Cambridge University Press
978-1-107-08389-9 - Climate System Dynamics and Modelling
Hugues Goosse
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1 Description of the Climate System and Its Components

OUTLINE

This first chapter describes the main components of the climate system as well as some processes that will be necessary to understand the mechanisms analysed in the chapters that follow. Complementary information is available in the Glossary for readers not familiar with some of the notions introduced here.

1.1 Introduction

Climate is traditionally defined as a description in terms of the mean and variability of relevant atmospheric variables such as temperature, precipitation and wind. Climate thus can be viewed as a synthesis or aggregate of weather. This implies that portrayal of the climate in a particular region must contain an analysis of mean conditions, of the seasonal cycle and of the probability of extremes such as severe frost, storms and so on. In accordance with the standard of the World Meteorological Organisation (WMO), thirty years is the classic period for performing the **statistics** used to define climate. This is well adapted for studying recent decades because it requires a reasonable amount of data along with a good sample of the different types of weather that can occur in a particular area. However, when analysing the more distant past, such as the last glacial maximum around 21,000 years ago, climatologists are often interested in variables which are characteristic of longer time intervals. As a consequence, the thirty-year period proposed by the WMO should be considered more as a practical indicator than as a norm that must be followed in all cases. This definition of climate as representative of conditions over several decades should not, of course, obscure the fact that climate can change rapidly. Nevertheless, a substantial time interval is needed to observe a difference in climate. In general, the smaller the difference between two periods, the longer is the time required to confidently identify any climate changes between those periods.

We also must take into account the fact that the state of the atmosphere used in the preceding definition of climate is influenced by numerous processes involving not only the atmosphere but also the oceans, sea ice, vegetation and so on. Climate is therefore now defined with increasing frequency in the wider sense of a description of the **climate system**. This includes an analysis of the behaviour of

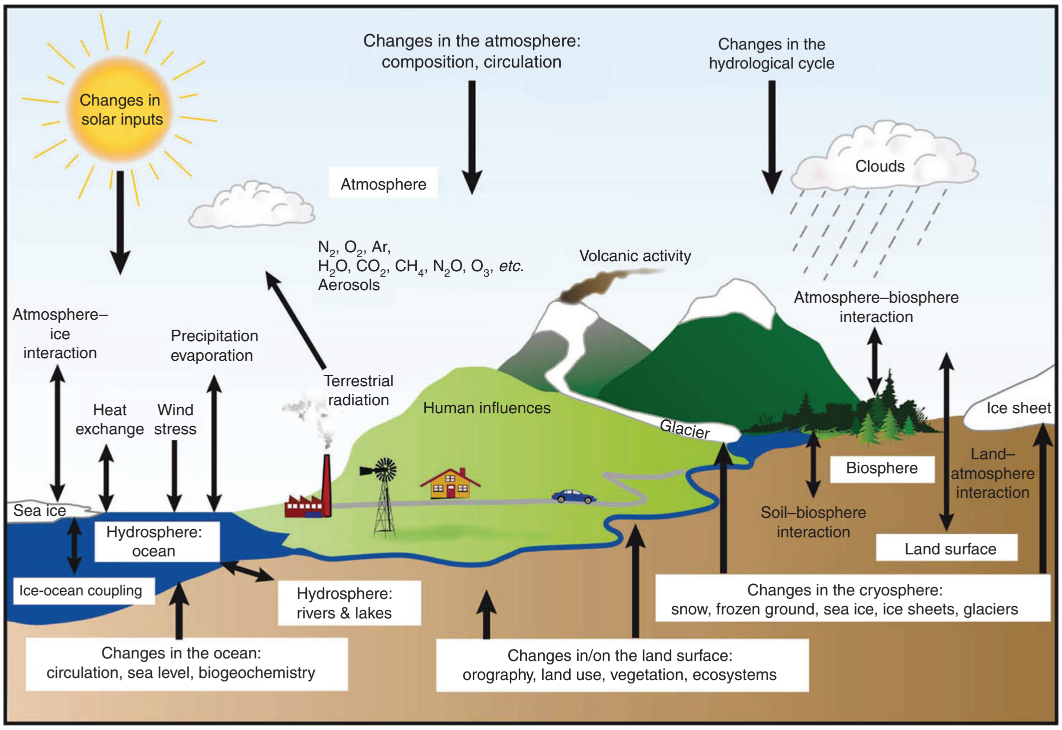

Fig. 1.1 Schematic view of the components of the climate system and their potential changes. (Source: IPCC 2007).

its five major components – the atmosphere (the gaseous envelope surrounding the Earth), the **hydrosphere** (liquid water, i.e., oceans, lakes, underground water, etc.), the **cryosphere** (solid water, i.e., **sea ice**, glaciers, ice sheets, etc.), the land surface and the **biosphere** (all the living organisms) – and of the interactions between them (IPCC 2007) (Figure 1.1). Here we will use the word 'climate' to refer to this wider definition. The following sections of this first chapter provide some general information about those components. Note that the climate system itself is often considered to be part of the broader **Earth System**, which includes all the parts of the Earth, not only the elements that are directly or indirectly related to the temperature or precipitation.

1.2 The Atmosphere

1.2.1 Composition and Temperature

Dry air is mainly composed of nitrogen (78.08% by volume), oxygen (20.95% by volume), argon (0.93% by volume) and to a lesser extent carbon dioxide (395 **ppm** or 0.0395% by volume in 2013) (see Section 2.3). The remaining fraction is made up of various trace constituents such as neon, helium, methane and krypton.

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In addition, a highly variable amount of water vapour is present in the air. This water content can be measured by the mixing ratio w, defined as the ratio between the mass of vapour and the mass of dry air; the **specific humidity** q, given by the ratio between the mass of vapour and the mass of air (i.e., including water vapour); or the partial pressure of the water vapour e, all these variables being, of course, related. When the water content is high enough to reach condensation in equilibrium conditions, the water vapour pressure is equal to the **saturation vapour pressure** e_s, and the mixing ratio is by definition w_s. The **relative humidity** is then given by the ratio w/w_s. It is nearly identical to e/e_s, which is also used directly to define the relative humidity. e_s can be expressed using the **Clausius-Clapeyron equation**, which shows that the amount of water vapour in the air at saturation strongly depends on temperature. For instance, the amount of water vapour that can be present in the atmosphere before saturation at a temperature of 20°C is more than three times higher than at 0°C. Consequently, the relative volume of water vapour in the air is close to 0% in the driest and coldest parts of the atmosphere but can reach 5% in hot regions at or close to saturation. On average, water vapour accounts for 0.25% of the mass of the atmosphere (Wallace and Hobbs 2006).

In nearly all the cases studied in climatology, the dry air and water vapour can be considered in good approximation as ideal gases (also called 'perfect gases'). The density, temperature and pressure thus are related through the **equation of state** of perfect gases, also referred to as the '**ideal gas law**'

$$p = \rho R_g T \tag{1.1}$$

where p is the pressure, ρ is the density, R_g is the gas constant (which is equal to 287.0 J K^{-1} kg^{-1} for dry air) and T is the temperature.

On a large scale, the atmosphere is very close to **hydrostatic equilibrium**, meaning that at a height z, the force due to the pressure p on a 1-m^2 horizontal surface balances the force due to the weight of the air above z. The atmospheric pressure thus is at its maximum at the Earth's surface, and the surface pressure p_s is directly related the mass of the whole air column at a particular location. Pressure then decreases with height, following approximately an exponential law:

$$p \simeq p_s e^{-z/H} \tag{1.2}$$

where H is a scale height (which is generally between 7 and 8 km for the lowest 100 km of the atmosphere). As a result of this clear and monotonic relationship between height and pressure, pressure is often used as a vertical coordinate for the atmosphere. Indeed, pressure is easier to measure than height, and choosing a pressure coordinate simplifies the formulation of some equations.

The temperature in the **troposphere**, roughly the lowest 10 km of the atmosphere, generally decreases with height. The rate of this decrease is called the '**lapse rate**' Γ:

$$\Gamma = -\frac{\partial T}{\partial z} \tag{1.3}$$

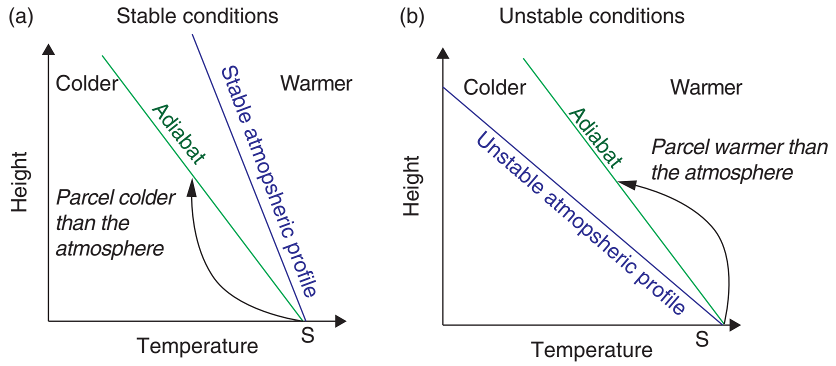

Fig. 1.2 Schematic illustration of the importance of the adiabatic lapse rate for the stability of the air. (a) If the air parcel following an adiabatic uplift becomes colder and thus denser than the surrounding air, it will tend to go downward, and the atmosphere will be stable. (b) If an air parcel becomes warmer and thus lighter than the surrounding air, it will continue to move upward, and the atmosphere will be unstable.

where T is the temperature. Its global mean value is about 6.5 K km^{-1}, or 0.0065 K m^{-1}, but Γ varies with location and season. The lapse rate depends mainly on the radiative processes in the atmosphere (see Section 2.1) and on the vertical exchanges in the air column (the resulting balance is often referred to as the 'radiative-convective equilibrium') but also on the horizontal heat transport.

The observed lapse rate determines the vertical stability of the atmosphere (Figure 1.2). If an air parcel moves upward because of a perturbation, its temperature does not remain constant: as pressure decreases with height, the parcel expands and thus cools. For dry air, this decrease in temperature with height, which is only due to expansion without any additional exchange of heat with the surrounding air, is referred to as the 'dry **adiabatic** lapse rate'. It has a value of 9.8 K km^{-1}.

For an observed lapse rate which is locally smaller than the adiabatic lapse rate (Figure 1.2a), the air parcel after an upward vertical shift will be colder than the environment and thus denser [Eq. (1.1)]. It will tend to move back to its original position, inhibiting vertical movements and leading to a stable atmosphere. Negative lapse rates (i.e., temperature increasing with height), called 'temperature inversions', therefore correspond to highly stable conditions. By contrast, if the observed lapse rate is locally larger than the adiabatic lapse rate (Figure 1.2b), an air parcel displaced upward will be warmer than the environment and less dense. Consequently, it will be further entrained upward, leading to instability, **convection** and mixing between the air parcels at different altitudes until the profile becomes identical to the adiabatic profile, restoring the stability of the air column.

If the air includes water vapour, an additional term has to be taken into account as the cooling during raising motions may induce condensation (and cloud formation) following the Clausius-Clapeyron equation. The **latent heat** released by this condensation partly compensates for the cooling that is due to the expansion. Consequently, the so-called saturated adiabatic lapse rate is lower than the dry adiabatic lapse rate. Its value depends on the temperature and pressure. It is lower

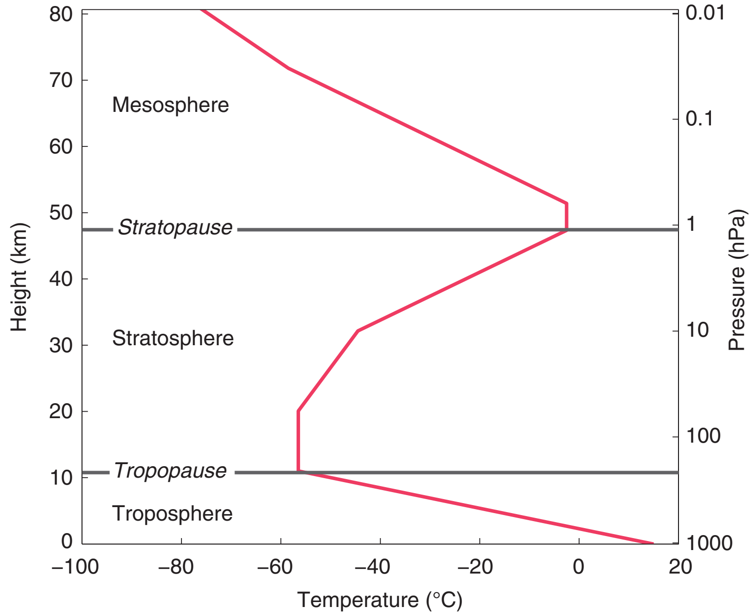

Fig. 1.3 Typical vertical temperature profile as given by the International Standard Atmosphere. (Source: From V. Zunz; reproduced with permission.)

at higher temperature because the air may contain more water vapour and is typically between 4 and 7 K km^{-1}. Convection leading to condensation is referred to as 'moist convection'.

Although complex mechanisms are involved, an important point to note is that an atmosphere in radiative equilibrium is unstable, in particular, because of the warming at the surface (see Section 2.1.6). Consequently, the air close to the surface is generally less dense than above and tends to rise. The convection processes thus are very important for the vertical structure of the atmosphere, and in many regions, particularly in the tropics, the observed lapse rate is very close to the saturated adiabatic lapse rate**.** The lapse rate is also involved in **feedbacks**, playing an important role in the response of the climate system to a perturbation (see Section 4.2.1).

At an altitude of about 10 km, a region of weak vertical temperature **gradients**, called the '**tropopause**', separates the troposphere from the **stratosphere**, where the temperature increases with height until the stratopause at around 50 km (Figure 1.3). The stratosphere therefore is generally very stable, being subject to conditions similar to temperature inversions in the troposphere. Above the stratopause, temperature decreases strongly as height increases in the mesosphere, until the mesopause is reached at an altitude of about 80 km, and then increases again in the thermosphere above this height.

The vertical temperature gradients above 10 km are strongly influenced by the absorption of solar **radiation** by different atmospheric constituents and by chemical reactions driven by the incoming light. In particular, the absorption

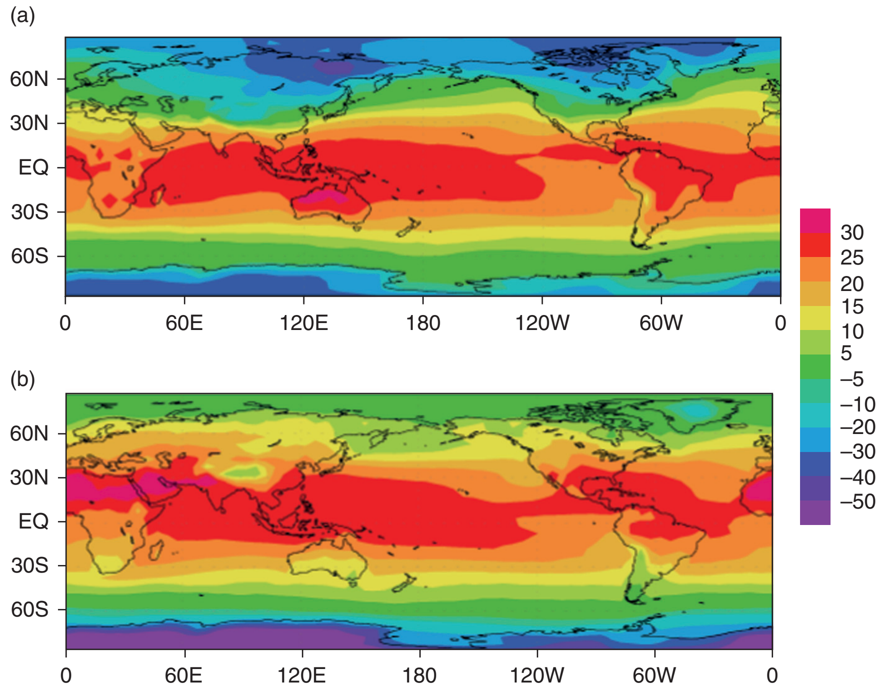

Fig. 1.4 Surface air temperature (in °C) averaged over (a) December, January and February and (b) June, July and August. (Source: Data from Brohan et al. 2006.)

of ultraviolet (UV) radiation by stratospheric **ozone**, which protects life on Earth from this dangerous radiation, plays a critical role in stratospheric warming.

Atmospheric **specific humidity** also displays a characteristic vertical profile with maximum values in the lower levels and a marked decrease as height increases. As a consequence, the air above the tropopause is nearly dry. This vertical distribution is mainly due to two processes. First, the major source of atmospheric water vapour is evaporation at the surface. Second, the warmer air close to the surface can contain a much larger quantity of water before saturation occurs than the colder air further aloft; saturation leads to the formation of water or ice droplets, clouds and eventually precipitation.

At the Earth's surface, the temperature reaches its maximum in equatorial regions (Figure 1.4) because of the higher incoming solar radiation in annual mean (see Section 2.1). In those regions, the temperature is relatively constant throughout the year. Given the much stronger seasonal cycle at middle and high latitudes, the temperature gradient between the equator and the polar regions is much larger in winter than in summer. The distribution of surface temperature is influenced by atmospheric and oceanic heat transport as well as by the thermal inertia of the ocean (see Section 2.1.5). Furthermore, the role of topography is important, with a temperature decrease at higher altitudes associated with the positive lapse rate in the troposphere.

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1.2.2 General Circulation of the Atmosphere

Not only is **convection** important for the vertical structure of the atmosphere (Section 1.2.1), but it is also responsible for horizontal movements on every scale from local or regional to global. The high temperatures at the equator make the air there less dense. It thus tends to rise before being transported poleward at high altitudes in the **troposphere**. This motion is compensated for at the surface by an equator-ward transport of air. On a motionless Earth, this big convection cell would reach the poles, inducing direct exchanges between the warmest and coldest places on Earth. However, owing to the Earth's rotation, such an atmospheric structure would be unstable (see, e.g., Marshall and Plumb 2008). Consequently, the two cells driven by the **ascendance** at the equator, called the '**Hadley cells**', close with a downward branch at about 30° latitude (Figure. 1.5). The poleward boundaries of these cells are marked by strong westerly winds in the upper troposphere called the 'tropospheric jets' or 'tropospheric jet streams'. At the surface, the Earth's rotation is responsible for a deflection of the flow coming from the middle latitudes to the equator towards the right in the northern hemisphere and towards the left in the southern hemisphere (due to the **Coriolis force**). This gives rise to the easterly **trade winds** characteristic of the tropical regions (Figure 1.6).

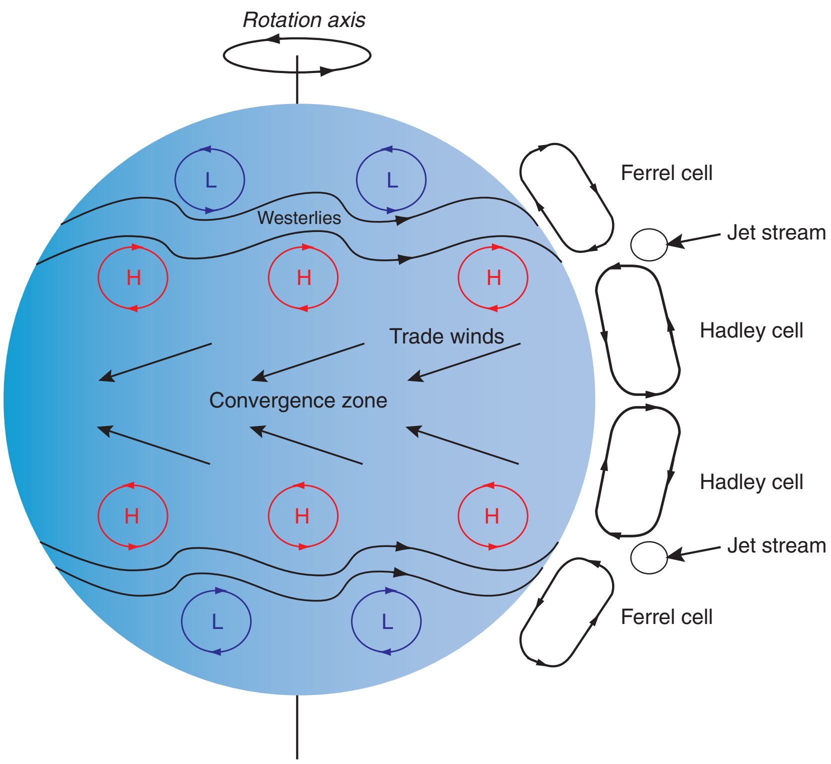

Fig. 1.5 Schematic representation of the annual mean general atmospheric circulation. H (L) represents high (low) pressure systems.

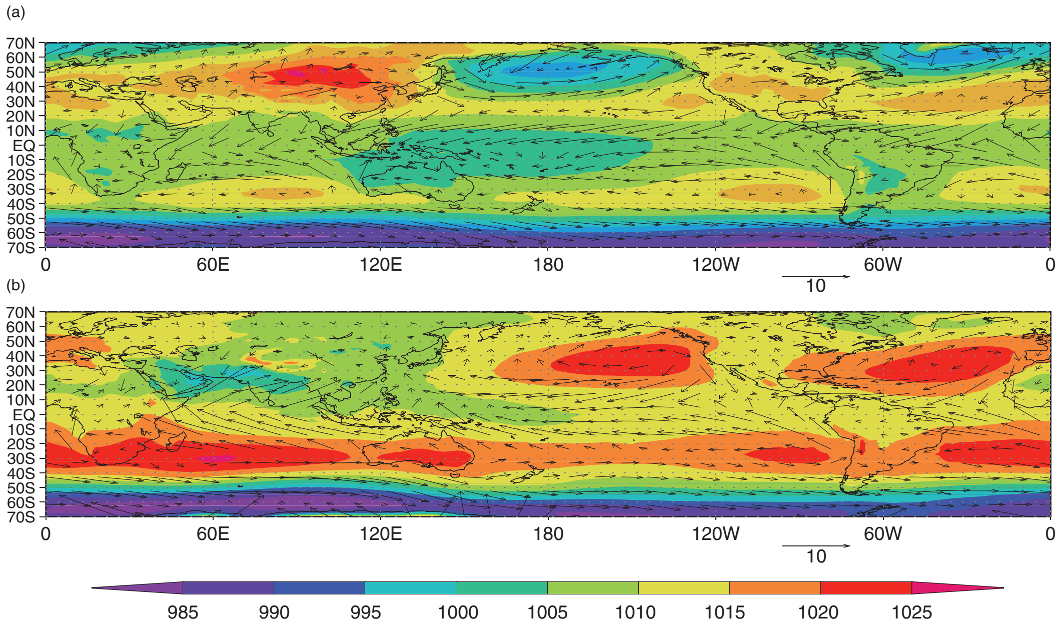

Fig. 1.6 10-m winds (arrows, in m/s) and sea-level pressure (colours, in hPa) in (a) December, January and February and (b) June, July and August. (Source: From Wouter Lefebvre using data from NCEP/NCAR **reanalyses**; Kalnay et al. 1996.)

The convergence of these surface winds and the resulting ascendance do not occur exactly at the equator but in a band called the '**Intertropical Convergence Zone**' (ITCZ). Presently, it is located around 5°N on average, with some seasonal shifts.

At the surface, the circulation at middle latitudes is characterised by westerly winds. Their **zonal** symmetry is perturbed by large wavelike patterns caused by an instability in the flow called the '**baroclinic instability**'. Such disturbances are associated with the formation of low- and high-pressure systems that govern the day-to-day variations in the weather in these regions. The dominant feature of the **meridional** circulation at these latitudes is the 'Ferrell cell', which is weaker than the Hadley cell. It is characterised by rising motion in its poleward branch and downward motion in the equator-ward branch. This contrasts with the Hadley cell, which is driven by convection and ascendance in the warmest regions near the equator. As a consequence, the Hadley cell is termed a 'direct cell', whereas the Ferrell cell is termed an 'indirect cell'.

Outside a narrow equatorial band and above the layer that interacts directly with the surface (the so-called **surface boundary layer)**, the large-scale atmospheric circulation is close to **geostrophic equilibrium**. This means that surface pressure and winds are closely related. In the northern hemisphere, the winds rotate clockwise around a high pressure and counter-clockwise around a low pressure, whilst the reverse is true in the southern hemisphere. Consequently, the middle-latitude westerlies are associated with high pressure in the subtropics and low pressure at around 50 to 60° in both hemispheres. Rather than a continuous structure, this subtropical high-pressure belt is characterised by distinct high-pressure centres,

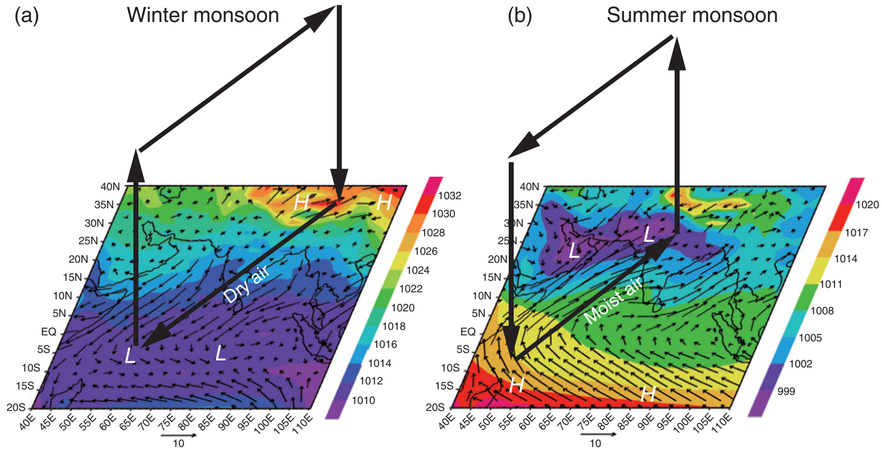

Fig. 1.7 10-m winds (arrows, in m/s) and sea-level pressure (colours, in hPa) in (a) January and (b) July illustrating the wind reversal between the winter and summer monsoons (zoom from Figure 1.6). The arrows represent a schematic view of the three-dimensional overturning circulation. The symbols *L* and *H* indicate the location of low and high pressures. (Source: NCEP/NCAR reanalyses; Kalnay et al. 1996.)

often referred to by the name of a region close to their maximum (e.g., 'Azores high' and 'St. Helena high'). In the northern hemisphere, low pressures at around 50 to 60°N manifest on **climatological** maps as cyclonic centres called, for example, the 'Icelandic low' and the 'Aleutian low'. In the Southern Ocean, because of the absence of large land masses in the corresponding band of latitude, the pressure is more zonally homogeneous, with a minimum surface pressure at around 60°S.

The preceding discussion briefly mentioned the potential impact of the presence of continents on the large-scale circulation, but the role of land surfaces becomes critical in explaining **monsoons** (Figure 1.7). In summer, the continents warm faster than the oceans because of their lower thermal inertia (see Section 2.1.5). The warming of the air close to the surface is associated with a decrease in pressure there whilst the surface pressure is higher over the ocean. This pressure difference between land and sea then induces the transport of moist air from the sea to the land. In winter, the situation reverses, with high pressure over the cold continent and a surface flow generally from land to sea. The monsoon circulation therefore exhibits clear similarities to the Hadley cell, both being driven by thermal differences, and also can be referred to as a 'direct circulation'.

Such a monsoon circulation, with seasonal reversals of wind direction, is present in many tropical areas of Africa, Asia and Australia. Nevertheless, the most famous monsoon is probably the South Asian one, which is a major feature of the Indian sub-continent climate (see the reversal of the winds in this region between Figures 1.7a and 1.7b).

1.2.3 Precipitation

Precipitation and temperature are the most important variables in defining the climate of a region. Precipitation is strongly influenced by the large-scale atmospheric circulation that transports water vapour horizontally and vertically. In particular, vertical movements are responsible for large temperature variations that play an important role in the condensation processes and therefore in precipitation.

In the upward branch of the **Hadley cell** along the ITCZ, the cooling of warm and moist surface air as it rises leads to condensation and heavy precipitation. For instance, the Western Tropical Pacific receives more than 3 m of rainfall per year. By contrast, the downward branch of the Hadley cell in the subtropics is associated with the **subsidence** of relatively dry air from the upper levels of the troposphere and thus very low precipitation rates. As a consequence, most of the large deserts on Earth are located in the subtropical belt.

The **monsoon** has a significant impact upon the precipitation over subtropical continents. During the winter monsoon, the inflow of dry continental air is associated with low precipitation. However, the summer brings moist air from the ocean which induces rainfall up to several metres in a few months.

The topography also plays a large role as it can generate significant vertical motion. Where the **ascendance** of moist air is topographically induced, massive precipitation can occur, as it does on the slopes of the Himalaya during the summer monsoon. By contrast, the subsidence of dry air, generated, for instance, because of the presence of mountains nearby, will tend to suppress precipitation, contributing to the occurrence of deserts. Mountains are also barriers to moist air coming from oceanic regions. Within this framework, the distance from the oceanic source also must be taken into account when studying the precipitation regime in a region. This explains why, for example, there is less rainfall in central Asia than in Western Europe at the same latitude.

Notable features are also present over the ocean, for instance, the South Pacific Convergence Zone (SPCZ) associated with the high precipitation rates in a northwest–southeast band from Indonesia towards 30°S, 130°W. In the middle latitudes, precipitation in winter is mainly due to **cyclones**, which tend to follow a common path at about 45°N in the Pacific and the Atlantic. This **storm track** manifests as maximum rainfall in this region. These effects are visible in the precipitation maps reproduced in Figure 1.8.

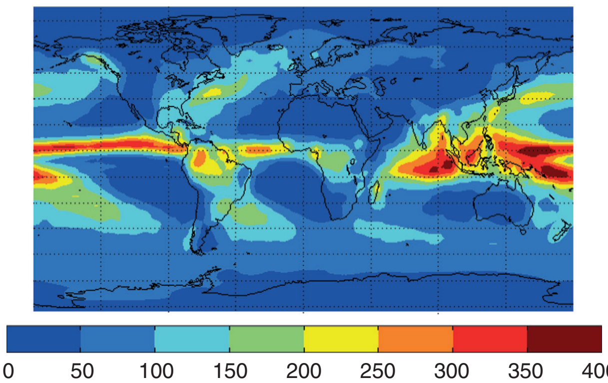

Fig. 1.8 Annual mean precipitation in centimetres per year. (Source: Xie and Arkin 1997 and updates.)