Chapter 2

An introductory view of the ocean

Many aspects of the ocean are challenging to understand, including how the ocean stores and redistributes heat over the globe, how life has colonised the ocean and how carbon is cycled between the ocean, atmosphere and land. To answer these questions, we need to adopt a holistic view to understand the relevant physical, chemical and biological processes, and how they are connected to each other. For example, a western boundary current, like the Gulf Stream, has a range of signatures: a rapid transfer of heat, nutrients and carbon along the current, enhanced contrasts in physical, chemical and biological properties across the current, and increased exchanges of heat, moisture and dissolved gases with the overlying atmosphere.

A difficulty in understanding the ocean, as compared to the atmosphere, is the problem of taking observations due to the ocean being a more inaccessible and hostile environment. To counter that difficulty, a range of different techniques have been developed to unravel how the ocean circulates, drawing on a combination of ship-based measurements, remote sensing from space and freely drifting floats spreading throughout the ocean.

In this chapter, we provide an introductory view of the large-scale ocean circulation, and basic property distributions (Plates 2 to 7), as well as briefly introduce the atmospheric circulation, and then discuss how life flourishes in the ocean and, finally, how carbon is cycled in the ocean. This material provides a starting point for the rest of the book.

2.1 Ocean circulation

Water has a very low viscosity compared with most other liquids and moves easily whenever there is forcing, rather than moving sluggishly like it does in a soggy marsh. The ocean circulates due to a combination of mechanical and density forcing:

- Air moving over the surface water provides a frictional drag, depending on the difference in the speed of both fluids. Since the atmosphere moves so much faster than the ocean, there is a transfer of momentum into the ocean, ultimately driving most of the currents in the upper ocean.
- Variations in the gravitational acceleration across the Earth from the Moon lead to bulges of water forming around the Earth, orientated towards and away from the Moon. As the Earth rotates about its own axis, these bulges appear as a regular rise and fall of sea level at the coast, occurring typically once or twice a lunar day (24 hours 50 minutes). There are similar, albeit weaker, tidal oscillations formed by the gravitational forcing from the Sun. While tides are most pronounced near the coast, they occur throughout the water column and can lead to enhanced mixing when the tidal flows interact with rough topography.
- Surface density is altered through the exchange of heat and fresh water with the atmosphere; increasing with surface cooling or evaporation.

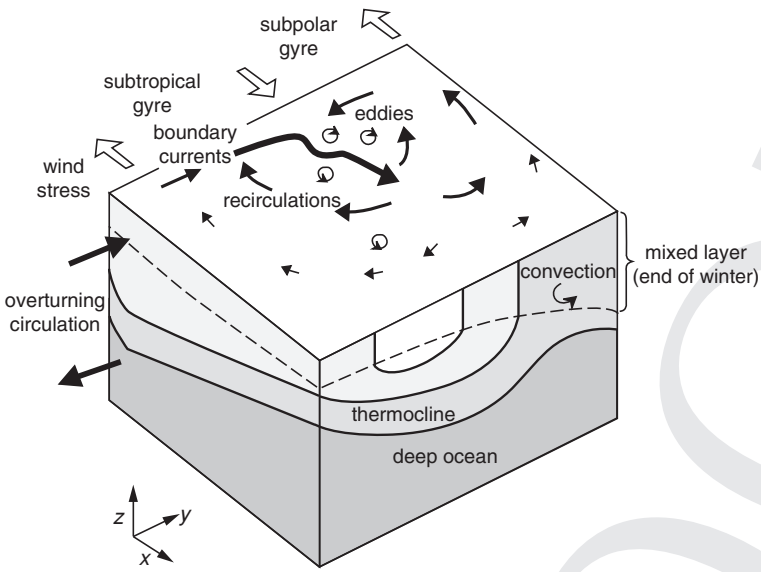

Figure 2.1 A schematic view of the circulation within an ocean basin, including recirculating gyres and overturning. Surface winds induce the gyres containing western boundary currents, interior flows and eddies. The horizontal circulation is connected with sloping temperature and density surfaces (full line) in the interior. In the vertical, there is a surface mixed layer (dashed line) where convection occurs, together with a thermocline, where there are strong vertical temperature gradients, and deep waters with more uniform properties.

In polar regions, density also increases through cooling by any overlying ice or by the atmosphere over gaps in the ice, as well by the addition of salt released when ice freezes. Whenever surface waters become denser than deeper waters, convection occurs and the fluid vertically overturns. Density contrasts lead to horizontal pressure gradients, which then accelerate the flow. Dense waters formed at high latitudes can sink to great depths and spread horizontally over the ocean, replaced at the surface by lighter waters, which then generate an overturning circulation.

With the exception of the tidal oscillations, the ocean moves relatively slowly, so that water takes many days to cross a basin, leading to currents being strongly constrained by the Earth's rotation; this rotational constraint leads to ocean flows persisting, rather than being quickly dissipated (as instead occurs for water sloshing from side to side in a bath).

The pattern of ocean circulation is also affected by topography, since deep flows move around any physical barriers and follow the pattern of channels and gaps in the topography. Surface currents are also sometimes influenced by the underlying topography, since currents often prefer to move around a submerged bump, rather than move directly over the bump.

The combination of forcing, rotation and topography leads to an intriguing range of physical phenomena, as illustrated in Fig. 2.1: there are strong boundary currents alongside some land boundaries, but not others; basin-scale circulations confined within continental barriers; overturning circulations, including deep boundary currents; and vibrant, time-varying eddies occurring throughout the ocean.

Next we consider observational views of the surface circulation and their connection to the underlying temperature and density distributions.

2.1.1 How does the surface flow vary?

Beneath a thin surface boundary layer, ocean currents follow the dynamic height of the sea surface in the same manner that air moves along pressure contours, as depicted on a weather chart; the dynamic height is the displacement of any pressure surface from a reference surface, called the geoid. A glance at a global map of dynamic height, as shown in Fig. 2.2, reveals several dominant, large-scale regimes:

- The ocean circulates on a horizontal scale of several thousand kilometres between the continents; these basin-scale circulations are called gyres (Fig. 2.3a). These gyre circulations rotate in a clockwise manner in the subtropics of

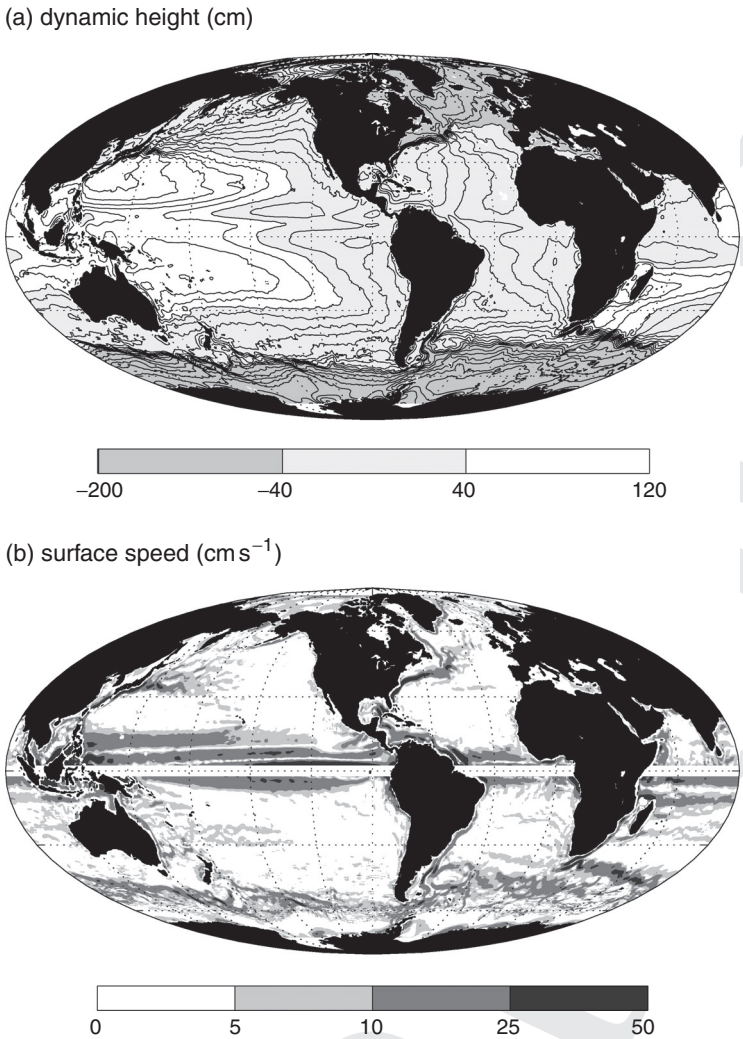

Figure 2.2 Different regimes for the near-surface circulation (excluding a frictional contribution from the wind) revealed in (a) mean dynamic height (contours with an interval of 10 cm and increased to 20 cm over the Southern Ocean) and (b) surface speed (cm s^{-1}) diagnosed from the gradient in mean dynamic height (assuming geostrophic balance with values excluded within 2° of the equator where this balance does not apply; see later Section 4.2.1). The dynamic height is the displacement of a pressure surface from a reference surface, called the geoid. Below a thin surface boundary layer, the surface flow is directed approximately along contours of dynamic height. There are large-scale recirculations within basins, referred to as gyres, including intense western, boundary currents, over much of the globe. In the Southern Ocean, the surface flow is directed broadly eastward and circumnavigates the globe, and is referred to as the Antarctic Circumpolar Current. The dynamic height was taken from a combined dataset integrating information from surface drifters in the ocean, remotely sensed measurements of sea-surface height from altimetry, surface winds and measurements of the geoid from the GRACE gravity mission (Niiler *et al.*, 2003; Maximenko and Niiler, 2005).

the northern hemisphere and anticlockwise in the subtropics of the southern hemisphere. The gyre circulations rotate in the opposite sense at higher latitudes. This change in rotation of the gyres reflects the pattern of the overlying winds. The rotation of the flow is defined as cyclonic whenever the rotation is in the same sense as the rotation of the Earth, anticlockwise in the northern hemisphere and clockwise in the southern hemisphere.

- Accompanying these gyre circulations are vigorous and narrow boundary currents along the western sides of the ocean basins, such as the Gulf Stream in the North Atlantic, the Brazil Current in the South Atlantic, and the Kuroshio in the North Pacific (Fig. 2.2b).
- In the tropics, there is an equatorial current system consisting of reversing, zonal flows with high speeds. Over the tropics, currents are forced by the prevailing westward winds: on either side of the equator, there are two westward flowing equatorial currents, together with a narrow eastward return flow in an equatorial counter current located between them; note that the speed is not mapped within 2° of the equator in Fig. 2.2b.
- Within the southern hemisphere, there are gaps between Antarctica and the northern land masses, allowing an uninterrupted flow to circumnavigate the globe roughly along latitude circles (Fig. 2.3b). This eastward current, the Antarctic Circumpolar Current, extends from

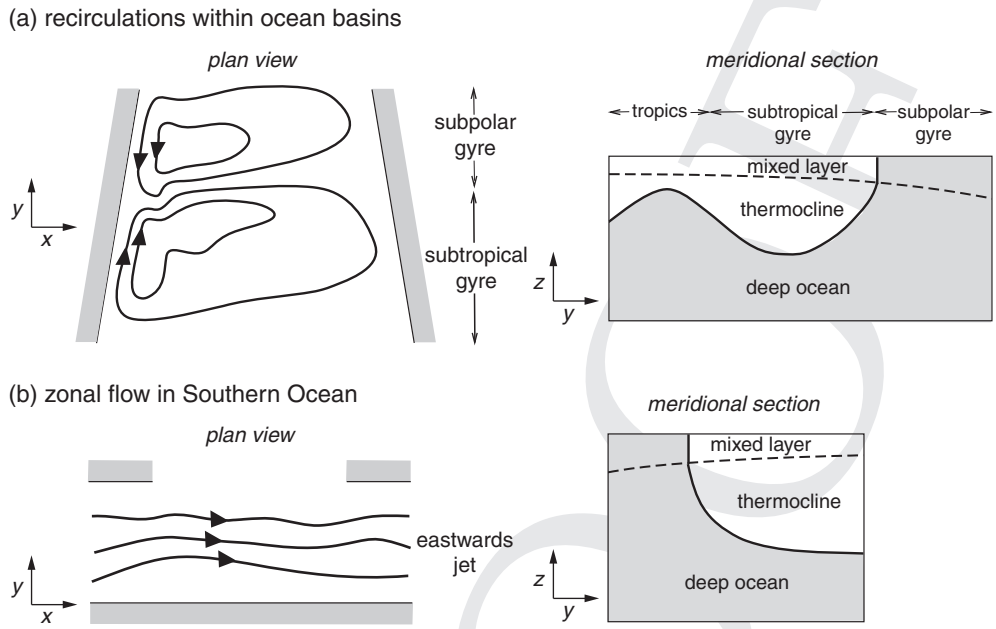

Figure 2.3 A schematic figure depicting (a) recirculating flows within ocean basins in the northern hemisphere and (b) near zonal flows within the Southern Ocean for a plan view (left panel) and a meridional section (right panel), including the base of the mixed layer (dashed line). In (a), the circulations are associated with a thin thermocline in the tropics and a thicker thermocline in the subtropical gyre, and a thin or seasonal thermocline over the subpolar gyre. In (b), an eastward wind-driven current in the Southern Ocean is associated with cold water outcropping on the poleward side of the current. On the equatorial side of the Southern Ocean, the near-zonal flows can interact with the gyre circulations in the Atlantic, Indian and Pacific basins.

the surface to the sea floor and is made up of a series of narrow jets (regions of fast-moving flow). The Antarctic Circumpolar Current is analogous to the fast moving Jet Stream in the upper atmosphere.

In addition to this network of persistent currents covering the globe, there are also intense time-varying circulations on smaller scales of several tens to a hundred kilometres, referred to as mesoscale eddies; although much smaller in size, these ocean eddies are analogous to atmospheric weather systems. Ocean eddies are preferentially formed along intense western boundary currents, such as the Gulf Stream, and within the jets making up the Antarctic Circumpolar Current.

2.1.2 How does the ocean vary in the vertical?

The ocean has a characteristic vertical structure (Fig. 2.4a) consisting of a surface mixed layer, overlying a stratified upper ocean and a weakly stratified, deep ocean:

- The surface layer is well mixed with nearly homogeneous properties in the vertical. This mixed layer varies in thickness from typically 30 m to 500 m, as depicted in Fig. 2.4b–e. The surface layer is in direct contact with the atmosphere or, in polar regions, with overlying ice. The mixing in this layer is driven by turbulence generated by the wind blowing over the sea surface and by convection from the sinking of dense waters formed by surface cooling or evaporation to the atmosphere, or by brine release from underneath ice.
- In the upper ocean, there is usually a strong vertical temperature gradient, referred to as the thermocline, which is typically accompanied by similar vertical gradients in salinity and density, such as seen extending from 500 m to 1500 m in Fig. 2.4. This strong stratification inhibits vertical mixing below the surface mixed layer. The thermocline can be separated into an upper, seasonal thermocline and a deeper, permanent or main thermocline. The seasonal thermocline strengthens from the

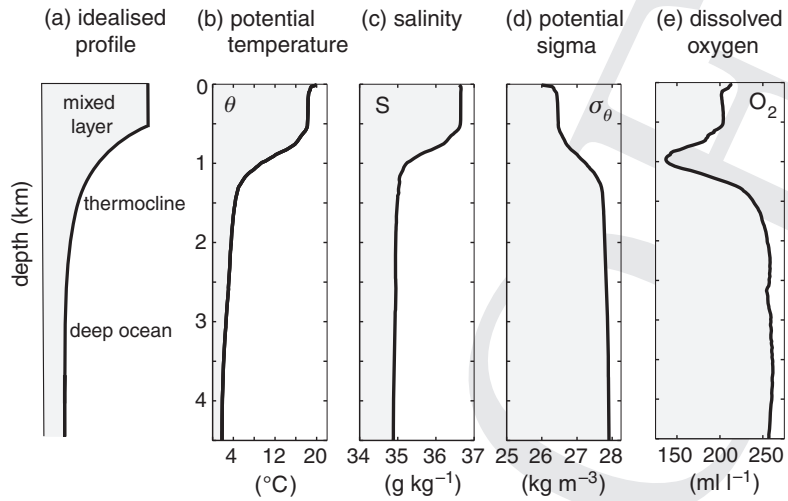

Figure 2.4 (a) A schematic vertical profile identifying the mixed layer, thermocline and deep ocean, together with actual profiles to the east of the Gulf Stream (74.28°W, 35.65°N) in May 2005, (b) potential temperature, θ (°C), (c) salinity, the concentration of salt (g kg^{-1}), (d) potential density minus 1000 kg m^{-3}, σ_θ and (e) dissolved oxygen (ml l^{-1}) versus depth (km). The mixed layer is very thin in summer, perhaps less than 30 m, and there is a weak seasonal thermocline over the upper 100 m. The mixed layer has previously been thick in winter, as suggested by the relatively uniform properties from 100 m to 500 m; in this case, with temperatures of 18 °C and high oxygen content. Below this seasonal boundary layer, there is a permanent thermocline, halocline and pycnocline with strong vertical gradients, extending from 500 m to 1200 m, together with an oxygen minimum zone. Below, in the deep ocean, there are relatively uniform physical properties and higher oxygen concentrations in the colder waters.

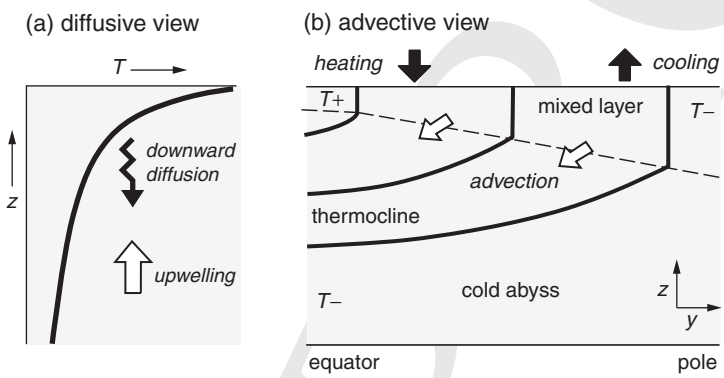

Figure 2.5 A schematic figure showing different views as to how the thermocline is sustained: (a) a diffusive view where there is a downward diffusion of heat (curly arrow) competing with an upward advection of cold, deep water (white arrow), and (b) an advective view where the thermocline is formed by the tilting of the surface temperature contrast into the vertical, involving both vertical and horizontal advection (white arrows).

warming by the penetrating rays of the Sun during summer, but weakens and becomes eroded by surface cooling during winter. The rest of the thermocline can persist over the entire year and there is no direct heating at these depths.

- In the deep ocean, there are relatively weak vertical gradients in temperature and density, such as seen below 1500 m in Fig. 2.4.

Why is there a thermocline?

In many ways, one of the most surprising features in the ocean is the thermocline, the region of warm waters with a large vertical temperature gradient. The persistence of the thermocline over the ocean is thought-provoking given the tendency of diffusion to erode property gradients and make tracers more uniform.

Traditionally, the thermocline has been explained in terms of a balance between the upward movement of cold water and downward diffusion of heat (Fig. 2.5a). However, there is little observational support for this view, since upwelling of cold water is too small to be measured and the vertical diffusivity turns out to be very small within the thermocline.

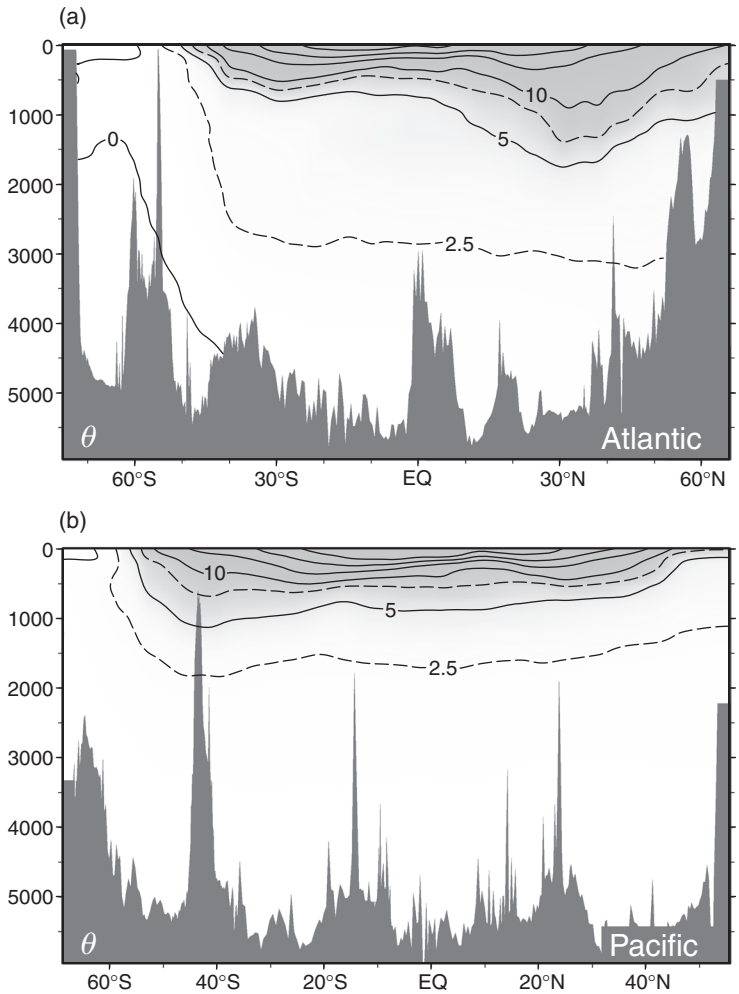

Figure 2.6 Observed meridional sections, latitude versus depth (m), of potential temperature, θ (°C), for (a) the Atlantic along 20°W, and (b) the Pacific along 170°W; see Plates 2a and 3a for colour and cruise track. There is generally a warm upper layer, referred to as the thermocline, overlying cold deep waters. The cold waters connect to the surface over the Southern Ocean. The sea floor is shaded in dark grey, revealing ridges and fracture zones. This section and many others are plotted from hydrographic data using Ocean Data View developed by Reiner Schlitzer.

Instead, the thermocline is a consequence of the three-dimensional circulation, effectively the surface temperature gradient is tilted into the vertical by the vertical and horizontal flows (Fig. 2.5b).

Over the globe, the thermocline is shallow in the tropics, deep at the mid latitudes, and eventually disappears at high latitudes (Fig. 2.6). This thermocline variation reflects the pattern of the atmospheric forcing and the resulting horizontal circulation. At mid latitudes, a thickening of the thermocline is associated with the subtropical gyres rotating anticyclonically (Fig. 2.3a), while at high latitudes, a poleward thinning and eventual vanishing of the thermocline is associated with the subpolar gyres rotating cyclonically and the strong eastward flow in the Antarctic Circumpolar Current (Fig. 2.3b); this dynamical connection is discussed further in Sections 8.1.2, 8.4.1 and 10.2.3.

2.1.3 How do water masses spread over the globe?

Inferring the interior circulation is more challenging than might be expected due to the difficulty of measuring the flow below the surface. Clues as to the interior circulation are provided by the distribution of water masses, defined in terms of a collection of physical and biogeochemical properties, including temperature, salinity, dissolved oxygen and nutrient concentrations.

Water masses are formed in the surface mixed layer where their properties are determined

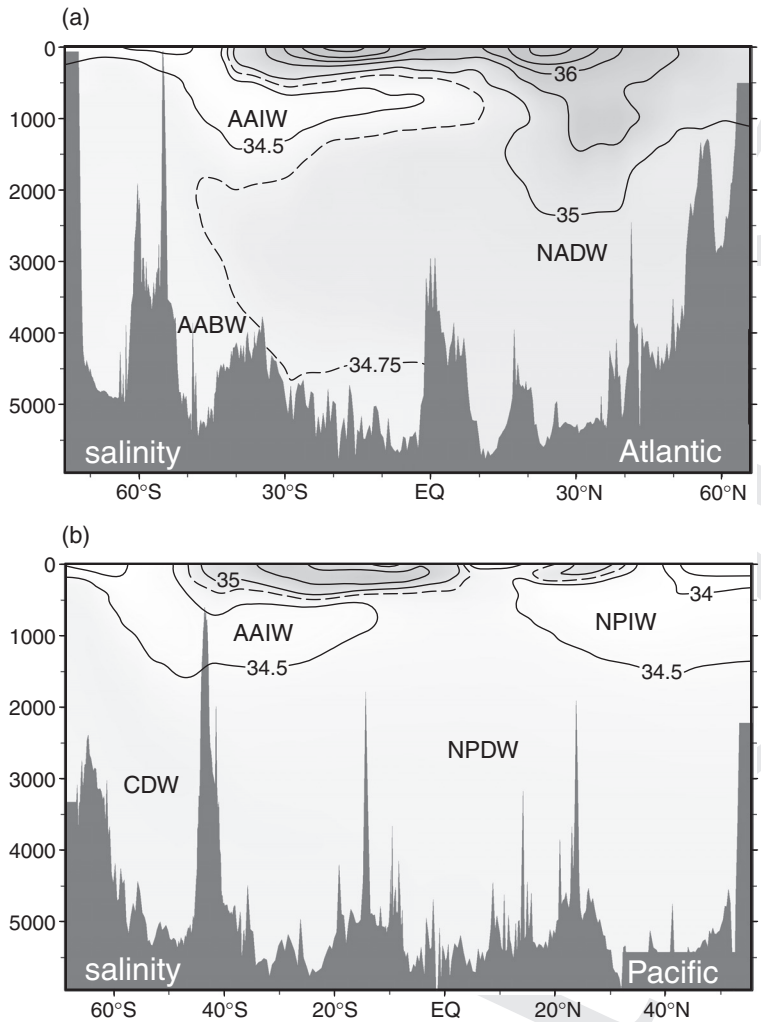

Figure 2.7 Observed meridional sections of salinity, S ($g\,kg^{-1}$), the concentration of salt in (a) the Atlantic along 20°W, and (b) the Pacific along 170°W; see Plates 2b and 3b. The salt distribution is used to define the major water masses: AAIW, Antarctic Intermediate Water; AABW, Antarctic Bottom Water; CDW, Circumpolar Deep Water; NADW, North Atlantic Deep Water; NPIW, North Pacific Intermediate Water; and NPDW, North Pacific Deep Water. These water masses are formed in discrete regions so that the water at any particular location is usually a composite of several water masses.

through the exchange of heat, moisture and dissolved gases with the atmosphere and, in polar regions, with ice, as well as biogeochemical properties affected by biological activity (Fig. 2.4).

Once the water masses spread in the ocean interior (Fig. 2.5b), they preferentially conserve their salinity, as well as pressure-corrected variants of temperature and density, referred to as potential temperature and potential density, respectively. There are striking contrasts in surface salinity between subtropical and subpolar gyres, as well as between the Atlantic and Pacific basins. Consequently, the interior salinity distribution is very useful in revealing how water masses spread. For example, salinity sections through the Atlantic and Pacific, as depicted in Fig. 2.7, reveal the following pathways:

- Plumes of low salinity suggest a northward spreading at mid depths and along the bottom from the Southern Ocean into the Atlantic and Pacific basins, as well as a southward spreading in upper waters from high latitudes in the northern Pacific.
- Broader regions of higher salinity suggest a southward spreading over several kilometres depth in the Atlantic, as well as a more localised salty intrusion spreading at a depth of typically 1 km, ultimately originating from the Mediterranean Sea.

Spreading of mode waters

The water-mass distribution can be viewed as layers of water with nearly uniform properties,

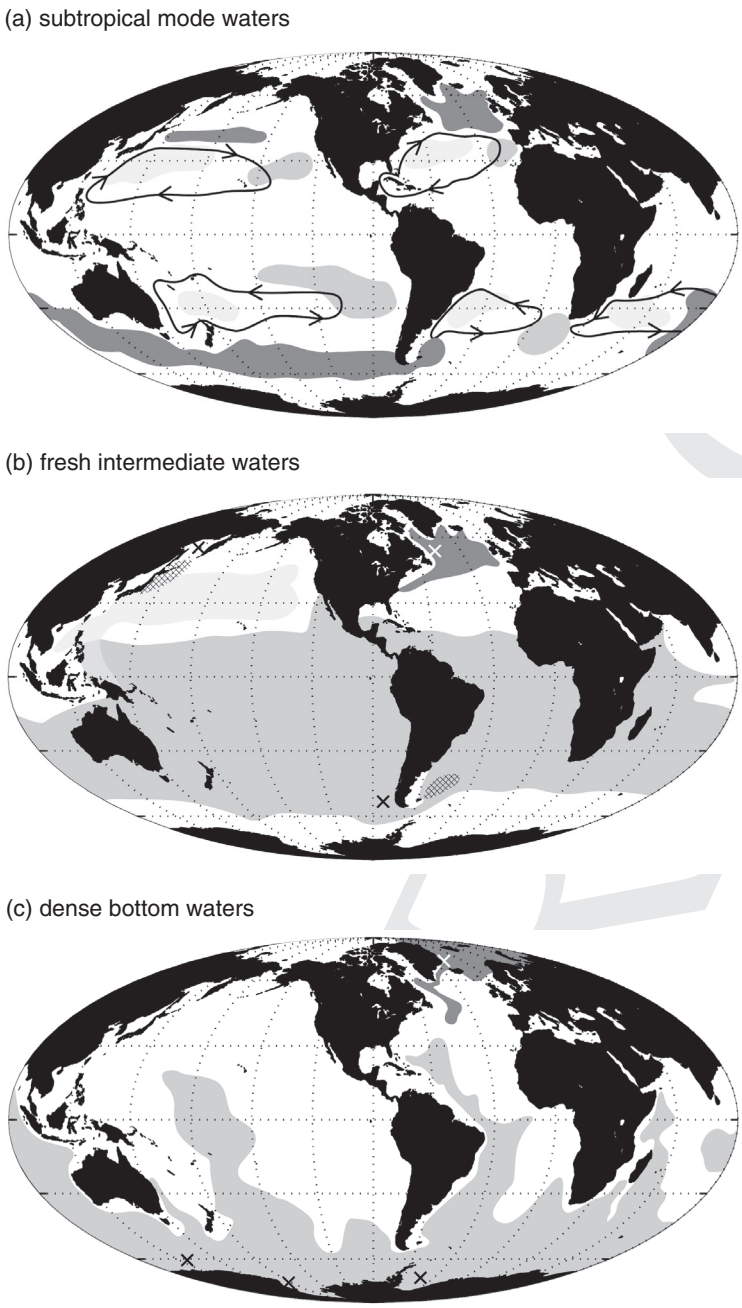

Figure 2.8 Global distribution of mode waters: (a) subtropical mode waters formed on the western side (light shading), eastern side (medium shading) and the poleward side (dark shading) of the gyre, including subpolar mode water in the North Atlantic, North Pacific central mode water and Sub-Antarctic mode water in the southern hemisphere, together with cartoons of the gyre circulation. (b) Fresh intermediate mode waters including Labrador Sea Water (dark shading, $\sigma_\theta = 27.8$), North Pacific Intermediate Water (light shading, $\sigma_\theta = 27.0$) and Antarctic Intermediate Water (medium shading, $\sigma_\theta = 27.1$), together with their formation sites (X's) and neighbouring regions of strong mixing (hatching). (c) Spreading of dense bottom waters originating from the Antarctic (dark shading) or the North Atlantic (light shading), as defined by the $\sigma_4 = 45.92$ surface (formation sites are marked by X's). Redrawn from figures courtesy of Lynne Talley; further details see Talley (1999).

referred to as mode waters, stacked on top of each other; the lightest mode waters lying above denser mode waters. The distribution of mode waters then provides clues as to the underlying interior circulation over the globe, as illustrated in Fig. 2.8 from Talley (1999):

- Light mode waters are formed in the mid latitudes, generally within the wind-driven subtropical gyres (Fig. 2.8a). For example, in the North Atlantic, the subtropical mode water is formed within the mixed layer close to the Gulf Stream, as illustrated by the 18 °C water in Fig. 2.4b, then

spreads into the thermocline over the western side of the subtropical gyre.

- Intermediate mode waters form in high latitudes or in neighbouring semi-enclosed seas, and spread at depths of typically 1 to 2 km (Fig. 2.8b). For example, in the North Atlantic, warm and salty Mediterranean Sea Water spreads out from the Straits of Gibraltar (after experiencing much mixing) at a depth of ~1200 m, while the colder and fresher Labrador Sea Water spreads at depths ranging from 500 m to 2000 m. In the southern hemisphere, Antarctic Intermediate Water, formed in the southeast Pacific, spreads around the Southern Ocean and extends northward, reaching as far as the tropics of the northern hemisphere (Fig. 2.8b).
- Dense mode waters making up the bottom waters over the globe are formed in the high latitudes of the North Atlantic and off Antarctica (Fig. 2.8c). Their spreading is steered by topography and their properties are gradually diluted by mixing, particularly in fracture zones and regions of rough topography.

Schematic view of the overturning

Based on these water-mass distributions (as in Figs. 2.7 and 2.8), a simplified cartoon view of the ocean overturning can be constructed, as schematically set out in Fig. 2.9:

- Light or less dense water circulates in upper ocean cells associated with the wind-driven gyre circulations, which are confined within the mixed layer and upper thermocline. These cells provide a poleward transport of warm water from the tropics to high latitudes.
- Dense water circulates in a bottom cell and spreads from the Southern Ocean into the northern basins.
- Between these two cells, there is a northward transport of water from the Southern Ocean at mid depths, as well as a southward transport of mid-depth and deep waters from the northern basins.

The relative extent of each of the overturning cells varies within each basin. In the Atlantic, there is a southward spreading of deep water, overlying a more limited northward spreading of bottom water originating from Antarctica (Fig. 2.9a). In contrast, in the Pacific, there is a much weaker southward spreading of water masses formed in the northern basin, together with deep and bottom water spreading northward from the Antarctic Circumpolar Current (Fig. 2.9b).

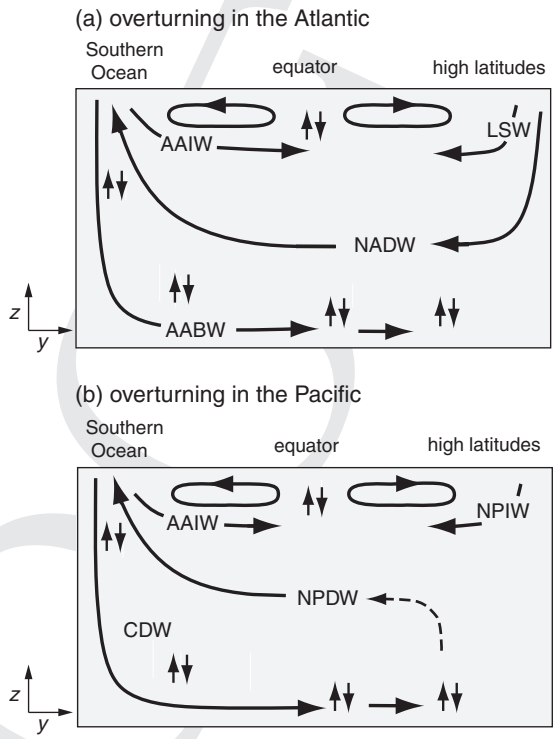

Figure 2.9 A schematic figure depicting a highly idealised, zonally averaged view (depth versus latitude) of the meridional circulation for (a) the Atlantic, and (b) the Pacific. Transport pathways (full lines) are based upon the preceding water-mass sections revealing characteristic water masses (definitions given in Fig. 2.7 together with Labrador Sea Water, LSW). Possible regions of strong vertical mixing (pairs of vertical arrows) are in the Southern Ocean, in the bottom waters and at the equator. The dominant components of the circulation are: subtropical gyre circulations in the upper ocean; a southward spreading of intermediate waters (LSW, NPIW) from the northern high latitudes; a southward spreading of dense water (NADW) in the Atlantic; and a northward spreading of intermediate water (AAIW) and dense bottom water (AABW, CDW) from the Southern Ocean. The bottom water in the Pacific lightens and forms deep water (NPDW) somewhere over the basin (dashed line).

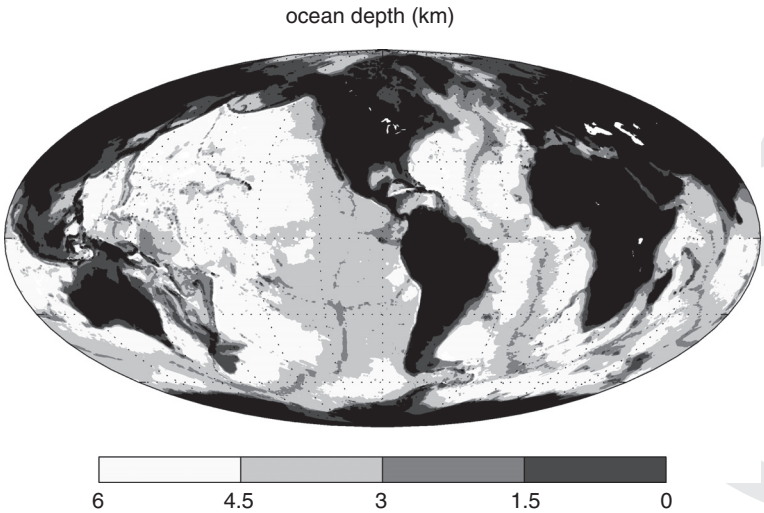

Figure 2.10 Map of ocean topography (km). The deepest ocean basins (unshaded) are defined by the position of topographic ridges (shaded), such as the mid-Atlantic ridge (30°W) and East Pacific Rise (110°W). The topography affects how the deep and bottom waters spread over the global ocean.

2.1.4 What is the effect of topography?

Topography affects the pattern of the deep circulation over the globe, sometimes providing a barrier that prevents water spreading into particular regions, as revealed by comparing the depth of the ocean and extent of bottom water (Figs. 2.10 and 2.8c). The effect of the topography on the surface circulation is less clear and consistent (Fig. 2.2a): the surface flow is not strongly deflected by the mid-Atlantic ridge (along 30°W) or by the East Pacific Rise (along 110°W), yet is deflected by ridges in the Southern Ocean.

Why does the topography appear to affect the surface flow in some cases and not in others? The answer lies in how the thermocline separates the surface and deep flows. When there is a thermocline and strong vertical contrasts in density, the surface and deep flows can be very different, even flowing in opposite directions, so that surface flows do not resemble the pattern of the underlying topography. Conversely, when there is no thermocline and vertical contrasts in density are weak, the surface and deep circulations flow in the same direction, hence such top to bottom circulations pass around ridges or bumps in the sea floor.

2.1.5 Summary

The ocean circulation is due to a combination of mechanical and density forcing, involving the surface winds, exchanges of heat, fresh water and salt, as well as tide-inducing gravitational accelerations. The ocean circulation can be viewed in the horizontal in terms of recirculating gyres and boundary currents within basins and near zonal flows close to the equator and in the Southern Ocean, as well as in the vertical by overturning cells connecting each of the basins with the Southern Ocean. Clues as to the interior circulation are provided by the distribution of water masses and tracers over the globe (see colour plates and later Section 10.1).

2.2 Atmospheric circulation

While this book focusses on the role of the ocean, it is important to consider how the atmosphere circulates in order to understand the primary forcing of the ocean and how the atmosphere and ocean interact in the climate system.

2.2.1 How does the atmosphere circulate?

Remotely sensed pictures of the planet, such as in Fig. 2.11, provide clues from the pattern of clouds as to how the atmosphere circulates. There is always persistent high cloud in the tropics (23°S to 23°N), contrasting with normally cloud-free skies outside the tropics (at typically 30°N and 30°S), and more variable cloud cover at mid and high latitudes. Clouds form as moist air rises; air cools with height, holding less water vapour, leading to water condensing and forming cloud drops.

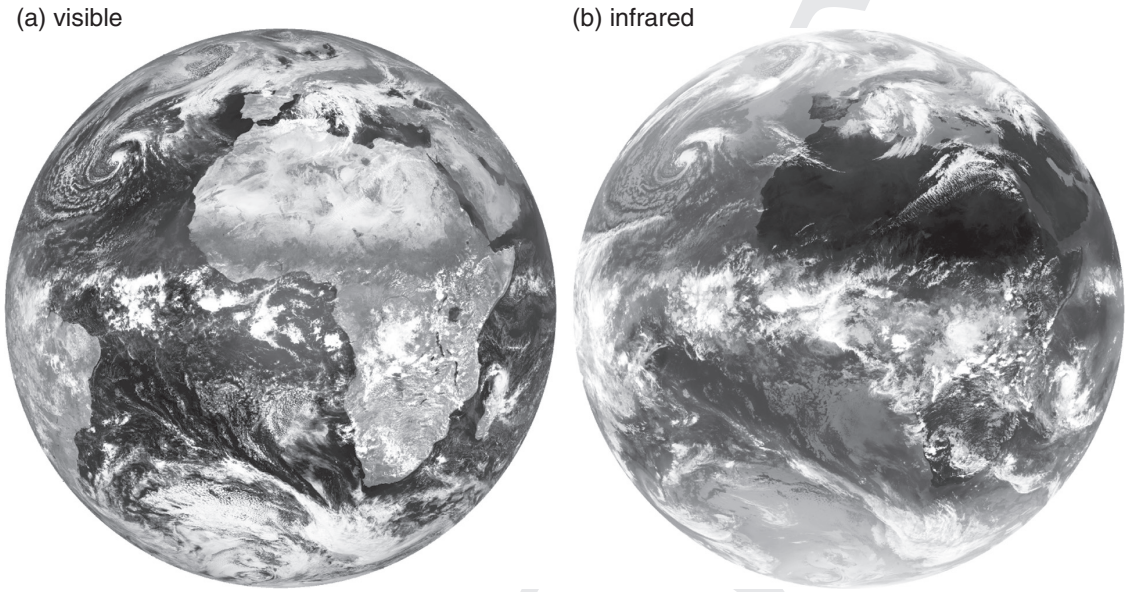

Figure 2.11 Atmospheric cloud revealed over the Atlantic sector for 6 March 2008 at 1200 UTC: (a) visible image of reflected light where white represents high reflectance from cloud (0.74–0.88 μm band) and (b) infrared image of long-wave radiation emitted, where white represent cold and dark represents warm temperatures, such as over the land (11–13 μm band). There is a band of cold, high clouds over the tropics, identifying the Intertropical Convergence Zone, together with clearer sky over northern Africa. At mid and high latitudes, there are variable cloud structures associated with the weather systems, such as the spiral cloud of a cyclone over the North Atlantic. From Meteosat Second Generation satellite MSG-2, image copyright: EUMETSAT, NERC Satellite Receiving Station, University of Dundee.

Conversely, sinking air is often associated with an absence of clouds. Individual bands of moving clouds also reveal the presence of fast-moving jets or storms, as seen when watching animated weather forecasts.

Consider now how the atmospheric circulation is controlled over the globe.

Overturning cells

The atmospheric circulation is ultimately driven by the latitudinal variation of the Sun's heating over the globe. Heating leads to a narrow band of warm, moist air rising over the tropics, indicated by thick cloud and strong precipitation (Fig. 2.11). This warm air moves poleward at heights of 10–20 km and is replaced by equatorward moving air in the lower few kilometres – an overturning referred to as the Hadley cell, as depicted in Fig. 2.12.

This overturning cell does not extend in a simple manner over the entire globe. Instead the circulation is strongly constrained by the Earth's rotation. The warm tropical air initially moves poleward aloft, but is deflected by the Earth's rotation into fast-moving, westerly upper air jets (Fig. 2.12; see Q2.4). The tropical air eventually descends just outside the tropics (typically 30°N and 30°S); hence, the Hadley cell only extends from the equator to just outside the tropics. The descending air warms and any water remains in vapour form, leading to clear skies and little precipitation (Fig. 2.11); hence, the great desert belts are formed along these latitude bands over the globe.

Mid-latitude jets and weather systems

In the mid latitudes, the fast-moving westerly jets are naturally expected to transfer heat zonally, rather than poleward. Hence, there is a problem: how does the atmosphere continue to move heat towards the poles across the mid latitudes?

These westerly jets turn out to be unstable, meandering and forming weather systems on horizontal scales of a thousand kilometres, as

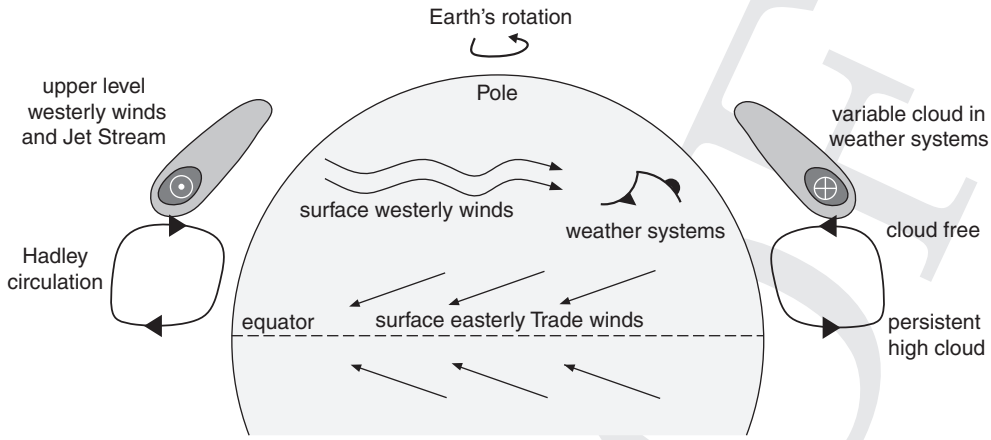

Figure 2.12 A schematic of the atmospheric general circulation. Heating in the tropics leads to rising warm air and persistent high cloud. This tropical air moves poleward aloft, but is deflected eastward and eventually descends at 30°N and 30°S, leading generally to cloud-free regions; this air is replaced on the ground through easterly Trade winds and towards the equator. On the poleward flank of this overturning cell (called the Hadley circulation), there are fast moving, upper level, westerly winds with the core of the subtropical Jet Stream (shading represents strong eastward flow); the surface winds at mid latitudes are also westerly. These fast jets in the mid latitudes are unstable, forming weather systems containing warm and cold fronts, associated with variable cloud, and lead to a polewards heat flux. Modified from article published in Marshall and Plumb (2007), © Elsevier.

illustrated by the cyclones with spiralling high cloud over the North Atlantic in Fig. 2.11. There is an exchange of air within these weather systems linked to the warm and cold fronts: warm air rises and moves poleward at the warm front, while cold air sinks and moves equatorward at the cold front. Hence, the instability of the zonal jets and formation of weather systems leads to a poleward heat flux at mid latitudes over the planet.

At mid latitudes, you can see the passage of these fronts from their characteristic cloud structures: early warning of an oncoming warm front is given by the arrival of high-level ice clouds (cirrus), identified by a hooked or hair-like profile from falling ice crystals in the sky. The later approach of the warm and cold fronts is associated with thickening grey cloud and drizzle, and the onset of the cold front by heavy rain. The eventual passing of the cold front is heralded by colder air, sometimes scattered showers, and clearer visibility.

In summary, the atmospheric general circulation is driven by the Sun's differential heating over the globe, although the response is constrained by the rapid rotation of the planet (Fig. 2.12). Heat is transferred poleward in the atmosphere by a combination of the Hadley overturning cell, extending from the equator to just outside the tropics, and then by eddy circulations formed along the zonal jets in mid and high latitudes.

2.2.2 How do the atmosphere and ocean transfer heat together over the globe?

Both the atmosphere and ocean lead to a poleward heat transport, reaching 5 PW at mid latitudes as displayed in Fig. 2.13. The heat gained in the tropics is transported to the mid and high latitudes, where the heat is ultimately radiated back to space at the top of the atmosphere (Fig. 1.3a). The atmosphere provides the dominant contribution, reaching more than 4 PW at mid latitudes (Fig. 2.13, dashed line). The ocean provides a smaller heat transport reaching 1 to 2 PW in the tropics (at ±20°N) and decreasing to less than half this value by the mid latitudes (±40°N) (Fig. 2.13, full line).

Concomitant with these transport patterns, the ocean gains heat in the tropics, but releases heat to the atmosphere at mid and high latitudes (Fig. 2.14). The strongest release of heat to the atmosphere occurs over the western side of the

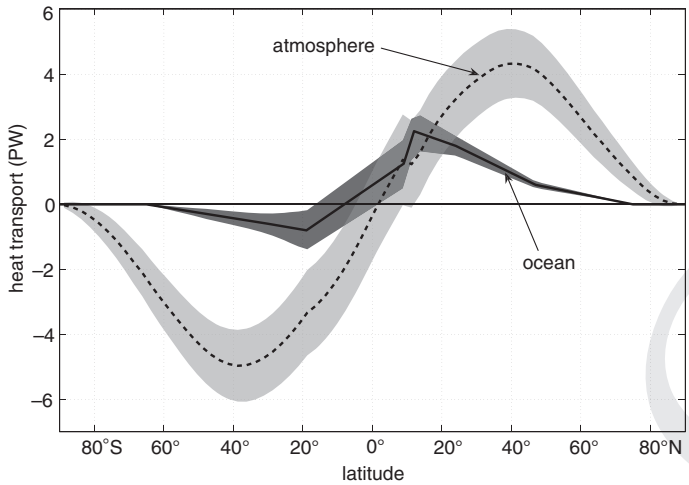

Figure 2.13 The northward transfer of heat (PW ≡ 10^{15} watts) over the globe by the atmosphere (dashed line) and ocean (full line), with shading for the error range. The ocean transport is diagnosed using a global inverse model based upon ocean section measurements, while the atmospheric transport is diagnosed as a residual using a heat budget from the mismatch between the measured radiation at the top of the atmosphere (from the Earth Radiation Budget Experiment over 3 years from 1987 to 1989) and these ocean measurements. From C. Wunsch; further details, see Wunsch (2005).

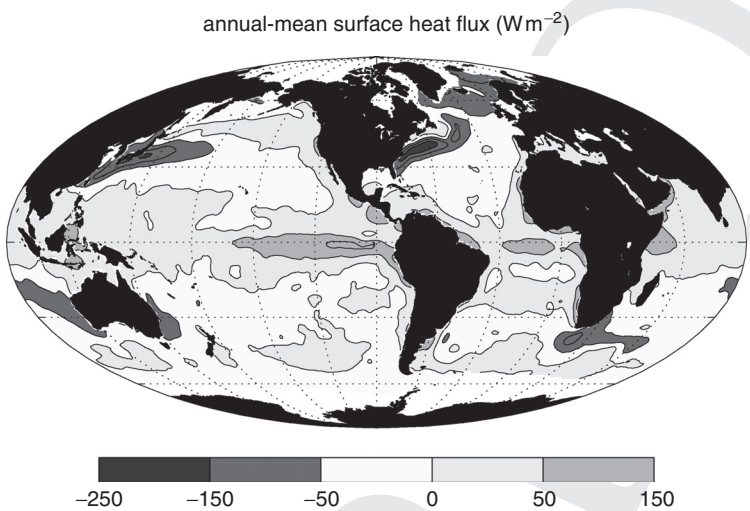

Figure 2.14 Air–sea heat flux into the ocean (W m^{-2}) from the NOCS climatology (Josey *et al.*, 1999) with contours every 50 W m^{-2}. There is a heat input in the tropics (light shading) and a heat loss over high latitudes and the western side of basins at mid latitudes (dark shading).

ocean basins in the mid latitudes, where cold, dry air passes from the continents over the warm waters in the ocean boundary currents.

The overall poleward heat transport by both fluids can partly be viewed as a relay race where the ocean transports heat poleward at low latitudes, passing the heat onto the atmosphere at mid latitudes, which is then transferred further poleward by the atmosphere, until ultimately the heat is radiated back to space.

2.2.3 Summary

The atmosphere and ocean combine together to transfer heat poleward, reducing latitudinal temperature contrasts and making the Earth's climate more equitable. There are some common phenomena in both fluids: the strong jet streams in the atmosphere and the intense currents in the ocean, including the Antarctic Circumpolar Current, equatorial currents and western boundary currents. Instability of these intense flows leads to strong temporal variability, generating weather systems in the atmosphere and ocean eddies on the horizontal scale of several tens of kilometres.

There are also some important differences between the atmosphere and ocean: the continents provide barriers to zonal flow in the ocean, leading to gyre circulations within ocean basins; and the fluids are heated at opposing boundaries,

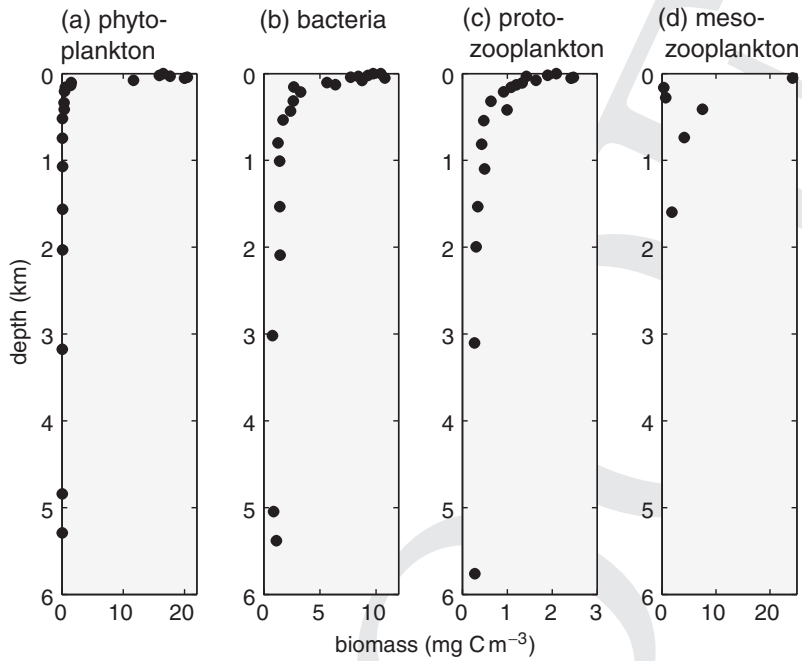

Figure 2.15 Observed vertical profiles of biomass (mg C m^{-3}) over the full water column depth in the North Pacific Ocean (39°N, 147°E): (a) phytoplankton, 0.2–200 μm, which perform photosynthesis; (b) bacteria, 0.2–2 μm, which respire organic detritus; (c) protozooplankton, unicellular and small predators, 2–200 μm, which consume bacteria and phytoplankton; and (d) mesozooplankton, 200–2000 μm, including tiny shrimp-like copepods, which prey upon the smaller organisms. Phytoplankton are restricted to the surface waters where sunlight can penetrate. Bacteria and their protozooplankton predators are ubiquitous, since organic detritus is found throughout the water column. They are more abundant at the surface in the region where photosynthesis provides a strong source of new organic material. Mesozooplankton are seen here throughout the upper water column (although observations were not made in the deepest waters). Replotted from Yamaguchi *et al.* (2002).

the atmosphere at its lower boundary and the ocean at its upper boundary. This contrast leads to the strongest atmospheric convection in the tropics, while the strongest ocean convection is at high latitudes.

Following these descriptive views of how the atmosphere and ocean circulate, we now turn to questions of how the ocean ecosystem and the carbon cycle operate.

2.3 Life and nutrient cycles in the ocean

The oceans sustain an enormous diversity of living creatures. While we are most familiar with the larger organisms, such as fish or whales, the vast majority of the living biomass is in the form of microbes; tiny creatures that cannot be seen without the aid of a microscope.

Phytoplankton are the tiny plants of the ocean performing photosynthesis, converting the energy in light to chemical energy through the formation of organic molecules. They produce chlorophyll and other pigments in order to absorb light. Since light penetrates only 100 m or so in seawater, phytoplankton are confined to live in near-surface waters, as revealed in Fig. 2.15a. They are extremely diverse in form and function, ranging in size from one to several hundred microns.

Bacteria and archaea are also extremely small, typically less than a micron in size. They acquire energy and nutrients by breaking down and respiring pre-existing organic matter. They live throughout the water column since particles of organic detritus continually rain down from the

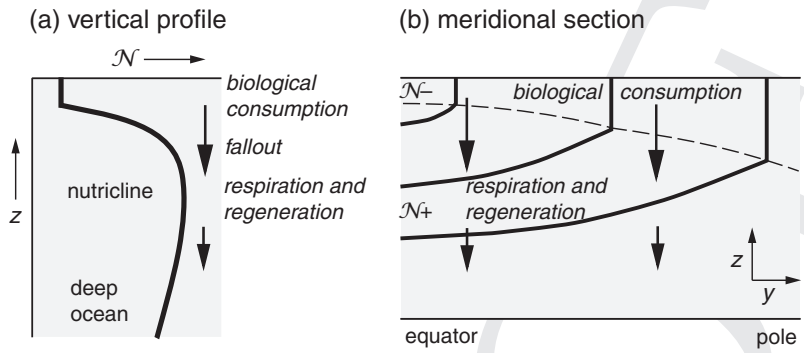

Figure 2.16 A schematic view of the biological cycling of nutrients in the ocean: (a) a vertical profile, and (b) a meridional section for a typical inorganic nutrient, $\mathcal{N}$, such as nitrate, phosphate and carbon. The nutrient profile is separated into a mixed layer overlying the nutricline and the deep ocean. Inorganic nutrients are consumed within the surface, sunlit ocean, then organic particles fall out of the surface layer. Most of the organic fallout is respired, leading to inorganic nutrients being regenerated and forming a subsurface maximum in concentration within the mid depths of the water column. In addition, the physical circulation also transports nutrients along density surfaces (black lines), leading to lower nutrient concentrations in younger waters (more recently in contact with the mixed layer) and higher concentrations in older waters.

productive surface layers, providing them with a means of sustenance (Fig. 2.15b).

Both phytoplankton and bacteria are eaten, or grazed, by zooplankton; unicellular grazers, perhaps some tens of microns in size, prey upon them. Some of these grazers also perform photosynthesis, hedging their bets for survival. Larger zooplankton, like submillimetre-sized shrimp, have more complex physiologies, and a variety of hunting strategies, and may mechanically break down particles of detritus. Like the bacteria, zooplankton can thrive throughout the water column (Fig. 2.15c,d).

2.3.1 Biological cycling of nutrients

Phytoplankton and other organisms need carbon, nitrogen, phosphorus, sulphur, iron and other elements to create their structural and functional organic molecules. The elemental composition of phytoplankton, in a bulk average sense, is relatively uniform (C:N:P = 106:16:1) reflecting the common biochemical molecules from which they are made.

Phytoplankton die through viral infection or are grazed by zooplankton, which in turn provide a food source for fish. This organic matter eventually either sinks or is transported from the sunlit, surface waters into the dark interior. As particles sink, they aggregate and disaggregate. They are consumed by zooplankton and filter feeders like jellyfish which, in turn, produce new detritus. They are inhabited by colonies of bacteria which attack and respire their organic components. Nearly all of the organic matter is oxidised within the water column. Less than one per cent of the sinking organic matter reaches the sea floor, where most of the remaining organic material is rapidly reworked by the benthic ecosystem.

Surface waters transferred into the interior ocean contain inorganic nutrients and dissolved organic material. As these waters move through the deeper ocean, the respiration of dissolved organic matter and sinking organic particles leads to a regeneration of inorganic nutrients, increasing the concentrations of phosphate and nitrate, as illustrated schematically in Fig. 2.16.

The effect of the interior source of inorganic nutrients varies according to the elapsed time since the waters were in the surface mixed layer. In the Atlantic, deep waters are relatively young (having been relatively recently formed in the mixed layer) and the nitrate distribution resembles that of a physical tracer, such as salinity; compare Figs. 2.17a and 2.7a. Conversely, in the deep Pacific, the deep waters are relatively old and the accumulated effect of biological fallout and respiration has created strong gradients in nitrate compared with salinity; compare Figs. 2.17b and 2.7b.

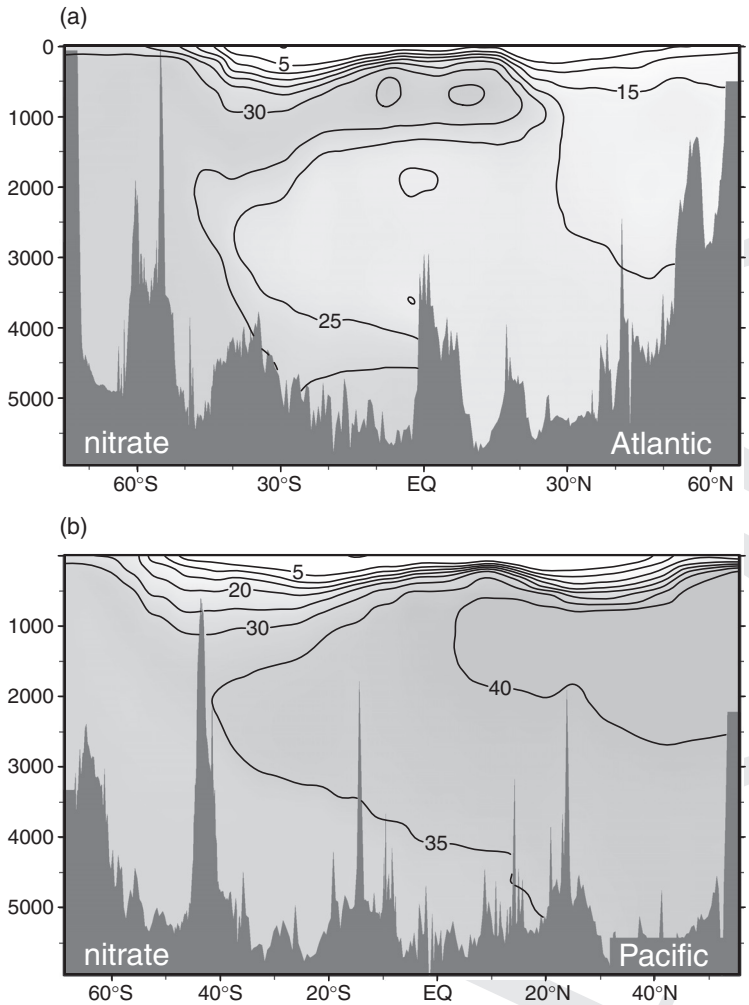

Figure 2.17 Observed meridional sections of nitrate (NO_3^-; $\mu mol\,kg^{-1}$) for (a) the Atlantic along 20°W, and (b) the Pacific along 170°W. The nitrate concentrations are generally depleted at the sea surface and increase to a mid-depth maximum. There is a strong vertical gradient in the upper waters, referred to as the nitracline, which undulates in the same manner as the thermocline. Horizontal contrasts in the nitrate concentrations reflect the physical transport of water masses, as revealed by salinity (Fig. 2.7).

Consequently, there are strong contrasts between the distributions of inorganic nutrients in the Atlantic and Pacific.

2.3.2 Where is organic matter produced?

Light is essential for photosynthesis, but visible wavelengths are absorbed very effectively by seawater, as well as any suspended particles, within a few tens of metres from the sea surface. Hence, organic matter is produced only close to the surface of the ocean. Phytoplankton absorb sunlight using pigments, notably chlorophyll *a*, which absorb visible light most effectively in blue and red wavelengths. Thus, the ocean appears with a green tinge in waters where phytoplankton are abundant. The 'greenness' of the ocean, measured by comparing the relative transmission and backscattering of green and blue wavelengths, can then be used to infer the concentration of chlorophyll and, hence, provide a gross measure of the surface phytoplankton distribution, as depicted in Fig. 2.18a.

Though the incident solar radiation increases towards the tropics, the highest surface concentrations of chlorophyll are in the shelf seas and high-latitude open ocean. Phytoplankton require not only sunlight, but also inorganic nutrients. Hence, the chlorophyll distribution broadly resembles the underlying nitrate distribution, with high concentrations in the high latitudes, shelf seas and parts of the tropics, and low or depleted concentrations in the mid latitudes (Fig. 2.18b); this connection is explored further in Sections 7.2 and 11.1.

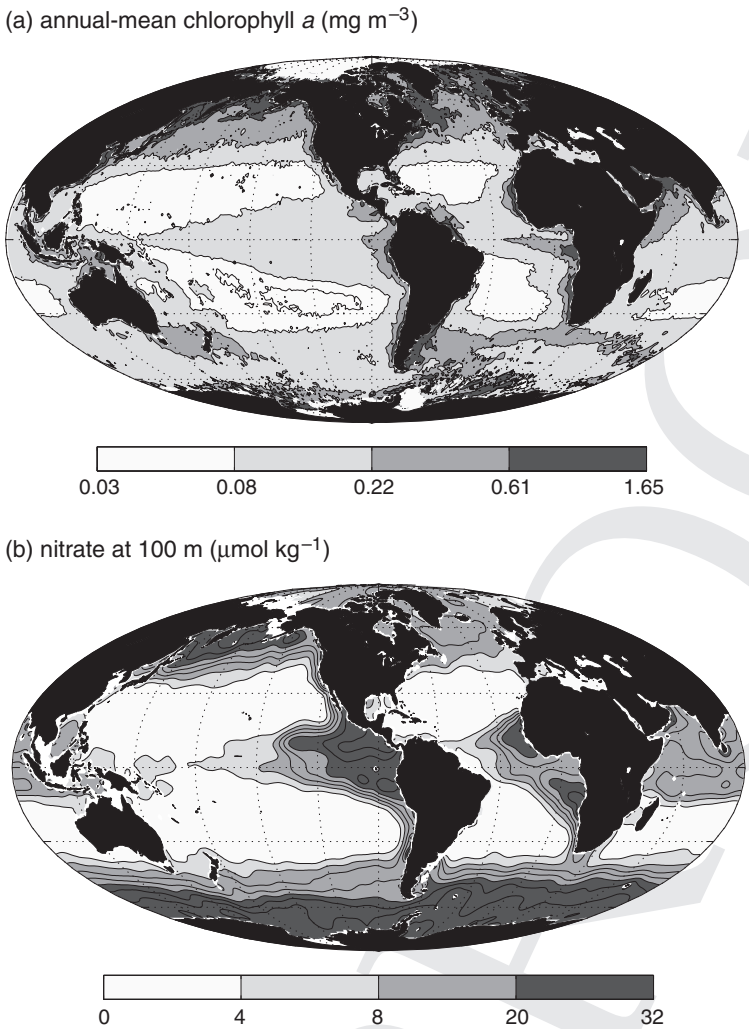

Figure 2.18 (a) The global distribution of phytoplankton inferred from annual-mean concentration of the green pigment, chlorophyll *a* ($C_{55}H_{72}O_5N_4Mg$; mg m^{-3}), inferred from remotely sensed backscatter of light observations (from NASA Sea-viewing Wide Field-of-view Sensor, SeaWiFS) at 1° resolution for 2006. Phytoplankton growth requires a supply of nutrients, so the pattern of chlorophyll broadly resembles (b) the distribution of nitrate at 100 m (contours every 4 μmol kg^{-1}) (Conkright *et al.*, 2002). There are higher concentrations of nitrate and chlorophyll *a* at high latitudes and the tropics, and along the coast, but much lower values over the open ocean in the mid latitudes.

2.3.3 Diversity in phytoplankton types

Many different types of phytoplankton contribute to the gross measure of phytoplankton distribution provided by remotely sensed chlorophyll (Fig. 2.18a). Each phytoplankton group and species has evolved to exploit slightly different conditions, some preferring relatively stable situations and others preferring regions of strong seasonality.

For example, diatoms are a fast growing group requiring silica to form parts of their protective shell (Fig. 2.19a). Thus, their distribution reflects that of the surface silica distribution, which has higher concentrations in high latitudes and parts of the tropics, but very low concentrations in the subtropics (Fig. 2.19b). In contrast, tiny *Prochlorococcus* cells are found mainly in the stable, nutrient-depleted subtropical waters over the globe; the diversity of phytoplankton types and the effect of seasonality is explored further in Sections 5.4 and 7.2.

2.3.4 Summary

This interplay of life and nutrient cycling in the ocean is very different from on land. As terrestrial plants grow, they take up nutrients and trace metals from the soil. When they die these elements are returned to the soil as the organic matter is

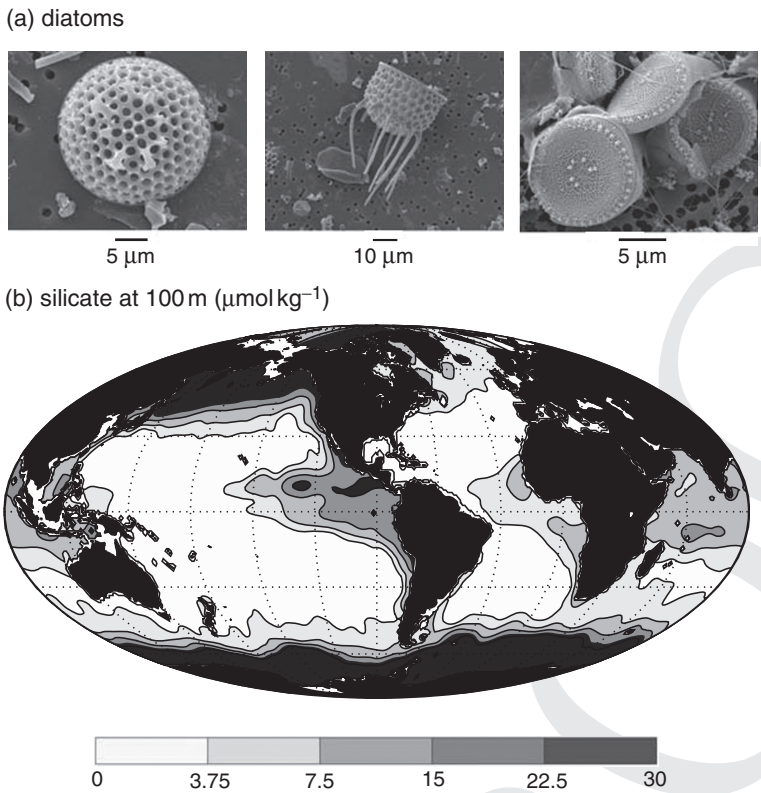

Figure 2.19 (a) Electron micrographs of the silica-based structural parts of fossil and modern diatoms (images courtesy of Zoe Finkel). There are thousands of species of diatom, spanning several orders of magnitude in cell size, some forming chain-like colonies. The availability of silicic acid in surface waters shapes their habitat. (b) Climatological, annual silicic acid concentration (μmol kg^{-1}) at a depth of 100 m (Conkright *et al.*, 2002).

respired locally by bacteria, and are assimilated the next year, and so the cycle goes on. In the oceans, sunlight is rapidly absorbed with depth and most of the water column is dark. When phytoplankton grow in the thin, sunlit surface layer, they consume inorganic nutrients. Most of the organic matter formed in the surface ocean is recycled locally, but a small fraction sinks through the water column, appearing like falling snow. Ultimately this organic matter is respired and regenerates inorganic nutrients at depth. This biological cycling then acts to transfer nutrients from the surface to the deep ocean and, ultimately, photosynthesis ceases unless nutrients can be returned to the sunlit surface layer. This resupply of nutrients to the surface ocean is principally achieved by physical processes acting within the ocean.

2.4 The carbon cycle in the ocean

Carbon dioxide dissolves and reacts in seawater forming dissolved carbon dioxide, CO_2^* (defined by the sum of the aqueous form of carbon dioxide, CO_2^{aq}, and carbonic acid, H_2CO_3), bicarbonate ions, HCO_3^-, and carbonate ions, CO_3^{2-}, which collectively are referred to as dissolved inorganic carbon, DIC:

$$\mathrm{DIC} = [CO_2^*] + [HCO_3^-] + [CO_3^{2-}], \tag{2.1}$$

where square brackets denote concentrations in seawater defined per unit mass in μmol kg^{-1}.

While carbon is exchanged between the atmosphere and ocean in the form of carbon dioxide, most of the carbon dioxide is transferred into bicarbonate and carbonate ions within the ocean, such that typically 90% of DIC is made up of bicarbonate ions, about 9% as carbonate ions, and only a small remainder, up to 1%, as dissolved carbon dioxide (Fig. 2.20).

This transfer of carbon dioxide into bicarbonate and carbonate ions then leads to the ocean holding 50 times as much carbon as in the overlying atmosphere. This inorganic carbon in the ocean is about 40 times larger than the amount held as organic carbon.

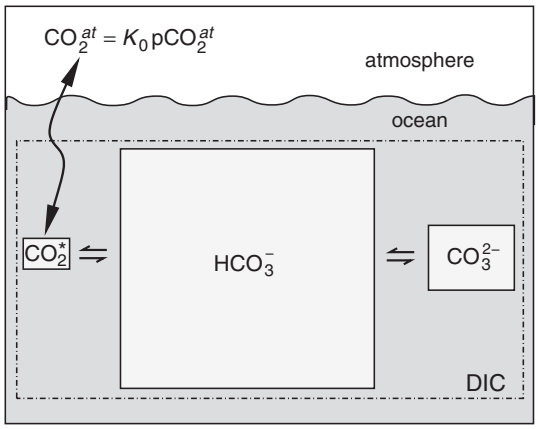

Figure 2.20 Carbon dioxide dissolves in and reacts with seawater, making up three dissolved inorganic forms of carbon: dissolved carbon dioxide, CO_2^*, bicarbonate, HCO_3^-, and carbonate CO_3^{2-} ions, which together make up dissolved inorganic carbon, DIC. For the present day, most of the dissolved inorganic carbon is in the form of bicarbonate ions. The air–sea exchange of carbon dioxide depends on the difference in the dissolved concentration of carbon dioxide in the ocean and the atmosphere, CO_2^{at}, which is related to the solubility, K_0, and the partial pressure of carbon dioxide, pCO_2^{at}.

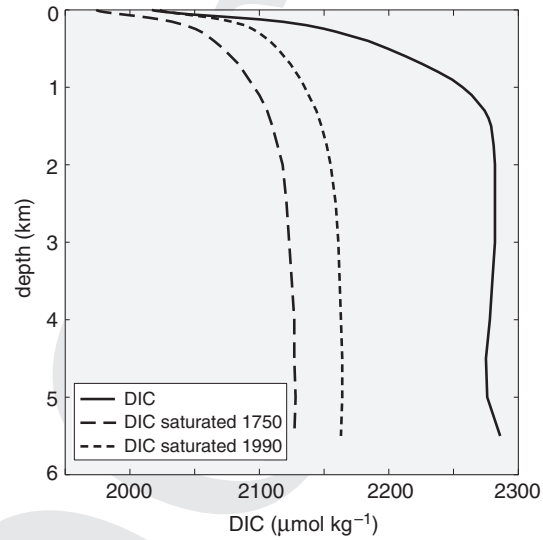

Figure 2.21 Vertical distribution of dissolved inorganic carbon, DIC (μmol kg^{-1}): globally averaged: observed vertical profile (solid line), together with equilibrium profiles *if* there were no biological processes and surface waters were in complete equilibrium with an atmospheric carbon dioxide for the pre-industrial era (long dashed line) or 1990 (short dashed line).

2.4.1 The vertical distribution of carbon

Dissolved inorganic carbon generally increases in concentration with depth, as depicted in Fig. 2.21 (solid line) with vertical contrasts of typically 200 μmol kg^{-1} and a much larger depth average of nearly 2300 μmol kg^{-1}. Two factors maintain the vertical gradient in DIC:

- At equilibrium with the atmosphere, cooler waters hold more DIC. Since the density structure over the globe is largely controlled by temperature, cool, carbon-rich waters slide under warm, carbon-depleted waters. This physical enhancement of the carbon stored in the deep ocean is referred to as the solubility pump; discussed further in Section 6.3.
- In addition, phytoplankton take up carbon dioxide in the sunlit surface waters, creating organic matter by photosynthesis. A small fraction of this organic carbon sinks to the deep ocean before being respired and returned to inorganic form by bacteria, which then increases DIC at depth. This biological enhancement of the carbon stored in the deep ocean is referred to as the biological pump; discussed further in Sections 5.6 and 6.4.

If there was no biological cycling of carbon, then the interior DIC distributions would resemble their equilibrium values given by their temperature and salinity at depth and an atmospheric value of carbon dioxide; as depicted for the pre-industrial or present era by dashed and dotted lines in Fig. 2.21, respectively. The actual DIC profile (in Fig. 2.21, full line) are greater than the equilibrium profiles, but comparable in magnitude. Hence typically 90% of the carbon stored in the ocean is due to the reactivity of carbon dioxide in seawater and the solubility pump, while the remaining 10% is due to the biological pump.

Atlantic and Pacific contrasts in DIC

The DIC distribution differs in each basin: in the Atlantic, there is a layered structure resembling that of physical tracers (Fig. 2.22a), while in the Pacific there is a relatively large vertical contrast (Fig. 2.22b). These DIC distributions are again due to a combination of physical transport and biological transfers:

- The physical transfer leads to the layering structure. Surface waters carry their properties with them as they pass from the mixed layer

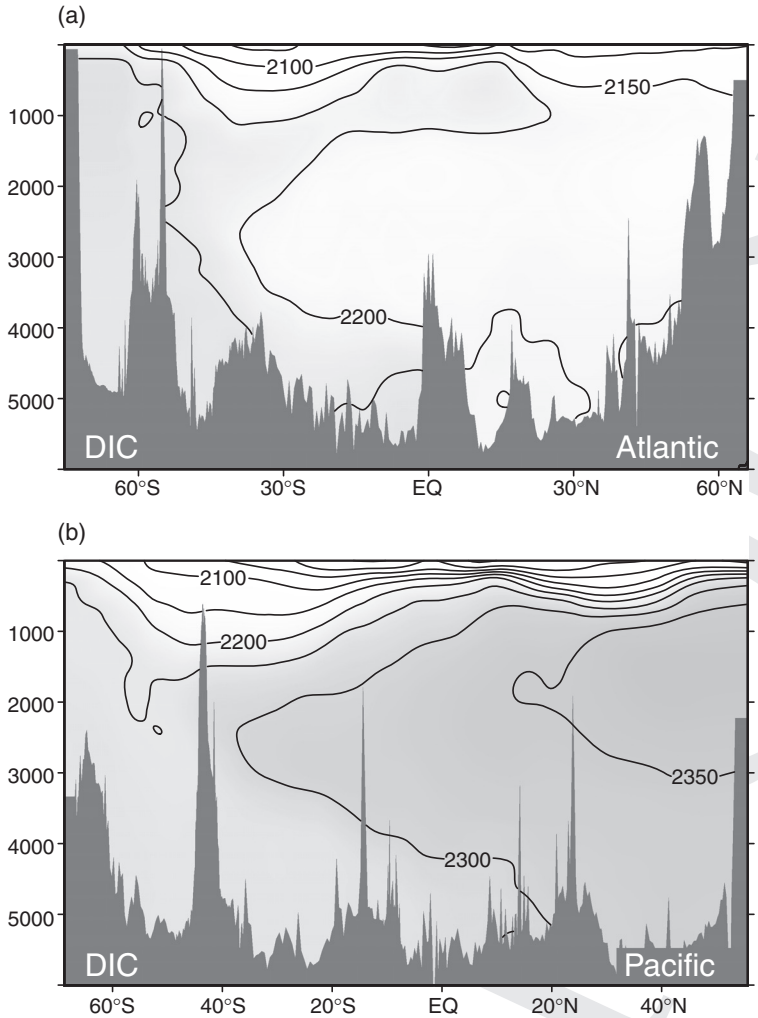

Figure 2.22 Observed meridional sections of dissolved inorganic carbon, DIC (μmol kg^{-1}) in (a) the Atlantic, approximately along 20°W, and (b) the Pacific, approximately along 170°W; see Plates 6a and 7a. DIC generally increases with depth, reflecting both the higher solubility of CO_2 in cold waters and the biological formation of sinking organic particles, which are respired and regenerated to inorganic nutrients and carbon at depth. Large-scale structures partly reflect physical tracers, like salinity.

into the ocean interior. Cool waters originating in the high latitudes bring high concentrations of DIC, while warmer waters originating in the mid latitudes bring low concentrations of DIC.

- The accumulated effect of the biological cycling and regeneration of carbon gives rise to the strong Atlantic–Pacific contrast in deep DIC, such that younger waters in the deep Atlantic have low DIC and older waters in the deep Pacific have high DIC.

2.4.2 Air–sea exchange of carbon dioxide

The atmosphere and ocean reservoirs of carbon are connected through air–sea exchange of carbon dioxide. Ocean uptake of CO_2 occurs over much of the mid and high latitudes and, conversely, ocean outgassing of CO_2 occurs in the tropics (Fig. 2.23, light and dark shades, respectively). The transfer of carbon between the atmosphere, the surface mixed layer and ocean interior involves the effects of physical transport, solubility, and biological cycling, as depicted respectively in Fig. 2.24a,b:

- At high latitudes, cooling increases the solubility and there is high biological productivity, both processes drawing down CO_2 from the atmosphere into the ocean.
- In the tropics, warming decreases the solubility and induces outgassing of CO_2. In addition, an upward transfer of deep waters to the surface can either enhance biological productivity from

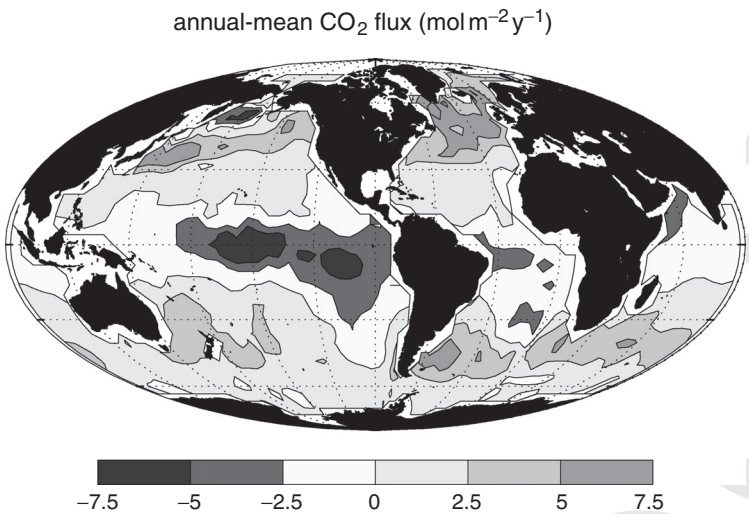

Figure 2.23 Climatological, annual-mean map of air–sea flux of CO_2 into the ocean (mol m^{-2} y^{-1}), data from Takahashi *et al.* (2002). There is an ocean uptake of CO_2 (light shading) in the high latitudes and an ocean outgassing in the tropics (dark shading); however, there is considerable uncertainty in the direction of the flux over the Southern Ocean. This estimate is based on a compilation of about a million measurements of surface-water pCO_2 obtained since the International Geophysical Year of 1956–59. The climatology represents mean non-El Nino conditions with a spatial resolution of $4^\circ \times 5^\circ$, normalised to reference year 1995.

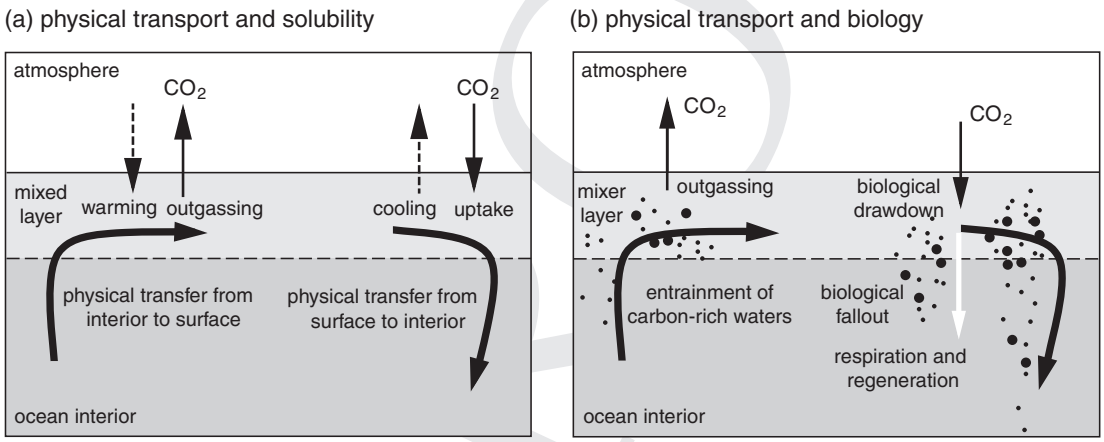

Figure 2.24 A schematic figure depicting how air–sea exchange of CO_2 is affected by an interplay of physical and biological processes involving the cycling and transport of DIC. In (a), warming of surface waters leads to an outgassing of CO_2, while a cooling of surface waters leads to an ocean uptake of CO_2. The physical transport (thick black arrows) of surface waters into the interior is then associated with an uptake of CO_2, while the return of deep waters to the surface leads to an outgassing of CO_2. In (b), the biological formation of organic matter leads to an ocean drawdown of atmospheric CO_2, which is then transferred to the ocean interior through the fallout of organic matter (white arrow). The respiration of the organic matter then regenerates the inorganic nutrients and carbon, increasing their concentrations in the deep waters. The return of deep water to the surface leads to an outgassing of CO_2.

the supply of nutrient-rich waters leading to an ocean uptake of CO_2 or, conversely, the supply of carbon-rich waters can lead to an outgassing of CO_2.

The air–sea fluxes of carbon dioxide integrated over the separate outgassing and uptake regions of the globe approximately balance each other, typically reaching 90 Pg C y^{-1} in each direction (based upon Fig. 2.23). In comparison, the anthropogenic increase in atmospheric CO_2 leads to a net air–sea flux directed into the ocean of 2 Pg C y^{-1}, much smaller than the regional variations in the annual flux. This anthropogenic increase may, though, tip the balance over parts of the Southern Ocean, possibly changing a pre-industrial net outgassing to a present-day or future uptake.

2.5 Summary

This chapter provides a preliminary view of the open ocean. The ocean circulation is driven by mechanical forcing from the surface winds and tides, as well as by pressure gradients from density differences formed through surface exchanges of heat, fresh water and salt. The ocean circulation can be viewed in terms of a horizontal circulation confined between the continents, made up of recirculating gyres and western boundary currents, together with intense, near-zonal currents running along the equator and circumnavigating the Southern Ocean. In addition, there is a vertical overturning with dense water sinking at high latitudes, especially in the North Atlantic and Southern Ocean, spreading over the globe at depth and replaced by lighter surface waters from lower latitudes.

The imprint of the physical circulation is seen in how a range of physical and biogeochemical properties resemble each other: plumes of fresh and nutrient-rich water spread northward from the Southern Ocean and the thickness of the relatively warm and nutrient-depleted, thermocline waters undulate together; as illustrated in the meridional sections in colour plates 2 to 6.

Phytoplankton absorb visible wavelengths of light and use the energy, along with nutrients containing essential elements, to create new organic molecules or reproduce by producing a copy of the cell. The organic matter provides the ultimate source of energy and nutrients for all other living creatures in the ocean. Phytoplankton growth leads to the consumption of inorganic nutrients in surface waters; some of the organic matter gravitationally sinks through the water column and is respired to regenerate inorganic nutrients at depth.

Carbon is stored within the ocean predominantly as dissolved inorganic carbon, consisting mainly of bicarbonate and carbonate ions with less than one per cent held as dissolved carbon dioxide. The ocean inventory of carbon is typically fifty times larger than the atmospheric inventory of carbon dioxide.

Carbon dioxide is exchanged between the atmosphere and ocean: the ocean takes up carbon dioxide when surface waters cool and sink, and when phytoplankton grow, forming new organic matter. Conversely, carbon dioxide is returned to the atmosphere when surface waters warm and when carbon and nutrient-rich deep waters are returned to the surface. Carbon is transferred into the ocean interior by the physical transfer of cold, carbon-rich surface waters and by the biological transfer of organic matter, falling through the water column and being respired at depth.

Following this descriptive overview, these themes are taken forward in a set of fundamental chapters addressing the controlling processes in more detail: how tracers are transported by the circulation; how the ocean circulates and is forced by the atmosphere; how phytoplankton cells grow and their implications for biogeochemistry; and how carbon dioxide is cycled in the ocean and exchanged with the atmosphere.

2.6 Questions

Q2.1. Heat storage of the atmosphere and ocean.

Estimate the thickness of the ocean that holds as much heat as the overlying atmosphere, where the amount of heat Q required to raise the temperature of the atmosphere or ocean by ΔT is given by,

$$Q = \rho C_p A D \Delta T, \tag{2.2}$$

where ρ is density ($\mathrm{kg\,m^{-3}}$), C_p is heat capacity ($\mathrm{J\,kg^{-1}\,K^{-1}}$), A is horizontal area ($\mathrm{m^2}$), and D is the vertical scale (m).

Assume $\rho \sim 1\,\mathrm{kg\,m^{-3}}$ for the atmosphere and $10^3\,\mathrm{kg\,m^{-3}}$ for the ocean, $C_p \sim 1000\,\mathrm{J\,kg^{-1}\,K^{-1}}$ for the atmosphere and $4000\,\mathrm{J\,kg^{-1}\,K}$ for the ocean, a vertical scale, D, of 10 km for the atmosphere (where the bulk of the atmosphere resides), $\Delta T = 1\,\mathrm{K}$ and a horizontal area $A = 1\,\mathrm{m^2}$.

Q2.2. Radiative heating and equilibrium temperature.

(a) For a planet with no atmosphere, derive how the equilibrium temperature, T, in kelvin, depends on the incident solar radiation, S_c, and the albedo, α, the fraction of reflected sunlight,

$$T = \left(\frac{(1-\alpha)S_c}{4\sigma_{sb}}\right)^{1/4}, \tag{2.3}$$

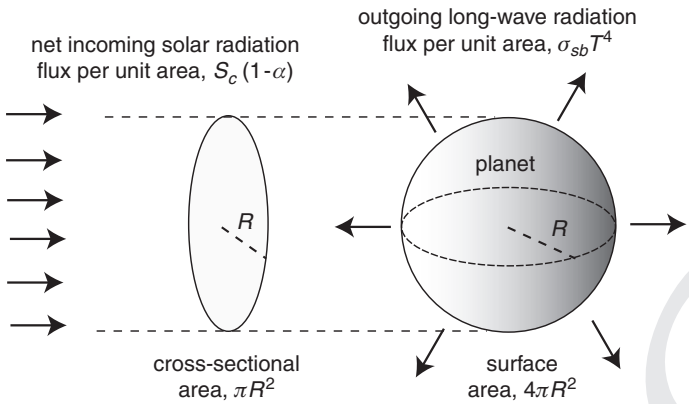

Figure 2.25 A schematic figure of the net incident solar radiation per unit horizontal area, the incident minus the reflected, $S_c(1-\alpha)$, which is absorbed over the cross-sectional area of a planet with radius R. The outgoing long-wave radiation, $\sigma_{sb}T^4$, is radiated over the entire surface area of the planet.

where σ_{sb} is the Stefan-Boltzmann constant. Assume a radiative balance, as depicted in Fig. 2.25, where (i) the net solar radiation is absorbed over a circular disc with a cross-sectional area of the planet, and (ii) the outgoing long-wave radiation per unit horizontal area in $\mathrm{W\,m^{-2}}$ is given by the Stefan-Boltzmann law, $\sigma_{sb}T^4$, integrated over the surface area of the planet.

(b) Estimate this equilibrium temperature in kelvin for Venus, Earth and Mars assuming that S_c is 2600, 1400 and 590 $\mathrm{W\,m^{-2}}$, and their albedos, α, are 0.8, 0.3 and 0.15, respectively, and $\sigma_{sb} = 5.7 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$. How do these temperatures compare with their respective observed surface values of typically 750 K, 280 K and 220 K? Why might there be a mismatch in some cases?

(c) If the planet is now assumed to have an atmosphere that is transparent to solar radiation, but absorbs and re-radiates long-wave radiation, then a local radiative balance suggests that the absorbed solar and long-wave radiation at the ground balances the outgoing long-wave radiation. The surface temperature is then given by

$$T = \left(\frac{(1-\alpha)S_c}{2\sigma_{sb}}\right)^{1/4}. \tag{2.4}$$

Use this relationship to estimate the implied temperature contrast between the tropics and the high latitudes. For simplicity, in the tropics, assume that the incident radiation is given by S_c, while at the high latitudes, the incident radiation is given by $S_c/3$. How does this estimate compare with the actual meridional temperature contrast of typically 30 K for the Earth?

Q2.3. Anthropogenic heating of the ocean by the increase in atmospheric CO_2.

Increasing atmospheric CO_2 leads to increasing radiative heating, $\Delta\mathcal{H}$ (in $\mathrm{W\,m^{-2}}$), which varies logarithmically with the increase in mixing ratio for atmospheric CO_2 (as the effect of increasing CO_2 on the absorption and emission of long-wave radiation gradually saturates),

$$\Delta\mathcal{H} = \alpha_r \ln\left(X_{CO_2}(t)/X_{CO_2}(t_0)\right), \tag{2.5}$$

where $\alpha_r = 5.4\,\mathrm{W\,m^{-2}}$ depends on the chemical composition of the atmosphere and $X_{CO_2}(t_0)$ and $X_{CO_2}(t)$ are the mixing ratios for CO_2 at times t_0 to t.

(a) Estimate the increase in implied radiative heating, $\Delta\mathcal{H}$, over the 50 years between 1958 and 2008 assuming an increase in X_{CO_2} from 315 ppmv to 386 ppmv; compare your answer with Fig. 1.11b.

(b) Given these estimates of radiative heating, then estimate how much the upper ocean might warm over 50 years. Assume that the temperature rise of the ocean is given from a simple heat balance by

$$\Delta T \sim \frac{\overline{\Delta\mathcal{H}}\,t}{\rho C_p h},$$

where $\overline{\Delta\mathcal{H}}$ is the average extra heating over the time period, t, of 50 years (convert to seconds) and h is the thickness of the upper ocean, taken as 1000 m; ρ and C_p are as in Q2.1. Compare this estimate with the reported change for the global warming of the Earth over the last 50 years (IPCC, 2007).

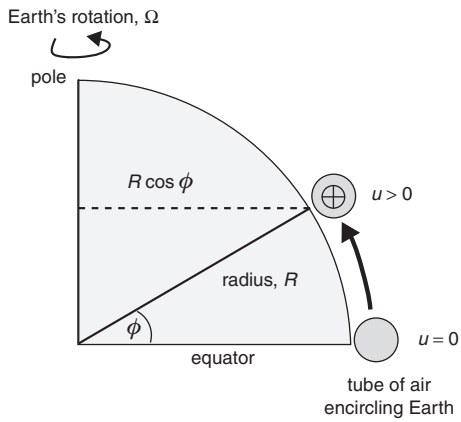

Figure 2.26 A schematic figure depicting a tube of air (dark shading) encircling the Earth along a latitude circle with the Earth rotating at an angular velocity Ω. The tube is at a distance $R\cos\phi$ from the rotational axis where R is the radius and ϕ is the latitude. As the tube moves from the equator towards the pole, the tube increases its zonal velocity, u, so as to conserve angular momentum. Adapted from Green (1981).

Q2.4. Atmospheric zonal jets and angular momentum.

Consider a tube of air circling the Earth at its equator that is uniformly displaced poleward, as depicted in Fig. 2.26. The angular momentum of the tube is given by

$$L_{ang} = (u + \Omega R\cos\phi)\, R\cos\phi, \tag{2.6}$$

where u is the zonal velocity, Ω is the angular velocity, R is the radius of the Earth, and ϕ is the latitude. $R\cos\phi$ represents the effective radius of the tube to its rotational axis, $\Omega R\cos\phi$ represents the velocity of the spinning Earth relative to a fixed point in space and u represents the velocity of the air relative to the Earth.

(a) Derive an expression giving the zonal velocity, u, as a function of latitude, ϕ, by assuming that angular momentum L_{ang} is conserved and the initial zonal velocity at the equator is zero.

(b) Calculate the implied zonal velocity for every 10° from the equator to 30°N for the Earth, assuming $\Omega = 2\pi/\text{day}$ and $R = 6340\,\text{km}$. What are the implications of your result?

2.7 Recommended reading

A descriptive view of the ocean circulation is provided in L. D. Talley, G. L. Pickard, W. J. Emery and J. H. Swift, *Descriptive Physical Oceanography: An Introduction*. 6th edition, Academic Press (due for publication 2011).

A comprehensive view of the ocean, including observations, theory and modelling, is provided in G. Siedler, J. Church and J. Gould (2001). *Ocean Circulation and Climate: Observing and Modelling the Global Ocean*. San Diego, CA: Academic Press, 693pp.

An introductory view of the physical processes operating in the atmosphere and ocean, drawing on laboratory experiments and relevant theory, is provided by J. Marshall and R. A. Plumb (2008). *Atmosphere, Ocean and Climate Dynamics: An Introductory Text*. Burlington, MA: Academic Press Elsevier, 319pp.

An introductory view of marine chemical processes is provided by S. Emerson and J. Hedges (2008). *Chemical Oceanography and the Marine Carbon Cycle*. Cambridge: Cambridge University Press, 468pp.

An advanced view of the biogeochemistry and carbon cycle is provided by J. L. Sarmiento and N. Gruber (2006). *Ocean Biogeochemical Dynamics*. Princeton, NJ: Princeton University Press, 526pp.