Ocean Circulation

Lynne D Talley

Volume 1, **The Earth system: physical and chemical dimensions of global environmental change**,
pp 557–579

Edited by

Dr Michael C MacCracken and Dr John S Perry

in

Encyclopedia of Global Environmental Change
(ISBN 0-471-97796-9)

Editor-in-Chief

Ted Munn

Ocean Circulation

Lynne D Talley

Scripps Institution of Oceanography, University of California, La Jolla, CA, USA

Ocean circulation is one of the major elements of climate, moving heat, freshwater, nutrients and dissolved gases. Ocean circulation at the largest spatial scales and long time scales contains many elements that are in common in the different oceans. Understanding the processes that drive these different elements has progressed a long way and is essential for verifying climate models. The ocean circulation is forced (1) by the winds, through a thin frictional layer called the Ekman layer and through the response of the interior ocean to mass convergences and divergences in this layer, and (2) by heating and cooling that change the temperature and processes that change the amount of fresh water and hence the salinity. The first type of circulation is wind driven and the second is the thermohaline circulation.

The prevailing, large-scale winds of the Earth are trades (easterlies) in the tropics, westerlies at mid-latitudes. The wind-driven ocean circulation that results contains subtropical gyres around ocean high pressure regions and subpolar gyres around low pressure regions. These gyres are asymmetric, with strong western boundary currents (velocities of about $1\,m\,s^{-1}$) that extend to great depth and relatively weak currents elsewhere. With increasing depth, each subtropical gyre shrinks poleward and westward towards its western boundary current. Shallow eastern boundary currents are created by local upwelling along eastern boundaries, driven by equatorward winds. In the tropics, the trade winds drive westward surface flow and create a strong and very thin eastward-flowing undercurrent on the equator. At the latitude of Drake Passage, between South America and Antarctica, the ocean is open all the way around the Earth. The westerly winds here create the strong, deep-reaching eastward flowing Antarctic Circumpolar Current.

Except in the strong western boundary currents and the Antarctic Circumpolar Current, the deep flow is driven by thermohaline forcing, in which dense waters are made in isolated locations at high latitudes and spread through the oceans via relatively strong deep western boundary currents, which differ in mechanism and often direction from the wind-driven western boundary currents. Abyssal flow away from the deep western boundary currents is dominated by topography, and is often found to move parallel to the topography, with low pressure in the center.

Ocean currents transport heat from the tropics to the poles. Most of the heat is lost at mid-latitudes, where vigorous western boundary currents bring warm water to the latitude of cold, dry air outbreaks from the continents. This part of the heat transport is associated with the shallowest, wind-driven part of the subtropical gyre circulations (Talley, 1999), where the thermohaline circulation, forced by heating and cooling, is overshadowed by the wind-driven circulation. The deep thermohaline circulation is asymmetric, with deep water formation only in the northern North Atlantic and its adjacent seas and along the margins of Antarctica. No deep water is formed at the sea surface in the North Pacific or northern Indian Ocean. The global thermohaline circulation thus consists of two intersecting cells, one with sinking in the North Atlantic and the other with sinking in the Antarctic.

INTRODUCTION

The circulation of the ocean has been of practical interest to fishermen, traders and navies for as long as humans have gone to sea. As knowledge about ocean currents and capabilities to observe it below the surface grew, curiosity about currents below the sea surface resulted in increasingly detailed descriptions of the ocean as a three dimensional fluid. These observations have led to an improved understanding of the processes that govern the circulation. In this day of concern about the Earth's climate and our burgeoning ability to understand and model it with ever better computer resolution and power, knowledge of ocean circulation is as essential as knowledge of atmospheric circulation and the biosphere. In this article, the basic circulation patterns as we understand them now are described briefly including some historical perspective on this knowledge, followed by a section describing the physical processes that we understand to govern the circulation. In the last section, ideas about the role of ocean circulation in climate are presented.

Fluid flows, including those of the ocean circulation, are continuous and turbulent, and contain a wide range of behaviors from very small waves to the global circulation that is described here. Thus approach them in a simplified way, focusing on the time and space scales for considering the potential for and character of global environmental change. What is described here is often called the *general circulation*, meaning that it is the circulation at the largest space scales and longest time scales. Most elements of the general circulation are nearly unchanging in time within a given major climate regime; while given currents may be somewhat stronger or weaker in different seasons or years, or found in a slightly different location, the currents are always present in some form. Thus, we can hope to understand the underlying reasons for their existence in the present interglacial state and therefore be able to surmise the circulation patterns in other major climate regimes.

Based on our understanding of the oceans, we have learned that, even with the massive climate shifts associated with glacial/interglacial switches, the basic elements of the ocean circulation would not have changed, with western boundary currents, wind driven gyres, eastern boundary currents, tropical circulation, and thermohaline circulation driven by dense water formation. However, strengths, depths, and horizontal positions of ocean currents would have shifted with large-scale shifts of wind patterns and changes in the relative strength of thermohaline forcing in each ocean basin. The circulation with quite different ocean geometry resulting from shifting continents can also be surmised given understanding of the processes that govern the present circulation.

The space scales of the ocean circulation range from about 100 km, which is the width of a strong current like the Gulf Stream off the east coast of North America, to several thousand kilometers, which is the width of the gyres that extend across each ocean basin, to global scale. Speeds of horizontal currents range from less than a centimeter per second in some areas of the deep ocean and far from the ocean sides to about $100 \, \mathrm{cm \, s^{-1}}$ ($1 \, \mathrm{m \, s^{-1}}$) in the strongest currents such as the Gulf Stream. Because the ocean depth averages around 5 km, and is at most about 10 km, the vertical circulation is limited and is in fact so much slower than the horizontal circulation that it is almost impossible to measure. Nevertheless, this very weak vertical circulation can be important for connecting the layers of the ocean, particularly for the circulation forced by water density changes.

Ocean currents are geostrophic (see section on processes below), like the large-scale atmospheric patterns that are associated with high and low pressure centers (*see* **Atmospheric Motions**, Volume 1). The force due to the difference in pressure pushes water from high to low pressure. However, because the Earth rotates, the water turns to the right in the Northern Hemisphere, and therefore circulates clockwise around high pressure centers and counterclockwise around lows; this tendency is due to the Coriolis acceleration (*see* **Coriolis Effect**, Volume 1). In the Southern Hemisphere, the water turns to the left and hence flows counterclockwise around lows, etc. High pressure at the sea surface in the ocean is due to water being piled up there, so the sea surface is slightly higher, up to one meter, than in the low pressure regions. Most surface height differences that drive the ocean circulation are much smaller, on the order of 5–10 cm. Spatial atmospheric pressure variations are too small to drive ocean circulation.

The ocean's circulation is forced almost exclusively at the sea surface by the winds and by changes in water density resulting from heating, cooling, evaporation and precipitation. These forcings create the high and low pressures through flows that are not geostrophic. We refer to the first type of forcing as *wind forcing*. The second is often called *buoyancy forcing*, because making water denser makes it less buoyant, or *thermohaline forcing*, because it results from changes in temperature and/or salinity (*see* **Thermohaline Circulation**, Volume 1). The ocean circulation can be reasonably well separated into portions that result from wind and from thermohaline forcing. Wind driven circulation is mainly confined to ocean basins while some parts of the thermohaline circulation extend from one end of the ocean to the other. Wind forcing dominates the surface circulation and creates the strongest currents, carrying the largest volumes of water, but thermohaline circulation dominates in the deep ocean. The wind driven and thermohaline circulations are of course connected with each other; a given water parcel will be subject to both forcings. The elements of the circulations are described in the next two sections. Our understanding of how the circulation is forced is described in more depth in the section on physical processes.

SURFACE CIRCULATION

The circulation at the ocean's surface is mainly produced by the winds. Mapping attempts have been made throughout the history of navigation, with a major growth in description and insight in the 17th century in the wake of global navigation. An excellent review, replete with reproductions of the significant charts of surface circulation, can be found in Peterson *et al.* (1996). The first to make the connection between winds and ocean currents was the German physician Varen (1622–1650), followed by the first global mapping and description of mid-latitude subtropical gyres in each ocean by the Dutch scholar Vos (1618–1689). Production of various schemes of surface circulation blossomed in the 19th century, with various degrees of realism, following invention of the chronometer and ability to determine longitude. Sobel (1995) provides a popular account of the British government's quest to solve the problem of measuring longitude, and the nearly unrewarded success of British clockmaker John Harrison in his lifelong work on the marine chronometer, a precise clock that worked even on moving and rolling ships.

The strongest currents that are part of the general circulation are about $1 \, \mathrm{m \, s^{-1}}$, or about 2 knots (1 knot = 1 nautical mile $\mathrm{h^{-1}}$). Tidal currents can exceed this, but are not found at this strength in the open ocean (*see* **Tides, Oceanic**, Volume 1). Currents of this strength can impact the progress of ships. An early map of the Gulf Stream (Figure 1a) was produced by Benjamin Franklin for the mail service from Britain to America, based on observations by his cousin Captain Timothy Folger, who was a whaling captain based in Nantucket. The original map was unearthed in an archive in Paris (Richardson, 1980) and is remarkably accurate, unlike a miscopied version that
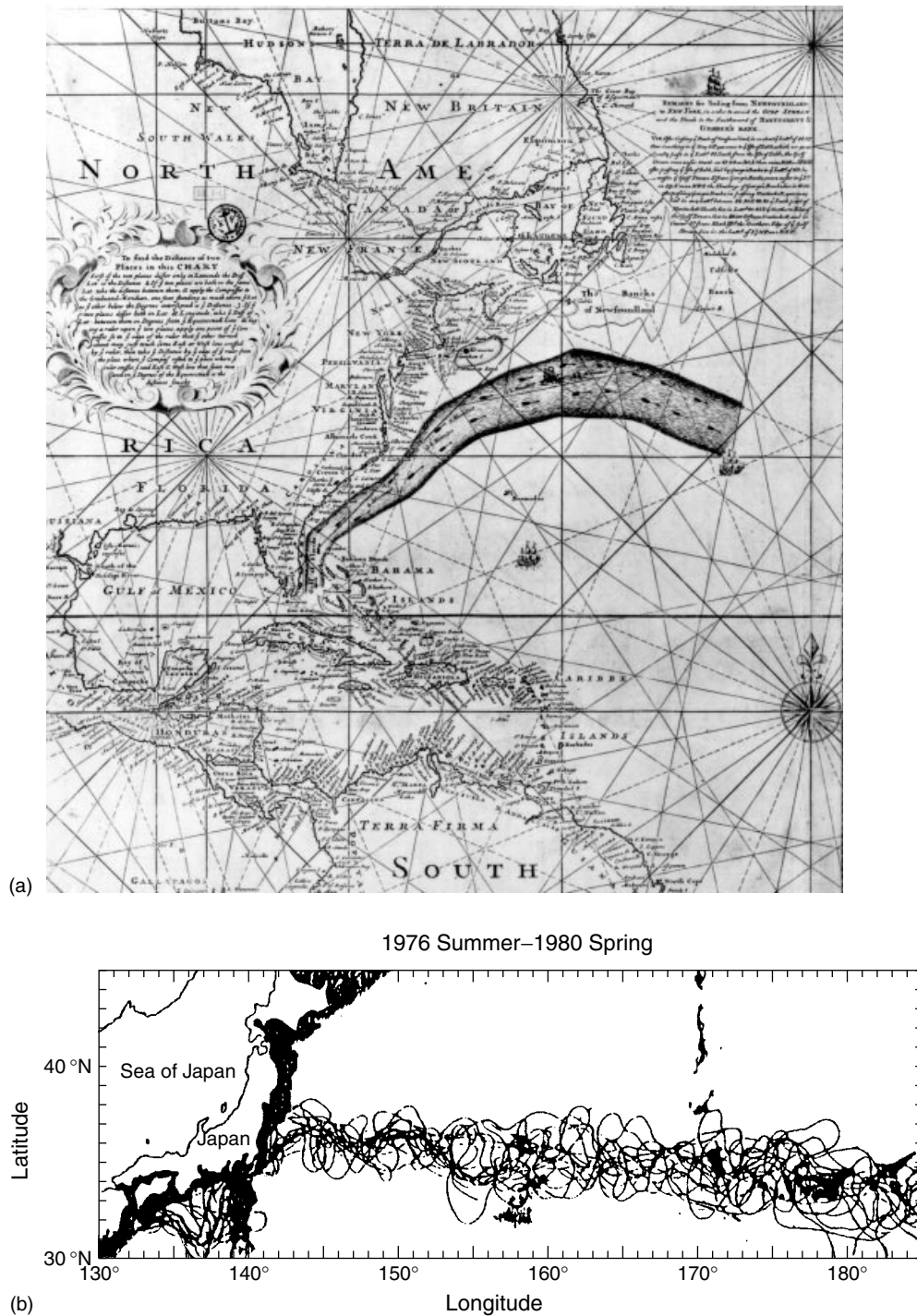
(a)

1976 Summer−1980 Spring

(b)

Figure 1 (a) Map of the Gulf Stream, produced by Benjamin Franklin based on information from Captain Timothy Folger, and rediscovered in a Paris archive by Richardson (1980). (b) Meanders of the Kuroshio after it separates from the Japanese coast (Mizuno and White, 1983). (c) Shown as a bright red band, the Gulf Stream is about 27 °C (∼80 F) in this sea surface temperature image of the western North Atlantic during the first week of June 1984. There is a large temperature difference between the Gulf Stream and the surrounding waters and so the current and it meanders and rings are visible in sea surface temperature. This image is based on data from NOAA-7 Advanced Very High Resolution Radiometer (AVHRR) infrared observations. Warmer hues denote warmer temperatures. (Courtesy of O. Brown, R. Evans and M. Carle, University of Miami, Rosenstiel School of Marine and Atmospheric Science, Miami, Florida.)
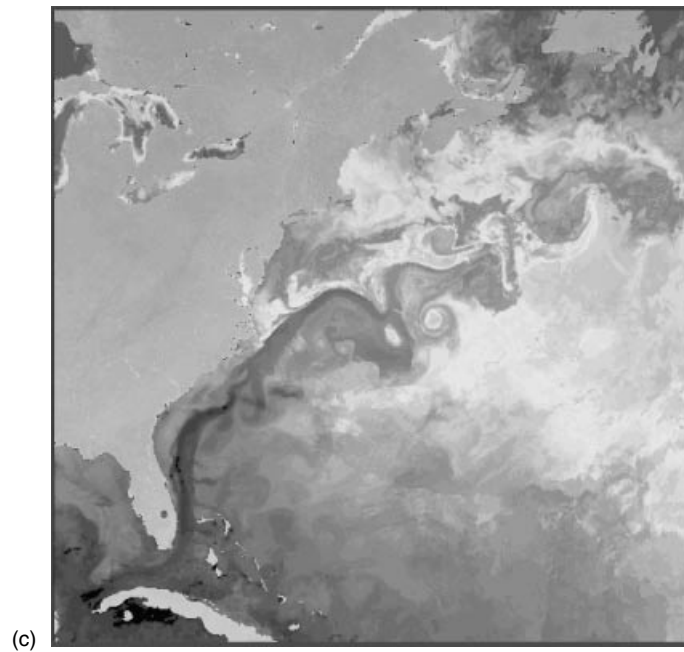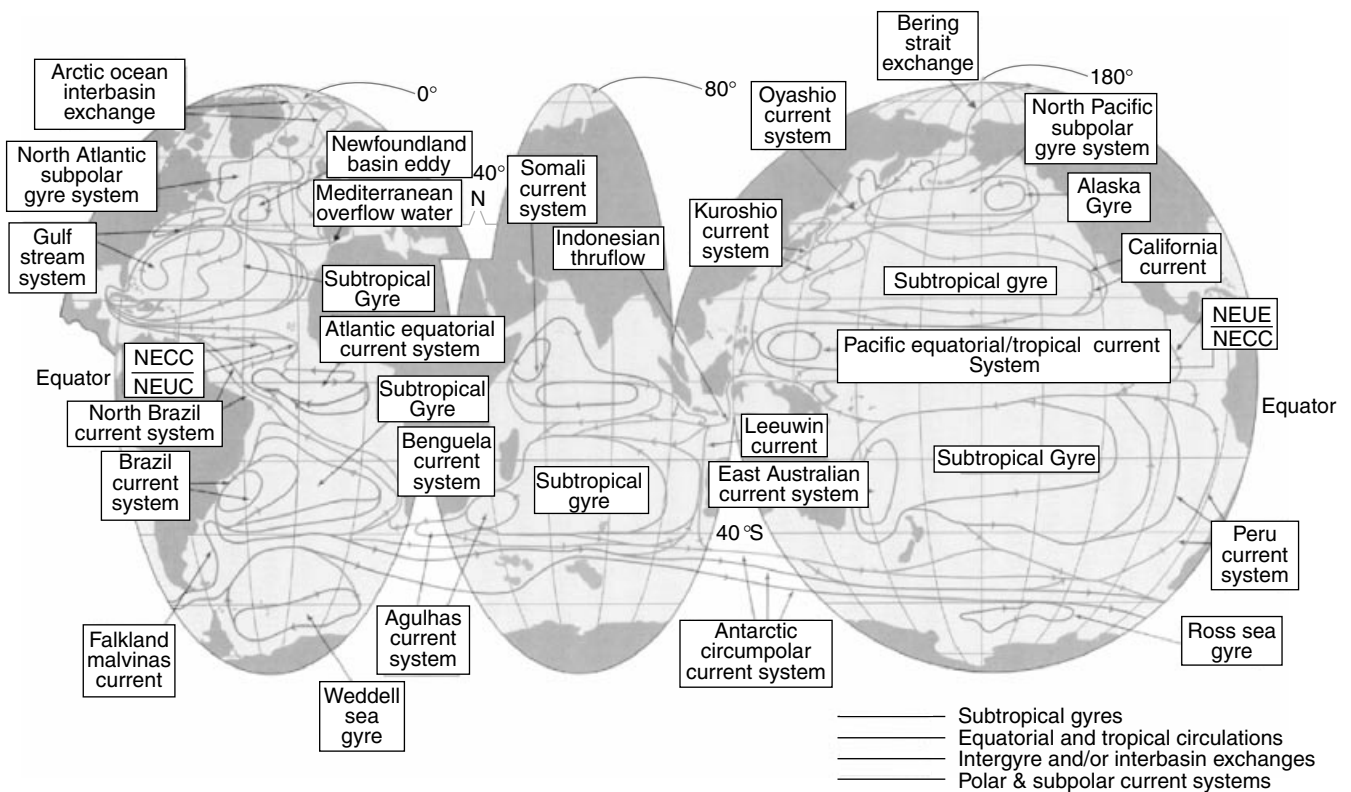
(c)

Figure 1   (*Continued*)

Figure 2   Schematic of the surface circulation of the world ocean (after Schmitz, 1996). NECC stands for North Equatorial Countercurrent, and NEUC stands for North Equatorial Undercurrent

had been widely circulated and which placed the Gulf Stream in the wrong location (*see* **Franklin, Benjamin**, Volume 1).

The surface circulation for the globe has been summarized in schematic form by Schmitz (1996), and is reproduced here with some changes (Figure 2). Each of the largest ocean basins (North and South Atlantic, North and South Pacific, Indian) has a subtropical gyre, in which the currents circulate clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. These gyres extend all the way from the western to the eastern boundary. The subtropical gyres are driven by westerly winds at mid-latitudes and easterly trade winds equatorward of about 30°latitude. The communication of this forcing to the general circulation is roundabout and is described in the section on physical processes below.

The subtropical gyres are asymmetric; the currents at the western boundaries are much stronger in all oceans than anywhere else in the gyres. These western boundary currents are the fastest currents in the ocean circulation. (Some tidal and strait flows are faster.) The western boundary currents include the Gulf Stream (North Atlantic), the Kuroshio (North Pacific), the Brazil Current (South Atlantic), the East Australia Current (South Pacific) and the Agulhas Current (Indian). All of these subtropical western boundary currents carry warm water away from the tropics towards the cooler regions at mid-latitudes; each loses a tremendous amount of heat during the process. In fact, the ocean's total heat loss is dominated by the losses in these western boundary currents. (The ocean's total heat gain is dominated by input from the Sun in the tropics.) As can be seen from the Franklin Gulf Stream chart (Figure 1a) and the modern satellite image of the Gulf Stream (Figure 1c), the Gulf Stream separates from the coast at North Carolina and moves out into the open ocean, remaining a strong current for a long distance out to sea. This behavior is characteristic of all of the other western boundary currents as well, in that all of them flow strongly along the boundary, then separate and flow offshore as strong and narrow currents for about 1000 km.

Western boundary currents are approximately 100 km in width, even after they separate from the coast and flow out to sea. After the currents leave the coast, they retain their width, but they also meander widely before they finally lose their energy. The meanders fill a wide swath, which for the Gulf Stream is almost exactly the envelope shown in the Franklin/Folger chart (Figure 1a). The meandering paths of the Kuroshio are shown in (Figure 1b). The meanders are about the same length and fill about the same envelope most of the time but are very dynamic, with a time-scale for change of two to four weeks. The meanders pinch off regularly both north and south of the current axis to form rings, which are 100–200 km across (e.g., Gulf Stream rings seen in Figure 1c). The rings that form on the north side of the current contain water from the south, warm side of the current and so are called *warm core rings*, while rings on the south side contain cold water and are called *cold core rings*. The rings generally migrate westward and sometimes rejoin the current.

All of the western boundary currents extend to the ocean bottom when they are along the boundary. The western boundary is not a vertical wall but is rather the continental slope between the land and the abyssal ocean. Before the separation point, the boundary current rides along the slope, not in the deepest water. After separation, the boundary currents flow out into the deeper part of the ocean and many of them then extend down to the ocean bottom at 4000–6000 m due to their tremendous energy. The highest speeds are always at the ocean surface, with decay to very low speeds at depth, reflecting the source of the western boundary currents as part of the wind driven circulation.

While every ocean has a subtropical gyre with a western boundary current, each western boundary current has its own peculiarities, arising from the shape or existence of the western boundary and its intersection with the part of the wind pattern that dictates the separation point (see the section on physical processes below). The Brazil Current of the South Atlantic is the most canonical, following a single boundary until the separation point. The North Pacific's Kuroshio is nearly as simple, although it is complicated along the western boundary because of the numerous island chains and marginal seas. In the North Atlantic, there are actually two subtropical western boundary currents: the Gulf Stream, which we have discussed, and the North Atlantic Current, which is in a sense a continuation of a portion of the Gulf Stream, with the coast of Newfoundland as its western boundary because the winds dictate that the subtropical gyre should extend as far northward as Newfoundland. In the South Pacific, the East Australia Current flows southward to the latitude of New Zealand and then shoots straight eastward across the Tasman Sea to the northern tip of New Zealand, where it then recommences its southward flow for a short distance as a western boundary current before separation. In the Indian Ocean, the Agulhas flows southward to the Cape of Good Hope and would keep going if Africa extended farther south because the wind pattern does not dictate a separation until farther south. So the Agulhas essentially runs past the western boundary and makes a tremendous loop south of Africa back towards the east into the Indian Ocean. This is called the Agulhas retroflection. The retroflection is highly unstable and large rings of Indian Ocean water are produced at the place where it curves back. These rings move westward into the South Atlantic and are a major source of Indian Ocean water for the Atlantic Ocean.

The strength of the western boundary currents is related to the strength of the wind forcing that drives the subtropical gyres. A useful measure of the strength of a current is the total amount of water that it transports, which means looking for the maximum depth and horizontal extent of the current and computing the net flux of water. The transport is the sum (integral) of all of the velocities through a chosen area. The usual unit for ocean transport is the Sverdrup, where 1 Sverdrup $= 1 \times 10^6 \, \mathrm{m^3 \, s^{-1}}$, which is velocity times area. Subtropical western boundary currents typically carry 60–100 Sverdrups. All of this water must be returned equatorward in the interior of the ocean, away from the boundary, and then rejoin the western boundary current.

Most of the area of the subtropical gyres is occupied by the broad, slow eastward flow on the poleward side of the gyre, broad, slow equatorward flow all the way across the ocean, and broad, slow westward flow on the equatorward side of the gyre. The eastward flows are sometimes referred to as the west wind drift; in the Northern Hemisphere they are also referred to as the North Atlantic Current and the North Pacific Current. The westward flows in both Northern Hemisphere oceans are called the North Equatorial Current, not because they are along the equator but because they are on the tropical side of the gyres. The westward flows in the Southern Hemisphere oceans are called the South Equatorial Current in each ocean. The eastward and westward slow flows are divided at the sea surface by a narrow front called the Subtropical Front, which is at around 30°latitude.

An important feature of the upper ocean subtropical circulation, called *subduction*, is associated with the density structure of the ocean, in which density increases with depth, and surface density increases with distance from the equator, that is, surface water is warmest in the tropics and colder at higher latitudes. When the warm surface water flows poleward in the western boundary current, it is subjected to intense cooling. The water that emerges from the separated western boundary current is then cooler than the water to the south. This emerging water circulates back towards the south, but when it does so, it encounters warmer surface water. The warming rate at the surface is not large enough to change the temperature and density of this equatorward flow, and so it submerges beneath the lighter surface layers. This process occurs for all waters that enter the subtropical gyres and that must make their way back towards the tropics. The submerging process was named subduction by Luyten *et al.* (1983) who were the first to fully describe the process and provide a theory for it.

Each subtropical gyre has an eastern boundary current as well as a western boundary current. All eastern boundary currents are significantly weaker and much shallower than western boundary currents. In snapshot observations, eastern boundary currents often resemble a string of highly time dependent eddies, with weak flow moving around and between the eddies. Eastern boundary currents are important for the ocean's biological productivity because they are driven by winds that cause upwelling of nutrient rich waters from about 100 m depth, below the sunlit euphotic zone where most ocean life occurs. The upwelled water is also cool, and is responsible for foggy coastal conditions in the eastern boundary regions. The simple theory for eastern boundary currents is included in the section on physical processes below. The eastern boundary currents are: Canary Current (North Atlantic), California Current (North Pacific), Benguela Current (South Atlantic), Peru Current (South Pacific) and the Leeuwin Current (Indian). Alone among the eastern boundary currents, the Leeuwin Current flows poleward, against the flow of the subtropical gyre.

A second major feature of the surface circulation that occurs in most ocean basins is a subpolar gyre. These are counterclockwise circulations found in the North Atlantic and North Pacific, and clockwise circulations in the Weddell Sea (Atlantic sector of the Antarctic) and Ross Sea (Pacific sector of the Antarctic). The subpolar gyres are driven by westerly winds that are strongest at the subtropical/subpolar gyre and weaker at higher latitudes. The Northern Hemisphere subpolar and subtropical gyres are separated from each other by a narrow front, called the Subarctic or Subpolar Front. The Antarctic region is open to ocean circulation all the way around Antarctica, at the latitude of the maximum westerly winds. A major current, the Antarctic Circumpolar Current, flows eastward around Antarctica at this latitude. Its maximum speeds are about half those of the Gulf Stream. The subpolar circulations in the Southern Hemisphere are south of the Antarctic Circumpolar Current, and are separated from the subtropical circulations by the two main fronts of the Antarctic Circumpolar Current, called the Subantarctic and Polar Fronts (*see* **Southern Ocean**, Volume 1).

The subpolar gyres, like the subtropical gyres, are asymmetric, with strongest circulation at the western boundary. Subpolar circulation extends more vigorously throughout the gyre to the ocean bottom than does subtropical circulation, and so there tend to be focused currents along all boundaries in subpolar gyres, but the western boundary currents are the strongest. The western boundary current in the Weddell Sea is the simplest, flowing northward along the Antarctic Peninsula. The Ross Sea gyre is less confined by a western boundary and part of the flow continues westward west of the Ross Sea along the coast of Antarctica. The North Atlantic's subpolar gyre has two western boundaries, one along Greenland and the other along Labrador. There are western boundary currents in each of these areas, called the East Greenland Current and the Labrador Current. The North Pacific's subpolar gyre has a leaky boundary

due to the many island chains. The main western boundary current is the Oyashio, which flows southward along the southern part of the Kuril Islands and along the coast of Hokkaido before it separates and flows eastward. The western boundary current in the Bering Sea and along the Northern Kuril Islands is referred to as the East Kamchatka Current because it mainly flows along the coast of Kamchatka. There is also a western boundary current along the coast of Sakhalin within the Okhotsk Sea, called the East Sakhalin Current. And finally, the shape of the Gulf of Alaska (the area of the eastern subpolar gyre surrounded by Alaska and Canada) allows a weak western boundary current called the Alaskan Current along the Alaska Peninsula and the eastern Aleutian Islands.

The third major surface circulation feature is the complex circulation in the tropics, which is described separately in the section on tropical circulation, including its vertical structure, which is also complex. At the sea surface, the Pacific and Atlantic tropical circulations include a North Equatorial Current (which was already mentioned as being part of the subtropical gyre circulation and which is not close to being on the equator, but rather is a westward flow north of $10\,^{\circ}$N), the North Equatorial Countercurrent (eastward flow at about $5\,^{\circ}$N), and the South Equatorial Current (westward flow at the surface on the equator and in the Southern Hemisphere).

The surface circulation in the tropical Indian Ocean is the most complex of all of the general circulation patterns because it responds to the strong seasonal monsoonal winds, which reverse direction. West of India, in the Arabian Sea, the circulation is clockwise (around a high pressure center) during the southwest monsoon in late summer, i.e., like a normal Northern Hemisphere subtropical gyre. The northward flowing western boundary current is called the Somali Current. During the Northeast Monsoon, which occurs in late winter, the wind blows from the land to the sea and the Arabian Sea circulation is counterclockwise around a low. The Somali Current then reverses direction and flows southward.

Finally for the surface circulation, the connections between basins and gyres are shown in Figure 2. The most important connecting currents for the global ocean circulation are the westward flow from the Pacific to the Indian Ocean through the Indonesian passages, the northward flow from the Pacific to the Arctic through Bering Strait, and the connection of Agulhas Current waters into the South Atlantic's circulation. These flows are not necessarily part of the wind driven circulation, but are part of the shallow portion of the thermohaline circulation, described below.

THE MID-DEPTH CIRCULATION

The surface circulations described in the previous section do not extend unchanged to depth. One major change

with increasing depth is that current speeds of the general circulation generally decrease considerably. This is most true for currents that are driven by the winds. There is another part of the circulation that is driven by high latitude cooling and resulting deep convection. These currents, although weak in comparison with the surface flows of western boundary currents such as the Gulf Stream, may be strongest close to the ocean bottom. A second major change is that the gyre circulations shift and shrink with depth. A third major change is that topography becomes more important in steering the flows (most apparently for the abyssal circulation described in the section on deep circulation). And fourthly, the equatorial circulation has a large amount of vertical structure (see section on tropical currents below), differing in this way from circulation outside the tropics.

The mid-depth circulation of the Atlantic and Pacific Oceans, from Reid (1994, 1997) is depicted in Figures 3(b) and 4(b). For comparison, Reid's maps for the surface circulation are included because the circulations in Figure 2 were schematic. As of now there is no product like those of Figures 3 and 4 for the Indian Ocean, but many of the comments made here about the subsurface circulation apply also to the Indian Ocean.

The subtropical gyres of all of the four basins (North and South Atlantic, North and South Pacific) in Figures 3 and 4 can be seen to have contracted significantly towards the western boundary and also poleward. This contraction begins just below the sea surface, as can be seen in Reid (1994, 1997), where by 200 m depth, the gyres are already considerably smaller than the surface gyres. What is constant with increasing depth in these gyres is the position of the western boundary current and that the gyres remain asymmetric, with the western boundary current still the most vigorous part of the flow.

The subpolar gyres of the North Pacific and North Atlantic also change with depth. According to Reid (1997) the North Pacific's subpolar gyre becomes more strongly confined to the north, with the subtropical gyre shifting farther north with increasing depth. The North Atlantic's subpolar gyre is not very different at 1500 m than at the surface, although the flow is more strongly constrained by topography because the Mid-Atlantic Ridge encroaches on the northern side of the gyre. The Weddell Sea gyre at 1500 m is about half the strength of the surface Weddell gyre, but with no change in location or horizontal extent. The clockwise Ross Sea gyre is not strongly defined at any level in Reid (1997) in that a western boundary current is not indicated, possibly due to lack of observations. However, the clockwise flow is apparent and is little changed in location at 1500 m compared with the surface.

The Antarctic Circumpolar Current is at about the same location at 1500 m as at the sea surface.

Outside the surface intensified gyres, weaker gyre, which, have no parallels at the sea surface appear at 1500 m in Reid's (1994, 1997) analyses. Most of these gyres have not been studied carefully or mapped outside this particular set of studies. Direct observations of flow in most mid-latitude regions, using either current meters or drifting floats, show that currents are dominated by time dependent eddies rather than by the very slow broad flows depicted in Figures 3 and 4. The primary exception to this eddy dominance is in the tropical region where the flows are nearly due east-west and somewhat faster than in the mid-latitudes (Davis, 1998).

Because the mid-depth circulation outside the tropics is so weak, the most useful measures of its impact and direction may be the total amounts of water moving from one latitude to another, as summarized in the section on global thermohaline circulation.
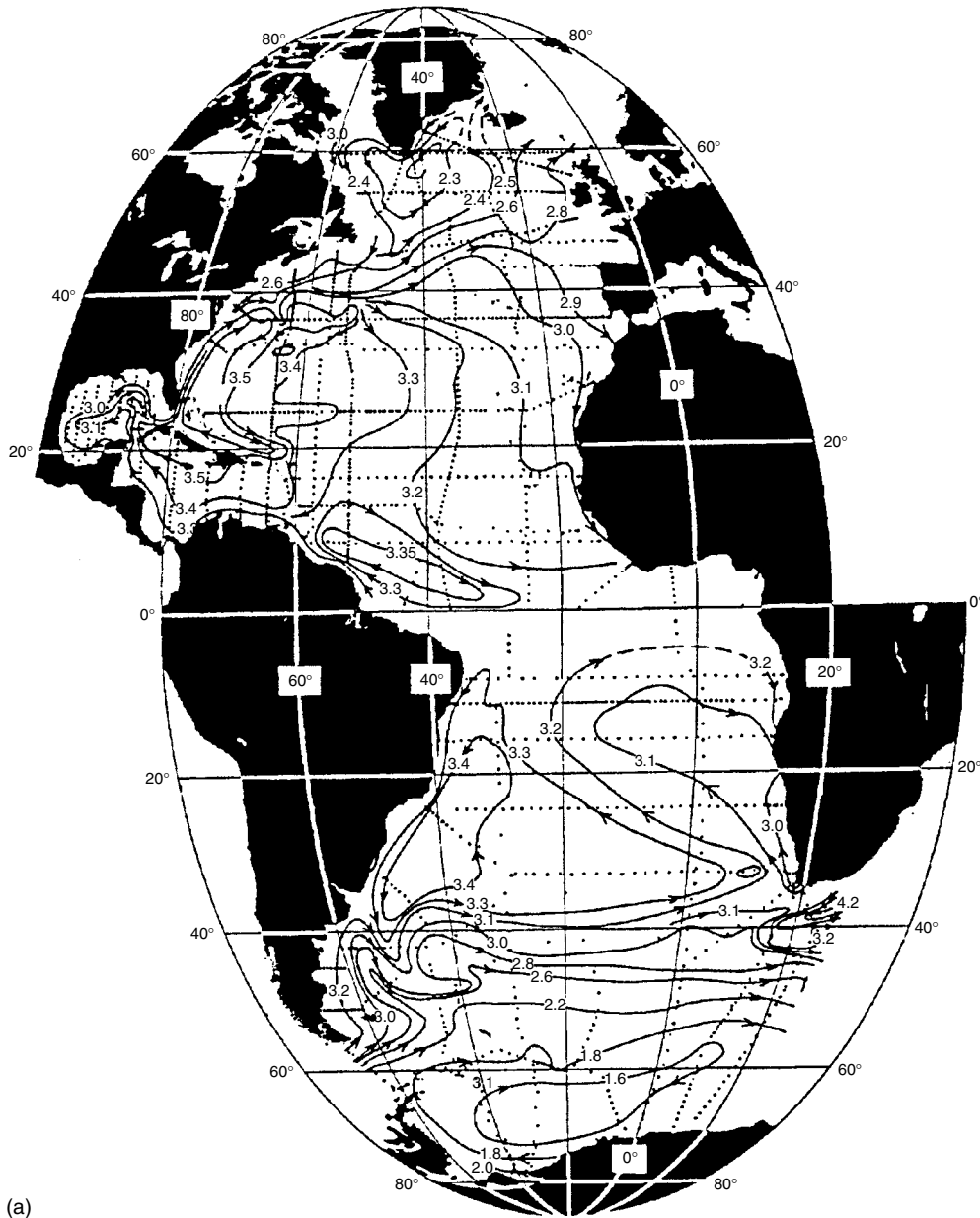

Figure 3 Depictions of the circulation of the Atlantic Ocean at the (a) sea surface, (b) at 1500 m depth, and (c) at 4000 m depth (all from Reid, 1994). The contoured quantity in each map is called the streamfunction. Where the contoured surface is high is the high pressure region described in the text. The general circulation follows the contours, that is, with flow around the highs and lows. The strength of the currents is proportional to the distance between the curves; the closer they are the stronger the current
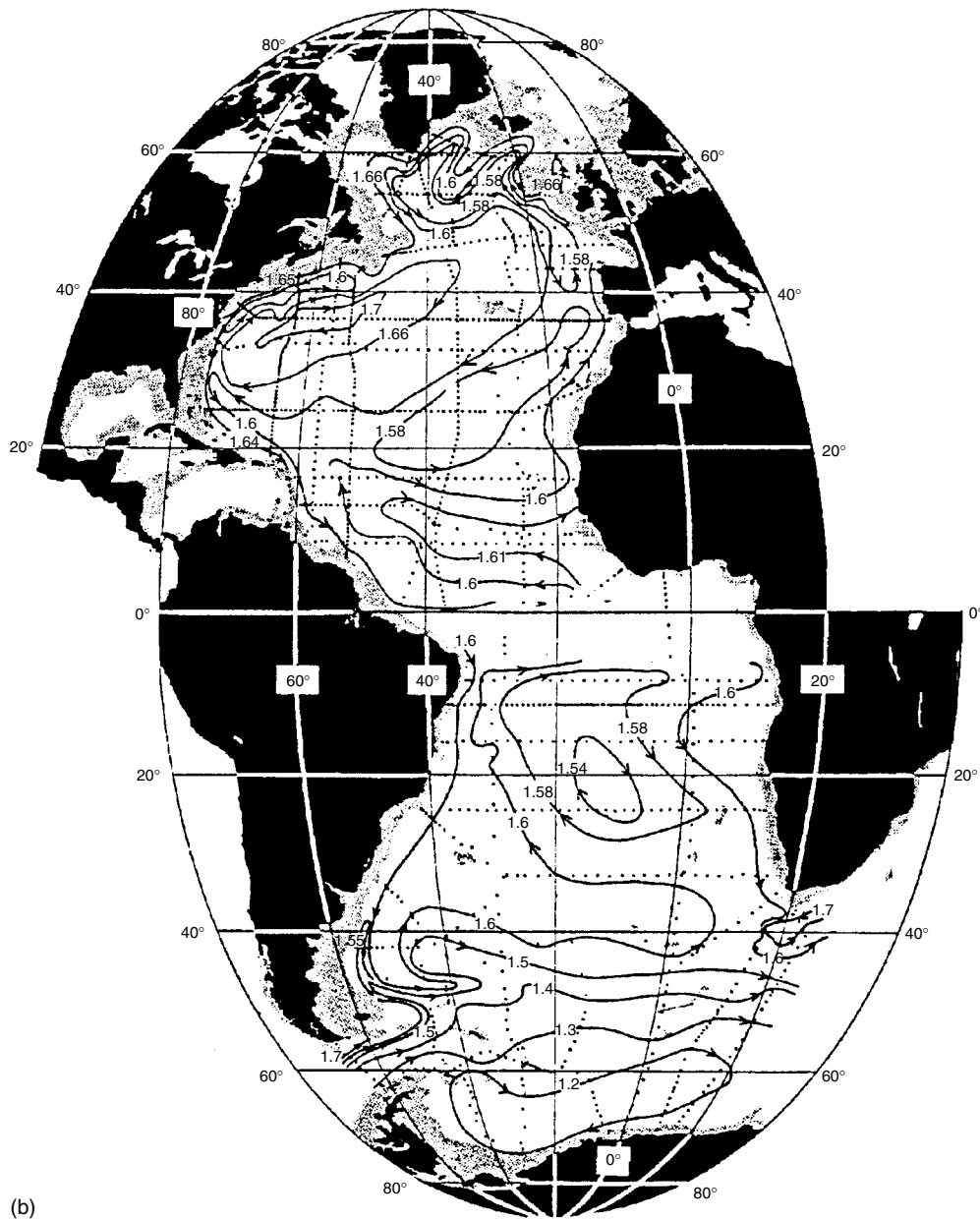
(b)

Figure 3 (*Continued*)

THE DEEP CIRCULATION

Near the ocean bottom, the topography of the ocean bottom significantly affects the circulation, because the mid-ocean ridges and island chains have larger and larger cross-sections. As seen in Figures 3(c) and 4(c) from Reid (1994, 1997) the subtropical gyres shrink all the way to the western boundary and separation region, where the most vigorous surface currents are found and which therefore penetrate to great depth. The subpolar circulations reach to the ocean bottom, but must wind their way around the various topographic obstacles, particularly in the North Atlantic. The Weddell Sea gyre shrinks to the westernmost part of its basin.

The most significantly different aspects of the abyssal ocean's circulation are concentrated western boundary currents that have nothing to do with the surface intensified gyre's western boundary currents. The clearest of these deep western boundary currents in Figures 3(c) and 4(c) is a southward flow along the western boundary of the Atlantic, extending from the northern subpolar region to South America to about 40 °S. This deep flow was first
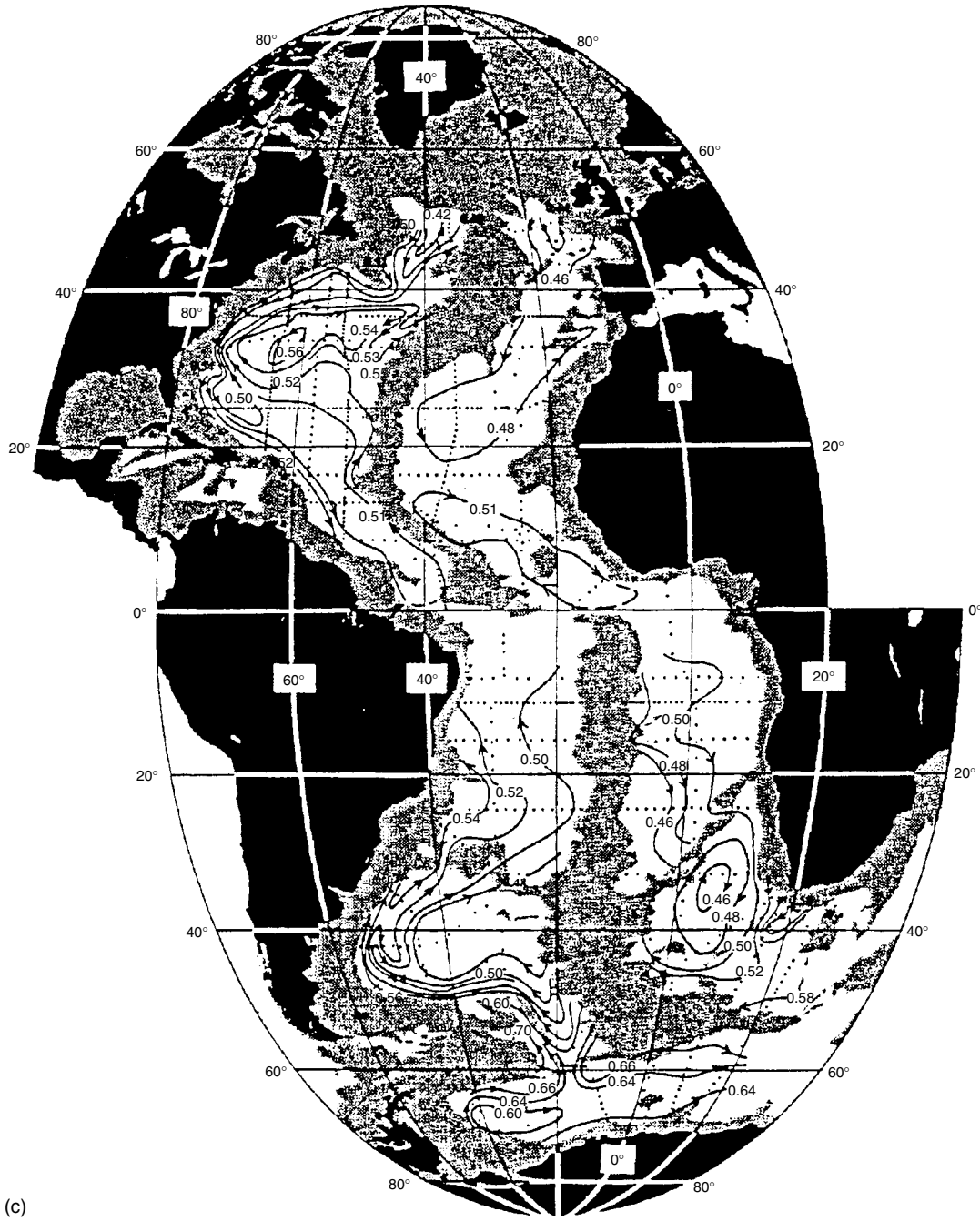
(c)

Figure 3 (*Continued*)

detected by Wüst (1935), although an explanation for its existence did not come along until the early 1960s (Stommel and Arons, 1960). The deep western boundary currents are the most obvious and measurable part of the deep thermohaline circulation, in which deep waters are formed at high latitudes and spread via the boundary currents and very slow interior flow to fill the whole of the deep oceans below about 1500 m depth.

A deep western boundary current flows northward in the Pacific Ocean (Figure 4c), bringing deep water from the Antarctic (of both North Atlantic and Antarctic origin) to the deep South Pacific, across the equator and into the deep North Pacific.

In mid-latitude regions of the Atlantic and South Pacific, the deep flow that is not associated with the deep extension of the subtropical gyres or with the deep western
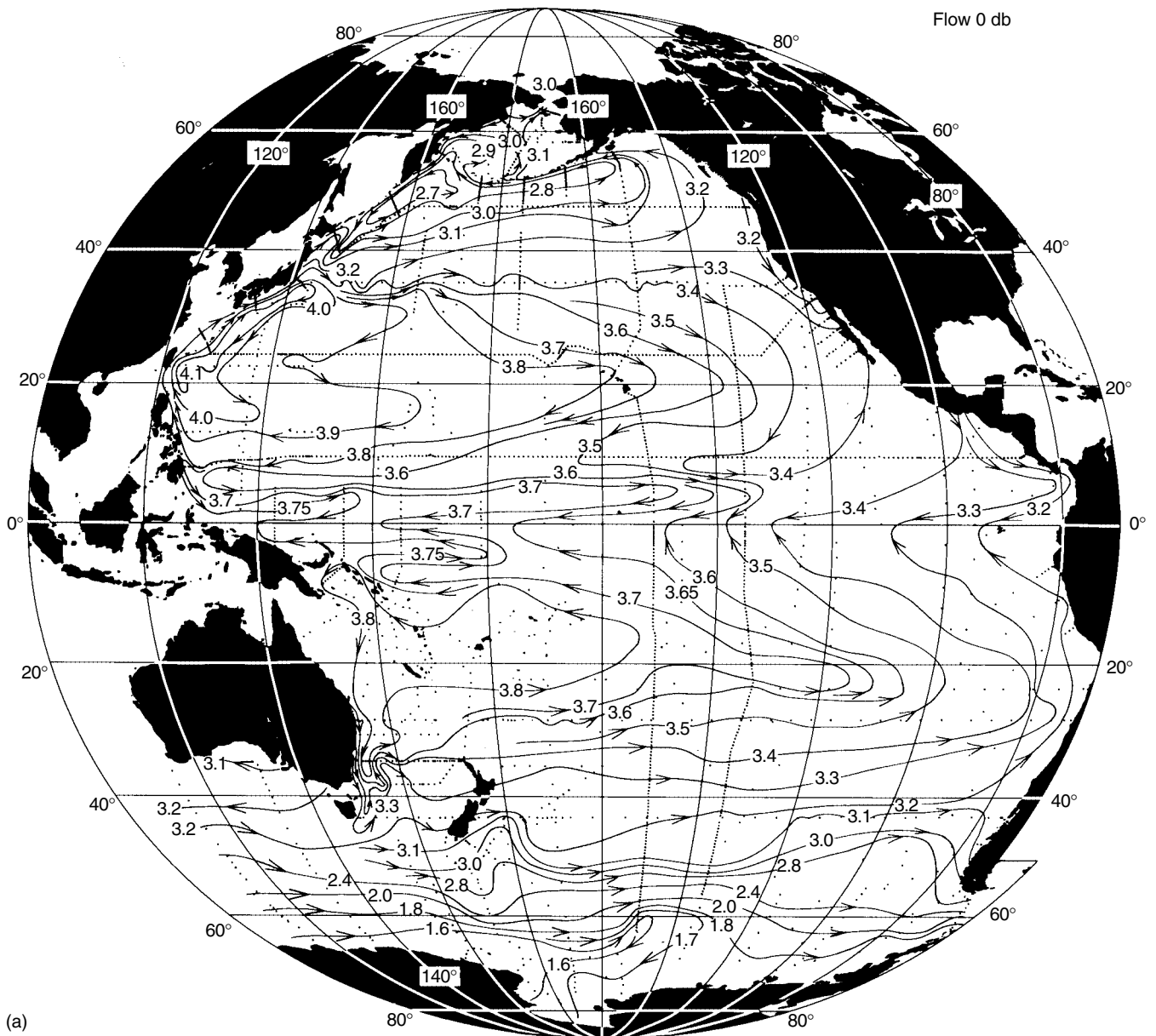
Flow 0 db

(a)

Figure 4   Depictions of the circulation of the Pacific Ocean at the (a) sea surface, (b) at 1500 m depth, and (c) at 4000 m depth, all from Reid (1997). The contoured quantity in each map is the streamfunction. Where the surface is high is the high pressure region described in the text. The general circulation follows the contours, that is, with flow around the highs and lows. The strength of the currents is proportional to the distance between the curves; the closer they are the stronger the current

boundary currents tends to be centered around low pressures (counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere), as shown by Reid in the maps reproduced in Figures 3(c) and 4(c).

THE TROPICAL CIRCULATION

The surface circulation in the equatorial Pacific and Atlantic is westward, and is directly connected to the westward

flow of the Southern Hemisphere subtropical gyres. The westward surface flow is separated from the Northern Hemisphere subtropical gyres by a narrow and strong eastward flow at about 5 °N, called the North Equatorial Countercurrent. This current probably reaches very deep into the ocean. It is the southern side of a very long and narrow counterclockwise circulation. The Southern Hemisphere does not usually have an equatorial countercurrent because the trade wind pattern is different from the
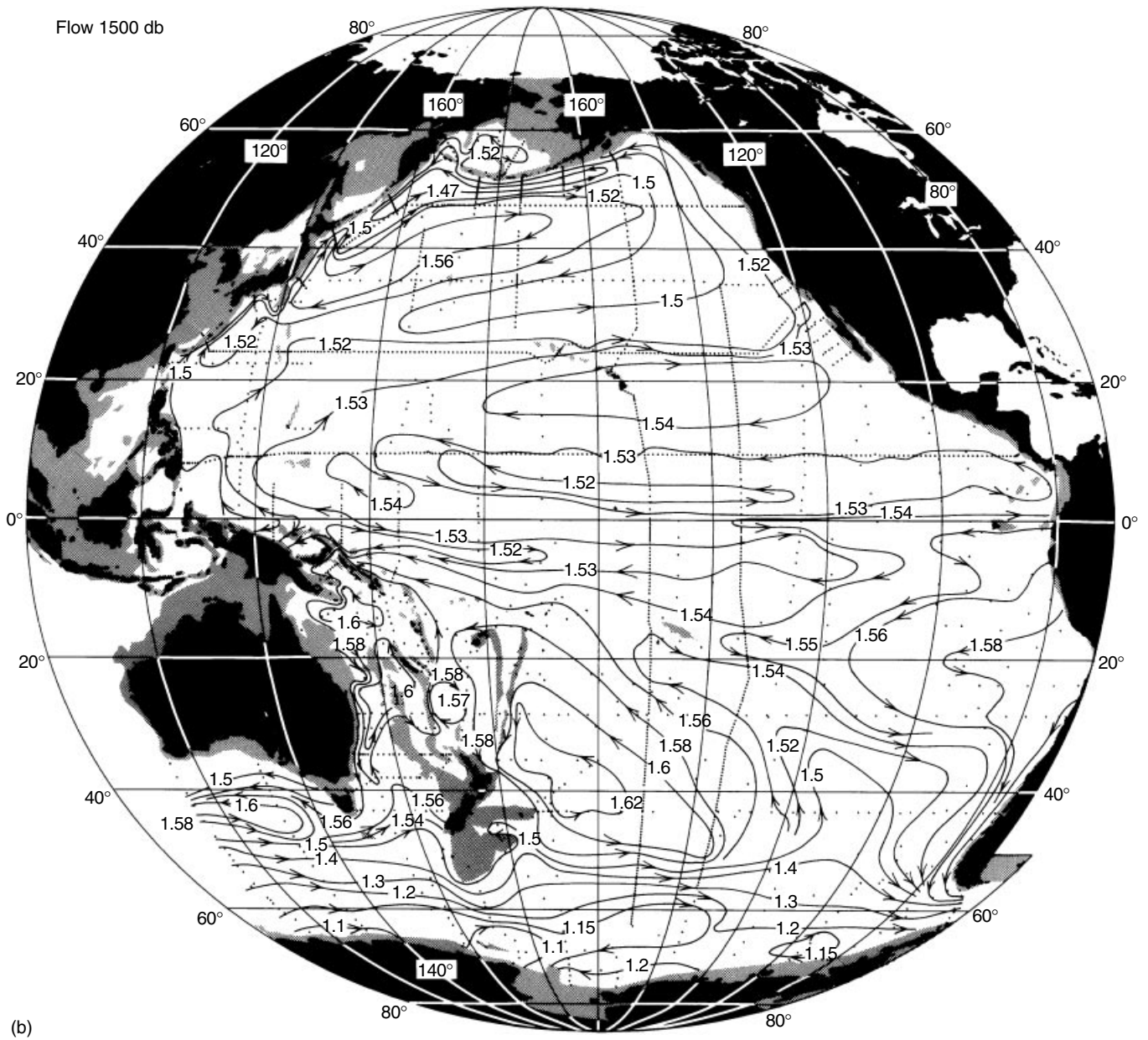
Figure 4   (*Continued*)

Northern Hemisphere. The difference lies in the average position of the Intertropical Convergence Zone north of the equator, which creates the North Equatorial Countercurrent (*see* **Intertropical Convergence Zone (ITCZ)**, Volume 1).

North of the North Equatorial Countercurrent lies the North Equatorial Current, which is also the westward flow of the subtropical gyre and which was mentioned in the section on surface circulation. The North Equatorial Current flows westward to the western boundary and splits into the northward western boundary current for the subtropical gyres (Gulf Stream and Kuroshio) and into a southward

flow for the long, narrow clockwise tropical gyre (Antilles Current in the North Atlantic and Mindanao Current in the North Pacific).

The westward equatorial surface flow is driven by the trade winds (easterlies). When the trade winds weaken or even reverse, the flow of water westward at the equator weakens or reverses and upwelling weakens or stops. In the Pacific, this is the case during an El Niño (*see* **El Niño and La Niña: Causes and Global Consequences**, Volume 1). In the Indian Ocean, this change in the trade winds happens twice a year during the transitions between monsoons; when
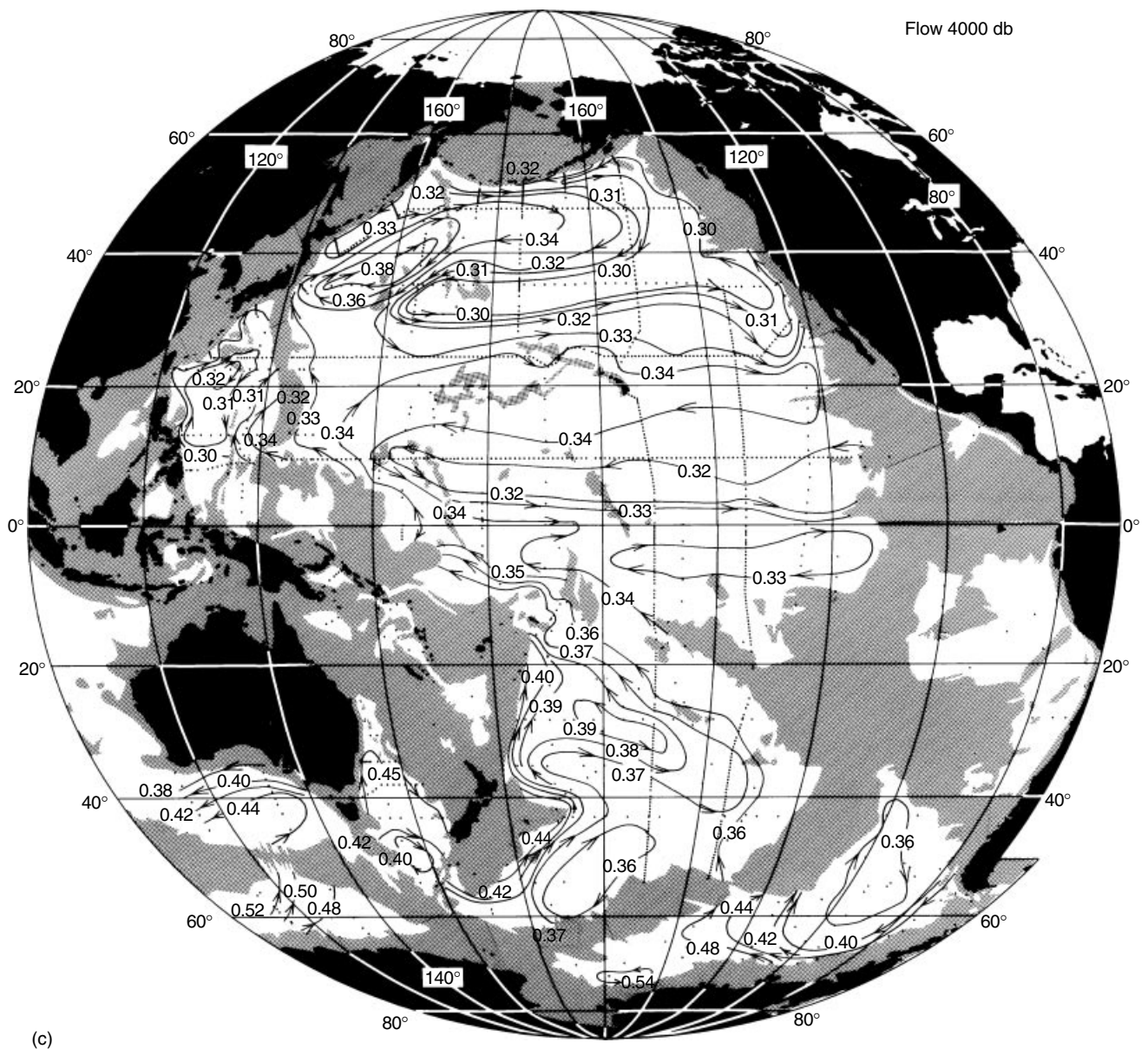
Flow 4000 db

(c)

Figure 4   (*Continued*)

the winds are from the west along the equator, the equatorial surface flow becomes strong and eastward and is referred to as the Wyrtki Jet.

Immediately below the surface currents, the flow directly on the equator is eastward and very strong (greater than $1 \, \text{m s}^{-1}$, that is, even stronger than the Gulf Stream). The strong flow is centered around 150 m depth, and is shallower to the east. It is no more than about 100 m thick and is confined to within 1 °latitude from the equator. This continuous flat ribbon of flow (100 m thick by 200 km wide) is called the Equatorial Undercurrent.

Below the Equatorial Undercurrent, the equatorial currents have complicated structure in the vertical.

Immediately below the undercurrent, the flow is westward and is called the Equatorial Intermediate Current. This current extends down to about 900 m depth. Between 900 and 1800 m, the flow reverses direction every 150 m. These are referred to as the stacked jets. Finally at the bottom on the equator is a layer several thousand meters thick that flows eastward (Firing, 1987).

The subsurface flow just north and south of the equator is eastward, down to at least 400 m (Figure 5). These flows are called the North and South Subsurface Countercurrents, or the Tsuchiya Jets. They sometimes appear to be slightly deeper poleward extensions of the Equatorial Undercurrent.
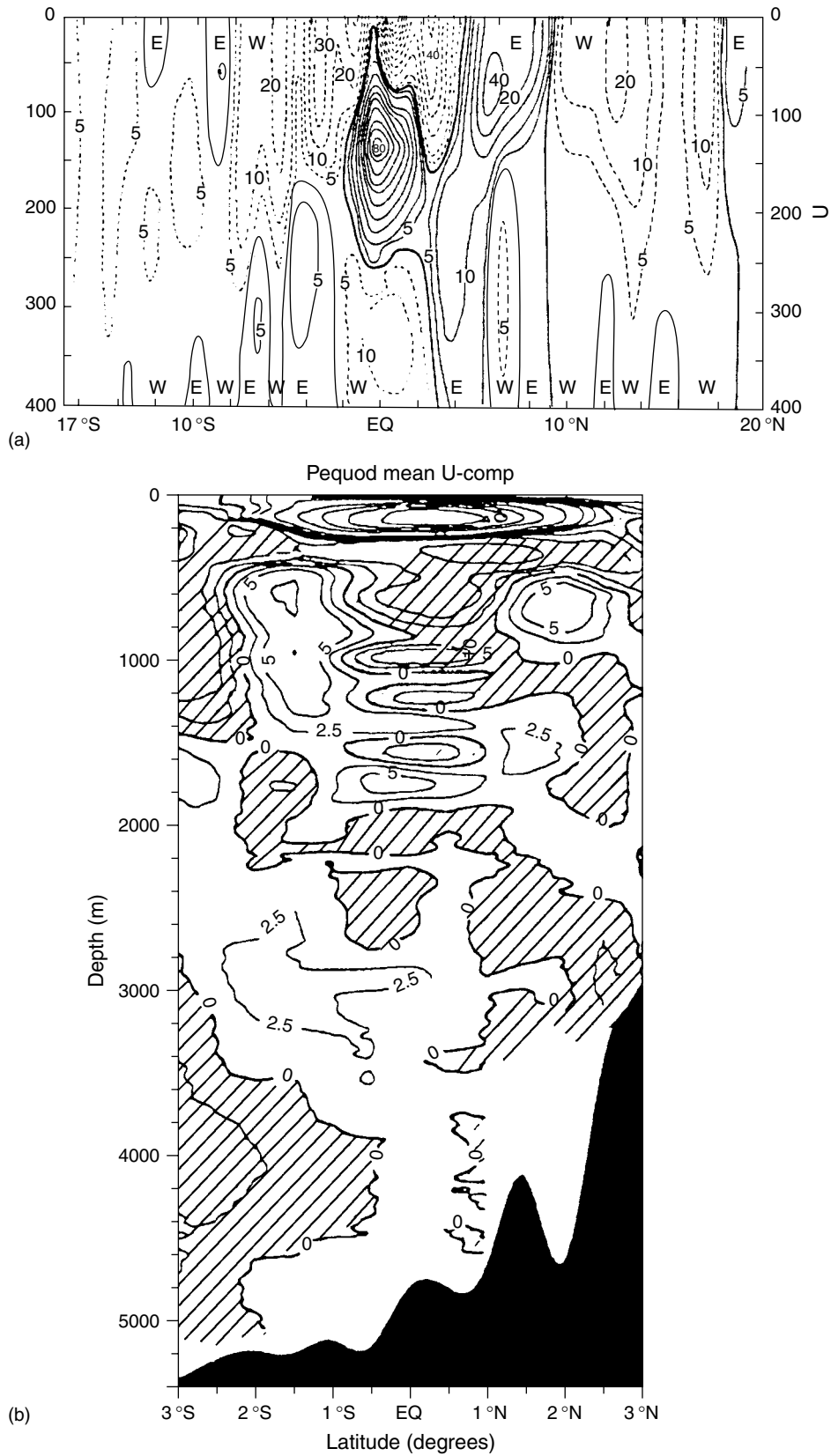
Figure 5   East–west circulation within several degrees of the equator in the central Pacific, based on Wyrtki and Kilonsky (1984) for the upper 1000 m and Firing (1987) for the full water column

THE GLOBAL THERMOHALINE CIRCULATION

The circulations at individual levels shown in the previous three sections provide a view of the complex horizontal flows that transport waters around the basins. It is difficult from these views to discern the throughputs of water that connect the basins and oceans. However when the ocean's properties are viewed along cross-sections from top to bottom and, say, from the northernmost to the southernmost part of each hemisphere, it is clear that there must be connections between the ocean basins. For instance, north–south cross-sections of salinity through the middle of each ocean in the article on salinity patterns (*see* **Salinity Patterns in the Ocean**, Volume 1) show that waters can be detected for a long distance from their high or low salinity sources. These properties are carried by the circulation. The circulation that carries water properties from one wind driven regime to another is the thermohaline circulation.

The thermohaline circulation is driven by cooling, evaporation and salinization through sea ice formation, which increases density, and by heating and precipitation, which decrease density. The thermohaline circulation is much weaker than the wind driven circulation and so it is difficult to see its effect on currents in the upper ocean where the wind driven circulation is vigorous. In the deep ocean, the permanent circulation, particularly when quantified by transports of currents carrying water from one region to another, is dominated by thermohaline forcing.

Cooling is probably the most significant of the various processes in driving a discernible, easily quantified circulation. This is because deep convection created by cooling is localized to well-defined regions of the oceans, and produces waters that are injected below the depth of the vigorous part of the wind driven circulation. References to global thermohaline circulation usually mean this mechanism of driving circulation. While cooling at mid-latitudes in the separation regions of the wind driven western boundary currents is enormous, the associated convection is not deep. The cooled waters are not dense enough to penetrate below the thermocline. It is nearly impossible to quantify how much circulation is associated with mid-latitude cooling because the circulation above the thermocline is strong and associated mainly with wind driving. Mid-latitude cooling is, however, an important part of the global heat budget (poleward heat transport), resulting in a shallow overturn, with somewhat cooler waters returning towards the equator, to be heated and returned poleward in the western boundary current.

Evaporation is large in the subtropical gyres, under the atmospheric high pressure zones (*see* **Salinity Patterns in the Ocean**, Volume 1). As with mid-latitude cooling, the effect of the evaporation can be seen almost exclusively in changes of properties and in calculations of net freshwater transport into the evaporation regions, but not in the effect of evaporation on the circulation itself. On the other hand, brine rejection associated with sea ice formation also increases salinity and is a significant, measurable thermohaline process in producing denser water because it is localized, like high latitude deep convection driven by cooling (*see* **Sea Ice**, Volume 1).

Globally the heating of the Earth balances cooling. Most heating is in the tropics. This means that deep waters cooled at high latitudes must eventually be heated and returned to the surface. This process is probably distributed over most of the oceans, with deep waters gradually warming by downward mixing of heat until the water parcels rejoin the surface circulation and move to the cooling regions. The upwelling rates are very low and regions of larger and smaller upwelling have yet to be well determined.

Precipitation's role in the thermohaline circulation is mostly to inhibit convection. A large-scale example is the subpolar North Pacific, where the surface layer is relatively fresh due to local precipitation, large amounts of runoff from the coast of North America and sea ice melt in the Bering and Okhotsk Seas. The result is a surface layer that cannot convect through the resulting halocline even when the surface temperature is reduced to the freezing point. There are also well-documented events of large (hundreds of square kilometers) patches of low salinity at the sea surface in the subpolar North Atlantic that can inhibit the usual convection as they move slowly counterclockwise with the wind driven circulation (Dickson *et al.*, 1988).

Because the thermohaline circulation is weaker than the wind driven circulation and because it inherently involves overturning, a common way to quantify it is to compute how much water in different layers is transported into and out of given areas. A common choice is to calculate how much water flows northward/southward across an east–west section that goes completely across an ocean basin. For instance, transports are often calculated at 24 °N in the Atlantic and Pacific, where there are data sets that go from coast to coast. To be most meaningful, oceanographers divide the water column up into layers associated with surface waters, thermocline water, and various intermediate and deep layers. The resulting transports can show water moving northward across that latitude in say the uppermost, warm layer, and returning southward in a deep, but not bottom layer.

Transport calculations were introduced above for the wind driven circulation. It was seen above that the wind driven western boundary currents carry about 50–100 Sverdrups. The global thermohaline circulations described here carry about 15–20 Sverdrups, which is considerably less than the wind driven circulation. However, the thermohaline circulation is important because it connects all regions of the global ocean.

The recognizable elements of the thermohaline circulation are the deep western boundary currents, which
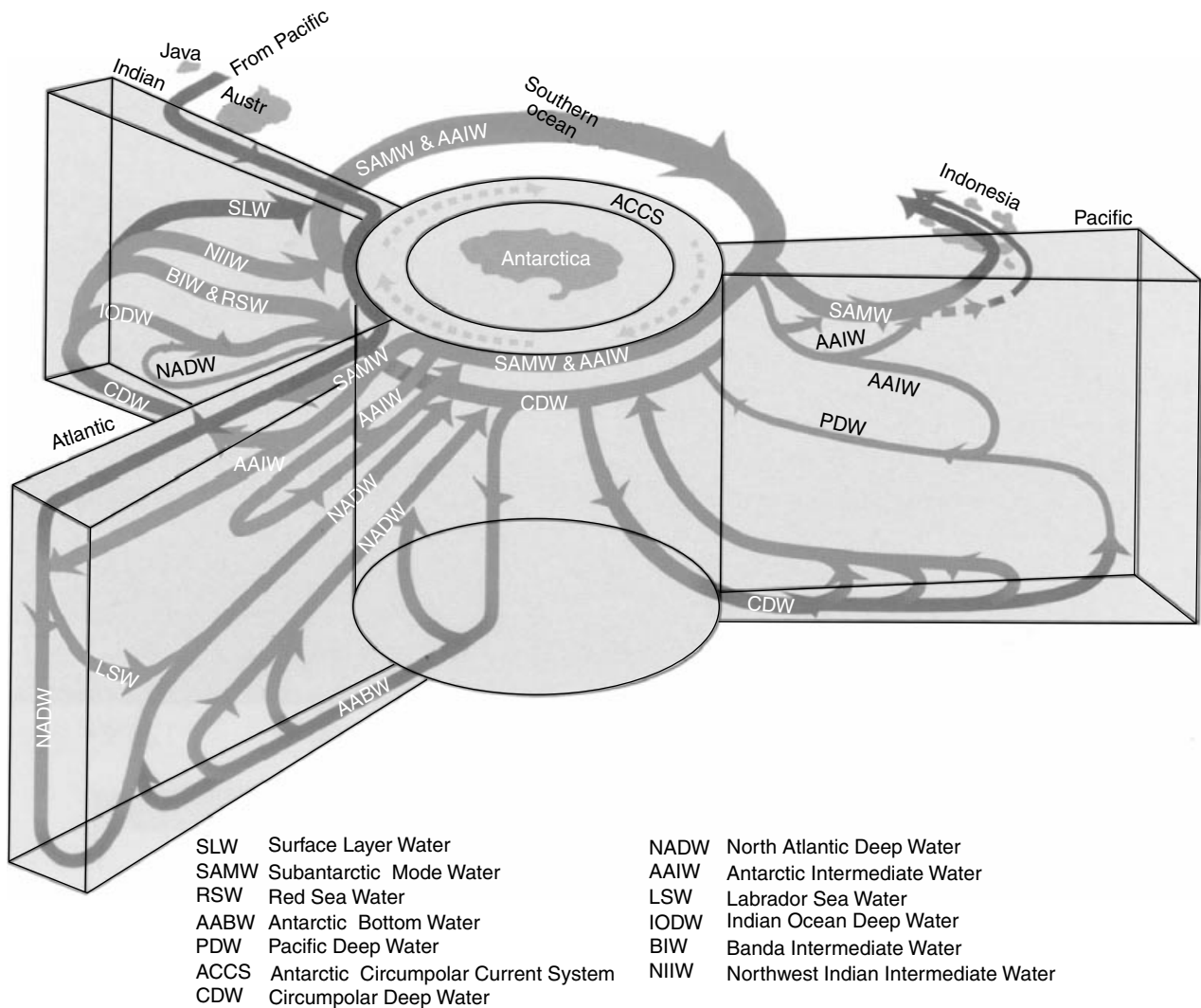
SLW	Surface Layer Water	NADW	North Atlantic Deep Water
SAMW	Subantarctic Mode Water	AAIW	Antarctic Intermediate Water
RSW	Red Sea Water	LSW	Labrador Sea Water
AABW	Antarctic Bottom Water	IODW	Indian Ocean Deep Water
PDW	Pacific Deep Water	BIW	Banda Intermediate Water
ACCS	Antarctic Circumpolar Current System	NIIW	Northwest Indian Intermediate Water
CDW	Circumpolar Deep Water		

Figure 6 Schematic circumpolar view of the global overturning circulation, modified from Schmitz (1996). The water mass names have been changed to match the nomenclature used here, some additional branches in the NADW formation have been added to match the text, and some details of Southern Hemisphere intermediate waters have been omitted

were mentioned in section 4 in reference to Figures 3(c) and 4(c). The pattern of flow that fills in the interior of the basins from the deep western boundary currents is strongly affected by topography.

There are two global thermohaline circulations involving deep convection at high latitudes: the North Atlantic Deep Water (NADW) and the Antarctic Bottom Water (AABW) cells (Figure 6). These two global overturning circulations are interconnected. Smaller thermohaline cells are more connected with the wind driven circulation and quite complex (see Schmitz, 1995) and will not be described here. NADW and AABW are water masses, that is, particular types of waters that can be recognized by their special properties, such as a salinity maximum (for NADW) or extreme cold temperature (AABW). Because thermohaline circulation involves transformation

of surface waters into something else, it is often and perhaps most easily described in terms of the plethora of recognized water masses. Besides the NADW and AABW, we will introduce several other water masses that are associated with the overturning circulations of these two.

The NADW global overturning has sometimes been referred to as the global conveyor belt (Broecker, 1991), referring to its extent from the northern North Atlantic to the Pacific Ocean via the Antarctic, its upwelling, and its return of warm water through the Indonesian archipelago, across the Indian Ocean and back up into the Atlantic (*see* **Ocean Conveyor Belt**, Volume 1). The actual NADW cell has considerably more detail and many additional elements (Worthington, 1976; Schmitz and McCartney, 1993; Reid and Lynn, 1971; review in Talley, 1996). The NADW

cell is depicted reasonably well by Schmitz (1995, 1996) (Figure 6).

The deep water in the NADW cell is formed in the northern North Atlantic. There are three northern North Atlantic sources. First is localized deep convection in the Nordic Seas, which lie north of the sills connecting Greenland, Iceland and Scotland. The renewed Nordic Seas water is dense and plunges over the sills into the subpolar North Atlantic; the most important pathway is through Denmark Strait between Greenland and Iceland. This dense Nordic Sea overflow water (NSOW) reaches the bottom of the northern North Atlantic and flows westward and southward, following the continental slope. The second source is localized intermediate depth convection in the Labrador Sea, between Labrador and Greenland. This area is part of the subpolar North Atlantic. The convected LSW fills the mid-depth of the Labrador Sea and is moved eastward out of the Labrador Sea to fill the subpolar North Atlantic and southward along the continental slope, just above the Nordic Seas water. Together the LSW and Nordic Seas overflow water form the deep western boundary current in the North Atlantic. At mid-latitudes these waters are joined by those from the third source, which is the Mediterranean Sea. Within the Mediterranean, deep convection driven by heat loss and evaporation creates a dense, very saline water that flows back into the North Atlantic, plunging down over the sill at the Strait of Gibraltar and filling intermediate depths at mid-latitudes. This MOW is recognized by its high salinity (*see* **Salinity Patterns in the Ocean**, Volume 1). The MOW joins the deep western boundary current in the tropics and together the renewed North Atlantic waters flow southward into the South Atlantic as the NADW. By the time they reach the southern South Atlantic, the individual sources of the NADW have mixed together and cannot be recognized; the NADW becomes a generally saline and well-oxygenated water mass.

NADW in the South Atlantic joins the Antarctic Circumpolar Current and flows eastward towards the Indian Ocean. As the NADW is carried eastward in the Antarctic Circumpolar Current, its high salinity gradually becomes fresher through mixing with other waters, but it remains recognizable as a salinity maximum all the way through the Indian and Pacific Oceans. Some of the high salinity water flows northward into the Indian and Pacific Oceans, and is one of the two major components of the deep water for these oceans. In the Indian Ocean, the northward flow of deep water occurs in three regions, separated by mid-ocean ridges. In the southwestern Indian Ocean, a branch of NADW flows northward into the region east of the Agulhas. In the Pacific Ocean, the main northward flow is just east of New Zealand, in the deep western boundary current described in the section on tropical circulation. This deep flow fills the South Pacific and moves northward through

the Samoan Passage into the tropics and then into the North Atlantic.

Where does NADW upwell to eventually complete the circuit back to the northern North Atlantic? This is a very difficult question to answer. Within the North Atlantic, by computing net transports across sections from coast to coast, one can show that the northward warm flow is in the Gulf Stream and is mostly in and above the thermocline. The calculations become more difficult the more removed they are from the North Atlantic. What is clear is that there is no wholesale upwelling in the Pacific Ocean that feeds the warm westward flow through the Indonesian archipelago, as depicted in Broecker (1991) (*see* figure in **Thermohaline Circulation**, Volume 1). In the Antarctic Circumpolar Current, NADW is joined by bottom waters formed in the Antarctic. These flow together into the deep Indian and Pacific Oceans and both upwell into older deep water layers (between 1500 and 3000 m depth) in those oceans. The IDW and PDW re-enter the Antarctic Circumpolar Current and dilute the NADW there. South of the Antarctic Circumpolar Current, these mingled deep waters upwell to near the surface, only to have some portion convected or salinified through brine rejection to become deep and bottom waters in the Antarctic (the second major global overturning cell).

Some of the upwelled deep waters might also take part in formation of the thick mixed layer waters just north of the Antarctic Circumpolar Current. These thick mixed layers vary in properties around Antarctica, becoming colder and fresher eastward from South America until reaching their most extreme values just west of Chile (McCartney, 1977). The thick mixed layers in general are called Subantarctic Mode Water (SAMW). The coldest, freshest SAMW is modified by intermediate depth convection just west of Chile and becomes a fresh intermediate water called Antarctic Intermediate Water (AAIW). AAIW fills all of the Southern Hemisphere oceans north of the Antarctic Circumpolar Current, by spreading northward by the broad subtropical gyre circulations. SAMW, which is warmer than AAIW, also spreads northward from the Antarctic Circumpolar Current, via the subtropical gyre circulations.

SAMW and AAIW have three sources: water that flows southward in the Southern Hemisphere's subtropical gyre western boundary currents, water from south of the Antarctic Circumpolar Current that is pushed northward across the current, and upwelling from the underlying deep waters, most likely as the SAMW and AAIW circulate around the subtropical gyres.

Schmitz (1995) shows an upwelling of deep water to intermediate water within the subtropical gyres and then a conversion of intermediate waters to warm upper waters, primarily in the eastern South Pacific and eastern South Atlantic. Many of the conjectures about upwelling are, however, based on mass balances that have significant

errors. What Schmitz's (1995) schematics and budgets do clearly show is that the global thermohaline system is fairly complicated.

The second major global overturning circulation is that of AABW (or Lower Circumpolar Deep Water, CDW), with formation of this water in localized regions in the Antarctic (Orsi *et al.*, 1999). Two primary sources of AABW have been found in the Weddell Sea and in the Ross Sea. Both of these areas have clockwise subpolar-type gyres, as described in the first section. Formation is through brine rejection, which occurs during formation of sea ice (*see* **Salinity Patterns in the Ocean**, Volume 1; **Sea Ice**, Volume 1). A third distributed source has been found along the Antarctic coast south of Australia (Rintoul, 1998), also presumed to be brine rejection in coastal areas. AABW spreads northward to the Antarctic Circumpolar Current through the gyre circulations south of the current. From the Antarctic Circumpolar Current, it enters the deep Atlantic Ocean as a northward flowing deep western boundary current, beneath the southward flowing deep western boundary current that carries NADW. AABW is the bottom water of the South Atlantic and also of the North Atlantic, extending as far north as Bermuda. By the time it reaches Bermuda, it all upwells into the NADW layer above it, and the bottom waters north of this region are NADW.

In the Indian and Pacific Oceans, AABW flows northward along with the high salinity NADW core that is also within the Antarctic Circumpolar Current. Oceanographers usually do not refer to these cores as AABW or NADW, because they are significantly diluted by the old deep waters from the Indian and Pacific Oceans, which themselves are formed from upwelled AABW and NADW. Throughout its path, AABW upwells into the deep water layer above it, and so it does not fill the northern reaches of either the Atlantic or Pacific Oceans.

What is the warm water return path to the AABW source? As mentioned above, the water that is found at the surface in the Antarctic regions where AABW is formed, is upwelled deep water from the Antarctic Circumpolar Current.

The magnitude of the currents associated with the NADW and AABW overturning circulations is about $5\,\mathrm{cm\,s^{-1}}$. The total transport of NADW out of the North Atlantic is 15–20 Sverdrups. The transport of AABW into the Antarctic Circumpolar Current is not as well measured, but is between 10–30 Sverdrups.

PHYSICAL PROCESSES THAT CREATE THE GENERAL CIRCULATION

Up to this point, the circulation of ocean waters has been described without much mention of the physical processes that oceanographers consider to be central. These are described here. The preceding sections could have been couched in these physical process terms, much as is the description of global circulation and water masses in a recent text by Tomczak and Godfrey (1994) and in a review of what is known about Pacific circulation with emphasis on comparison with theory (Talley, 1995). The central concepts for ocean circulation are: geostrophy, Ekman transport, Sverdrup transport, western boundary current theory, subduction, abyssal circulation theory, and equatorial circulation theory. These are not a complete set of the important concepts, but serve as a framework for much of what was described in the previous sections. Reasonable descriptions of these processes without extensive mathematical formulations can be found in texts such as Tomczak and Godfrey (1994).

A basic tenet for general circulation, which has very large spatial scales except in boundary currents, and long temporal scales, is that friction (viscosity) is very weak. Molecular-scale diffusion is indeed extraordinarily weak compared with the diffusion and friction that would be needed to have any impact on the general circulation. The ocean does have some frictional responses, including the Ekman transport and western boundary current formations described below. Because we observe these behaviors, which occur only in boundary layers at the sides of the ocean where flow changes rapidly with distance or depth, and because we also observe that most of the ocean acts as if there were no friction, we conclude that something like friction must be acting near boundaries. Based on where the pseudo-frictional behavior occurs, we empirically conclude that small-scale turbulence, and possibly eddies up to a scale of a few kilometers, create something like friction and diffusion for the general circulation.

The most central concept for the general circulation is *geostrophy*, which is a balance between the force resulting from differences in pressure and the Coriolis force resulting from the Earth's rotation (*see* **Coriolis Effect**, Volume 1). Geostrophic flows in the Northern Hemisphere are always exactly to the right of the pressure difference force, while in the Southern Hemisphere they are always exactly to the left. All general circulation flows, except in the very surface layer (top 50 m) and right on the equator, are geostrophic. Tests of geostrophy using direct current measurements have borne this out. The pressure differences, especially for the upper ocean circulation, arise from differences in sea surface height, which allows more water mass to appear in one place than another. The difference in sea surface height across the strongest currents such as the Gulf Stream are no more than about 1 m and height difference for weaker currents are much smaller. This cannot be measured directly with today's technology. Therefore, we have no direct knowledge of the absolute pressure difference between two points. What we can construct though from the ocean's density distribution is how the pressure differences change with depth, which means that the velocity changes with depth. For instance, for a wind driven circulation such

as the Gulf Stream, the velocity is highest at the sea surface, which means that the largest pressure difference is at the surface. Velocity decreases with depth, which has implications for the density distribution with depth, resulting in density surfaces that tilt across the current (see the textbook descriptions). We then must somehow measure or guess the true pressure difference at one reference depth. This is done in many ways, all of which require intensive work.

The second central concept is of *Ekman transport*, which is the flow driven directly by the winds in the surface layer. The winds push on the water through friction, and the frictional flow in the very top layer (say 1–2 m) is conveyed through friction to the next layer and so forth. This downward conveyance of the frictional driving dies out with depth, and vanishes around 50 m. Because of the Earth's rotation, each layer is driven slightly to the right of the one above in the Northern Hemisphere (left in the Southern Hemisphere). If the total flow in this 50 m thick layer, which was first described by Ekman (1905), is added up, the net flow is exactly to the right of the wind (Northern Hemisphere). This net flow is the Ekman transport and is how the wind directly forces the ocean.

Ekman transport has immediate consequences for the ocean when winds blow along a coast. If, as along California, they blow from the north in the Northern Hemisphere, then the Ekman transport is offshore (i.e., to the west), which results in upwelling at the coast to fill in waters pushed westward by the Ekman transport. The upwelled water comes from about 100–200 m depth and brings with it significant nutrients and colder water, as described in the first section. The offshore transport also causes the sea surface to be slightly lower at the coast than offshore, which results in a geostrophic flow along the west coast of North America, creating the California Current, which is the eastern boundary current described in the first section.

In the middle of the oceans, in places where the winds vary with position, the Ekman transport also varies with position. This results in upwelling if the Ekman transport has to be supplied (Ekman divergence), or downwelling if it has to be removed (Ekman convergence). In subtropical gyres, which are driven by westerly (west to east) winds on the higher latitude side and trade winds (east to west) on the lower latitude side, Ekman transport is towards the equator on the higher latitude side and towards the pole on the lower latitude side. This results in downwelling in the gyre. Subpolar gyres on the other hand are characterized by upwelling, which explains their much higher surface nutrient content and hence much higher productivity than subtropical gyres. The biology in surface waters uses up whatever food is available in a short time, and the dead parts and fecal pellets that fall below the surface layer provide nutrients back into the water through bacterial decay.

The surface circulation described in the first section was the purely geostrophic circulation, based on pressure differences and Coriolis acceleration. Ekman transport is not geostrophic, and is superimposed on top of the geostrophic flow. Thus, the surface circulation, as described here, is really more like the true circulation about 50 m below the sea surface, assuming only small changes in the pressure difference force between the surface and 50 m.

The third concept for general circulation is *Sverdrup transport*. This is the reaction of the full water column to Ekman convergence or divergence. When there is Ekman convergence, and hence downwelling, it is as if the whole water column is being squashed. Because the water column is effectively rotating because the Earth is rotating, squashing will cause its rotation rate to flow (much like a spinning ice skater who extends her arms and slows). The rotation rate change is accomplished by the movement of the water column to a different latitude, with slower spinning being towards the equator and faster spinning towards the poles. This response to the winds/Ekman transport was theorized by Sverdrup (1947). Sverdrup transport is geostrophic, i.e., the forcing that causes the northward or southward movement is very small, so that what we actually observe is that the pressure field is set up to produce geostrophic flow in the correct direction.

Because the average wind patterns for the globe are east–west, the mid-ocean Sverdrup transport tends to be all northward or almost all southward at a given latitude. Therefore Sverdrup transport can work only if there are north–south boundaries along which the flow can return in a narrow boundary current that has different physics from the Sverdrup transport. Support for the Sverdrup transport idea lies in the existence of the subtropical and subpolar gyres, in which flow is equatorward over most of the ocean in Ekman convergent regions, and poleward in Ekman divergent regions. This requirement for boundaries explains why the ocean does not exhibit a clear Sverdrup transport response to the winds in the Antarctic where there is no land (at the latitude of Drake Passage), and why the atmosphere which obeys the same physical laws as the ocean does not exhibit Sverdrup transport.

Narrow western boundary currents are required to return the Sverdrup transport back to its original latitude, to close the mass balance (because there are no holes in the ocean). These boundary currents are always on the western side of the ocean. This is because they are removing energy (or rotation changes) put in by the large-scale winds through the Sverdrup transport. The physics of the boundary currents that dictates that the currents must be on the western side is friction (or viscosity), which can act if there are large changes in flow in a short space scale. In a subtropical gyre of the Northern Hemisphere for instance, the winds put in negative rotation (reducing the rotation), with the response that the water moves equatorward. The water must move

back poleward to complete the gyre. This is only possible if there is a source of positive rotation. Friction acting on boundary current at the western wall can input positive rotation because the current will have no velocity at the wall, and maximum velocity at some distance offshore (about 50–100 km). This difference in velocity (zero at the wall and maximum offshore) will thus create positive rotation (think of a paddle wheel sitting in the flow next to the wall) and can balance the Sverdrup transport forcing. The theory of frictional western boundary currents was introduced by Stommel (1948) and Munk (1950) (*see* **Munk, Walter**, Volume 1). A frictional boundary current on the eastern wall however would have negative rotation with zero velocity at the wall and maximum poleward velocity offshore and so cannot balance the Sverdrup transport forcing. Like Sverdrup transport and unlike Ekman transport, the flow in western boundary currents is almost geostrophic i.e., the frictional part, while important in determining the direction and shape of the flow, is still much weaker than geostrophy, and so it is possible to use estimates of the horizontal changes in pressure to compute the speed of the currents.

Eastern boundary currents are of a completely different nature from western boundary currents, and were explained above when Ekman transport was introduced. They are the geostrophic flow that goes with the coastal upwelling created by offshore Ekman transport (equatorward if there is upwelling and poleward if there is downwelling). Upwelling is very shallow because the ocean is strongly stratified and it is not possible to bring very dense water all the way up to the surface, and so eastern boundary currents are also very shallow.

Returning to the interior circulation of the ocean gyres, we consider again the topic of subduction. As already explained, subduction is a phenomenon in the subtropical gyres in which the cooler surface waters from the higher latitudes in the gyre must be moving equatorward because of the Sverdrup transport. In the steady state ocean circulation, the surface waters are warmer and lighter towards the equator. Because there is no whole scale heating at the sea surface that would change the cooler water into exactly the right warmer waters, the cooler waters subduct beneath the warmer ones. Subduction is the main mechanism for setting the vertical temperature (and salinity) structure in the subtropical gyres, down to the depth of the coldest/densest water that outcrops at the highest latitude of the subtropical gyres. This latitude is around 40–55° depending on the wind patterns in each of the ocean basins. The subducted waters typically reach no deeper than about 500–1000 m in the centers of the subtropical gyres.

Part of the deep ocean circulation (i.e., the thermohaline part) is driven by sinking of cold, dense water at high latitudes and upwelling of these waters elsewhere. Abyssal circulation theory, introduced by Stommel and Arons (1960), explains the current structure that results from isolated sinks of water, the spread of these waters to fill the oceans, and a general upwelling to return the cold water back to the upper ocean. The latter process must be accompanied by downward diffusion of heat, of course, but is not included explicitly in the Stommel/Arons theory. The crux of their theory is that most of the ocean is responding to the general upwelling, because the sinking is so localized (for the ocean now this would be in the Nordic Seas and the Antarctic). If the upwelling is uniform, i.e., the same size everywhere in the ocean, and if the ocean had a flat bottom, then the upwelling acts on the deep ocean like Ekman divergence, acting to create Sverdrup transport in the wind driven ocean circulation. That is, the response is for the water columns to move poleward (slowly) because their rotation rate is being increased by the upwelling.

There must of course be a connection between the sources of dense water at high latitudes and all of this upwelling and poleward flow. Stommel and Arons showed that the connection is through deep western boundary currents, called deep because this mechanism has to do with circulation in the deep layer fed by the high-latitude sinking. Because the actual deep ocean is far from flat bottomed, and because the upwelling itself very likely is not uniform (although we do not yet know what its distribution is), the slow poleward flow of the Stommel/Arons theory is not really observed. However, the deep western boundary currents that connect all of the sinking and upwelling together are clearly observed and are the primary evidence that something like their mechanism is at work. Again, as for Sverdrup transport and wind driven western and eastern boundary currents, the abyssal circulation driven through this sinking and upwelling is basically geostrophic.

The equatorial circulation is a special topic because geostrophy breaks down at the equator. The breakdown is apparent because the Coriolis force has one sign in the Northern Hemisphere (flow is to the right of the pressure difference force) and the opposite sign in the Southern Hemisphere (flow to the left of the pressure difference). Right on the equator, the Coriolis force vanishes. Flow directly on the equator responds directly to the pressure difference force and flows from high pressure to low pressure (instead of being geostrophic). The surface frictional flow on the equator is pushed in the direction of the winds (rather than at right angles as in Ekman transport). In the Pacific and Atlantic, the equatorial winds are almost always easterlies (trades), and blow the surface water to the west. In the Indian Ocean, the winds reverse seasonally and sometimes blow the surface water to the west and sometimes to the east. Because all of these oceans have side boundaries at the equator, the wind blown water piles up at the boundary. This creates a sea surface height difference between the western and eastern parts of the oceans. In

the Pacific and Atlantic, the water is piled up at the western boundary. This creates a pressure difference force that drives water back towards the eastern boundary. The flow created by this pressure difference force is the equatorial undercurrent.

A small distance away from the equator (within 30 km), the Earth's rotation begins to assert itself on the equatorial circulation. Ekman transport (at right angles to the wind forcing) begins to assert itself and flow also begins to respond geostrophically to pressure differences. The Ekman response to the prevailing trade winds in the Pacific and Atlantic is transport away from the equator both north and south of the equator. This creates upwelling in the equatorial region, which is important for setting sea surface temperature there and hence has a direct impact on climate (see section below).

There are many additional aspects to ocean circulation theory at the equator, including reasons for the existence of such very complicated vertical structure, but there is not room here to go into them because explanations require introducing the very large-scale waves that adjust the ocean circulation to changes in forcing.

THE ROLE OF OCEAN CIRCULATION IN CLIMATE

The ocean affects the atmosphere and hence the Earth's climate by directly heating and cooling the atmosphere. The ocean also affects climate through its effect on the Earth's albedo, which is how strongly the Earth reflects solar radiation back to space, and by its role in greenhouse gas cycles. Climate change also has an expression in the ocean through changes in temperature and salinity distributions, largely due to circulation changes, which in turn affect biological distributions in the ocean.

The circulation, temperature and salinity changes that are associated with present day climate fluctuations, particularly those centered at mid to high latitudes such as the North Atlantic Oscillation and the Arctic Oscillation, are linked to only small perturbations to the general circulation (*see* **Arctic Oscillation**, Volume 1; **North Atlantic Oscillation**, Volume 1). The tropical climate fluctuations such as El Niño are associated with relatively larger changes, but even these are only large at the sea surface, with only small changes to the deeper tropical circulation. During the much longer glacial cycles, the circulation can be significantly perturbed, but the basic physical elements described in the section above remain intact even if altered.

Ocean circulation thus affects climate through its influence on sea surface temperature, ice formation, and the upwelling/downwelling that affect greenhouse gas sequestration. The ocean's surface temperature distribution is a major forcing factor for the atmosphere, particularly in the tropics where atmospheric convection to great heights

occurs. The ocean has a central role in the tropically centered climate cycle of El Niño (*see* **El Niño and La Niña: Causes and Global Consequences**, Volume 1). Sea surface temperature in the equatorial oceans is affected by the depth of the thermocline there, by the extent to which there is upwelling of colder waters from below the thermocline, and by east–west circulation that moves warm and cold waters around. The normal trade wind forced Pacific Ocean circulation, with surface waters pushed to the west and upwelling driven mainly by Ekman transport away from the equator, results in warm temperatures in the west and cold in the east. When the trade wind strength is reduced, and winds in the western Pacific even reverse, upwelling is reduced and warm water surges to the central Pacific. This reduces the temperature difference between the western and eastern equatorial regions, which in turn changes the strength and shifts the location of the atmosphere's tropical convective cell (the Walker circulation). This shift in the atmosphere further reduces the trade winds hence the cycle has positive feedback. Similar feedbacks between the ocean and atmosphere are being explored for the tropical Atlantic and Indian Oceans.

Within the lowest layer of the atmosphere the ocean has a major impact on local climate, as clearly seen in coastal regions. Water has a much larger heat capacity than air, which means that water temperature changes in response to gain or loss of a given amount of heat are much smaller than air temperature changes. Ocean temperatures vary within a much smaller range than air temperatures. Coastal climates have much smaller extremes than inland climates because of this relative oceanic stability.

At mid and high latitudes, the direct feedback of the ocean's surface temperature on climate is less marked because deep atmospheric convection is a tropical phenomenon. The general surface temperature difference from equator to pole, much of which is due to the ocean, drives the atmosphere's Hadley circulation with rising air in the tropics and sinking air at mid-latitudes. The large heat loss from the ocean (heat gain by the atmosphere) within the subtropical western boundary current separation regions affects the major storm tracks. Several climate modeling studies have suggested the importance of mid-latitude ocean temperature distributions in the mid-latitude decadal climate oscillations (e.g., Latif and Barnett, 1994). The lure of using ocean circulation to explain these variations is in the match of the time scale of the oscillations to the time scale of the upper ocean circulation. This takes a decade or two depending on ocean width, with the Pacific width and time scales about twice those of the Atlantic. Mid-latitude temperature distributions are affected by the strength of the western boundary currents, by sequestration of temperature anomalies through subduction in the subtropical gyres with reemergence of the anomalies either in the tropics or at the western boundary, and by westward, slow propagation of

large-scale planetary waves that carry information about the ocean's thermocline depth and hence upper ocean temperature distribution.

The ocean's various thermohaline circulations help to redistribute the Earth's heat, with the oceans carrying about one-third to one-half of the Earth's heat from the tropics to the poles and the atmosphere carrying the remainder. Where circulation is dominated by wind driving, such as in the North Pacific, almost all of the heat is carried in the shallow wind driven circulation, and is associated with subduction. Where circulation also has a major thermohaline component, and if the thermohaline component is associated with temperature changes of more than about 5 °C, both the wind driven and thermohaline circulations are important for heat transport. Such is the case in the North Atlantic, where there is both major heat loss in the Gulf Stream separation region, associated with the wind driven subtropical gyre, and heat loss in the eastern subpolar gyre and in the Nordic Seas, associated with the deep thermohaline overturning. Warmth is carried northward by the ocean circulation, and the cooled waters return southward in both the subducted upper layers of the subtropical gyre and in the intermediate and deep waters formed farther north.

Changing sea ice distributions (*see* **Sea Ice**, Volume 1) may have a stronger effect on climate than mid-latitude sea surface temperature changes, through both the much larger temperature contrast between the ice and water than between different waters and by the role of ice in the Earth's albedo (reflectivity). Sea ice distributions are affected to some extent by ocean circulation. A marked example is in the Nordic Seas between Greenland and Norway and north of Iceland, where warm northward flowing North Atlantic waters keep the eastern part of the Nordic Seas ice-free all year round. An increase in the flow would decrease the sea ice area and move the ice edge. During glacial periods, sea ice extended much farther equatorward and the atmospheric circulation would have had significantly different strength as a result.

Within the ocean, climate change, including interannual and decadal variations as well as much longer time scales, has a demonstrable effect on local biological productivity and species distribution. In the North Atlantic for instance, observed changes in temperature in the subpolar gyre over the past 100 years have had major consequences for fisheries, particularly spawning grounds which depend on temperature (Dickson *et al.*, 1988). In the North Pacific, El Niño changes the temperature distribution of the tropical and eastern boundary regions, including flow of warmer waters poleward along the eastern boundaries, which brings tropical species to higher latitudes, and which reduces the efficiency of local upwelling and hence reduces nutrient supply to the local fisheries.

At the western boundary of the North Pacific, the position of the subpolar western boundary current separation (that is, maximum southward intrusion of Oyashio waters) is carefully monitored by Japanese fisheries agencies because of the profound impact of presence or absence of the nutrient rich Oyashio waters. Within the subpolar North Pacific, the position of the subarctic front also has a large impact on fisheries, because of the large-scale upwelling of the subpolar gyre north of the front. The latitude at which the subpolar and subtropical gyre circulations split in the eastern North Pacific impacts the paths followed by returning spawning salmon. (North of Vancouver Island the fish belong to Canadian fisherman and south of Vancouver Island they belong to US fisherman.) Many other examples of how ocean and circulation changes impact biota abound, including in paleo-oceanographic studies, in which changes in marine species are used to reconstruct past ocean circulations.

The role of the ocean circulation in climate was the main subject of the international World Ocean Circulation Experiment (WOCE), whose field study phase was 1991–1998, and which has sponsored major advances in general circulation modeling and ocean data assimilation (*see* **WOCE (World Ocean Circulation Experiment)**, Volume 1). Because it was recognized that the general circulation, as it exists today, is central to climate, much of WOCE dwelt on describing and modeling the present ocean circulation. WOCE results have been reported in several thousand scientific publications, and have been summarized in a recent book (Siedler *et al.*, 2001). The WOCE data sets and model advances will be used for decades to come to continue to advance understanding of the general circulation and to provide necessary checks for climate models, which couple all of the relevant systems; atmosphere, ocean, land, and biology.

See also: **Atmospheric Motions**, Volume 1; **Ocean Observing Techniques**, Volume 1; **Sea Surface Temperature**, Volume 1; **Sverdrup, Harald Ulrik**, Volume 1.

REFERENCES

Broecker, W (1991) The Great Ocean Conveyor, *Oceanography*, **4**, 79–89.

Davis, R E (1998) Preliminary Results from Directly Measuring Mid-depth Circulation in the Tropical and South Pacific, *J. Geophys. Res.*, **103**, 24 619–24 639.

Dickson, R R, Meincke, J, Malmberg, S-A, and Lee, A J (1988) The Great Salinity Anomaly: In the Northern North Atlantic 1968–1982, *Prog. Oceanogr.*, **20**, 103–151.

Ekman, F W (1905) On the Influence of the Earth's Rotation on Ocean Currents, *Arkiv. Matem, Astr. Fys. (Stockholm)*, **2**(11), 53.

Firing, E (1987) Deep Zonal Currents in the Central Equatorial Pacific, *J. Mar. Res.*, **45**, 791–812.

Latif, M and Barnett, T P (1994) Causes of Decadal Climate Variability over the North Pacific and North America, *Science*, **266**, 634–637.

Luyten, J R, Pedlosky, J, and Stommel, H (1983) The Ventilated Thermocline, *J. Phys. Oceanogr.*, **13**, 292–309.

McCartney, M S (1977) Subantarctic Mode Water, *Deep-Sea Res.*, **24**, Suppl., 103–119.

Mizuno, K and White, W B (1983) Annual and Interannual Variability in the Kuroshio Current System, *J. Phys. Oceanogr.*, **13**, 1847–1867.

Munk, W H (1950) On the Wind Driven Ocean Circulation, *J. Meteor.*, **7**, 79–93.

Orsi, A H, Johnson, G C, and Bullister, J L (1999) Circulation, Mixing and Production of Antarctic Bottom Water, *Prog. Oceanogr.*, **43**, 55–109.

Peterson, R G, Stramma, L, and Kortum, G (1996) Early Concepts and Charts of Ocean Circulation, *Prog. Oceanogr.*, **37**, 1–115.

Reid, J L (1994) On the Total Geostrophic Circulation of the North Atlantic Ocean: Flow Patterns, Tracers, and Transports, *Prog. Oceanogr.*, **33**, 1–92.

Reid, J L (1997) On the Total Geostrophic Circulation of the Pacific Ocean: Flow Patterns, Tracers, and Transports, *Prog. Oceanogr.*, **39**, 263–352.

Reid, J L and Lynn, R J (1971) On the Influence of the Norwegian-Greenland and Weddell Seas upon the Bottom Waters of the Indian and Pacific Oceans, *Deep-Sea Res.*, **18**, 1063–1088.

Richardson, P (1980) Benjamin Franklin and Timothy Folger's First Printed Chart of the Gulf Stream, *Science*, **207**, 643–645.

Rintoul, S R (1998) On the Origin and Influence of Adelie Land Bottom Water, Ocean, Ice and Atmospheric Interactions at the Antarctic Continental Margin, *Ant. Res. Ser.*, **75**, 151–171.

Schmitz, W J (1995) On the Interbasin-scale Thermohaline Circulation, *Rev. Geophys.*, **33**, 151–174.

Schmitz, W J (1996) On the World Ocean Circulation: Volume II, The Pacific and Indian Oceans/A Global Update, Woods Hole Oceanographic Institution Technical Report WHOI–96–08.

Schmitz, W J and McCartney, M S (1993) On the North Atlantic Circulation, *Rev. Geophys.*, **31**, 29–49.

Siedler, G, Church, J, and Gould, J (2001) *Ocean Circulation and Climate: Observing and Modelling the Global Ocean*, Academic Press, New York.

Sobel, D (1995) *Longitude: The True Story of a Lone Genius who Solved the Greatest Scientific Problem of his Time*, Walker, New York.

Stommel, H (1948) The Westward Intensification of Wind Driven Ocean Currents, *Trans. Am. Geophys. Union*, **29**, 202–206.

Stommel, H M and Arons, A B (1960) On the Abyssal Circulation of the World Ocean, I, Stationary Planetary Flow Patterns on a Sphere, *Deep-Sea Res.*, **6**, 140–154.

Sverdrup, H U (1947) Wind Driven Currents in a Baroclinic Ocean; With Application to the Equatorial Currents of the Eastern Pacific, *Proc. Natl. Acad. Sci. USA*, **33**, 318–326.

Sverdrup, H U, Johnson, M W, and Fleming, R H (1942) *The Oceans: Their Physics, Chemistry and General Biology*, Prentice Hall, Englewood Cliffs, NJ.

Talley, L D (1995) Some Advances in Understanding of the General Circulation of the Pacific Ocean, with Emphasis on Recent US Contributions, *Rev. Geophys. Suppl.*, 1335–1352.

Talley, L D (1996) North Atlantic Circulation and Variability, Reviewed for the CNLS Conference, *Physica D*, **98**, 625–646.

Talley, L D (1999) Some Aspects of Ocean Heat Transport by the Shallow, Intermediate and Deep Overturning Circulations, in *Mechanisms of Global Climate Change at Millennial Time Scales*, eds P U Clark, R S Webb, and L D Keigwin, *Geophys. Mono. Ser.*, **112**, American Geophysical Union, 1–22.

Tomczak, M and Godfrey, J S (1994) *Regional Oceanography: an Introduction*, Pergamon, Oxford.

Worthington, L V (1976) *On the North Atlantic Circulation*, Johns Hopkins Oceanographic Studies, 6.

Wüst, G (1935) Schichtung und Zirkulation des Atlantisches Ozeans. Die Stratosphäre, *Wissenschaftliche Ergebnisse der Deutschen Atlantischen Expedition auf dem Forschungs-und Vermessungsschiff, Meteor 1925–1927*, **6**, 109–288.

Wyrtki, K and Kilonsky, B (1984) Mean Water and Current Structure during the Hawaii-to-Tahiti Shuttle Experiment, *J. Phys. Oceanogr.*, **14**, 242–253.