Part I

Introduction

1

Description of the world's oceans

The main focus of this book is the study of large-scale circulation in the world's oceans. As a dynamical system, the circulation in the world's oceans is controlled by the combined effects of external forcing, including wind stress, heat flux through the sea surface and seafloor, surface freshwater flux, tidal force, and gravitational force. In addition, the Coriolis force should be included, because all our theories and models are formulated in a rotating framework. In this chapter, I first describe surface forcing and the distribution of physical properties. I then discuss the classification of different kinds of motion in the world's oceans, and briefly review the historical development of theories of oceanic general circulation.

1.1 Surface forcing for the world's oceans

The ocean is forced from the upper surface, including wind stress, and heat and freshwater fluxes. In addition, tidal forces affect the whole depth of the water column, and geothermal heat flux and bottom friction also contribute to the establishment and regulation of the motions in the ocean. However, the surface forces are the primary forces for the oceanic circulation, and these are the focus of this section.

1.1.1 Surface wind forcing

Wind stress is probably the most crucial force acting on the upper surface of the world's oceans. The common practice in physical oceanography is to treat the effect of wind as a surface stress imposed on the upper surface of the ocean. The sea surface wind stress is usually calculated from the geostrophic wind 10 m above the sea surface, using bulk formulae. However, the air–sea interface is actually a transition zone between the atmospheric boundary layer and the oceanic boundary layer. Most importantly, the oceanic boundary layer is a wave boundary layer in which surface waves and turbulence play vitally important roles in regulating the vertical fluxes of momentum, heat, freshwater, and gases. Strictly speaking, therefore, the so-called wind stress on the sea surface should be replaced by the radiation stress between the wave boundary layer in the upper surface and the water below. Wind stress acting on the water below should depend on many dynamical aspects of these two boundary layers, such as the stability of the atmospheric boundary layer and the age of surface waves in the upper ocean. However, the discussion in this book follows the traditional approach, and the term "wind stress" is used for simplicity.

Furthermore, the distribution of wind stress on the upper ocean should be a final product of the atmosphere–ocean coupled system, and such interaction involves very complicated dynamical processes that are the subject of air–sea interactions and are beyond the scope of this book. Thus, in this book we will treat the wind stress as an external forcing for the oceanic general circulation.

Wind stress at sea level is the surface expression of the turbulent motions in the atmosphere, which occupy rather broad spectra in both space and time. It is common knowledge that wind stress changes over different time scales, from seconds to interannual and centennial time scales. The most important cycles in wind stress include the diurnal cycle and the seasonal cycle, in addition to changes on longer time scales, from interannual to decadal. Similarly, wind stress varies on spatial scales over a very broad spectrum. However, for the theory of oceanic general circulation, wind stress is normally referred to the smoothed wind stress data for large spatial scales and long time scales.

The dominant player in setting up the global wind stress pattern is the equator–pole temperature difference. Owing to this surface differential heating, atmospheric circulation is organized in the form of "Hadley cells." The prime feature of the surface wind stress is the strong westerlies associated with the Jet Stream at mid latitudes of both hemispheres. The existence of a quasi-steady circulation requires a near balance of the surface frictional torque; therefore, easterlies should exist at low latitudes. In fact, both the equatorial Pacific and equatorial Atlantic Oceans are dominated by easterlies (also known as trades) (see Fig. 1.1).

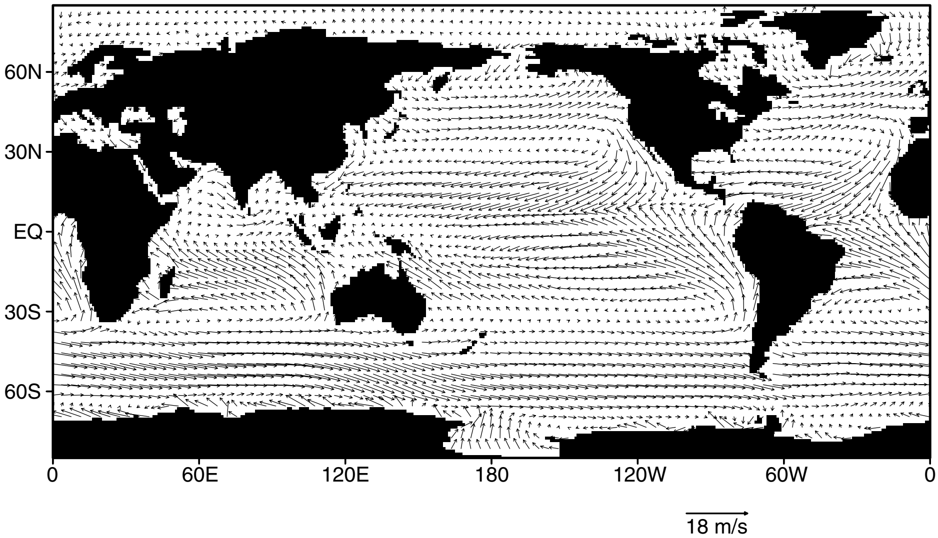

Fig. 1.1 Annual mean wind vector (in m/s) on the world's oceans, based on the European Centre for Medium-Range Weather Forecasts (ECMWF) dataset (Uppala *et al.*, 2005).

Owing to the large-scale distribution of land and ocean, wind stress on the surface is far from being zonally symmetric. Another major player in setting up the global wind stress pattern is the Earth's rotation. For example, at low latitudes the near-surface branch of the equatorward return flow of the Hadley cell is turned westward and appears as the northeast trade wind in the Northern Hemisphere and the southeast trade wind in the Southern Hemisphere. Under many such dynamical constraints, the sea surface wind stress pattern takes complicated forms. In fact, the most outstanding feature in the North Pacific Basin is the huge cyclonic wind stress pattern in the subpolar basin and the anticyclonic wind stress pattern in the subtropical basin. Similar features also exist in the Atlantic Basin and in the Southern Hemisphere, including the South Pacific, South Atlantic and South Indian Oceans.

Strong circulation is induced by wind stress in the upper kilometer of the world's oceans. The most outstanding features of wind-driven circulation include gigantic anticyclonic gyres in subtropical basins, and cyclonic gyres in subpolar basins, as shown in Figure 1.2. In addition, there is a strong circumpolar current system in the Southern Ocean, which is one of the most crucial branches of circulation in the world's oceans.

Wind stress is one of the most crucial driving forces of the oceanic circulation. As explained in Chapter 4, the westerlies at mid latitudes are responsible for the equatorward surface drift, the so-called "Ekman drift," and the easterlies at low latitudes are responsible for the poleward surface drift. The anticyclonic wind stress in the subtropical basin drives the anticyclonic circulation in the subtropical basin, and the cyclonic wind stress in the subpolar basin drives the cyclonic circulation there. In the Southern Ocean the westerly wind appears as a continuously strong belt around the whole latitudinal band; this wind

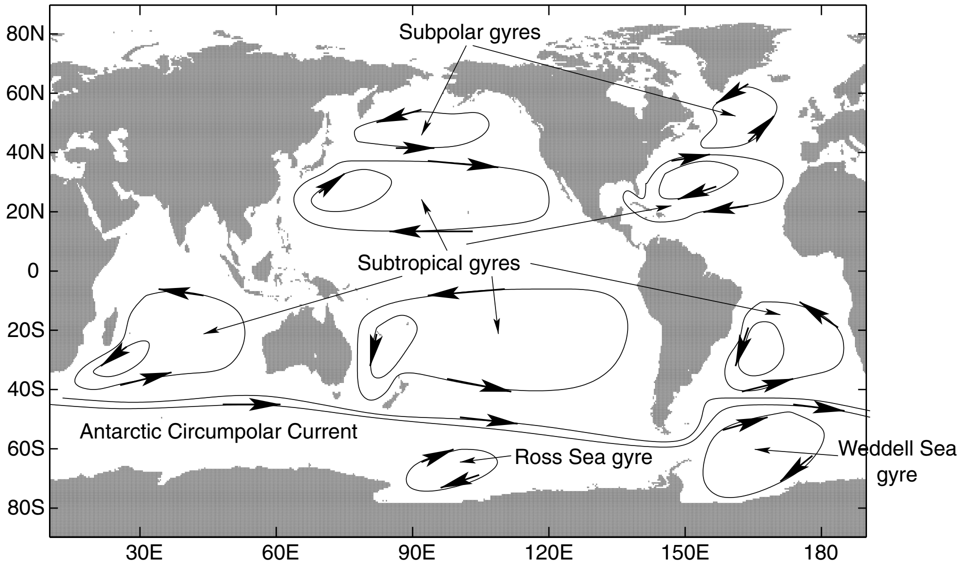

Fig. 1.2 Sketch of the major wind-driven gyres and currents in the world's oceans.

stress gives rise to the strong northward Ekman transport and upwelling, and is a direct driving force of the Antarctic Circumpolar Current (ACC).

Although a layman can observe the surface waves created by wind blowing on the sea surface, and a yachtsman can discern the surface wind drift from observation, the dynamical effects of wind stress include phenomena of huge spatial scales on the order of hundreds or thousands of kilometers – phenomena undetectable to the layman's eyes. A comprehensive understanding of the wind-driven motions in the oceans can only be achieved by systematic scientific research. In fact, the theory of the wind-driven circulation has to be developed side by side with the progress of *in situ* observations through the development of modern scientific instrumentation.

1.1.2 Surface thermohaline forcing

Heat and freshwater fluxes through the air–sea interface are the most critical forcing boundary conditions for the temperature and salinity distribution in the oceans. In addition, the oceans also receive geothermal heat from the seafloor; however, under present-day geological conditions, the amount of heat received from the seafloor is relatively small, approximately a thousand times smaller than that through the air–sea interface, so it is a rather minor contributor to the oceanic general circulation, except near the seafloor.

Surface heat flux

I first discuss the heat fluxes through the air–sea interface. The heat flux maps presented in this section are based on the NCEP-NCAR Reanalysis Project (Kistler *et al.*, 2001). In the following figures, downward heat flux into the ocean is defined as positive, and upward heat flux, leaving the oceans, is defined as negative.

The most essential forcing for the climate system on Earth is solar radiation, and this energy is in the form of short waves. Most of the energy required for maintaining the climate system can ultimately be traced back to solar radiation. Since the atmosphere is nearly transparent for solar radiation, most of it can penetrate the atmosphere and reach the lower boundary of the atmosphere, over both the land and the oceans. The amount of short-wave radiation reaching the sea surface depends primarily on the latitudinal location. Furthermore, cloudiness may be another major player in regulating the amount of solar radiation which can reach the sea surface. On the sea surface, part of the incoming short-wave radiation is reflected; thus, what the ocean receives is the net short-wave radiation, as shown in Figure 1.3. The global maxima of net short-wave radiation are closely related to the cold tongues in the eastern part of the equatorial Pacific and Atlantic Oceans.

The net short-wave radiation at each station on the sea surface is balanced by the heat transport within the ocean through advection and diffusion, plus upward heat flux through the air–sea interface. The major term in the heat flux from the ocean to the atmosphere is the latent heat flux associated with evaporation. The latent heat content for water vapor is approximately 2,500 kJ/kg; thus a relatively small amount of evaporation can transfer a large amount of heat from the ocean to the atmosphere.

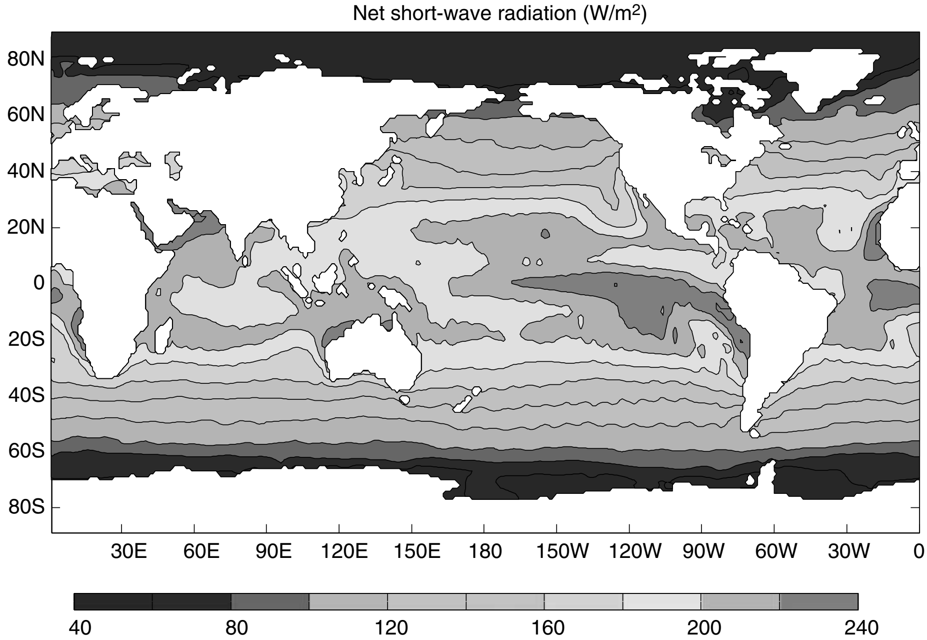

Fig. 1.3 Annual mean (NCEP-NCAR) net short-wave radiation (W/m^2). See color plate section.

There are several places where the latent heat flux is maximal (Fig. 1.4). First, the global maxima of latent heat loss exist in western boundary outflow regimes, such as the Kuroshio and the Gulf Stream, where the warm water brought by these strong western boundary currents meets cold and dry air from the continents, and strong latent heat loss is induced. These places are closely linked to a high rate of evaporation, as will be discussed shortly. Second, centers of strong latent heat loss exist at extratropics/subtropics in both hemispheres. There are areas of very low rate of latent heat flux associated with the cold tongues of surface waters in the eastern equatorial Pacific and Atlantic Oceans. Latent heat loss is generally small at high latitudes, where low sea surface temperature cannot sustain much evaporation.

The backward radiation from the ocean to the atmosphere and outer space is made up of two components: short-wave radiation reflected from the sea surface and long-wave radiation. The long-wave radiation is due to the fact that the equivalent radiation temperature of the Earth is rather low; this outgoing radiation is directly controlled by sea surface temperature and the local atmospheric conditions. The bulk formula for long-wave radiation is

$$IR \uparrow\downarrow = IR \uparrow - IR \downarrow$$

i.e., the heat flux associated with long-wave radiation is the outgoing long-wave radiation from the ocean to the atmosphere minus the backward long-wave radiation from the atmosphere to the ocean. Both these terms are proportional to the fourth power of temperature at

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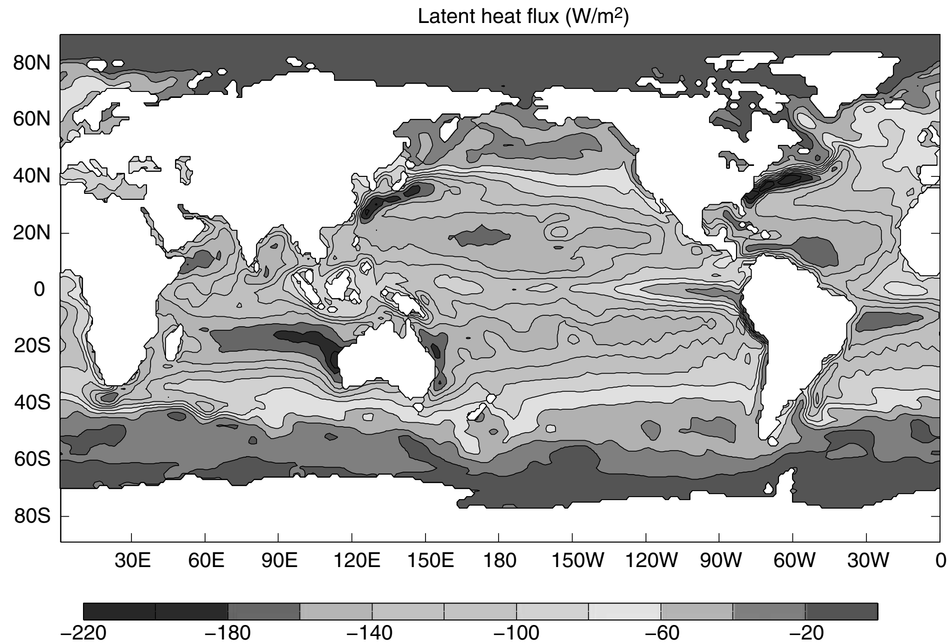

Fig. 1.4 Annual mean (NCEP-NCAR) latent heat flux due to evaporation (W/m^2). See color plate section.

the sea surface and of the atmosphere. The annual mean net heat flux of long-wave radiation is shown in Figure 1.5.

Owing to the competition of these two processes, the pattern of outgoing long-wave radiation is more complicated than other fluxes. In general, it is high near the western boundary outflow regimes in the subtropical basins of both hemispheres, especially in the Pacific Ocean. In comparison, it is much lower in the equatorial band and at high latitudes.

Sensible heat loss to the atmosphere is intimately related to the difference between the sea surface temperature and the atmospheric temperature. The most important sites of large sensible heat flux from the ocean to the atmosphere are over the Gulf Stream in the North Atlantic Ocean and the Kuroshio in the North Pacific Ocean. These high sensible heat flux regimes are clearly related to the warm water flowing in both the Gulf Stream and the Kuroshio. Note that, at high latitudes, the annual mean flux of sensible heat is actually from the atmosphere to the ocean. In particular, sensible heat flux in the Indian and Atlantic sectors of the Southern Ocean is from the atmosphere to the ocean, indicating that sea surface temperature is lower than the atmospheric temperature (Fig. 1.6). Such a low sea surface temperature is closely related to the cold deep water brought up by the strong Ekman upwelling driven by the Southern Westerlies under the current land–sea distribution.

The net air–sea heat flux, which is the sum of the four previous terms in the heat balance, is shown in Figure 1.7. As expected, there is a strong heat gain along the equatorial band, in

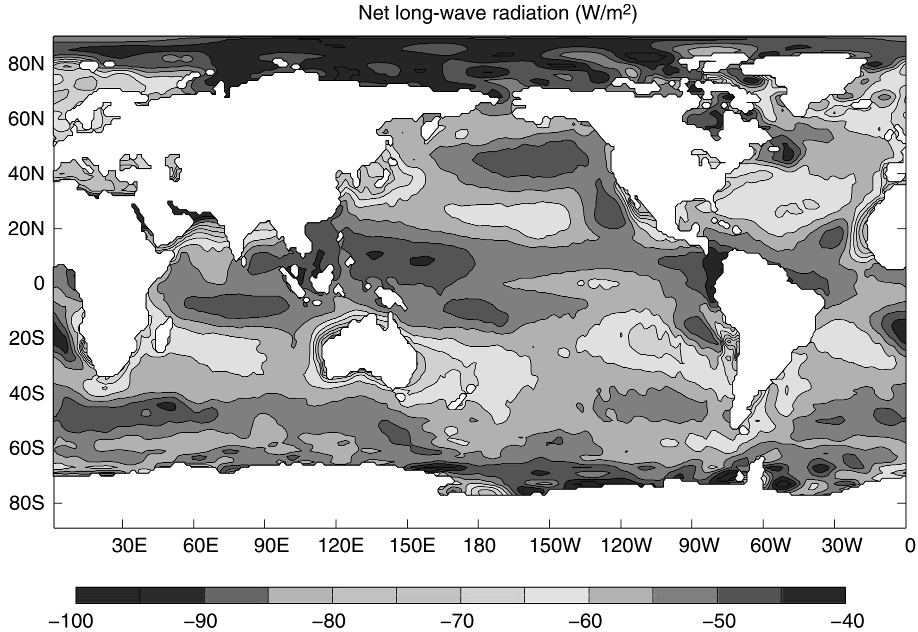

Fig. 1.5 Annual mean (NCEP-NCAR) net long-wave radiation (W/m^2). See color plate section.

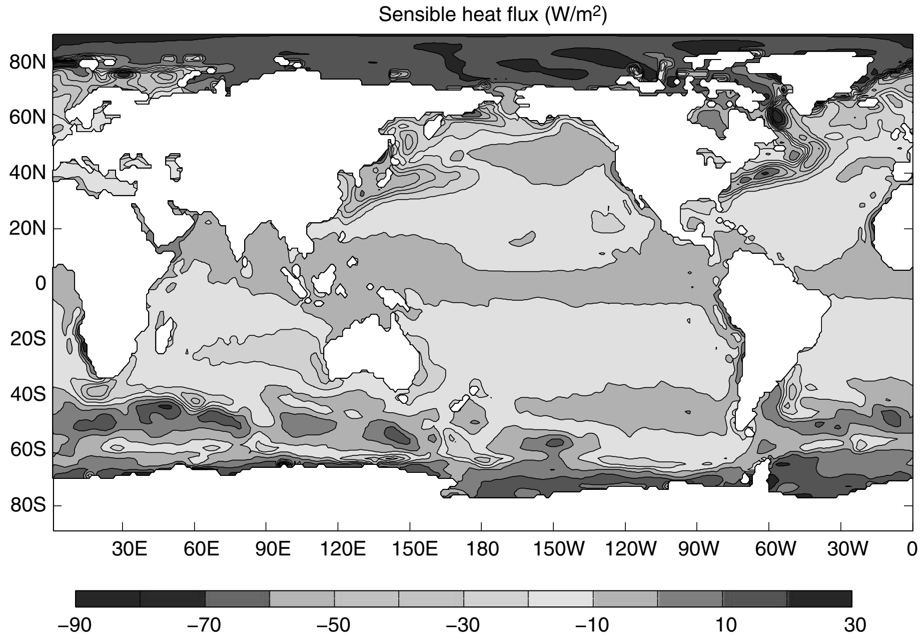

Fig. 1.6 Annual mean (NCEP-NCAR) sensible heat flux in the world's oceans (W/m^2). See color plate section.

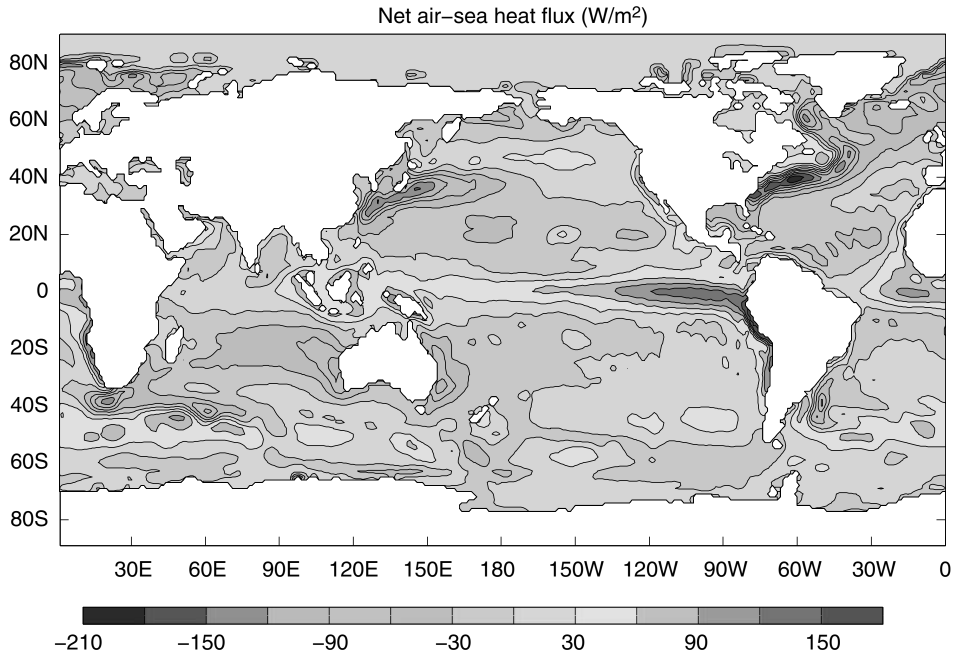

Fig. 1.7 Annual mean (NCEP-NCAR) net air–sea heat flux in the world's oceans (W/m^2). See color plate section.

particular the cold tongues in both the Pacific and Atlantic Oceans. In addition, the western coasts of South America and Africa appear as heat absorption bands, linked to the downward sensible heat flux, which is due to the low sea surface temperature associated with strong coastal upwelling. Both the Kuroshio and the Gulf Stream are major sites of heat loss in the world's oceans. The high-latitude Atlantic Ocean appears as another major site of heat loss in the world's oceans, which is related to the present-day strong meridional overturning in this basin.

Another major feature of this map is that the net heat flux is asymmetric with respect to the equator. Given the strong net heat loss at high latitudes in the Northern Hemisphere, one may expect a similar situation to occur in the Southern Hemisphere. However, a close examination reveals a different pattern. In fact, in the Indian sector and the South Atlantic sector of the Southern Ocean, the net heat flux is downward, i.e., the ocean there gains heat from the atmosphere, instead of losing heat. Comparing Figures 1.6 and 1.7, it is readily seen that these areas of net heat gain in the Southern Ocean are closely related to the downward sensible heat flux associated with the cold water upwelling driven by the strong westerlies in this latitudinal band.

The net air–sea heat flux distribution shown in Figure 1.7 implies that there is a meridional heat transport in the ocean, otherwise the underlying ocean would continuously cool or warm depending on the sign of the heat flux. In order to demonstrate the meridional heat flux, we first calculate the zonally integrated net air–sea heat flux, then integrate the zonal heat flux meridionally, starting from the South Pole, $H_f = \int_{\theta_S}^{\theta} \dot{q} a d\theta$, where a is the radius of the Earth, θ is the latitude, θ_S is the latitude of the South Pole, and $\dot{q} = \dot{q}(\theta)$ is the meridional distribution of net air–sea heat flux obtained by zonally integrating the flux shown in Figure 1.7. Accordingly, a positive slope of the curve shown in this figure indicates a downward heat flux into the ocean at the latitude of concern, and a negative slope indicates an upward heat flux at this latitude. For example, the strong positive slope over the equatorial band and the latitudinal band of 58° S–42° S indicates strong heat absorption by the ocean (Figure 1.8a).

On the other hand, a positive value of H_f indicates the northward heat flux in the oceans; thus, over the entire Northern Hemisphere, there is a poleward heat flux. As a matter of fact, in the Northern Hemisphere, the poleward heat flux reaches a maximum of nearly 2 PW (1 PW $= 10^{15}$ W) around 15° N. The corresponding poleward heat flux in the Southern Hemisphere is much smaller and changes its sign several times. In fact, the result obtained from this approach shows a northward heat flux in the latitudinal band of 58° S–20° S; however, values of poleward heat flux obtained from other more comprehensive methods indicate that meridional heat flux in the ocean is mostly southward in the Southern Hemisphere, as discussed in Section 5.3.1. Such a large discrepancy in poleward heat flux is due to the fact that the air–sea heat flux data obtained from observations is not very accurate, especially in the Southern Ocean, where reliable *in situ* observations are sparse.

Similarly, there is a strong zonal transport of heat in the ocean. In order to demonstrate the zonal heat transport, we integrate the net air–sea heat flux, starting from the longitude of the southern tip of South America. As shown in Figure 1.8b, there is a westward heat flux in the Pacific Basin, with a peak value of 1.4 PW. The zonal heat flux is a manifestation of the zonal asymmetric nature of the thermal forcing of the oceans. This zonal heat flux is intimately linked to the oceanic currents, which are discussed in later chapters.

Surface freshwater flux

The oceans exchange freshwater with the atmosphere through evaporation and precipitation, plus river run-off. The river run-off is the result of water vapor from sea surface evaporation and precipitation on land. Freshwater exchange with the atmosphere is one of the most important forcing conditions for both the oceanic general circulation and the climate system. Evaporation is the most crucial vehicle for bringing heat from low-latitude ocean to the atmosphere, where water vapor is carried poleward. Water vapor carries a large amount of latent heat, and this is one of the vital mechanisms of poleward heat transport in the climate system. Water vapor in the atmosphere eventually condenses and releases the latent heat content, returning to the oceans or land as precipitation.

Freshwater flux through the air–sea interface plays a vital role in regulating the hydrological cycle in the ocean. In particular, freshwater flux is the key ingredient in controlling the salinity distribution in the oceans. Water density is primarily controlled by temperature and salinity, and thus freshwater flux is one of the key players in regulating the thermohaline circulation through its direct connection with the salinity distribution, which is one of the most important topics related to thermohaline circulation and climate. Meridional transport