Chapter 1. Description of the climate system and its components

1.1 Introduction

Climate is traditionally defined as the description in terms of the mean and variability of relevant atmospheric variables such as temperature, precipitation and wind. Climate can thus be viewed as a synthesis or aggregate of weather. This implies that the portrayal of the climate in a particular region must contain an analysis of mean conditions, of the seasonal cycle, of the probability of extremes such as severe frost and storms, etc. Following the World Meteorological Organisation (WMO), 30 years is the classical period for performing the **statistics** used to define climate. This is well adapted for studying recent decades since it requires a reasonable amount of data while still providing a good sample of the different types of weather that can occur in a particular area. However, when analysing the most distant past, such as the last glacial maximum around 20 000 years ago, climatologists are often interested in variables characteristic of longer time intervals. As a consequence, the 30-year period proposed by the WMO should be considered more as an indicator than a norm that must be followed in all cases. This definition of the climate as representative of conditions over several decades should, of course, not mask the fact that climate can change rapidly. Nevertheless, a substantial time interval is needed to observe a difference in climate between any two periods. In general, the less the difference between the two periods, the longer is the time needed to be able to identify with confidence any changes in the climate between them.

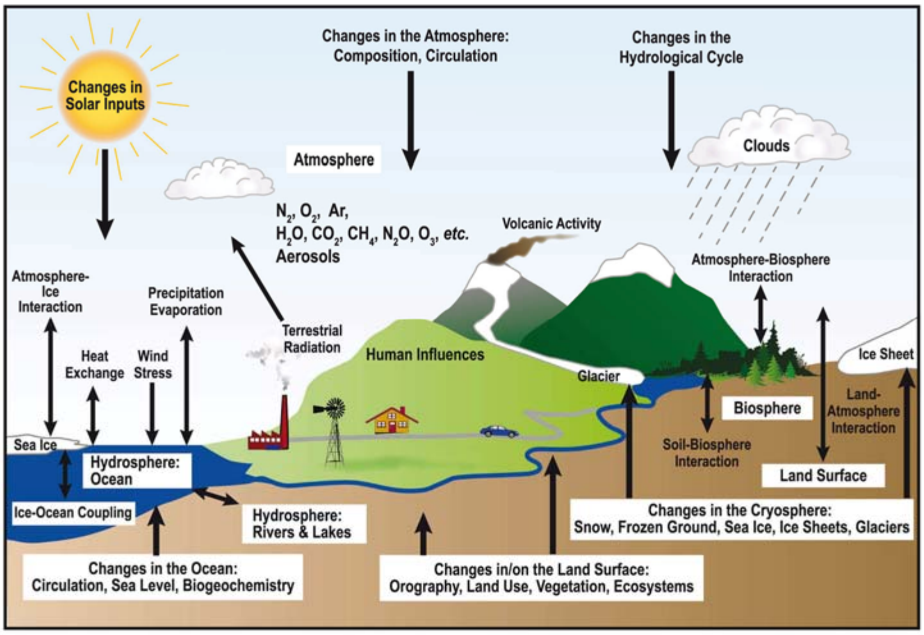

Figure 1.1: Schematic view of the components of the climate system and of their potential changes. FAQ 1.2, figure 1 from IPCC (2007) using a modified legend, published in: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, copyright IPCC 2007. Reproduced with permission.

We must also take into account the fact that the state of the atmosphere used in the definition of the climate given above is influenced by numerous processes involving not only the atmosphere but also the ocean, the sea ice, the vegetation, etc. Climate is thus now more and more frequently defined in a wider sense as the statistical description of the **climate system**. This includes the analysis of the behaviour of its five major components: the atmosphere (the gaseous envelope surrounding the Earth), the **hydrosphere** (liquid water, i.e. ocean, 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, Fig. 1.1). We will use this wider definition when we use the word **climate**. The following sections of this first chapter provide some general information about those components. Note that the climate system itself is often considered as part of the broader **Earth System**, which includes all the parts of the Earth and 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 % in volume), oxygen (20.95% in volume), argon (0.93% in volume) and to a lesser extent carbon dioxide[1] (380 ppm or 0.038% in volume). The remaining fraction is made up of various trace constituents such as neon (18 ppm), helium (5 ppm), methane[1] (1.75 ppm), and krypton (1 ppm). In addition, a highly variable amount of water vapour is present in the air. This ranges from approximately 0% in the coldest part of the atmosphere to as much as 5% in moist and hot regions. On average, water vapour accounts for 0.25% of the mass of the atmosphere.

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 is thus 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, closely following an exponential law:

$$p \simeq p_s e^{-z/H} \tag{1.1}$$

where *H* is a scale height (which is between 7 and 8 km for the lowest 100 km of the atmosphere). Because 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:

$$\Gamma = -\frac{\partial T}{\partial z} \tag{1.2}$$

where *T* is the temperature. The lapse rate depends mainly on the radiative balance of the atmosphere (see section 2.1) and on **convection** as well as on the horizontal heat transport. Its global mean value is about 6.5 K km^{-1}, but Γ varies with the location and season.

[1] The concentrations of carbon dioxide and methane are changing quickly (see section 2.3).

The lapse rate is an important characteristic of the atmosphere. For instance, it determines its vertical stability. For low values of the lapse rate, the atmosphere is very stable, inhibiting vertical movements. Negative lapse rates (i.e. temperature increasing with height), called temperature inversions, correspond to highly stable conditions. When the lapse rate rises, the stability decreases, leading in some cases to vertical instability and convection. 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 generally increases with height until the stratopause at around 50 km (Fig. 1.2). Above the stratopause, temperature decreases strongly with height 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 **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 warming in the stratosphere at heights of about 30-50 km is mostly due to the absorption of ultraviolet **radiation** by stratospheric **ozone**, which protects life on Earth from this dangerous **radiation**.

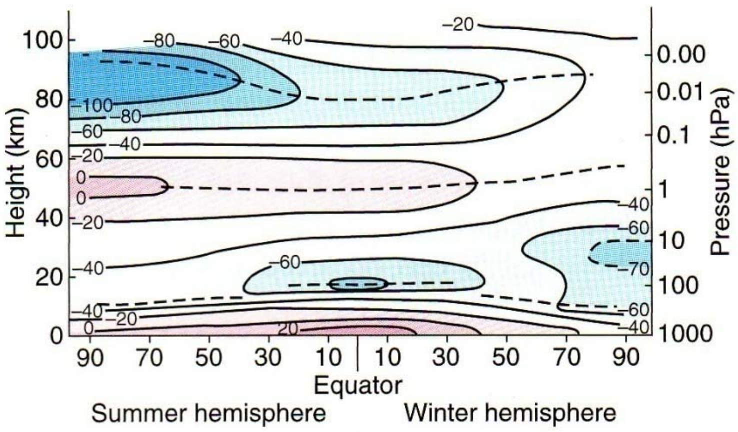

Figure 1.2: Idealised **zonal** mean temperature (in °C) in the atmosphere as a function of the height (or of the pressure). The dashed lines represent schematically the location of the tropopause, stratopause and mesopause. This figure was published in Atmospheric science: an introductory survey, Wallace and Hobbs, International Geophysics Series 92, Copyright Elsevier (Academic Press) 2006.

Atmospheric **specific humidity** also displays a characteristic vertical profile with maximum values in the lower levels and a marked decrease with height. 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. Secondly, the warmer air close to the surface can contain a much larger quantity of water before it becomes saturated than the colder air further away; saturation that 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 (Fig. 1.3) because of the higher incoming radiations (see section 2.1). In those regions, the temperature is relatively constant throughout the year. Because of the much stronger seasonal cycle at mid and high latitudes, the north-south gradients are much larger in winter than in summer. The distribution of the surface temperature is also 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.

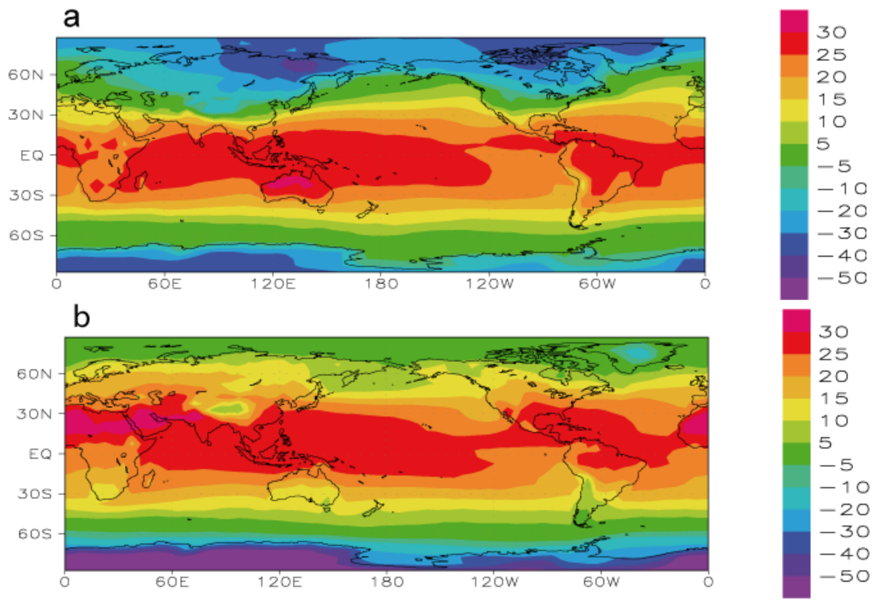

Figure 1.3: Surface air temperature (in °C) averaged over (a) December, January, and February and (b) June, July, and August. Data source : Brohan et al. (2005). http://www.cru.uea.ac.uk/cru/data/temperature/.

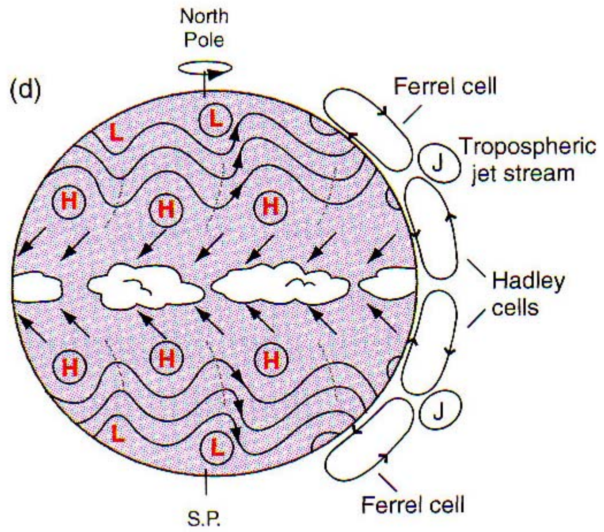

Figure 1.4: Schematic representation of the annual mean general atmospheric circulation. H (L) represents high (low) pressure systems. This figure was published in Atmospheric science: an introductory survey, Wallace and Hobbs, International Geophysics Series 92, Copyright Elsevier (Academic Press) 2006.

1.2.2 General circulation of the atmosphere

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 equatorward displacement of the 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, because of the Earth's rotation, such an atmospheric structure would be unstable. Consequently, the two cells driven by the **ascendance** at the Equator, called the **Hadley cells**, close with a downward branch at a latitude of about 30° (Fig. 1.4). The northern boundary of these cells is marked by strong westerly winds in the upper troposphere called the tropospheric jets. At the surface, the Earth's rotation is responsible for a deflection toward the right in the northern hemisphere and toward the left in the southern hemisphere (due to the **Coriolis force**) of the flow coming from the mid-latitudes to the Equator. This gives rise to the easterly **trade winds** characteristics of the tropical regions (Fig. 1.5).

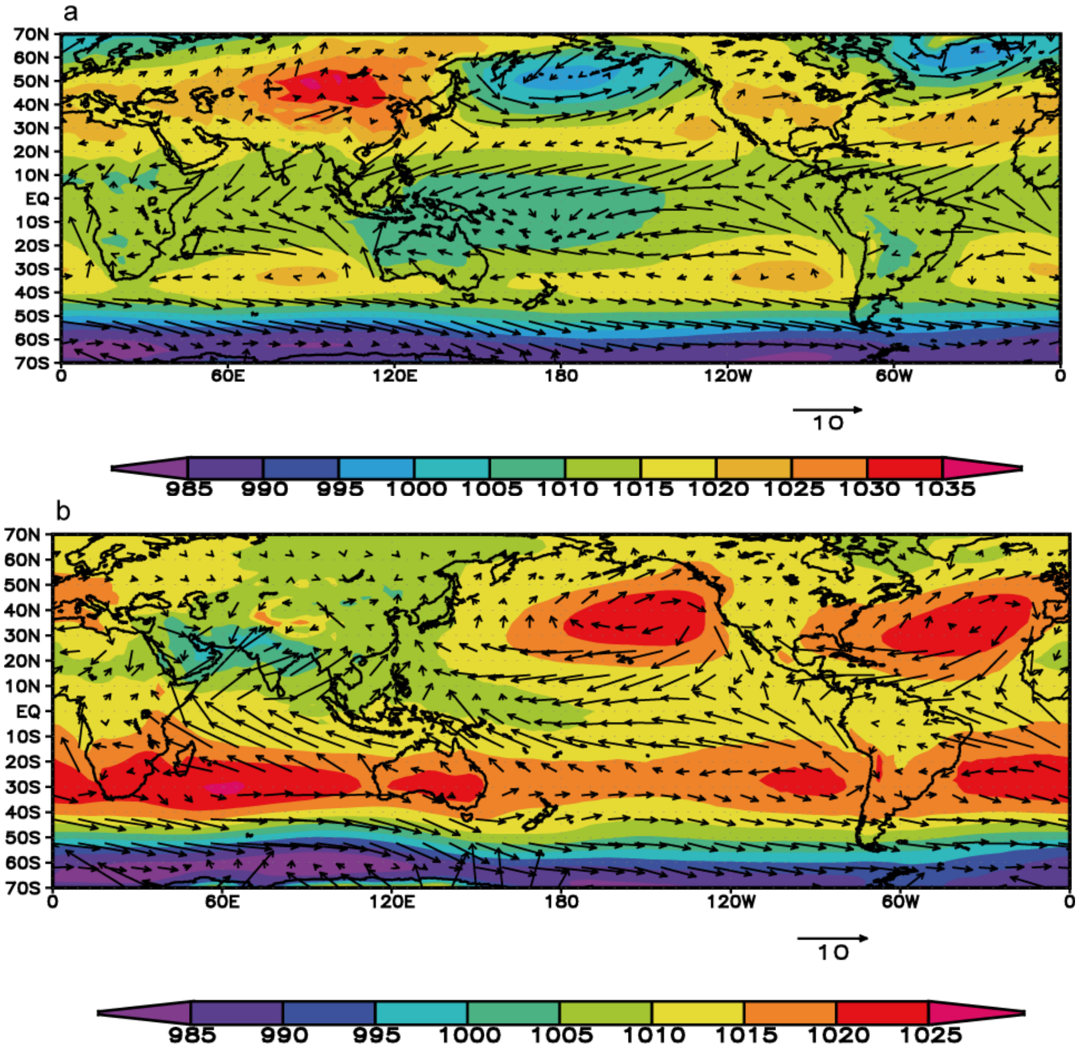

Figure 1.5: 10m winds (arrows, in m/s) and sea level pressure (colours, in hPa) in (a) December, January, and February and (b) June, July, and August. Data source: NCEP/NCAR **reanalyses** (Kalnay et al. 1996).

The extratropical circulation is dominated at the surface by westerly winds whose **zonal** symmetry is perturbed by large wave-like patterns and the continuous succession of disturbances that governs the day-to-day variations in the weather in these regions. The dominant feature of the **meridional** circulation at those latitudes is the Ferrell cell, which is weaker than the **Hadley cell**. As it is characterized by rising motion in its poleward branch and downward motion in the equator branch, it is termed an indirect cell by contrast with the Hadley cell, which is termed a direct cell.

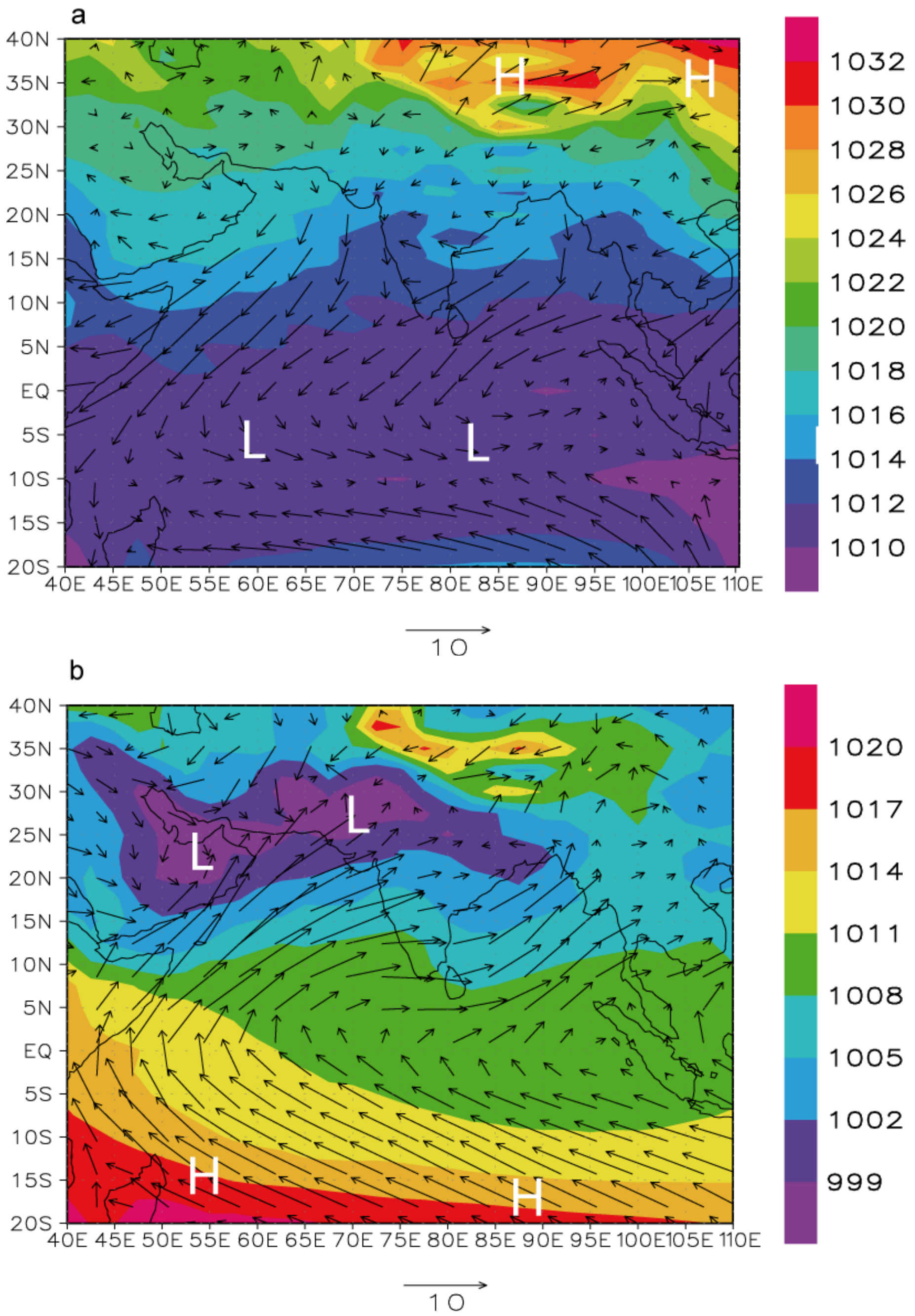

Figure 1.6: 10 m winds (arrows, in m/s) and sea level pressure (colours, in hPa) in (a) January and (b) July illustrating the wind reversal between the winter and the summer monsoon. Data source: NCEP/NCAR **reanalyses** (Kalnay et al. 1996).

Outside a narrow equatorial band and above the **surface boundary layer**, the large-scale atmospheric circulation is close to **geostrophic equilibrium**. The surface pressure and winds are thus closely related. In the Northern Hemisphere, the winds rotate clockwise around a high pressure and conterclockwise around a low pressure, while the reverse is true in the Southern Hemisphere. Consequently, the mid-latitude westerlies are associated with high pressure in the subtropics and low pressure at around 50-60°. Rather than a continuous structure, this subtropical high pressure belt is characterised by distinct high pressure centres, 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, the low pressures at around 50-60°N manifest themselves on **climatological** maps as cyclonic centres called 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 homogenous, with a minimum in surface pressure around 60°S.

In the real atmosphere, the convergence of surface winds and the resulting **ascendance** does not occur exactly at the equator but in a band called the **Intertropical Convergence Zone** (ITCZ). Because of the present geometry of the continents, it is located around 5°N, with some seasonal shifts. The presence of land surfaces also has a critical role in **monsoon** circulation. In summer, the continents warm faster than the oceans because of their lower thermal inertia (see section 2.1.5). This induces a warming of the air close to the surface and a decrease in surface pressure there. This pressure difference between land and sea induces a transport of moist air from the sea to the land. In winter, the situation reverses, with high pressure over the cold continent and a flow generally from land to sea. Such a monsoonal circulation, with seasonal reversals of the wind direction, is present in many tropical areas of Africa, Asia and Australia. Nevertheless, the most famous monsoon is probably the South Asian one that strongly affects the Indian sub-continent (Fig. 1.6).

1.2.3 Precipitation

The large-scale atmospheric circulation has a strong influence on precipitation, which is, with temperature, the most important variable in defining the climate of a region. Along the ITCZ, the cooling of warm and moist surface air during its rising motion leads to condensation and heavy precipitation in this area. For instance, the western tropical Pacific receives more than 3 m of rainfall per year. By contrast, the downward motion in the subtropics is associated with the presence of very dry air and very low precipitation rates. As a consequence, the majority of the large deserts on Earth are located in the sub-tropical belt.

The **monsoon** strongly affects the precipitation over subtropical continents. During the winter **monsoon**, the inflow of dry continental air is associated with low precipitation. On the other hand, the summer brings moist air from the ocean inducing rainfall that can reach several metres in a few months. The topography also plays a large role as it can generate significant vertical motion. Where **ascendance** of moist air is topographically induced, massive precipitation can occur, as on the slopes of the Himalaya during the summer monsoon. If **subsidence** of dry air is generated at a location because of the mountains nearby, the precipitation will be lower, even contributing to the presence of a desert. Mountains are also barriers to moist air coming from oceanic regions. In this framework, the distance from the oceanic source must also be taken into account when studying the precipitation regime in a region. This explains why, for example, rainfall is less 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 NW-SE band from Indonesia towards 30°S, 130W. In the mid-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 itself as maximum rainfall in this region. These effects are visible in the precipitation maps reproduced in Fig. 1.7.

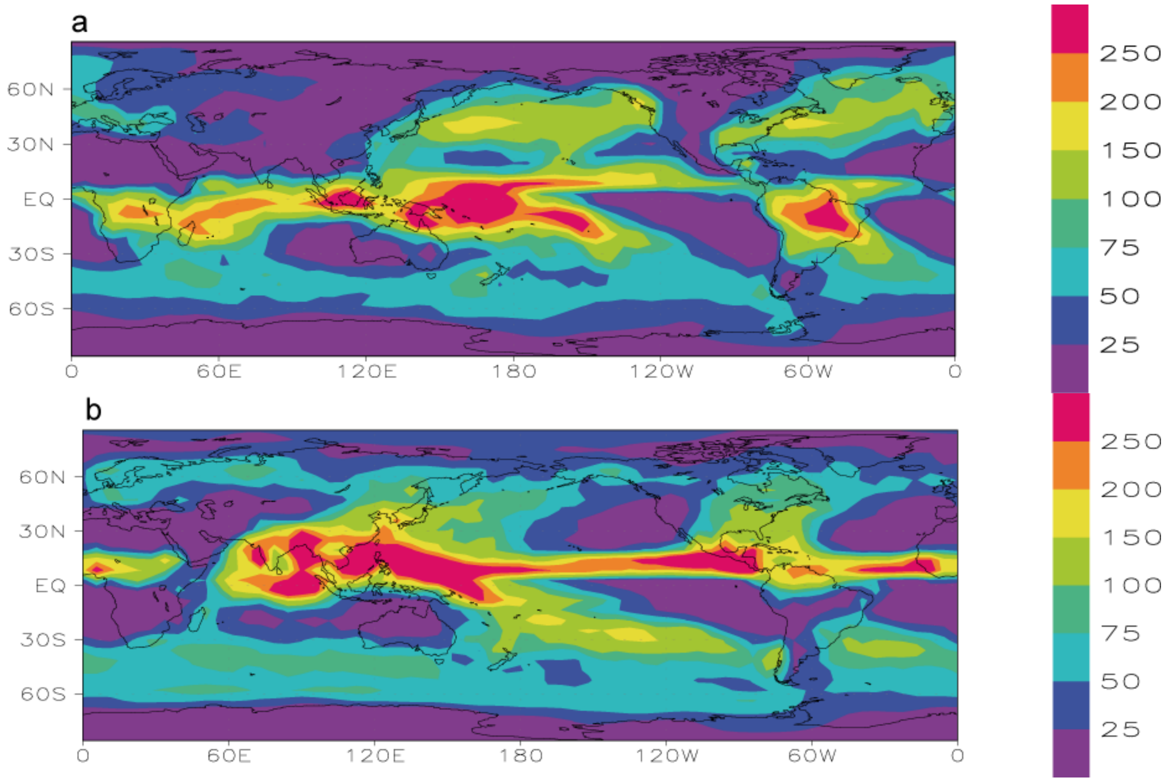

Figure 1.7: Precipitation in cm in (a) December, January and February; and (b) June, July and August. Data source: Xie, P., and P. A. Arkin, 1997 and updates

1.3 The ocean

1.3.1 Composition and properties

The ocean is a very important element of the climate system which covers about 71% of the Earth's surface and has an average depth of roughly 3700 m. Sea water is composed of 96.5% water and 3.5% dissolved salts, particles, gases and organic matter. The most important of these components are chloride and sodium, which represent about 85% of the dissolved material. Although the total quantity of dissolved salts varies from place to place, their relative contribution is very stable in sea water. Rather than specifying each of the components, it is thus very convenient to define a bulk salinity as the total amount of dissolved material (in grams) in a kilogram of sea water. This dimensionless salinity is then given in ‰ (parts per thousand). However, in practice, measuring the total amount of dissolved material is difficult. The salinity scale is thus based nowadays on the conductivity of sea water and given in psu (practical salinity unit). For simplicity, this new scale has been chosen so that the salinity in psu is very close to that in ‰.

The density of sea water increases with salinity as well as with pressure (thus with depth as the ocean is also in **hydrostatic equilibrium** on a large-scale), while it decreases with increasing temperature. In a very simplified picture, it is often considered that temperature dominates the density changes at high temperatures, while the role of salinity is larger at low temperatures. Salinity also influences the freezing point of sea water which, at the surface, decreases from 0°C for pure water to -1.8°C at a salinity of 35 psu.

1.3.2 Oceanic circulation

The surface ocean circulation is mainly driven by the winds. At mid-latitudes, the atmospheric westerlies induce eastward currents in the ocean while the **trade** winds are responsible for westward currents in the tropics (Fig. 1.8). Because of the presence of continental barriers, those currents form loops called the subtropical **gyres**. The surface currents in those gyres are intensified along the western boundaries of the oceans (the east coasts of continents) inducing well-known strong currents such as the **Gulf Stream** off the east coast of the USA and the Kuroshio off Japan. At higher latitudes in the Northern Hemisphere, the easterlies allow the formation of weaker subpolar gyres. In the Southern Ocean, because of the absence of continental barriers, a current that connects all the ocean basins can be maintained: the Antarctic Circumpolar Current (ACC). This is one of the strongest currents on Earth, which transports about 130 Sv (1 Sverdrup = 10^6 $m^3\ s^{-1}$). All these currents run basically parallel to the surface winds. By contrast, the equatorial counter-currents, which are present at or just below the surface in all the ocean basins, run in the direction opposite to the trade winds.

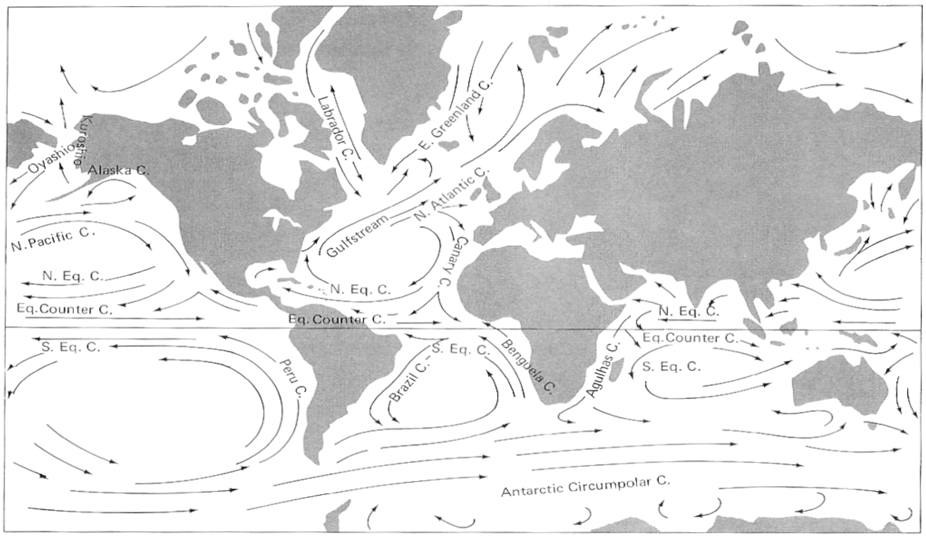

Figure 1.8: Schematic representation of the major surface currents. Eq. is an abbreviation for equatorial, C. for current, N. for North, S. for South and E. East. Reprinted by permission of Waveland Press, Inc. From Knauss, Introduction to Physical Oceanography. (Long Grove, IL: Waveland Press, Inc, 1997 (reissued 2005)). All rights reserved.

Because of the Earth's rotation, the ocean transport induced by the wind is perpendicular to the wind stress (to the right in the Northern Hemisphere, to the left in the Southern Hemisphere). This transport, known as the **Ekman transport**, plays an important role in explaining the path of the wind-driven surface currents (Fig. 1.8). Furthermore, along a coastline or if the transport has horizontal variations, this can lead to surface convergence/divergence that has to be compensated by vertical movements in the ocean. An important example is the equatorial upwelling (Fig 1.9). In the Northern Hemisphere, the **Ekman transport** is directed to the right of the easterly wind stress and is thus northward. By contrast, it is southward in the Southern Hemisphere. This results in a divergence at the surface at the equator that has to be compensated by an **upwelling** there. In coastal **upwelling**, the wind stress has to be parallel to coast, with the coast on the left when looking in the wind direction in the northern hemisphere (for instance, northerly winds along a coast oriented north-south). This causes an offshore transport and an upwelling to compensate for this transport.

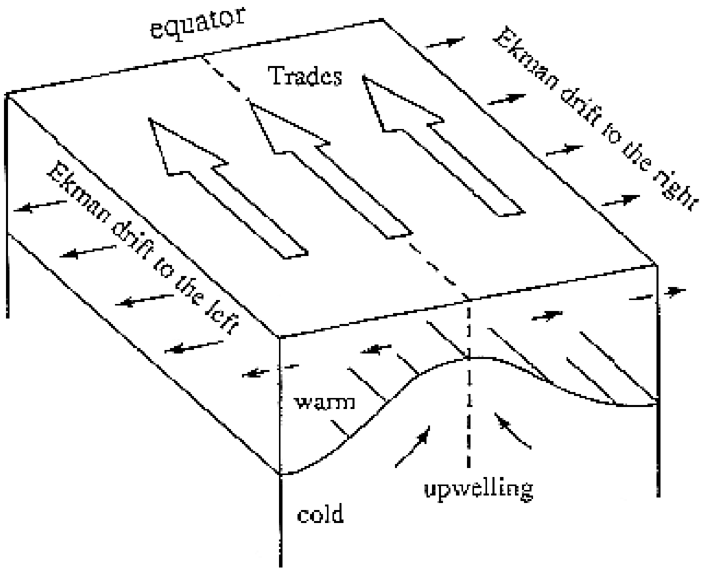

Figure 1.9: Schematic representation of the equatorial upwelling. Figure from Cushman-Roisin et al. (1994). Copyright B. Cushman-Roisin (1994), reproduced with permission.

At high latitudes, because of its low temperature and relatively high salinity, surface water can be dense enough to sink to great depths. This process, often referred to as deep oceanic **convection**, is only possible in a few places in the world, mainly in the North Atlantic and in the Southern Ocean. In the North Atlantic, the Labrador and Greenland-Norwegian Seas are the main sources of the North Atlantic Deep Water (NADW) which flows southward along the western boundary of the Atlantic towards the Southern Ocean. There, it is transported to the other oceanic basins after some mixing with ambient water masses. This deep water then slowly **upwells** towards the surface in the different oceanic basins. This is very schematically represented on Fig.1.10 by upward fluxes in the North Indian and North Pacific Oceans. However, while sinking occurs in very small regions, the **upwelling** is broadly distributed throughout the ocean. The return flow to the sinking regions is achieved through surface and intermediate depth circulation. In the Southern Ocean, Antarctic Bottom Water (AABW) is mainly produced in the Weddell and Ross Seas. This **water mass** is colder and denser than the NADW and thus flows below it. Note that, because of the mixing of water masses of different origins in the Southern Ocean, the water that enters the Pacific and Indian basins is generally called Circumpolar Deep Water (CDW).

This large-scale circulation (Fig. 1.10), which is associated with currents at all depths, is often called the oceanic **thermohaline circulation** as it is driven by temperature and salinity (and thus density) contrasts. However, winds also play a significant role in this circulation. First, they influence the surface circulation and thus the upper branch of the thermohaline circulation which feeds the regions where sinking occurs with dense enough surface waters. Secondly, because of the divergence of the **Ekman transport**, the winds influence the **upwelling** of deep water masses towards the surface in some regions. This plays a particularly important role in the Southern Ocean. Winds could also act as a local/regional preconditioning factor that favours deep convection.

The thermohaline circulation is quite slow. The time needed for water masses formed in the North Atlantic to reach the Southern Ocean is of the order of a century. If the whole cycle is taken into account, the **timescale** is estimated as several centuries to a few millennia, depending of the exact location and mechanism studied. On the other hand, this circulation transports huge amounts of water, salts and energy. In particular, the rate of NADW formation is estimated to be around 15 Sv. Uncertainties are larger for the Southern Ocean, but the production rate of AABW is likely quite close to that of NADW. As a consequence, the thermohaline circulation has a important role in oceanographic as well as in climatology (see section 2.2).

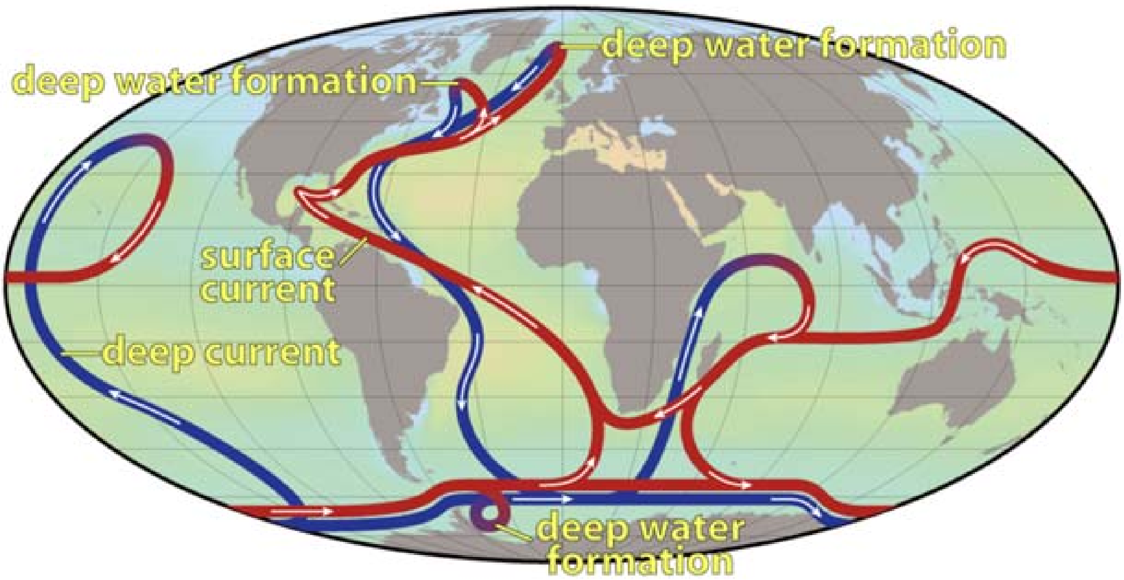

Figure 1.10: Schematic representation of the oceanic thermohaline circulation. *Source*: http://en.wikipedia.org/wiki/Thermohaline_circulation. Author Robert Simmon, NASA with Minor modifications by Robert A. Rohde. http://en.wikipedia.org/wiki/User:Dragons_flight. Not protected by copyright.

1.3.3 Temperature and salinity

1.3.3.1 Surface layer

Because of the strong interactions between the ocean and the atmosphere, the sea surface temperature (SST) (Fig.1.11) is very close to the temperature of the air above it (Fig. 1.3). One exception is the polar regions where **sea ice** (see section 1.4) insulates the ocean from the cold polar atmosphere.

The sea surface salinity is strongly influenced by the freshwater fluxes at the surface. The salinity reaches a maximum in subtropical areas because of the large evaporation and low rainfall there. The high precipitation rates induce lower salinity at the equator, while the weak evaporation is responsible for the lower salinity observed at mid and high latitudes. River input also has a large regional impact, as seen in Fig. 1.11, with low values close to the mouths of the Amazon and Mississippi rivers.

As with the surface temperature over land, the first tenths of metres of the ocean at mid- and high latitudes show a clear seasonal cycle (Fig. 1.12). However, this cycle is shifted by 1 to 3 months compared to that of land surface temperature and its amplitude is weaker because of the large thermal inertia of the ocean (see section 2.1). In winter, the stirring by the winds and the cooling at the surface, which tend to destabilise the water column and generate shallow **convection**, induce strong mixing in the ocean. This homogenises a surface layer, called the **oceanic mixed layer**. Its depth is generally about

50 to 100 metres in winter but can reach several hundred metres in some regions. When temperature rises in spring and summer, the density at the surface decreases. This tends to stabilise the water column. As the winds also tend to be weaker in spring and summer, generating less mixing, the mixed layer becomes shallower. The warming is thus concentrated in a shallow layer, whose depth is generally lower than 40m. Below this summer mixed layer, the temperature is insulated from the surface and thus still conserves the properties that it has acquired by contact with the atmosphere in winter. This seasonal process induces the formation of a region with strong vertical **gradients** at the base of the summer mixed layer referred to as the seasonal **thermocline** (Fig 1.12).

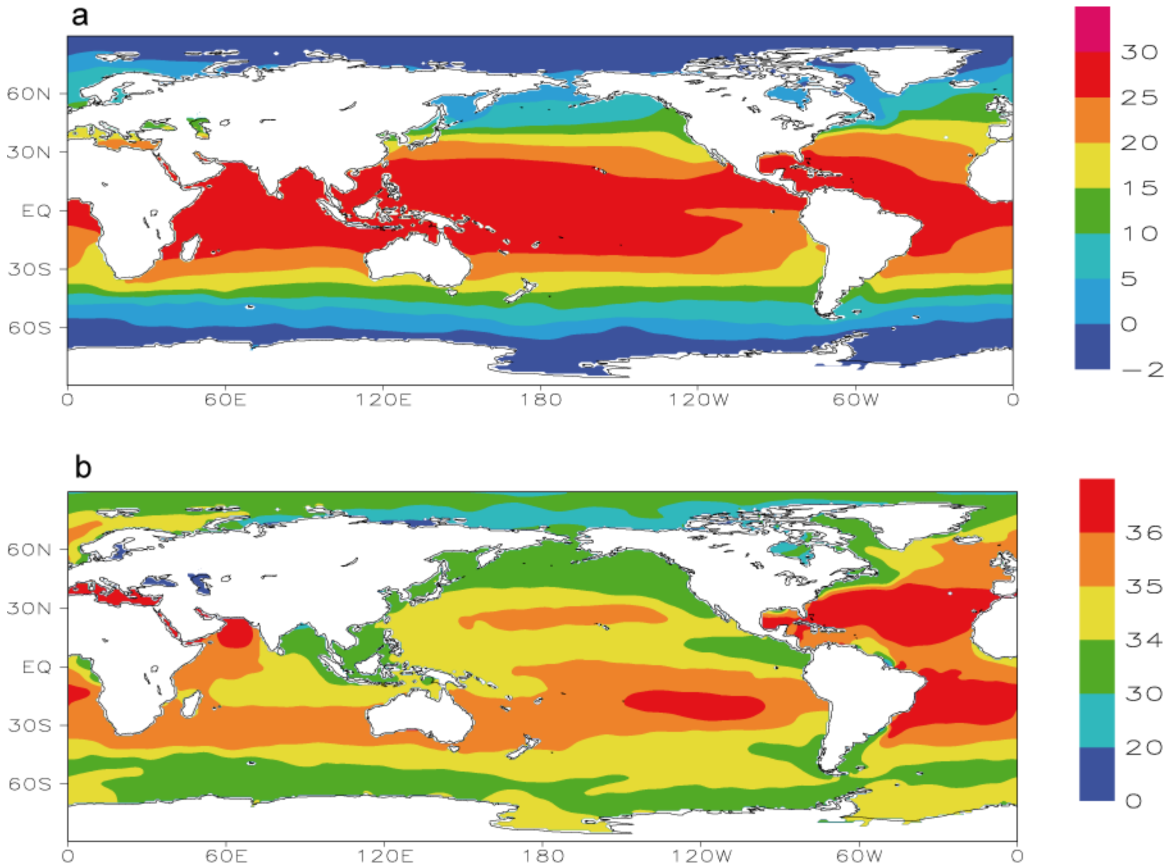

Figure 1.11: (a) Annual mean sea surface temperature (°C) and (b) surface salinity (psu). Data source: Levitus (1998).

The mixed layer dynamics, and in particular the seasonal changes in its depth, have a considerable influence on the surface ocean properties and on the exchanges of heat, water and gases between the ocean and the atmosphere (see section 2.1). Its development also has a large impact on the growth of phytoplankton, which is the basis of the whole oceanic food web. In order for photosynthesis to occur, phytoplankton need light, which is only available close to the surface. If the mixed layer is deep, as in winter, the phytoplankton is mixed over a large depth range by surface turbulence and thus spends a large part of its time in the dark, deep levels. If the relatively low flux of solar radiation at the surface during winter is also taken into account, it is easy to see why photosynthesis cannot take place in winter. By contrast, the availability of light is high in summer because of the shallow mixed layer and the large amount of incoming solar radiation. However, the shallow mixed layer limits the exchanges between the surface and the deep water which is rich in the nutrients required by the phytoplankton (see section 2.3). The concentration of those nutrients is thus generally too low to sustain a large production in summer. As a consequence, the phytoplankton growth generally reaches its maximum during spring **blooms**. The mixed layer is relatively shallow during this period, but the nutrient concentration is high enough, thanks to the exchanges with deeper layers that occurred during the previous winter.

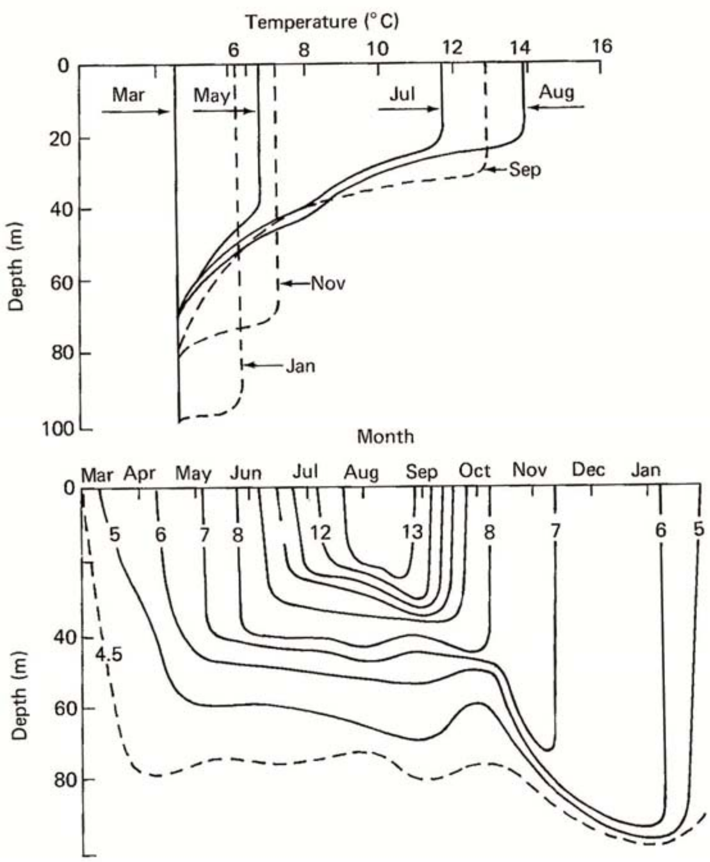

Figure 1.12: Typical growth and decay of the seasonal thermocline at a mid-latitude site in the Northern Hemisphere (50° N, 145° W). Reprinted by permission of Waveland Press, Inc. From Knauss, Introduction to Physical Oceanography. (Long Grove, IL: Waveland Press, Inc, 1997 (reissued 2005)). All rights reserved.

1.3.3.2 Intermediate and deep layers

Below the mixed layer, a strong vertical temperature **gradient** is observed (except in some regions at high latitudes), which defines the permanent **thermocline** (Figs. 1.13 and 1.14). This shows that the majority of the ocean is strongly stratified, meaning that light water sits above dense water as required by the vertical stability of the water column. In the deep ocean, the vertical gradients are much weaker. It is somehow surprising that, near the equator, the temperature difference between the surface and a depth of 1000 m could be more than 20°C while the temperature difference between 1000 m and the ocean bottom is only of the order of 3°C.

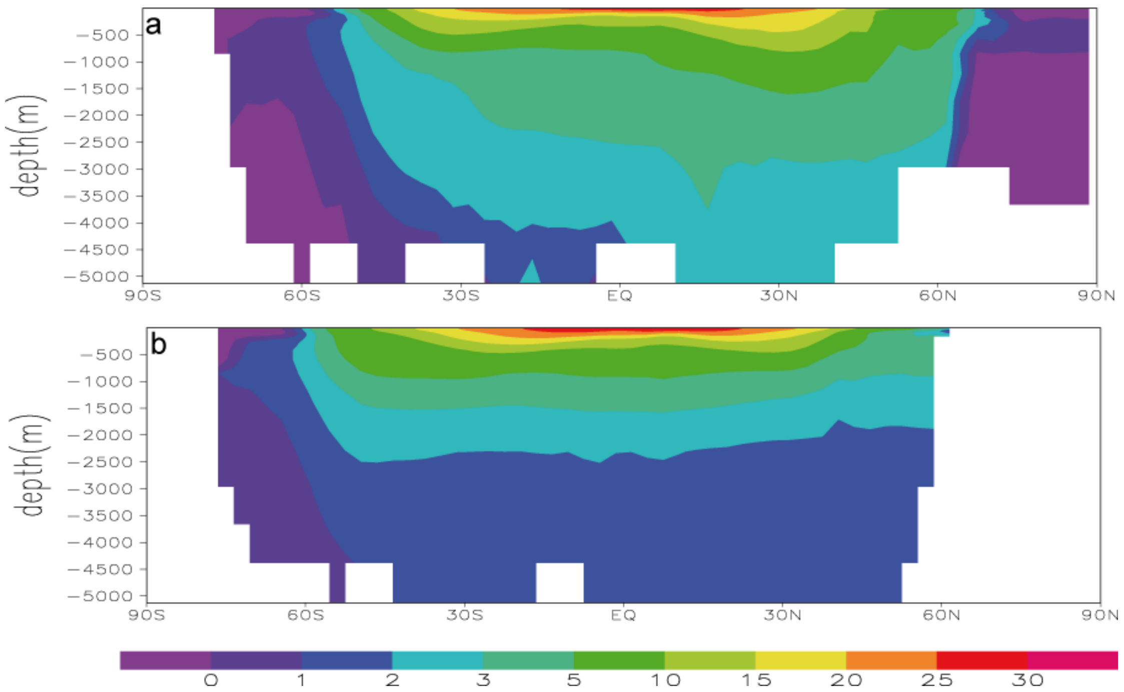

Fig. 1.13: Temperature (°C) averaged over all latitudes (i.e. zonal mean) in (a) the Atlantic and in (b) the Pacific. Data source: from Levitus (1998).

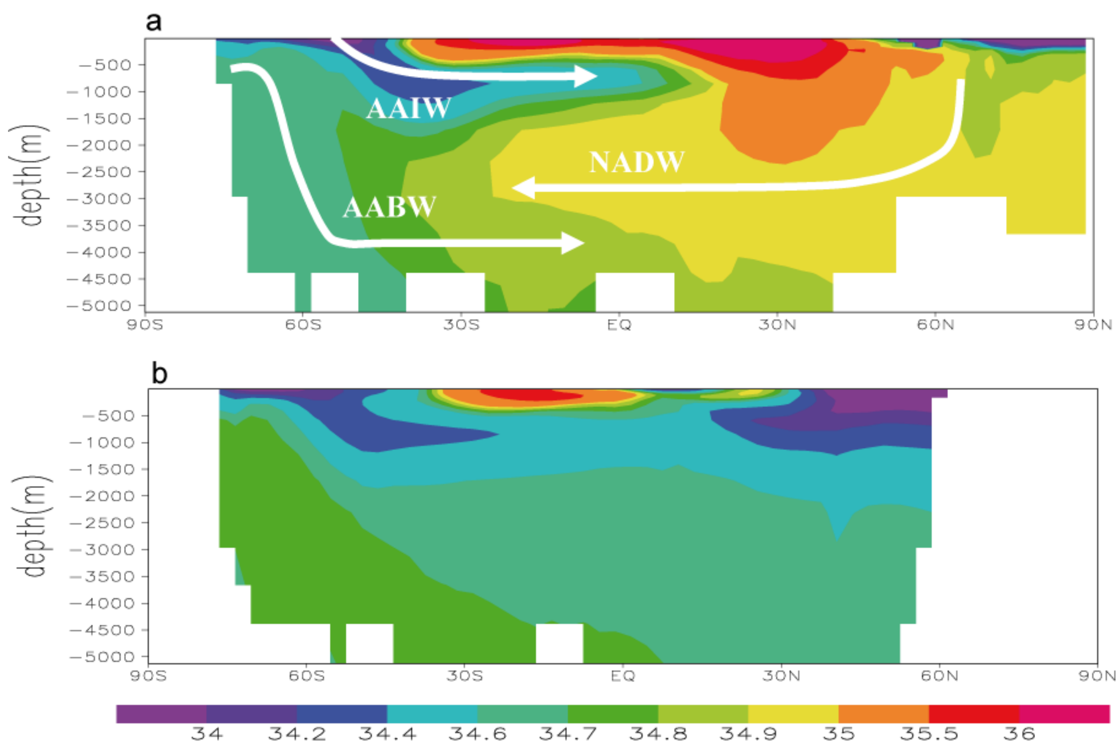

Figure 1.14: Salinity (psu) averaged over all latitudes (i.e. **zonal** mean) in (a) the Atlantic and in (b) the Pacific. The schematic paths of three important water masses are shown for the Atlantic. Data source: from Levitus (1998).

The temperature and salinity of sea water are modified by interactions with the atmosphere only in the oceanic mixed layer. The mixed layer is also the area where the greatest mixing occurs, diffusion being weaker in the ocean interior. **Water mass formation** and transformation thus mainly occur close to the surface. When these waters flow beneath the mixed layer, they tend to keep the properties they have acquired close to the surface. This is particularly clear in the deep ocean. As a consequence, the path of important water masses, like NADW and AABW, can easily be followed from their region of formation on temperature and salinity vertical sections (Fig.1.13 and 1.14). The influence of Antarctic Intermediate Water (AAIW), originating from the Southern Ocean, is also clearly identified as a low salinity tongue reaching the equator at intermediate depth. More generally, in the **thermocline**, water can originate from a nearby location (generally poleward), where surface density in winter is high enough to allow water to sink to intermediate depths.

1.4 The cryosphere

1.4.1 Components of the cryosphere

The cryosphere is the portion of the Earth's surface where water is in solid form. It thus includes sea ice, lake ice and river-ice, snow cover, glaciers, ice caps and ice sheets, and frozen ground. The snow cover has the largest extent, with a maximum area of more than 45 10^6 km^2 (Table 1.1). Because of the present distribution of continents, land surfaces at high latitudes are much larger in the Northern Hemisphere than in the Southern Hemisphere. As a consequence, the large majority of the snow cover is located in the Northern Hemisphere (Figs. 1.15 and 1.16). The same is true for the freshwater ice that forms on rivers and lakes in winter. Both the snow cover and freshwater ice have a very strong seasonal cycle, as they nearly disappear in summer in both hemispheres (Table 1.1).

Component	Maximum area (10^6 km^2)	Minimum area (10^6 km^2)	Maximum Ice volume (10^6 km^3)	Minimum Ice volume (10^6 km^3)
Northern Hemisphere Snow cover	46.5 (late January)	3.9 (late August)	0.002	
Southern Hemisphere Snow cover	0.83 (late July)	0.07 (early May)		
Sea ice in the Northern Hemisphere	14.0 (late March)	6.0 (early September)	0.05	0.02
Sea ice in the Southern Hemisphere	15.0 (late September)	2.0 (late February)	0.02	0.002

Table 1.1: Areal extent and volume of snow cover and sea ice. Data compiled in Climate and Cryosphere (CliC) project science and co-ordination plan (2001).

Sea ice, which is a moving medium formed when sea water freezes, generally does not cover the whole oceanic surface in a region. Relatively narrow elongated areas of open water inside the pack are called **leads**, while larger areas of open water are called **polynyas**. The sea-ice concentration is defined as the fraction of the surface of interest (pixel from a satellite image, area surrounding a boat, etc) that is effectively covered by sea ice. A concentration of ice of 1 (or 100%) thus corresponds to a continuous ice pack, while a value of 0 corresponds to open ocean.

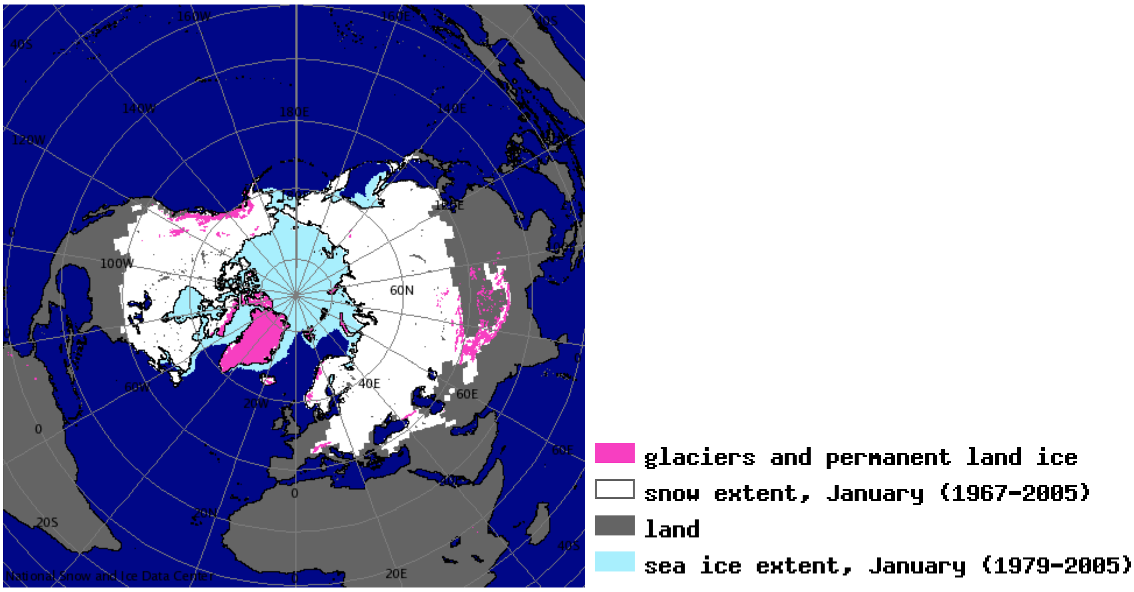

Figure 1.15: The distribution of sea ice, snow and land ice in January in the Northern Hemisphere. Source: Atlas of the Cryosphere, National Snow and Ice Data Center (NSIDC), http://nsidc.org/data/atlas/ .

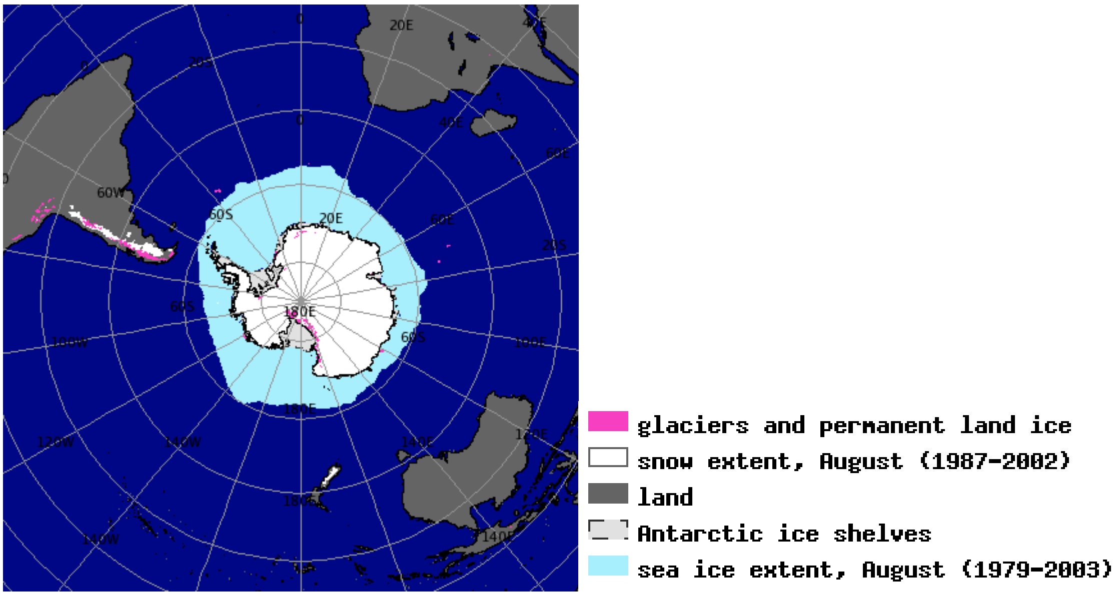

Figure 1.16: Location of sea ice, snow and land ice in August in the Southern Hemisphere. Source; Atlas of the Cryosphere, National Snow and Ice Data Center (NSIDC), http://nsidc.org/data/atlas/ .

Sea ice covers a similar area in both hemispheres (Table 1.1). Its seasonal cycle is larger in the Southern Ocean (Fig. 1.17) where the majority of the ice cover is first-year sea ice (i.e. sea ice that has not survived one summer). Because of the large thermal inertia of the ocean (see section 2.1.5), the minimum and maximum sea ice extent are shifted by about two months compared to the snow cover on land, with maximum/minimum values around March and September in both hemispheres (Fig. 1.17). The sea ice is thinner in the Southern Hemisphere, with a mean thickness of less than 1 m, while the mean ice thickness in the central Arctic is around 3m.

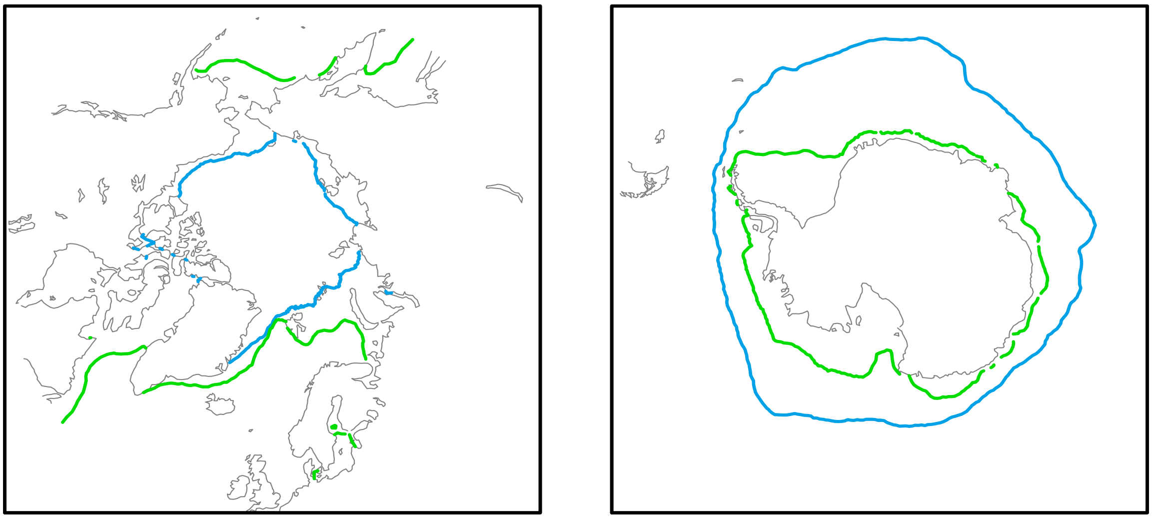

Figure 1.17: Location of the ice edge in March (green) and September (blue) in both hemispheres. The ice edge is commonly defined as the line where the ice concentration is 15%. Data from Rayner et al. (2003).

Component	Area (10^6 km^2)	Ice volume (10^6 km^3)	Sea level equivalent (m)
Continuous permafrost	10.69	0.0097-0.0250	0.024-0.063
Discontinuous permafrost	12.10	0.0017-0.0115	0.004-0.028
East Antarctica	10.1	22.7	56.8
West Antarctica and Antarctic Peninsula	2.3	3.0	7.5
Greenland	1.8	2.6	6.6
Small ice caps and mountain glaciers	0.68	0.18	0.5
Ice shelves	1.5	0.66	-

Table 1.2: Areal extent and volume of permafrost and land ice. Data compiled in the Climate and Cryosphere (Clic) project science and co-ordination plan (2001). *N.B.* The sea-level equivalent is computed as the thickness of a water layer corresponding to the ice mass distributed over the whole World Ocean. This is not directly equal to the resulting sea-level rise as parts of the Antarctic and Greenland ice sheets are presently below sea level.

Like snow, the seasonally frozen ground covers a large fraction of the continents in the Northern Hemisphere. Where the annual mean temperature is below -1°C, the ground can be perennially frozen below an active layer which melts in summer. This is the **permafrost**, which is estimated to cover more than 20% of the land area in the Northern Hemisphere (Table 1.2). The thickness of the frozen layer can exceed 600 m at high latitudes. Further south this layer thins, and the permafrost becomes discontinuous close to its margins (Fig. 1.18).

A large majority of the ice present on Earth today is located in two big ice-sheets: the Greenland and Antarctic ice sheets. The Antarctic ice sheet is itself commonly divided into two parts, East Antarctica and West Antarctica, roughly corresponding to the eastern and western hemispheres relative to the Greenwich meridian. The thickness of ice on these ice sheets can reach several kilometres (Fig 1.19 and 1.20). Ice sheets are formed by the accumulation of snow layers over tens of thousand years. As snow falls at the surface, the pressure on the older snow layers increases, transforming them into ice. Ice sheets (like glaciers) are not stagnant and generally flow slowly towards their margins. However, in some regions (called ice streams), the flow is much faster than in other parts of the ice sheet, sometimes reaching several kilometres per year.

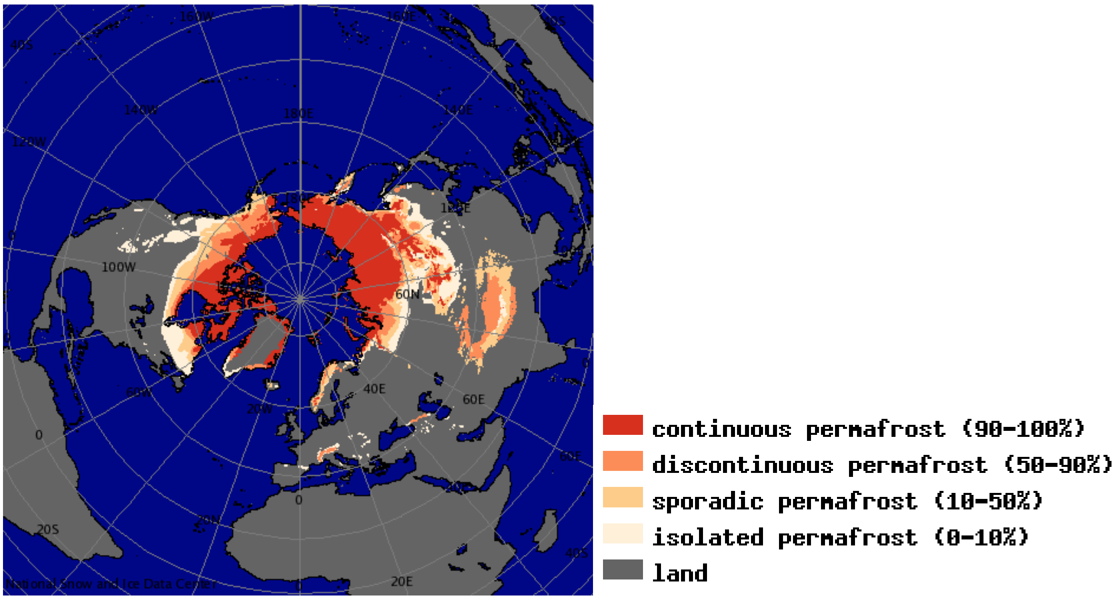

Figure 1.18: Location of permafrost in the Northern Hemisphere. Source: Atlas of the Cryosphere, National Snow and Ice Data Center (NSIDC), http://nsidc.org/data/atlas/ .

Because of the weight of the ice, the bedrock is depressed and, in some areas, is well below the sea level. For instance, most of the East Antarctic ice sheet is a high ice plateau that rests on bedrock, but large areas of the West Antarctic ice sheet are grounded below sea level. The total volume of the West Antarctic ice sheet that is below sea level has been estimated at around 1.9 10^6 km^3.

Antarctica is surrounded by **ice shelves**. These are floating platforms made of ice originating from the continent which has flowed down the coastline into the ocean. The two largest ice shelves are the Ross and Filchner shelves, which together cover more than 800 000 km^2. Ice shelves and glaciers which reach the shore are able to release **icebergs** that can drift over long distances, pushed by the ocean currents and winds. **Icebergs** are thus found in the open ocean, but they should not be confused with sea ice. They are usually much thicker (sometimes more than 100 m) and consist of freshwater, while sea ice is salty and is formed directly from sea water

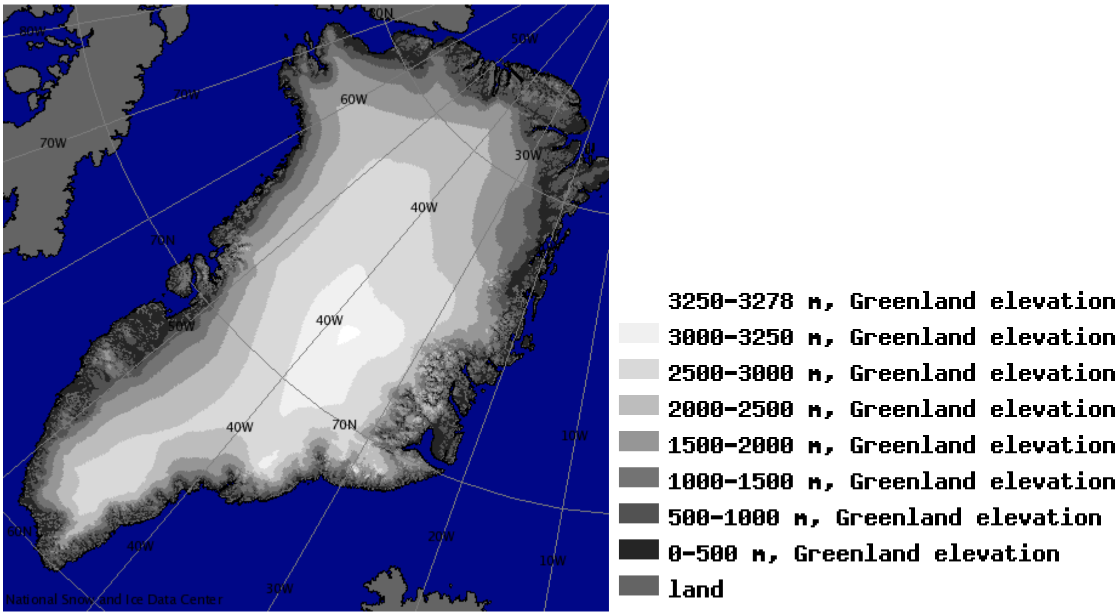

Figure 1.19: Greenland surface elevation. To obtain the ice thickness, the bedrock elevation has to be subtracted from this figure. Source: Atlas of the Cryosphere, National Snow and Ice Data Center (NSIDC), http://nsidc.org/data/atlas/ .

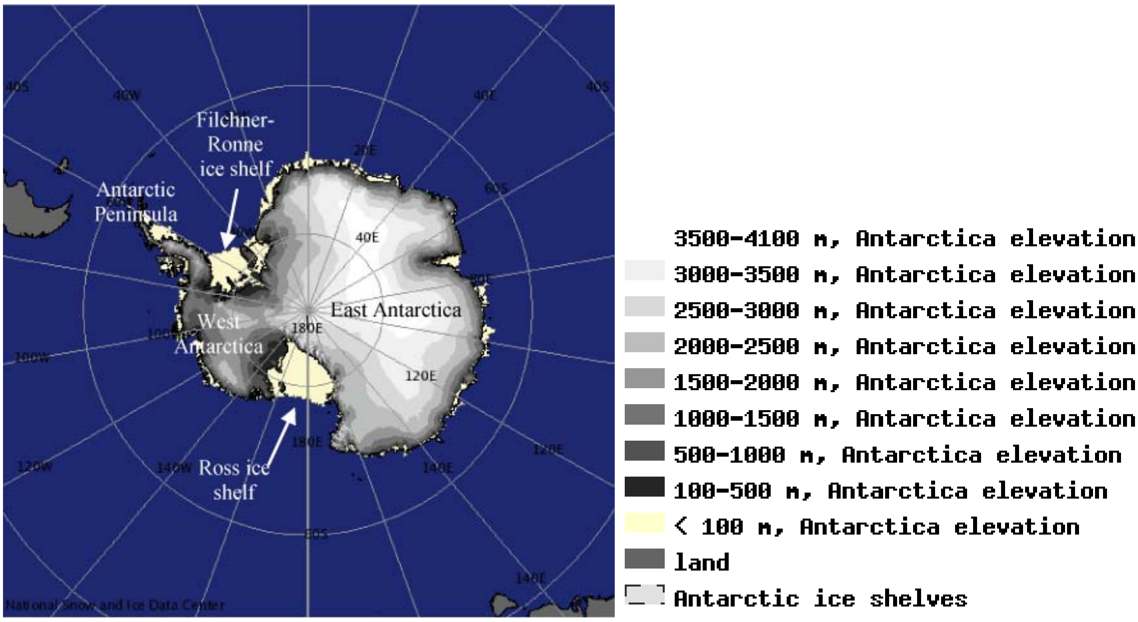

Figure 1.20: Antarctica surface elevation. To obtain the ice thickness, the bedrock elevation has to be subtracted from this figure. Source: Atlas of the Cryosphere, National Snow and Ice Data Center (NSIDC), http://nsidc.org/data/atlas/.

1.4.2 Properties of the cryosphere

Snow and ice have a very large **albedo**, i.e. they reflect the majority of the incoming solar radiation. They thus play a major role in the global heat balance of the Earth (see section 2.1). By storing and releasing **latent heat**, they affect the seasonal cycle of the surface temperature (see section 2.1). They are also good insulators that reduce the heat loss from the underlying surface (land or ocean) towards the cold atmosphere in winter. More generally, the presence of sea ice restricts the exchanges of heat and gases between the ocean and the atmosphere. When sea ice forms, only a fraction of the salt present in the ocean is trapped in the ice, the remainder being ejected towards the ocean (this is called brine rejection). The resulting sea ice salinity is between 10 psu in relatively young ice and less than 2 psu in very old ice (compared to around 35 psu for the ocean, see section 1.3). Because of this brine rejection, sea ice formation increases the salinity at the ocean surface while, melting sea ice is associated with surface freshening. Sea ice drift is also associated with a horizontal freshwater transport. If there is a net convergence of the sea-ice transport and intense ice melting in a region, this will decrease the salinity of surface water there. On the other hand, in regions such as coastal **polynyas**, the strong winds in winter continually push the newly formed ice off shore, leading to a strong divergence of the sea-ice transport. This implies very high ice formation rates in these **polynyas** (up to 10 m per year at some locations) and thus large amounts of brine rejection which can lead to very high ocean salinities in those regions.

Ice sheets store large amounts of water on land. Any change in their volume thus has a considerable effect on the sea level. It is estimated that, if all the ice sheets melted completely, taking into account the fact that some ice sheets are grounded below sea level, the sea level would rise by more than 60 m. On the other hand, if we neglect the effect of dilution on sea water density and volume, the melting of sea ice and **ice shelves** does not influence sea levels. Indeed, because of Archimedes' law, floating ice displaces its own weight of sea water and the melt water thus simply replaces the volume of ice previously below sea level. Ice sheets are also big mountains that, because of their height, help to maintain cold conditions on the surface. The presence of cold air on the ice sheet also has a regional influence, cooling the surrounding areas.

1.5 The land surface and the terrestrial biosphere

As discussed above, many characteristics of the climate are influenced by the distribution and topography of land surface. For instance, mountain chains such as the Andes or the Rocky mountains (Fig. 1.21) are formidable barriers to the westerly winds that influence the climate on a continental scale. Mountains also have an important role at the hemispheric scale, by affecting planetary waves and the global atmospheric circulation (section 1.2). The distance to the coast influences the temperature and aridity of a region. The presence of land boundaries to the ocean (and more generally the ocean bathymetry) affects the location of the strong western boundary currents and of the straits that allow water exchanges between the different basins (section 1.3). The shape and even the existence of an ice sheet is strongly conditioned by the underlying bedrock (section 1.4).

In addition to the influence of the land geometry, the type of vegetation present on land also has a critical influence on climate at all spatial and temporal scales. One of the most important roles of terrestrial vegetation is related to its **albedo** (Fig.1.22), (see section 2.1). Vegetation usually has a lower albedo than soil (Table 1.3), in particular much smaller than that of deserts. This is why subtropical deserts such as the Sahara appear as regions of particularly high albedo on global maps (Fig. 1.22). A maximum is also observed at high latitudes because of the presence of snow and ice. At these latitudes, the vegetation modulates the influence of the snow. In the absence of vegetation or in the presence of low-growing vegetation such as grass, the snow can cover the whole area, leading to highly reflective white areas with a high **albedo**. If snow falls on a forest, relatively dark trunks, branches and possibly needles or leaves will partially emerge from the snow, resulting in a much lower **albedo** than with an homogenous snow blanket.

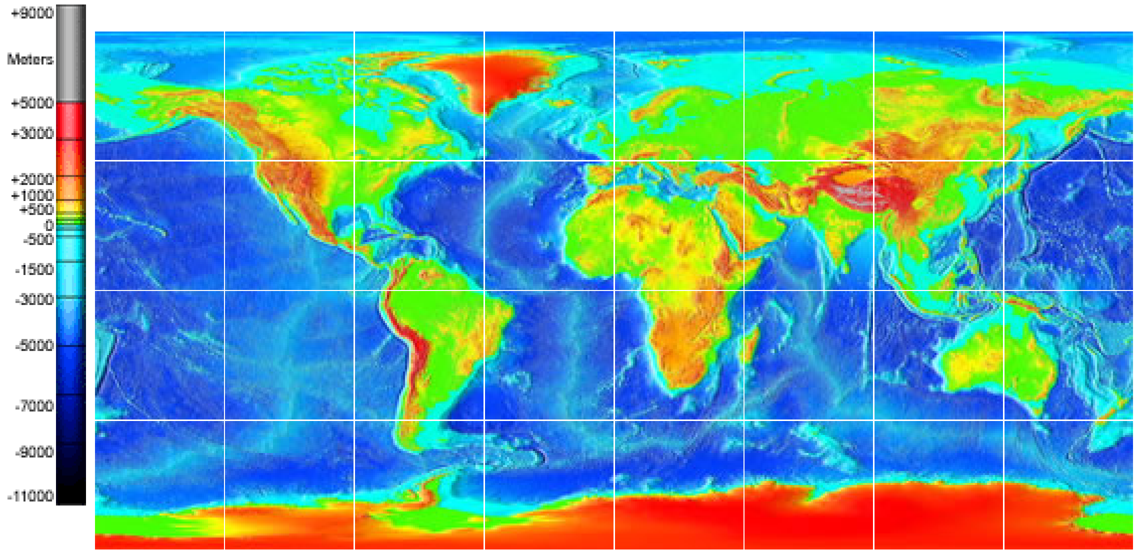

Figure 1.21: High resolution map of the surface topography. *Source* : Etopo2v2, http://www.ngdc.noaa.gov/mgg/image/2minsurface. Following the policy of U.S. governments agencies, this figure is not subject to copyright protection.

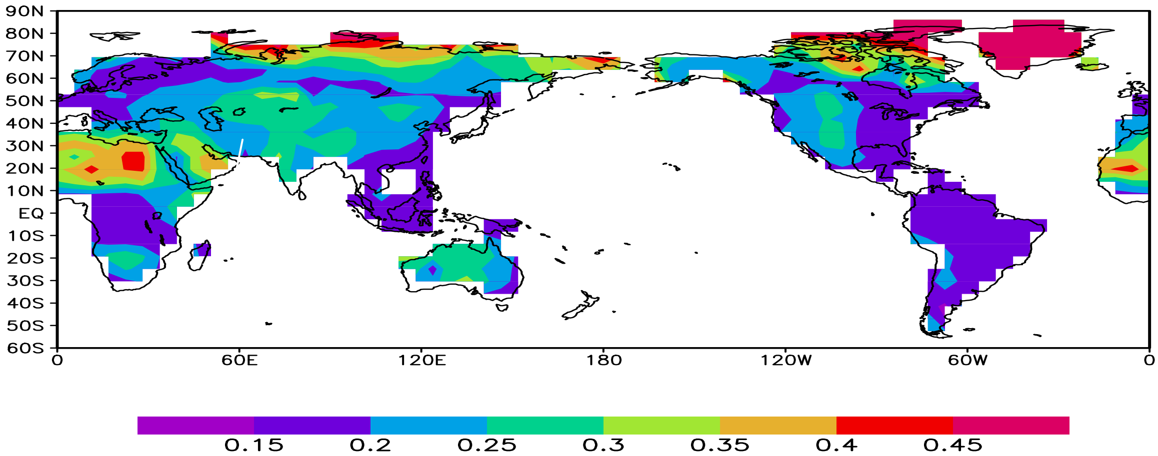

Figure 1.22: Surface **albedo**. Average of visible and near infra-red albedo. Data from Global Soil Wetness Project (GSWP2). Data source http://www.monsoondata.org:9090/dods

Surface type	**Albedo**
Ocean	0.05-0.15
Fresh snow	0.75-0.90
Old snow	0.40-0.70
Sea ice	0.3-0.6
Soil	0.05-0.40
Desert	0.20-0.45
Cropland	0.18-0.25
Grassland	0.16-0.26
Deciduous forest	0.15-0.20
Coniferous forest	0.05-0.15
Snow covered coniferous forest	0.13-0.3

Table 1.3: Typical range of the **albedo** of various surfaces.

The terrestrial **biosphere** also has a clear impact on the hydrological cycle (see section 2.2). Water storage is generally greater in soil covered by vegetation than on bare land where direct runoff often follows precipitation. Stored water can later be taken up by plant roots and transferred back to the atmosphere by **evapotranspiration**. A third effect of the vegetation cover is related to the surface roughness that influences the stress at the atmosphere-land interface and the turbulent exchanges at the surface (see section 2.1). Finally, the role of the terrestrial biosphere in the global carbon cycle will be discussed in section 2.3.

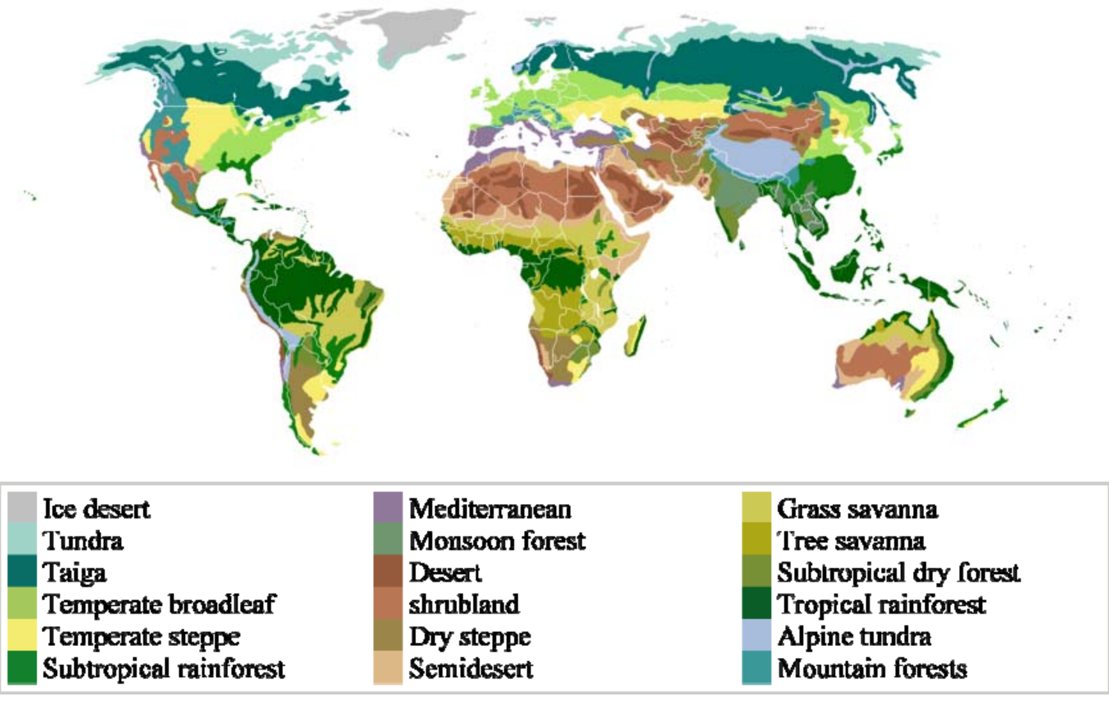

Figure 1.23: Terrestrial biomes. *Source*: http://en.wikipedia.org/wiki/Biome#Map_of_Biomes.

Because of this climatic role of vegetation, it is useful to describe the general distribution of the different **biomes**, which are regions with distinctive large-scale vegetation types (Fig. 1.23). Their exact definition, as well as the number of important biomes that are considered, differ from one study to the other. Nevertheless, it is generally considered that the natural biomes can be classified, according to their typical percentage of grass and trees, into five groups: desert, grassland, shrubland, woodland and forest. Cropland and built-up areas can be added to take into account the role of land use associated with human activities.

Deserts are characterised by a very small amount of vegetation. Grassland, as indicated by its name, is mainly covered by grass and lichens. It can be found at various latitudes and includes **tundra**, steppe and savannah. In shrubland, low woody plants are present in addition to grass. The fraction of trees is higher in woodland, but there are still significant areas covered by grass and often relatively large distances between trees. Finally, in forests, a dense cover of trees is observed (as in tropical rainforest and boreal conifer forest, also called **taiga**).

We have discussed above how vegetation influences **climate**, but of course climate also influences vegetation. This leads to powerful **feedback** loops that will be described in more detail in section 4.3.3. The dominant features of the climate are achieved through the distribution of incoming solar radiation, temperature and precipitation. If precipitation and/or temperature are too low, desert **biomes** dominate (as in the Sahara or Antarctica). At higher temperatures, forests can be maintained if a sufficient supply of water by rainfall is available. Between those two extremes, different combinations of grass and trees are found (see also Fig. 3.9).

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Exercises

Exercises are available on the textbook website (http://www.climate.be/textbook) and on iCampus for registered students.