6.1. INTRODUCTION: CLIMATE VARIABILITY

The ocean is host to a vast diversity of life; it is of immense social and economic value to us and it plays a crucial role in our climate system over synoptic to multimillennium timescales. Consequently, it is of great importance to understand changes that oceans have gone through in the past and how they might change in the future as we increasingly perturb the climate system. In the context of anthropogenic climate change, the ocean acts as an important buffer. The majority of extra heat entering the climate system through enhanced greenhouse forcing (~80% since 1960; Levitus et al., 2005) and almost half of the anthropogenic carbon dioxide (CO₂) since pre-industrial times (Sabine, 2004; or ~30% since 1960, Canadell et al., 2007) have been sequestered and stored in the ocean. While mitigating against more rapid temperature rise in the atmosphere, the oceanic absorption of heat and CO₂ has impacts on the ocean environment. For example, the absorption of heat has caused thermal expansion of the ocean, which is a major contributor to global sea-level rise (e.g., Domingues et al., 2008); temperature and stratification increase can directly (via physiological changes) and indirectly (via changes in ocean mixing and nutrient supply) affect marine species. The sequestration of anthropogenic CO₂ has measurably altered ocean chemistry, reducing water pH and aragonite saturation (Raven et al., 2005). An increasing number of studies are showing that such chemical changes can have major impacts on both marine calcifiers and higher marine animals (Fabry et al., 2008; Doney et al., 2009). As a result of the physical and biochemical changes that have occurred in the ocean, some evidence is emerging that the ability of the ocean to mitigate atmospheric temperature rise, via the sequestration and removal to the deep ocean of CO₂, is declining (Canadell et al., 2007; Le Quéré et al., 2009).
Compared to pre-industrial times, atmospheric CO₂ concentrations have increased by over 30%, which, together with increases in other greenhouse gases (GHGs) have driven a rise in globally-averaged ocean surface temperatures of >0.6°C (IPCC, 2007a), lagging slightly behind increases in atmospheric temperatures. Over the same time, average surface pH has reduced from ~8.2 to 8.1, corresponding to a ~30% increase in hydrogen ion content (Raven et al., 2005). The oceanic changes already observed are likely to be compounded in the future, given that GHG emissions are still growing and show little sign of an imminent slowdown (e.g., Le Quéré et al., 2009). It is unlikely, however, that future change will scale in the same way as historical change, as the radiative forcing perturbation increasingly diverges from pre-industrial levels. Feedbacks within the system mean that future responses will probably be non-linear and may result in tipping points or abrupt changes to components of the system (e.g., Schneider, 2004; Lenten, 2012, this volume). While some of this non-linearity is inherent in the climate models (e.g., water vapour and ice–albedo feedbacks), we know that certain feedbacks are poorly represented (in particular those related to clouds) or entirely missing. The climate models used to inform AR4 projections do not explicitly include carbon cycle interactions or feedbacks related to terrestrial or marine biological systems. Some of the models taking part in AR5 will include ocean and terrestrial carbon cycle components (Taylor et al., 2009).

The goals in this chapter are threefold. Our first goal is to examine the large-scale changes to the oceans that have been observed over recent decades and examine any link to anthropogenic climate change (Section 6.2). Two fundamental challenges exist here. Firstly, compared to terrestrial or atmospheric systems the ocean is relatively poorly sampled, particularly in the Southern Hemisphere and as we go further back in time prior to the advent of remote measurement. In Section 6.2.1, we describe some of the important new observing systems and the challenges they face. Secondly, the oceans are highly dynamic and, like the atmosphere, exhibit natural variability on a wide range of timescales. Given the brevity of the observational records, it therefore becomes difficult to disentangle long-term natural variability from change that is driven by increased greenhouse forcing, particularly at regional and smaller scales. In Section 6.2.2, we briefly discuss the primary modes of variability that are important for the ocean.

The second goal is to examine projections of future physical and chemical changes in the ocean (Section 6.3). This is primarily based on the intercomparison of multiple coupled climate models. Such a review is necessarily biased towards changes in the physical environment, given that most climate models are yet to include sophisticated biochemical components. There is, however, a more limited set of models that provide insights into chemical and biological change, and attempts have been made to infer how ecosystems may respond to given projections of the physical system. We focus our discussion of projections on three key areas: (i) the tropical Pacific Ocean, where changes in the El Niño–Southern Oscillation (ENSO) can have major regional and global repercussions (see Latif and Park, 2012, this volume); (ii) the Southern Ocean, which plays an important role in both the global overturning circulation and the two-way transfer of carbon between the ocean and atmosphere; and (iii) sea-level rise, where some post AR4 research points towards considerably larger changes than those derived from climate models.

The third goal is to discuss how the observed and projected changes to the ocean will affect the regulatory function of the ocean within the climate system. We review a number of feedback mechanisms, including those associated with the carbon pumps (physical and biological processes that can transport carbon from the surface ocean to the deep ocean or sediment) that can modulate the ability of the ocean to sequester atmospheric CO₂ (Section 6.4). In Section 6.5, we draw conclusions and briefly examine what the next generation of climate models have in store.

6.2. OBSERVED OCEAN VARIABILITY AND CHANGE

6.2.1. Observing the Global Ocean

Our understanding of the ocean and its role in the climate system is underpinned by observations. However, large parts of the ocean remain poorly sampled, partly because the ocean can be a hostile and inaccessible place, making it a costly challenge to take observations. In addition, unlike in meteorology where there is obvious day-to-day justification for the collection of large-scale, high-resolution atmospheric data (e.g., weather forecasting, aviation), this is not the case for the ocean.

Over recent decades, there have been considerable improvements in our ability to observe the ocean as new technologies have been deployed, especially under the auspices of various international collaborations. A number of particular projects have led to major advances in our understanding of the ocean’s role in climate. These include: The Tropical Ocean – Global Atmosphere (TOGA) study (Hayes et al., 1991), which has implemented an array of upper ocean moorings along the equatorial Pacific Ocean. This has provided valuable insights into ENSO mechanisms and has led to the ability to produce long-term forecasts of ENSO-related climate variability. The World Ocean Circulation Experiment (WOCE), an unprecedented collaboration of 30 countries during
the 1990s to produce a three-dimensional view of the structure of the ocean (Siedler et al., 2001), using a variety of technologies, including an extensive network of hydrographic sections, current moorings, Expendable Bathythermograph (XBT) deployments, floats, and so on.

Rapid Climate Change (RAPID) mooring array, an array of moorings spanning the North Atlantic at 26.5°N that provides an accurate measure of the upper and lower branches of the Atlantic Meridional Overturning Circulation (AMOC or ‘thermohaline circulation’). This circulation carries heat, CO₂, and other properties throughout the global ocean. It includes both upper and deep ocean circulation pathways, connected by regions of sinking or ascent at high latitudes. As such, it plays a major role in modulating the climate system. In fact, it has been implicated as a major driver of rapid climate change at certain times in the palaeo record (e.g., McManus et al., 2004).

Over recent years, two new technologies in particular have produced step-changes in our ability to observe the global ocean:

Satellite technology now gives us near-global coverage of the surface ocean. Variables that are now routinely measured from space include sea surface temperature (SST), sea-ice cover, wind speed (that can be used to derive the surface flow in the upper few tens of metres of the ocean), sea-level anomaly (which provides information about the depth-integrated ocean circulation), and ocean colour (which can be used to infer primary productivity and total biomass). Global coverage of ocean colour has only been available since the mid-1990s as part of projects like the Sea-viewing Wide Field-of-view Sensor (SeaWiFS); although limited measurements from other instruments were available from 1978 to 1986 (McClain, 2009). In 2009, a satellite mission, Soil Moisture Ocean Salinity, was launched that utilizes the sensitivity of microwave emissions to the dielectric properties of water (which is sensitive to salinity) to provide the first exploratory global sea-surface salinity (SSS) measurements from space. Further satellite missions are planned for the near future (Lagerloef et al., 2009).

ARGO profiling floats provide near-global coverage of the subsurface ocean. Over the last decade, a working array of ~3000 floats has been built up (http://wojcommops.org/cgi-bin/WebObjects/Argo). Floats typically descend to ~1000 m–2000 m measuring temperature, salinity, and pressure, remain submerged for ~8–10 days, and then ascend to the surface where the stored data are relayed via satellite to a data centre. In its lifetime of a few years, each float can provide hundreds of profiles. Recent improvements to some of the floats include an ice-detection algorithm and interim storage that means that the ARGO network is being extended to the poorly sampled sub-sea-ice regions (Klatt et al., 2007). In addition, the advent of new stable oxygen sensors (Tengberg et al., 2006) has led to pilot deployments of ARGO floats with oxygen measuring capability, with limited results now becoming available (e.g., Riser and Johnson, 2008).

While there has been an order of magnitude jump in our observing capability over recent years, we are still limited to relatively short data records over most of the ocean. Satellite-derived SST extends back as far as the 1970s (Guan and Kawamura, 2003) and altimetry with accuracies of a few cm since the early 1990s (Fu et al., 1994), but large-scale global coverage of the subsurface ocean, provided by ARGO, is available for less than a decade. Furthermore, there are large discrepancies in the type of data available. Ocean temperature is the most well-observed ocean property, due a long history of hydrographic and XBT measurements and, more recently, the inclusion of satellite data. Even so, data gaps mean that identifying reliable long-term trends is still not possible in some regions (Deser et al., 2010; see also Figure 6.1). Salinity has also been routinely measured. However, as a result of XBT temperature measurements in the salinity database is only a fraction of the size of the global temperature record (de Boyer Montegut et al., 2007).

Ocean chemical (e.g., oxygen, pH, nutrients) and biological (e.g., chlorophyll/productivity) properties are much more poorly sampled. A striking consequence of this is that over 28,000 significant biological changes were noted for the terrestrial system in the IPCC AR4, whereas only 85 were noted for marine and freshwater systems (Richardson and Poloczanska, 2008). There is a vital need to maintain our existing observational network and to improve it in some important key areas. In particular, the Southern Ocean is very poorly sampled due to its remoteness and its inhospitable nature.

6.2.2. Natural Modes of Variability

In addition to data scarcity, examining long-term ocean change and its possible link to anthropogenic climate change is made much more challenging because of the signal-to-noise problem. The oceans are dynamic and, like the atmosphere, exhibit natural variability on a wide range of timescales: ocean ‘weather systems’ or eddies are pervasive throughout the ocean, there is strong seasonality in ocean properties and circulation and, on longer timescales, the ocean exhibits intrinsic inter-annual variability, spanning a few years to multiple decades. In addition to this intrinsic variability, the ocean is also affected by non-anthropogenic
external factors, including solar variability and volcanoes. Given that natural variability may possess significant low-frequency components, long observational records and sophisticated statistical techniques are often required to extract any long-term signals from anthropogenic interference (see Latif and Park, 2012, this volume).

Climate variability is largely understood in terms of recurring regional patterns (or ‘climate modes’) related to natural internal dynamics of the ocean and atmosphere, and much effort goes into understanding the mechanisms behind, and the impacts of, these modes. A particular mode is associated with a distinct set of responses in a variety of climate properties both locally (where the dynamic interactions are taking place) and remotely (via changes propagated to other regions via the atmosphere and ocean). A further complication in separating anthropogenic signal from climate noise exists since some evidence suggests that the spatial pattern of anthropogenic climate change may project on to existing modes of variability (Corti et al., 1999; Stone et al., 2001; Latif and Park, 2012, this volume). A clear example of this is the long-term historical trends in highlatitude atmospheric circulation that exhibit strong similarities to the positive phase of the naturally occurring Southern and Northern Hemisphere annular modes (Stone et al., 2001; Brandefelt and Källén, 2004). Before turning to the observed long-term changes, we introduce the dominant ‘modes’, how they impact ocean properties, and whether there is evidence for any long-term changes in these modes.

6.2.2.1. El Niño—Southern Oscillation (ENSO)

ENSO is a coupled oscillation in the tropical Pacific Ocean and atmosphere, characterized in the ocean by a shift in the position of the very warm water that makes up the tropical Western Pacific Warm Pool. During the El Niño phase of ENSO, as the tropical easterly winds relax, the equatorial thermocline flattens and the Warm Pool waters flood eastwards, shifting the centre of maximum atmospheric convection. This perturbation in the tropical Pacific propagates via the atmosphere and ocean to remote regions around the world (Bjerknes, 1969), causing significant changes to regional weather conditions (including tropical cyclones and the monsoon systems) and to ocean ecosystems and fisheries (e.g., Holland 2009 and references therein). The tropics represent the major source of natural CO₂ outgassing from the ocean to the atmosphere and ENSO-related changes strongly modulate this flux (Feely et al., 1999). As such, changes in the frequency or strength...
of this mode would have major impacts on global and regional climate.

A multidecadal increase in the strength of ENSO events has been reported over the last 30 years with the two largest recorded events having occurred in 1982 and 1998 (Fedorov and Philander, 2000). The shortness of the instrumental record, however, precludes attributing these changes to anthropogenic warming, especially given that ENSO exhibits substantial low frequency modulation (see PDO in Section 6.2.2.2). Furthermore, in a multimillennial climate model simulation, Wittenberg (2009) finds substantial low frequency natural variability, with multidecadal to inter-centennial periods of both, almost no ENSO activity, and much enhanced ENSO activity. If this is representative of the real climate, it suggests that untangling human-induced trends from natural variability would be extremely hard without very long climate records.

Since the 1970s, the ‘typical’ ENSO pattern, with maximum El Niño (La Niña) warming (cooling) centred in the eastern Pacific has often been replaced by a distinctly different pattern where maximum warming (cooling) is shifted to the central Pacific (Latif and Park, 2012, this volume). This modified variability, termed El Niño Modoki (Ashok et al., 2007), produces very distinct changes to the Walker Circulation (Ashok et al., 2007; Wang and Hendon, 2007) and remote responses compared to canonical ENSO events (e.g., Ashok et al., 2007; Taschetto et al., 2009, 2010).

Air–sea fluxes of CO$_2$ in the tropical Pacific (particularly in the eastern and central basin) dominate inter-annual variations in oceanic CO$_2$ and may constitute up to 30% of atmospheric variability (Feely et al., 2002). Upwelling of deep CO$_2$-rich water in the eastern Pacific drives a net CO$_2$ flux to the atmosphere. During El Niño events, the weaker winds, a reduction in upwelling, and the eastward expansion of the Warm Pool waters can considerably reduce normal outgassing (from ~0.8–1 PgC per year to 0.2 PgC–0.4 PgC per year; Feely et al., 2002). Decadal variability in ENSO activity can cause significant low frequency variations in the build-up of atmospheric CO$_2$ levels. Thus, long-term climate changes in the tropical Pacific may be associated with significant modulation of the global carbon cycle.

### 6.2.2.2. Pacific Decadal Oscillation (PDO)

Closely linked to low frequency modulations of ENSO is the Pacific Decadal Oscillation (PDO: Mantua et al., 1997) or closely related Inter-Decadal Pacific Oscillation (Power et al., 1998). (These phenomena are defined differently but express essentially the same thing.) The associated pattern of surface ocean warming and cooling is similar to ENSO (although the temperature anomaly tends to be broader in a north–south sense). However, changes in the PDO phase occur on multidecadal, rather than inter-annual, timescales. Untangling the influence of this low-frequency variability and anthropogenic long-term change requires care (see discussion of SST in Section 6.2.3). The widespread impact of this mode on marine and terrestrial systems has led to transitions between its phases being described as ‘regime shifts’. A number of Pacific marine species are thought to be sensitive to changes in the PDO (see review by Mantua and Hare, 2002). Increases in tropical wind strength and equatorial upwelling associated with the most recent PDO phase change in the late 1990s are consistent with increased CO$_2$ outgassing from the tropical Pacific in recent years (Feely et al., 2006). The brevity of the data record precludes the identification of any long-term trends in the PDO (Henson et al., 2010). Pacific decadal variability is further explored in Latif and Park (2012, this volume).

### 6.2.2.3. Southern Annular Mode (SAM)

The Southern Annular Mode (SAM) or Antarctic Oscillation is the dominant pattern of natural variability in the Southern Hemisphere outside the tropics. The SAM is characterized by a poleward intensification (equatorward weakening) of the mid-latitude westerly winds that extends from the surface to the upper jet, in its positive (negative) phase. It has an important influence on rainfall over high-latitude countries (e.g., Gillett et al., 2006; Hendon et al., 2007), Southern Ocean circulation, SST, and sea-ice concentration (e.g., Sen Gupta and England, 2006) as well as biological productivity (Lovenduski and Gruber, 2005). Modelling studies suggest that inter-annual changes in the SAM can account for ~40% of the inter-annual variability in CO$_2$ flux in the Southern Ocean. During a positive SAM, increased winds drive greater upwelling of natural carbon from the deep ocean to the surface, diminishing the ability for the ocean to absorb CO$_2$ (Lenton and Matear, 2007). The SAM has shown a robust long-term trend towards a more positive state over recent decades, with the strongest surface trend occurring in austral summer (Thompson and Solomon, 2002; Gillett and Thompson, 2003). Observational and modelling studies have attributed the trend to anthropogenic factors, in particular a combination of stratospheric ozone depletion and enhanced GHG forcing (e.g., Fyfe et al., 1999; Arblaster and Meehl, 2006; Cai and Cowen, 2007). Observational (Le Quéré et al., 2007) and modelling studies (Lenton et al., 2009) suggest the ocean circulation changes associated with the SAM trend have partially offset the increase in the Southern Ocean sink of CO$_2$ that would otherwise have occurred due to rising atmospheric CO$_2$ concentrations.

### 6.2.2.4. North Atlantic Oscillation (NAO)

The North Atlantic Oscillation (NAO) is a prominent mode of variability with important impacts from the polar to...
subtropical Atlantic and surrounding landmasses (Latif and Park, 2012, this volume). It is most pronounced during boreal winter. The NAO is associated with changes in wind strength and direction, heat and moisture transport, and the frequency and strength of storms (see Hurrell et al., 2003). It also plays an important role in modulating ocean properties, such as SST, mixed layer depth, and, on long timescales, basin-wide changes in circulation (Visbeck et al., 2003) including modulation of North Atlantic overturning (e.g., Biastoch et al., 2008). Low frequency biological activity has been related to the NAO variability. In the subpolar regions for example, the spring phytoplankton bloom is delayed by 2–3 weeks during positive NAO phases, related to changes in wind-driven mixing (Henson et al., 2009). A positive trend in the winter NAO index over the last 40 years of the twentieth century has been attributed in part to increases in GHGs (see the review by Gillett et al., 2003). However, the NAO is subject to strong multidecadal variability (particularly in winter) and the NAO index has subsequently decreased to relatively quiescent conditions over the last decade (e.g., http://www.cpc.ncep.noaa.gov/products/precip/CWlink/pna/new.nao.shtml). Modelling studies suggest that the phase of the NAO can drive sub-basin-scale changes in surface CO2 concentrations, driven largely by modulations in the advection of dissolved inorganic carbon by the North Atlantic Current. These NAO-driven changes exceed anything driven by secular trends in atmospheric CO2 (Thomas et al., 2008).

6.2.2.5. Atlantic Multidecadal Oscillation (AMO)

Multidecadal variability is also present in the Atlantic in the form of the Atlantic Multidecadal Oscillation (AMO) (Kerr, 2000; see also Latif and Park, 2012, in this volume). The AMO appears to be intimately tied to variations in the Atlantic overturning circulation, over timescales of ~60 years. Changes in the AMO are manifested as slow changes to Atlantic SST (Knight et al., 2005). The timescales involved in this natural oceanic oscillation mean that great care must be taken in distinguishing the low-frequency changes associated with this mode and any human-induced signal. This can be a non-trivial task, as can be seen in the recent controversy regarding the relative importance of the AMO compared to global warming on the melting of Swiss glaciers (Huss et al., 2010a, 2010b; Leclercq et al., 2010; Cogley, 2012, this volume).

6.2.3. Surface Temperature and Salinity

SST is of fundamental importance to the climate system. SST plays a major role in determining the rate of heat, moisture, and gas flux between the ocean and atmosphere. It represents a proxy for upper ocean mixed-layer temperature and, as a result of the large thermal inertia of this upper layer, makes it an important quantity for monitoring climate change. SST is also a major factor affecting marine ecosystems.

Recent estimates suggest that globally-averaged SST has increased by ~0.68°C per century (linear trend, 1901–2004, HasSST2), with more rapid warming occurring in the later part of the time series (Rayner et al., 2006). It should be noted that such estimates incorporate sparse data in certain extended regions, including much of the Southern Ocean (Figure 6.1). The spatial pattern of warming is highly non-uniform. It is well-known that ocean temperature increases lag behind the land surface (as a consequence of different thermal capacities and the ocean’s greater ability to lose energy via latent heat fluxes: Sutton et al., 2007; Dommengen, 2009). Furthermore, the high-latitude Arctic regions have amplified warming (as a result of sea-ice interactions: IPCC, 2007a). In addition to this, there is a high degree of spatial variability in trends across the oceans. Deser et al. (2010) analysed a variety of global SST (and air temperature) datasets to identify robust spatial trends over a period of time sufficient to average over much of the inherent natural variability (1900–2008). Warming has been occurring in most datasets over all areas where a trend could be computed, except in the North Atlantic where there is a significant cooling, possibly a consequence of long-term changes in the NAO or the overturning circulation (Deser et al., 2010). Insufficient data were available for high-latitude Arctic and Southern Ocean regions.

The various SST products present conflicting trends in the central and eastern Pacific (Vecchi et al., 2008; Deser et al., 2010), an area important for ENSO processes. However, examination of independent air temperature measurements is consistent with the uninterpolated SST datasets that more closely reflect the in-situ data, suggesting that warming is occurring across the entire tropical Pacific.

It is interesting to examine equivalent trends for a shorter time period. Figure 6.1b shows a 30-year trend from the NOAA Optimum Interpolation SST V2 dataset (Reynolds et al., 2002), which includes satellite data. While the globally-averaged rate of warming is larger than for the longer period, significant cooling is evident over large parts of the central and eastern Pacific Decadal Oscillation. The pattern in the Pacific is consistent with the observed downward swing in the phase of the PDO (Section 6.2.2.2). The cooling in the Southern Ocean is consistent with the observed intensification of the mid-latitude westerlies (Section 6.2.2.3), which force cold high-latitude surface water northwards (see Sen Gupta and England 2006, their figure 12).

Salinity, like temperature, plays a key role in ocean processes, affecting water density, water column stability and the overturning circulation. It also provides valuable information for understanding the marine hydrological
cycle (e.g., Durack and Wijffels, 2010; Helm et al., 2010). The large-scale salinity distribution is largely determined by local precipitation minus evaporation rates and modified by ocean circulation and mixing, and at coastal locations by freshwater runoff or sea-ice formation or melt. As tropospheric temperatures increase, theoretical arguments suggest a resultant intensification of the hydrological cycle (e.g., Held and Soden, 2006). Over the ocean, such changes are difficult to measure directly due to the scarcity of sampling stations, which are primarily located on land areas, and the highly heterogeneous nature of the rainfall distribution. Examination of historical salinity data, however, is able to resolve changes that are uncertain in the rainfall datasets. Even prior to the expansion of salinity observations afforded by ARGO, studies were able to identify a freshening (reduced salinity) at high latitudes and increased subtropical water salinity (Antonov et al., 2002; Boyer et al., 2005). More recent studies incorporate the independent ARGO dataset. Hosoda et al. (2009), Helm et al. (2010), and Durack and Wijffels (2010) demonstrate these results to be robust. Using subsurface salinity data, Helm et al. (2010) find a salinity increase within the near surface (~100 m) salinity maximum and a freshening of the intermediate depth salinity minimum (for ~1970–2005). The upper waters are primarily sourced from the subtropics (regions of relatively weak net precipitation) while intermediate waters are sourced at higher latitudes of the North Atlantic and Southern Ocean (regions of relatively high net precipitation). These changes imply an intensification of the hydrological cycle. Durack and Wijffels (2010) have constructed a new salinity database that attempts to remove the effect of inter-annual ENSO variability. The 1950–2008 trend (Figure 6.2) bears a strong resemblance to the mean salinity field and, in turn, to the mean freshwater flux field (see Durack and Wijffels, 2010, their figure 5). This is again indicative of a spin-up of the hydrological cycle. In precipitation-dominated low-salinity regions, including the tropical west Pacific Warm Pool, under the Atlantic ITCZ, and at high latitudes (particularly in the Southern Ocean and the North Pacific), significant freshening has occurred. Conversely, in the subtropical gyres, located in regions of mean atmospheric subsidence, salinity has been increasing. While globally-averaged surface salinity remains nearly unchanged, the normal gradient between the relatively fresh Pacific basin and salty Atlantic basin has increased (Durack and Wijffels, 2010). Considerable interest has been focused on the Pacific Warm Pool. In the region of the warmest waters (>28.5°C), temperatures have increased on average ~0.3°C and salinity has dropped by ~0.34 psu over the last 50 years (1955–2003: Cravatte et al., 2009). The magnitude and distribution of the observed salinity changes are consistent with the long-term warming signal (as theory suggests that net precipitation will change in proportion to the SST change; Held and Soden, 2006; Cravatte et al., 2009). An important consequence of these changes is a considerable expansion of Warm Pool area, mean depth and volume, and an eastward shift of the Warm Pool edge (as defined by either the 34.6‰ or 34.8‰ isohalines or the 29°C isotherm) of between 15° and 20° of longitude (Cravatte et al., 2009). This region plays an important role in ENSO dynamics (e.g., Picaut et al., 1997). In addition, the distribution of important marine species (e.g., skipjack tuna) are highly sensitive to the location of the Warm Pool boundary (Lehodey et al., 1997, 2010).

6.2.4. Heat Content and Sea Level

Given that ~80% of the total additional heat related to anthropogenic warming now resides in the ocean (Levitus et al., 2005; Bindoff et al., 2007), ocean heat content is an important measure of the change in the Earth’s radiative balance (Harvey, 2012, this volume). It is also a major contributor (via thermal expansion) to sea-level rise (e.g., Domingues et al., 2008). In addition to problems of data scarcity, estimating global heat content has been subject to data bias issues, most importantly in relation to
the XBT devices, which have historically been the most common instruments for measuring subsurface temperature (>70% of the global temperature archive: Wijffels et al., 2008). XBTs are dropped from ships while attached to a thin wire and fall from the surface to the deep ocean, measuring temperature. These devices estimate depth using a fall-rate equation and the time elapsed from release. It has long been recognized that errors in the fall-rate calculation produce significant biases in heat content estimates. A number of recent studies have used different techniques to correct for the XBT bias (see review by Lyman et al., 2010). Results from three such studies are shown in Figure 6.3 (Levitus et al., 2009). Linear trends for 1969–2003 range from 0.24 to 0.41 times \(10^{22}\) J per year of additional heat (Ishii and Kimoto, 2009; Levitus et al., 2009; Domingues et al., 2008). Based on the higher estimate of Domingues

**FIGURE 6.3** (a) Global 0 m–700 m ocean heat content from three recent studies (Source: Levitus et al., 2009 © Copyright 2009 American Geophysical Union. Reproduced/modified by permission of American Geophysical Union.). (b) Breakdown of total sea-level rise into upper ocean (0 m–700 m, red) and deep (orange) thermal expansion, ice sheets (blue), and terrestrial storage (green). (c) Observed sea-level rise from two independent studies (black, yellow dashed) and satellite altimetry (red-dashed) compared to the sum of the contributions from the middle panel (blue). (Source: Panels (b) and (c) from Domingues, et al., 2008 (their figure 3). Reprinted by permission from Macmillan Publishers Ltd.)
et al. (2008), this corresponds to an additional globally-averaged air–sea heat flux of just under 0.4 W m\(^{-2}\) with about 90% of the additional heat stored in the upper 300 m. This additional heat corresponds to a sea-level rise of ~0.5 mm per year due to upper-ocean thermal expansion. Domingues et al. (2008) examine literature-based estimates for the various factors contributing to sea-level rise, including: (i) ~0.2 mm per year from Antarctica and Greenland since 1993 (with little data to constrain contributions prior to this time); (ii) ~0.5 mm per year from glaciers and ice caps for 1961–2003 (Dyurgerov et al., 2005), the importance of glaciers being dealt with in greater detail in Cogley (2012, this volume); (iii) terrestrial water storage, which lacks any long-term trend but exhibits large inter-annual variability; and (iv) penetration of heat into the deep ocean, particularly at high southern latitudes in regions of deep convective mixing (Rintoul, 2007; Johnson and Doney, 2006; Johnson et al., 2007, 2008). Figure 6.3b shows time series for these contributing components (Domingues et al., 2008). Despite recognized uncertainty in these terms, the observed estimate of total sea-level rise and the sum of independently estimated components (Figure 6.3c) show very similar trends ~1.6 mm per year (1961–2003), providing some confidence that the sea-level budget is approaching closure.

In addition to changes in global inventories of heat content, regional changes in temperature within different water masses are also apparent. Levitus et al. (2005), updated in Bindoff et al. (2007), reported on the zonally-integrated heat uptake into the top 1500 m of the global ocean. A number of interesting features are evident, and are discussed below.

In the Northern Hemisphere, there is a deep warming signal (centred at 40°N, Figure 6.4) associated with a warming of the North Atlantic subtropical gyre, the Atlantic current, and the Gulf Stream. In the subpolar region, there is an associated cooling and a strong freshening that affects the properties of Labrador Sea Water, an important component of the North Atlantic Deep Water that forms the Atlantic branch of the global overturning circulation. However, on inter-annual timescales, changes in water mass properties in the far North Atlantic are sensitive to the state of the NAO, which modulates the intensity of deep mixing and the circulation of the subpolar gyre (Sarafanov, 2009). Since the mid-1990s, the observed trends in salinity and temperature that took place over the preceding ~30 years and reported in Levitus et al. (2005), have largely reversed in tandem with the change in the phase of the NAO (Holliday et al., 2008). This highlights the difficulty in separating natural variability from anthropogenic factors.

In the Southern Hemisphere, a deep warming signal is evident, particularly in the vicinity of the Antarctic Circumpolar Current (Levitus et al., 2005; and Figure 6.4).

**FIGURE 6.4** (a) Linear trend in the zonally-averaged temperature for the world ocean (1955–2003, contour interval is 0.05°C per decade; dark red or blue shading indicates regions where the magnitude of warming or cooling exceeds 0.025°C, respectively (Source: modified from IPCC, 2007a and Levitus et al., 2005 © Copyright 2005 American Geophysical Union. Reproduced/modified by permission of American Geophysical Union.). Linear trend (1950–2008) in subsurface temperature at (b) 160°E and (c) 160°W using the new dataset of Durack and Wijffels (2010). (Source: Figure courtesy of Paul Durack. © American Meteorological Society. Reprinted with permission.)
This pattern of warming has been confirmed in more recent studies that include large amounts of new ARGO data (Gille, 2008; Böning et al., 2008). The mechanism driving these and other changes are still a matter of debate. As noted by Gille (2008), the pattern of change is consistent with a southward displacement of the ACC and the observed southward shift in the overlying westerly winds (Thompson and Solomon, 2002). The observed mid-latitude warming is also well-reproduced in the current coupled climate models. While the simulated response in the models is indeed driven by a southward shift of the ACC, the southward displacement is not purely adiabatic (i.e., an additional input of energy is required), and a large amount of heat is absorbed to the south of the ACC and advected northwards in the surface Ekman layer (Cai et al., 2010). However, as Cai et al. (2010) and others have pointed out, the lack of meso-scale eddy activity in climate models may undermine any conclusions with regards to the processes driving change in the ACC (see Section 6.2.5).

In the tropics, there is a distinctive cooling in the thermocline straddling the equator, sandwiched between strong surface and weaker deep warming (Figure 6.4a). This pattern is evident in both the Pacific and the Indian Oceans (Levitus et al., 2005) and appears to be related to a shoaling of the tropical thermocline associated with a weakening of the Walker circulation and the equatorial trade winds (Mcphaden and Zhang, 2004; Vecchi et al., 2006; Han et al., 2006). Over the last decade, there appears to have been a partial recovery associated with changes in the PDO (Mcphaden and Zhang, 2004). However, analysis that includes the most recent ocean observations and explicitly attempts to extract natural variability from the long-term trend (Durack and Wijffels, 2010) still shows a strong subsurface cooling along the equatorial Pacific (Figure 6.4b, 6.4c). Furthermore, the subsurface cooling is a robust feature of the coupled climate models that include anthropogenic forcing, suggesting that this pattern is indeed a fingerprint of human-induced change.

### Section 6.2.5. Ocean Circulation

A key role of the ocean is the redistribution of heat between the low and high latitudes. This is mediated by the large-scale circulation and the meso-scale activity that permeates the ocean. The large-scale circulation is often compartmentalized into an upper ocean circulation, including the expansive subtropical and subpolar gyres and the overturning (or thermohaline) circulation that includes some of the upper ocean pathways, but also comprises deep circulation fed by the sinking of dense high-latitude water. The circulation is controlled by a number of factors including: (i) fluxes of heat and freshwater to and from the ocean; (ii) wind forces, which are the major driver of upper ocean circulation and play an important role in ocean mixing; and (iii) tidal forces, which drive the daily tidal currents and also play an important role in ocean mixing. Significant long-term trends have been observed in both heat fluxes, freshwater fluxes, and wind fields in certain regions (Trenberth and Josey, 2007; Bindoff et al., 2007), leading to the expectation of changes to the ocean circulation.

A key metric that has been the focus of much scientific and public interest is the strength of the North Atlantic overturning circulation. The upper branch of this circulation brings large amounts of heat, via the Gulf Stream and Atlantic Current, from low latitudes to the North Atlantic. This plays an important role in warming the climate of northern Europe via heat lost to the atmosphere from the current as it travels northwards. Evidence also suggests that dramatic changes in the overturning circulation strength and location have played an important role in abrupt climate change (e.g., Clark et al., 2002; McManus et al., 2004). Model simulations also suggest that the warming and freshening in the North Atlantic, which would accompany anthropogenic climate change, could cause a significant slowdown of this circulation in the future (Meehl et al., 2007, their figure 10.15).

Considerable excitement accompanied the release of findings, based on five repeat hydrographic sections in the North Atlantic at 25°N (Bryden et al., 2005), which suggested a 30% slowdown in the overturning circulation between 1957 and 2004, a rate considerably larger than anything simulated by climate models. However, a slowdown of this magnitude should have left a strong cooling signature in SST (~1°C) in the North Atlantic, which was not observed (Kerr, 2005). Based on results from the RAPID continuous monitoring array (see Section 6.2.1), Cunningham et al. (2007) found a mean (±standard deviation) transport of −19 ± 6 × 10⁶m²s⁻¹ with a full range of 4.0 × 10⁶ to 34.9 × 10⁶m²s⁻¹. This degree of variability precludes drawing robust results from the few available hydrographic sections. Independent analysis, using an SST proxy for overturning strength, also finds strong coherence between the circulation and low-frequency NAO variability, but finds no evidence of a long-term trend (Latif et al., 2006; Latif and Park, 2012, this volume).

While the lack of pertinent observations precludes direct estimation, model experiments and indirect observational evidence suggest an increase in the transport of warm salty water from the Indian Ocean to the South Atlantic via the ‘Agulhas leakage’, another important part of the thermohaline circulation. Biastoch et al. (2009) use a realistically-forced eddy-resolving ocean model (in the vicinity of South Africa) nested in a global model (for 1958–2004) to simulate the circulation changes. The observed southward shift of the westerlies (Section 6.2.2.3)
drives a simulated southward shift of ~2° in the position of the Southern Hemisphere subtropical gyre. This, in turn, leads to a significant increase in the leakage of water into the Atlantic. The hindcast simulation also shows an increase in salinity of the Brazil current that is consistent with observations. A significant portion of this leaked water is not entrained back into the subtropical gyre and instead moves northward, leading to a 25% increase in the simulated transport of salt into the Northern Hemisphere. Biastoch et al. (2009) postulate that such changes could moderate future changes in the North Atlantic overturning circulation. The advection of the greater amounts of salty (Indian Ocean) water into the North Atlantic could act to partially offset any future freshening in the North Atlantic, potentially helping to stabilize the sinking branch of the overturning circulation (Weijer et al., 2002).

The mechanism behind this southward shift of the circulation relates to the southward intensification of the mid-latitude westerlies. While the latitude of maximum wind stress change sits well south of the subtropical gyres, it is the wind stress curl (in particular north–south gradients in the east–west wind) that, via Sverdrup dynamics, control the gyre circulations (Saenko et al., 2005; Cai et al., 2005). The maximum change in the wind stress curl occurs considerably further north over the subtropical gyres. Using historical wind observations, Cai (2006) computes the theoretical change in the depth-integrated ocean circulation. He finds a southward shift of the ‘supergyre’ that encompasses all the Southern Hemisphere subtropical gyres. Using a number of independent observational datasets including satellite altimetry, hydrography, and drifting floats, Roemmich et al. (2007) find a robust intensification of the South Pacific subtropical gyre (1993–2004) consistent with the wind stress changes. Altimetry also suggests an intensification of the gyres in the other basins. Using a unique 60-year time series (1944–2002) off southeastern Tasmania, Hill et al. (2008) show dramatic trends in temperature (2.23°C per century, well above global average increases) and salinity (0.34 psu per century) with both the decadal variability and the long-term trend consistent with changes in the wind field driving a southward intensification of the East Australian Current. The circulation and temperature change have also been implicated in important changes to local ecosystems. For example, it has permitted the invasion of the spiny sea urchin (*C. roderusii*) from mainland Australia to Tasmania, where overgrazing by the invading urchins has led to large localized species losses (Ling et al., 2009).

Theoretical arguments suggest that the southward shift in the maximum winds should also drive a southward shift in the location of the Antarctic Circumpolar Current (ACC) (Saenko et al., 2005, and references therein). Based on hydrographic and ARGO data, Sallée (2008) finds coherence between the SAM and the position of the fronts defining the ACC at certain longitudes, depending on the relative position of the mean westerlies, the ACC, and topographic constraints. Recently, Sokolov and Rintoul (2009) documented a zonally-averaged southward shift in the position of the ACC fronts (as determined by examining sea-surface height) of ~0.6° of latitude since the early 1990s. The southward shift is consistent with the observed southward displacement of density surfaces and large positive trends in temperature centred at ACC latitudes (Gille, 2008; Böning et al., 2008). While there is some observational evidence that supports a coherence between the Southern Annular Mode (see **Section 6.2.2.3**) and ACC transport through the Drake Passage on inter-annual timescales based on moorings (Meredith et al., 2004) and satellite gravity measurements (Böning et al., 2010), observations are insufficient to determine any long-term trends. The issue of ACC intensification and the associated spin-up of the Southern Ocean overturning is a matter of considerable contention as CMIP3 climate models and eddy-resolving ocean models produce different responses to intensified winds (Meredith and Hogg, 2006; Screen et al., 2009). In the climate models, increased westerly winds drive increased northward Ekman transport and an acceleration of the ACC, while in eddy-resolving models much of the additional energy input goes into an intensification of the eddy field, rather than into the mean circulation.

### 6.2.6. Oxygen

Dissolved oxygen plays a vital role for biological processes both directly through its effect on organism physiology and indirectly through its effect on the nitrogen and carbon cycles. Low oxygen conditions (hypoxia) can lead to significant die-offs in animal species (e.g., Chan et al., 2008) and result in extensive ‘dead zones’ in the ocean (Keeling et al., 2010). Patterns of oxygen concentrations also hold important clues for understanding changes in ocean biogeochemistry and circulation (e.g., Joos et al., 2003).

Dissolved oxygen concentrations are controlled by air–sea fluxes (which are strongly related to temperature–dependent solubility), ocean mixing (which helps to ventilate subsurface waters), internal advection, and biochemical processes, including primary production in the upper ocean and the bacterial breakdown of organic matter in the deep ocean. Recent attention has focused on the oxygen minimum zones, regions of poor ventilation, and/or high oxygen utilization that exist in the eastern tropical Pacific and Atlantic and the northern Indian and Pacific Oceans at intermediate depths. These regions can be hypoxic for certain species (i.e., low oxygen levels cause stress and ultimately death if maintained for extended periods). The threshold for hypoxia varies considerably depending on the species, although a relatively consistent lower bound exists (pO2 < 0.8 kPa) beyond which only species with specific...
low oxygen adaptation can survive (Seibel, 2011). While observations are much sparser than for physical variables, time series have been constructed that show a robust vertical expansion of the Atlantic and Pacific oxygen minimum zones (Stramma et al., 2008). These changes are consistent with model simulations with interactive ocean carbon cycles (e.g., Matear et al., 2000; Matear and Hirst, 2003). An extreme expansion in the tropical North Atlantic of 85% is observed since the 1960s, together with a decrease in minimum concentrations. Brandt et al. (2010) use extensive cruise data to both confirm the reduction in oxygen concentration (15 μmol kg⁻¹ between 1972–1985 and 1999–2008), and to propose a mechanism for the change. They show that ventilation of the tropical North Atlantic is from eastward-flowing jets of relatively high-oxygen concentration. Although no direct current measurements are available, hydrographic calculations suggest a weakening of the jets. While lack of observations precludes the determination of any trend in the northern Indian Ocean oxygen minimum zones, a time series has been constructed in the eastern tropical region, where no significant trend was found (Stramma et al., 2008).

Another region that has exhibited robust oxygen changes is the North Pacific, a vast region of relatively low concentration. Station data from off the Japan coast and in the Alaskan gyre, together with a number of repeat sections across the Northern Hemisphere basin, all show consistent declines in oxygen concentration across subpolar and eastern subtropical regions and in many marginal seas around Japan and western North America (Keeling et al., 2010, and references therein). Deutsch et al. (2006) use a hindcast, physical—biogeochemical model to show that the large changes in the lower subpolar thermocline are primarily due to weakened ventilation of dense water in the North Pacific. While significant fluctuations in biological export production are simulated, the influence of this on oxygen levels is confined to the upper thermocline. Earlier biogeochemical modelling suggests that, in general, thermocline concentrations are influenced primarily by solubility at the surface source region, while deeper oxygen decreases are more influenced by reductions in interior ventilation (Matear and Hirst, 2003).

Given the biological and oceanographic importance of oxygen, a concerted effort is under way to expand the observational network, including pilot deployments that include oxygen sensors on some of the ARGO floats. Initial data will be carefully compared with shipboard measurements, with large global deployments being subsequently planned (Gruber et al., 2007).

### 6.2.7. Carbon and Biogeochemistry

The ocean holds about 50 times the amount of carbon than the atmosphere and, therefore, has a profound influence on atmospheric CO₂ levels on both short and long timescales (cf. Harvey, 2012, this volume). The ocean has two different mechanisms by which it can affect atmospheric CO₂ via modulations in the surface-ocean CO₂ concentration and resulting changes in air–sea CO₂ fluxes. These will be discussed below.

#### 6.2.7.1. Anthropogenic Ocean CO₂ Budget

Detailed records since the start of the Industrial Revolution give us the ability to determine the amount of CO₂ emitted globally from anthropogenic fossil fuel use and land-use changes. In this time, about 244 ± 20 PgC has been emitted into the atmosphere (Sabine et al., 2004). In 2008, annual anthropogenic CO₂ emissions were 8.7 ± 0.5 PgC, which corresponds to a rate increase of 40% since 1990 (Le Quéré et al., 2009). Atmospheric CO₂ measuring stations are located across the world and have accurately determined the in-situ concentrations of CO₂ in the atmosphere over the last 60 years. Comparing the estimated concentration of CO₂ solely from fossil fuel emissions and the actual concentration in the atmosphere shows a large difference (Figure 6.5), with a shortfall in the atmospheric concentration of at least 55 ppmv. On average, only 43% of the total anthropogenic CO₂ emissions each year between 1959 and 2008 has remained in the atmosphere (Le Quéré et al., 2009). A large part of the ‘missing’ atmospheric CO₂ has been absorbed by the ocean (Sabine, 2004).

The ocean has a very large capacity to absorb atmospheric CO₂ due to both its volume and its ability to

![Figure 6.5](Image)

**Figure 6.5** Average annual atmospheric CO₂ concentrations (dotted line) and the atmospheric CO₂ concentrations that should be in the atmosphere solely due to fossil fuel emissions and land-use change (dashed line). Blue shading is the proportion of the anthropogenic CO₂ absorbed since the Industrial Revolution. Most of this CO₂ is absorbed by the ocean, with the terrestrial biosphere playing a minor role with regard to net changes over this timeframe (Source: Sabine et al., 2004.)
redistribute CO₂ into other inorganic forms of carbon. A wide variety of measuring programmes and methods have accurately determined the anthropogenic CO₂ sink (Sabine, 2004), which indicates that, since the Industrial Revolution, the ocean has absorbed about half of the total anthropogenic CO₂ emissions (120 ± 15 PgC up until 1995). More recent observational constraints suggest the oceanic sink was 2.2 ± 0.4 PgC per year between 1990 and 2000 (Keeling and Garcia, 2002; McNeil et al., 2003; Dickinson, 2012, this volume).

6.2.7.2. The Solubility CO₂ Pump

Oceanic CO₂ enters the ocean via air—sea gas exchange. The dissolved CO₂ then dissociates into bicarbonate (HCO₃⁻) and carbonate ions (CO₃²⁻) as shown in Equation (6.1). The vast majority of oceanic inorganic carbon exists as either bicarbonate (~90%) or carbonate (~9%), with CO₂ only 1% of the total inorganic carbon pool. The presence of inorganic carbon in three different forms means that the ocean has an immense capacity to alter oceanic CO₂ equilibration and to influence atmospheric CO₂ on long timescales. In fact, even if humans release all of the estimated 5000 PgC of current reserves of fossil fuel carbon (both conventional and unconventional), the ocean will eventually (2000–20,000 years) absorb between 65%–80% of all anthropogenic CO₂ after equilibration (Archer et al., 2009).

\[
CO_2 + H_2O \rightleftharpoons H_2CO_3 \rightleftharpoons HCO_3^- + H^+ \\
\downarrow \quad \Downarrow \quad \Updownarrow \quad \Updownarrow \\
CO_3^{2-} + 2H^+ \quad \text{(6.1)}
\]

The solubility of CO₂ in seawater is dependent on ocean temperature, with higher dissolution in cooler seawater (Weiss, 1974). If the oceanic CO₂ distribution were based on solubility alone, a quasi-linear relationship between the partial pressure of carbon dioxide (pCO₂) and SST would exist. Oceanic CO₂, however, is only weakly correlated with SST (Takahashi et al., 2002), which demonstrates the importance of other mechanisms in driving CO₂ variations in the ocean. The ocean overturning circulation is one of those important mechanisms for sequestering atmospheric CO₂. Surface waters exchange CO₂ with the atmosphere. As water is transferred from the surface ocean to the deep ocean via density transformation, there is also a transfer of large amounts of CO₂ that become sequestered (or isolated) from the atmosphere for long timescales (>1000 years). The rate of sequestration is therefore strongly tied to the rate of overturning. The transfer of CO₂ into the interior of the ocean via physical mixing of ocean waters is termed the ocean’s solubility carbon pump and has been an important contributor to atmospheric CO₂ variations in the past (Sigman and Boyle, 2000; Figure 6.6).

Recent observational evidence suggests that the efficacy of the physical pump may have changed over recent decades in certain regions. In the Southern Ocean, using a network of observational stations combined with a carbon model, Le Quéré et al. (2007, 2008) found a net reduction in CO₂ uptake since 1981. These authors suggest that the stronger overturning circulation, which might accompany the intensified winds (see Section 6.2.5), would allow more

anthropogenic CO$_2$ to be sequestered. At the same time, however, it would also bring to the surface deep water that has high concentrations of old natural CO$_2$. Le Quéré et al. (2007, 2008) find this latter process to be dominating, resulting in a net decrease in Southern Ocean CO$_2$ sequestration. However, this is at odds with some modelling studies (Law et al., 2008) and shows some sensitivity to the choice of stations (Zickfeld et al., 2008).

6.2.7.3. The Biological CO$_2$ Pump

The other ocean mechanism that can significantly alter the atmospheric CO$_2$ concentrations is the biological CO$_2$ pump (Figure 6.6). Via photosynthesis in the sunlit ‘euphotic’ zone, phytoplankton utilize CO$_2$ to form biomass. This primary productivity has the effect of reducing the concentration of CO$_2$ in the surface ocean and creates a concentration gradient, which causes an additional air to sea flux of CO$_2$ into the ocean. Primary production in the ocean fixes about 45 PgC per year of biomass, which is about eight times the annual anthropogenic CO$_2$ emissions (~8 PgC per year). However, 75% of the fixed organic carbon in the surface ocean is respired (~34 PgC per year) back into the atmosphere while the remaining ~11 PgC per year is biologically exported into the interior of the ocean (Laws et al., 2000). This biological carbon pump has been in quasi-steady state throughout the twentieth century and the first decade of the twenty-first century. Although biological productivity drives very large gross CO$_2$ fluxes across the air–sea interface, the steady-state fluxes have resulted in little influence on the net changes in atmospheric CO$_2$ during the industrial period (Matear and Hirst, 1999). The process that could potentially drive significant changes to atmospheric CO$_2$ is related to biological carbon export. When phytoplankton utilize CO$_2$ to create organic material, a portion of this organic material is grazed on by zooplankton and higher marine organisms. Following death or aggregate formation, this organic material sinks from the surface ocean into the deeper ocean. This process, known as carbon export, is an efficient way of sequestering CO$_2$ from the atmosphere. If the rate of carbon export is equal to the physical supply of carbon by upwelling or mixing, then there is no net change on atmospheric CO$_2$. Any change in the balance between the physical supply of carbon and the biological carbon pump could in principle drive large changes in atmospheric CO$_2$ concentrations. Variation in the strength of this pump is a critically important factor in controlling CO$_2$ variations between glacial and interglacial periods (Martin, 1990). We will discuss later how non-linear changes to the biological pump are starting to emerge as an important factor for future change.

Calcium carbonate (CaCO$_3$) is a mineral produced by many marine organisms (coccolithophores, foraminifera, pteropods, etc.). The production of particulate inorganic carbon (PIC) in the form of CaCO$_3$ via calcification has a smaller, but not insignificant, role to play in modulating atmospheric CO$_2$ on centennial timescales. Unlike photosynthesis, the process of CaCO$_3$ production actually acts to raise the surface-dissolved CO$_2$, and thus retards the sequestration of atmospheric CO$_2$. Of the 11 PgC per year of biological carbon export in the form of particulate organic carbon (POC), about 1 PgC per year is in the form of CaCO$_3$ or PIC. Although the biological carbonate pump does not play a significant role in the short-term fluctuations of atmospheric CO$_2$, carbonate processes and changes start to play an important role over multimillennial timescales, the timescale over which ocean/terrestrial weathering becomes important (Zondervan et al., 2001).

6.2.7.4. Ocean Acidification

The absorption of nearly half of the anthropogenic CO$_2$ load since the Industrial Revolution has resulted in chemical changes within the ocean, that pose potential threats to certain marine organisms (Fabry et al., 2008; Gattuso et al., 1998; Langdon et al., 2000). As anthropogenic CO$_2$ enters the ocean, the pH of the seawater decreases (i.e., acidity increases) and the surface ocean carbonate ion (CO$_3^{2-}$) concentration decreases (see Equation (6.1)). These changes, together known as ‘ocean acidification’, have direct consequences for the marine CaCO$_3$ cycle and those species that interact, exploit, and secrete the mineral. Ocean observations already show surface pH to be lowered by 0.1 (Raven et al., 2005; Figure 6.10). Palaeoproxy data suggests that current pH levels are more extreme than anything experienced over the glacial–interglacial cycles of the last ~2 million years, with levels projected for the end of the century likely to be unprecedented for ~40 million years. Moreover, the rates of change that are likely to occur during this century are probably two orders of magnitude greater than those experienced over the glacial–interglacial cycles (Pelejero et al., 2010). At these rates of change, the natural buffering mechanisms that can act to moderate pH changes on millennium timescales are unable to keep up. The role of CO$_2$ in the palaeorecord is discussed in more detail in Berger and Yin (2012, this volume).

Marine biological calcification requires adequate availability of seawater carbonate ions (Raven et al., 2005), which is often represented using the CaCO$_3$ saturation state of seawater: $\Omega_x = \frac{[Ca^{2+}][CO_3^{2-}]}{\lambda_x}$, where $\lambda_x$ is the solubility coefficient appropriate for the particular form of CaCO$_3$ under consideration (i.e., $x$ is either calcite or aragonite). Aragonite is the more soluble form of CaCO$_3$ and is secreted by many marine organisms. The decrease
in Ω from ocean acidification reduces the ability of some species (e.g., corals, calcifying algae, echinoderms, molluscs) to form their skeletal material and, therefore, future ocean acidification is expected to adversely affect these species and their associated communities and ecosystems (Raven et al., 2005). Experiments have shown that calcifying marine organisms are highly susceptible to changes in the aragonite saturation state (Ωarag) (Fabry et al., 2008; Langdon et al., 2000). Furthermore, direct dissolution of aragonite occurs when Ωarag < 1. The Southern and Arctic Oceans are most vulnerable to reaching the corrosive point for aragonite. Recent work suggests these regions will start to become corrosive for aragonite once atmospheric CO2 reaches 450–470 ppmv (McNeil and Matear, 2008; Steinacher et al., 2009). Despite the changes occurring within the ocean carbonate system, much uncertainty remains over the likely impacts to the marine ecosystem. Although some aragonitic species are likely to be detrimentally impacted by ocean acidification, along with those that graze on these species, other species that were predated on will likely ‘win’. The ecosystem-wide effects of ocean acidification are still uncertain.

### 6.2.8. Ocean Biology

Climate variability and change affect ocean ecosystems through a large variety of mechanisms (see review by Drinkwater et al., 2010). Temperature changes, for example, may affect physiology and growth rates, activity rates (e.g., swimming speed), reproductive fecundity, phenology, and species distribution. Surface temperature changes also drive changes in vertical stratification and mixing, which, in turn, affect the upwards entrainment of nutrients from the deeper ocean and light availability for phytoplankton at high latitudes. Changes in ocean circulation can affect dispersal of larvae and fish eggs and their subsequent recruitment (Drinkwater et al., 2010). Compounding these changes related to the physical environment are changes in ocean chemistry, which affect both calcification and physiology in a variety of marine species (Fry et al., 2008). Below, we examine some examples of biological changes that have been associated with long-term anthropogenic warming.

Phytoplankton form the basis for the vast majority of marine ecosystems and make up approximately half of the planet’s primary productivity (Field et al., 1998). They are sensitive to their physical and chemical environments and so are vulnerable to long-term climate change. They also represent an important feedback onto the climate system, modulating the absorption of energy in the upper ocean and affecting CO2 sequestration (e.g., Murtugudde et al., 2002; Sabine et al., 2004). Large-scale measurement of both the rate of phytoplankton production and the standing stock are generally achieved through the measurement of chlorophyll concentration. While measurements of chlorophyll have been taken since the start of the twentieth century (Jeffrey et al., 1997), synoptic global-scale measurements have only been available since the mid-1990s (Section 6.2.1).

Based on 9 years of satellite data (1998–2007), Polovina et al. (2008) find a dramatic increase in the area of the productivity minimum zones. These are regions of oceanic downwelling and poor nutrient supply, collocated with the subtropical gyres. They find area increases of between 0.8% and 4.3% per year along with corresponding increases in SST. These changes are consistent with increased stratification and a weakening of the nutrient supply from the deep ocean. Vantrepotte and Melin (2009) also note a significant reduction in productivity within the subtropical gyres, but find both increasing and decreasing trends in other locations. Behrenfeld et al. (2006) also find a strong negative relationship between SST and chlorophyll, but show that the relatively short satellite record is dominated by inter-annual variability, particularly associated with ENSO. Henson et al. (2010) demonstrate that the brevity of the data record precludes unequivocal attribution of satellite-derived change, which would require ~40 years of sustained observations.

The transparency of the water column can also be related to surface chlorophyll concentrations and can be derived from satellite, in situ optical measurements and Secchi disk measurements (Boyce et al., 2010). Secchi disks are simple black-and-white patterned disks that are lowered down through the water column until their pattern becomes indistinct to the surface observer. The associated depth can be converted to a chlorophyll estimate in low-turbidity open-ocean regions. Such measurements have been routinely made since the late nineteenth century. Boyce et al. (2010) have collated available transparency measurements to obtain centennial-scale estimates of primary productivity trends. Based on a set of climatically similar regions, the authors show that in 80% of these regions, productivity shows a long-term decline, with a globally-averaged rate of decline of ~1% per year. They also show a correspondence between productivity and SST, suggesting that stratification associated with global warming is a major driver in these changes (at mid and low latitudes). Considerable controversy surrounds this result however, as biases may be introduced when combining Secchi disk and in situ chlorophyll measurements (Mackas, 2011; Rykaczewski and Dunne, 2011; McQuatters-Gollop et al., 2011; Boyce et al., 2011).

A growing number of studies have found widespread poleward or downward shifts in the distribution of various marine species (Drinkwater et al., 2010, and references therein). Perry et al. (2005), for example, examined a large number of fish species in the North Atlantic and found that half of the species had moved significantly poleward over a 25-year period (independently of changes in fishing...
Another sensitive indicator of climate-related change is phenology — the seasonal timing of important events within a species lifecycle (e.g., migration, reproduction, feeding). However, the numbers of species for which sufficiently long time series are available are limited, with few data outside the Northern Hemisphere (and the North Atlantic in particular: Ji et al., 2010). Investigating a large number of planktonic species in the North Atlantic over the second half of the twentieth century, Edwards and Richardson (2004) find that larval release and development, which is related to ocean temperature, occurred significantly earlier, while the lifecycle of many diatom species, which form the nutritional basis for many of the higher animals and whose lifecycle timing is more closely tied to light availability, has remained relatively stable. This timing mismatch produces a reduction in the efficiency of energy transfer through the food chain and may have significant impact on regional fisheries (Edwards and Richardson, 2004). Focusing on zooplankton in the North Pacific, and species of pelagic fish and seabirds feeding on them, Mackas et al. (2007) also find shifts of up to a few weeks in zooplankton phenology between warm and cold years. They also postulate the strong likelihood of ecosystem disruption, as the phenology changes are species-specific. Temperature-related phenological changes have been observed in a number of other species, although these changes are usually related to inter-annual or regime shift changes (Drinkwater et al., 2010), with insufficient data to attribute changes to long-term anthropogenic warming.

Most of the evidence linking lowered pH with changes to biological systems is based on laboratory-based experiments (e.g., Fabry et al., 2008). However, a more limited number of studies provide direct in situ evidence. For example, Moy et al. (2009) find that planktonic foraminifera in the Southern Ocean have shell weights that are 30%—35% lighter than background Holocene levels, consistent with increased acidification. De’ath et al. (2009) show that a particular group of corals that span the Great Barrier Reef exhibit a 14% reduction in calcification since 1990 that is unprecedented for at least 400 years (based on the available coral proxy record). Again, they find this to be consistent with increased stress from both temperature and declining saturation state.

6.3. PROJECTIONS FOR THE FUTURE

As detailed in the preceding sections, the ocean has changed profoundly over the last few decades. While low-frequency natural variability plays an important role in these changes, human-induced changes are also apparent for many properties.

To obtain quantitative estimates of how the ocean might respond to further greenhouse emissions requires the use of coupled climate models. Most of these models, those that comprise the CMIP3 repository used for informing the IPCC AR4, lean strongly towards the physical climate system. At the core of a climate model lies the numerics to describe the dynamics and thermodynamics of the ocean and atmosphere. Forcing by GHGs in the atmosphere is prescribed and, as such, explicit feedbacks associated with changes in the carbon or other cycles and with ecosystem changes are precluded (Harvey, 2012, this volume). A smaller number of coupled climate—carbon cycle models are also available (comprising the Coupled Climate—Carbon Cycle Model Intercomparison Project C4MIP; Friedlingstein et al., 2006). While incorporating components of the ocean and terrestrial carbon cycles, the physical components of these models tend to be less sophisticated than the CMIP3 models and simulations are generally carried out at coarser resolutions. While carbon cycle feedbacks may only represent second-order corrections to future change, this may not be the case. In the C4MIP models, for instance, feedback effects associated with the land and the ocean will increase atmospheric CO₂ concentrations by between 20 ppmv to 200 ppmv (full inter-model spread) by 2100, equivalent to an additional 0.1°C to 1.5°C warming (Friedlingstein et al., 2006). Such feedbacks increase the likelihood of surprises—of tipping points or abrupt change (Smith et al., 2009; Lenton, 2012, this volume).

The IPCC AR4 (IPCC, 2007a) presents a thorough review of climate model-derived projections. While we cannot update all aspects of oceanic change dealt with in the IPCC report, we will review some of the latest work on a selection of key ocean regions and processes. This will be followed by an examination of some of the potential oceanic feedback mechanisms absent from the CMIP3 models that will be incorporated some of the models employed for simulations assessed in the future by the IPCC.

6.3.1. Tropical Pacific

Given that ENSO has such widespread importance, any changes to the tropical Pacific are of great interest. As such, the assessment of the fidelity of ENSO in coupled climate models and projected changes have received significant attention in the scientific literature (Collins et al., 2010, and references therein). These studies have generally found little agreement across the models with respect to how the strength or frequency of ENSO will evolve in the future (e.g., van Oldenborgh et al., 2005; Guilyardi, 2006). This is
because characteristics of ENSO events are associated with a number of processes that compete to either dampen, or enhance, ENSO variability. Yang and Zhang (2008), for example, used model experiments to demonstrate the competing effects of a reduced Walker circulation, which tends to weaken variability, and an increase in the equatorial stratification, which acts to enhance variability. While their model exhibited enhanced ENSO variability as a result of the dominance of the latter process, this does not hold true for all climate models. A number of additional feedback processes that affect ENSO strength are discussed by Collins et al. (2010). Using a long model simulation, Wittenberg et al. (2009) also suggest that a large natural inter-centennial modulation of ENSO could also make the isolation of any anthropogenic signal very hard.

Despite lack of clarity on the future evolution of ENSO, the models do show considerable agreement in their projections of changes to the mean state. Many of the projected oceanic changes stem from the robust simulated slowdown of the equatorial trade winds, which are, in turn, a consequence of the tropospheric warming (Held and Sodden, 2006). The warmer atmosphere is able to hold additional moisture; this moisture is precipitated out as the air rises over the Warm Pool in the western Pacific and in the convergence zones. A result of this, a robust projected western Pacific freshening, occurs across the models (Ganachaud et al., 2011). The rate of precipitation increase is, however, constrained by the rate at which the additional latent heat released by the condensing water can be radiated away and the rate of moisture increase exceeds the rate of precipitation increase. As a result, the upward transport of humid air must weaken. This drives a general slowdown of the atmospheric overturning circulation, including the surface trade winds (Held and Sodden, 2006; Vecchi et al., 2006). As the upper ocean equatorial circulation is primarily driven by local wind changes, this also causes a slowdown of the complex array of equatorial currents and counter-currents and the tropical overturning circulation (Ganachaud et al., 2011, and references therein). One exception to this is the Equatorial Undercurrent, a fast eastward flowing jet that sits between depths of 200 m and 400 m and plays an important role in transporting dissolved iron to the iron-limited eastern Pacific (Mackey et al., 2002). The undercurrent shows a robust projected increase in strength, driven primarily by an increased western boundary current input in the Southern Hemisphere (Luo et al., 2009; Ganachaud et al., 2011).

Most of the models show a tropical warming pattern that is largest near the equator (Figure 6.7), particularly in the central Pacific. Dinezio et al. (2009) show that the pattern of warming stems from a complex interaction of processes. This includes: (i) reduced poleward divergence

![Figure 6.7 CMIP3 multimodel mean linear trend (2000–2099) in SST (a) °C per 100 years and surface salinity (b) psu per 100 years for the ‘business as usual’ SRES A2 emissions scenario.](image-url)
of surface waters, which slows the removal of heat; (ii) cloud and evaporative processes in the west having a cooling effect; (iii) reduced westward flow that alters the lateral advection of heat; and (iv) altered vertical heat transport in the east, which causes a relative cooling as a result of increased thermal stratification, despite weakened upwelling. The causes for projected SST pattern formation have been extended to the wider extratropics (and other basins) by Xie et al. (2010). They show, for example, that the very weak simulated warming projected in the southeastern extratropical Pacific (Figure 6.7) is driven by an increase in the wind strength that allows extra latent heat to escape from the ocean.

Despite the weakening of the trade winds and the associated thermocline changes, it is a misnomer to describe the changes as El Niño-like, as is often done. In particular, the projected changes are not accompanied by a projected decrease in the zonal SST gradient, an important part of the Bjerknes feedback that characterizes ENSO variability. Consequently, the mechanisms driving the projected tropical Pacific changes, which are ultimately tied to long-term warming, are distinct from those that drive inter-annual ENSO variability (Vecchi and Soden, 2007; Collins et al., 2010).

Bell et al. (2011) provide a comprehensive review of the projected physical changes in the tropical Pacific and the effects these changes are likely to imply for biological systems and fisheries.

### 6.3.2. Southern Ocean

The Southern Ocean is a region where profound changes are occurring rapidly and major additional changes are projected. Many of these changes are intimately tied to possible climate feedbacks associated with the ocean carbon cycle. However, it is also a region where major uncertainties in the science and, therefore, in model projections are known to exist.

Assessment of the CMIP3 models shows qualitatively consistent projections for the ocean, despite sometimes large quantitative inter-model differences (Sen Gupta et al., 2009, and references therein). The changes are driven by a projected intensification of the hydrological cycle, a southward intensification of the westerlies (and the associated wind stress curl), and an increase in the radiative heat flux into the ocean, primarily southward of ∼50°S. The shift in the wind drives a southward migration and intensification of the Antarctic Circumpolar Current (ACC), with an associated increase in northward Ekman transport and increased upwelling of circumpolar deep water. The shift in the maximum wind stress curl occurs at lower latitudes over the subtropical gyres, causing them to shift southwards and, in some cases, intensify. The southward shift in the mid- and high-latitude lateral circulation is manifested as a large increase in the depth-integrated temperature between the southern limbs of the subtropical gyre and the core of the ACC (Sen Gupta et al., 2009). The projected shift in the circulation and enhanced deep mid-latitude warming is consistent with changes already observed over recent decades (see Section 6.2.4). Despite most of the extra heat entering the ocean at high latitudes, this heat is advected northwards and subducted (Cai et al., 2010). Hence, the high-latitude ocean is projected to warm at a substantially lower rate than that further to the north (Figure 6.7). Increases in high-latitude precipitation, together with increased Antarctic runoff, causes a significant freshening of the Southern Ocean (Figure 6.7). This high-latitude warming and freshening and the intensification of the westerlies is consistent with a simulated slowdown of Antarctic Bottom Water formation and subsequent northward export at abyssal depths (Sen Gupta et al., 2009; Saenko et al., 2011). Like Bottom Water, Mode and Intermediate Water formed at mid-latitudes in the Southern Ocean also play an important role in the removal of surface water into the ocean interior (subduction) and, therefore, the sequestration of anthropogenic CO₂. Projections suggest that, as the climate warms, the rate of subduction of intermediate water and denser mode waters will be reduced, while there is greater uncertainty with regard to the lighter Mode Waters (Downes et al., 2010; Saenko et al., 2011). As noted by Saenko et al. (2011), the role of ocean eddies (not included in the climate models) may affect these results.

The simulated wind-driven intensification of the ACC and associated spin-up of the overturning circulation (making up the Deacon circulation) in climate models has been the subject of considerable contention recently. High-resolution ocean models that can simulate small-scale ocean eddies (e.g., Meredith and Hogg, 2006; Screen et al., 2009; Spence et al., 2010) and some empirical evidence (Böning et al., 2008) suggests that the additional wind-derived energy from the strengthened westerlies would energize eddy activity. This would act to oppose the enhanced Ekman transport and so dampen any change in the strength of the circulation on long timescales. A second issue with the picture presented by the climate models relates to the projected trend in the westerly winds. Recent evidence, based on models with more realistic depictions of the stratosphere, suggests that the poleward intensification of the wind field may halt, or even reverse, at least during austral summer, as a result of the recovery of stratospheric ozone over Antarctica (Son et al., 2008). Only a few of the AR4 models incorporate any form of ozone recovery and those that do tend to underestimate the resulting modification in the atmospheric circulation compared to models with more realistic stratospheres (Son et al., 2008).

Each of the Southern Ocean water masses transports significant quantities of heat and dissolved gases into the
ocean interior. Independent modelling studies suggest that the Southern Ocean is the most important oceanic region for anthropogenic CO$_2$ uptake and the region likely to undergo most change due to global warming (Sarmiento and Le Quéré, 1996; Sarmiento et al., 1998; Matear and Hirst, 1999; Caldeira and Duffy, 2000). This conclusion was further supported by results from 13 different OGCMs in the Ocean Carbon Model Intercomparison Project (OCMIP; Orr, 2002). Figure 6.8 shows OCMIP-2 results. The Southern Ocean (south of 40°S) is important in all models, absorbing from 30%–50% of the total global anthropogenic CO$_2$ taken up by the ocean. Any changes in Southern Ocean uptake could therefore have a large effect on the global CO$_2$ budget.

6.3.3. Sea-Level

Based on climate model projections, the IPCC provides a range of values for globally-averaged sea-level rise of 18 cm to 59 cm over the next 100 years (depending on the choice of emissions scenario) with an additional 10 cm–20 cm for rapid dynamical changes in ice flow of the ice sheets (Meehl et al., 2007). There is an important proviso attached to these estimates, however. The CMIP3 models do not represent processes related to ice-sheet dynamics, while recent observational evidence suggests a growing importance for these processes (Rignot, 2006; Allison et al., 2009b). Furthermore, model estimates of sea-level rise tend to underestimate recent observations (Rahmstorf et al., 2007; Allison et al., 2009a).

A number of recent studies have attempted to provide an independent estimate of future sea level based on various semi-empirical techniques that use historical records to calibrate simple statistical models (Rahmstorf, 2007; Horton et al., 2008; Vermeer and Rahmstorf, 2009; Jevrejeva et al., 2010; Grinsted et al., 2009). These studies generally suggest a future sea-level rise with upper bounds that are larger than the IPCC values (for the same amount of globally-averaged warming; see Figure 6.9). One shortcoming of these statistical models is that they only take into account processes that are operating within the calibration period. For example, non-linear changes in ice flow that may occur at higher temperatures will not be accounted for (Rahmstorf, 2010). Using mass budget estimates, Pfeffer et al. (2008) consider the case of a rapid, but physically

![Figure 6.8](image1.png)

**FIGURE 6.8** Latitudinal variations in anthropogenic CO$_2$ uptake (a), and inventory (b) from 1765 to 1996 for models taking part in the Ocean Carbon Model Intercomparison Project. (Source: Orr, 2002, their figure 1.18.)

![Figure 6.9](image2.png)

**FIGURE 6.9** Estimates of projected sea-level rise from the IPCC (AR4), based on output from the CMIP3 climate models, semi-empirical sea-level models, and mass budget estimates (see text). (Source: Rahmstorf, 2010, his figure 1; all projections span the B1 to A1FI range of emissions scenarios. Reprinted by permission from Macmillan Publishers Ltd.)
plausible acceleration of ice-sheet melt to put an upper bound on possible sea-level rise of 2 m. Using more conservative rates of increase they still find an increase of 0.8 m, which is in excess of the IPCC estimates.

Recent research has focused on understanding the regional variations of future sea-level rise. Yin et al. (2009), for example, find that in a selection of CMIP3 models there is a particularly large projected sea-level increase off the northeast coast of North America (of the order of 20 cm above the global mean in the A2 scenario for 2100). Almost all the models project warming and freshening in the far North Atlantic that drives a slowdown in the overturning circulation (Meehl et al., 2007). The projected slowdown, characterized by the weakening of southward flowing North Atlantic Deep Water, is associated with a warming along the flow pathway. This, in turn, causes a thermally driven sea-level rise, which, together with a dynamical adjustment, produces the rapid projected sea-level rise response (Yin et al., 2009).

Timmerman et al. (2010) examine regional changes in the Tropical Pacific. Using a idealised ocean model forced by the wind projections from a selection of CMIP3 models, they were able to attribute most of the spatial variability in the projected sea-level rise pattern to purely wind-driven convergences and divergences in the surface circulation (i.e., Ekman pumping anomalies). These wind-driven adjustments cause regional deviations of up to 30% of the projected global mean sea-level rise (Timmerman et al., 2010).

Gravitational readjustment of the ocean due to changes in the mass of land-locked ice can also significantly affect regional sea level. This is particularly pertinent for the large masses associated with the Antarctic and Greenland ice sheets. The gravitational attraction of these ice sheets tends to draw water towards them. As the ice sheet melts, the attraction is reduced and there is a redistribution in sea level, together with a small adjustment in the Earth’s axis of rotation (Mitrovica et al., 2009). For Antarctica, mass-losses drive an absolute sea-level reduction within ~2000 km of the ice sheet, despite an increase in the global average. In addition, regions to the east and west of North America and sites bordering the Indian Ocean experience sea-level rise as much as 30% greater than would be expected from a uniform redistribution of water from the ice-sheet melt (Mitrovica et al., 2009).

6.4. OCEAN BIOGEOCHEMICAL FEEDBACKS

The current generation of climate models include many of the processes required to investigate physical feedbacks within the climate system (see Harvey, 2012, this volume). Primary among these are water vapour feedbacks, which play a dominant role in determining climate sensitivity (see Harvey, 2012, this volume). Ocean feedbacks associated with sea-ice also receive considerable attention given their role in the amplification (approximately twice the global mean) of surface temperatures over the Arctic Ocean and surrounding areas. However, there still appears to be considerable contention around the role of various feedback processes in driving the polar amplification. Hall et al. (2004) find that the sea-ice—albedo feedback, while having relatively little effect on inter-annual timescales, can substantially amplify high-latitude warming on longer timescales. However, based on output from a selection of CMIP3 models, Holland and Bitz (2003) describe other contributing mechanisms related to increases in ocean heat transport, changes in cloud cover, and the mean state of the ice thickness. Moreover, using CMIP3 output, Winton (2006) finds that, while ice—albedo feedbacks play a role, feedbacks associated with changes in cloud cover, longwave and shortwave radiation are equally important. Based
on an analysis of the observed seasonal variability of the polar amplification, Screen and Simmonds (2010) find that ocean heat losses associated with reduced ice cover have historically played a major part in polar temperature amplification in autumn and winter.

Another oceanic feedback process that receives considerable attention relates to the Atlantic overturning circulation and the associated poleward transport of heat. Changes to this circulation driven by density changes in the high-latitude North Atlantic (usually associated with increased freshwater input) have been implicated in rapid climate change in the palaeorecord (e.g., McManus et al., 2004). In the context of climate change over the next century, while most climate models project some slowdown of the overturning, none show a collapse of the circulation (Meehl et al., 2007) and any associated Northern Hemisphere cooling is more than offset by increased radiative forcing.

While representing many of the physical processes operating in the climate system, the CMIP3 models do not explicitly include climate feedbacks that involve changes in the carbon cycle: CO$_2$ (and other GHG) levels in the atmosphere are prescribed in the models, based on a set of future emissions scenarios (Nakicenovic et al., 2000). These feedbacks may substantially alter the efforts required to reach CO$_2$ stabilization. For example, based on the results of the C4MIP models, emissions may need to be cut by an additional 30% when trying to reach a 450 ppmv CO$_2$ stabilization, when the effects of changes in the land and ocean sinks are factored in (Friedlingstein et al., 2006; Friedlingstein, 2008). While the feedbacks associated with terrestrial processes are more important in the C4MIP models (Friedlingstein et al., 2006), large uncertainties in ocean carbon cycle processes still abound.

### 6.4.1. Solubility Carbon Pump

Given the ability of the ocean to modulate anthropogenic CO$_2$ in the atmosphere, it is fundamentally important to understand the ocean’s stability and the extent to which it can continue acting as a significant sink for anthropogenic CO$_2$. Climate change itself will impact physical oceanic processes in two main ways that will alter the capacity of the ocean to absorb anthropogenic CO$_2$. Firstly, business-as-usual emissions scenarios are projected to warm the upper ocean by up to 4°C by the end of the century, which reduces the solubility of CO$_2$. The warming effect on CO$_2$ solubility reduces the cumulative absorption by ~10%—14% over the next century (Matear and Hirst, 1999; Plattner et al., 2001; Sarmiento et al., 1998), equivalent to a ~20 ppmv increase in the atmosphere. Secondly, surface warming combined with regional freshening (at high latitudes and in tropical regions) is projected to cause enhanced stratification of the upper ocean, with impacts varying according to region. Stronger density gradients through the water column inhibit vertical mixing, impeding the transfer of CO$_2$ into the interior of the ocean where it can be sequestered on long timescales. This circulation feedback is projected to also slow the uptake of anthropogenic CO$_2$ by between 5% and 17% (Matear and Hirst, 1999; Plattner et al., 2001; Sarmiento et al., 1998; Friedlingstein et al., 2006). Both direct effects of ocean warming on solubility and the reduction in the rate of overturning are projected to slow the cumulative uptake of anthropogenic CO$_2$ by at least 15% by the end of this century, resulting in higher atmospheric CO$_2$ accumulation (Matear and Hirst, 1999; Plattner et al., 2001; Sarmiento et al., 1998). Despite our understanding of the stability of the oceanic carbon sink, many uncertainties remain, particularly with regard to model uncertainties and regional differences, particularly in the Southern Ocean.

### 6.4.2. The Biological Pump

The oceanic biological pump, as described earlier (the export of organic carbon from the euphotic zone into the ocean interior: Figure 6.6), is a fundamental process regulating atmospheric CO$_2$ (Martin, 1990; Volk and Hoffert, 1985). The biological carbon pump has remained close to a steady state in the ocean for at least the last 10,000 years prior to industrialization, since atmospheric CO$_2$ levels changed little over this period (Petit et al., 1999). However, future non-linear changes to the biological pump (via climate change or ocean acidification) may be associated with a significant future ocean carbon/climate feedback (Riebesell et al., 2009). Higher biological drawdown at the surface would lower CO$_2$ and allow the ocean to take up additional carbon from the atmosphere (and vice versa for lower biological drawdown). However, the magnitude and extent to which the biological carbon pump will respond to both higher CO$_2$ and climate change (nutrient supply) is still uncertain, although recent research has brought new insight into this evolving field of research.

#### 6.4.2.1. Traditional View of Biological Carbon Feedback

Nutrients and light are fundamental when considering the biological carbon pump, since photosynthetic growth at the surface is dependent on them. In most of the ocean, the amount of light available to phytoplankton will not be the limiting factor in the future, so most of the traditional views on the biological carbon pump have a nutrient-centred viewpoint. This comes about because, in vast areas of the surface ocean (e.g., the subtropical gyres), nutrient levels are very low, thereby limiting photosynthetic growth. In some regions, however, (such as the Southern Ocean) there are ample macronutrients. Here, light availability and micronutrients (such as iron) also limit growth. Inorganic carbon
has never been limiting in the ocean, so the traditional view of biological feedbacks ignores carbon as being important in the alteration of the ocean’s biological pump. In this view, the biological carbon pump is a balance between the upward supply of nutrients from the interior of the ocean and the downward export of carbon via POC. This balance can be altered by physical climate change since the upward supply of nutrients to the euphotic zone is likely to be significantly reduced under enhanced upper ocean stratification. Hence, stratification in some regions will reduce the biological carbon export. However, in some regions, where nutrient supply is not limiting, stratification may actually induce higher carbon export (for a finite period of time) (Matear and Hirst, 1999; Plattner et al., 2001; Sarmiento et al., 1998). This increase in carbon export is because warming may directly increase the rate of photosynthetic activity by 1%—8% (Sarmiento et al., 2004) and stratification creates more stable conditions for biological growth and export. Taken together, the traditional view of changes to biological carbon dynamics under global warming is nutrient-centred and will be regionally specific.

6.4.2.2. A Potential New Paradigm for the Biological Pump

A recent mesocosm experiment identified a potentially profound new paradigm for understanding ocean biological carbon feedbacks in a high CO$_2$ world. It was found that rising CO$_2$ levels may stimulate the biological carbon pump, thereby providing a negative feedback on future climate change (Riebesell et al., 2007). This study suggested that enhanced biological carbon consumption in a high CO$_2$ ocean would increase the ratio of carbon to nutrients (nitrogen and phosphorus) for the exported particulate organic matter (POM). This means that carbon would be acting as a limiting nutrient. This has significant ramifications for understanding future carbon feedbacks. Based on a simple box model, Riebesell et al. (2007) estimated that enhanced carbon export of POM with rising CO$_2$ would increase oceanic uptake of CO$_2$ from the atmosphere by between 74 PgC and 154 PgC by 2100 under the IS92a scenario from a present-day range of biological carbon export of between 8 PgC and 16 PgC per year, respectively. If correct, this would represent a large negative feedback on future atmospheric CO$_2$. For example, the positive feedback related to changes in the solubility pump due to ocean warming is ~50 PgC per year over the same period of time (Plattner et al., 2001). The biological carbon feedback could potentially offset these solubility-related changes threefold, thus enhancing the combined CO$_2$ drawdown by the carbon pumps over time.

Riebesell et al.’s (2007) estimate of the atmospheric drawdown of CO$_2$ uses a highly simplified box model approach. Many details of their ocean model are also missing, which makes it difficult to confirm their results. Furthermore, the physical resupply of carbon and nutrients from the deep and intermediate ocean to the surface ocean may not be accurately simulated using simple box models, as demonstrated by Oschlies et al. (2008). As with biological carbon pump perturbations using iron fertilization or macronutrient fertilization, the atmospheric uptake efficiency of the biological carbon pump is quite inefficient over large spatial-scales. This inefficiency dampens the impact of a stimulated biological pump associated with high atmospheric CO$_2$. It is important to emphasize, however, that the Riebesell et al. (2007) mesocosm is still an area of contention, since other mesocosm work on coccolithophore species Emiliania huxleyii did not find equivalent results (Iglesias-Rodriguez et al., 2008). Despite this, non-linear biological changes have the potential to dramatically alter the ocean’s role in modulating atmospheric CO$_2$ and will be the subject of many mesocosm and modelling experiments to better constrain the biological carbon pump.

There are other interactions between the marine carbon and nutrient cycles that have the potential to affect CO$_2$ feedbacks. These are reviewed in Matear et al. (2010). They find that feedbacks associated with the biological pump are an order of magnitude smaller in the ocean than on land on century-scale timescales, although they note that considerable uncertainty remains.

6.4.3. Ocean Acidification Feedbacks

Projections of future decreases in both pH and $\Omega$ (the CaCO$_3$ saturation state) due to higher atmospheric CO$_2$ levels have traditionally been obtained from ocean-only models (Caldeira and Wickett, 2003; Kleypas et al., 1999), although these modelling studies have not considered the effect of climate-change feedbacks on the carbon chemistry of the ocean. Recently, however, Orr et al. (2005) explored the significance that physical climate change (i.e., changes over and above those directly associated with CO$_2$ change) plays on the extent of ocean acidification. Using three separate climate models, they found that the physical components of climate change have little impact on the projected future decreases of pH, while having a significant impact on $\Omega$. That is, climate change has the potential to alter the surface-ocean carbon characteristics. As discussed earlier, the three main processes that will influence carbon feedbacks are circulation-driven changes; biologically-induced changes (both organic and inorganic); and ocean warming (direct warming effect and CO$_2$ solubility-driven effect). Recent work has found that pH and $\Omega_{\text{arag}}$ respond differently to these climate change feedbacks due to non-linearities in ocean CO$_2$ chemistry (McNeil and Matear, 2007). In simulations with physical climate change suppressed (i.e., with CO$_2$ increases only),
pH is projected to decline by ~0.3 (Figure 6.10). With physical climate change feedbacks also included, there is little additional impact on the projected changes because of non-linearities in seawater CO₂ chemistry (McNeil and Matear, 2007). For Ω_{arag}, with physical climate change suppressed, surface values are projected to decrease substantially across the ocean, although inclusion of climate-change feedbacks buffers the simulated decline by as much as 15% (Figure 6.10). This buffering is due to a temperature-driven chemical change in the surface ocean that causes increased CO₂ outgassing. The buffering of Ω_{arag} is most pronounced in regions that have a higher ability to absorb anthropogenic CO₂, that is, subtropical regions.

### 6.4.3.1. Calcium Carbonate Ballast Feedback

Chemical changes associated with ocean acidification (lowering pH and Ω) are expected to have the largest impact on the CaCO₃ cycle in the ocean. CaCO₃ is produced by a range of phytoplankton (e.g., coccolithophores and pteropods), marine organisms (e.g., molluscs, oysters), and coral reefs. Acidification causes a broad-scale reduction in calcification and an increase in the dissolution of the CaCO₃ mineral (see Section 6.2.7.4). Klaas and Archer (2002) found that CaCO₃ may act as a ballast for sinking POC export. The POC is protected within the sinking CaCO₃ shells of microorganisms, thereby transporting organic matter to much greater depths. Reduced calcification or enhanced dissolution of CaCO₃ in the surface ocean, via ocean acidification, could therefore inhibit the organic matter flux into the deeper ocean interior, acting as a large positive feedback on atmospheric CO₂ (Heinze, 2004; Riebesell et al., 2009). The importance of this CaCO₃ feedback is most likely to be regionally and species-specific. For example, corrosive conditions for aragonite are expected to occur seasonally when atmospheric CO₂ reaches 450 ppmv in the Southern Ocean (McNeil and Matear, 2008). Although this ballasting effect would probably begin in the Southern Ocean, the biological proportion of CaCO₃ to POC is very small in comparison to the subtropical oceans, so the effect is likely to be small. The magnitude and spatial effects of this ballasting feedback is therefore poorly understood, as is our understanding of the global feedback related to CaCO₃ ballasting changes (Riebesell et al., 2009).

### 6.4.4. Other Climate Feedbacks

In addition to carbon-cycle–related feedbacks, other chemical cycles may be influenced by, and in turn influence, future climate change. For example, nitrous oxide, another important GHG, is produced in the bacterial breakdown of organic matter with air–sea fluxes constituting an important source to the atmosphere (Suntharalingam and Sarmiento, 2000). Interestingly, model studies suggest that any benefits related to the increased sequestration of CO₂ resulting from iron fertilization initiatives may be offset by increased remineralization and release of nitrous oxide to the atmosphere (Jin and Gruber, 2003). This result could presumably apply to any increase in ocean productivity.

The ocean also releases aerosols, the rate of which is sensitive to the physical and biological environment. Sea-salt aerosol, for example, affects cloud formation and rainfall and is sensitive to factors including wind speed and surface temperature. Recent model experiments indicate that, in the Arctic, as sea-ice extent reduces, there is an increase in sea-salt aerosol flux. This in turn causes an increase in cloud albedo and an associated reduction in incoming radiation (Struthers et al., 2010). As such, the air−salt−sea-ice interaction may constitute a negative feedback to Arctic warming.

Dimethylsulfide (DMS) produced primarily by phytoplankton emissions is a major contributor to atmospheric aerosols (~43% of global sulfate aerosols: Chin and Jacob, 1996). Changes in DMS production are related to multiple physical and biological factors and, as such, both negative and positive climate feedbacks have been suggested, as reviewed in Rice and Henderson-Sellers (2012, this volume). Despite considerable research, the effect of DMS on climate remains uncertain (Ayers and Cainey, 2007; Carslaw et al., 2009).

### 6.5. OCEANIC VARIABILITY AND CHANGE

#### 6.5.1. Oceans and the Future Climate

The ocean changes on a wide range of timescales, with certain natural climate cycles varying over periods of multiple decades or longer. Nevertheless, we are now able to identify changes to the ocean that are not consistent with natural variability. Figure 6.11 presents a number of robust changes that have been attributed to anthropogenic influences via increases in atmospheric CO₂ and other GHGs (including ozone). Robust physical changes include: (i) a surface-intensified warming across most of the global ocean with an associated increase in stratification and ocean heat content; (ii) broad-scale coherent trends in surface salinity, associated with an intensification of the hydrological cycle; (iii) a dramatic reduction in Arctic sea-ice extent (and probably thickness); and (iv) an acceleration in global sea-level rise, driven by both thermal expansion and land-ice melt. In addition, there is some evidence to suggest modified water mass properties and pathways, an acceleration of ice melt from both Greenland and the Antarctic, together with modified circulation, associated with large-scale wind changes.
Combined with these physical changes has been a robust increase in dissolved CO$_2$ concentration in the upper ocean and an associated lowering of pH and CaCO$_3$ saturation. The observational record of chemical and biological change is considerably more limited than for many of the physical variables. For example, worldwide, only a handful of sites have continuous biogeochemical monitoring that starts before 1990 (Chavez et al., 2011). Maintaining and expanding our current observational network, together with biogeochemical process studies at key locations, must be a priority if we are to adequately understand the biological and chemical changes to the ocean and more accurately project the future evolution of our climate system. Nevertheless, evidence is growing for a number of other long-term changes to the ocean. These include: (i) long-term decreases in primary productivity and an expansion of the low-productivity subtropical gyre zones; (ii) reduced calcification in certain phytoplankton and corals (in certain regions); (iii) regional expansion of the oxygen minimum zones; (iv) poleward shifts in the location of some species; and (v) physiological and phenological changes in some species (Figure 6.11).

### 6.5.2. Future Unknowns

Carbon cycle feedbacks have played an important role in mitigating against an increased greenhouse effect in the past. However, evidence suggests that the efficacy of these feedbacks is changing and may change more dramatically in the future as the climate system is increasingly perturbed (Le Quéré et al., 2007; Friedlingstein, 2008). Marine carbon cycle feedbacks can be examined in terms of various carbon ‘pumps’. Physical ocean changes affect the solubility pump and the transfer via ocean overturning, of CO$_2$ saturated surface water, into the deep ocean. Warming-related responses to the solubility pump stem from a number of factors including the reduced ability for warmer water to absorb CO$_2$; and increased stratification, which acts to isolate the surface and deep ocean. At the same time, regional wind-driven changes can generate enhanced overturning, increasing the transport of old, natural CO$_2$ from the deep to the surface ocean. Some evidence suggests that the wind-driven changes appear to be dominating in the Southern Ocean (Le Quéré et al., 2007), although considerable uncertainty exists with regards to future projections. While there is little evidence that the biological pump has changed significantly over recent decades, carbon cycle models suggest that changes to this pump may be more important in the future. Physical and chemical changes to ocean productivity would alter the export of carbon from the surface into the isolated ocean interior. Complexities related to how biological processes respond, especially in light of increased ocean acidification, and changes in nutrient cycles and to the intriguing possibility that CO$_2$ may actually have a fertilizing effect add additional uncertainty.

Based on the CMIP3 climate models, all of the physical changes (discussed above) are projected to intensify substantially over the coming century. However, we know...
that many important processes within the models are poorly represented or entirely absent, adding uncertainty to any quantitative assessment of change. In particular, carbon cycle and other biogeochemical feedbacks are excluded from the CMIP3 models. Some of these uncertainties will hopefully be resolved in the next generation of models (taking part in CMIP5) that will provide input to the IPCC AR5.

Although at the time of writing there are few new results available for CMIP5, the new simulations are likely to be a major boost for understanding the future response of the climate system, including carbon cycle feedbacks. While peer-reviewed assessments are not yet available, there are anecdotal reports of significant improvements in the fidelity of many of the models in reproducing the current climate, for example in the representation of ENSO (a major failing in the CMIP3 models; e.g., Collins et al., 2010). The number of climate models will expand from ~24 to ~40 (from CMIP3 to CMIP5\(^1\)) and the experiments being run will diverge considerably from those undertaken as part of CMIP2 and CMIP3. In particular the SRES pathways (Nakicenovic et al., 2000) that form the basis for previous projections have been replaced by a set of Representative Concentration Pathways. These cover both unmitigated and alternative mitigation scenarios (Figure 6.12) and are based on a better understanding of recent environmental changes, economics, and emerging technologies (Moss et al., 2010). In addition to all forcing experiments, some modelling groups are conducting single forcing runs to identify the impacts of given climate forcings (e.g., CO\(_2\) only, anthropogenic aerosol only). A new core set of experiments will also examine decadal climate forecasts where climate simulations are initialized from observed ocean and ice conditions. The long-term persistence of anomalies within the ocean and ice systems has the potential to provide additional skill when producing decadal projections.

Some of the new models will now also include chemical and biological components, for both the ocean and the land, which will provide new insights into feedback processes absent in the CMIP3 models (CLIVAR, 2009; Taylor et al., 2009). Historical simulations and future projections (based on the high emissions RCP8.5: Figure 6.12) will be supplemented by experiments where: (i) the physical effects of climate change are suppressed (i.e., changes due to CO\(_2\) increase alone are isolated); and (ii) warming occurs in response to increasing CO\(_2\), but the CO\(_2\) change is ’hidden’ from the carbon cycle model components (Taylor et al., 2009). These experiments should provide more quantitative information regarding the strength of certain carbon cycle feedbacks and, as a result, provide more credible estimates of future change.

**ACKNOWLEDGEMENTS**

We would like to thank the four scientific reviewers, Richard Matear, Nathan Bindoff, Martin Heimann, and one anonymous reviewer for their valuable comments.

---

1. WCRP has opted to skip CMIP4 to reduce confusion with C4MIP and to mesh the numbering of the IPCC assessments and its model inter-comparisons.