductions in Arctic sea-ice, rapidly rising temperatures at the Antarctic Peninsula, and a break-up of a number of Antarctic ice shelves. Arctic sea-ice has retreated rapidly in recent years, whereas Antarctic sea-ice has shown a more regional pattern of change—decreasing in some sectors, but increasing in others, and with an overall small increase. Much of the old multi-year ice in the Arctic has been discharged so that the ice now found there is thinner and younger. Sea-ice loss is acting as a trigger for further regional warming, potentially contributing to melting of the Greenland ice sheet and release of methane, a potent greenhouse gas. In the Arctic, release of methane from marine and terrestrial sources is particularly likely to contribute to positive feedback effects to climate change. In the Southern Ocean, the regional sea-ice changes have the potential to modulate the formation of dense waters, with implications for the uptake of CO_2 from the atmosphere, as well as oceanic fluxes of heat and freshwater. The carbon storage capability of the circumpolar Southern Ocean is reported to have decreased in recent decades, leaving more CO_2 in the atmosphere, although investigations are ongoing into this phenomenon.

If the regional average temperature rise above Greenland increases above some threshold, estimated as 3 °C above pre-industrial values (which equates to a global average temperature of 1–2 °C), it is projected that the ongoing contraction of the Greenland ice sheet would be irreversible. Without effective mitigation of carbon emissions, global warming could exceed this value during the twenty-first century, leading to a total melting of the ice sheet and a rise of several metres in sea-level over a timescale that is estimated to take centuries to thousands of years. The rate of loss of Arctic sea-ice was underestimated in the IPCC report in 2007, which, along with omission of some feedbacks, may have led to an underestimate of the cuts in emissions of greenhouse gases necessary to stabilise climate change at given atmospheric levels. The current rate of change in the Arctic, and its active feedbacks, have been triggered by a relatively small increase in global average temperature rise.

1.7. Plankton productivity, oxygen content and upwelling

Evidence is accumulating for increases in the intensity of upwelling in the major upwelling regions of the world, leading to a rise in phytoplankton production, anoxia and release of greenhouse gases. Anoxia is the lack of oxygen (O_2), an element that plays a direct and important role in the biogeochemical cycling of carbon and nitrogen. It is fundamental to all aerobic organisms, including those living in the dark deep sea. Areas of the ocean that stagnate can become anoxic due to the continual consumption of O_2 by living organisms. The main feedbacks to climate from plankton are via potential reductions in CO_2 drawdown and in the efficiency of the biological pump.

1.8. Microbes

The role of microbes in climate and climate change is crucially important, but little understood and poorly quantified, especially in terms of their contribution to biogeochemical and nutrient cycling, microbial diversity and feedbacks. A considerable increase in research effort is required to improve understanding of the impacts that microbes have on the planetary-scale climate system.

1.9. Nutrients

The contrast between biological and nutrient interactions within oceanic and terrestrial systems means that the oceans respond much more rapidly to climate change and feedbacks from oceanic biology. Therefore, biogeochemical interactions are likely to take effect more quickly. Strong regional changes in nutrients are expected in the future, dependent on variability in wet precipitation, evaporation, wave storminess, mixing and the depth of stratification. Precipitation is expected to increase especially in tropical regions. At present, it is not possible to predict future trends because of the localisation of the changes and our lack of knowledge of complex ecosystem interactions. It is also not clear how all the regional responses will add up to a global mean. The subtropical gyres play a large role in carbonate production and export to depth (carbonate and biological pumps) and are predicted to expand in area, but not in productivity, in a warming world.

1.10. Sea-level rise

Sea-level has been rising at the upper end of the IPCC AR4 projections and can contribute to coastal erosion, inundation and salinification of aquifers. Sea-level rise will affect humans in many ways, including the potential displacement of millions of people. Migration of populations and loss of coastal lands will likely lead to changes in land and resource use that have the potential to establish further positive feedbacks to climate change.

1.11. Structure of the chapter

The chapter has been organised into sections that reflect the main ocean drivers of climate change and the variables that contribute to them, as shown schematically in Fig. 1.1. Note that this figure focuses on factors interacting with nutrients; the real situation is more complex as the drivers may also directly impact other processes independently of nutrients. Denitrification may also act independently and be linked to atmospheric concentrations of CO_2. The physics starts the process with recycling feedbacks at all levels. The other sections examine key elements of ocean–climate interactions covering: Ocean Physics, Circulation and the Hydrological Cycle, Primary Production: Plankton, Light and Nutrients, the Oceanic Carbon Cycle, Ocean Acidification and Modelling. An additional special focus has been placed on the critically important, but still under-studied polar regions, with separate sections on the Arctic and Southern Oceans.

Throughout the chapter, our aim has been to provide an assessment of the key processes and feedbacks from the oceans to climate and climate change and, where possible, prioritise their importance. Gaps in knowledge are identified in modelling and research programmes, with a particular reference to observing systems that are needed to adequately assess the scale and speed of change. Some of the positive feedback mechanisms from the oceans to climate change have been insufficiently included in climate modelling and calculations for stabilisation targets. Without these, it is possible that the stabilisation targets for climate mitigation underestimate the action needed to limit global temperature rise within any given limit. The chapter also includes in places a discussion of tipping points (sudden, possibly irreversible changes that might lead to rapid climate change) and a brief discussion on iron fertilisation.

The work to produce this chapter was initiated by a Worldwide Fund for Nature (WWF) sponsored workshop in London during March 2008 that was attended by 30 international researchers who are experts in aspects

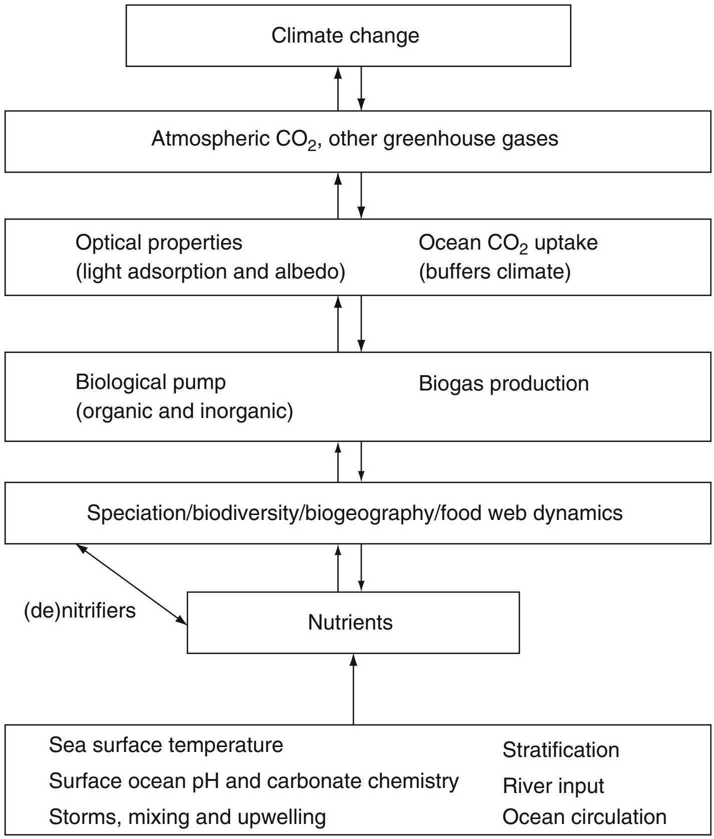

Figure 1.1 Schematic of potential relationship and links between key nutrient drivers and climate change (produced by Carol Turley, Plymouth Marine Laboratory).

of the field. A list of the participants and the themes addressed at the workshop are appended as an Appendix. The science of the chapter has built on the workshop outcomes, recent reports of the IPCC plus new information from the literature, as well as correspondence with experts selected to cover (where possible) all aspects of ocean science.

While other activities, such as fishing, whaling, pollution and habitat destruction, also impact the oceans, here we focus only on the interaction between the oceans and climate, without detailed account of these additional impacts. The extent to which positive feedbacks may lead to a potential acceleration of climate change is assessed. Where possible an update and expansion on ocean information covered by IPCC is included. The chapter aims to stimulate and inform debate, provide a useful complement to the work of IPCC and contribute to the preparations for the next

IPCC review. It is hoped that it will also be of value to other international and national organisations working on climate change and to the research and modelling community in helping to prioritise improvements that need to be included in future research, modelling and observing programmes.

1.12. Summary conclusions and recommendations

This chapter demonstrates that the oceans are vital in regulating our climate. They have buffered climate change substantially since the beginning of the industrial revolution, acting as a sponge to carbon dioxide and heat from global warming. While it was assumed this would continue, our chapter gives a warning—even at current warming levels to date, changes underway in our oceans may accelerate warming and its consequences to organisms, and have the potential to intensify climate change itself. In some examples, such as sea-ice loss, this process may already be underway.

A concerted effort to better understand the implications of the role of the oceans in regulating the climate is essential to better predict climate change. Where complete understanding is not possible, feedbacks from the oceans to climate change need to be taken account of when planning responses to climate change. It is necessary to apply the precautionary principle in both marine and climate management until a fuller understanding is achieved. Most ocean observing programmes are still funded from research budgets and, other than for some aspects of the physics, have a poor global coverage, especially for deeper waters and for biological and biogeochemical processes. Implementation of an improved ocean observing system is urgently needed to monitor changes in the interactions between the oceans and climate change.

2. Ocean Physics, Temperature, Circulation, Sea-Level Rise and the Hydrological Cycle

This section describes how the large changes that have taken place in SST, ocean heat content and salinity over the last century are altering ocean density, with effects on stability (stratification), circulation, mixing and feedbacks to the atmosphere. The consequences of these changes for sea-level, polar ice, the frequency and intensity of tropical storms (hurricanes, cyclones and typhoons) are then examined as are connections to the monsoons and modes of variability such as the El Niño/Southern Oscillation (ENSO). The physical changes in the oceans were well covered in the IPCC AR4 reports as much more is known about the physics of the oceans than other subjects and more data have been collected on temperature and salinity than any other variable.

Historically, the climate has undergone large natural change, independently of man's influence, at global and regional scales through geological time, at alternating time intervals ranging from millions to decadal to annual periodicity (Crowley, 1996; CLIVAR brochure: http://www.clivar.org/publications/other_pubs/latest_clivar_brochure.pdf). Natural climate variability can be forced by many factors including changes outside the Earth in the Sun and in the orbit of the Earth in relation to the Sun, and by natural events such as volcanic eruptions and oscillatory regional modes of variability such as El Niño, the North Atlantic Oscillation (NAO) and Pacific Decadal Oscillation (PDO) and the MOC (e.g. Chen *et al.*, 2008a,b; Keenlyside *et al.*, 2008; Shindell *et al.*, 2003). Natural changes may also occur very rapidly, as evident in the ice core record of Greenland where the return to cold glacial temperatures in the Younger Dryas abruptly changed around 12,000 years ago with a rapid rise in temperature of approximately 8 °C in less than a decade (Brauer *et al.*, 2008). Against this background, the rise in temperature over the last 50 years cannot be explained without including human forcing. Most of the warming since the mid-twentieth century was considered by IPCC AR4 to be very likely due to the observed increase in anthropogenic greenhouse gas concentrations (Alley *et al.*, 2007).

2.1. Changes in ocean temperature

2.1.1. Sea surface temperature

On a global scale, SST (the temperature of the upper few metres of the ocean) observations have shown a progressive warming trend of ~0.64 °C over the last 50 years. A steady increase has been recorded since 1910 other than an apparent peak centred on 1940 (Trenberth *et al.*, 2007). Thompson *et al.* (2008) have shown recently that this peak is an artefact due to sampling biases. Their results alter the variability, but not the long-term trend. Modelling studies predict that the trend in SST is likely to continue in the twenty-first century, with regional variability. The regional differences include enhanced warming in the Arctic, in the Indian Ocean and along the equator in the eastern Pacific, with a lower rate of warming in the Northwest Atlantic and in the Southern Ocean (Meehl *et al.*, 2007). Warming has been more pronounced in the Southern Ocean over the last 50–70 years (Gille, 2002, 2008) and has changed locally around the Antarctic Peninsula where the very rapid atmospheric warming has been paralleled by an increase in surface ocean temperature of >1 °C in summer months since the 1950s (Meredith and King, 2005).

Superimposed on the global trend are natural interannual and decadal variability. This is associated in the Atlantic, for example, with the NAO and the Atlantic Multi-decadal Oscillation (AMO), and in the Pacific with the PDO/ENSO. Regional variability may also be marked. In the North

Atlantic there is asymmetry across the basin with cooling in the Northwest and warming in the Northeast, until recently when the Northwest region also showed strong warming (Hughes *et al.*, 2008). In the tropical Pacific, there is a general warming trend, with reduced zonal patterns and more El Niño type east to west patterns of change.

2.1.2. Ocean heat content

The ocean's main role in climate variability and change is its huge capacity for the transport and storage of heat. On a global scale, ocean warming accounts for more than 90% of the increase in the Earth's heat content between 1961 and 2003 (Bindoff *et al.*, 2007). For the upper 700 m of the ocean (the water column from the surface to a depth of 700 m inclusive), the latter study estimates an average increase in temperature of 0.1 °C, equivalent to a flux of heat into the ocean of 0.2 ± 0.06 W m^{-2}. This large increase in heat storage has implications for seasonal and decadal variability in climate, transport and circulation by ocean currents, stratification, biology and biogeochemistry. All of these factors can lead to feedbacks to climate change.

Because of its fundamental importance, there have been many studies of changes in ocean heat content. These have revealed deficiencies in both historical and recent global ocean datasets. Analyses have demonstrated significant time-dependent biases in the expendable bathythermograph (XBT) data that dominates the historical archive since the early 1970s until the recent advent of Argo profiling floats. Wijffels *et al.* (2008) have shown that biases in the fall rate of XBTs are the dominant source of error and that they can be reduced substantially. In addition, the recent cooling of the ocean (Lyman *et al.*, 2006), reported following the introduction of the new Argo observing system, has now been shown to be incorrect and was a result of inadequate quality control in some of the new Argo floats as well as biases in XBTs (Willis *et al.*, 2007).

In addition to instrumental biases, there are also significant sampling problems associated with an inadequate ocean database. Palmer *et al.* (2007) demonstrated that the accuracy of heat content estimates can be improved by determining changes in heat content relative to an isotherm rather than a fixed depth level. They estimate a warming trend of 0.12 ± 0.04 W m^{-2} relative to the 14 °C isotherm.

Both the XBT instrumental biases and the sampling issues were addressed by Domingues *et al.* (2008). Compared to the assessment in the most recent IPCC report (Bindoff *et al.*, 2007), their improved estimate of upper-ocean warming is ~50% larger for 1961–2003 and ~40% smaller for 1993–2003. From 1961 to 2003, their estimate of heat flux into the upper 700 m of the ocean is 0.36 ± 0.06 W m^{-2}. The new results for near globally averaged anomalies of ocean heat content (Fig. 1.2) show similar multi-decadal variability to SST. They also reduce the large (but spurious) warming in the early 1970s and the subsequent cooling in the early 1980s that was a feature of previous estimates and which could not be reproduced in climate simulations. Climate models which include the full range of natural and anthropogenic forcing factors reproduce this observed long-term trend and the decadal variability, and demonstrate that violent volcanic eruptions are responsible for significant variability in ocean heat content (Fig. 1.2). However, this new analysis suggests that climate models may slightly underestimate the amount of ocean heat uptake in the upper 700 m for the period 1961–2003 (Domingues *et al.*, 2008).

Observations indicate that the deep and abyssal ocean may be absorbing large amounts of heat (Johnson and Doney, 2006a,b; Johnson *et al.*, 2007, 2008; Köhl *et al.*, 2007). Unfortunately, our historical observations and our current observing systems are inadequate to calculate quantitatively this storage on a global scale (Domingues *et al.*, 2008).

There are pronounced regional patterns in ocean warming, including indications of warming of the subtropical ocean gyres in both hemispheres and a poleward expansion of these gyres. For example, Palmer *et al.* (2007) suggest that the North Atlantic is a region of net heat accumulation over the period 1965–2004. Pronounced decadal variability is evident as a result of wind stress changes with a deepening of the North Atlantic subtropical gyre from 1981 to 2005 following an earlier period from 1959 to 1981 when the thermocline shoaled (Leadbetter *et al.*, 2007). There is also a significant deep warming near the poleward boundary of the subtropical gyre in the South Pacific Ocean (Roemmich *et al.*, 2007). Recent reanalysis of the sparse

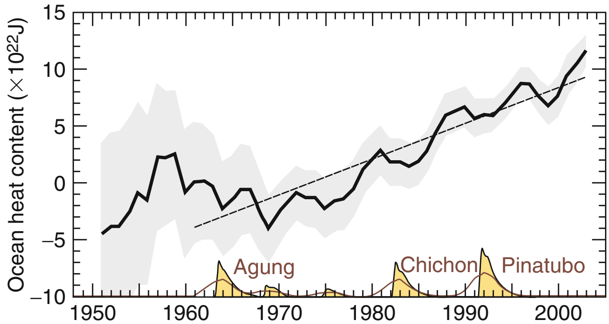

Figure 1.2 Upper-ocean heat content (grey shading indicates an estimate of one standard deviation error) for the upper 700 m relative to 1961. The straight line is the linear fit for 1961–2003. The global mean stratospheric optical depth (Ammann *et al.*, 2003) (arbitrary scale) at the bottom indicates the timing of major volcanic eruptions. The brown curve is a 3-year running average of these values, included for comparison with the smoothed observations. Figure modified from Domingues *et al.* (2008).

Southern Ocean dataset has revealed significant warming (Gille, 2008). However, much remains to be done to identify clearly these regional patterns to understand the dynamics underlying the changes and to evaluate the ability of climate models to simulate the variability.

2.2. Changes in salinity

One of the clear statements in the IPCC AR4 report is that while it is impossible to determine the precise origin of recent changes in regional patterns of freshening and salinification of the global ocean they are consistent with an enhanced hydrological cycle (Bindoff *et al.*, 2007). This is largely a consequence of the much smaller volume of observational data available for salinity compared to temperature, especially for the oceans in the Southern Hemisphere, which form two-thirds of the global ocean area. Salinity is, however, still the most measured property in the ocean after temperature and provides important information on the hydrological cycle, including rates of surface freshwater fluxes, transport and ocean mixing, all of which are important components of climate dynamics. Boyer *et al.* (2005) reinforced at a global scale the basin-wide message of Curry *et al.* (2003), who showed that a systematic freshening had occurred in high-latitude regions of the Atlantic at all depths in both the southern and northern hemispheres between the periods 1955–1969 and 1985–1999 (Fig. 1.3).

The freshening was especially pronounced in the intermediate depth waters of the Labrador Sea and in the deep outflows from the Nordic Seas via the Faroe–Shetland Channel and Denmark Strait. In contrast, higher salinities have been recorded in the intermediate depth (1000–1200 m) waters flowing out of the Mediterranean reflecting the rising deep water salinities recorded from this sea. It is expected that freshening will continue in the Arctic due to ice loss, but the Northwest Atlantic has undergone a rapid change to higher salinities post-1998 due to changes in the circulation of the sub-polar gyre (Hátún *et al.*, 2005; Holliday *et al.*, 2008) and increases in the salinity of the top 500 m have occurred in the subtropical gyre.

Freshening has also occurred in the subtropical gyres of the Indian Ocean (e.g. Bindoff and McDougall, 2000). In general, surface waters in the subtropical gyres of the Indian and Pacific Oceans have a higher salinity although there is evidence of freshening in the tropical Pacific (Delcroix *et al.*, 2007). A large-scale freshening of waters in the Southern Ocean close to Antarctica has been observed, including upper layer waters in the Ross Sea (Jacobs *et al.*, 2002) and Antarctic Bottom Water, adjacent to a large part of East Antarctica that is derived from the Ross Sea (Rintoul, 2007). Exact causes for the overall freshening are unknown, but glacial ice melt from the West Antarctic ice sheet has been suggested, along with changes in the sea-ice field of the Weddell Sea. Contributing factors to the changes in

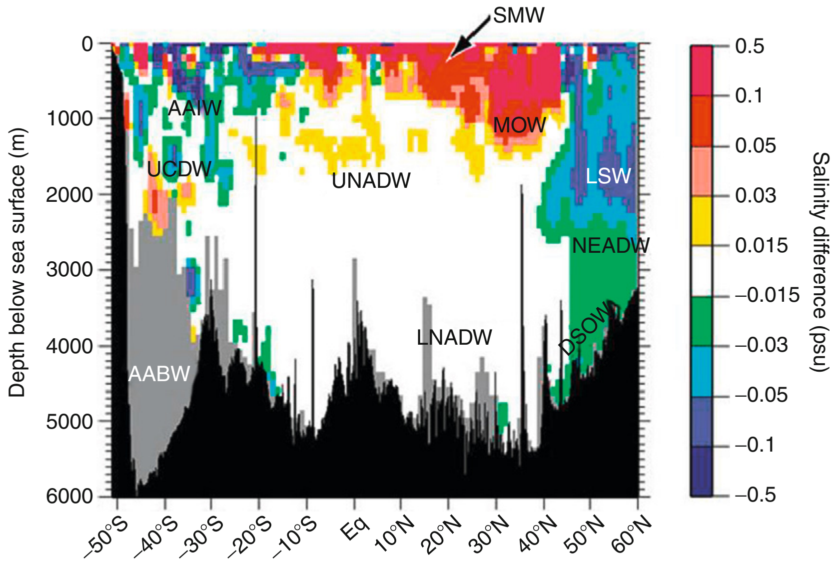

Figure 1.3 South-to-north vertical section of salinity versus depth for the western Atlantic basin, plotted as Salinity difference averaged for the period 1985–1999 minus 1955–1969. Grey colour means that sampling was not sufficient to estimate mean salinity. Acronyms are for the different water masses; see original paper. From Curry *et al.* (2003).

salinity are alterations in precipitation/evaporation, freshening from melting of ice, reduced ice formation and changes in ocean circulation.

The relative contributions of these factors to the large observed changes are still a matter of debate, although changes in evaporation/precipitation are shown to be important by Curry *et al.* (2003). The increasing differences in the salinity budgets of the Atlantic and Pacific suggest a change in the freshwater budget of the two basins. Bindoff *et al.* (2007) conclude that pronounced changes in salinity reflect a modification of the Earth's hydrological cycle with enhanced transport of water in the atmosphere between low and higher latitudes. Combined together, the salinity and temperature changes alter the density distribution and thus stability (stratification) as well as the THC of the ocean, with large potential feedbacks on regional climate and weather conditions such as temperature, storminess and rainfall patterns.

2.3. Global circulation

The world's large-scale ocean circulation is driven by a range of forcing mechanisms (e.g. winds, heating/cooling/salinity-density) and it is technically not possible to separate the currents based on their respective forcing. Nevertheless, there is a strong tradition in oceanography to consider the upper-ocean circulation as wind driven and that which reaches the deep oceans as density-driven. Thus, we often speak of the world's THC as the density-driven circulation that interconnects all the world's basins and all the ocean depths (see IPCC AR4, 2007 for a definition of the THC). The THC cannot be measured directly in contrast to the sinking and spreading of cold water through the MOC, which is an observable quantity.

Even so, the MOC is only observable in principle—in practice, it is prohibitively expensive to observe this circulation in all but a few limited places. This is rather restrictive because the MOC does not circulate in a pipe (in which case it only needs to be observed in one location), rather it recirculates vigorously both in the surface and deep ocean. So most of the inferences about the MOC are from indirect measurements taken from the far more abundant observations of temperature, salinity, pressure, altimetry, etc., rather than from direct current measurements. Changes in heat and freshwater storage can be used to derive changes in transport. Throughout this chapter both concepts (THC and MOC) are used, but what really is meant is the ocean's large-scale vertical and horizontal overturning circulation.

2.3.1. Meridional overturning circulation in the North Atlantic/Arctic

2.3.1.1. Subtropical measurements Direct measurements of the heat transport associated with the Atlantic MOC indicate a maximum transport at the 26.5°N latitude (e.g. Ganachaud and Wunsch, 2000). At the same latitude, the Gulf Stream component of the MOC is channelled through the Florida Strait, where robust transport measurements have been maintained since 1980 (Baringer and Larsen, 2001). This latitude has been suggested as one of the optimum locations at which to monitor the MOC – to both establish how the system varies naturally, and to seek evidence of any long-term change that may be underway.

Based on measurements from ship transects over the past six decades an apparent 30% reduction in the strength of the MOC was calculated by Bryden *et al.* (2005) (Fig. 1.4). This appears to be driven by an enhanced southwards re-circulation of upper waters by the subtropical gyre, and a compensatory reduction in the deep southwards return leg of the MOC fed by high-latitude, cold, dense waters. The size and rate of this reduction received much attention, exceeding the limits of projected changes for the same time period based on climate model simulations (see Section 8). In response to the ongoing threat of an abrupt MOC change, and the societal implications this could have for Europe, an international collaborative monitoring system was launched to provide a continuous record of Atlantic MOC strength at 26.5°N, as part of the UK-led RAPID Climate Change Programme. Initial results reveal significant short-term (daily) variability in the strength of the MOC implying that the decrease evident in the ship-based

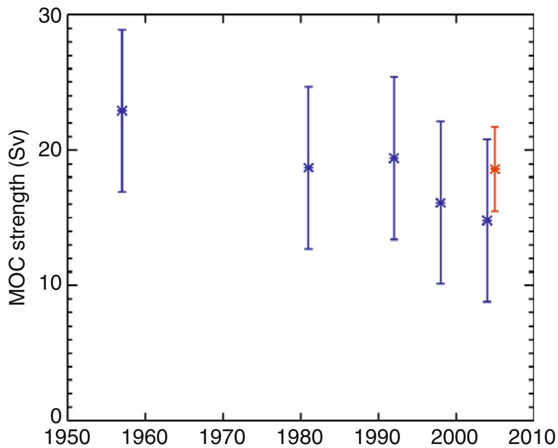

Figure 1.4 Mean strength of the Atlantic MOC at 26.5°N between 1957 and 2005 and associated error bars. Blue data points are for measurements taken from ships (Bryden *et al.*, 2005). The red data point is an average of observations taken in the first full year of the RAPID monitoring array, plus error bar (Cunningham *et al.*, 2007). Units are Sv (1 Sv = 1 million $m^3 s^{-1}$ of water passing the 26.5°N line). Values indicate a northwards net transport for water shallower than 1000 m.

measurements may be, at least in part, an artefact due to high-frequency 'noise' (Cunningham *et al.*, 2007).

2.3.1.2. Arctic/subarctic measurements The oceanic exchanges of surface and deep waters 'that connect the Arctic and Atlantic oceans through Subarctic Seas are of fundamental importance to climate' (Dickson *et al.*, 2008). In particular, changes that have taken place in the poleward ocean heat flux are likely to have played a central role in the decline of Arctic sea-ice (see Section 6). The signal from the changes in the Arctic has, and is expected to continue to, propagate south through Subarctic Seas on either side of Greenland, to modulate the Atlantic thermohaline 'conveyor' (Dickson *et al.*, 2008). To measure these changes lines of moorings, supplemented in the last decade by ADCPs (Acoustic Doppler Current Profilers) and other measurements between (1) Iceland and Greenland, (2) Iceland and the Faroe Islands, (3) The Faroe Islands and Shetland, (4) Greenland, Spitsbergen and Norway, and more recently (5) in the Canadian Archipelago, have been in place for some years through the Arctic–Subarctic Flux Study (ASOF) (see http://www.asof.npolar.no) and its predecessors (Fig. 1.5). The aim of ASOF was to observe the inflow and outflow of water to and from the Arctic.

A successor integrated Arctic/Subarctic Seas international programme (The integrated Arctic Ocean Observing System, iAOOS) is now in place as part of the International Polar Year (Dickson, 2006; Dickson *et al.*, 2008).

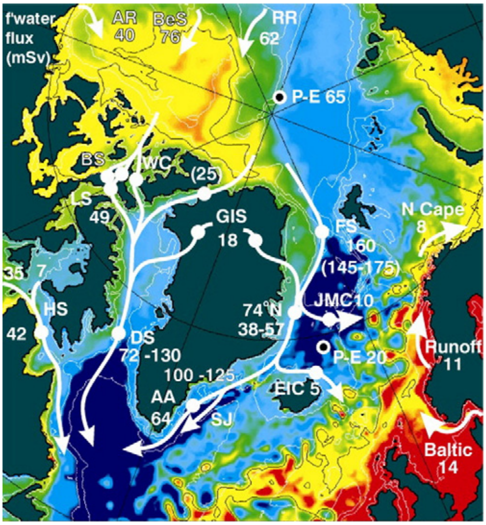

Figure 1.5 Estimates of freshwater flux relative to $S = 34.8^{\star}$ in Arctic and Subarctic Seas as determined during the ASOF project. Units are mSv and the base map is a snapshot of modelled sea surface height courtesy W. Maslowski, NPS, Monterey (1 mSv = 31.546 km^3 year^{-1}; * the numbers for P_E, runoff and ice melt are independent of the choice of reference salinity). From Dickson *et al.* (2007).

The longest current meter records presently just exceed a decade, so it is difficult to determine any evidence for a long-term trend.

There has been a pronounced increase in heat transport to the Arctic in the last 10 years (Holliday *et al.*, 2008; Hughes and Holliday, 2007), with the maximum being reached 5 years ago and with another pulse of heat on its way. As the warmer water delivered to the Arctic is leaving already, the total heat content in the Arctic is slightly decreasing, but with high interannual variability (Dickson *et al.*, 2008; Schauer *et al.*, 2008).

2.3.2. Meridional overturning circulation in the Southern Ocean/Antarctica

The Southern Ocean is a key region in the THC/MOC where the products of deep convection in the North Atlantic are upwelled and mixed into shallower layers. These waters are then converted into shallow and deep return flows to complete the overturning circulation (see Section 7).

Profound physical changes have been observed in the water masses of both the shallow and deep return flows. The shallow limb of the MOC is sourced towards the northern flank of the Antarctic Circumpolar Current (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. These waters show variability in properties on a range of timescales (seasonal to decadal and longer), reflecting global and regional climate variability in their source regions. The formation and subduction of the mode and intermediate waters (Fig. 1.6) is believed to be a critical process that removes anthropogenically produced CO_2 from the atmosphere and likely contributes to

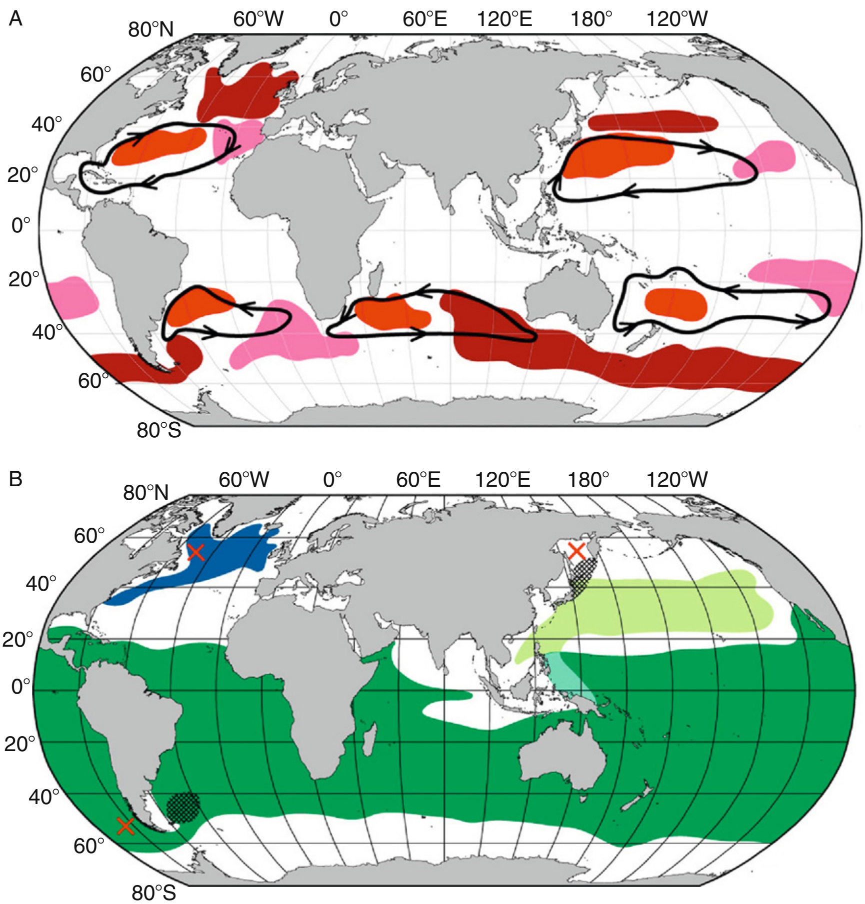

Figure 1.6 (A) Location of mode and intermediate waters in the global ocean. Low-density mode waters of the eastern subtropical gyres—pink. The highest density mode waters, which subduct in the subtropical gyres—red. Atlantic Sub-polar Mode Water, North Pacific central mode water and Subantarctic Mode Water (SAMW)—dark red. (B) Covering a large area of the ocean, intermediate waters are found below the mode water, Labrador Sea intermediate water (LSW)—blue, North Pacific intermediate water (NPIW)—pale green, Antarctic intermediate water (AAIW)—green. These waters eventually re-emerge at the surface far from their origin. Primary formation areas for the intermediate waters are indicated with red crosses. From Talley (1999): http://www-pord.ucsd.edu/~ltalley/papers/1990s/agu_heat/talley_agu_heat.html.

internannual variability in global oceanic uptake. For example, 40% of the global ocean inventory of anthropogenic CO_2 is found south of 30°S and most of that is in the intermediate and mode water (Sabine *et al.*, 2004a).

Changes have also been observed in the deep return flow of Antarctic Bottom Water (AABW), the deepest water on the Earth (Fahrbach *et al.*, 2006; see Fig. 1.3). This cold water forms via intense air/sea/ice interaction at the surface, sinks and then spreads northwards towards the Arctic. A freshening of the AABW has occurred off a large sector of East Antarctica that may in part reflect melting at depth (~700 m) of Antarctic glaciers that extend over the sea (see Section 7). The densest component of the AABW has shown a warming trend until very recently (Fahrbach *et al.*, 2006), while the less dense variety that can escape the Weddell Sea and penetrate north in the Atlantic has shown a marked decadal warming (Meredith *et al.*, 2008).

2.3.3. Slowing down of the MOC and cooling of NW Europe

The general consensus from modelling projections for the twenty-first century is that there is likely to be a reduction in the strength of the Atlantic MOC of up to 50% of its current strength. This will not lead to a cooling of Europe, but less warming. This is because the general atmospheric warming 'wins' over the cooling expected from a reduced MOC. The impacts associated with a reduced MOC are contained in the projections of global and regional climate change provided by the IPCC AR4 WG report. These include a continuation of already observed changes in precipitation that include droughts in the subtropics and increased rainfall in equatorial and high-latitude regions. The results indicate that it is unlikely that there will be a large abrupt change in the MOC during this period (Meehl *et al.*, 2007), although changes beyond 2100 cannot be confidently assessed.

2.4. Upwelling

Wind-driven Ekman pumping with the Coriolis force drives the four major eastern boundary upwelling regions of the world: Peru, Benguela, California and Northwest Africa, supplemented by a region off Northeast Africa in the Arabian Sea that is driven by monsoonal wind forcing. These regions are possibly the most productive locations in the oceans (Thomas *et al.*, 2004) due to the high concentrations of nutrients that are brought to the surface. Poleward divergence of water driven by the trade winds also causes upwelling to either side of the equator. Upwelling has a dual role in climate modulation as regions of strong outgassing of CO_2 and other greenhouse gases (Bakun and Weeks, 2004) and as areas where the biological pump is especially strong as a consequence of the high productivity and rapid sedimentation of planktonic material to the ocean floor. A further consequence of this productivity is a reduction in oxygen levels, with at times the establishment of extensive areas of bottom anoxia (Bakun and Weeks, 2004; Neretin, 2006; Tyson and Pearson, 1991).

It has been postulated, on the basis of palaeo-evidence, that increases in coastal upwelling and an intensified biological pump reduced levels of atmospheric CO_2 in the lead up to the Pleistocene glaciations (Berger, 1985). Bakun (1990) reported an intensification of equatorward alongshore winds and an associated upward trend in upwelling from the 1940s to 1988 in all four of the eastern boundary regions. He attributed the changes to rising global temperatures and predicted an increase in upwelling intensity as global warming progresses. A similar substantial increase in upwelling and a >300% increase in chlorophyll has occurred in the Arabian Sea due to intensified summer monsoon winds in recent years due to warming of the Eurasian landmass (Goes *et al.*, 2005). A modelling study by Hsieh and Boer (1992), however, suggests that upwelling may respond in the opposite way to that suggested by Bakun in a warming world. Their model analysis showed that reduced latitudinal gradients would lead to weaker upwelling and less productivity.

2.5. Changing physics of tropical seas in a warming ocean

SSTs in the tropics determine where the upward branch of the Hadley Circulation in the atmosphere is located over the oceans and the strength of the circulation is related to the ENSO (IPCC AR4, WG 1, 2007, p. 296). For example, the Asian-Australian (AA) Monsoon (see WCRP/CLIVAR flyer on the AA Monsoons, available from CLIVAR http://www.clivar.org/) is strongly influenced by changes in SST in the Indian Ocean that are modulated by ENSO. The potential effects of tropical seas on climate change have only been discussed briefly in this chapter and should form a follow-up study.

2.5.1. Tropical storms (hurricanes, cyclones, typhoons)

Tropical storms play a vital role in climate by pumping a considerable quantity of heat from the ocean into the atmosphere each year, by generating mixing that brings cold deep water to the surface and, through evaporation (Trenberth and Fasullo, 2007). During the storm, precipitation releases latent heat that is rapidly transported high into the atmosphere where it may radiate into space (Emanuel, 2006). These storms act as a release valve for solar heat caught above the sea in the humid, cloudy conditions of the summer tropics and are generated when surface water temperatures reach a threshold of ~26 °C over a depth of ~50–100 m. The contribution that tropical storms may make to climate change through feedbacks related to a possible increase in their frequency and intensity is still unclear.

The intensity of tropical storms has increased by 75% from 1970 to ~2004 in the North Atlantic and western North Pacific and a global increase in their destructiveness is documented by Trenberth *et al.* (2007). They also note that the first recorded hurricane ever to cross the coast of South America occurred in March 2004. Atlantic hurricane activity is highly correlated with SST and a rise of only 0.5 °C can lead to an increase of ~40% in hurricane frequency and activity (Saunders and Lea, 2007). Regional variability in the occurrence of tropical storms is closely linked to ENSO and decadal environmental changes so that there is often an alternation between basins in the number of storms. For example in El Niño years hurricane intensity decreases in the North Atlantic, far west Pacific and Australasian regions, but increases in the remainder of the Pacific. As global temperatures rise it is expected that precipitation will be enhanced, as well as the extent of the geographical area suitable for seeding storms so that global storm intensity and possibly frequency will likely increase.

2.6. Sea-level rise

Sea-level rise is a major impact of climate change. Ocean thermal expansion was an important component of sea-level rise during the latter half of the twentieth century and models project it is likely to be the largest contributing factor in the twenty-first century.

During the Pleistocene sea-level varied from metres above to over 120 m below present-day values as major ice sheets waxed and waned, particularly in the Northern Hemisphere (Berger, 2008). At the time of the last interglacial period about 125,000 years ago, sea-level was likely 4–6 m higher (Overpeck *et al.*, 2006) than it was during the twentieth century, at polar average temperatures 3–5 °C higher than present values.

The Third IPCC Assessment Report, TAR (Church *et al.*, 2001), reported that during the disintegration of the Northern Hemisphere ice sheets at the end of the last glacial maximum, sea-level rose at an average rate of 1 m per century, with peak rates of about 4 m per century.

In the longer term, these ice sheets have the potential to make the largest contributions to sea-level rise and there is increasing concern about the potential instability of the West Antarctic and Greenland ice sheets.

The current projections of sea-level rise are based on the SRES emission scenarios. However, global emissions are already above (Canadell *et al.*, 2007; Raupach *et al.*, 2007) the highest of these scenarios and well above stabilisation scenarios of twice pre-industrial values. Since the start of the IPCC projections in 1990, sea-level is actually rising at near the upper end of the highest IPCC Third Assessment Report projections of 2001 (Rahmstorf *et al.*, 2007).

There will also be regional changes in sea-level with some areas showing a decrease relative to the global average rise, due to circulation changes, but there is little understanding of such variability. One regional change that is likely to have a substantial impact is that many deltaic regions around the world are sinking as a result of reduced sediment supply, compaction of sediments and water (and/or oil or gas) extraction.

Sea-level rise will be felt most acutely through extreme events, such as Hurricane Katrina and Cyclone Nargis. Rising sea-level on its own (without any change in the intensity or frequency of extreme weather driving coastal storm surges) will result in extreme sea-level thresholds of a given value being crossed more frequently. This change in frequency can be pronounced. Any change in the frequency or intensity of meteorological conditions will also change the frequency/intensity of extreme sea-level events.

2.7. Destabilisation of ice sheets/glaciers

It is possible that rising sea-levels might destabilise buttressing ice shelves and/or increase the proportion of glaciers that float. A retreat of the grounding line of these glaciers may allow ice streams to speed up and potentially contribute to a large discharge of ice from an ice sheet although the mechanisms involved are still little understood. Recent research, however, has documented the production of a wedge of sediment that stabilises the position of the grounding line indicating that sea-level rise may be implicated in recent retreats (Alley *et al.*, 2007; Anandakrishnan *et al.*, 2007). A combination of basal melt and rising sea-level might, however, allow seawater to extend into sub-ice sheet basins that are presently isolated from the sea and lead to accelerated subsurface melting (D. Martinson, personal communication). Enhanced submarine melting causes the grounding line of glaciers to retreat, reduces the buttressing of frontal ice on inland ice, and allows faster rates of ice flow to the sea (Thomas, 2004).

The melting of the glaciers in the western Antarctic Peninsula is more influenced by rising temperature than by changes in sea-level. It is unclear if ocean temperature or air temperature is the more important factor, but the ocean has a larger heat capacity and is in subsurface contact with the ice. Half a degree of temperature change in ocean temperature is more significant than half a degree change in air temperature. The recent decadal warming of the ocean adjacent to the western Antarctic Peninsula (>1 °C in summer months since the 1950s) is mooted to have played a significant role in the retreat of its tidewater glaciers (Meredith and King, 2005).

Surface melt of the Greenland ice sheet has increased and is projected to increase at a faster rate than additions from higher precipitation as temperature rises. If Greenland air temperatures rise an average of 3 °C, it is predicted that the ongoing contraction of the ice sheet may be irreversible (ACIA, 2005). Global warming could exceed this value during the twenty-first century without effective mitigation of emissions. If these temperatures were maintained, they would lead to a virtually complete elimination of the Greenland ice sheet and a contribution to sea-level rise of up to about 7 m in the coming centuries to thousands of years.

Some recent observations suggest a (rapid) dynamic response of the Greenland and West Antarctic ice sheets (WAIS) that could result in an accelerating contribution to sea-level rise. This is only included in an *ad hoc* fashion in the current IPCC projections. For the Greenland ice sheet, this is hypothesised to involve surface melt water making its way to the base of the ice sheet and lubricating its motion enabling the ice to slide more rapidly into the ocean. Glaciers in Greenland are already retreating. The sea-ice there shows no buttressing in the way that it does in Antarctica. Even so, sea-ice in the Greenland Sea has rapidly decreased over the last two decades and the Oden ice tongue between 70 and 75°N has disappeared. In Antarctica, the WAIS is grounded below sea-level, allowing warmer ocean water to melt the base of the ice sheet and potentially leading to significant instability. Understanding of these processes is limited. As a result, they are not adequately included in current ice sheet models and there is no consensus as to how quickly they could cause sea-level to rise. Note that these uncertainties are essentially one sided. That is, they could lead to a substantially more rapid rate of sea-level rise but they would not lead to a significantly slower rate of sea-level rise.

Current projections suggest that the East Antarctic ice sheet will remain too cold for widespread surface melting and that it is expected to gain mass from increased snowfall over the higher central regions. Net loss of mass could occur, however, if there was a more rapid ice discharge into the sea around East Antarctica due to a higher rate of accumulation from snowfall over the interior or due to a warming of the coastal waters in contact with the glaciers. The latter would increase submarine melting, which in turn would release the grounded glaciers from their bed and allow them to flow faster towards the sea.

2.8. Concluding comments

- Global SST has shown a progressive increasing trend over the last century with warmer water extending into the Arctic and parts of the Southern Ocean adjacent to Antarctica.
- There has been a large increase in the heat content of the ocean down to 700 m depth.
- The deep ocean appears to be absorbing heat at an increasing rate, but the amount of heat stored is inadequately quantified because of poor sampling.
- Pronounced changes in salinity have occurred in many regions of the world, likely reflecting a modification of the Earth's hydrological cycle.
- Combined together, the salinity and temperature changes alter the density distribution, stratification and THC/MOC with large potential feedbacks to climate.
- There is clear evidence of large changes and pronounced daily to decadal variability in the MOC in different areas of the world.

- It is not possible at present to say if these changes on a global scale are a consequence of a reduction in the strength of the circulation due to climate change.
- The general consensus from modelling projections for the end of the twenty-first century is that there is likely to be a reduction in the strength of the Atlantic MOC by 0–50% of its current strength. This will not lead to a cooling, but less warming in Europe, with perhaps more warming in the tropics.
- It is unlikely that there will be a large abrupt change in the MOC during the next century, although changes beyond 2100 cannot be confidently assessed.
- There is evidence for increases in the intensity of upwelling, leading to large increases in phytoplankton production, anoxia and release of greenhouse gases.
- The intensity of tropical storms (hurricanes, cyclones, typhoons) has increased by 75% in the North Atlantic and western North Pacific and a global increase in their destructiveness is documented.
- With rising sea temperature and enhanced precipitation, the area for seeding tropical storms will expand possibly leading to an increase in storm frequency and intensity.
- Since the start of the IPCC projections in 1990, sea-level is rising at near the upper end of the highest IPCC Third Assessment Report projections of 2001.
- Historical evidence adds credibility to the possibility of an increase in the rate of sea-level rise at the upper end of and beyond IPCC projections.
- If global average temperature in Greenland increases by ~3 °C above pre-industrial values, a level that could be reached during the twenty-first century without effective mitigation of emissions, the ongoing contraction of the Greenland ice sheet may not be reversible and could result in several metres of sea-level rise over hundreds or thousands of years.
- Recent rapid dynamic responses of the Greenland and West Antarctic ice sheets might result in a future accelerating contribution of their ice melt to sea-level rise.
- Feedbacks to climate change from sea-level rise are uncertain.
- Feedbacks from sea-level rise can accelerate ice sheet loss at the coast.

3. Primary Production: Plankton, Light and Nutrients

Microscopic marine phytoplankton form the base of the marine food web. They use energy from the Sun to fix CO_2 and account for around 45% of global primary production. Most of the organic carbon formed is consumed by herbivores or respired by bacteria, the remainder, about 35% (16 Gt, Falkowski *et al.*, 1998; 11 Gt, Denman *et al.*, 2007; Fig. 1.7), sinks

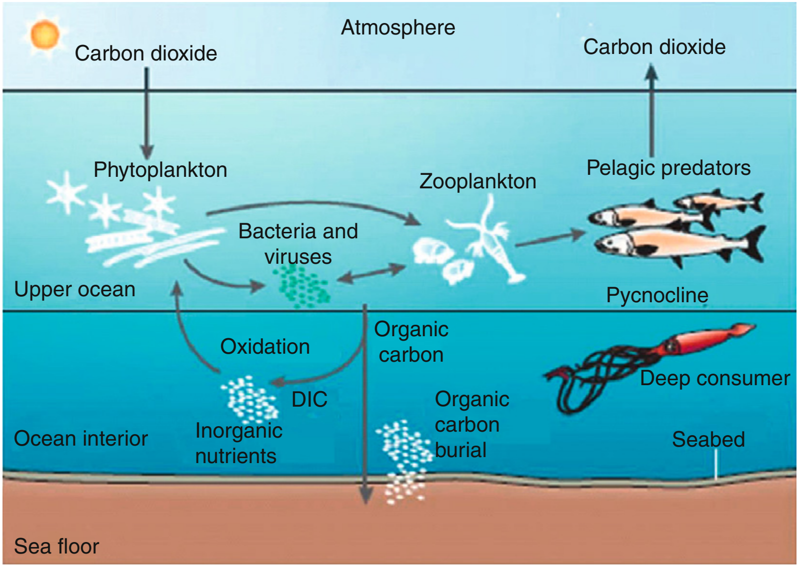

Figure 1.7 Cartoon of the Biological pump modified from Falkowski and Oliver (2007). Note that CO_2 is emitted from all heterotrophic organisms (e.g. zooplankton, fish and squid) and O_2 is produced by phytoplankton as well as other gases such as methane and DMS.

below the upper sunlit layer every year. This section addresses the contribution that planktonic and benthic organisms make to carbon cycling in the ocean with a commentary on the biogeochemical and other controls on primary production. An attempt is made to synthesise and prioritise potential feedbacks to climate change from the many complex processes involved. It should be remembered that any feedbacks to climate are now taking place against a background of a very changed biology that has been impacted by eutrophication and hypoxia (Diaz and Rosenberg, 2008), removal of top predators (Pinnegar *et al.*, 2000) and overfishing (Myers and Worm, 2003).

3.1. Oceanic primary production

Production of atmospheric oxygen and fixation of carbon during photosynthesis by phytoplankton enables the Earth to support a rich diversity of marine life and has strongly influenced changes in climate through geological time (Diaz and Rosenberg, 2008; Mackenzie and Lerman, 2006). Phytoplankton biomass and primary production is determined by light availability and access to nutrients (nitrogen, phosphate, silicic acid, iron) as well as grazing and viral lysis. Light varies with the angle of solar insolation (latitude), season, cloud cover, level of water clarity and mixing and is variably absorbed by different phytoplankton pigments. Superimposed on these growth-limiting factors is a physical regulation by ocean circulation, mixed-layer dynamics and upwelling. Since ~1% of light penetrates to 100 m (a very small proportion may reach as far as 1000 m) in the open ocean, and in productive coastal seas may only extend to 30 m or less, photosynthesis is confined to this upper layer.[2] The contributors to primary production vary from cyano- and eubacteria [e.g. *Synechococcus*, *Prochlorococcus*, *SAR 11* (SAR 11: a dominant cluster of marine bacterial phylotypes first described from the Sargasso Sea)] and eukaryotic picoplankton (0.2–2 μm in size), especially in tropical and oligotrophic oceanic waters, to eukaryotic nannoflagellates (2–10 μm) elsewhere with larger eukaryotic phytoplankton (10 to ~150 μm) such as diatoms and dinoflagellates forming an important component of the biomass in upwelling regions and in boreal and temperate seas. A new paradigm for primary production now exists (see Fig. 1 in Karl, 2007), which includes the above new microbial contributors as well as photolysis (PL) of dissolved and particulate organic matter by sunlight (Fuhrman *et al.*, 2008; Karl, 2007). This means that total primary production is likely to exceed the traditional view of chlorophyll-based gross primary production.

Plankton also plays a key role in the Biological pump (see Section 4) that moves organic and inorganic carbon to the deep ocean. Grazing and recycling of nutrients by zooplankton, bacteria, archaea and viruses including reprocessing and packaging of planktonic detrital material as it sinks through the water column, are key processes in determining the export rate of C fixed by primary production (Steinberg *et al.*, 2008; Yamaguchi *et al.*, 2002). Viruses (also fungi) have an important role as terminators of plankton blooms and because of their role in the mortality of marine organisms are key players in nutrient and energy cycles and in the structuring of microbial communities (Suttle, 2007). It is believed that changes in the relative strengths of the two fluxes (Primary Production and export flux) strongly influence climate and have been responsible for many of the changes in climate in the geological past (Falkowski *et al.*, 1998). The plankton also change surface albedo, increase retention of heat in the upper ocean by absorbance and contribute to the production of other potent greenhouse gases such as methane and nitrous oxide, and to reactive gases, such as dimethylsulphide (DMS) and halocarbons.

It is worth noting here some important differences between oceanic and terrestrial ecosystems. Most marine organisms are small, have rapid turnover times, are able to react quickly to changes in temperature, and are easily distributed by changing ocean currents in contrast to their terrestrial equivalents (Sarmiento *et al.*, 2004). In the upper waters of the open ocean,

[2] http://oceanexplorer.noaa.gov/explorations/04deepscope/background/deeplight/deeplight.html

temperature and nutrients are strongly related (Kamykowski and Zentara, 2005a,b); this is not the case on land. As a consequence, oceanic organisms react more directly to changes in seawater composition resulting from climate change than terrestrial systems, and feedbacks from oceanic biology and biogeochemical interactions are likely to take effect more quickly.

3.2. Microbial plankton

A number of major scientific breakthroughs have greatly improved understanding of oceanic microbial diversity and ecology over the last decade (Karl, 2007). Using genetic sequencing, it has been possible for the first time to determine the microbial (bacterial, archaeal and protist) composition of seawater samples (e.g. Fuhrman and Davis, 1997). One litre of seawater may contain as many as 20,000 species of bacteria, but only a very few ~20 dominate. The remainder form what is termed the 'rare biosphere' (Karl, 2007; Sogin *et al.*, 2006). Archaea are also an important component of the picoplankton in shallow and deep waters (Herndl *et al.*, 2005; Massana *et al.*, 2000). It is estimated that the global oceans contain approximately 1.3×10^{28} archaeal cells, equivalent to 30–40% of the estimated abundance of bacteria (DeLong, 2007; Quiñones *et al.*, 2009). By combining genetic and isotope techniques with membrane lipid research, a range of new ecological roles for microbes in the biogeochemical cycling of C, N, S, Fe and many other trace elements have been demonstrated that are important to climate change. These include anoxic oxidation of methane (AOM) (Boetius *et al.*, 2000; Michaelis *et al.*, 2002; Stadnitskaia *et al.*, 2008), anaerobic ammonium oxidation (Anammox process) that releases nitrogen gas from the oceans (Galán *et al.*, 2008; Jensen *et al.*, 2008; Kuypers *et al.*, 2003; Sinninghe Damste *et al.*, 2002), ammonia as an energy source for Crenarcheota so that they function as chemolitho-autotrophic organisms in the nitrogen cycle (Agogue *et al.*, 2008; Konneke *et al.*, 2005; Nicol and Schleper, 2006), the discovery of new nitrogen fixing microbial organisms in the oceans (Montoya, 2004; Zehr *et al.*, 2008), use of light as an energy source enabling massive fixation of CO_2 by SAR bacteria (Eiler, 2006), widespread anoxygenic photoheterotrophy in marine bacteria, including bacteria with bacteriochlorophyll and with proteo-rhodopsin (PR) (Beja *et al.*, 2002; Eiler, 2006; Gomez-Consarnau *et al.*, 2007; Moran and Miller, 2007), close syntrophic partnerships of anaerobic methane oxidising archaea and sulphate-reducing bacteria (Boetius *et al.*, 2000; Pernthaler *et al.*, 2008). These new findings demonstrate the crucial importance of microbes in climate and climate change and highlight a virtual complete absence of understanding of how microbial systems will change and impact biogeochemical cycling with climate change. Developing an understanding of the role of microbial diversity and functioning in biogeochemical and nutrient cycling is a major challenge for the future.

3.3. Phyto- and zooplankton

The distribution, abundance, production and biodiversity of different plankton species/groups are likely to be profoundly affected by projected climate-driven changes in the physical and chemical properties of the ocean including circulation, stratification, nutrients, light, trace metals (e.g. iron) and carbonate chemistry (Richardson, 2008). The converse will also occur as changes in plankton ecology and biodiversity can have a large and rapid impact with important feedbacks to climate variability through their role in the Biological, Continental Shelf and Carbonate Counter carbon pumps (see Section 4). Despite this importance, variability of these organisms on a global scale, other than for satellite measurements of chlorophyll, has been poorly studied. Long-term plankton observation programmes, other than the Continuous Plankton Recorder (CPR) survey (see http://www.sahfos.ac.uk/) in the North Atlantic and Southern Ocean, are non-existent in many oceanic regions of the world.

At a global scale there is a strong negative relationship between satellite-derived primary production and SST (Behrenfeld *et al*., 2006), but see the qualification of Sarmiento *et al.* (2004), that reflects the closely coupled relationship between ocean productivity and climate variability. One of the main reasons for this coupling is that the availability of nitrate (the principal nutrient limiting phytoplankton growth in much of the ocean) has been found to be negatively related to temperatures globally (Kamykowski and Zentara, 2005a,b). In the North Atlantic and over multi-decadal periods, changes in phytoplankton species and communities have been associated with Northern Hemisphere temperature trends and variations in the NAO index (Beaugrand and Reid, 2003). While at the interannual timescale correlations between temperature and phytoplankton are weak, due to high variance inherent in phytoplankton populations, at decadal intervals they are well correlated. Over the whole Northeast Atlantic there has been an increase in phytoplankton biomass in cooler and a decrease in warmer regions (Richardson and Schoeman, 2004). This relationship is likely to be a trade-off between increased phytoplankton metabolic rates caused by higher temperatures in cooler regions and a decrease in nutrient supply in warmer regions (Doney, 2006). The floristic shifts associated with this warming move a diatom-based system towards a more flagellate-based one (Leterme *et al*., 2005). In this scenario, however, it is assumed that the carbon sequestration will be less efficient because, unlike boreal diatom-based systems, much of the flagellate and nanoplanktonic production is remineralised near the well-mixed surface. In a warming ocean, microbial activity is also likely to be faster leading to more rapid recycling of carbon and a less efficient Biological pump. However, due to the underlying complexity of biological communities and their quite often non-linear responses to environmental variability, this makes predicting both floristic and faunistic changes and their feedbacks to climate fraught with uncertainty.

A number of groups in the plankton, including some benthic larval stages, have calcareous body parts (calcite, Mg calcite and aragonite). These organisms, including the algal coccolithophores (photosynthetic) and protist foraminifera (some of which are photosynthetic through symbiosis) are major contributors to pelagic carbonates especially in the subtropical gyres. Calcification (see Section 5) has the opposite effect to primary production, releasing some CO_2 to the water that may outgas adding to the CO_2 concentration in the atmosphere.

The effects of projected changes in the pH of the oceans (acidification), in combination with rising temperature, on plankton community structure and calcification/dissolution processes are likely to have profound implications for biodiversity, living marine resources and again with likely feedback to the carbon cycle (see Sections 4 and 5). At the microbial level, the dominance of different picoplankton taxa may be affected by changes in pH and temperature with impacts on food web structure, especially in oligotrophic waters (Fu *et al.*, 2007).

Changes in ecosystem composition, that appear to be driven by climate variability, are already underway: examples include desertification around the Mediterranean Sea (Kéfi *et al.*, 2007), shifts in North Atlantic plankton biomass (Beaugrand *et al.*, 2002), regime shifts in the North Pacific and North Sea (Chavez *et al.*, 2003; Reid *et al.*, 2001; Roemmich and McGowan, 1995) and observed shifts in phytoplankton pigment distributions as seen from satellite (Alvain *et al.*, 2008). Large productivity crashes are also associated with ENSO events in the Pacific (e.g. Lavaniegos and Ohman, 2007; Peterson *et al.*, 2002). These changes when seen together indicate that marine ecosystems are reacting beyond what might be expected from interannual variability. The changes also indicate that there may be locations in the ocean that act as biological hot spots that interact with climate change. If this is true, the identification and monitoring of such locations might be crucial for biodiversity, the maintenance of marine ecosystem goods and services, and for carbon drawdown.

As oceans warm, primary productivity is peaking earlier in the season in some areas, with less distinctiveness between the spring, summer and autumn seasons (Edwards *et al.*, 2001; Reid, 2005). Other parts of the food chain are shifting geographically poleward in response to thermal stimuli, rather than tracking the seasonal shift in primary productivity (Beaugrand *et al.*, 2002; Mackas *et al.*, 2007). These changes are creating a ‘mismatch between trophic levels and functional groups’ with implications for ocean–climate interactions and living marine resources (Edwards and Richardson, 2004).

3.4. Chlorophyll and primary production

Chlorophyll *a* is the dominant pigment in phytoplankton. As it is relatively easy to determine, it has been widely used in field studies as an *in situ* measure of phytoplankton biomass. With the advent of satellite instruments such as the Coastal Zone Color Scanner (CZCS; 1978–1986) and Sea-viewing Wide Field-of-view Sensor (SeaWiFS; 1997 to the present) and development of algorithms for the estimation of chlorophyll *a* and primary production, we have a synoptic viewpoint that allows study of phytoplankton at the global scale (McClain *et al.*, 2004; Fig. 1.8). It should be noted that the satellites only measure phytoplankton in the top few metres of the ocean and do not reflect the estimated ~10% of marine primary production that takes place in the Deep Chlorophyll Maximum. In addition, it is becoming increasingly clear that chlorophyll *a*, on the basis of the new microbial evidence cited above, is not a good proxy for primary production.

Using SeaWiFS data for 1997–2006, Behrenfeld *et al.* (2006) showed that global chlorophyll and calculated net primary production (NPP) increased sharply at the beginning of the period and then declined. The dominant signal in the data reflected changes in the 74% of the global ocean that comprises low-latitude permanently stratified tropical and subtropical waters with mean annual SSTs above 15 °C. The pattern of change was highly correlated with SST and an index of ENSO climate variability. The link between ocean biology and the physical environment was shown to operate via warmer upper-ocean temperatures enhancing stratification, which reduced the availability of nutrients for phytoplankton growth and *vice versa*.

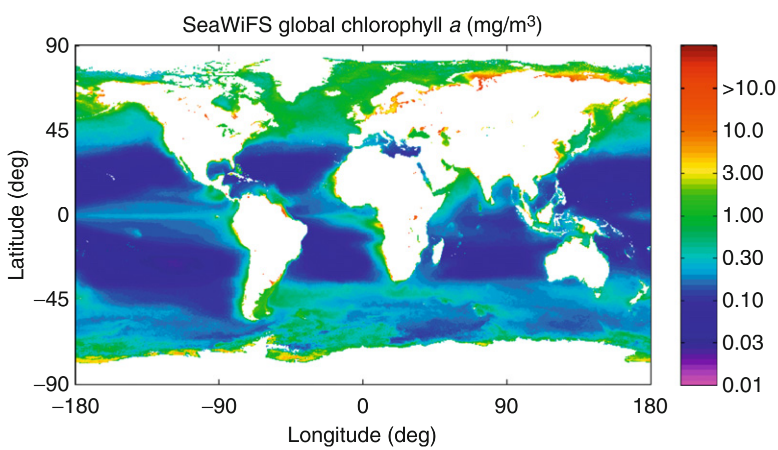

Figure 1.8 Global image of mean surface chlorophyll for the period 1998–2007. Processed from SeaWiFS data by Takafumi Hirata, PML.

Polovina *et al.* (2008), using the same SeaWiFS data, for the period January 1998–February 2007 have shown a similar expansion, especially in winter, of the least productive areas of the ocean, the subtropical gyres. In the North and South Atlantic and Pacific Oceans, outside the equatorial zone, approximately 6.6 million km^2 of former higher chlorophyll habitat have been replaced by low chlorophyll water. It appears that the subtropical gyres are expanding on their current position, becoming warmer and more oligotrophic and are likely to continue to expand as temperatures rise further. In the Southern Ocean, Le Quéré *et al.* (2002) show in a modelling study that the response of primary production to increased stratification and temperature varies regionally and that light availability is more important than temperature and dust as a forcing factor. The net effect for the whole ocean is not known.

If the reductions in NPP seen within the SeaWiFS period are extrapolated on the basis of projected changes in SST and nutrients, it could lead to a substantial reduction in the productivity of the oceans over the next 100 years. If these changes coincided with a reduction in the net input of CO_2 and a reduction in export fluxes, there could be a large impact on the biological pump.

A decade or more ago (e.g. Falkowski *et al.*, 1998) modelling had already predicted that primary production would reduce as stratification increased in the oceans. From a climate change perspective, it is important to note that the rates of expansion of the subtropical gyres measured by Polovina *et al.* (2008) already far exceed recent model predictions. As a corollary, Behrenfeld *et al.* (2006) also note that modelling has shown that changes in ecosystem structure (e.g. taxonomy, physiology and light absorption) due to climate variations may be as or more important than the changes in bulk integrated satellite measures of chlorophyll. The global observing systems needed to measure such variability are rudimentary and concentrated in the Northern Hemisphere at present.

A range of mesoscale processes such as eddies, fronts and their interaction with wind are important in mixing and bringing nutrients to the surface and may stimulate blooms (e.g. McGillicuddy *et al.*, 2007). Similar mixing may be induced by the passage of hurricanes (Son *et al.*, 2007). Understanding of mesoscale variability in oceanic waters and its potential impact on NPP, export flux and climate is poor.

3.5. Plankton biodiversity functional groups and ocean biomes

There are tens of thousands of different species of viruses, bacteria, archaea, cyanobacteria, phyto- and zooplankton and other organisms in the plankton. Together they play a key role in ecological and biogeochemical processes (Falkowski *et al.*, 2003) that modulate the cycling of CO_2.

These roles include the regulation of the settling flux of organic and inorganic carbon to the deep ocean and determining levels of reactive trace gases and aerosols in the atmosphere as well as proxy tracers for the carbon cycle such as O_2, $^{13}CO_2$ and $O^{18}O$. The biodiversity also contributes to variability in physical processes in the ocean, including temperature, stratification and mixing (Dewar *et al.*, 2006; Le Quéré *et al.*, 2005).

To manage the complex variety of these different plankton modes in modelling and other studies, assemblages of species are generally consolidated, for example, into ‘Plankton Functional Groups’ (Boyd and Doney, 2002; Jin *et al.*, 2006; Sarmiento *et al.*, 2004). Changes in the relative importance of different functional groups in the plankton can strongly impact the biological pump. Diatoms, relatively heavy and fast sinking, would be efficient in transporting sinking carbon to the deep ocean (Smetacek, 1998; Tréguer and Pondaven, 2000; Yool and Tyrrell, 2003), whereas nanoplankton (including the calcareous coccolithophores), being smaller as well as being rapidly ingested by heterotrophs, would be less efficient in the biological pump. Thus, relative fluxes of diatoms versus calcareous plankton have been implicated as one of the causes for the changes in CO_2 between glacial and interglacial periods (Harrison, 2000; Tréguer and Pondaven, 2000). Several feedback effects are conceivable: higher wind speeds, that favour diatom growth, would result in increased export (Le Quéré *et al.*, 2007, 2008); on the other hand, increasing temperatures, which favour the growth of smaller phytoplankton, are thought to reduce export flux. New developments in the interpretation of SeaWiFS data (e.g. Aiken *et al.*, 2008; Alvain *et al.*, 2005, 2006, 2008; Raitsos *et al.*, 2008) are making possible the identification of some phytoplankton functional groups on a global scale.

The next step in an analysis of biological variability in the ocean and its importance to climate change is to determine change in different ocean provinces (Biomes: Longhurst, 1998). Sarmiento *et al.* (2004) in a multi-model comparison study have divided the ocean into six different Biomes based on physical characteristics that reflect nutrient supply (Fig. 1.9). It is clear from carbon export studies in the Southern Ocean and elsewhere (Boyd and Trull, 2007) that there is a need to increase the number of Longhurst provinces with more precise/sensitive definitions based on multi-disciplinary information systems including plankton assemblage data (e.g. Beaugrand *et al.*, 2002; Devred *et al.*, 2007). To achieve this aim an improved knowledge is needed of chemistry, mesoscale properties, spatial and temporal variability in plankton composition and production versus recycling and export rates.

In a review by Le Quéré *et al.* (2005), plankton functional types (PFTs) were selected among other criteria on the basis of their clear biogeochemical role, quantitative importance in terms of biomass and production, well-defined environmental, physiological and nutrient control and biological

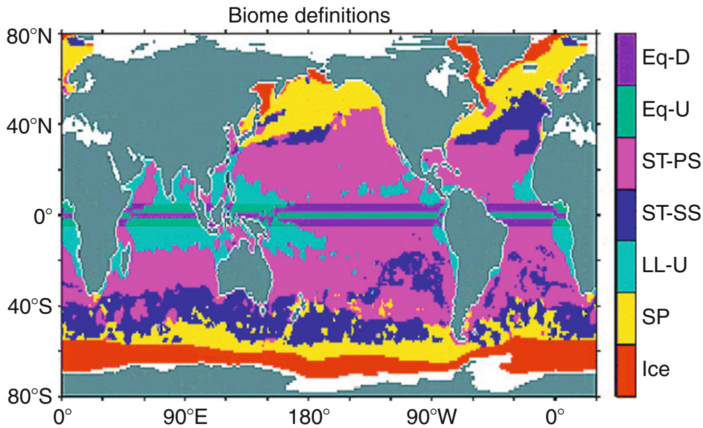

Figure 1.9 The distribution of six different ocean biomes: (1a) equatorial—downwelling (Eq-D), (1b) equatorial—upwelling (Eq-U), (2) subtropical gyre—permanently stratified (ST-PS), (3) subtropical gyre—seasonally stratified (ST-SS), (4) low latitude—upwelling (LL-U), (5) sub-polar (SP) and (6) marginal sea-ice (Ice). From Sarmiento *et al.* (2004).

interactions. A provisional grouping of 10 different PFTs that need to be simulated in a new generation of linked ocean physics and ecosystem models was selected by Le Quéré *et al.* (2005) to 'capture important biogeochemical processes in the ocean'. The groups selected are involved in, for example, bacterial remineralisation, N_2 fixation, phytoplankton calcification, silicification and DMS production, and include three different size fractions of zooplankton that contribute to export via a range of processes including faecal pellet and mucilaginous packages. The modelling and research strategy outlined by Le Quéré *et al.* (2005) addressed the urgent need to improve understanding of the interactions between the different types of plankton, food web structure and export efficiency of carbon. To successfully progress such a strategy will require a new level of international collaboration between modellers and marine ecologists, which has been historically absent.

3.6. Benthos

Benthic ecosystems play an important role in global carbon cycles as they are sites for remineralisation, burial and calcification. They also are places where many of the nutrients that sustain planktonic production are regenerated. While benthic studies have been carried out worldwide, they are largely confined to shelf waters and are spatially and temporally patchy. The inshore biota of Western Europe and parts of the coast of North America is well known but that of most of Asia, Africa and South America, especially in the tropics and subtropics, is poorly studied and the deep sea (around 97% of the world ocean) has scarcely been sampled. As a consequence it is difficult to quantify the role of the benthos in the global carbon cycle or to identify the regional areas that are most important. Furthermore, the majority of benthic studies have been focussed on larger bodied animals; sedimentary microbiology is a relatively young science and thus it is equally difficult to partition carbon cycling between the different elements of the benthic biota.

While there has been considerable concern about the impacts of ocean acidification on animals with calcareous body parts, some species appear to be able to increase the rate of calcification at a lower pH (Wood *et al.*, 2008b) although this is at some metabolic cost and may not be sustainable. Non-calcareous species are also likely to be affected as their physiology is finely regulated and has evolved to function within relatively narrow pH and CO_2 ranges (Michaelidis *et al.*, 2005). The scale of impacts on populations and assemblages resulting from a potential decline in growth and reproductive rates from ocean acidification has yet to be quantified.

3.7. Migration of plankton, fish and benthos towards the poles

A pronounced consequence of a warming ocean has been a poleward expansion of the range of many species in both the Southern and Northern Hemispheres. The resulting marked changes in community structure that are reflecting warming oceans and changes in circulation (e.g. Hátún *et al.*, 2009) have implications for the biological pump and CO_2 drawdown. Some of the strongest evidence for large-scale biogeographical changes in the oceans comes from the Continuous Plankton Recorder survey. In the Northeast Atlantic warmer water, zooplanktonic copepods have moved to the north by 10° latitude (1000 km) within 50 years while colder water plankton has retreated in the same direction (Beaugrand *et al.*, 2002). This represents a mean poleward movement of between 200 and 250 km per decade. The speed of this migration, due to advective processes, is more pronounced than any documented terrestrial study. Responses of zooplankton to changing water temperature have also been observed in the North Pacific (Mackas *et al.*, 2007), but it remains unclear if these are systematic responses to climate change or simply related to shifts in climatic state as reflected by major climate indices such as the Pacific Decadal Oscillation. Many species of fish have also shown apparently similar northerly range extensions in the eastern Atlantic and North Sea, at estimated rates that are up to three times faster than terrestrial species (Brander *et al.*, 2003; Perry *et al.*, 2005). One of the largest biogeographical shifts ever observed for fish species is the dramatic increase and subsequent northerly geographical spread of the snake pipefish (*Entelurus aequoreus*). Prior to 2003 this fish was confined to the south and west of the British Isles, but it now extends as far north as the Barents Sea and Spitzbergen (Kirby *et al.*, 2006; Harris *et al.*, 2007). The pelagic environment is of course three-dimensional and recent research has observed a movement of fish species towards deeper cooler waters in response to climate warming (Dulvy *et al.*, 2008). This change can be seen as analogous to the upward altitudinal movement of terrestrial organisms in alpine environments.

There is evidence for poleward migration of benthic species in temperate and sub-polar latitudes of both hemispheres; in all areas, changes that have taken place within the last 20 years. Range expansions have been described, for example, from around California (Barry *et al.*, 1995), the British Isles (Mieszkowska *et al.*, 2007), in the Bering Sea (Grebmeier *et al.*, 2006) and off the Antarctic Peninsula (Clarke *et al.*, 2005; Thatje, 2005). There is minimal information to indicate if tropical and subtropical species, other than possibly corals on the eastern margin of Florida, are expanding poleward.

3.8. Oxygen

One of the most critical variables in the world's ocean is the distribution of dissolved O_2. Oxygen plays a direct role in the biogeochemical cycling of carbon and nitrogen as well as being fundamental for all aerobic life, including organisms living in the dark ocean interior. If the oceans were to stagnate, many regions of its interior would be devoid of O_2 within a few decades as oxygen is continually being consumed by deep-dwelling organisms (Feely *et al.*, 2004; Whitney *et al.*, 2007).

A critical threshold is reached when O_2 levels reach $\sim$60 μmol kg^{-1}, below which most macro-organisms become hypoxic, that is, severely O_2 stressed (Gray *et al.*, 2002). A second threshold is crossed when O_2 drops below $\sim$5 μmol and nitrate becomes important in respiration, a condition termed 'suboxic'. When O_2 levels drop to zero, the water is termed 'anoxic', and biogeochemical processes are then dominated by sulphate-reducing microbes.

Ocean anoxic events (OAEs) have occurred episodically throughout the geologic record (Cohen *et al.*, 2007; Jones and Jenkyns, 2001; Wignall and Twitchett, 1996). These episodes are defined by sedimentary evidence of widespread anoxia and are often associated with evidence of warmer climate conditions, rises in sea-level and occasionally with mass extinctions. Although the cause of the events remains a matter of speculation, their existence underscores the potential vulnerability of oceanic O_2 supply in warmer climates.

Anoxia is rare in the modern open ocean, but is important in enclosed basins such as the Black and Baltic Seas. Hypoxic conditions occur, however, at mid-depths over wide expanses of the North Pacific, in smaller regions of the north Indian Ocean, and in the eastern tropical Atlantic and Pacific Oceans. These regions are known as oxygen minimum zones (OMZs), with the low levels of O_2 mostly attributable to the sluggish rate at which the subsurface is renewed by mixing with well-aerated surface waters (Karstensen *et al.*, 2008). Suboxic conditions are restricted to more limited regions of the north Indian and eastern Pacific OMZs. Hypoxic and suboxic conditions frequently occur in coastal waters, where low subsurface O_2 levels can be generated by natural high biological productivity in the overlying waters or by eutrophication from agricultural runoff or sewage inputs (Diaz and Rosenberg, 1995).

Significant reductions in the O_2 supply to the ocean interior and expansion of OMZs may result from continued anthropogenic global warming. Under business-as-usual type emission scenarios, climate models suggest that the global ocean O_2 inventory will decrease by 4–7% over the next century with continued reductions after that (Matear and Hirst, 2003; Schmittner *et al.*, 2008). The main mechanism is increased stratification of the surface ocean due to warming and freshening of high-latitude surface waters which reduces renewal rates. The details differ between models, with several other processes also being relevant including the direct reduction of O_2 solubility in warmer water and changes in rates of photosynthesis influencing the sinking flux of organic detritus into the ocean interior. The models suggest that detectible changes in O_2 content due to global warming may already have occurred (Matear *et al.*, 2000; Sarmiento *et al.*, 1998).

Marked declines in subsurface O_2 concentrations have been noted in 30-year records from the western North Pacific (Ono *et al.*, 2001) and 50-year records from the eastern North Pacific (Whitney *et al.*, 2007). The largest long-term declines (of the order 10 μmol kg^{-1} per decade) have been found in layers occupied by North Pacific intermediate water (150–600 m depth), which is renewed by contact with the surface in the Sea of Okhotsk and neighboring regions. The declines have been tied to freshening surface waters in the renewal regions associated with a reduction in renewal rates (Mecking *et al.*, 2006; Nakanowatari *et al.*, 2007) and appear superimposed on decadal variability of natural origin (Andreev and Baturina, 2006; Mecking *et al.*, 2008). As a consequence the oxic/hypoxic boundary has shoaled from 400 to 300 m over the past 50 years (Whitney *et al.*, 2007).

Significant O_2 declines have been found in 50-year records from the OMZs in the eastern tropical Pacific and Atlantic (Stramma *et al.*, 2008). The declines are of the order 1–3 μmol kg^{-1} per decade and are associated with a vertical expansion of the hypoxic layers. Clear evidence of long-term trends is lacking in other regions, however, and the ability to resolve long-term trends is impaired by sparse coverage of reliable historical O_2 measurements and by short-term or decadal variability (Johnson and Gruber, 2007; Min and Keller, 2005). Climate models suggest that the O_2 levels may actually increase in some regions despite the decline in the global average due to the complex response of dissolved O_2 to circulation changes (Matear and Hirst, 2003; Schmittner *et al.*, 2008). Changes in climate may also contribute to increasing coastal hypoxia (Grantham *et al.*, 2004).

The implications of a continuing global decline in global oceanic O_2 to climate change are unclear as the science is still at an early stage. Potential impacts of O_2 on organisms and ecosystems must be considered in concert with changes in acidity and temperature (Portner and Farrell, 2008). Further expansion of the O_2 minimum zones will likely adversely impact the distribution of fish and other commercially valuable species (Gray *et al.*, 2002).

3.9. Nutrients in general

In addition to CO_2 and light, phytoplankton growth and productivity requires the availability of a range of nutrients. In the 1930s, Redfield found that the bulk elemental composition of particulate organic matter in seawater is constrained and reflects the concentration of the major elements in seawater. This led to the adoption of the Redfield ratio 106C (carbon):16N (nitrogen):1P (phosphorus) (and 16Si for silicic acid that is essential for diatoms) as the average elemental composition. It should be stressed that there is considerable variability around these average ratios in time, space and by species/taxa (Arrigo, 2005; Klausmeier *et al.*, 2008). Spatial and temporal variability in nutrient availability has a profound influence on the composition, biomass, seasonal cycle and spatial variability of phytoplankton.

The ocean basins of the world show very variable spatial concentration patterns for different nutrients. There have been observed changes through time, but these show no consistent basin-scale patterns (Bindoff *et al.*, 2007). This lack of coherence in the data has been attributed to poor sampling coverage and limited compatibility between the methods used through time and by different laboratories (Bindoff *et al.*, 2007). Kamykowski and Zentara (2005a,b), in contrast, using a calibrated temperature/nitrate relationship from a range of locations, have produced anomaly charts of the difference between nitrate availability between 1909 and 2002 (Fig. 1.10). The figure shows the clear expansion of a shallower thermocline over much of the ocean (blue) with reduced availability of nitrate. Equatorial nitrate availability linked to El Niño is evident in the Pacific and a pronounced contrast between the two sides of Canada is seen, in the east reflecting increased haline stratification.

Three large regions that together cover ~20% of the ocean's surface, the Southern Ocean, eastern equatorial Pacific and subarctic Pacific, are characterised by high levels of nutrients and low chlorophyll (HNLC regions) (Aumont and Bopp, 2006). The stable chlorophyll levels in these regions have been attributed to iron limitation or grazing control by

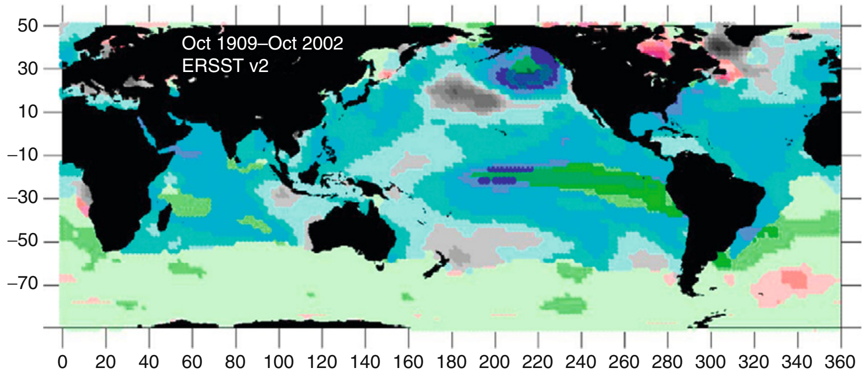

Figure 1.10 Map showing modelled difference in nitrate availability based on a temperature nitrate relationship, between October 1909 and 2002. Darker colours represent greater contrasts between the years. From Kamykowski and Zentara (2005a,b). Green, nitrate in 1909 present at the surface >2002; red, nitrate in 2002 present at the surface >1909; blue, stratification in 1909 between nitracline and surface <2002; grey, stratification in 1909 between nitracline and surface >2002.

microzooplankton (see Section 4). It is possible that increased desertification and land use changes in the future may lead to more aerial input of dust, including micro-nutrients such as iron, into the oceans (Mahowald *et al.*, 1999). Other factors than iron limitation may be behind the low chlorophyll levels, however, and so the addition of iron alone may not lead to an increase in production; it is the cumulative effect of the many changes that is important.

There has been an enhanced riverine input of N and P to near shore regions over the last century and especially since ~1950 that in some cases have caused eutrophication and elsewhere has been buried in organic carbon in sediments (Smith *et al.*, 2003). The latter study derived estimates based on population relationships that are three times higher than those derived in the 1970s. As population levels rise in the future, this pattern is likely to be reinforced as higher global temperatures are expected to lead to an increased mobilisation of N and P from sediment (Mackenzie *et al.*, 2002). Sequestration of carbon in sediments is likely to especially apply in regions subject to anoxia or hypoxia that appear to be increasing in extent (Diaz and Rosenberg, 2008). Atmospheric inputs of fixed nitrogen to the ocean have also increased and have contributed to higher algal production and nitrogen oxide (N_2O) emissions from the ocean (Duce *et al.*, 2008). Langlois *et al.* (2008) have shown that nitrogen fixing cyanobacteria in the North Atlantic were most abundant at higher temperatures (see also Stal, 2009) and with enhanced inputs of atmospheric dust.

Strong regional changes in nutrients are expected in the future dependent on variability in wet precipitation, wave storminess, expanding OMZs, mixing and the depth of stratification. Precipitation is expected to increase especially in tropical regions. At present, it is not possible to predict future trends in nutrients because of the localisation of the changes. It is also not clear how all the regional responses will add up to a global mean and influence climate change.

3.9.1. The oceanic nitrogen cycle

Nitrogen is a fundamental component of all organisms and essential in the chemical forms that are needed for assimilation in primary production. Understanding of the nitrogen cycle (Fig. 1.11), and especially of the role played by microbes in the cycle, has increased substantially in the last decade. The predominant form in the ocean is nitrogen gas (N_2); this gas can only be utilised by a few specialist nitrogen fixers that include the

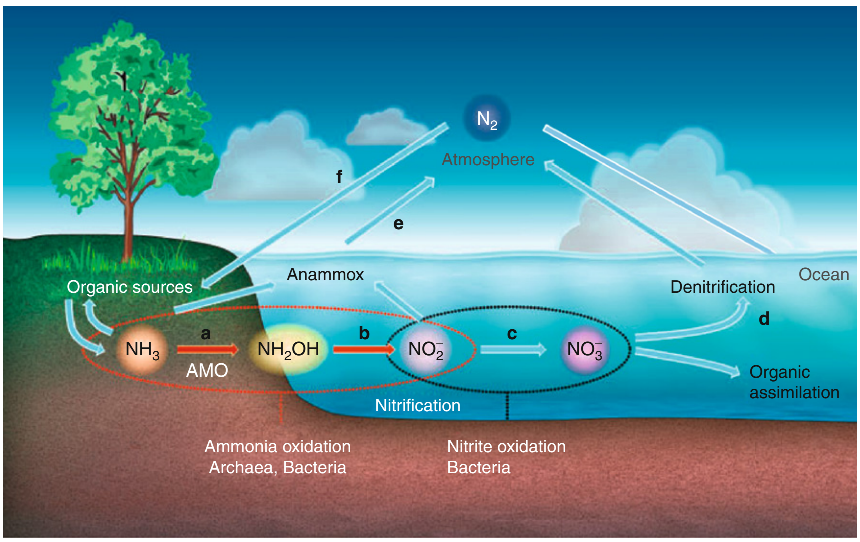

Figure 1.11 The global nitrogen cycle. Nitrification is a step-wise process by which ammonia (NH_3) of organic origin is oxidised to nitrate ions (NO_3^-). The first step (a) involves the ammonia monooxygenase (AMO) enzyme in Crenarchaeota and bacteria that convert ammonia into hydroxylamine (NH_2OH). (b) This is then processed by bacteria and possibly archaea into nitrite ions (NO_2^-). (c) Other specialised bacteria complete nitrification by converting nitrite to nitrate. (d) Nitrate is then assimilated into organic matter during primary production or denitrified by other organisms to form nitrogen some of which escapes to the atmosphere. (e) Anammox bacteria can also convert ammonia and nitrite into nitrogen for release to the atmosphere. (f) Nitrogen is fixed by specialised organisms and converted to ammonia that is converted into organic matter or oxidised as the cycle continues. The organic sources on land also apply to the ocean and additional sources to the ocean are contributed from rivers and the atmosphere. Figure modified from Schleper (2008).

important tropical/subtropical cyanobacteria, *Trichodesmium* and a number of groups of unicellular cyanobacteria (Capone *et al*., 2005; Stal, 2009). In addition, endosymbiotic associations between some diatoms and cyanobacteria (Gomez *et al*., 2005) and some bacteria (Johnston *et al*., 2005) may also fix N_2. 'A new paradigm for nitrogen fixation' (Arrigo, 2005; Karl, 2007) means that it is now recognised that the oceans contribute at least 50%, and likely more, to global fixation than the minor contribution previously believed (Stal, 2009). This percentage is likely to increase in the future as tropical and subtropical waters expand, with considerable consequences for climate through changes in nutrient ratios (stochiometry), plankton communities and the biological pump. Other phytoplankton groups utilise a range of reduced and oxidised forms of nitrogen (NH_4^+, NO_2^-, NO_3^- plus organic nitrogen). The concentration and spatial and temporal variability of these N forms is a key global determinant of phytoplankton biomass and rates of primary production. In energetic terms, when available, ammonium (NH_4^+) is the preferred nitrogen source. Ammonium is also the primary form of N used by phytoplankton in the large regions of the ocean (subject to permanent stratification) where recycling-based communities prevail and the phytoplankton often have higher surface area-to-volume ratios. Other important breakthroughs in understanding the nitrogen cycle (Fig. 1.11) include the recognition that Archaea (Crenarchaeota) as well as bacteria are capable of oxidising ammonia and the role of the Anammox process in delivering nitrogen gas to the atmosphere (see Section 3.2).

Conversion of dissolved ammonia, nitrite and nitrate to particulate forms against the reverse process of denitrification appear to be generally in balance within a 3000-year time period. There are, however, opposing views at present on the size and relative balance between biological nitrogen fixation and denitrification in the ocean. Yool *et al.* (2007), for example, have shown that about half of the global uptake of nitrate by marine phytoplankton is produced by denitrification. On the basis of their results they suggest that the biological pump may be less efficient than previously estimated. Note the additional complication, by reference to Section 3.2 earlier, of the important discovery of the new Anammox denitrification process and the use of NH_4 as an energy source by Archaea, especially in the surface ocean. An increased contribution to the nitrogen pool due to utilisation of N by cyanobacteria and bacteria is thought unlikely as iron is also needed in this process (Falkowski *et al*., 1998).

The extent to which the availability of nitrogen will affect the ability of the biosphere to absorb increasing levels of atmospheric CO_2 in the future is not clear (Gruber and Galloway, 2008). There has been a large increase in the anthropogenic input of nitrogen to the environment deriving from industrial processes fertilisers, animal husbandry, sewage and fossil fuel emissions. In the marine environment, many of these inputs have caused problems in coastal waters in the form of eutrophication. Anthropogenic emissions contribute to the build-up of greenhouse gases such as nitrous oxide, (N_2O) and nitrogen trifluoride (NF_3) (Forster *et al.*, 2007; Prather and Hsu, 2008) as well as contributing to a loss of ozone in the atmosphere. On a global scale the deposition in the ocean of anthropogenic biologically available nitrogen from atmospheric sources is likely to have stimulated phytoplankton growth (Duce *et al.*, 2008). The total input of these anthropogenic sources at 160 Tg N $year^{-1}$ accounts for more than the natural fixation of nitrogen on land (110 Tg N $year^{-1}$) or in the ocean (Gruber and Galloway, 2008), but not all may be available for marine photosynthesis. An observed parallel development of trends in atmospheric CO_2, N_2O and temperature over the last 250 years (Fig. 1.12) with good evidence for parallel changes in the Pleistocene (Flückiger *et al.*, 2004) emphasises the close relationship between these two gases and their links to climate change.

A recent study has shown that cell division rate doubled and the Redfield ratios (106C/16N/1P) C/P and N/P (not C/N) of the planktonic cyanobacteria *Trichodesmium* changed markedly, as a response to increasing levels of CO_2 leading to an enhancement of nitrification and potential enhanced CO_2 drawdown (Barcelos e Ramos *et al.*, 2007). It is believed unlikely, but if this proved to be a selective response in phytoplankton, in general it would establish a strong negative feedback to climate, provided the produced POC sank below the mixed layer.

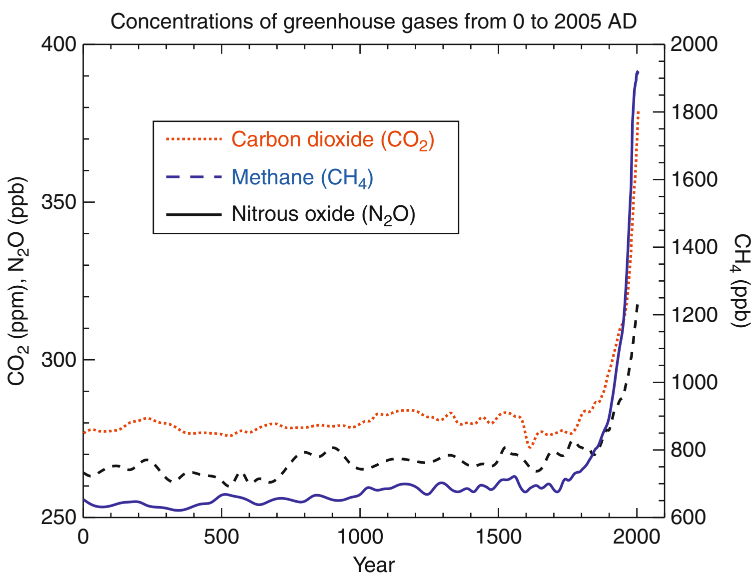

Figure 1.12 Atmospheric concentrations of N_2O, CO_2 and CH_4 over the last 2000 years showing the close parallel nature of their trends. From Forster *et al.* (2007) IPCC AR4 WG 1.

3.9.2. The oceanic phosphorus cycle

Phosphorus is an essential nutrient in primary production (Froelich *et al.*, 1982) and is used as an energy carrier (in ATP/ADP molecules) in all organisms (Föllmi, 1996). The marine phosphorus cycle involves uptake and assimilation of phosphate by plankton in surface waters, its release back into the water by processes such as cell lysis and bacterial degradation and its subsequent transport to the deep ocean via sinking of released organic material, and return to the upper ocean by slow mixing (diffusion, up- and downwelling) and circulation (Tyrrell, 1999; Williams and Follows, 2003). There has been a considerable debate on the influence of P and N on ocean productivity during glacial cycles and consequently on atmospheric concentrations of CO_2. Recent results from deep sea cores (Tamburini and Föllmi, 2009) suggest that burial of reactive P in glacial conditions is reduced and that richer P conditions characterise glacial terminations, both with a possible feedback to the carbon cycle. Models of phosphate cycling by Tyrrell (1999) indicate that while in the steady-state nitrate is more deficient than phosphate, external inputs of P control the longer term primary production of the global ocean. Sources of P in the oceans are dominated by river input with *circa* 90% of all inputs consisting of organic debris. Outside coastal waters, P concentrations in the euphotic zone are dependent on rates of upwelling and diffusion from deep water in the ocean's interior as well as the concentration of P in the source water (Froelich *et al.*, 1982). In HNLC (high nutrient low chlorophyll) and upwelling regions, PO_4 is in plentiful supply, but generally not elsewhere, except during the winter in temperate and sub-polar latitudes. An exception is the eastern Mediterranean where a high nitrate-to-phosphate ratio has been observed resulting in phosphate limitation of the primary production (Krom *et al.*, 2004; Rees *et al.*, 2006). A similar situation has been described from Station ALOHA in the northwest Pacific sub-polar gyre where N_2 fixation is possibly increasing over time due to climate-coupled changes, leading to an intensification of P stress in this P-limited ecosystem (Karl, 2007).

3.9.3. The oceanic silicon cycle

In the oceans silicon (Si) is primarily in the dissolved inorganic oxidised form silicic acid. When silicic acid is available, diatoms dominate phyto-plankton communities and are important because of their high sinking rates, in the export of carbon via the Biological pump. The proportions of silicate versus carbonate sedimentation have been implicated as a factor in the reduction of atmospheric CO_2 concentrations by ~100 ppm in glacial periods. Work by Kohfeld *et al.* (2005), however, suggests that increased growth of diatoms and other biological processes could account for no more than 50% of the drawdown. Over the last 40 million years since the Late Oligocene silica-rich upwelling regions have been increasing as the planet cooled, a trend that is closely associated with the evolution of whales (Broecker and Kunzig, 2008); implications for whales and climate in a warming planet are unclear.

The supply of silicic acid to the upper illuminated layer derives from weathering, riverine fluxes and upwelling from the ocean interior (Falkowski *et al.*, 1998). Concentrations of silicic acid in water are highest in the Northern Hemisphere, off major river basins, in Subarctic Seas and the Arctic Ocean, but also in the Southern Ocean where silica wells up from deeper water and is carried northwards by the ACC. In contrast, in the central ocean gyres levels of silicic acid are very low. This distribution is reflected in the distribution of the production of biogenic silicon (opal) as diatoms and radiolaria (Fig. 1.13). As a contrast to the Northern Hemisphere, more than two-thirds of the global sedimented silica is deposited south of the Polar Front under the ACC (Smith *et al.*, 2003; Wischmeyer *et al.*, 2003). In a modelling study of the silicon cycle, Yool and Tyrrell (2003) show that the ecological success of diatoms varies inversely with the concentration of silicic acid and thus through a negative feedback controls the cycle. However, total primary production is shown to be controlled by phosphate and not silicic acid availability.

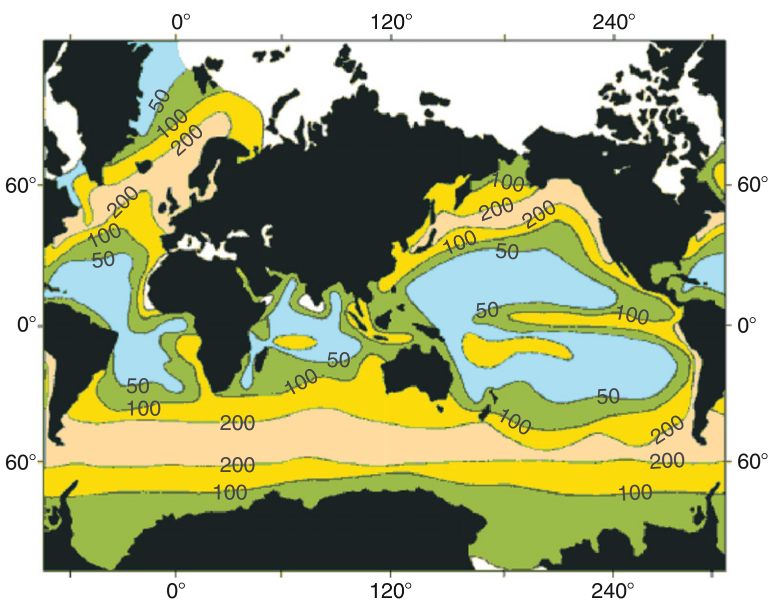

Figure 1.13 Annual production of biogenic silicon in the oceans (g m^{-2} year^{-1}). Source www.radiolaria.org.

3.9.4. Iron and dust

Iron is an essential element for all organisms because it is needed in a wide range of enzyme systems for processes including photosynthesis, respiration and nitrogen fixation. In HNLC regions, iron deficiency reduces growth rates, cellular chlorophyll levels, net CO_2 fixation and active uptake of CO_2 and HCO_3^- (Schulz *et al*., 2007). The late John Martin helped to focus attention on the importance of iron supply to the Southern Ocean by highlighting that this was higher in the past, which could account for a substantial proportion of the 80-ppm drawdown in atmospheric CO_2 observed between glacial and interglacial periods (Martin, 1990). In the contemporary ocean, iron limits the growth of phytoplankton over broad areas where macronutrient concentrations remain high (Boyd *et al*., 2000). These include the so-called HNLC regions of the Southern Ocean, as well as parts of the North and the Equatorial Pacific.

Iron is supplied to the oceans with soil alumino-silicates in riverine inputs, upwelling and to the open ocean via aeolian inputs of dust that originate from various deserts, with lesser inputs from volcanic, anthropogenic and meteoric sources. Drought conditions as well as changes in land use and agricultural practice can lead to increased dust emissions. Dust inputs are particularly important for the vast open ocean regions and a key concern is that arid regions are very sensitive to climate change and this has the potential to change ocean productivity and global climate in turn (Jickells *et al*., 2005). However, aside from the issue of climate change, a recent study by Wagener *et al*. (2008) underlines the need for in-depth comparisons of model and *in situ data* and a re-evaluation of predictions of present and past dust inputs. They examined aerosol deposition of iron to two remote oceanic areas in the Southern Hemisphere and concluded that current dust deposition models overestimated iron inputs and that dust deposition is not the dominant source of iron for this large and important HNLC region.

3.10. Other gases and aerosols

3.10.1. Methane, nitrous oxide and halocarbons

Man-made (livestock, arable farming, landfill, industry) and natural emissions from terrestrial environments and the sea contribute to the atmospheric burden of methane (CH_4), nitrous oxide (N_2O) and a suite of volatile halocarbons (compounds containing chlorine, iodine and/or bromine). Methane and N_2O are potent greenhouse gases with global warming potentials about 21 and 310 times that of CO_2. Methane oxidises into CO_2, and continues to have a long-term global warming effect. The halocarbon gases are key sea-to-air transfer compounds for the global biogeochemical cycles of bromine, chlorine and iodine.

Global net methane emissions from oceans and freshwaters are estimated at 10 Tg per annum. While this is only about 2% of the total global source, the marine methane cycle is nevertheless considered highly significant because the concomitant anaerobic methane oxidation sink ensures that the net flux of methane to the air is a minor fraction of the total methane produced (Reeburgh, 2007). Methane is produced and consumed in seawater via microbial reactions, and also arises from geological sources including methane clathrates, hydrothermal vents, cold seeps, mud volcanoes and anaerobic methanogenesis. Considerable regional variability is evident over the ocean that has much lower levels than over the land (Fig. 1.14). The Black Sea stands out for having high surface methane concentrations and fluxes. Until recently, methane production was thought to require strictly anaerobic conditions. The origin of the methane distributed in oxygenated surface ocean waters was considered a paradox, and it was assumed that methane production was limited to anoxic environments in digestive tracts and faecal pellets. However, recently Karl *et al.* (2008) have shown that methane is a by-product of the aerobic microbial breakdown of methylphosphonate (CH_5O_3P), and they suggest that marine methane production could increase with global warming-induced increases in stratification and nutrient limitation.

The oceans, particularly the sediments of continental slope regions, are estimated to harbour 2000 Gt of carbon as methane gas and icy solids known as methane clathrates or hydrates (Buffett and Archer, 2004). It has been suggested that catastrophic release of methane from this store caused abrupt

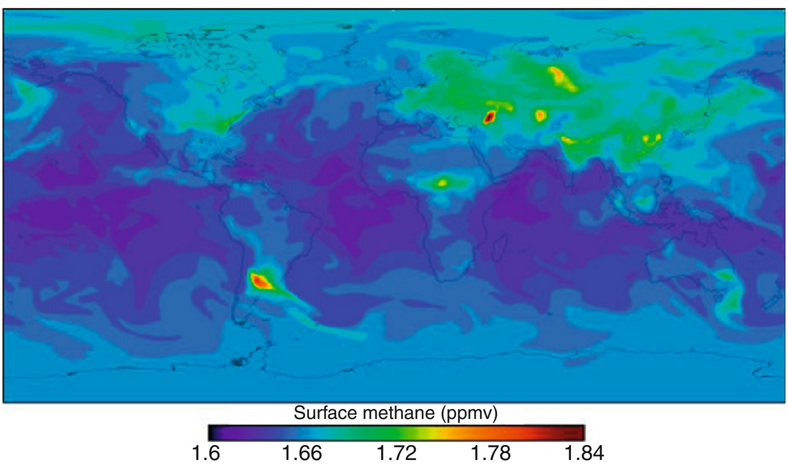

Figure 1.14 Global map of surface methane concentrations. From NASA: Credit: GMAO Chemical Forecasts and GEOS GHEM NRT Simulations for ICARTT.

climate warming in the past and there are concerns that a warmer future climate could destabilise this reservoir and trigger further warming (see Section 6).

N_2O production is enhanced in areas where oxygen levels are depleted and nitrate fuels denitrification. These conditions are combined in upwelling areas and it has been suggested that the global expansion of hypoxic/anoxic zones (Chan *et al.*, 2008) is leading to increased production of N_2O and the accumulation of this radiatively active trace gas in the atmosphere (Naqvi *et al.*, 2000).

If the mid-water depths of the ocean are shut off from ocean ventilation, CH_4 and N_2O will be increasingly produced and as a gas will bubble up towards the surface. An increase of even a small amount of N_2O entering the atmosphere could have large implications. The possible role that an increase in sea areas subject to hypoxia/anoxia (as in the Black and Baltic seas or in upwelling regions such as the Arabian Sea) might play as a contribution to climate change is not clear.

Some halocarbons are man-made, but many are derived from seaweeds, microalgae and/or extracellular photochemical reactions. They break down photochemically in the troposphere to form halogen radicals which destroy ozone and produce halogen oxides. Ozone is a major precursor for hydroxyl (OH) radicals and its removal from the system impairs the atmospheric cleansing of pollutants including methane. It has been suggested that BrO interacts with DMS (see below) reducing its cooling effect on climate via aerosol production (von Glasow *et al.*, 2004) whereas iodine oxides contribute to the formation and growth of marine aerosol (O'Dowd and Leeuw, 2007). The deep ocean may be an important source region for halocarbons and other gases, but little work has been done in this area.

3.10.2. Aerosols

The burning of fossil fuels, including in shipping, releases SO_2 to the air where it is oxidised to form sulphate aerosol. This acts to cool the climate directly, because the aerosol particles reflect solar radiation back into space, and indirectly as sulphate particles can also act as condensation nuclei influencing cloud formation and the radiative properties of clouds. Natural emissions of sulphur are dominated by marine biological production of dimethylsulphide [DMS; $(CH_3)_2S$] with sporadic minor emissions from volcanoes. Other sources of marine aerosol such as sea salt, other biogenic gases (e.g. iodinated gases and isoprene) or organic matter produced in phytoplankton blooms may also serve as cloud concentration nuclei (CCN). The combined response of all marine aerosol sources as a potential feedback to climate change is still unclear.

Between 15 and 33 $\times$ 10^{12} g of the volatile sulphur trace gas DMS are emitted annually from the ocean to the atmosphere. This is equivalent to ~27–60% of the estimated flux of sulphur from anthropogenic sources and makes DMS a significant compound in the global sulphur cycle (Kettle and Andreae, 2000), especially when set against the evidence that man-made emissions in many areas have declined since 1989 (Andreae *et al.*, 2005). Using a modelling approach, Vallina *et al.* (2007) estimated that globally the DMS contribution to CCN was ~30% of the total CCN numbers determined from satellite data.

In the oceans, DMS is derived from dimethylsulphoniopropionate [$(CH_3)_2S^+H_2CH_2COO^-$; DMSP] that is produced by some marine phytoplankton, especially the Prymnesiophyceae and Dinophyceae (Stefels *et al.*, 2007). Release of DMSP and DMS to seawater occurs during phytoplankton cell death, grazing and viral mortality and may be passively and/or actively released. Most is utilised by bacteria or oxidised via photochemical processes and the amount of DMS emitted to the air is a few percent of the total DMSP and DMS pool. Recent work suggests that UV light enhances DMS production by phytoplankton and decreases bacterial turnover of DMSP. Climate change could impact on DMS production and emissions via changes in wind intensity, ocean circulation, light field, the relative abundance of DMSP-rich and DMSP-poor phytoplankton types, levels of marine productivity and food web functioning.

It is well accepted that sulphate aerosols arising from volcanic and fossil fuel-derived sulphur emissions influence the climate by reducing radiative forcing by direct reflection of radiation back into space and through CCN (Andreae *et al.*, 2005), but as yet modelling studies have not reached true consensus on whether DMS has a significant influence on the Earth's climate. Gunson *et al.* (2006) used a coupled ocean–atmosphere general circulation model with an atmospheric sulphur cycle and tested how climate might respond to altered DMS emissions. They altered DMS emissions to half the control simulation value and this increased radiative forcing by 3 W m^{-2} and surface air temperature by 1.6 °C. Using a 2× CO_2 scenario, Bopp *et al.* (2004) estimated a 3% increase in global DMS flux that would give a minor negative feedback of about −0.05 W m^{-2}, but the regional variation in the model output was large (−15% to 30%) and a substantial radiative forcing of −1.5 W m^{-2} was suggested for 40–50°S in summer. Kloster *et al.* (2007) also found large regional-scale variability and their models predicted a 10% reduction in the global annual mean DMS sea surface concentration and the DMS flux for 2061–2090 compared to 1861–1890, but the atmospheric DMS burden was reduced by only 3% because DMS would have a longer lifetime in air with a warmer climate. Again the Southern Ocean was identified as a vulnerable region, with DMS levels reduced by 40% and DMS concentrations were also reduced in the mid- and low-latitude regions because of nutrient limitation associated with increased stratification.

3.11. Concluding comments

- Marine methane production by plankton could increase with global warming-induced increases in stratification and nutrient limitation.
- There has been a pronounced expansion of the large low productivity regions of the world (subtropical gyres), which already far exceeds the predictions of models.
- Extrapolation on the basis of projected changes in sea surface temperature and nutrients could lead to a substantial reduction in the productivity of the oceans and the efficiency of the Biological pump over the next 100 years.
- The role of microbes in climate and climate change is crucially important, but little understood and poorly quantified. Developing an understanding of their contribution to biogeochemical and nutrient cycling and microbial diversity is a major challenge for the future.
- Shifts from a diatom to a flagellate dominated system in temperate latitudes and increased microbial remineralisation in a warming ocean are expected to lead to a less efficient Biological pump.
- If these changes caused a reduction in the net input of atmospheric CO_2 to the oceans, there would be a strong positive feedback to climate change.
- Large changes have been observed in marine ecosystems in many different parts of the oceans; when seen together they indicate that they are reacting beyond what might be expected from interannual variability.
- Modelling has shown that changes in ecosystem structure (e.g. types of plankton, physiology, light absorption, food web structure) and export efficiency may be as or more important to understanding interactions with climate than changes in bulk-integrated satellite measures of chlorophyll. The global observing systems needed to measure such variability are rudimentary and concentrated in the Northern Hemisphere at present.
- The oceans are a major producer of sulphur particulates which seed cloud formation. Changes in the production of all aerosols as seas warm has implications for global warming, but the net effect is unclear.
- Mismatch between trophic levels and functional groups has implications for ocean–climate interactions, including CO_2 drawdown.

4. The Solubility, Biological and Continental Shelf Carbon Pumps

4.1. The ocean carbon cycle

The carbon cycle is crucial to climate because it governs the amount of the important greenhouse gases such as CO_2 and CH_4 in the atmosphere. Methane provides a continuous, transitory supplement as it is slowly converted to CO_2 in the atmosphere over approximately a 10-year period. The oceans play a crucial role in this cycle as the main reservoir for carbon (32,000 Pg estimated as stored in the deep ocean), other than the long-term storage of carbon in the Earth's crust. Feedbacks from the ocean carbon cycle and relevant processes are discussed Denman *et al*. (2007). To quote from the IPCC report, ''small changes in the large ocean carbon reservoir can induce significant changes in atmospheric CO_2 concentration'' and the oceans can also buffer ''perturbations in atmospheric pCO_2''.

In the pre-industrial Holocene there was an approximate time- and space-averaged equilibrium between CO_2 in the atmosphere and dissolved in the surface ocean. The regional differences in partial pressure in seawater CO_2 are due to interactions between biological, chemical and physical processes. Anthropogenic CO_2 release to the atmosphere has resulted in a net flux of CO_2 from the atmosphere to the ocean that occurred on top of an already active oceanic carbon cycle (Fig. 1.15). Anthropogenic CO_2 is absorbed into the water by direct solubilisation, with the dissolved carbon subsequently distributed to depth by mixing and ocean currents.

The contribution that biology makes is still far from understood. For example, it is not known if CO_2 drawdown increases if plankton are more productive and/or if functional groups such as diatoms are more dominant. Introduced CO_2 reacts with water to produce carbonic acid. Subsequent re-equilibration of the dissolved inorganic carbon (DIC) system results in an increase in the concentration of CO_2 and carbonic acid, a smaller proportionate (but greater in absolute terms) increase in bicarbonate ions, and a decrease in carbonate ions and pH. There is a marked difference in the concentration of DIC between the deep ocean and the mixed layer at ~500 m (Raven and Falkowski, 1999; Fig. 1.16) reflecting net autotrophy of surface waters and net heterotrophy in deep waters that results in a huge reservoir of DIC in the deep ocean. The DIC is transported, directly or as dissolved organic (DOC), particulate organic (POC) or inorganic (PIC) carbon, to the deep ocean by four processes collectively known as 'carbon pumps'. In the upwelling regions of the world, cold DIC-rich waters from the deep ocean recirculate to the surface where CO_2 outgases to the atmosphere to complete the ocean carbon cycle.

The four 'carbon pumps' (Solubility, Biological, Continental Shelf and Carbonate Counter) sequester CO_2, largely as DIC at the surface of the ocean, with additional transfer through the intermediarie POC, DOC and $CaCO_3$ PIC to the deep ocean reservoir that is mostly comprised of DIC. To some extent the Continental Shelf and Carbonate Counter pumps can be considered as subsidiary versions of the Biological pump. The Carbonate Counter pump will be covered more fully in Section 5.

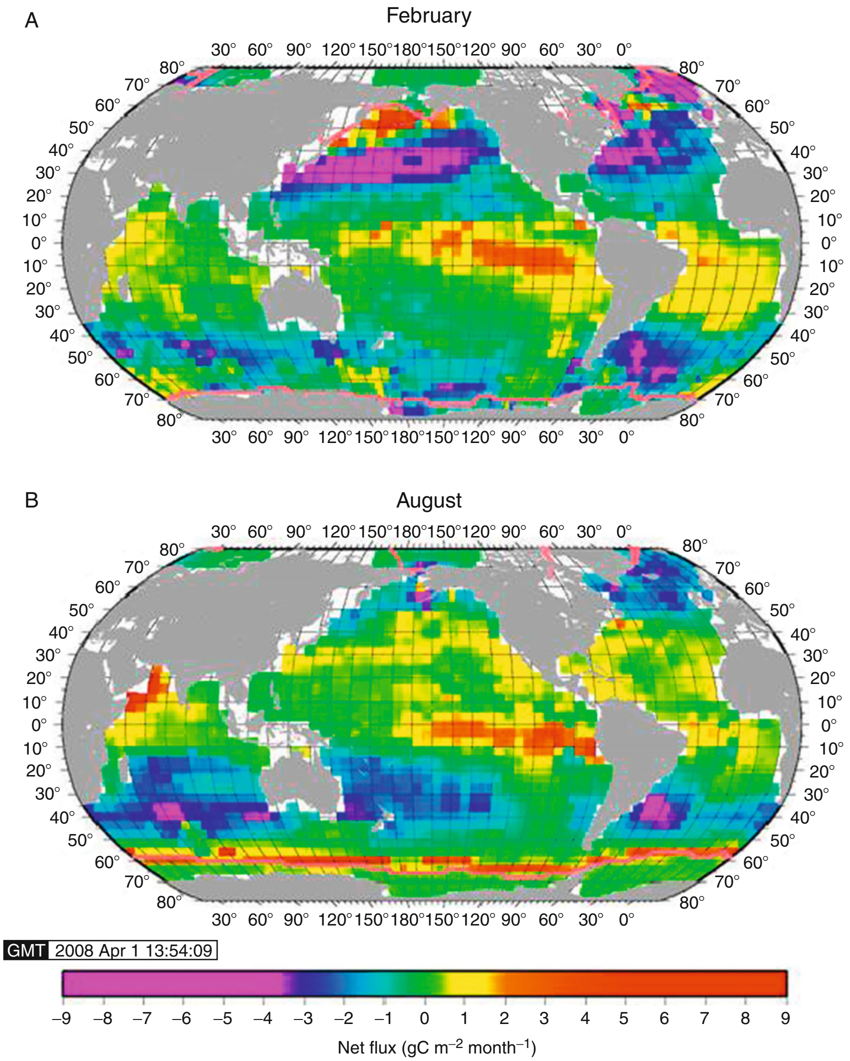

Figure 1.15 Climatological mean distribution of CO_2 flux (g C m^{-2} $month^{-1}$) between the air and sea or *vice versa* for February (A) and August (B) in the reference year 2000. The wind speed data are from the 1979–2005 NCEP/DOE AMIP-II Reanalysis, and the gas transfer coefficient is computed using a (wind speed) squared dependence. Positive values (yellow–orange–red) indicate sea-to-air fluxes, and negative values (blue–magenta) indicate air-to-sea fluxes. Ice field data are from NCEP/DOE-2 Reanalysis Data (2005). An annual flux of 1.4 ± 0.7 Pg C $year^{-1}$ is obtained for the global ocean by a summation of 12 monthly maps that were produced from approximately 12 million measurements. Figure from Takahashi *et al.* (2009).

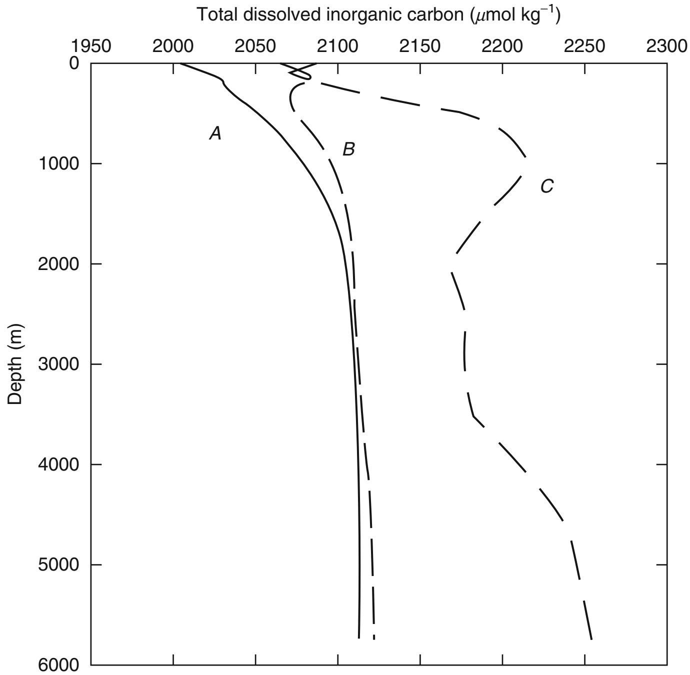

Figure 1.16 Vertical profiles of dissolved inorganic carbon (DIC) in the ocean. Curve *A* is a theoretical profile from prior to the industrial revolution with an atmospheric CO_2 concentration of 280 ppm. Curve *B* is a similar calculated profile for the year 1995, with atmospheric CO_2 at 360 ppm. The difference between these two curves is the integrated oceanic uptake of CO_2 from anthropogenic emissions since the beginning of the industrial revolution, assuming that biological processes have been in steady state (i.e. not materially affected by the net influx of CO_2). Curve *C* is a representative profile of measured DIC from the central Pacific Ocean. The difference between curves *C* and *B* is the contribution of biological processes to the uptake of CO_2 in the steady state (i.e. the contribution of the 'Biological pump' to the DIC pool). Figure redrawn from Raven and Falkowski (1999).

4.2. Ocean carbon pumps

4.2.1. Solubility pump

This pump operates most efficiently at low temperatures where the uptake of CO_2 as DIC is much higher due to increased solubility and at high latitudes where water downwells. This process only occurs in the sub-polar seas of the North Atlantic (not in the North Pacific) and in the Southern Ocean. When ice is formed in these polar regions, the released dense brines sink rapidly carrying with them DIC-rich water. Dense water may also be formed below pancake ice, for example in the Greenland Sea or in Arctic polynyas (a polynya is a large area of open water surrounded by sea-ice). A similar process takes place over the Arctic shelf as new ice is formed each year with the carbon-rich brines flowing along the bottom and over the shelf edge into the deep ocean. In such regions of deep water formation, carbon is delivered at high concentrations to the deep ocean where the deep circulation (MOC) carries it around the world and keeps it out of contact with the atmosphere for up to 1000 years. It has been estimated that about 25–50% of the steep vertical gradient in DIC (Fig. 1.16) is contributed by this pump. In regions where subtropical mode and intermediate waters are formed (see Section 2), usually by wintertime convective mixing, uptake of CO_2 by the Solubility pump provides an intermediate (up to decades) carbon sink (Bates *et al.*, 2002; Sabine *et al.*, 2004a). Sabine *et al.* (2004a) show, for example, that 40% of the global ocean inventory of anthropogenic CO_2 is found south of 30°S and most of that is stored in intermediate and mode water.

The fact that CO_2 solubility reduces with higher temperatures and salinity is of key relevance to climate change. It is estimated that the Solubility pump has become less efficient in the northern North Atlantic (Sabine *et al.*, 2004a) due to the warmer temperatures that have occurred over the last decade or more and supported by the observed reduction in the density of the deep water found in the Norwegian Sea. A similar reduction in uptake has recently been described for the shallower Japan Sea (Park *et al.*, 2008).

Changes over the last few decades in the large-scale atmospheric circulation of the Southern Hemisphere (Thompson and Solomon, 2002) are reflected in the leading mode of Southern Hemisphere climate variability, the Southern Annular Mode (SAM; Thompson and Wallace, 2000). Interannual variability and trends in the SAM also have been shown to drive substantial variability in ocean circulation with a poleward shift and intensification of westerly winds, in upper-ocean biology, and in the uptake and release of CO_2 to and from the Southern Ocean (Lovenduski and Gruber, 2005; Lovenduski *et al.*, 2007, 2008). Model simulations suggest that the trend towards more positive SAM conditions has led to a reduction in the strength of the Southern Ocean CO_2 sink (Lenton and Matear, 2007; Lovenduski *et al.*, 2007, 2008) by anomalous outgassing due to an increase in upwelling. This hypothesis has been supported by the inversion of atmospheric CO_2 data (Le Quéré *et al.*, 2007), but remains a subject of intense discussion. While doubts have been raised about the sensitivity of the inversion method to the choice of stations used (Law *et al.*, 2008), of the ocean model to the forcing used (Law *et al.*, 2008; Lovenduski *et al.*, 2008), and whether the 'saturation' of the Southern Ocean sink is likely to continue in the future (Zickfeld *et al.*, 2008), the results, for example, Le Quére *et al.* (2008) underscore the potential sensitivity of the global carbon cycle to changes in the circulation of the Southern Ocean. A number of climate change experiments reinforce this message by suggesting that increased greenhouse gases may, in turn, drive long-term changes towards a more positive SAM state (e.g. Kushner *et al.*, 2001; Miller *et al.*, 2006). Thus, the Southern Ocean carbon cycle, in connection with Southern Hemisphere atmosphere–ocean circulation, winds and stratification could give rise to a positive feedback that would enhance global warming (Friedlingstein, 2008; Lovenduski and Ito, 2009).

4.2.2. Biological pump

Through this pump CO_2 fixed by photosynthesis is transferred to the deep ocean primarily as dead organisms (including the organic skeletal), faecal material (POC), and carbonate skeletons (PIC; note that calcification produces CO_2). This results in sequestration (storage) of carbon for periods of decades to centuries (depending on the depth of remineralisation) or even more permanently in the sediments. Longer-term sequestration may be in the form of organic matter, such as the type of material that is ultimately the source of oil and natural gas. A small proportion of the total annual production of the plankton ends up in the deep ocean, but there is strong evidence to suggest that this pump contributes importantly to the different levels of atmospheric CO_2 found between glacial and interglacial periods (Raven and Falkowski, 1999). Plankton can act as ballast for the export of carbon to the deep ocean with the organisms that have mineralised skeletal parts playing an important role. Siliceous diatoms and calcareous foraminifera, coccolithophores and molluscan pteropods and cephalopods are important ballast organisms. Other forms of settling occur via faecal pellets or aggregates and gelatinous plankton. Exopolysaccharide aggregation can increase sinking POC and PIC at a given overall density by decreasing the surface area per unit volume (Engel *et al.*, 2004) and terrigenous materials such as clay may also contribute (Klaas and Archer, 2002). It has even been suggested that POC fluxes may drive mineral fluxes rather than *vice versa* (Passow, 2004). Although there is much data on the rate of organic carbon sinking in the Biological pump and its determinants, there is still uncertainty as to the nature of a predictive model (Boyd and Trull, 2007; De La Rocha and Passow, 2007; Passow, 2004).

To predict future CO_2 concentrations in the atmosphere there is a need for a much improved understanding of the way that the Biological pump varies both geographically and temporally and the effects on the pump of changes in temperature, ocean circulation and ocean chemistry (e.g. acidification due to increased CO_2). It is not known, for example, if earlier spring blooms or higher Fe input into HNLC areas (e.g. the Southern Ocean) will affect carbon drawdown, or if CO_2 drawdown will reduce during prolonged periods of recycled production due to longer summers, nutrient limitation and expansion of the subtropics (e.g. Bopp *et al.*, 2001). Recent studies, for example, van Hoof *et al.* (2008) indicate that natural decadal variability in atmospheric concentrations of CO_2 as measured from leaf stomata in the pre-industrial period from 1000 to 1500 AD were more pronounced and faster than proposed in IPCC AR4. They suggest that the variability is driven by oceanic perturbations in temperature and salinity. The extent to which the oceans may contribute to such 'short-term' variability is not known.

4.2.3. Continental Shelf pump

Continental shelf seas comprise ~7% of the surface ocean but provide a disproportionately large fraction (15–30%) of oceanic primary production (Bozec *et al.*, 2005). Thus these regions have a strong impact on the global carbon cycle and provide a net flux to the deep ocean reservoir calculated at ~1 Pg C year^{-1} by Tsunogai *et al.* (1999).

Cold, denser water with lower $p\mathrm{CO_2}$ is formed in many coastal shelf seawaters at temperate and sub-polar latitudes during the colder periods of the year. As a consequence these are regions of net uptake of atmospheric CO_2 by solubilisation that may be enhanced by higher levels of phytoplankton production. Shelf seas may be totally mixed throughout the year or have a pycnocline/thermocline that separates stratified waters from the mixed waters below that are isolated from the atmosphere. A range of complex processes transfer DIC through the pycnocline via the intermediaries POC, DOC and PIC. DIC is then transferred by isopycnal mixing (advection and diffusion) off the shelf to the deep ocean. The transfer to the deep ocean may continue even while the surface layer is isolated by stratification. Material may also be transferred to the deep ocean as POC, DOC and PIC via nepheloid layers and by transport of organic material as fluff along the bottom. In strongly mixed waters as in the southern North Sea (Bozec *et al.*, 2005), the whole water column is in regular contact with the atmosphere and bacterial regeneration ensures that these regions are generally net sources of CO_2, especially if they are enriched with nutrients.

During stratified summer conditions, carbon export to the mixed waters below the pycnocline is probably reduced and so higher temperatures and the resultant stronger stratification will likely feedback to a reduced export of CO_2. It is also estimated that higher nutrient input to these regions, especially in eutrophicated areas, will contribute to increased CO_2 drawdown if more nutrients are available. Major works to improve water and sewage treatment in Europe, for example, will thus reduce CO_2 drawdown by the Continental Shelf pump.

4.2.4. Carbonate Counter pump

This pump operates in parallel with the (organic carbon) Biological pump and covers the production and dissolution of marine organisms with body parts made up of inorganic $CaCO_3$. The phytoplankton (coccolithophores) and zooplankton (foraminifera, pteropods, planktonic larval stages of benthic organisms) and some benthic algae plus many benthic animals, including corals, produce body parts made of calcite, aragonite or Mg carbonates. Production of carbonates leads to CO_2 release (see Section 5).

4.3. Role of the four ocean carbon pumps

Some idea of the importance of these carbon pumps can be gauged from a comparison of the present estimated transfer of carbon by the Biological pump to the deep ocean (see IPCC AR4 WG 1, 2007, Fig. 7.3). A net reduction of only 10% (1.1 Pg year^{-1}) would virtually counterbalance the current estimated net input (1.4 Pg year^{-1}) (Takahashi *et al.*, 2009) of atmospheric CO_2 to the ocean. The relative contributions and importance of the Solubility, Biological, Continental Shelf and Carbonate Counter pumps and their geographical and temporal variability is poorly constrained and needs to be better defined to facilitate modelling efforts. In particular, the importance of mesoscale variability in the carbon pumps is poorly understood at present.

4.4. Species biodiversity and functional groups

The diversity of species present in the plankton – from the viruses, bacteria and archaea to the largest zooplankton and fish – is immense and with modern genetic studies the true diversity is expected to be even larger. In addition to this genetic diversity there is also a diversity of function as pertains to the role a species or group of species plays in the ecosystem, including the contribution to carbon turnover by the biological pump. Over long time scales the relative dominance of functional groups is thought to have modulated carbon cycling between the ocean and atmosphere (Falkowski *et al.*, 2003).

Plankton assemblages that characterise particular biogeochemical functions are important: in the production, turnover and release of radiatively active gases and their exchange with the atmosphere (e.g. CO_2, DMS), in the relative proportion of organic material that is respired near the surface or is sequestered to the deep ocean, and in the cycles of major elements such as nitrogen and silica (Boyd and Doney, 2002, Jin *et al.*, 2006). The concept of functional groups is particularly applied in models to simulate the present and future role (in a changing environment) of biology and to estimate the contribution of organisms to global-scale element cycles (Le Quéré *et al.*, 2005). Changes in the relative importance of different functional groups in the plankton can strongly impact the Biological pump; for example, relative fluxes of diatoms versus calcareous plankton have been implicated as one of the causes for the changes in CO_2 between glacial and interglacial periods. The changes are attributed to substantial differences between the periods in nutrient inputs to the ocean from dust and rivers, sourced especially during glacial times from loess and coastal erosion (Harrison, 2000; Tréguer and Pondaven, 2000).

4.4.1. Changes in the benthos and sea bottom sediment

Benthic organisms and bottom sediments also contribute to the oceanic carbon cycle. Animals with carbonate skeletal systems live over a huge shelf area. It has been estimated (Andersson *et al.*, 2005) that coastal ocean surface water carbonate saturation state will decrease by 46% by 2100 due to acidification, leading to a decrease of 42% over the same period in the biogenic production of $CaCO_3$. Their modelling results also show that the carbonate saturation state of pore water in sediment will decrease in the future due to a greater deposition of both land derived, recycled and locally produced organic matter. This will lead to an increased dissolution of carbonate minerals in the sediments. The future reintroduction of carbon from sediments on the sea floor to seawater due to global warming will have a considerable impact on the atmosphere. Warming of shelf seas will change the rates of microbial production and thus gas exchange and nutrient supply—but potential impacts are largely unknown. Changes in the composition, biomass and production of the benthos of both shelf seas and the deep oceans are also likely to be important—but again the impacts are unknown.

4.5. Global and regional information

For modelling evaluation, validation and other studies of the processes involved in the ocean carbon cycle comprehensive information is needed on the spatial and temporal coverage of key parameters over a long period (Boyd and Trull, 2007; Le Quéré *et al.*, 2005). Information is available for mean fluxes of CO_2 (Takahashi *et al.*, 2002, 2009) and DMS (Kettle and Andreae, 2000; Kettle *et al.*, 1999), but there is limited temporal information. Global-scale observations of chlorophyll did not begin until 1978 with the operation of satellite measurements by the CZCS. SeaWiFS satellites have provided a global coverage of chlorophyll since 1997, although this is constrained by cloud cover in many parts of the world so that the coverage is piecemeal in places and at certain times of the year. New approaches to processing the multi-spectral characteristics of SeaWiFS data means that some individual plankton groups such as cyanobacteria and diatoms may also be estimated on a global scale (Raitsos *et al.*, 2008). High reflectance from coccoliths released into the water after coccolithophore blooms signifies that these phytoplankton can in part be determined on a global scale from satellites (Brown and Yoder, 1994; Iglesias-Rodriguez *et al.*, 2008). Nonetheless, satellite information is inadequate to clarify how these post-bloom events relate to carbon export. Also, the most important calcifying species are restricted to deeper water in the subtropics that cannot be detected from satellites. *In situ* data to calibrate the satellite measurements of phytoplankton are limited, and satellites provide no information on zooplankton. The Continuous Plankton Recorder surveys in the North Atlantic (Richardson *et al.*, 2006) and Southern Oceans (Hosie *et al.*, 2003) provide the only comprehensive coverage of selected phytoplankton classes and zooplankton diversity and abundance at monthly and regional, but not global scales.

Global information is available from more than 100 sediment trapping experiments (Francois *et al.*, 2002; Klaas and Archer, 2002, Boyd and Trull, 2007) used to determine downward fluxes, regeneration and detrital composition at single-site moorings throughout the ocean. Trap sampling methodologies have not been standardised, so there are problems of interpretation. Furthermore, trap coverage is very restricted, especially on continental shelves. A comparison between modelled estimates of export flux between the last glacial maximum and the present, with superimposed measurements from satellites is shown in Fig. 1.17. This figure further emphasises the limited spatial information that is available from sediment traps for some regions of the world.

It is clear that global coverage of key parameters needed to understand the ocean carbon cycle is limited and in most cases restricted to a short time series. This applies particularly to routine synoptic measurements at a species level of plankton that are needed for validation of satellite measurements. Finally, the modelled results in some locations are at odds with other calculations of palaeo-productivity, for example, estimated high productivity in the central Pacific during the last glacial maximum (LGM) determined

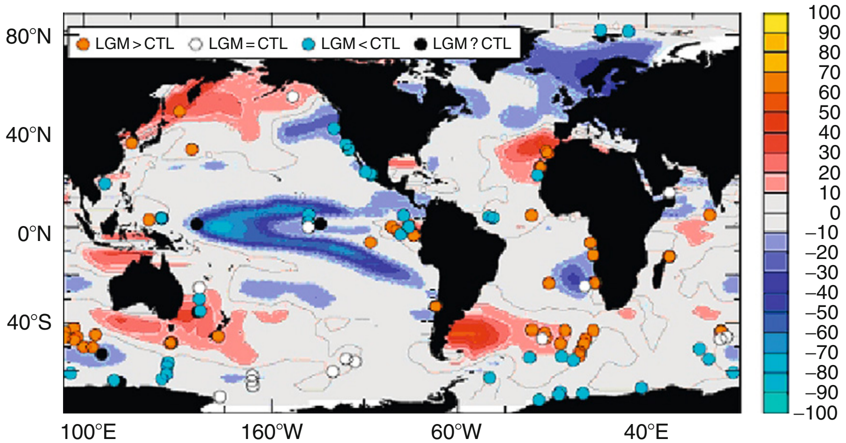

Figure 1.17 Observed (superimposed circles) and modelled changes in export at the LGM compared to the late Holocene (Bopp *et al.*, 2003). Model results are in percent. Observations are qualitative only and indicate a higher (red), lower (blue) or similar (white) export in the LGM compared to the present day. From Le Quéré *et al.* (2005).

from barite measurements (Paytan *et al.*, 1996) and from organic matter deposition (Perks and Keeling, 1998).

4.6. Ocean fertilisation

Seeding the oceans with iron as a micro-nutrient, an essential nutrient for healthy growth of most phytoplankton species, has been proposed as a mitigation measure against rising levels of atmospheric CO_2. General concerns exist that the science behind large-scale fertilisation of the oceans by nutrients (including iron) to increase the sequestration of atmospheric CO_2 by the oceans is still poorly understood. These concerns have been debated worldwide, for example, at the Woods Hole Oceanographic Institution in September 2007 (http://www.whoi.edu/page.do?pid=14617) and at the 30th meeting of the London Dumping Convention and associated London Protocol in December 2007. The latter meeting endorsed the concerns expressed by scientists, declared an intention to develop international regulations to oversee such activities, and advised that large-scale fertilisation schemes are currently not justified. An example of the latest scientific view is given in a press statement by the Scientific Committee on Oceanic Research (SCOR) and the Group of Experts on the Scientific Aspects of Marine Environmental Protection (GESAMP), which can be found here: http://www.imo.org/includes/blastDataOnly.asp/data_id%3D21214/INF-2.pdf. However, given the urgency of potential climate change impacts, there is a need to continue smaller scale experiments. Such experiments should have similar controls to those outlined by the SCOR and GESAMP statement to determine if manipulation of the oceans might be an effective means of mitigation to help reduce the effects of rising atmospheric CO_2.

4.7. Concluding comments

- If the combined efficiency of the ocean carbon pumps showed a marked decrease, there would be a strong positive feedback on atmospheric CO_2.
- It is estimated that the Solubility pump may already have become less efficient due to warmer temperatures.
- There is strong evidence to suggest that the Biological pump contributed importantly to the marked variation in levels of atmospheric CO_2 found between Pleistocene glacial and interglacial periods.
- There is limited understanding of processes and spatial and temporal variability in the Biological pump at the present day and how it may change in the next century and impact climate change.
- Drawdown of CO_2 by the Continental Shelf pump is likely to reduce over the next century due to warmer seas, compounded by improvements to urban waste water treatment.

- Changes in the relative importance of different functional groups in the plankton may impact the Biological pump.
- Changes in the composition, biomass and production of the benthos and in associated sediments of shelf seas and the ocean are likely to be important to climate change, but the impacts are difficult to assess due to limited data.
- Small-scale, well controlled experiments in ocean fertilisation should be continued as long as they adopt the controls outlined by SCOR and GESAMP.
- Global and regional coverage of many of the key biological measurements needed to determine fluxes to the deep ocean, including spatial and temporal variability of plankton functional groups and sediment trapping is limited.
- Lack of inclusion of some ocean carbon feedbacks in climate change modelling may lead to underestimates of the action required to stabilise emissions at given targets.

5. Ocean Acidification and the Carbonate Pump

The important role that the oceans play in the carbon cycle and in the uptake of atmospheric CO_2 is described in the previous section. As levels of CO_2 in the atmosphere increase due to anthropogenic emissions there is a larger uptake of CO_2 by the oceans across the air/sea interface. This transfer leads to higher levels of carbon in surface waters and by reaction, more acidic seawater, which is reflected in a lower pH (pH is a measure of acidity). This process is known as 'ocean acidification' (Denman *et al.*, 2007; IPCC AR4 WG 1, 2007, Box 7.3; Raven *et al.*, 2005) and is an independent consequence of rising levels of anthropogenic CO_2 separate from the Greenhouse Effect. Additional acidification in some coastal waters derived from anthropogenic nitrogen and sulphur deposition from fossil fuel combustion and agriculture may also increase in the future to further exacerbate the problem (Doney *et al.*, 2007). Levels of pH have declined at an unprecedented rate in surface seawater over the last century and are predicted to undergo a further substantial fall by the end of this century as anthropogenic inputs of CO_2 continue to rise sharply (Caldeira and Wickett, 2003). This is against a background where we know that emissions are already going up even faster than the maximum modelled projections of IPCC (Canadell *et al.*, 2007). There is real concern over the impact that such a large, rapid and unprecedented rise in acidification might have on marine organisms (Guinotte and Fabry, 2008; Hall-Spencer *et al.*, 2008; ICES, 2007), but little emphasis has so far been placed on the potential feedbacks from acidification to climate change. Acidification may cause positive or negative feedbacks to climate change through alterations in biogeochemical processes, nutrient speciation, trace metal availability and ecosystem biodiversity (Milliman *et al.*, 1999; Raven *et al.*, 2005), changes that may be accentuated in combination with rising temperatures.

5.1. The buffering of climate change by the oceans

The oceans have taken up ~40% of the anthropogenic CO_2 produced from fossil fuel burning and cement manufacture since the industrial revolution (Sabine *et al.*, 2004a). In doing so, the ocean is essentially buffering the effects of climate change from the even more elevated atmospheric CO_2 concentrations that it would be experiencing if it was not carrying out this important role. The costs of uptake of anthropogenic CO_2 by the surface of the world's ocean are higher bicarbonate ions, lower carbonate ions, higher hydrogen ions and reduced pH (i.e. a more acidic surface ocean). It should be noted that the oceans will not become truly acidic as their pH will remain above 7, even with the worst case scenario, due to the intrinsic buffering capacity of the oceans (this is the ability of a fluid to sustain a certain pH; in this particular case while absorbing CO_2).

Anthropogenic emissions cause an increase in the partial pressure of atmospheric CO_2 ($pCO_{2,\mathrm{atm}}$). As $pCO_{2,\mathrm{atm}}$ is typically larger than its equivalent over most of the upper mixed layer of the ocean ($pCO_{2,\mathrm{ocean}}$), there is a net flow of CO_2 from the atmosphere to the ocean. Note, however, that there is considerable spatial variability in relative net fluxes (see Section 4 and Fig. 1.18). During the dissolution of atmospheric CO_2 in seawater, most reacts rapidly with the water (H_2O) to produce carbonic acid at the same time as lowering pH. The reaction continues to produce bicarbonate ions and carbonate ions. This chain of reactions forms the carbonate buffer system that enables the ocean to take up much more CO_2 than would be possible from solubility alone. Only the remaining unreacted carbon dioxide fraction of DIC in the seawater takes part in ocean–atmosphere interactions (Denman *et al.*, 2007; Zeebe and Wolf-Gladrow, 2001). In typical seawater, the products of the reactions (Fig. 1.18) occur in the approximate proportions and ratios:

Bicarbonate (HCO_3^-)	~90%
Carbonate ions (CO_3^{2-})	~10%
Remaining aqueous carbon dioxide (CO_2)	~1%
Remaining carbonic acid (H_2CO_3)	Negligible

The sum of these various breakdown products of former atmospheric CO_2 are termed DIC:

$$\mathrm{DIC} = [CO_2] + [HCO_3^-] + [CO_3^{2-}] + [H_2CO_3]$$

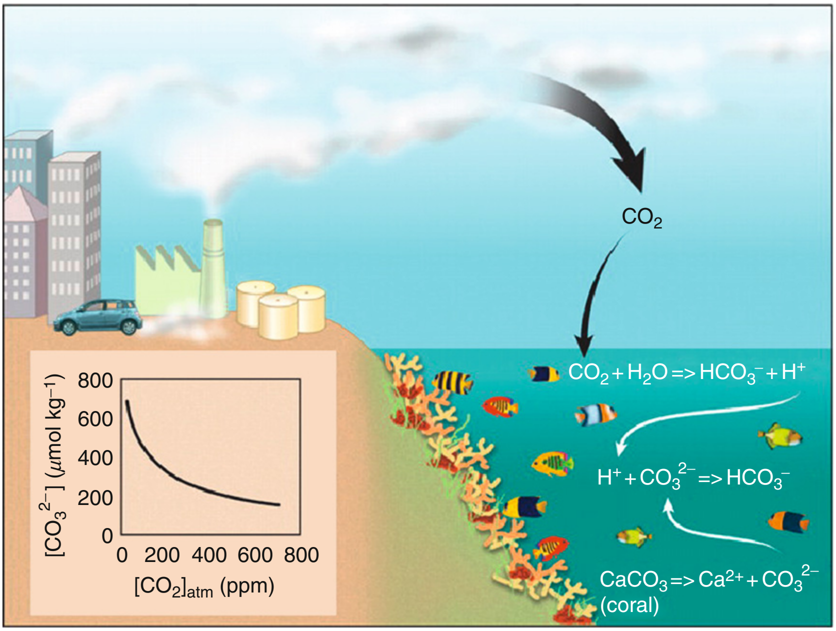

Figure 1.18 Linkages between the build-up of atmospheric CO_2 and the slowing of coral calcification due to ocean acidification. Approximately, 25% of the CO_2 emitted by humans in the period 2000–2006 was taken up by the ocean where it combined with water to produce carbonic acid, which releases a proton that combines with a carbonate ion. This decreases the concentration of carbonate, making it unavailable to marine calcifiers such as corals. Figure from Hoegh-Guldberg *et al.* (2007).

A net result of the chain of chemical reactions is that carbonate ions (CO_3^{2-}) are neutralised and reduced as a proportion of the DIC. On a global scale, this process means that the overall buffering capacity is reduced as levels of atmospheric CO_2 rise and more hydrogen ions (H^+) remain in solution. This will increase acidity (reduce pH). A summary explanation of the terms pH, DIC, carbonate buffer, carbonate saturation horizon, and a more detailed outline of the various reactions is given in Appendix 1 of Raven *et al.* (2005).

In equilibrium, the increase in dissolved CO_2 in the surface ocean is proportional to the atmospheric pCO_2, but the increase in DIC is not proportional to $pCO_{2,atm}$. This is due to the carbonate buffering capacity of seawater, which results in a smaller pH change, and can be explained by the Revelle factor (Zeebe and Wolf-Gladrow, 2001). The Revelle factor (or, previously buffering capacity factor) ranges between 8 and 15 units, depending on temperature and pCO_2. Due to the buffering capacity, an increase in DIC caused by acidification does not correlate with a 1:1 ratio to

Table 1.1 Changes in surface ocean inorganic carbon chemistry assuming equilibrium with atmosphere

	Pre-industrial	Present	Twice pre-industrial	Thrice pre-industrial
Atmospheric CO_2[a]	280	380	560	840
Surface ocean CO_2[b]	9	13	19	28
Surface ocean HCO_3^-[b]	1766	1876	1976	2070
Surface ocean CO_3^{2-}[b]	225	185	141	103
Surface ocean total dissolved inorganic C[b]	2003	2065	2136	2201
Surface ocean pH	8.18	8.07	7.92	7.77

[a] μmol mol^{-1}.
[b] μmol kg^{-1}.
Total alkalinity 2324 μmol kg^{-1}, 18 °C (modified from Table 1 in Raven *et al.*, 2005).

the increase of atmospheric CO_2 but rather ~1:10. Thus, the increase to the present day in atmospheric CO_2 of ~100 ppm from the pre-industrial 280 ppm represents a rise of ~36% whereas DIC has only increased by 3.1% (Table 1.1), approximately a 10-fold difference.

5.2. Carbonate formation

A second major process within the carbon chemistry of the ocean that has a crucial long-term role (see later section) in modulating levels of atmospheric CO_2 is the production of carbonates. Three types of minerals may be formed: calcite ($CaCO_3$), aragonite ($CaCO_3$) and magnesium (Mg) calcites, each with different solubility characteristics defined by their saturation state (symbolised by omega, Ω). Since calcium (Ca^{2+}) is extremely abundant in seawater and as a result its concentration difficult to alter, the saturation state of seawater with respect to calcium carbonate ($\Omega = [CO_3^{2-}][Ca^{2+}]/K_{sp}$) is almost always most strongly influenced by changes in the carbonate ion concentration. The formation of the different forms of $CaCO_3$ (see equation below) requires the presence of water that is supersaturated with carbonate ions, as is typically found at present in most of the upper mixed layer of the ocean.

$$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O$$

In the reaction that produces carbonates, DIC is reduced, alkalinity consumed, CO_2 released, acidification increased and pH lowered.

The solubility of calcium carbonate increases with pressure (depth) and with lower temperature. In consequence each of the three minerals has a different depth where Ω changes from saturation (>1) to undersaturation (<1). Below this depth, known as the saturation horizon, the minerals will dissolve unless they are protected by an organic membrane as part of a living organism or detrital aggregate. The depth horizon at which $CaCO_3$ starts to disappear from the sediments is known as the lysocline, and the depth at which it (almost) completely disappears is known as the 'compensation depth' in sediments. The lysocline and compensation depth are shallower for Mg carbonate, aragonite and calcite in that order. As the seawater saturation state of the North Pacific is lower than the North Atlantic, the aragonite saturation horizon almost reaches the surface in the North Pacific while it is at approximately 3000 m in the North Atlantic (Fig. 1.19). For the calcite form of $CaCO_3$ the saturation horizon varies between less than 1000 m in the North Pacific and more than 4500 m in the North Atlantic.

While calcification by carbonate minerals may, in favourable conditions, occur by precipitation, the vast majority is secreted by pelagic and benthic organisms to form complex tests and skeletal structures. Important calcifying groups include the microscopic protist foraminifera, algal coccolithophores that utilise calcite, corals and bivalves (including the pelagic pteropods with aragonitic structures and coralline algae supported by Mg calcite). Some of these organisms play a key role in the biological and carbonate pumps and form extensive areas of calcareous ooze on the bottom of the ocean.

In polar regions, the lysocline comes much closer to the surface and, as carbonates dissolve more readily in cold water, polar and sub-polar waters are particularly vulnerable to future changes in ocean carbonate chemistry. There is already evidence that the saturation horizons for aragonite and calcite are shoaling (Orr *et al.*, 2005) and organisms such as pteropods are thus especially under threat. This is an additional vulnerability for Antarctic and Arctic waters over the next century (Fig. 1.20) to those caused directly by climate change (Andersson *et al.*, 2008).

There have been observations to suggest that some pelagic and benthic marine organisms may increase their calcification rates with increasing acidification (Iglesias-Rodriguez *et al.*, 2008; Wood *et al.*, 2008a). However, the majority of experiments, models and field observations to date show a deleterious impact on calcifiers from acidification. It should be noted that the 'sudden' changes in pH in these short-term experiments may not be representative of nature and that to some extent organisms may be able to adapt to the slower longer term projected changes. This is a key area for research as the implications either way are important.

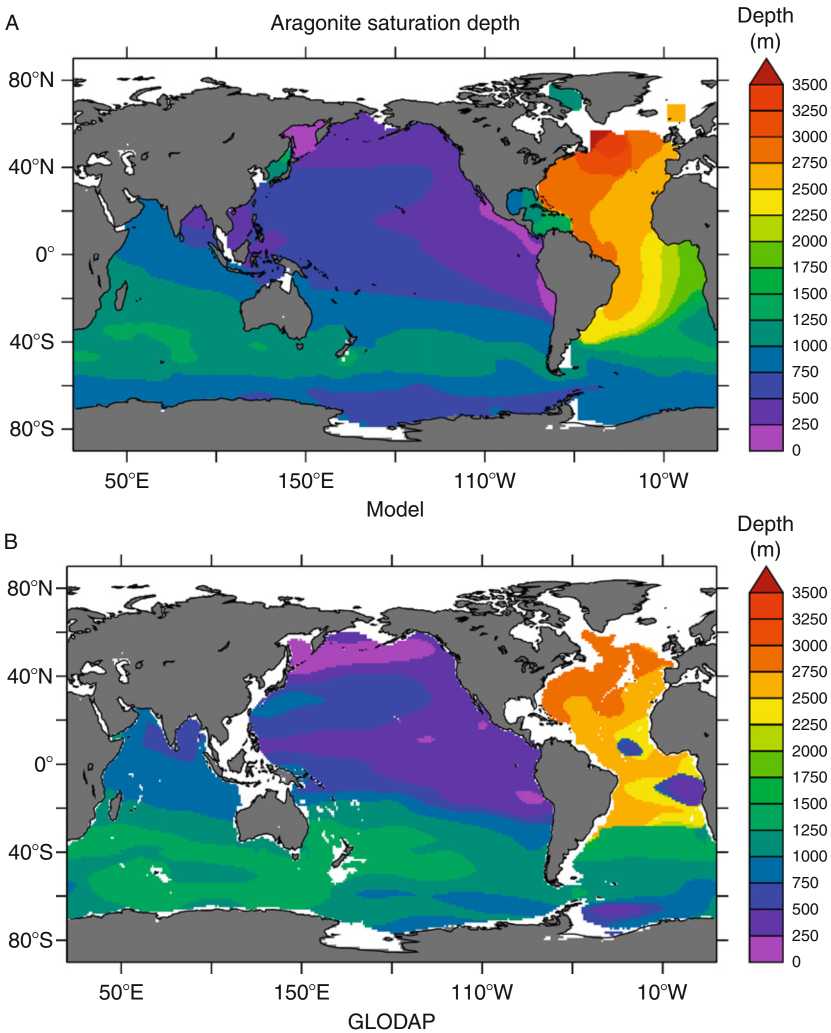

Figure 1.19 Depth of aragonite saturation horizon: lower map from measurements recalculated from GLODAP after Key *et al.* (2004) and upper map modelled calculations. Figure from Gangstø *et al.* (2008).

5.3. Carbonate dissolution

As atmospheric CO_2 increases in the long-term and penetrates deeper into the ocean due to the THC and downwelling and via the solubility, Biological, Carbonate and, ultimately, the shelf sea pumps, the lysoclines for calcite, aragonite and Mg calcite will shoal (come closer to the surface). Previously sedimented carbonates will start to dissolve increasing dissolved

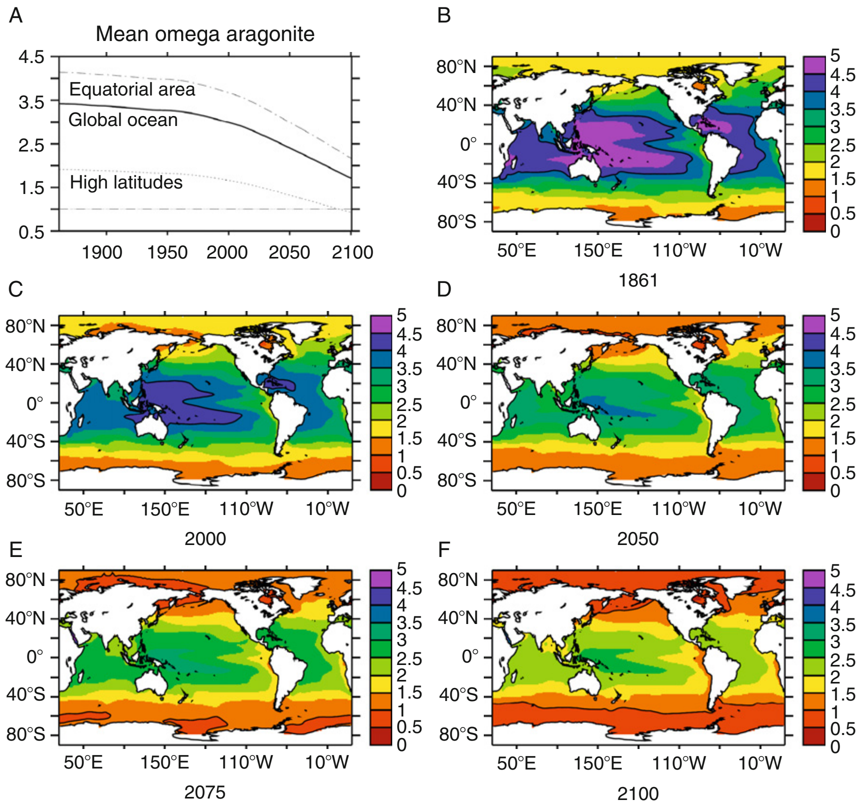

Figure 1.20 Saturation state with respect to aragonite of surface waters (0–100 m): (A) time series of mean Ω for the global ocean, the equatorial area and for high latitudes, and maps in year (B) 1861, (C) 2000, (D) 2050, (E) 2075 and (F) 2100. Figure from Gangstø *et al.* (2008).

carbonate alkalinity. Total alkalinity, which in terms of ocean buffer capacity is more important, will also increase. Thus, when the water returns to the surface over a timescale of several centuries, further CO_2 can be removed from the atmosphere and acidification of surface waters will be partly reversed.

In a world of increasing CO_2 the dissolution of calcium carbonate can be expressed as

$$CO_2 + CaCO_3(s) + H_2O \rightarrow 2HCO_3^- + Ca^{2+}$$

Dissolution of $CaCO_3$ minerals in the surface layers of the oceans acts as a further buffer of pH and carbonate saturation state against acidification from an increasing ocean uptake of atmospheric CO_2 (Andersson *et al.*, 2006). However, "$CaCO_3$ dissolution has a negligible impact on atmospheric

pCO_2 or the atmospheric stabilisation of CO_2 emissions" over the next few centuries (Archer *et al.*, 1998). Even if dissolution of $CaCO_3$ was taking place at a rate equivalent to the estimated total annual production of $CaCO_3$ in surface waters, it would still only be partially buffered (estimated maximum 6%). This is especially so as most $CaCO_3$ production is pelagic and sinks to deeper depths (Andersson *et al.*, 2006). Dissolution from bottom sediments as the lysocline shoals in response to the reduction in pH over the next century and longer will enable the oceans to increase their CO_2 sink.

5.4. Uptake of CO_2 by the ocean

There is evidence that key ocean sinks for CO_2 (the North Atlantic and Southern Ocean) may already be reducing their rate of CO_2 uptake (Le Quéré *et al.*, 2007, 2008; Schuster and Watson, 2007). However, these measurements have not yet been taken over sufficiently long periods to distinguish whether this is from natural variation, changes in buffering or other causes. In the North Atlantic, for example, the reduced uptake has been linked to a decline in mixing and ventilation between surface and subsurface waters. This is due to increasing stratification associated with changes in the NAO, exacerbated by the changing buffer capacity of seawater as the carbon content of surface waters increased (Schuster and Watson, 2007). Sustained backbone observations such as carried out in these studies are important to continue but are under threat.

Over time periods of <10,000 years, the ocean is particularly sensitive to increases in pCO_2. On a longer term basis, the buffering capacity of the ocean will become greater (as measured by the Revelle factor), as buffering caused by the dissolution of carbonate sediments moderates the effect of pH change. On a 1000–100,000-year timescale, it is estimated that $CaCO_3$ dissolution will absorb 60–70% and the oceanic water column 22–33% of anthropogenic CO_2 emissions (Denman *et al.*, 2007). Thus, eventually, the oceans and their sediment $CaCO_3$ will buffer or neutralise the CO_2. However, the new concentration level of CO_2 in the atmosphere will never return to pre-industrial levels (Andersson *et al.*, 2003, 2005; Archer *et al.*, 1998).

High atmospheric CO_2 does not automatically correlate with lower pH, because it can vary while pH remains constant if DIC changes. This fact is particularly relevant to the interpretation of palaeo-evidence where acidification and calcification were thought to be high at the same time. If acidification takes place gradually, as appears to have occurred at times in the geological record, then some of the potential change in pH may be absorbed by dissolution of sediments. During the Palaeocene–Eocene thermal maximum (PETM), it is calculated that following the initial acidification there was a widespread dissolution of sea-floor carbonates, a pattern that replicates modelled response to the anthropogenic rise in CO_2 (Zachos *et al.*, 2005). Changes in temperature may also alter pH, but any feedback to climate is likely to be trivial. Any effect is likely to be in the other direction as warmer waters absorb less CO_2. Warmer temperatures will increase stratification and thus reduce the volume of mixed water available for CO_2 absorption from the atmosphere (Raven *et al.*, 2005). And as a further consequence the increased stability will lead to a reduction in the return flow of carbon and nutrients from the deep ocean, reduced primary production and thus lower uptake of CO_2.

5.5. Projected future levels of acidification

Recorded on a logarithmic scale, pH has reduced as a global average in surface seawater since the beginning of the industrial revolution by ~0.1 units (current mean level pH ~ 8.08, pre-industrial ~8.18, last glacial maximum ~8.35). This is equivalent to a 30% increase in the concentration of hydrogen ions (pH is a measure of the free positive hydrogen ion concentration; measured in seawater as the total concentration; see Zeebe and Wolf-Gladrow, 2001). With continued 'business-as-usual' use of fossil fuels, pH is estimated to decrease by a further 0.4 by 2100 and 0.77 units by 2300 (Caldeira and Wickett, 2003). The rate of change and degree of change are unprecedented for likely the last 20 million years (Raven *et al.*, 2005; Fig. 1.21) and possibly since the PETM, 55 million years ago (Zachos *et al.*, 2005). The most pronounced changes in acidification are seen in the North Atlantic extending down to 5000 m, a much deeper depth than previously thought, due to the deep water formation that occurs there

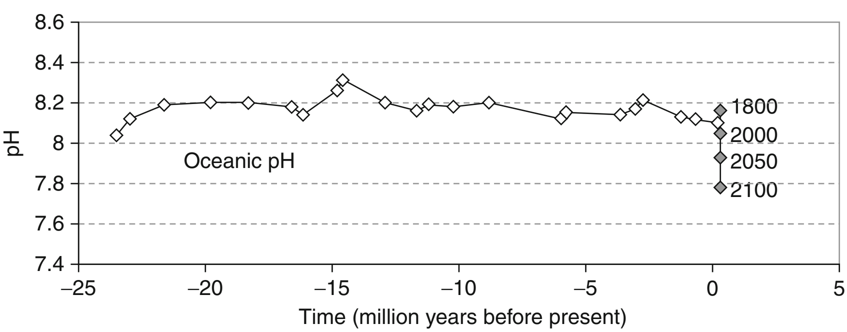

Figure 1.21 Changes in the level of seawater pH over more than the last 20 million years. Geological estimates (white diamonds) taken from Pearson and Palmer (2000) (method considered by some researchers to be unreliable when older than a few million years). Calculated mean oceanic pH levels for 1800 and 2000 shown on a vertical line against dates with modelled future predictions for 2050 and 2100 based on IPCC mean scenarios (grey diamonds with dates). Figure from Turley *et al.* (2006).

(Tanhua *et al.*, 2007). These changes reflect a high input of anthropogenic carbon and possibly indicate that the oceans can take up more carbon than previously thought over the next century.

5.6. Regional variation in acidification

Changes in acidity measured in the open ocean appear to be extending to some shelf seas. For example, Feely *et al.* (2008) have shown a pronounced shoaling and an increase in the extent of upwelled undersaturated water on the western continental shelf of North America. This region of the North-east Pacific has one of the shallowest aragonite saturation horizons in the globe and it appears to have shoaled by about 50 m as a response to acidification now allowing acidified waters to penetrate extensively onto the shelf. The impact has been pronounced with deoxygenation and mass mortality of benthic organisms. If a similar situation arose in the Arctic it might further exacerbate the vulnerability of calcareous organisms in this ocean to acidification. Within the second half of the last century input of bicarbonate (DIC) by the Mississippi river increased substantially, primarily as a consequence of human agricultural changes. This marked change in one of the major rivers of the world may have provided a localised buffering system for ocean acidification in parts of the Gulf of Mexico (Raymond *et al.*, 2008). It is not clear how such regional variability in acidification might feedback to climate change.

5.6.1. Carbonate biology: Plankton

Laboratory and field observations suggest that ocean acidification enhances photosynthetic carbon fixation in the major phytoplankton functional groups of the modern ocean namely cyanobacteria nitrogen fixers (Hutchins *et al.*, 2007), diatoms (Tortell *et al.*, 2000) and coccolithophores (Iglesias-Rodriguez *et al.*, 2008; Riebesell *et al.*, 2000, 2007). Unlike in corals, the effect of ocean acidification on calcium carbonate-producing phytoplankton is, however, unclear, and shows a non-uniform response across species in laboratory experiments (Iglesias-Rodriguez *et al.*, 2008; Langer *et al.*, 2006; Riebesell *et al.*, 2000) although most experiments, including recent ones (Feng *et al.*, 2008), have shown a decline in the ratio of inorganic/organic carbon in coccolithophores at higher CO_2. These laboratory experiments may, however, not reflect natural oceanic conditions. In the open ocean, the effect of ocean acidification on calcification remains an open question. It is the balance between calcification and photosynthetic carbon fixation that controls whether calcifying phytoplankton represent a sink or a source of CO_2 to their surrounding environment (see Frankignoulle *et al.*, 1994), and this information is crucial to elucidate changes in the contribution of taxa to changes in the magnitude and direction of CO_2 fluxes. Assessing the variability of responses across taxa is a challenge, particularly in the light of evidence of variability in other calcifying groups within marine invertebrates (Ries *et al.*, 2008; Wood *et al.*, 2008b). Whether intraspecific physiological variability adds to the complexity of these responses or whether natural populations respond uniformly to changes in carbonate chemistry is the next step in assessing the reciprocal interactions between changing pCO_2 and the carbon signature associated with biotic responses to ocean acidification. Higher ocean pCO_2 will lead to increased acidity, a lower pH and lower relative calcium carbonate saturation (omega). A temperature rise will, however, increase the relative calcium carbonate solubility, omega. Any effects from a combination of a reduction in pH and rise in temperature will depend on the relative rates of change of the two variables.

5.6.2. Carbonate biology: Coral reefs

While only covering ~2% of the area of continental shelves, corals through their calcification account for ~33–50% of the global production and accumulation of inorganic $CaCO_3$ (PIC) (Borges, 2005). There is now, however, considerable evidence that these levels of coral calcification will be severely impacted by future projected ocean acidification (Tyrrell, 2008). A linear relation has been demonstrated between saturation state and calcification for coral reefs, but these experiments were performed in biospheres and tanks where the corals were stressed and growing at a slower rate. Recent research has shown, however, that in combination with rising temperatures calcification rates have already declined (De'ath *et al.*, 2009). It is estimated that lower rates of calcification will lead to a reduction in coral $CaCO_3$ on a global scale of between 9% and 30% over the next 50–100 years (Gattuso *et al.*, 1999; Kleypas *et al.*, 1999).

A range of experimental results show that coral calcification, structure and growth will be reduced by up to 40% for a doubling of pre-industrial atmospheric CO_2 to 560 ppm (Hoegh-Guldberg *et al.*, 2007; Wood *et al.*, 2008b). These authors also showed, both experimentally and by comparison with present distributions, that aragonite formation ceases at saturation values of 3.3. Acidification, however, is not the only process impacting coral reefs; other phenomena such as extreme temperatures (coral bleaching), viral attacks, starfish predation, dust and precipitation as well as over fishing, pollution and physical damage also need to be taken into account. The prognosis for reef corals is dire with serious consequences for the many millions of people who depend on them for their homes and livelihoods, as biodiversity hot spots, for shore protection, local fisheries and tourism (Carpenter *et al.*, 2008; Hoegh-Guldberg *et al.*, 2007; Pandolfi *et al.*, 2003). Because of shoaling of the aragonite lysocline, cold water corals are also seriously threatened by acidification. However, the ability of coral species to adapt to change, and especially to the rapid rate of change in pH is not yet clear.

While corals only cover a small area of the global ocean, they are expected to continue to be major players in the carbon cycle over the next 100–1000 years because the formation and deposition of $CaCO_3$ during their growth is so intense and because rather little of it dissolves. The breakdown of coral reefs in glacial periods due to lower sea-levels is considered. (Coral reef hypothesis of Berger (1982) as one of the possible causes of the alternation of CO_2 levels between glacial and interglacial times Kleypas *et al.*, 2006.)

5.6.3. Carbonate biology: Benthos

In attempting to identify the impact of ocean acidification on the marine benthos, it would be unrealistic to focus excessively on the impacts of lowered pH on animals with calcareous body parts. The physiology of the marine biota is finely regulated and has evolved to function within relatively narrow pH and CO_2 ranges (Michaelidis *et al.*, 2005). In a more acid ocean, animals operate under sub-optimal conditions and hence energy apportionment between respiration, repair, growth and reproduction will change. The latter two processes will suffer as more energy is consumed by repair and respiration. While there has been considerable concern about the impacts of ocean acidification on animals with calcareous body parts, some species appear to be able to increase the rate of calcification at a lower pH (Wood *et al.*, 2008b) although this is at some metabolic cost and may not be sustainable. There are therefore large uncertainties in the adaptation capabilities of marine species and functional groups and, in consequence, any feedback effects to marine climate.

The scale of impacts on populations and assemblages resulting from a decline in growth and reproductive rates has yet to be quantified, but it is believed that coralline algae are particularly vulnerable as they utilise Mg calcite. Experimental studies by Albright *et al.* (2008), Jokiel *et al.* (2008) and Kuffner *et al.* (2008), for example, have shown a marked reduction in the growth and recruitment of both coralline algae and corals at elevated levels of pCO_2 comparable to those likely to be experienced near the end of the century. Hall-Spencer *et al.* (2008) have demonstrated the effects of acidification by studying shallow benthic ecosystems adjacent to volcanic CO_2 vents along pH gradients. Rocky shore communities with abundant calcareous organisms showed significant reductions in sea urchins, coralline algae and the absence of scleractinian corals at the extreme of the gradient with evidence of dissolution of gastropod shells.

5.7. Carbonate pump

This pump involves the production and dissolution of the three calcium carbonate minerals (primarily calcite and aragonite), their transport to the deep ocean by sedimentation and a contribution to increased levels of pCO_2

in the surface ocean. Some of the pCO_2 will escape out of the ocean to contribute in a small way to increased levels of atmospheric CO_2. This pump is termed the $CaCO_3$ counter pump in Denman *et al.* (2007). Sedimenting coccolithophores, calcareous resting cysts of dinoflagellates, foraminifera and pteropods form most of the settling material. The mineral component may, because of its higher density, provide an important ballast that ensures that detrital material from the dead organisms settles rapidly within aggregates, mucus nets and faecal pellets. Measurements from sediment traps have shown that the net deposition rate in the carbonate pump is comparable to the organic matter that is deposited by the biological pump. The highest production of carbonate minerals is in coastal upwelling areas and within subtropical gyres. Any reduction in calcification due to acidification of planktonic organisms could have a serious impact on the rates of settling out of both organic and calcareous material from the plankton with important feedback implications.

The subtropical gyres play a large role in carbonate production (Carbonate pump) and are predicted to expand in area, but not in productivity, in a warming world (Behrenfeld *et al.*, 2006). They are sensitive areas, but not the focus for much research or monitoring. There is limited understanding of the different regions where both the biological and carbonate pumps are most active, but some indication can be seen from bottom sediment distribution (Fig. 1.22).

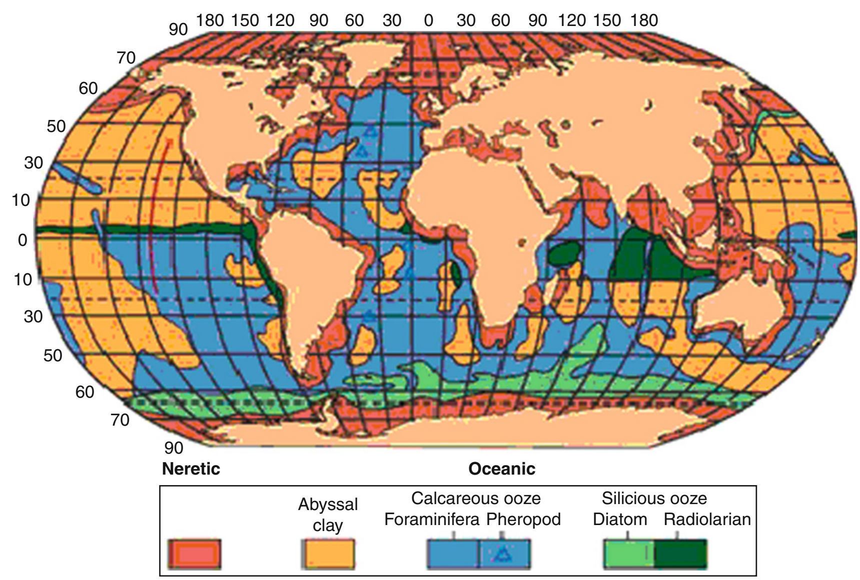

Figure 1.22 Global map of the distribution of different sediment types on the bottom of the ocean. Source: www.radiolaria.org.

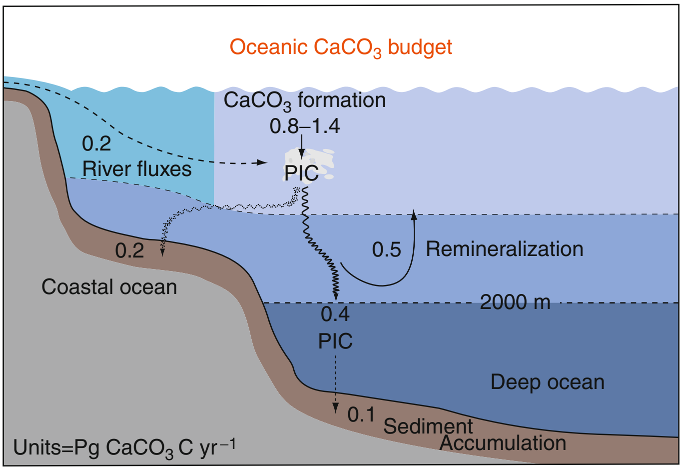

Figure 1.23 Schematic $CaCO_3$ budget. Figure from Sabine *et al.* (2004b).

Based on the global carbonate budget calculated by Milliman *et al.* (1999), about 20% of the pelagic carbonate production accumulates in deep sea sediments (Fig. 1.23). These authors also estimate that 60–80% of the planktonic calcium carbonate formed is dissolved in the upper 500–1000 m of the ocean by reprocessing and packaging in the guts of zooplankton and settling aggregates. Previous to this study it was thought that dissolution did not take place until particles settled below the carbonate lysocline. This dissolution buffers acidification and if it is within the penetration layer for anthropogenic CO_2 would allow the oceans to take up more CO_2 (Milliman *et al.*, 1999).

Shallow-water benthic production of carbonate minerals is also important both within and above the sediment, but little of this material will end up in the deep ocean. Potential dissolution of carbonate minerals on the shelves in the future as pH levels fall further may have important implications for shelf ecosystems with unknown feedbacks to the carbon cycle.

The Biological pump is also likely to be seriously affected by acidification through change in plankton ecology and in the physiological processes involved in organic and carbonate production.

5.8. Nutrients

It is known that there are interactions between nutrients (including DIC), and photosynthetically active radiation (PAR) in the way that algae take up and assimilate DIC. Most algae, as well as corals and seagrasses, have concentrating mechanisms which can increase the rate of CO_2 assimilation, especially at low DIC concentrations (Giordano *et al.*, 2005). The discovery of the widespread and abundant occurrence of PR genes in the oceans opens up the possibility of 'a previously unrecognised pathway of energy capture on Earth' by heterotrophic bacteria (Karl, 2007); the consequences for biogeochemical cycles are as yet unknown. Changes in the availability of ammonium or nitrate as the nitrogen sources and of the supply of phosphorus and iron can also affect uptake rate and growth. It is expected that these effects will modulate the way in which growth by photosynthetically different primary producers respond to increased CO_2, but much more research is needed before confident predictions can be made. In considering the effects of increased CO_2 on growth above it was pointed out that CO_2 assimilation continues to increase with rising CO_2 even after growth (cell division rate) has saturated. This will alter the food quality to the next, and perhaps higher, trophic levels. It is not clear if there are biologically important effects of the changes in ionisation state with decreasing pH of, for example, ammonium and phosphate. Finally, pronounced changes in N_2 and CO_2 fixation rates have been described from experiments on the cyanobacteria *Trichodesmium* at atmospheric levels of CO_2 up to 1500 ppm, implying potentially large impacts on the N and C cycles and on phosphate availability (Hutchins *et al.*, 2007).

5.9. Palaeo-comparisons

By measuring CO_2 in bubbles of air (Fig. 1.24) in layered ice cores from both Antarctica and Greenland, the cyclic alternation of CO_2 and temperature between Pleistocene glacial and interglacial periods has been documented and recently extended back to 800,000 years before present (Luthi *et al.*, 2008). From the last glacial maximum to prior to the industrial

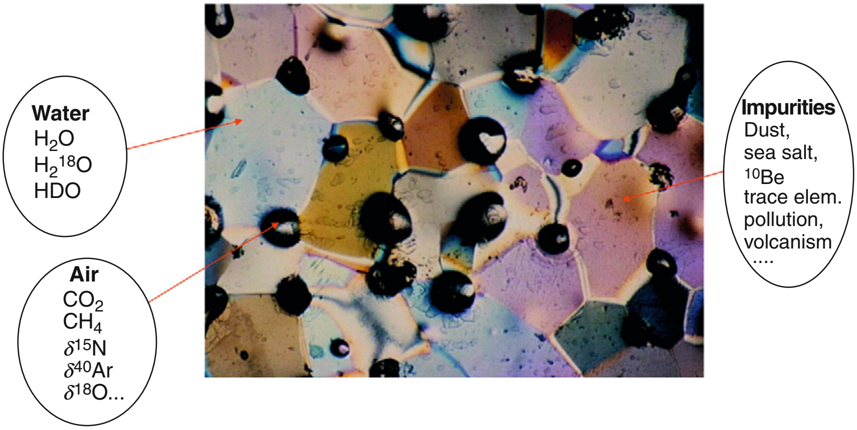

Figure 1.24 Bubbles of air in polar ice observed in a thin section under polarised light. Text redrawn from Raynaud D. EPICA lecture (2008 Ocean Sciences Meeting, Orlando, USA). Image: Copyright Michel Creseveur, CNRS/LGGE.

revolution CO_2 levels in the atmosphere increased by 40%, likely representing a release from the ocean reservoir. It should be noted, however, that ice core CO_2 measurements are 'smoothed' by diffusion and dating problems. Higher resolution data from oak leaf stomata reveal more 'natural' variation (up to 34 ppmv) of CO_2 in the millennium (van Hoof *et al.*, 2008). The authors show that their stomata-based CO_2 trends correlate with changes in Atlantic SST trends and suggest that this may indicate that changes in oceanic sources/sinks may be the mechanism behind the recorded CO_2 variability.

Prior to the Pleistocene, atmospheric concentrations of CO_2 have to be determined from proxies or are calculated using geochemical models. At timescales of millions of years during the Phanerozoic atmospheric CO_2 was dependent on the balance between volcanic sources (there were major periods of volcanic activity in Earth history) and consumption of CO_2 by weathering of silicate minerals in terrestrial rocks (this process is rate dependent on temperature) followed by deposition of carbonate sediments on the sea floor (François *et al.*, 2005) and by changes in photosynthesis. On timescales of tens of thousands of years, weathering does not affect major ocean chemistry to any great degree; such changes take place in the longer term. Rates of organic carbon deposition in rocks are also important. Both the latter rates and weathering are highly dependent on evolution, especially of the angiosperms, as well as extinction events and tectonic activity (Berner and Kothavala, 2001).

The geological record provides evidence of major changes in ocean chemistry that are linked to levels of atmospheric CO_2 such as fluid inclusion and other evidence that calcium concentrations approximately halved and magnesium concentrations approximately doubled over the last 100 million years (e.g. Tyrrell and Zeebe, 2004). Mackenzie and Pigott (1981) were the first to note that the carbonate oolites and cements in calcareous sediments oscillated between calcite/dolomite and aragonite through geological time (subsequently termed calcite–dolomite or aragonite seas). Later work showed that biological skeletal precipitates show the same alternation. These periods reflect changing environments and climate as well as Mg to Ca ratios in seawater, and atmospheric and seawater CO_2 concentrations (Arvidson *et al.*, 2006).

Using the MAGic model of Arvidson *et al.* (2006) and a relatively small number of calculated chemical parameters, Mackenzie *et al.* (2008) characterised the history of atmosphere and ocean composition during the Phanerozoic Eon (the last 545 million years). The two major oscillatory chemostatic modes (Fig. 1.25) are distinguished by differences in seawater carbonate saturation state, major ion chemistry, especially SO_4/Ca and Mg/Ca ratios, degree of ocean acidification, and atmospheric CO_2. The computed trends agree with fluid inclusion data for Mg/Ca and SO_4/Ca ratios through Phanerozoic time and the mineralogy of the dominant carbonate precipitates. During the earlier part of the Phanerozoic (the Palaeozoic),

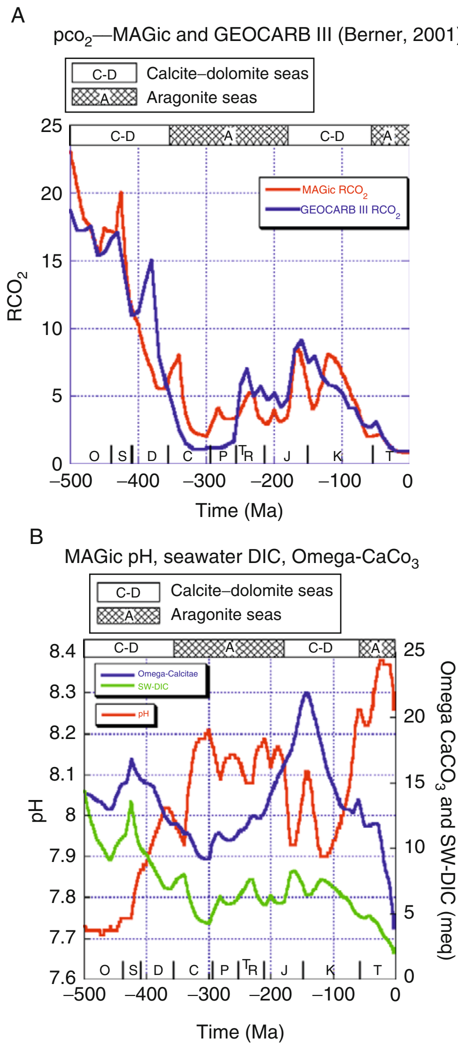

Figure 1.25 (A) Atmospheric levels of CO_2 through the Phanerozoic calculated from two different models with the periods of geological time distinguished from O (Ordovician) to T (Tertiary) on the bottom of the figure and alternations between calcite–dolomite and aragonite seas at the top. (B) Plots of pH, Omega $CaCO_3$ and DIC plotted over the same period. From Mackenzie *et al.* (2008).

seafloors were covered in carbonates and atmospheric levels of CO_2 were very high in contrast to the succeeding Mesozoic and Tertiary when terrigenous silicilastic sediments predominated (Peters, 2008). In the Cretaceous (Fig. 1.25) high levels of atmospheric CO_2 coincided with high rates of $CaCO_3$ formation, including the proliferation of coccolithophorids. An important element of the high CO_2 world at that time was the higher sea-level, which meant that the area covered by shelf seas was much larger allowing the deposition of extensive layers of chalk. The modelling of Arvidson *et al.* (2006) suggests that pH was low and carbonate saturation state higher in the Cretaceous. There is however, a contrasting view, that the Cretaceous ocean may not necessarily have been acidic and that the modelled saturation states proposed above would have made the calcite compensation depth too deep.

Andersson *et al.* (2008) have proposed the hypothesis that the more modern Earth system, in terms of the mineral composition of biogenic calcifiers and carbonate sediments, and because of rapidly increasing levels of atmospheric CO_2 and ocean acidification, may currently be in process of a transition from an aragonite sea to a condition that is more characteristic of a calcite sea. If such an event occurs it will be without a change in the Mg/Ca or SO_4/Ca ratios of seawater. It is predicted however, that the Mg content of calcitic hard parts in marine organisms is likely to decrease, the proportion of stable carbonates formed (e.g. calcite) increase and the Mg content of carbonate sediments decrease. Such changes have occurred in geological time and have been accompanied by profound alterations in marine ecosystems and biogeochemical processes.

5.10. Concluding comments

- The impact on climate change from ocean acidification is unclear.
- The structure of marine ecosystems and the physiological responses of marine organisms are expected to be severely impacted by acidification with potential extinctions, primarily because of the speed of change that is taking place.
- Reduction in carbonate mineralisation due to acidification may have an important impact on ballast in sedimenting particles, likely leading to more recycling in the upper ocean and a lower uptake of atmospheric CO_2.
- Changes in the relative location and intensity of the biological and carbonate pumps may have important feedbacks to climate change.
- While corals cover only a small part of global shelf systems they are still expected to be major players in the carbon cycle over the next 100–1000 years due to their intense growth and limited dissolution.
- Acidification is expected to have very serious consequences for the survival and growth of corals.
- Over a timeframe of thousands of years the oceans will be able to continue to take up much, but not all, anthropogenic CO_2 due to carbonate dissolution in the deep sea.

6. A Special Case: The Arctic and Seas Adjacent to Greenland

6.1. Climate change in the Arctic Ocean and Subarctic seas

The Arctic Ocean (Fig. 1.26) has a central role in global climate. Its key attributes are its high latitude, marked seasonality of insolation, unique enclosed nature and high reflectance of sunlight (albedo) from sea-ice, adjacent glaciers and snow cover. Enclosing the ocean is a terrestrial

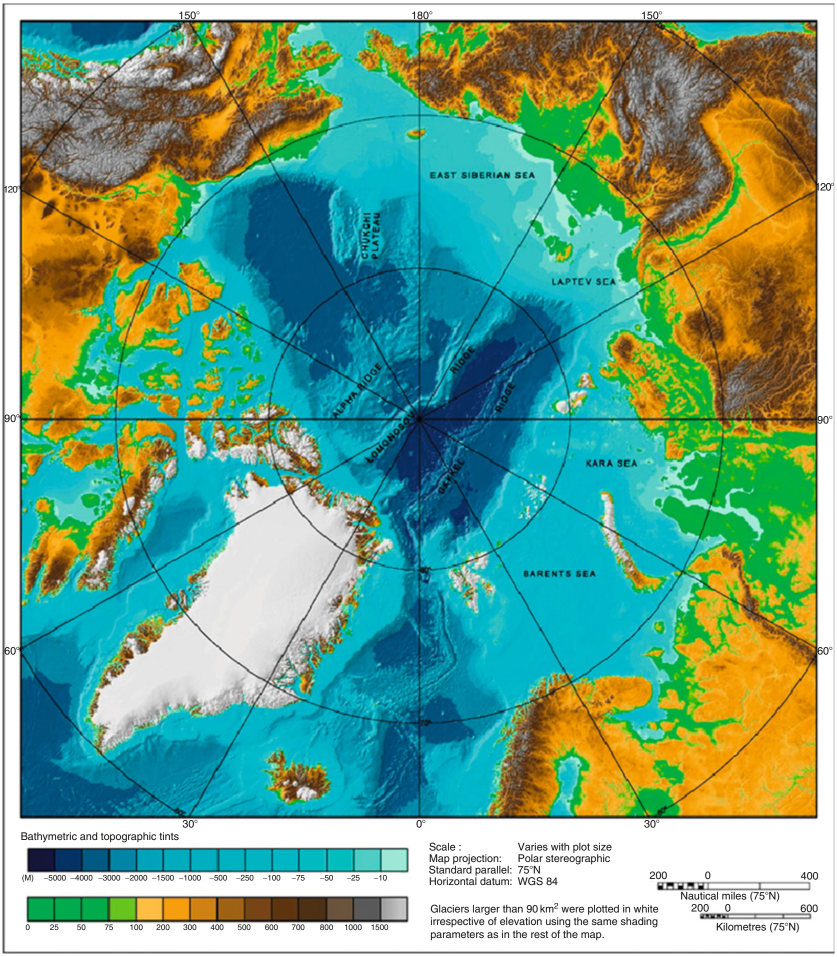

Figure 1.26 Map showing the geographical features and bathymetry of the Arctic Ocean and adjacent seas. From Jakobsson *et al.* (2008).

environment that is dominated by the cryosphere, either seasonally on the surface, or permanently just below the surface (i.e. permafrost). As a result of a strong ice–ocean influence, small changes in temperature and salinity may trigger large and sudden changes in regional climate with potential downstream feedbacks to the climate of the rest of the world.

It is clear from the Arctic Climate Impact Assessment (ACIA, 2005), the IPCC AR4 and more recent publications that the Arctic region as a whole is changing rapidly. While there are few long-term measurements, it is thought likely that Arctic air temperatures have been increasing since the beginning of the last century and certainly since the 1950s, when more observations became available. During the twentieth century, it is estimated that the Arctic warmed at a rate that was 50% faster (0.09 °C compared to 0.06 °C per decade) than the average for the whole of the Northern Hemisphere (ACIA, 2005; Fig. 1.27). However, Polyakov *et al.* (2002) consider that the Arctic long-term trend may not have been amplified with respect to the global trend, and instead that the difference is a consequence of poor seasonal sampling coverage in the Arctic that is hiding pronounced interdecadal variability. New research by Kaufman *et al.* (2009) has shown that the Arctic cooled progressively over the last 2000 years until ~1900 since when the trend reversed sharply to give from 1950 four of the warmest decades in two millenia. Precipitation has also increased and, together with the temperature increase, has led to a chain of other rapid changes within the last two decades including rising river flows, changes in ocean salinity, thinning of permafrost, declining snow cover, melting of glaciers and the Greenland ice sheet, rising sea-levels and most markedly rapid retreat of summer sea-ice extent and a reduction in its thickness.

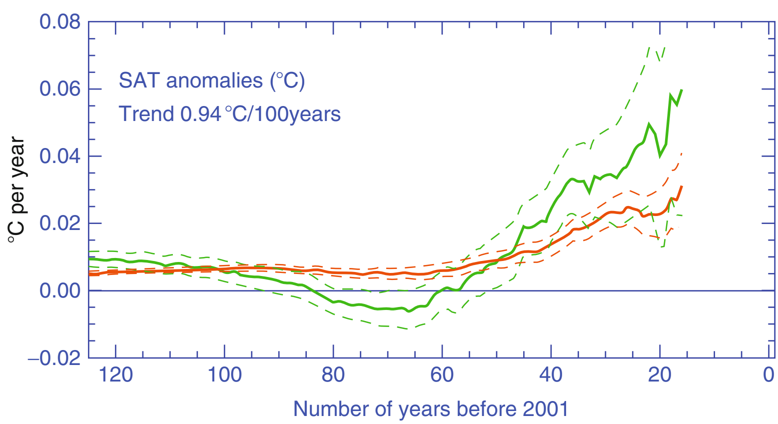

Figure 1.27 Surface atmosphere air temperature trends (°C per year) averaged for the Arctic (green) and Northern Hemisphere (red) (Jones *et al.*, 1999) with 95% significance as dashed lines from Polyakov *et al.* (2002).

6.2. The circulation of the Arctic Ocean and sub-polar seas

Warm input to the Arctic Ocean comes from extensions of the Gulf Stream (Fig. 1.28). The North Atlantic Current and the European Slope current releasing heat and water to the atmosphere en route. These currents continue north into the Arctic via the Norwegian Sea, as an outer meandering and a topographically constrained current at the edge of the shelf. As the warm saline Atlantic water moves into the Arctic Ocean and loses heat it becomes denser and sinks beneath a cold halocline layer (~200 m) and circulates throughout the Arctic Ocean. Mixing and diffusion spread both heat and salt upward into the surface waters.

The counterbalancing deep outflow from the system is fresher and primarily driven by temperature with sources from the Arctic shelf seas and deep convection sites in the Greenland and Labrador seas. This water forms the southern out-flowing limb of the MOC in the North Atlantic. The exchange of water and heat is delicately balanced and highly dependent on the rate of sea-ice formation that in turn is governed by temperature and salinity.

The upper surface of the cold and dense Arctic-sourced bottom water in the Norwegian Sea has lowered markedly over the last two decades (Dickson *et al.*, 1996). This suggests that dense water formation has declined, and implies that the MOC might also have been reduced. At present, however, there is no indication of a slowing of the MOC (see Section 2), but some strong evidence for increased inflow of warm saline Atlantic water into the Barents Sea and Arctic that falls counter to this suggestion.

The cold, dense water emanating from the Arctic has a further hurdle to cross before it becomes incorporated into the main circulation of the MOC in the North Atlantic. The relatively shallow sills that extend between

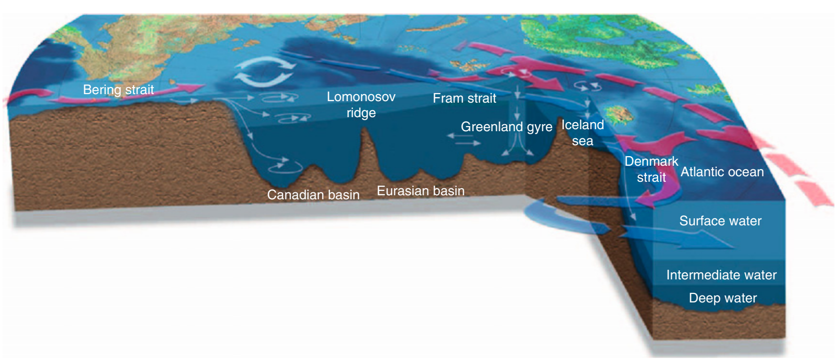

Figure 1.28 Schematic of Arctic circulation (ACIA, 2005).

Greenland, Iceland the Faroes and Scotland with the two main overflow points through the Denmark Strait and the Faroe Shetland Channel mean that there is no direct connection between the Arctic and the global ocean (Hansen *et al.*, 2008). Understanding and measuring the variability of outflow and inflow at these two sites as part of the MOC has and will continue to be a major area of research into the future (Dickson, 2006; Dickson *et al.*, 2008).

6.3. Runoff from Arctic rivers

Increased melting of permafrost and higher levels of precipitation in Russia and Canada (ACIA, 2005) has led to a considerable increase in river runoff to the Arctic. This is in turn leading to changes in nutrients and circulation. Ice-free coastal waters are likely to be more turbid and less productive due to light limitation, in addition to showing increased stratification due to riverine inflow. Basin wide, higher river flows will increase the intensity of the Arctic's haline stratification. Both increased turbidity and enhanced stratification will reinforce the absorption of the Sun's energy into these coastal waters and put more heat in contact with any remaining ice to hasten melting and warm the region.

6.4. Ice formation in the Arctic

The development and seasonal sequence of sea-ice in the Arctic is very different to the Antarctic, because of its enclosed nature. Due to its constrained movement, ice that survives summer melt may continue to thicken from below year after year to form multi-year ice. The relative proportions of young and multi-year ice, and the characteristic double halocline, have a strong influence on the role of the Arctic in climate.

The surface halocline layer in the Arctic is maintained by melting sea-ice. When this ice reforms at about -1.8 °C (due to the depressed freezing point of saltwater), seawater fractionates producing brine that sinks rapidly downward, because of its density. This leaves both ice and fresher water behind at the surface, with additional freshening provided by the contributions from Arctic rivers. This effect helps to explain why the Arctic is so different from the Antarctic: a fresher ice-covered Arctic Ocean is insulated from saltier warmer water below by the density differences (analogy of oil and vinegar). Cold, dense water that is formed seasonally on the shelves is also contributed to the deep basins. These processes are important components of the MOC/THC. The multi-layered haline system is still a key element of the Arctic Ocean. As ice retreats and a more open water ocean starts to develop, strong mixing will remove the haline stratification leading to a step change in the whole Arctic system. Wind mixing will increase, biological production will be enhanced, a biological carbon pump will develop and the solubility pump may become less important among other changes. However, even in an ice-free Arctic a surface freshwater layer will be maintained in coastal zones due to the riverine input.

6.5. Observed changes in Arctic sea-ice cover

The most evident and rapid change that has taken place in the Arctic Ocean is the decline in summer Arctic sea-ice cover.

A marked decline has been measured, from both *in situ* and satellite observations, in summer Arctic sea-ice cover over the last three decades (Fig. 1.29). Since 1995 approximately, the decline has accelerated reaching the lowest recorded area ever in September 2007 (~4.13 million km^2; Fig. 1.30) (Stroeve *et al.*, 2007). In September 2007 sea-ice extent was nearly 50% lower than during the 1950s and 1960s and the 2008 sea-ice area was also significantly below the long-term average (Fig. 1.31), and similar to but not as low as 2007.

The thickness and volume changes were estimated to have been twice as fast as the changes in sea-ice extent (Maslowski *et al.*, 2000). There is now

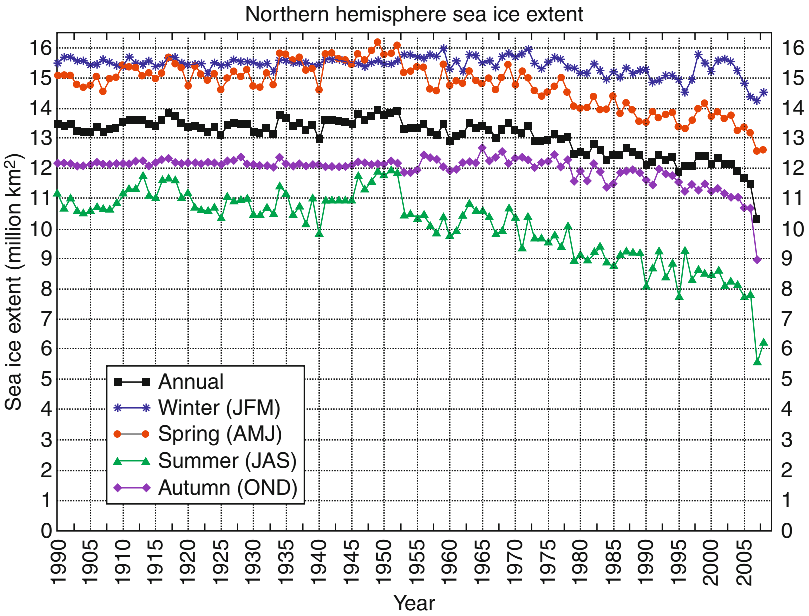

Figure 1.29 Seasonal and annual mean sea-ice extent averaged for the whole of the Northern Hemisphere, January 1900–September 2007. Source: The Cryosphere Today, University of Illinois, Polar Research Group, based on data from the U.S. National Center for Environmental Prediction/NOAA (http://arctic.atmos.uiuc.edu/cryosphere/).

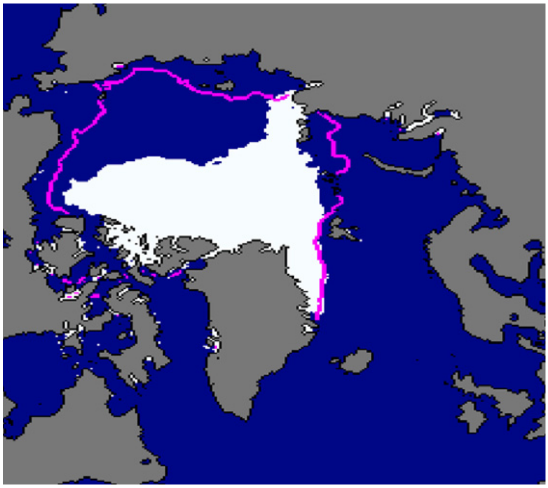

Figure 1.30 Sea-ice in September 2007. Total extent = 4.1 million km^2. The grey line shows the average position of the ice edge (median). Source: U.S. National Snow and Ice Data Center, Boulder, CO (http://nsidc.org/news/press/2007_seaiceminimum/images/20071001_extent.png).

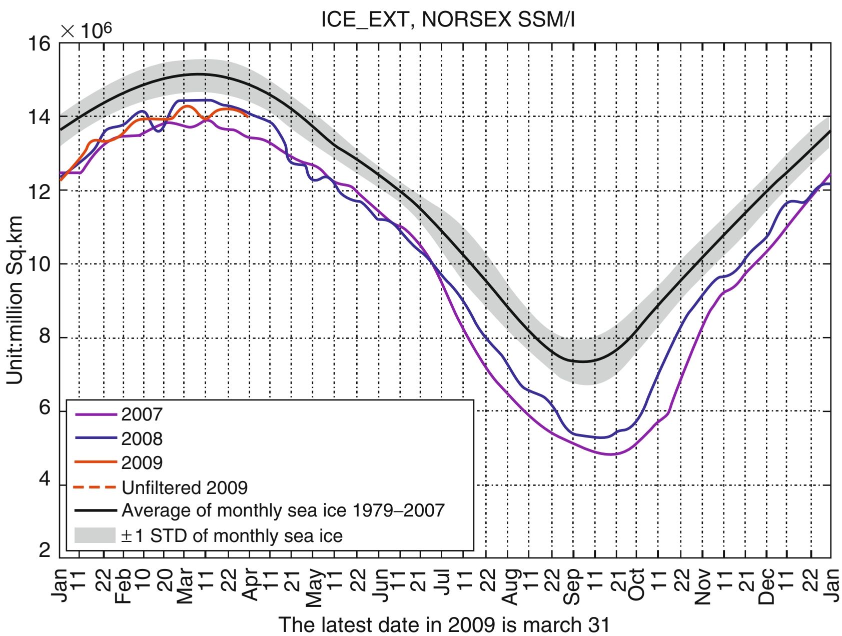

Figure 1.31 Extent of Arctic sea-ice (area of ocean with at least 15% sea-ice) in 2007, 2008 and part of 2009 with the long-term average. Source: Nansen Environmental and Remote Sensing Center, via the Arctic-ROOS web site (http://arctic-roos.org).

little of the thick, old ice left, which could make the region increasingly vulnerable to further ice loss (Rigor and Wallace, 2004). In the winter of 2008 measurements taken by Envisat showed that the thickness of winter sea-ice reduced by 0.26 m compared to the previous 6 years, averaged for the whole circumpolar region (Giles *et al.*, 2008). Thinning and less coverage leads to a reduction in the overall volume of sea-ice, determining its disappearance in the future. Inflow of warm salty water from the Atlantic has likely contributed to the overall decline Polyakov *et al.* (2008) as well as changes in atmospheric circulation (Maslanik *et al.*, 2007) and cloud cover (Francis and Hunter, 2006) plus a marked increase in export of old ice via the Fram Strait (Nghiem *et al.*, 2007). While the major focus of reduction since 1995 has been in the Eurasian Arctic, there has also been an important contribution to the melting from warm water originating from the Pacific and advection into the Chukchi Sea and adjoining deep basins (Shimada *et al.*, 2006). In 2008, the main melt occurred in the Beaufort, Laptev and Greenland Seas.

6.6. Trigger factors for initial sea-ice reductions

The North Atlantic Current, the west European shelf edge current and their extension in the Norwegian Sea (the Norwegian Current) have warmed markedly over the last two decades (Holliday *et al.*, 2008). This increased input of heat into the Arctic Ocean may have contributed to the trigger for the start of the decline in ice extent. However, there is still considerable debate on the relative role of oceanic versus atmospheric forcing of the changes. The Atlantic inflow is mainly related to a strong increase in the Arctic Oscillation (AO) and the North Atlantic Oscillation (NAO) at the end of the 1980s. However, even after 1996 when the NAO is average, temperatures, and possibly flow have increased, and there has been no return to the sea-ice state of pre-1988.

Warming in the North Pacific and Bering Strait in ~1995 led to the first major reduction in the extent and thickness of ice in the western basin. Contrary to the North Atlantic side of the Arctic, which is insulated by a deep halocline layer, the North Pacific surface water is close to the ice, affecting it directly in the winter. This warming has been reinforced since ~1998 by warmer temperatures in the West Greenland current. The coincidence of warmer conditions in the Canada Basin and in Baffin Bay led in September 1998 to a complete retreat of ice from the north of Alaska and Canada for the first time in recorded history.

6.7. Projected changes in Arctic sea-ice cover

Sea-ice loss is 30–50 years ahead of the modelling used in IPCC AR4 (Stroeve *et al.*, 2007). If the present rate of reduction in sea-ice continues, some models project an ice-free ocean in the Arctic summer by 2030

(Stroeve *et al.*, 2007) or 2040 (Holland *et al.*, 2006) compared to the more conservative estimates of a loss of greater than 40% in the area covered by sea-ice by 2050 as suggested by the majority of IPCC AR4 models (Overland and Wang, 2007). A more recent modelling analysis of trends in ice extent, thickness and volume (Maslowski *et al.*, 2007); Whelan *et al.* (2007) estimated that the Arctic may be ice free in the summer as early as 2013; however, more recent studies suggest a date of 2037 (Kerr, 2009). Such a rapid reduction will result in changes to many components of the Arctic environment as well as to adjacent seas. This will include a change to the ocean/atmosphere energy balance, affecting weather patterns, an increase in the freshwater budget from melting ice, supplemented by an increase in river runoff. Traditional patterns of salt and freshwater mixing will change with a likely reduction in the strength of the MOC/THC because of reduced deep water convection. A reduction in deep convection will in turn lead to lower fluxes of CO_2/DIC to the deep ocean. In the near-term, further sea-ice loss and increases in marine phytoplankton growth rates are expected to increase the uptake of CO_2 by Arctic surface waters (Bates *et al.*, 2006), although mitigated somewhat by warming in the Arctic (Bates and Mathis, 2009). Each of these changes has the potential to have a global effect on climate and climate change.

6.7.1. Sea-ice retreat and feedbacks

A positive feedback from the ice reduction already appears to be operating and leading to an acceleration of the retreat. Preconditioning of the sea was identified as a contributory factor to further sea-ice loss by Lindsay and Zhang (2005). Historically, the high reflectivity (albedo) of ice has reflected much of the sunlight during the long Arctic summers back into space, but once the ice starts to break up, it exposes large areas of dark open water where sunlight further heats the ocean. The scale of the effect from a change in albedo is very marked. Perovich (2005), for example, has calculated that a 500% increase in solar heat anomaly, due to the extensive area of open water in the summer of 2007, contributed to an increase in basal melting of ice in the Beaufort Sea and its accelerated retreat. The ponding of meltwater on the surface of sea-ice further leads to reduced reflectance and increased absorption of solar heat. The loss of sea-ice accelerates the warming of the dark ocean below, which distributes the heat to the surrounding water, deeper waters, sea bed and atmosphere. Recent modelling has demonstrated that during rapid sea-ice loss episodes, warmth is released back to the air and can penetrate up to 1500 km from the coast (Lawrence *et al.*, 2008). This can destabilise permafrost and lead to the release of methane, thus accelerating climate change. Methane released from shallow shelf seas has recently been reported as reaching the surface and off-gasing to the atmosphere (Westbrook *et al.*, 2009) although most methane released from sediment is converted to CO_2 by microbial anaerobic oxygenation of methane (AOM) before it reaches the surface.

6.8. The Greenland ice sheet

Changes in the mass balance (snow accumulation–melting) of the Greenland ice sheet will be strongly impacted by adjacent Arctic and Subarctic seas. The recent acceleration in ice reduction may in part be affected by the higher SST found in adjacent waters since ~1998 (Holland *et al.*, 2008). Changes in the circulation of the sub-polar gyre (Hátún *et al.*, 2005, 2009) are likely to have contributed to the higher sea temperatures. Warmer temperatures are increasing the number of summer days when portions of the surface of the Greenland ice sheet melt. Along the margins of the ice sheet, up to 20 additional days of melting occurred in 2005 compared to the average since 1988 (Fig. 1.32).

Because of the high elevation of its central mass, the ice sheet has a major impact on Northern Hemisphere atmospheric circulation and storm track location. Observed changes in the ice sheet as summarised in IPCC AR4 (2007) are:

- Inland thickening over the higher elevations
- Faster thinning around the coastal periphery
- Recent accelerated shrinkage of the total mass
- Northerly movement of the main ice zone from 66 to 70°N between 2000 and 2005

Model simulations indicate that the Greenland ice sheet will decrease in volume and area over the next few centuries, if a warmer climate continues. A threshold beyond which the ice sheet will continue to melt over many centuries (3000 years, Ridley *et al.*, 2005) is expected to be crossed if global annual mean temperature exceeds 3.1 ± 0.8 °C or the annual mean temperature for Greenland exceeds 4.5 ± 0.9 °C (Gregory and Huybrechts, 2006; Lowe *et al.*, 2006), or 3 °C (ACIA, 2005). Temperatures of this order are well within the IPCC A1B Scenario estimates for 2100 (IPCC, 2007), and unless global temperatures decline, the threshold for a complete melting of the Greenland ice sheet is likely to be crossed within this century. Once crossed, it is believed that the ice melt will be irreversible, resulting in sea-level rise of several metres over the coming centuries. This is in addition to any contribution from melting of the West Antarctica ice sheet. Lowe *et al.* (2006) suggested that complete or partial deglaciation of Greenland may be triggered for even quite modest CO_2 stabilisation targets.

6.9. Methane and feedbacks to climate change

The importance of methane hydrates (methane gas trapped in an ice-like solid) is becoming increasingly recognised. Methane is ~25 times more potent as a greenhouse gas than CO_2, thus the release of this gas is potentially a large feedback to climate change. While elsewhere on Earth, methane hydrates are maintained in place by the pressure of the overlying water,

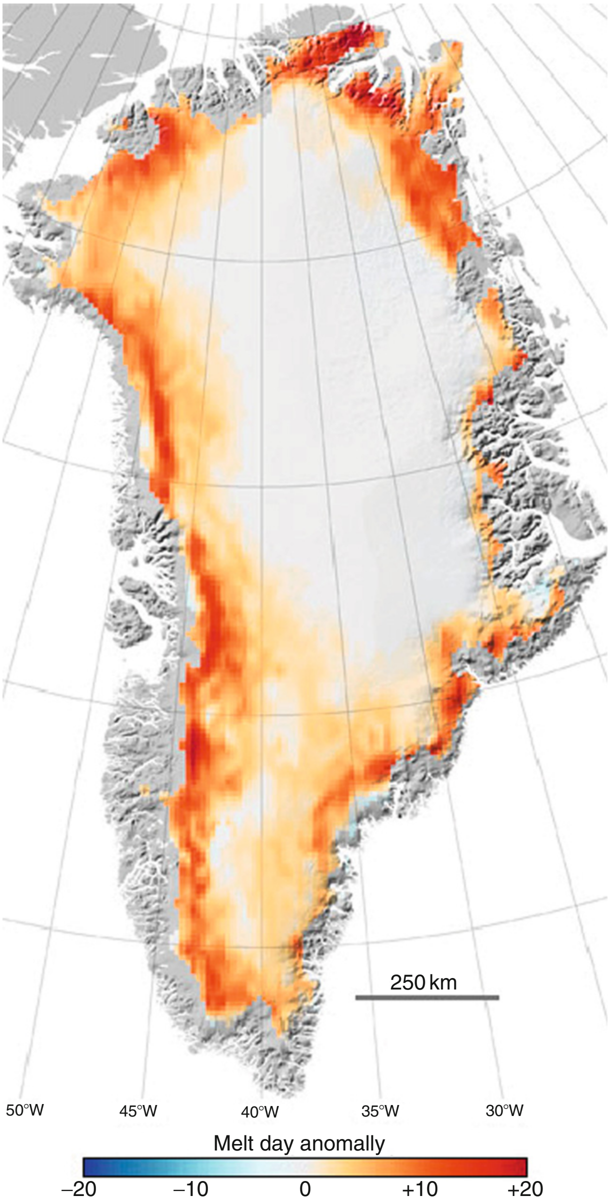

Figure 1.32 NASA map (by Robert Simmon and Marit Jentoft-Nilsen, based on data from Marco Tedesco, GSFC) indicating especially the increase in melt days in coastal regions (http://earthobservatory.nasa.gov/Features/Greenland/greenland3.php).

in the Arctic they are mainly stabilised by temperature (and occur at shallower depth). This fact makes them potentially vulnerable to climate change, especially in a warming Arctic Ocean.

6.9.1. Methane hydrates

In the Arctic, and on continental shelves and intercontinental rises elsewhere, sediments entrap major deposits of this greenhouse gas. There is considerable variability and uncertainty over the size of reserves of methane with estimates for methane stored in marine hydrates and sediment ranging from 10,000 Gt (GtC) (approximately twice all other carbon fossil fuels; Buffett and Archer, 2004) to 500 GtC and in permafrost from 450 to 7.5 GtC (Brook *et al.*, 2008). The impact of a sizeable release of these reserves into the atmosphere would be large. It is estimated, for example, that a 10% release of global methane stores to the atmosphere over a few years would be equivalent to a factor of 10 increase in atmospheric CO_2 (Archer, 2007). Fortunately, release of methane hydrates over the next century is thought to be significant, but not catastrophic (Archer, 2007).

Methane hydrates are sensitive to temperature and geostatic/hydrostatic pressure changes, but will be partially stabilised by the increased pressure from a rise in sea-level. There are two main mechanisms that affect methane hydrates: The first is sensitivity to sea-level rise: if a shelf region is flooded with warm water, a large thermal wave propagates into shallow sea bottom sediments and into the soils of flooded low-lying terrestrial regions. Many gigatonnes of methane hydrates are stored on the Arctic shelves of Alaska, Canada and Siberia, which could be reactivated by this type of warming. However, there are many uncertainties: including the level of temperature increase and timescales needed to melt permafrost. Advective inflow of warmer water from the Pacific or Atlantic would lead to a faster release of gas and could be a positive feedback to climate change. It is worth noting, however, that AOM will have converted most methane to CO_2 before it enters the atmosphere although this process will be slowed in the cold waters of the Arctic. The anaerobic micro-organisms responsible substantially reduce oceanic emissions of methane and are a key component of the carbon cycle (e.g. Pernthaler *et al.*, 2008). Secondly, methane outcropping at deeper depths can be affected by the temperature structure on the sea floor. If there is a feedback, it could come from a destabilisation of the hydrates in sediments; any loss could lead to a slope failure. If a turbidity flow was generated as a consequence, it might lead to a large pulse release of hydrates that would float to the surface of the ocean and be released to the atmosphere. Such a process has been suggested as one of the possible trigger mechanisms for the huge Støregga Slide on the shelf slope off Norway and its resulting tsunami *circa* 8000 BP (Beget and Addison, 2007).

For the Arctic, hydrate records show that they were released in previous glacial oscillations, but at least the present Canadian gas hydrate reserves have probably been recharged since the last ice age. In off-shore regions of the Beaufort Sea, methane can be found bubbling from hydrates which have been destabilising for thousands of years—this is not a recent change, but contributes to climate change. Post 2000, large volumes of methane have been observed bubbling from the sea bed of the Laptev and East Siberian Shelf Seas and were measured in elevated concentrations above the sea surface (Shakhova *et al.*, 2007, 2008) and attributed to warming conditions. Escape of methane has also been measured off Spitzbergen (Westbrook *et al.*, 2009). It is not clear if these new findings are a response to an anthropogenic signal, and it is also unclear if hydrates have been involved in previous changes of climate. For example, it is still unclear whether the PETM (55 million years ago) was brought on solely by a major methane hydrate release (Panchuk *et al.*, 2008).

6.9.2. Permafrost methane release

Warming of the Arctic, through direct warming and through heat released by the ocean, can degrade and melt permafrost and potentially lead to the release of any methane stored within it. The rate of increase of this gas into the atmosphere has slowed down over the last 20 years; a rate change that is not well understood. In 2007, however, the concentrations of methane in the air increased, particularly in the Arctic, suggesting a release from Arctic permafrost among other sources (the NOAA annual greenhouse gas index, Hoffman, http://www.esrl.noaa.gov/gmd/aggi/).

6.10. Arctic ocean ecosystems

As surface sea-ice continues to be lost, there are likely to be further large changes in the ecosystems and primary production of the Arctic (Carmack and Wassmann, 2006). Temperature, sea-ice cover and light penetration in the enlarged ice-free zones are expected to change, but not the light season, so phytoplankton primary production will increase within the same growing months. Reductions of sea-ice cover in the last decade, particularly in the western Arctic Ocean, have resulted in a longer marine phytoplankton growing season and an ~30–60% increase in the rate of primary production (Pabi *et al.*, 2008). Over the last several years, a similar ~10–40% increase in phytoplankton primary production has been observed in the Beaufort and Chukchi Seas (Arrigo *et al.*, 2008). In the Bering Sea, reduced sea-ice cover is thought to favour a 'phytoplankton–zooplankton' dominated ecosystem over the typical 'sea-ice algae–marine benthos' ecosystem (Piepenburg, 2005).

The life cycles of most Arctic species are highly adapted and intimately linked to the timing of sea-ice melt. At present, most phytoplankton primary production takes place as the sea-ice is melting and retreating towards the pole, with primary production rates typically lower in the older open waters of the central basin. It is difficult to estimate what the balance will be between planktonic production in a summer sea-ice-free ocean and sea-ice margin production. Indications are that a climate-induced reduction of sea-ice cover duration on Arctic shelves will favour the population growth of several key zooplankton species (e.g. Ringuette *et al.*, 2002), notably the predominant calanoid copepods, perhaps with a transition to a 'phytoplankton–zooplankton' dominated ecosystem rather than a 'sea-ice algae–marine benthos' ecosystem. In the Arctic's marginal seas, ecosystem changes could be profound if changes in benthic–pelagic coupling lead to increased pelagic production and a reduction of benthic production (e.g. Grebmeier *et al.*, 2006; Wassmann, 2006). Regardless of any changes in benthic–pelagic coupling, an enhanced seasonal penetration of the generally smaller subarctic species is expected, although the degree to which Arctic species may be displaced is uncertain. Such reorganisation in the way the ecosystem operates will ultimately alter the pathways and magnitude of energy that passes into upper trophic levels such as fish, sea birds and marine mammals, and impact the people dependent on those resources. Potential feedbacks from all these biological changes are unclear, as an integrated ocean-wide view on the structure and function of Arctic Ocean food webs is not yet available (Carmack and Wassmann, 2006).

The large reduction in sea-ice cover to the north of Alaska and Canada (including the archipelago) for the first time on record in 1998, is likely linked to evidence for an increased inflow to the Atlantic via this route seen in the first record of the Pacific diatom *Neodenticula* in the Northwest Atlantic in the following year (Reid *et al.*, 2007). Further incursions of Pacific water are likely to impact biological diversity and the carbon pump in the northern North Atlantic.

6.11. Modelling

The current range of Arctic ice and climate models are too conservative and do not adequately reproduce current changes taking place in the Arctic Ocean (Stroeve *et al.*, 2007). Some work suggests that a threshold (tipping point) has been passed in the loss of Arctic sea-ice (Lindsay and Zhang, 2005). There are inadequacies in model forcing, parameterisation of sea and ice processes and model structure. An Arctic model intercomparison project is underway jointly between the EU DAMOCLES and US SEARCH projects (Proshutinsky *et al.*, 2008) for a wide range of global and regional models that focus on different aspects of the Arctic environment. The Global Green Ocean model (Le Quéré *et al.*, 2005) indicates that there will be an increase in the Biological pump but this model does not consider acidification. The predicted ocean uptake of anthropogenic CO_2 using the IPCC (Intergovernmental Panel on Climate Change) scenarios (e.g. Solomons *et al.*, 2007) is expected to lower pH by 0.3–0.5 units over the next century and beyond (Caldeira and Wickett, 2003, 2005), with the

Arctic Ocean impacted before other regions due to the relatively low pH of polar waters (Bates *et al.*, 2009; Orr *et al.*, 2005; Steinacher *et al.*, 2009). Work within the International Polar Year is currently fostering improvements in many aspects of modelling within the region.

6.12. Concluding comments

There are many feedback loops within the Arctic, some of which have serious implications for global climate and climate change.

- Increasing heat flux of oceanic water to the Arctic as well as higher atmospheric temperatures is contributing to an accelerating retreat of Arctic sea-ice cover.
- The ice retreat removes the insulation between the ocean from the atmosphere enhancing ocean/atmosphere interaction and influences atmospheric circulation.
- The ice-albedo effect is large and provides a strong positive feedback from sea-ice loss.
- Increased warming of the wider Arctic oceanic region is likely to be contributing to Greenland ice sheet reduction.
- Methane is a potent greenhouse gas; warming of the Arctic Ocean and surrounding tundra may lead to its destabilisation and release from hydrates and permafrost with the capacity to accelerate global warming.
- Unless the trend in global temperature rise reduces, the temperature threshold for an eventual complete melting of the Greenland ice sheet may be crossed this century. Melting of the Greenland ice sheet alone could lead to 7 m of sea-level rise over the coming centuries.
- Model predictions for the disappearance of Arctic sea-ice during summers vary between 2013 and the end of the century.
- Most climate models underestimate the rate of ice loss over recent decades.
- Some work suggests that a threshold (tipping point) has been passed in the loss of Arctic sea-ice and recent low ice conditions may persist for some time.
- The impacts of climate change on Arctic biology and the carbon pump, and *vice versa* (any feedback to Climate Change) was not addressed by IPCC AR4. The scale and rate of any feedback to climate remain unclear.
- Deep water formation, the westward retraction of the sub-polar gyre (positive feedback to ice) and biological impacts were not adequately covered by IPCC AR4.

7. The Southern Ocean and Climate

This section focuses on the key role that the Southern Ocean (Fig. 1.33) plays in global climate, through its role in the MOC and interaction with sea-ice, Antarctic ecosystems and carbon uptake. Major

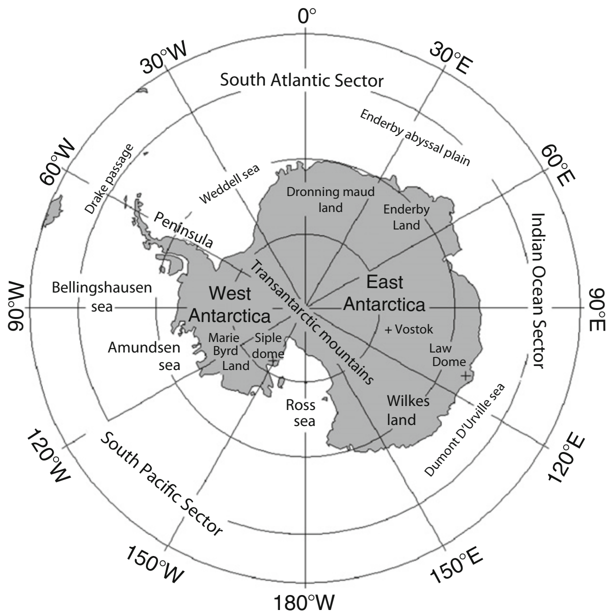

Figure 1.33 **A map showing the regional geography of the Southern Ocean and Antarctica that includes the names of locations referred to in the text.** ***Figure courtesy*****: Mike Meredith, British Antarctic Survey.**

changes that have taken place over recent decades in the forcing and response of the Southern Ocean are outlined, along with the impacts of these changes. As an example some of the rapid changes observed at the Antarctic Peninsula are described. Loss of ice shelves is addressed as well as evidence for net reductions of ice in western Antarctica. Finally, some recent modelling prognoses are presented alongside some of the technological and observing challenges that need to be addressed to monitor such a large and extreme environment. Changes in the cryosphere were comprehensively addressed in the IPCC AR4 reports with less coverage of the Southern Ocean. A detailed analysis of the state of the Antarctica and Southern Ocean climate system has recently been completed for the Scientific Committee on Antarctic Research (SCAR; Mayewski *et al.*, 2009).

7.1. Role of the Southern Ocean in climate

The Southern Ocean plays a critical role in driving, modifying and regulating global climate change. This is partly due to its unique configuration: it is the only ocean that circles the globe without being blocked by land. As a consequence, it is home to the largest of the world's ocean currents: the ACC, which is driven by the strong westerly winds and buoyancy fluxes over the Southern Ocean. This current transports 150 times more water around Antarctica than the flow of all the world's rivers combined.

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 (Fig. 1.34). The resulting global ocean circulation redistributes heat, salt, freshwater and other climatically and ecologically important properties. It has a global impact on patterns of temperature, rainfall and ecosystem functioning.

The Southern Ocean is a key region in the oceanic MOC/THC, 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 (Fig. 1.34). This upwelling brings carbon and nutrient-rich waters to the surface, acting as a source of CO_2 for the atmosphere and promoting biological production.

The lower limb of the MOC comprises the cold, dense 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 (Fig. 1.34), travelling

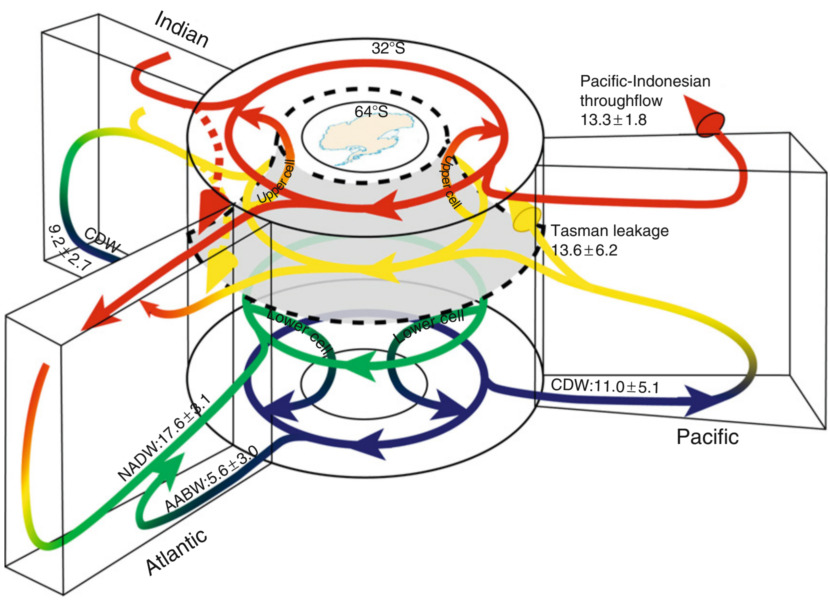

Figure 1.34 Schematic of the global ocean circulation, emphasising the central role played by the Southern Ocean. Figure from Lumpkin and Speer (2003). See Fig. 1.35 for an explanation of the acronyms.

as far as the North Atlantic and North Pacific. This cold water also absorbs atmospheric gases, including oxygen and carbon dioxide, which enables it both to aerate the bed of the global ocean and to act as a temporary (hundreds of years) sink for natural and anthropogenically produced CO_2.

The upper limb of the MOC is sourced towards the northern flank of the ACC. Here, the water that is upwelled within the ACC is converted into mode waters and nutrient-rich intermediate waters that permeate much of the global ocean basin south of the equator. Mode waters (like the Subantarctic Mode Water, SAMW, Fig. 1.35) form at the surface in winter via convective processes, and are relatively homogeneous water masses of uniform density. They are undercut by intermediate waters (AAIW in Fig. 1.35) that are renewed by subduction near the Polar Front. The formation and subduction of the mode and intermediate waters is believed to be a critical process that removes anthropogenically produced CO_2 from the atmosphere (Fig. 1.36).

Closer to the continent, ocean processes are strongly controlled by sea-ice, the formation of which is the largest single seasonal phenomenon on Earth. The freezing of the sea around the continent as sea-ice each year effectively

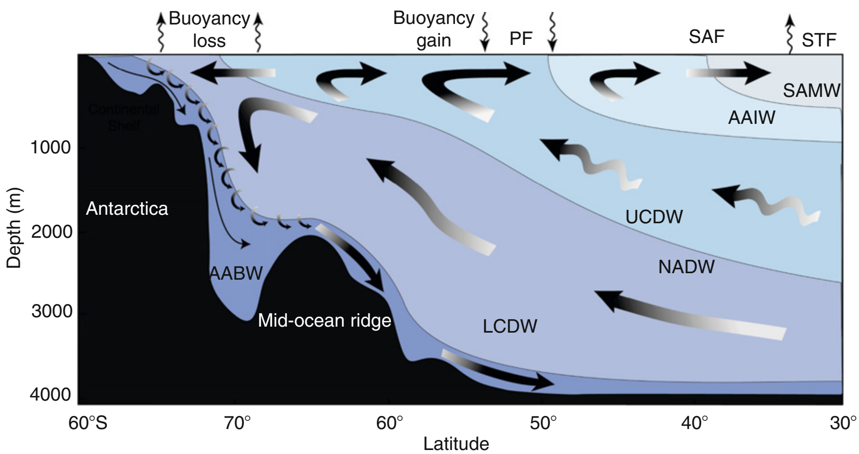

Figure 1.35 Schematic of the overturning circulation in the Southern Ocean, whereby southward-flowing products of deep convection from the North Atlantic are converted into upper-layer (mode and intermediate) waters and deeper (bottom) waters and returned northward. Marked are the positions of the main fronts (PF, Polar Front; SAF, Subantarctic Front; STF, Subtropical Front) and water masses (AABW, Antarctic Bottom Water; LCDW and UCDW, Lower and Upper Circumpolar Deep Waters; NADW, North Atlantic Deep Water; AAIW, Antarctic Intermediate Water; SAMW, Subantarctic Mode Water). Note that as well as the north–south water movement shown in the figure water is also generally moving towards the observer (i.e. west to east), except along the coast where coastal currents are moving water away from the observer (east to west). Figure from Speer *et al.* (2000).

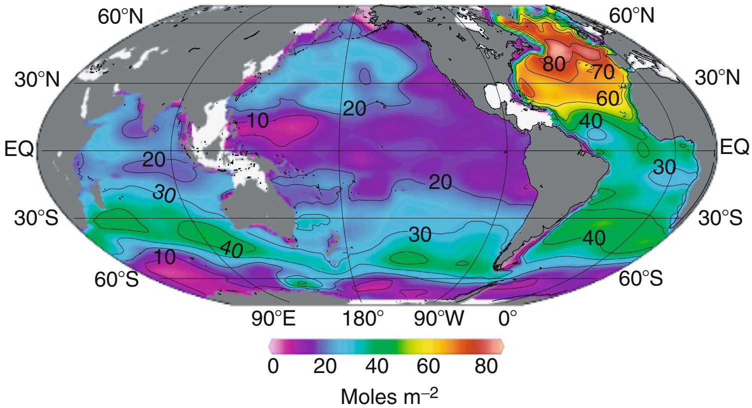

Figure 1.36 Water column inventories of anthropogenic CO_2 in the ocean. Note in particular the band of high levels flanking the northern side of the ACC, associated with mode and intermediate waters. Dissolved CO_2 is lost to the atmosphere south of the Polar Front, where NADW wells up to the surface close to the coast (purplish colours) and gained from the atmosphere north of the Polar Front where mode waters and intermediate waters sink in the subduction process (green colours), making the Southern Ocean both a source and a sink for atmospheric CO_2. From Sabine *et al.* (2004a).

doubles the size of Antarctica and has a profound effect on climate. Because of its high albedo (whiteness), it reflects the Sun's heat back into space, cooling the planet. However, the sea-ice also acts as a 'patchy blanket', limiting heat loss from the ocean to the atmosphere and restricting air–sea exchange of climatically important gases. The formation of sea-ice, as noted above, plays a key role in the production of AABW and its annual melt supplies a thin layer of freshwater to the surface ocean that stabilises the stratification and can promote phytoplankton blooms. 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 highly biologically productive, although it is not as productive as it could be. This is because the productivity is limited by the low availability of micro-nutrients such as iron, except in a few areas such as near the isolated islands that are scattered within the ACC. Nevertheless, the Southern Ocean is a key region for the Biological pump with diatoms as ballast playing an important role in the sedimentation of organic material to the deep ocean (see Section 4).

7.2. Observed changes in the Southern Ocean region

The Southern Ocean has shown many marked changes in recent decades, highlighting its sensitivity to global processes and illustrating different aspects of its control on regional and global change. The most conspicuous of these changes is probably the profound warming within the ACC (Gille, 2002, 2008; Fig. 1.37). In this area, a strong, surface-intensified temperature increase has been noted that exceeds that of the global ocean as a whole. The exact causes of this warming are not yet understood, though most theories seek to relate it to the intensification and southward shift of the band of westerly winds that overlie the circumpolar Southern Ocean. Potential candidate mechanisms include a latitudinal shift in the ACC, greater air–sea heat fluxes, an intensification of the circumpolar eddy field, and possibly other processes also, almost certainly in some combination (Fyfe, 2006; Fyfe and Saenko, 2006; Gille, 2002, 2008; Hogg *et al.*, 2008; Meredith and Hogg, 2006). The large-scale circulation patterns of the Southern Hemisphere atmosphere over the past few decades reveal changes that are reflected in the leading mode of Southern Hemisphere climate variability, the SAM (Thompson and Wallace, 2000). Interannual variability and trends in the SAM have been shown to drive substantial variability in ocean circulation, upper-ocean biology, and the uptake and release of CO_2 to and from the Southern Ocean (Lovenduski and Gruber, 2005; Lovenduski *et al.*, 2007, 2008). The intensification and movement of the wind field as expressed in SAM changes is known to be at least partially due to anthropogenic processes (ozone depletion and greenhouse gas emissions; Marshall, 2003; Thompson and Solomon, 2002). This suggests that the activities of mankind are perturbing the ocean around Antarctica on a large scale. Noteworthy also is the observation that the current generation of climate

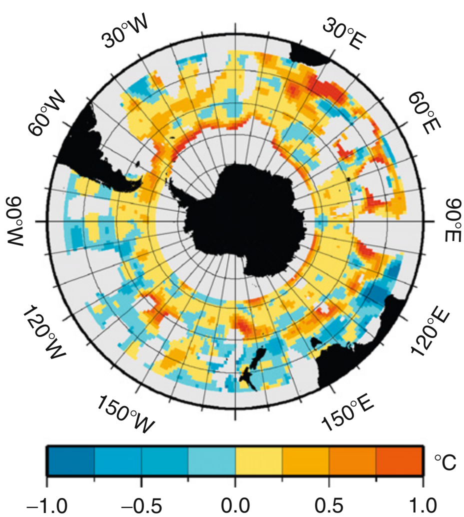

Figure 1.37 Temperature differences in the Southern Ocean between the 1990s and earlier decades, at approximately 700–1100 m depth. Note in particular the marked warming around the circumpolar band. Figure from Gille (2002).

models can produce a warming in the Southern Ocean comparable to that observed only if anthropogenic gases and sulphate and volcanic aerosols are included (Fyfe, 2006). If the role of volcanic aerosols is neglected, the simulated warming is nearly double, implying that the potential human impact on Southern Ocean warming is only partially realised at present in our sequence of observations.

There are countless likely feedback mechanisms from this circumpolar warming on regional and global climate, including impacts on sea-ice formation (and hence albedo; see below), solubility of carbon dioxide and other climatically important gases, and modulations to primary production and associated 'biological pumping' of carbon. Each of these are the subject of ongoing investigation; however, it should be noted here that the circumpolar winds are predicted to increase further over the next few decades, hence (if theories relating the warming to strengthening circumpolar winds are indeed correct) the effects are likely to be persistent rather than transient.

The strengthening circumpolar westerly winds have been highlighted as the potential root cause of another important observed change in the Southern Ocean region, namely a saturation of the Southern Ocean CO_2 sink. Ocean inverse analyses (Gloor *et al.*, 2003; Mikaloff Fletcher *et al.*, 2007) indicate that the pre-industrial Southern Ocean (south of 44°S) was a source of natural CO_2 to the atmosphere. But the rise in atmospheric CO_2 from pre-industrial levels of about 280–380 ppm at present, has led to a strong perturbation of the air–sea CO_2 balance, that is, it induced a flux of anthropogenic CO_2 that is directed into the ocean. This sink of anthropogenic CO_2 in the Southern Ocean takes up nearly 10% of the CO_2 emissions to the atmosphere. However, based on atmosphere and ocean measurements, and the analysis of model output, Le Quéré *et al.* (2007) argued that this sink has not increased since 1981, in spite of a >40% increase in CO_2 emissions. A reduction in CO_2 uptake would likely lead to an increase in the amount of CO_2 in the atmosphere, with clear implications for climate change. The theory proposed to explain this observation was that upwelling in the Southern Ocean, which brings natural carbon from the deep ocean to the surface layers, has been accelerated by the strengthening winds. This has increased surface concentrations of CO_2, and precluded further absorption of anthropogenic CO_2. Also noteworthy is that the increased upwelling of CO_2 will increase the rate of ocean acidification, with consequences for the ecosystem. Although Le Quéré's conclusions have been supported by another modelling study (Lovenduski *et al.*, 2008), it should be noted that Böning *et al.* (2008) have questioned this saturation of the Southern Ocean CO_2 sink, arguing that the effect of increased eddy formation could compensate for the extra energy imparted to the ocean by the winds, with no significant change in the overturning. Much remains to be done on this important subject, including improved monitoring of CO_2 and better understanding of the physical processes that mediate the vertical motion in the ocean, and it is imperative that the research community tackles these subjects as a matter of priority.

On a regional level, the area in the Southern Hemisphere whose atmospheric climate has been changing most rapidly is that of the Antarctic Peninsula (e.g. Turner *et al.*, 2005). On its western side, a wintertime atmospheric warming of 5 °C since 1950 has been observed with a smaller (but still significant) warming seen in summer. This warming has been noted to be strongly linked with the reduction of sea-ice extent and duration in the adjacent Bellingshausen Sea since the 1950s, and has also been shown to be connected to a very strong summer warming of the upper ocean (Meredith and King, 2005; Fig. 1.38). These authors used a large compilation of *in situ* hydrographic profiles collected between the 1950s and 1990s to demonstrate a surface-intensified warming (of >1 °C in the shallowest levels), and a coincident strong summer salinification. It is worth emphasising that these oceanographic changes constitute positive feedbacks that act to sustain and enhance the atmospheric warming and further reductions in sea-ice formation—a clear example of ocean processes exerting a strong influence on regional climate, in an area of very rapid

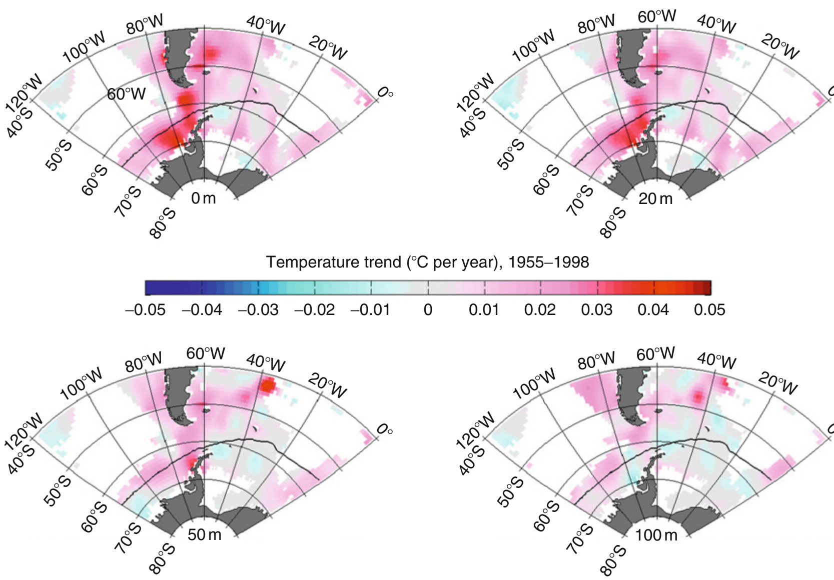

Figure 1.38 Trends in temperature during the second half of the twentieth century in the vicinity of the Antarctic Peninsula. Four different depth levels are shown, namely the surface, 20, 50 and 100 m. Note the strong, surface-intensified warming at the western side of the Peninsula. From Meredith and King (2005).

change. Meredith and King (2005) also noted that the rising temperatures have played an important role in the accelerating retreat of the tidewater glaciers on the Peninsula that have led to an increasing contribution of glacial meltwater to the adjacent ocean. In accord with this, Cook *et al.* (2005) showed that the majority of glaciers on the western side of the Peninsula are retreating, and that retreat rates are accelerating.

Meredith and King (2005) also noted the profound consequences for the ocean ecosystem in this sector, where benthic organisms are generally well adapted to cope with low temperatures, but poorly adapted to cope with changes in temperature. Indeed, based on a study of temperature tolerances, Peck *et al.* (2004) asserted likely 'population and species level losses' of marine organisms at the western Peninsula associated with a proposed 2 °C change in ocean temperature. The observed change of >1 °C in 50 years illustrates that such changes are very possible within the next few decades. It is also worth noting that a key species in Southern Ocean food webs, namely Antarctic krill, has been undergoing a dramatic decline in numbers in the South Atlantic in recent decades (Atkinson *et al.*, 2004). This population is sourced at least partially from breeding and nursery grounds near the western Peninsula, and it was argued that the loss of sea-ice and warming of the ocean may be the cause of their decline (Atkinson *et al.*, 2004; Meredith and King, 2005). Krill are a critical component of the Southern Ocean marine food web (Hill *et al.*, 2006; Knox, 2007) with most higher trophic levels depending on them and some such as the baleen whales feeding exclusively on these crustaceans. Krill are also now targeted by commercial fishing and so are particularly vulnerable. Establishing the long-term impacts of their removal from regional ecosystems and on climate is a high priority research area.

To the south of the retreating glaciers on the Peninsula referred to above and extending as far as the Trans-Antarctic Mountains is the West Antarctic ice sheet. Much of this ice sheet rests on rock that is well below sea-level and parts of its margin are in direct contact with the ocean. Using satellite radar interferometry and regional climate modelling, Rignot *et al.* (2008) have estimated a widespread net loss of the ice sheet in western Antarctica adjacent to the Bellingshausen and Amundsen seas; with the rate of loss increasing by 59% in the decade to 2007. The current reduction in mass from the Amundsen Sea embayment of the West Antarctic ice sheet is equivalent to that from the entire Greenland ice sheet (Lemke *et al.*, 2007). The process believed responsible for this rapid and large change is a progressive thinning of the fringing ice shelves seaward of the Amundsen Sea outlet glaciers. The likely cause of this melting is a greater penetration onto the shelf, typically at a few hundred metres depth and upwelling of 'warm' Circumpolar Deep Water from the ACC. Why this warmer water is now reaching the ice shelves more readily is still not fully understood, but it is believed to be caused by the increase in westerly winds. At a few key locations these warmer waters flow towards the outlet glaciers along deep glacially scoured submarine troughs carved by ice in past glacial periods. Once the 'warmer' water reaches the base of the floating ice shelves, it causes extremely high melting rates of many tens of metres per year due to the large temperature contrast between the seawater and ice. 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 towards the interior of West Antarctica (e.g. Holt *et al.*, 2006), so a contribution from this sector alone of some tens of centimetres to sea-level rise by the end of the twenty-first century cannot be discounted.

On the eastern side of the Antarctic Peninsula the greatest warming is during the summer months, and appears to be directly related to the strengthening of the circumpolar westerly winds (Marshall *et al.*, 2006). These have resulted in more relatively warm, maritime air masses crossing the Peninsula from the west and reaching the low-lying ice shelves in the east. The impacts of this change have included the break-up of large parts of the Larsen ice shelf, which has progressively disintegrated from north to south. The remaining part of this shelf ('Larsen-C') is being closely watched, and may well disintegrate within the next few years or decades. Again, with the root cause of the change being the strengthening winds, there is a direct connection between anthropogenic processes (greenhouse gas emission and ozone depletion) and a large response in the Antarctic.

In IPCC AR4 it was predicted that the Antarctic ice sheet as a whole will increase in mass over the next century due to higher snowfall as a consequence of a warmer climate (see also Krinner *et al.*, 2007). Build-up of snowfall in the interior of Antarctica is balanced by wastage due to melting and calving of ice along the coast and this balance is an important component of sea-level rise. Measurements made by satellite altimetry confirm a growth in ice mass from snowfall in East Antarctica over an 11-year period since 1992 (Davis *et al.*, 2005) and another study has shown a doubling in snowfall in the western Antarctic Peninsula since 1850 (Thomas *et al.*, 2008). In contrast, the long records of Monaghan *et al.* (2006) and van den Broeke *et al.* (2006) for the whole of Antarctica reveal large decadal to multi-decadal variability in snowfall and yet no significant trend. The absence of an observed long-term change in snowfall, when averaged for the whole continent, has taken place against a background where significant warming is not just confined to the Antarctic Peninsula, but extends over much of western Antarctica and positive trends in temperature are recorded for the whole of Antarctica since the 1950s (Steig *et al.*, 2009). There is therefore no clear evidence that snowfall has changed in Antarctica as a whole in response to rising temperature, contrary to the predictions of climate models. The only exception is in the Peninsula, but there the increase in mass has been small compared to the loss of mass from the glaciers, so that precipitation plays only a minor role in mitigating the contribution to sea-level change from the Peninsula.

In terms of ice wastage, a rapid increase in glacier flow and retreat of ice sheets has occurred in both the Antarctic Peninsula and West Antarctica; the speed of the observed changes has demonstrated the important role that ice shelves play in controlling the mass balance of ice sheets (Rignot, 2006). In East Antarctica, glaciers grounded well below sea-level are also thinning (Rignot, 2006). The main driver for the glacier changes in the Antarctic is the ocean. There is clear evidence that the Southern Ocean has warmed (Gille, 2008) and it is likely that the ocean waters along some coastal sectors of the Antarctic are warmer than in the past, for example, the western Peninsula and Amundsen Sea, but unfortunately there are not enough measurements to show the timing of these changes. These warmer waters condition the evolution of ice shelves much more than air temperature; subsurface melting of ice by the ocean is orders of magnitude larger than what is happening on the surface. Rignot (2006) concludes that the mass balance of ice in a warmer climate will be more affected by the evolution of its ice streams and glaciers than on changes in the precipitation of snow in the interior.

7.3. The future

The future climate evolution of Antarctica and the Southern Ocean is especially hard to predict, since many of the coupled climate models that are traditionally used for such predictions do not represent well some of the key processes, and there is also a dearth of data with which to validate and challenge the model-based results. Notwithstanding this, it is possible to generate some understanding of likely future change using such models. For example, Bracegirdle *et al.* (2008) considered the output of the 20 coupled climate models used in the IPCC Fourth Assessment Report, and produced a 'scaled average' of their predictions, with the scaling for each individual model being in accord with the skill shown by that model at reproducing previous (observed) climate change. A 'middle-of-the-road' scenario for fossil fuel emissions was adopted for this.

Using this approach, the models predicted a continuing warming of the circumpolar Southern Ocean in the next 100 years, but with markedly stronger warming in the sub-polar gyres (Weddell Sea and Ross Sea; Fig. 1.39). This stronger regional warming is associated with a 25% decrease in sea-ice extent. Given the locales, this will almost certainly impact on AABW production in these two key formation sites, with possible consequences for the lower limb of the ocean overturning circulation. The circumpolar westerly winds are also predicted to continue strengthening, with likely consequences in line with the discussions above.

In practice, the predicted ubiquitous warming over Antarctica and the Southern Ocean is in line with changes predicted and beginning to be observed in the Arctic. The equilibrium response of the two polar regions to planetary-scale climate change is comparable; the differences observed so far

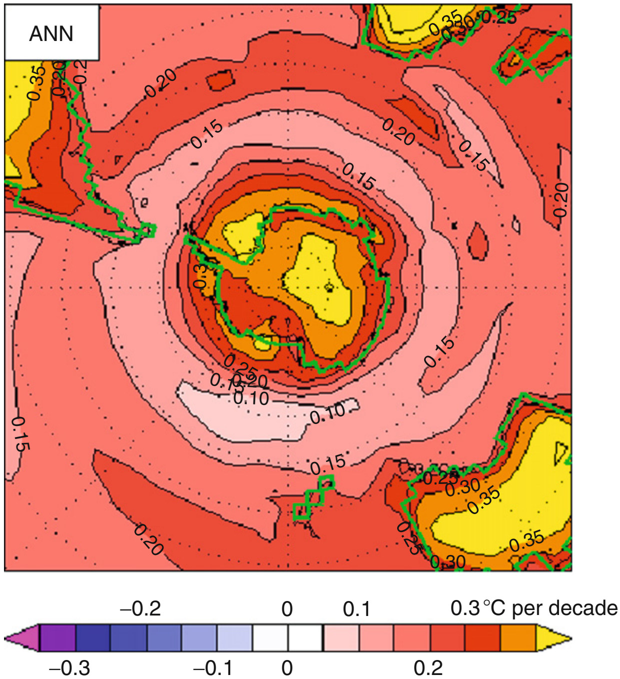

Figure 1.39 Predicted trends in surface temperatures over the next 100 years from a weighted average of the 20 coupled models used in IPCC AR4. Note the ubiquitous Southern Ocean surface warming, with ocean 'hotspots' in the Weddell and Ross Seas. From Bracegirdle *et al.* (2008).

are concerned with the transient response. Accordingly, the rapid changes observed in the Arctic and at the Antarctic Peninsula can be viewed as potential harbingers of what is to come around Antarctica and the Southern Ocean as a whole.

The future evolution of the Southern Ocean CO_2 sink is hard to predict. It depends on how the ocean circulation will change as forcing by the atmosphere evolves, in particular its overturning component. However, knowledge of the evolution of Southern Ocean overturning alone is not sufficient to predict the future of the oceanic CO_2 sink because under very high CO_2 an increase in upwelling could favour more uptake of anthropogenic CO_2 and over-compensate the enhanced ventilation of natural carbon coming from the deep ocean. Furthermore, changes in circulation, temperature and acidification will certainly impact the downward transport of organic carbon by biological activity, but we do not know the direction, amplitude or the rate of the potential changes. However, based on the behaviour of the Southern Ocean carbon cycle during glaciations, it appears that its response to a warmer climate would be to permanently outgas some of its natural CO_2 to the atmosphere while likely reducing the uptake of anthropogenic CO_2, which will add to the challenge of stabilising atmospheric CO_2.

Regarding the future of the Antarctic ice sheet, changes in the Southern Ocean, especially along the coast of Antarctica, are a major control on the evolution of Antarctic glaciers and on the mass balance of the ice sheet as a whole. This is something that glaciologists and modellers have known about for some time, but it has not been included as a factor in the predictions of the future state of the Antarctic ice sheet. The oceans matter even more than anticipated, but it is a domain of study that is currently fundamentally limited by a lack of basic observations, such as the shape of sub-ice-shelf cavities and oceanic conditions (temperature, salinity, currents, etc.) along the coast of Antarctica, near the glacier grounding lines. Only limited progress will be possible in the prediction of the evolution of Antarctica over the next 100 years until an understanding has been developed of how the Southern Ocean is changing now and into the future. This important new theme is emerging strongly from all recent ice studies and needs to be addressed by the scientific community.

7.4. Concluding comments

- The Southern Ocean, including the ACC, plays a critical role in driving, modifying and regulating global climate and climate change.
- An increase in westerly wind speeds due, at least partly, to human influences (increases in greenhouse gases and ozone depletion) has been observed.
- A continuing warming of the Southern Ocean and strengthening of the westerly winds is predicted.
- Rapid warming of the ocean west of the Antarctic Peninsula since the 1950s has been measured and associated atmospheric and cryospheric changes observed.
- Changes in the Southern Ocean are closely connected to the production and melting of sea-ice, the formation of which is the largest seasonal phenomenon on Earth. Sea-ice has a major effect on the Earth's energy budget and thus climate.
- A strong decrease in sea-ice extent over the next 100 years is predicted.
- The evolution of the Southern Ocean along the coast of Antarctica is a major control on the stability of Antarctic glaciers and on the mass balance of the ice sheet as a whole.
- Deflation of the northern sector of the West Antarctic ice sheet induced by a thinning of the fringing ice shelves is most likely associated with greater subsurface penetration of Circumpolar Deep Water onto the continental shelf.
- Reduction of glaciers in the Peninsula and in West Antarctica is predicted to accelerate.

- The Southern Ocean is an important sink for both natural and anthropogenic carbon dioxide, a sink that has been reported as possibly saturating, and which urgently requires further investigation.
- Further modification to the CO_2 sink is predicted.
- Marked changes have occurred in Southern Ocean ecosystems including a substantial decline in krill numbers.
- Acidification is predicted to increase with likely important modifications to unique Antarctic ecosystems.

8. Climate Models

This section provides a brief review of the 'state of the art' in modelling the feedbacks of the ocean on climate change. It notes existing limitations and offers some suggestions for important research priorities in model development and associated observations.

8.1. Ocean–climate feedbacks

Figure 1.40 summarises the key ocean feedbacks that contribute to climate. This section attempts to identify and summarise the modelling issues and limitations that exist for these feedbacks and in particular how well the models currently used for climate projections simulate the carbon cycle and sea-ice. The following key questions need to be considered against each of the feedbacks: What are the consequences for future climate prognoses if these feedbacks are not adequately understood and how can they be prioritised?

8.2. Heat uptake

The oceans are the main heat reservoir for the world, particularly over longer time periods, and strongly influence the rate of climate change as they have a large capacity to absorb heat compared to land. As a consequence, the oceans warm up slowly down to depths of kilometres and act as a delay on anthropogenically forced global temperature rise. A corollary is that ocean warming will continue for a long time into the future, even if greenhouse gas concentrations are stabilised in the atmosphere. While damping the rate of surface climate change, this warming of the ocean also leads to sea-level rise through the thermal expansion of seawater.

State-of-the-art coupled atmosphere–ocean global circulation models (AOGCMs) include the primary physics that controls ocean heat uptake, but there are still substantial differences in the results obtained by different models. For example, the efficiency of ocean heat uptake varies by a factor of over 5 among the AOGCMs used in the IPCC AR4, although the majority of the models are more tightly clustered (Randall *et al.*, 2007). Some models are able to reproduce the broad picture of increasing

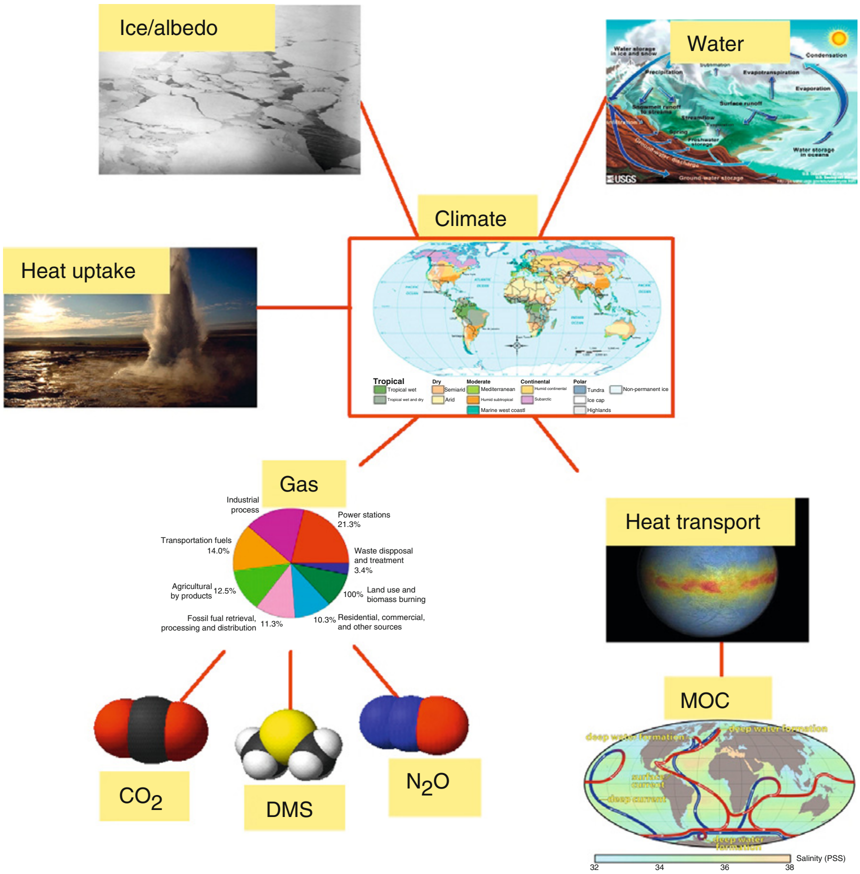

Figure 1.40 Factors involved in the interaction of the ocean with climate.

heat content that is deduced from historical observations over the past 50 years (e.g. Barnett *et al.*, 2001). Models generally do not reproduce the apparent maximum in global heat content in the 1970s, apparently calling into question whether the models have sufficient amplitude of internal climate variability; however, recent analysis by Domingues *et al.* (2008) suggests that the heat content maximum in the observations may be partly an artefact of instrumental errors, thus reducing the discrepancy between models and observations. Uncertainty in the observed heat content estimates also arises from the limited sampling, especially in the era before the Argo buoy network (Gregory *et al.*, 2004). Assessment of modelled and observed heat content changes remains an active research area.

Modelling of changes in the heat content of the North Atlantic by Banks and Gregory (2006) has shown that regional distributions of heat uptake are crucially dependent on the changes in the large-scale circulation and mixing of the oceans and not just along lines of equal density as might be indicated by tracers. In another study, Lozier *et al.* (2008) showed large regional differences in the heat uptake of the North Atlantic, which has increased on average at a rate equivalent to a surface heat flux of 0.4 ± 0.05 W m^{-2} over the last 50 years. This basin-wide increase disguises a large contrast between the sub-polar gyre which experienced a net loss of heat between the periods 1950–1970 and 1980–2000 against a large heat gain in the tropical and subtropical North Atlantic. The changes were attributed at least in part to recent decadal variations of winds and heat flux linked with the NAO. The present generation of climate models, in general, do not model recent NAO changes well, so that it can be concluded that there is still considerable uncertainty in modelling heat uptake at this level of detail.

Recent analyses by Sriver and Huber (2007) have demonstrated that 'tropical cyclones are responsible for significant cooling and vertical mixing of the surface ocean in tropical regions'. They calculated that ~15% of the transport of heat by the ocean may be associated with this downward mixing of heat. Furthermore, the strength of mixing is correlated with SST so that future increases in tropical temperatures may have important consequences for ocean heat transport and circulation. Since tropical cyclones are poorly resolved in models that are in use at present, their effects must be represented in gross form (parameterised). The size of errors in mixing projections associated with possible future changes in tropical cyclones has not been assessed.

8.2.1. Main limiters to heat uptake modelling progress

- The wide range in the present generation of model estimates of heat uptake efficiency.
- A greater understanding of the reasons for inter-model differences at the process level.
- An improvement in temporal and spatial coverage of observational data needed to evaluate models.
- Good estimates of historical water mass changes, including complete error estimates.
- Shortness of time series (and limits to modelling of the main modes of climate variability) makes the distinction of Climate Change from natural variability difficult.
- Better sampling (from Argo) will over time improve global observational coverage.
- More sophisticated data assimilation (reanalysis) methods are needed to extract maximum information from limited historical data.
- Improved estimates of atmospheric aerosol forcing would have a knock-on impact by further constraining the calculation of the efficiency of ocean heat uptake.

8.3. Heat transport

The poleward transport of heat from the tropics, by extensions of the Gulf Stream and the North Atlantic Current and the southerly directed deep counter currents as part of the MOC, has major implications for climate. Around 10^{15} W of heat is moved northwards in the North Atlantic and is dissipated to the atmosphere northwards of about 24°N to represent a substantial heat source for Northern Hemisphere climate. It has been postulated that a slowing down or cessation of the MOC could lead to a sudden and marked reduction in heat transport to the region, leading to a cooling of Europe's climate. However, total shutdown of the MOC is generally considered to be a high impact, low probability event, especially in the twenty-first century. The present generation of climate models suggests a slowdown ranging between 0% and 50% during the twenty-first century, under the IPCC A1B scenario, but none of the models suggests a shutdown (Meehl *et al.*, 2007). In the models where the MOC weakens, warming will continue in Europe as any reduction in the MOC will be counterbalanced by warming due to increasing greenhouse gases. At present it is not possible to give precise quantitative advice, especially on longer timescales, due to a large range of uncertainty in the modelling.

Density contrasts caused by spatially differing air–sea heat exchange are one of the three driving forces behind the MOC; others are density contrasts due to spatially differing freshwater exchange (haline forcing) (Saenko *et al.*, 2002) and surface flux of momentum (wind stress forcing) (Beena and von Storch, 2009; Chelton *et al.*, 2001; Delworth and Greatbatch, 2000). Using models to help determine the relative importance of these three driving forces is an important research area for Climate Change. Overall, models suggest that the response of the MOC to Climate Change is initially driven by changes in thermal forcing, with fresh water/salinity effects taking on an increasing role at longer timescales. Freshwater supply from melting of the Greenland ice sheet is not properly modelled in the current generation of climate models. Evidence from recent studies with improved ice sheet models (Fichefet *et al.*, 2003; Ridley *et al.*, 2005) is mixed as to whether this extra water source would have a significant impact on the MOC, but still no model suggests an MOC shutdown during the twenty-first century. Since important branches of the MOC pass through narrow straits that are not fully resolved in present climate models, the sensitivity of model projections to model resolution is an important open question.

Many changes have been observed in the North Atlantic recently, for example, in salinity and in some elements of the MOC flow. The MOC at the latitude (25°N) of maximum heat transport has been estimated only a few times from direct observations (Bryden *et al.*, 2005) making it difficult to know if there have been any long-term trends. However, a programme of continuous monitoring has recently begun (Cunningham *et al.*, 2007), which should considerably improve our knowledge of the MOC and its variability. Overall there is not yet a clear picture of how the North Atlantic has been changing in recent decades so that it is not yet possible to separate out the effects of climate change and natural climate variation. This remains an active and important research area.

Because of the circumpolar surface circulation and the presence of major sites of deep water formation, the Southern Ocean is also a key contributor to the overturning of the world's ocean. Its large surface area and potential for strong mixing make it an important area for heat and carbon uptake. The existing generation of climate models tend to have some biases in their simulations of the Southern Ocean circulation, believed to be partly a consequence of errors in the simulated winds (Randall *et al.*, 2007). Eddies and boundary current processes may also play an important role in the circulation and tracer transports, and it is an open question as to whether their effects can be adequately parameterised in coarse resolution models (e.g. Banks *et al.*, 2007). Lack of available observations to test the models in this remote part of the world remains an important constraint, although the Argo float programme is now helping to fill data gaps. The tropics are another area where improved understanding of modes of heat transport variability and links to ENSO requires further development in global climate models (GCMs).

8.3.1. Main limiters to heat transport modelling progress

- Poor historical time series information on the MOC and its components.
- Complex patterns of variability make the disentangling of natural and anthropogenic influences difficult.
- A wide range of responses to increasing greenhouse gases in models of the MOC. Detailed process-level understanding of the different responses is required.
- Important flows through narrow channels are poorly resolved in present-day models although the importance of this for the modelled response is unknown.
- Some common errors are found in model simulations of the Southern Ocean. In this region, poor resolution of eddies and boundary currents may be a particular modelling issue and observational gaps limit understanding.
- The limited observational evidence available does not suggest any radical change to the existing picture of the ocean's role in the climate system. While there are deficiencies in climate models, there is no clear evidence that the models on average would over- or underestimate large-scale climate change.

8.4. Water cycle

Many of the effects of climate change will be seen through the hydrological cycle. The hydrological cycle also feeds back on the ocean circulation through the impacts of fresh water on the THC (see Section 8.3). In modelling there is an increasing focus on the prediction of regional changes in precipitation. Overall, climate models predict a drying of the subtropics and increased precipitation at high latitudes, but beyond these broad indications there is considerable variation among model projections.

A key issue from a modelling perspective is a lack of available observations of freshwater fluxes over the oceans. Substantial changes in salinity have been seen, and these have been interpreted as indirect evidence of changes in the hydrological cycle (Bindoff *et al.*, 2007). However, because of the ability of the ocean circulation to transport large amounts of fresh water, the interpretation of the salinity observations remains a matter of debate (e.g. Pardaens *et al.*, 2008; Wu and Wood, 2008). New datasets from satellites suggest a stronger response of the hydrological cycle to temperature changes than is seen in climate models (Wentz *et al.*, 2007), but the datasets are still new and require further scrutiny.

8.4.1. Main limiters to water cycle modelling progress

Limited observations of precipitation and evaporation over the ocean, and non-quantified error bars.

- Limitations to the use of historical salinity observations, and possibly large natural variability, may restrict the use of salinity to quantify changes in the hydrological cycle.

8.5. Sea-ice

In the Arctic, most climate models simulate slower losses of sea-ice in recent decades than have been seen in measurements made from satellites. A few models are able to simulate the observed long-term reducing trend (Stroeve *et al.*, 2007), but it has been suggested that even these models are misrepresenting key processes of ocean heat transport into the Arctic due to limited resolution (Maslowski lecture 2008, http://www.ees.hokudai.ac.jp/coe21/dc2008/DC/report/Maslowski.pdf; see also Maslowski *et al.*, 2007, 2008). Record low sea-ice extents were observed in summer 2007, but it is important not to read too much into an individual season, since year-to-year variability is large and not all observed trends are necessarily anthropogenically forced.

In contrast, in the Antarctic a decrease in ice extent is simulated over recent decades by some models, but other than the Antarctic Peninsula no such decrease has been observed. Clearly there is much research needed to understand recent observed changes and to model them adequately. The sea-ice components of climate models have improved considerably over the past decade, but the overall quality of sea-ice simulation depends also on the driving atmospheric and ocean simulations; and these may now be the limiting factors. Nonetheless a number of important thermodynamic and dynamic processes are still absent from most climate models, and this may be playing a role in some of the model–observation discrepancies (see next paragraph; Hegerl *et al.*, 2007; Randall *et al.*, 2007).

Pronounced changes take place in the albedo of the ice-covered Arctic and Southern Ocean when sea-ice melts or is covered with snow or water. The physics behind the changes is still not fully understood and in particular interactions with the atmosphere, with surface melt water that can form ponds on top of the ice, with varying thicknesses of surface snow and with the freshwater surface layer on top of seawater once the ice has melted. Observations of these parameters are very limited, especially historically. In addition to the above difficulties, present-day modelling may not be adequately representing the dynamics of sea-ice, despite important developments in recent years.

8.5.1. Main limiters to sea-ice modelling progress

- A lack of observations of sea-ice thickness.
- Poor understanding of processes controlling sea-ice distribution, including important driving variables in the atmosphere and ocean.
- Potentially large year-to-year variability makes it difficult to distinguish a climate change signal.
- A lack of understanding of a number of key sea-ice processes.

8.6. Gas exchange/carbon uptake (CO_2, N_2O, DMS)

Understanding the transfer of CO_2 from the atmosphere to the oceans and the carbon cycle is critical to the development of accurate future predictions. Eventually carbon from the atmosphere will end up in the oceans; the problem is in determining the quantity, rates of transfer and location of the fluxes. There is a poor (but improving) knowledge of how the oceanic carbon cycle works. This is a key issue in predicting climate change as the amount of carbon dioxide absorbed by the ocean will strongly affect the impacts of particular atmospheric CO_2 emission pathways on climate. Carbon cycle processes are not yet routinely included in climate general circulation models; hence feedbacks of climate change on carbon uptake are not explicitly modelled. However, in recent years a number of modelling groups have developed simple models of both the land and ocean carbon cycles that have been coupled into GCMs to estimate these feedbacks (Friedlingstein *et al.*, 2006).

Unlike physical processes, there is no convergence of scientific opinion on what are the key processes required to model the role of ocean biology and microbial ecology in carbon uptake and the production of radiatively active gases. Part of this debate involves the complexity that is required to adequately model feedbacks between the biology and climate. Processes (and hence parameterisations) of gas exchange and sinking fluxes are poorly understood (see the 'Science Plan and Implementation Strategy' of The Surface Ocean—Lower Atmosphere Study: SOLAS, 2004; http://www.uea.ac.uk/env/solas), yet models are very sensitive to these parameters. Coastal processes, which are not explicitly included in global carbon models, are likely to be highly dynamic in terms of gas exchange and carbon flux, although their overall importance for long-term carbon storage is uncertain. The debate extends to the physical part of the models: for example, eddies may play an important role in the carbon cycle through vertical transport of nutrients, and it is not known whether such transports can be adequately modelled with the existing resolution that is feasible in climate models.

In summary, process-level understanding is poor so that predictive output differs greatly between models. However, most existing models suggest that the fraction of CO_2 emissions absorbed by the ocean will decrease as climate warms (Denman *et al.*, 2007). This is likely partly due to increased stratification and lower solubility of CO_2 as the ocean surface warms. Recent observations have suggested reductions in carbon uptake in both the Southern Ocean and the North Atlantic; however, it is not clear whether these changes are global in extent or can be related to climate change (Le Quéré *et al.*, 2005; Schuster and Watson, 2007). It therefore remains an open question whether such analyses of recent carbon uptake changes provide a useful constraint on future model predictions.

8.6.1. Main limiters to gas exchange modelling progress

- A lack of quantitative and global understanding of driving biogeochemical processes.
- There is poor understanding of how to incorporate into models the complex biodiversity and functioning of microbial systems and their impact on biogeochemical cycles.
- Uncertainty over what level of complexity is required to adequately model the global effects of the ocean ecosystem.
- A potential high sensitivity of model results (especially vertical tracer fluxes) to resolution.

8.7. Retro-modelling of past climate change

While palaeoclimate scenarios have been only marginally covered in this chapter, they provide some analogies to the rapid increases in temperature and pCO_2 that are currently taking place due to anthropogenic forcing.

Possibly the closest analogues to the present situation are the changes that took place prior to and during the PETM, ~56 million years ago. In this event, global temperatures increased by 5 °C within 1000 years and >2000 GtC as CO_2 was injected into the atmosphere with profound impacts on, and feedbacks from, the oceans (Nunes and Norris, 2006; Sluijs *et al.*, 2007; Zachos *et al.*, 2008). The source of the CO_2 remains controversial, but the most likely candidates are methane hydrates, volcanic emissions and oxidation of sedimentary organic carbon (Sluijs *et al.*, 2007). It should be noted that even this event, considered 'rapid' in geological terms, was a significantly slower change than is projected over the twenty-first century as a result of anthropogenic greenhouse gas emissions. Retro-modelling of the PETM has failed as the models show a strong gradient between the equator and poles, whereas palaeodata convincingly indicate a weak gradient with subtropical conditions in the Arctic (Moran *et al.*, 2006; Sluijs *et al.*, 2006). This implies that some key processes/phenomena that were operating in the PETM are not being taken account of in the current generation of models. The PETM provides us with information on the feedbacks that operate in the Earth system on longer (multi-century) time-scales. It is not clear how important these feedbacks are for the more rapid twenty-first century response to anthropogenic forcing. A greater use should also be made of palaeodata not only to test models but also to investigate the coupling of carbon cycling and climate, and the role of feedbacks and the sensitivity of climate to extreme changes in greenhouse gases (see Zachos *et al.*, 2008).

8.8. Final comments

- There is a need for an improved understanding of the sensitivity of model results to resolution. This will require the development of higher resolution models and/or improved parameterisations of unresolved processes (e.g. vertical mixing, sill through-flows, boundary currents, eddies). Developments of this nature will be highly dependent on the availability of appropriate computer power.
- An integrated global ocean observing programme needs to be implemented to include continuous time series of key ocean–climate variables. Such time series need to be maintained for a sufficient length of time to enable a climate change signal to be distinguished from internal variability (e.g. Argo, Altimetry, RAPID MOC array, Continuous Plankton Recorder, CPR).
- Development of improved and integrated observational datasets of sea-ice thickness is needed.
- A better observational structure is required to measure the large-scale hydrological cycle that includes error estimates.

- Observational constraints on large-scale ocean carbon uptake need to be resolved with an improvement in the understanding of key processes controlling the ocean carbon cycle (leading to development of models at the appropriate level of detail).
- Model development is a painstaking, lengthy and continuous process. Long-term investment commitments for both model development and observational time series must be maintained if current demands for an increasing level of detail and reliability in climate predictions are to be met.
- A greater use should be made of palaeodata not only to test models but also to investigate the coupling of carbon cycling and climate, and the role of feedbacks and the sensitivity of climate to extreme changes in greenhouse gases.

9. Conclusions and Recommendations

The Earth is a blue planet, with two-thirds of its surface covered by oceans. It is home to many hundreds of thousands of organisms ranging from the important microbial viruses, bacteria and Archaea to the microscopic and beautiful siliceous, frustuled diatoms to magnificent whales. Some indication of this diversity and beauty has been captured by the Census of Marine Life.[3] This chapter has been produced to draw attention to the key role that the oceans play in regulating climate as the main heat engine, water reservoir and carbon sink of the planet. It is worth noting as well that the oceans are greatly impacted in turn by climate change with considerable consequences for coastal communities and urban centres from sea-level rise and storms to fisheries and marine transport.

The oceans have been buffering (neutralising) climate change over the past two centuries by absorbing carbon dioxide and heat from the atmosphere generated by both natural variability and man's contribution via increased levels of greenhouse gases in the atmosphere. This key role in climate has helped substantially reduce the rate of climate change. There has not, however, been any reduction in the independent and parallel effects of ocean acidification due to increasing concentrations of CO_2. In recent decades changes have occurred that could alter and possibly undermine the buffering role of the oceans via negative and positive feedbacks. There is a need for a better understanding of these feedbacks, not all of which are fully included in modelling. Some studies indicate that incomplete accounting of land and ocean carbon cycle feedbacks may already have led to an underestimate of the measures needed to mitigate climate change.[4] The

[3] http://www.coml.org
[4] IPCC, SPM, WGIII Legend to Table.

wide range and rates of changes now underway in the seas and the potential for abrupt changes to occur that may be triggered by feedbacks from the oceans, raises concern. It should be noted, however, that it is still impossible, due to a lack of appropriate long-term measurements, to establish the extent to which some of these changes are due to natural variability or directly a consequence of man-made climate change.

9.1. A decade ago

In 1998 in a seminal paper in *Science*, Falkowski *et al.* noted that their intention was ''not to make quantitative predictions of the feedbacks, but to call attention to the sensitivity of marine ecosystems, on all time scales, to climatic and geophysical processes external to the ocean, and the role marine ecosystems have played in regulating the chemistry of the Earth. Our predictive capabilities will improve only when the need for an international network of coordinated long-term (multidecadal) observations of oceanic biology is addressed, and our ability to incorporate the biological processes and feedbacks in coupled ocean–atmosphere models is dramatically improved''. These words are as true today as they were then. A decade has passed, climate change has become a much more urgent issue, and yet the resources to develop an understanding and to measure, through ocean-scale observing programmes, these key feedbacks for climate change have not been made available. Progress has been made, but not at the scale and rate that is needed.

9.2. Warming waters

The main role that the ocean plays in climate variability and change is its huge capacity for the transport and storage of heat that reaches the surface of the planet from the Sun. Some of the heat is transferred to the deep ocean by mixing and some is released back to the air-driving weather systems and warming adjacent coasts (Bindoff *et al.*, 2007). Over recent decades the oceans have warmed rapidly at the surface (~0.64 °C over the last 50 years) and in the whole water column in terms of heat storage. Some idea of the scale of the change is clear when it is realised that warming of the global oceans accounted for more than 90% of the increase in the Earth's heat content between 1961 and 2003. Surface warming has been most pronounced in the Arctic and around the western Antarctic Peninsula where winter temperatures have increased by 5 °C in winter months since the 1950s. Globally, most of the increase in ocean heat content has very likely been caused by increasing greenhouse gases. Heat is the main driver of change within the oceans and leads to the biggest feedbacks to climate change. It has pronounced effects on global ocean circulation, sea-level rise, the concentrations of a major greenhouse gas, water vapour in the air (through increased evaporation), the occurrence of tropical storms, and the melting of polar sea-ice. Increased temperatures and storms could also alter the sea-to-air transfer of sea salt particles and gases that contribute to climate-cooling aerosols and clouds. Warming also affects the water's ability to absorb carbon dioxide and the amount of this greenhouse gas removed from the atmosphere. Finally, there is evidence for increases in the intensity of upwelling at the major upwelling sites around the world leading to large increases in phytoplankton production, anoxia and release of greenhouse gases.

9.3. Freshening waters

Salinity, the second factor that changes the density of seawater besides temperature has shown a remarkable freshening in many regions of the world, including in deep water surrounding Antarctica. The pattern of change is consistent with an enhanced hydrological cycle, a response that has been predicted by climate modellers as a consequence of a warming ocean. In the case of the deep waters around Antarctica the reduced salinities almost certainly reflect the measured deflation of the West Antarctic ice sheet, retreat of glaciers in the Antarctic Peninsula and enhanced basal melt of sea glaciers.

9.4. Changing ocean circulation and sea-level

Warmer water is less dense; as it heats up, a warmer upper layer is established and 'floats' above cooler, denser water. This 'stratification' of seawater is increasing globally, isolating the surface warmer layer from the nutrient-rich deeper waters. As a consequence, the large central tropical/subtropical areas deficient in nutrients are expanding in most oceans. Associated with this change is an expansion of the OMZs in the tropical oceans that has a pronounced effect on the carbon and nitrogen cycles and impacts on marine ecosystems. Combined, all these factors can limit the production of plankton and reduce the amount of carbon dioxide that is removed from the air. Intensified stratification and oxygen depletion may also lead to better preservation of carbon in bottom sediments, thus acting as a sink for carbon dioxide. The net global balance between these opposing processes is likely to leave more carbon dioxide in the air and contribute to increasing rates of global heating.

As rainfall patterns change and ice melts, the freshwater inputs into many seas have increased. The saltiness of the sea has declined markedly in deeper waters of the Southern Ocean and in waters at all depths flowing from the Arctic into the Atlantic. Global circulation in the oceans, the 'conveyor belt', relies upon the formation of cold and salty water sinking in high-latitude seas, and ultimately drives the transfer of heat, nutrients and dissolved gases around the world's oceans. Warmer and less saline polar seas are less effective at driving this process, thereby affecting the way heat is transported around the world. Current models predict a reduction in the intensity of this global overturning circulation of up to 50% by 2100, but no abrupt shutdown,[5] as has been occasionally suggested in the media.

Both the expansion of water due to heating and the melting of glaciers and ice caps cause sea-level to rise. Sea-level is currently tracking the rise in global SST. There are major concerns over the likely contribution that the Greenland, and possibly the West Antarctic, ice sheets might make to sea-level over the next few centuries. The processes involved in ice sheet destabilisation are not well understood and have not been adequately taken into account in current ice sheet models. Historical evidence adds credibility to the possibility of rises at the upper end of and beyond the IPCC AR4 projections by 2100 and a rise of several metres within several hundred to thousands of years. Sea-level rise will affect humans in many ways, including the potential displacement of millions of people. Displacement of populations and loss of coastal lands will likely lead to changes in land and resource use that have the potential to further increase climate change.

9.5. The MOC and cooling of NW Europe

Combined together, changes in salinity and temperature alter density distribution, stratification and the Meridional overturning circulation with large potential feedbacks to climate. However, there is no evidence as yet that the THC/MOC has been changed by the observed salinity and temperature changes. Modelling projections predict that the MOC will reduce by between 0% and 50% by the end of the century, but that this will not lead to a cooling of Northwest Europe, but a slowing down of the warming associated with a rise in global mean temperature.

9.6. Tropical storms

The intensity of tropical storms (hurricanes, cyclones, typhoons) has increased by 75% in the North Atlantic and western North Pacific and a global increase in their destructiveness has been documented. With rising sea temperature and enhanced precipitation the area for seeding tropical storms may expand. These storms may feedback to climate as they have a major impact on the mixing of the ocean. There is, however, at present, no scientific consensus on whether tropical storms will continue to increase in intensity and possibly frequency with rising global temperatures.

[5] IPCC AR4 WG 1, Chapter 10.

9.7. Primary production, biodiversity and non-native species

Production of atmospheric oxygen and fixation of carbon during photosynthesis by phytoplankton enables the Earth to support a rich diversity of marine life and has strongly influenced changes in climate through geological time. The many tens of thousands of planktonic species in the oceans play a key role in ecological and biogeochemical processes that are important in the carbon cycle and climate. Within the last decade major advances have been made in understanding oceanic microbial diversity and ecology, but the extent to which these newly discovered microbial systems will change and impact biogeochemical cycles and climate in a warming world is poorly understood. Changes in the composition of different functional groups in the plankton can strongly impact the biological pump that removes carbon from the upper ocean and have been implicated as one of the causes of the large changes in carbon dioxide between glacial and interglacial periods. There is limited knowledge of the spatial and temporal variability of plankton composition and production versus recycling and export rates in most oceanic geographical provinces. Improved understanding of the interactions between different types of plankton food web structure and the export efficiency of carbon is urgently needed.

Increased inflow of warmer water from both the North Atlantic and North Pacific into the Arctic Ocean has contributed to reductions in sea-ice. In 1998/1999, retreat of the ice from the north of Alaska and Canada allowed the first trans-Arctic migration of a Pacific organism (the phytoplankton *Neodenticula seminae*) into the North Atlantic, for more than 800,000 years. Further introductions of invasive species are expected following the ice reduction in the summer of 2007. Such non-native species could have a large impact on the plankton communities, biodiversity and ecosystems of the North Atlantic and the biological pump—with implications for the amount of CO_2 which is absorbed by the ocean from the atmosphere. Warming seawater is also allowing non-native species to extend their distributions polewards.

9.8. Oxygen

One of the most critical variables in the world's ocean is the distribution of dissolved oxygen (O_2) which is fundamental for all aerobic life. Significant reductions in the O_2 supply to the ocean interior and expansion of low oxygen areas may result from continued anthropogenic global warming, although there may be regional increases in O_2 levels. Models suggest that detectible changes in O_2 content due to global warming may already have occurred. Expansion of the regions of the ocean interior that are devoid of O_2 (anoxic) will adversely affect fish and other species.

9.9. Nutrients

A range of nutrients and micro-nutrients such as iron are essential for phytoplankton growth and production. Strong regional changes in nutrients are expected in the future dependent on variability in wet precipitation, wave storminess, expanding OMZs, increased nitrogen fixation by cyano-bacteria in tropical/subtropical waters, mixing and the depth of stratification. It is not possible at present to predict future trends in nutrients because of the localisation of the changes or how these regional responses will add up to a global mean and influence climate change.

9.10. Ocean uptake of carbon dioxide

The ocean carbon pumps together are possibly the second most important feedback to climate after rising temperatures. The ocean takes up carbon dioxide (CO_2) from the air through three major processes that buffer climate change. Each of these processes has the potential to become less effective as global warming impacts the oceans, leaving more carbon dioxide in the atmosphere, and further increasing climate change.

9.10.1. The Solubility pump

The gas CO_2 is soluble in water and enters the ocean by air–sea exchange. The solubility pump removes large quantities of CO_2 from the atmosphere each year, and stores them in the deep ocean where they cannot immediately contribute to the greenhouse effect. Over $\sim$1000 years, these deep waters are mixed back to the surface, allowing some gas to return to the atmosphere. At high latitudes, dense waters sink, transferring carbon to the deep ocean. Warming of the ocean surface inhibits the sinking and so reduces the efficiency of this pump. Furthermore, as waters warm, the solubility of CO_2 in seawater declines, so less gas can be held in the seawater and taken up from the atmosphere.

9.10.2. The Biological pump

CO_2 is used by phytoplankton to grow. While most organic material is recycled within the food chain, a small proportion of the plankton sinks, and carries with it carbon from the ocean's surface to the deep sea. In the very long term, much of this carbon is stored in sediments and rocks, eventually forming oil and gas deposits. Changes in temperature, acidification, nutrient availability, circulation, and mixing all have the potential to change the plankton productivity of our seas, and are expected to reduce the drawdown of CO_2 via the Biological pump.

9.10.3. The Continental Shelf pump

Water and particles containing carbon are transferred from shallow shelf seas to the deep ocean by this pump. Projected warmer water and higher rainfall (causing reduced salinity) will together lead to increased layering of shelf sea-waters and are expected to contribute to a decline in the efficiency of this pump.

9.10.4. The Carbonate Counter pump

This pump provides a relatively small offset to the above effects. Many marine animals and plants, such as some plankton and corals, use carbon to make calcium carbonate, a building block of their protective walls and shells. By this process, carbon is 'fossilised', but the net growth of these organisms typically does not draw down CO_2, but releases back a small proportion to the water and potentially to the atmosphere, in this way acting as a reverse pump. Acidification (see Section 9.11) in combination with rising temperatures is expected to have a pronounced effect on the efficiency of this pump and through dissolution of carbonate will allow the oceans, over several centuries, to take up slightly more CO_2.

For CO_2 to be transferred from the air to the sea, the level in air must be higher than in the surface water. There is mounting evidence that concentrations in surface seawater have increased faster than in the air in some regions. If this trend became global in extent and continued into the future, the efficiency of oceanic carbon uptake could be expected to reduce.

Given their importance, there is an urgent need to improve understanding of these carbon pumps and better include them in climate model predictions. IPCC AR4, for example, noted that "There are no global observations on changes in export production or respiration". Of great concern is evidence from observations and models that the uptake of carbon dioxide by the oceanic sink may be declining, and that the terrestrial sink may not be keeping pace with increasing emissions.

9.11. Acidification

As well as causing climate change through the 'greenhouse effect', carbon dioxide is having a profound effect on the ocean by making seawater more acidic (lower alkalinity). As this gas dissolves into the ocean, it reacts, forming carbonic acid and reduces the pH of seawater. The changes in acidity measured in the open ocean also appear to be extending to some shelf seas. Due to the rapid rate of acidification, the ocean is predicted to be less alkaline, within 50 years, than at any time in the past 20 million years and possibly since the PETM, 55 million years ago. There is concern that ocean organisms will not be able to adapt to the speed and scale of change now underway. Among organisms expected to be most affected are some plankton (e.g. small snail-like pteropods; Fig. 1.41) and corals. These

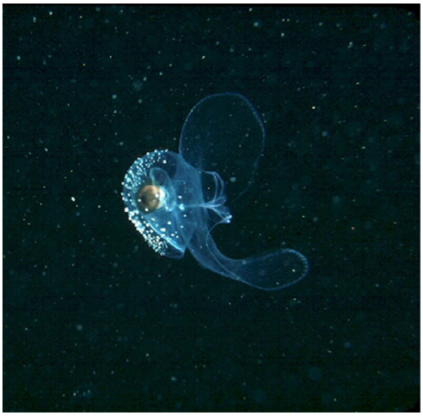

Figure 1.41 A pteropod. Image from http://pubs.usgs.gov/of/2000/of00-304/htmldocs/chap11/images/pelag10l.jpg (U.S. Geological Survey).

organisms may be vital to the whole food chain, but also to the way the oceans take carbon dioxide out of the atmosphere and store it in the oceans, thus affecting the Biological pump.

9.12. A special case: The Arctic

Covered by ice for much of the year, the Arctic Ocean is strongly influenced by relatively small changes in sea and air temperature. Warming may change Arctic winds, the thickness and extent of sea-ice, and the water's salinity by melting ice and driving higher precipitation. Alterations in each of these may trigger large changes in regional climate within decades, with downstream consequences for the rest of the world.

The Arctic has lost around 30% of its summer sea-ice in recent decades, with the most extreme reductions observed during the last decade. Sea-ice extent in 2007 was at a record low that was 40% below the recent long-term average. Despite being a cooler year than most in the past decade, the sea-ice extent in 2008 was also well below the long-term average, although it was not as low as the 2007 record. Sea-ice in 2008 was notable in that there is now little of the thick, old ice left, which could make the region increasingly vulnerable to further ice loss. The Arctic has been losing its sea-ice rapidly and it has been suggested that this may lead to a step change in the whole system due to a loss of the capping layer of fresh water in the Arctic Ocean. Ice is highly reflective and returns much of the solar radiation back to space; as it melts and exposes the dark ocean surface, the water rapidly absorbs and accumulates the incoming radiation over the long Arctic summer. This creates a feedback that tends to accelerate ice loss and warming of both sea and air, triggering further ice loss and regional warming.

Model predictions for the disappearance of Arctic sea-ice during summers vary between 2013 and the end of the century. However, most climate models underestimate the rate of ice loss over recent decades, and even those that simulate recent trends suffer from deficiencies in resolution that may lead to an underestimate of future change. Some model results suggest that a threshold (tipping point) has been passed in the loss of Arctic sea-ice, meaning that recent low ice conditions may persist for some time.

9.13. Methane

A warmer Arctic Ocean releases warmth back to the air, which can penetrate into adjacent coastal areas as far as 1500 km. This can melt permafrost, and potentially lead to the release of methane stored within it. Methane is 25 times more potent as a greenhouse gas than CO_2, thus the release of methane is potentially a large feedback to climate change. The rate of release of this gas into the atmosphere has slowed down over the last 20 years; a rate change that is not well understood. In 2007 the concentrations of methane in the air increased, particularly in the Arctic, suggesting a release from Arctic permafrost among other sources.

Extensive deposits of methane hydrate (methane gas trapped in an ice-like solid) are found beneath coastal Arctic seas, and within permafrost on the adjacent land. A large release of methane from warming of marine hydrates is thought to be unlikely, unless warming causes landslides on steep continental slopes which hold hydrates or inflow of warmer water from adjacent oceans intensifies. Recently, large volumes of methane have been observed bubbling from the sea bed of the Laptev and East Siberian shelf seas and off Spitzbergen. It is not clear, however, if these new findings are a response to an anthropogenic warming signal.

9.14. Greenland ice sheet

A recent accelerating reduction in the mass balance of the Greenland ice sheet may in part be due to higher temperatures in the adjacent ocean. While the precise threshold is not known, unless the trend in global temperature rise reduces, the temperature threshold for an eventual complete melting of the Greenland ice sheet may be crossed this century. Melting of the Greenland ice sheet alone could lead to 7 m of sea-level rise over thousands of years with implications for all coastal regions around the world.

9.15. The Southern Ocean

The Southern Ocean, including the ACC, plays a critical role in driving, modifying and regulating global climate and climate change. An increase in westerly wind speeds due, at least partly, to human influences (increases in greenhouse gases and ozone depletion) has been observed, and a continuing warming of the Southern Ocean and strengthening of the westerly winds is predicted. The ocean west of the Antarctic Peninsula has warmed rapidly since the 1950s in parallel with changes in atmospheric climate and the cryosphere. Changes in the Southern Ocean are closely connected to the production and melting of sea-ice, the formation of which is the largest seasonal phenomenon on Earth. Sea-ice has a major affect on the energy budget of the Earth and thus climate. A 25% decrease in sea-ice extent over the next 100 years is predicted. The evolution of the Southern Ocean along the coast of Antarctica is a major control on the stability of Antarctic glaciers and on the mass balance of the ice sheet as a whole. A reduction in the mass balance of the northern sector of the West Antarctic ice sheet induced by a thinning of the fringing ice shelves is most likely associated with greater subsurface penetration of Circumpolar Deep Water onto the continental shelf. The rates of retreat of glaciers in the Peninsula and in West Antarctica are predicted to accelerate. The Southern Ocean is an important sink for carbon dioxide, a sink that has been reported as possibly weakening and approaching saturation, and which urgently requires further investigation. In the future, it is estimated that the CO_2 sink in the Southern Ocean will undergo further modification. Marked changes have occurred in Southern Ocean ecosystems including a substantial decline in krill numbers. The changes in biodiversity, combined with the effects of acidification and rising temperature are likely to lead to important modifications to unique Antarctic ecosystems with associated feedbacks to the carbon cycle and climate.

9.16. Modelling

Pronounced changes in ocean processes are now being recorded, some of which through complex feedbacks may accelerate global warming. Models are an essential tool to help investigate these feedbacks and their role in future climate change. Global Climate Change models have proved to be especially reliable in predicting future changes in global temperature. Modelling of ocean processes is less advanced, however, and there are a number of general limitations to progress in modelling feedbacks including poor data, a lack of understanding of key processes and inadequate representation of the processes in models (parameterisation). Data are needed for input to and validation of models; a lack of historical measurements and time series of key variables and processes is a major restriction on modelling progress. To address this problem there is an urgent need to implement an integrated global ocean observing programme that includes continuous time series of key ocean–climate variables. Such time series need to be maintained for a sufficient length of time to enable a climate change signal to be distinguished from internal natural variability (e.g. Argo, Altimetry, RAPID MOC array, ADCP arrays, CPR).

In a number of cases, models representing key ocean feedbacks that contribute to climate have failed to represent observations or capture regional variation between different ocean basins, for example, the heat uptake, models used in IPCC AR4 and sea-ice models in the Arctic. Poor understanding of processes, inadequate representation of ocean/atmosphere drivers, a lack of inclusion of some important processes in models, scaling factors and spatial resolution as well as a lack of measurements are likely contributors to the failures. Some important ocean feedbacks such as the different ocean carbon pumps are not well represented in GCMs to date. Some studies of mitigation options omit feedbacks from the carbon pumps altogether. This omission could lead to an underestimate of the rate of future climate change, the stabilisation targets necessary to limit warming and, thus, the measures needed to achieve mitigation.

There is no clear scientific agreement on the key processes required to model the role of ocean biology and microbial ecology on carbon uptake and the production of radiatively active gases. The processes involved in gas exchange and sinking fluxes, and their parameterisation are especially poorly understood and yet models are very sensitive to these parameters. In particular, it is not yet clear how the complex biodiversity and functioning of microbial systems and their impact on biogeochemical cycles should be incorporated into models. In the case of acidification, open ocean models work well, but the models are less effective in upwelling, coastal and shelf sea regions, which could be especially vulnerable to increased acidification.

Observed changes in ocean feedbacks have occurred with a global average (land and sea) temperature rise of less than 1 °C. Further warming may increase the impacts of the oceans on climate change, and amplify feedbacks. Despite considerable progress in the development of ocean/climate models the above limitations mean that their output and prognoses need to be viewed with caution. It should be stressed, however, that while the models are not perfect, this does not reflect on their usefulness as they are an essential tool to look into the future.

9.17. Final concluding comments

This chapter demonstrates that the oceans are vital in regulating our climate. There is an urgent need to improve understanding of the interaction between the oceans and climate change and better include this in climate model predictions. Greater use should also be made of palaeodata to test and inform climate models.

The oceans have buffered climate change substantially since the beginning of the industrial revolution, acting as a sponge to carbon dioxide and heat from global warming. While it was assumed this would continue, this chapter gives a warning—changes underway in our ocean may accelerate warming or its consequences to organisms, and have the potential to increase climate change itself. In some examples, such as sea-ice loss, this process may already be underway. In this sense, and to quote a reviewer: "The ocean strikes back".

Appendix: Workshop Participants

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