

Introduction to the Climate System

1.1 ATMOSPHERE, OCEAN, AND LAND SURFACE

Climate is the synthesis of the weather in a particular region. It can be defined quantitatively by using the expected values of the meteorological elements at a location during a certain month or season. The expected values of the meteorological elements can be called the climatic elements and include variables such as the average temperature, precipitation, wind, pressure, cloudiness, and humidity. In defining the climate, we usually employ the values of these elements at the surface of Earth. Thus, one can characterize the climate of Seattle by stating that the average annual mean precipitation is 38 in. and the annual mean temperature is 52°F. However, one might need a great deal more information than the annual means. For example, a farmer would also like to know how the precipitation is distributed through the year and how much rain would fall during the critical summer months. A hydroelectric plant engineer needs to know how much interannual variability in rainfall and snow accumulation to expect. A homebuilder should know how much insulation to be installed and the size of the heating or cooling unit needed to provide for the weather in the region. Sailors might like to know that the wind blows in the winter, but not so much in the summer.

The importance of climate is so basic that we sometimes overlook it. If the climate were not more or less as it is, life and civilization on this planet would not have developed as they have. The distribution of vegetation and soil type over the land areas is determined primarily by the local climate. Climate affects human lives in many ways; for example, climate influences the type of clothing and housing that people have developed. In the modern world, with the great technological advances of the past century, one might think that climate no longer constitutes a force capable of changing the course of human history. It is apparent, on the contrary,

that we are as sensitive now as we have ever been to climate fluctuations and climate change.

Because food, water, and energy supply systems are strained to meet demand and are optimized to the current average climatic conditions, fluctuations or trends in climate can cause serious difficulties for humanity. Moreover, since the population has grown to absorb the maximum agricultural productivity in much of the world, the absolute number of human lives at risk of starvation during climatic anomalies has never been greater. In addition to natural year-to-year fluctuations in the weather, which are an important aspect of climate, we must be concerned with the effects of human activities in producing long-term trends in the climate. It is now clear that humans are affecting the global climate, and this influence is growing. The actions of humankind that can change the global climate include altering the composition of the atmosphere and the nature of Earth's surface.

The surface climate of Earth varies greatly with location, ranging from the heat of the tropics to the cold of the polar regions, and from the drought of a desert to the moisture of a rain forest. Nonetheless, the climate of Earth is favorable for life, and living creatures exist in every climatic extreme. The climate of a region depends on latitude, altitude, and orientation in relation to water bodies, mountains, and the prevailing wind direction. In this book, we are concerned primarily with the global climate and its geographic variation on scales of hundreds to thousands of kilometers. In order to focus on these global issues, climate variations on horizontal spatial scales smaller than several tens of kilometers are given only minimal discussion.

The climate of Earth is defined in terms of measurable weather elements. The weather elements of most interest are temperature and precipitation. These two factors together largely determine the species of plants and animals that survive and prosper in a particular location. Other variables are also important, of course. The *humidity*, the amount of water vapor in the air, is a critical climate factor that is related closely to the temperature and precipitation. Condensation of water in the atmosphere produces clouds of water droplets or ice particles that greatly change the radiative properties of the atmosphere. The occurrence of clouds is important in itself for aviation and other activities, but clouds also play a role in determining both precipitation and surface temperature. Cloudiness influences the transmission of terrestrial radiation through the atmosphere and the amount of solar radiation that reaches the surface. The mean wind speed and direction are important considerations for local climate, air-pollution dispersion, aviation, navigation, wind energy, and many other purposes. The climate system of Earth determines the distribution of energy and water near the surface and consists primarily of the atmosphere, the oceans, and the land surface. The workings of this global system are the topic of this book (Fig. 1.1).

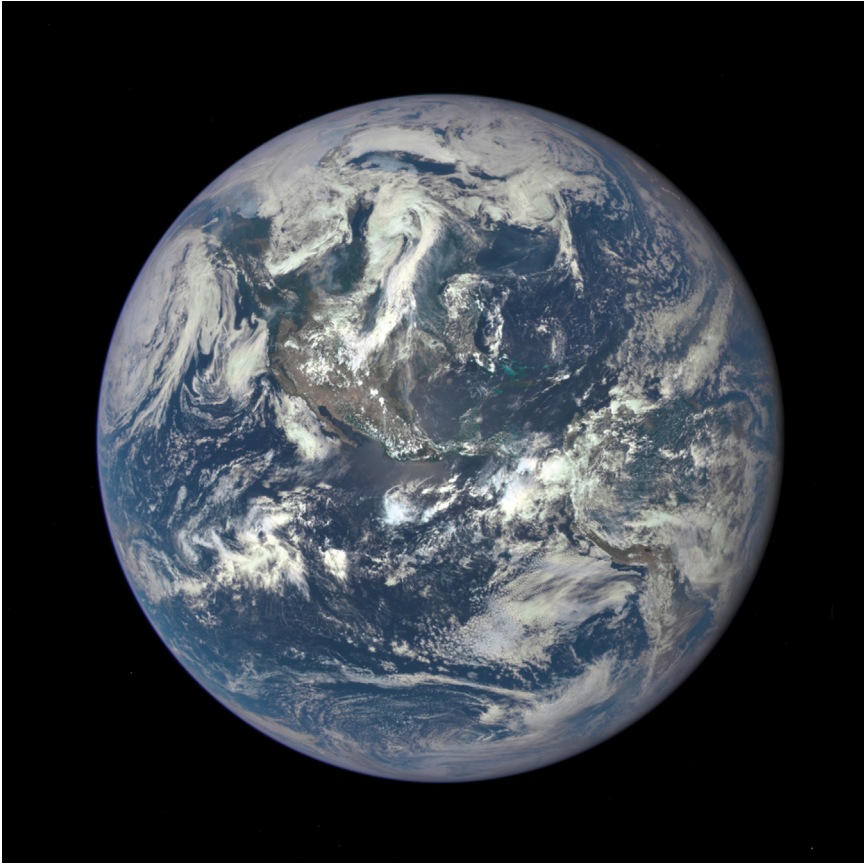


FIGURE 1.1 Earth as seen on July 6, 2015 by the NASA Earth Polychromatic Imaging Camera aboard the NOAA Deep Space Climate Observatory spacecraft one million miles from Earth.

1.2 ATMOSPHERIC TEMPERATURE

Temperature is the most widely recognized climatic variable. The global average temperature at the surface of Earth is about 288 K, 15°C, or 59°F. The range of temperatures encountered at the surface is favorable for the life forms that have developed on Earth. The extremes of recorded surface temperature range from the coldest temperature of -89.2°C (-128.6°F) at Vostok, Antarctica to the warmest temperature of 56.7°C (134°F) at Furnace Creek Ranch in Death Valley, California. These temperature extremes reflect the well-known decrease of temperature from the tropics, where the warmest temperatures occur, to the polar regions that are much colder. Both the warm temperature in Death Valley and the cold temperature at Vostok also

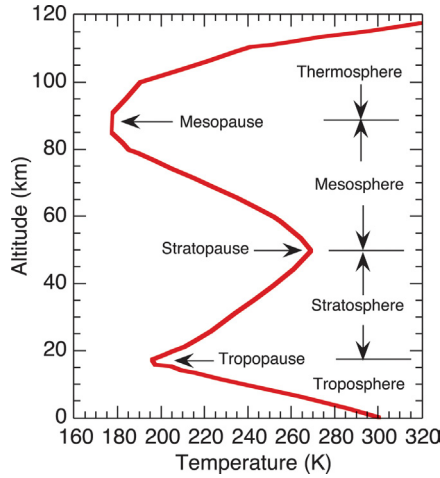


FIGURE 1.2 The main zones of the atmosphere defined according to the temperature profile of the standard atmosphere profile at 15°N for annual-mean conditions. Data from U.S. Standard Atmosphere Supplements (1966).

result partly from the decrease of temperature with altitude because Death Valley is below sea level and Vostok is 3450 m above sea level.

An important feature of the temperature distribution is the decline of temperature with height above the surface in the lowest 10–15 km of the atmosphere (Fig. 1.2). This rate of decline, called the *lapse rate*, is defined by

$$\Gamma \equiv -\frac{\partial T}{\partial z} \quad (1.1)$$

where T is the temperature and z is altitude and the deltas indicate a partial derivative. The global mean tropospheric lapse rate is about 6.5 K km^{-1} , but the lapse rate varies with altitude, season, and latitude. In the upper *stratosphere*, the temperature increases with height up to about 50 km. The increase of temperature with height that characterizes the stratosphere is caused by the absorption of solar radiation by ozone. Above the stratopause at about 50 km the temperature begins to decrease with height in the *mesosphere*. The temperature of the atmosphere increases rapidly above about 100 km because of heating produced by absorption of ultraviolet radiation from the sun, which dissociates oxygen and nitrogen molecules and ionizes atmospheric gases in the *thermosphere*.

The decrease of temperature with altitude in the *troposphere* is crucial to many of the mechanisms whereby the warmth of the surface temperature of Earth is maintained. The lapse rate in the troposphere and the mechanisms that maintain it are also central to the determination of

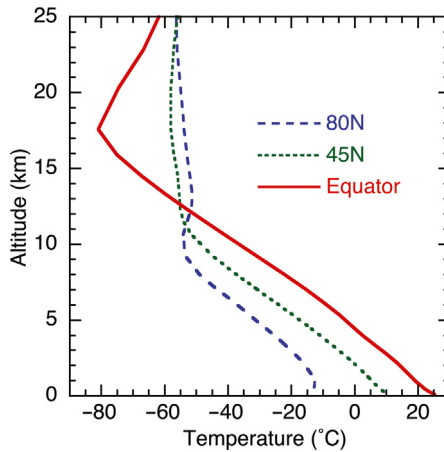


FIGURE 1.3 Annual mean temperature profiles for the lowest 25 km of the atmosphere in three latitude bands. Data from ERA-Interim.

climate sensitivity, as discussed in Chapter 10. The lapse rate and temperature in the troposphere are determined primarily by a balance between radiative cooling and convection of heat from the surface. The vertical distribution of temperature varies with latitude and season. At the equator, the temperature decreases with altitude up to about 17 km (Fig. 1.3). The tropical *tropopause* is the coldest part of the lowest 20 km of the atmosphere in the annual mean. In middle and high latitudes, the temperature of the lower stratosphere is almost independent of height. The tropospheric lapse rate in polar latitudes is less than it is nearer the equator. At high latitudes, the temperature actually increases with altitude in the lower troposphere in the winter and spring (Fig. 1.4). A region of negative lapse rate is called a *temperature inversion*. The polar temperature inversion has important implications for the climate of the polar regions. It arises because the surface cools very efficiently through emission of infrared radiation in the absence of insolation during the winter darkness. The air does not emit radiation as efficiently as the surface, and heat transported poleward in the atmosphere keeps the air in the lower troposphere warmer than the surface.

The variation of the zonal mean temperature with latitude and altitude is shown in Fig. 1.5. In Southern Hemisphere winter (June, July, and August, JJA), the polar stratosphere is colder than 180 K, and is the coldest place in the atmosphere, even colder than the tropical tropopause. The Northern Hemisphere stratosphere does not get as cold, on average, because planetary Rossby waves generated by surface topography and east–west surface temperature variations transport heat to the pole during *sudden stratospheric warming* events.

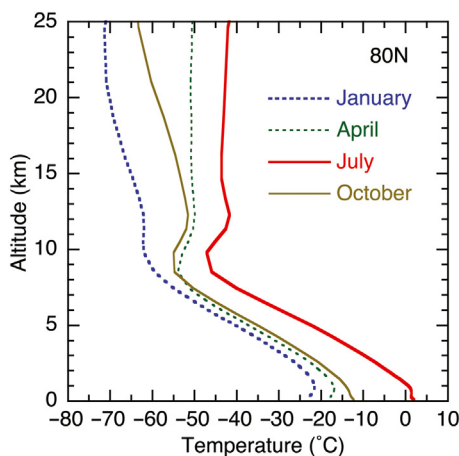


FIGURE 1.4 Season variation profiles at 80°N. Data from ERA-Interim.

The geographic and seasonal variation of surface temperature is shown in Fig. 1.6. The surface temperature is greatest near the equator, where it exceeds 296 K (23°C) across a broad band of latitudes in all seasons. Outside this belt, surface temperature decreases steadily toward both poles. The interiors of the northern continents become very cold during winter, but they are warmer than ocean areas at the same latitude during summer. Seasonal variations of surface temperature in the interiors of North America and Asia are very large (Fig. 1.6). Seasonal variation in the Southern Hemisphere is much smaller because of the greater fraction of the surface covered by ocean. The smaller seasonal variation of air temperature in mid-latitudes of the Southern Hemisphere is associated with the larger fraction of ocean-covered surface there. The ocean stores heat very effectively. During the summer season it stores the heat provided by the sun. Because a large amount of heat is required to raise the surface temperature of the oceans, the summer insolation raises the surface temperature by only a small amount. During winter, a large amount of heat is released to the atmosphere with a relatively small change in sea surface temperature. Land areas heat up and cool down much more quickly than oceans (see Chapter 4).

1.3 ATMOSPHERIC COMPOSITION

The composition of the atmosphere is a key determinant of Earth's climate. The interaction of atmospheric gases with radiant energy modulates the flow of energy through the climate system. The atmosphere has a mass of about 5.14×10^{18} kg, which is small compared to the mass of the ocean, 1.39×10^{21} kg, and the solid Earth, 5.98×10^{24} kg. Dry atmospheric air

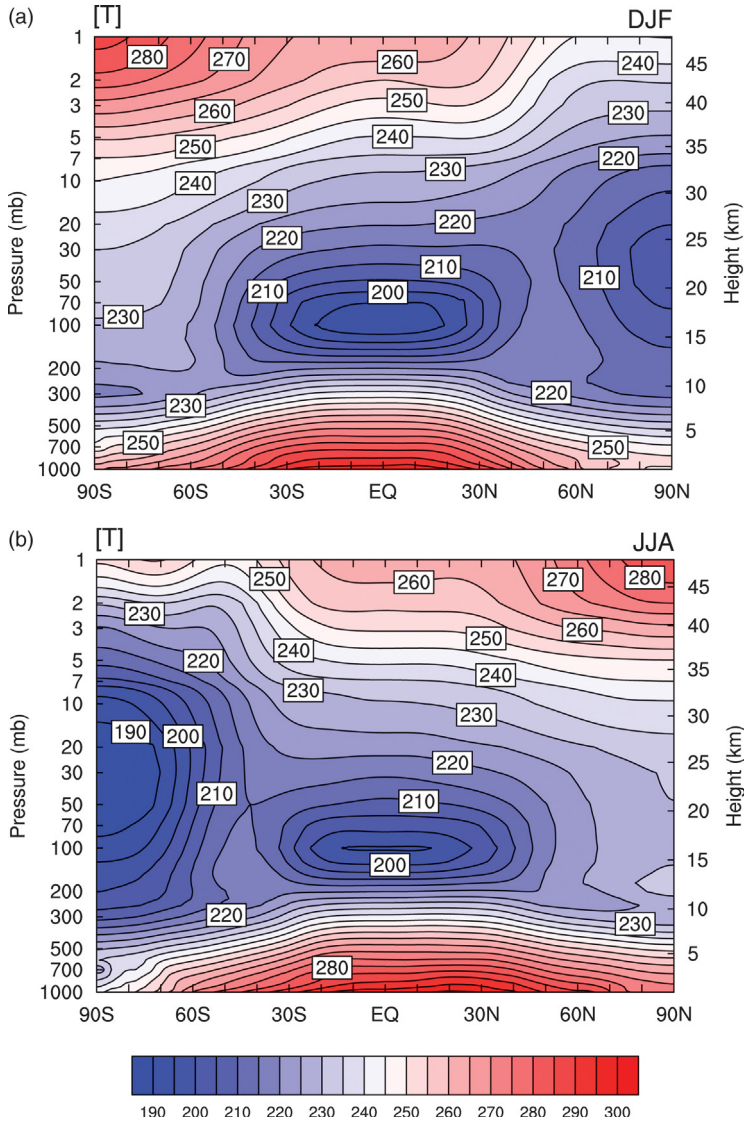


FIGURE 1.5 Zonal average temperature (K) as a function of latitude and altitude for the (a) December, January, February (DJF) and (b) June, July, August (JJA) Seasons. Data from ERA 40 reanalysis.

is composed mostly of molecular nitrogen (78%) and molecular oxygen (21%). The next most abundant gas in the atmosphere is argon (1%), an inert noble gas. The atmospheric gases that are important for the absorption and emission of radiant energy comprise less than 1% of the atmosphere's mass. These include water vapor (3.3×10^{-3} of the atmosphere's total mass), carbon dioxide (5.3×10^{-7}), and ozone (6.42×10^{-7}), in order of

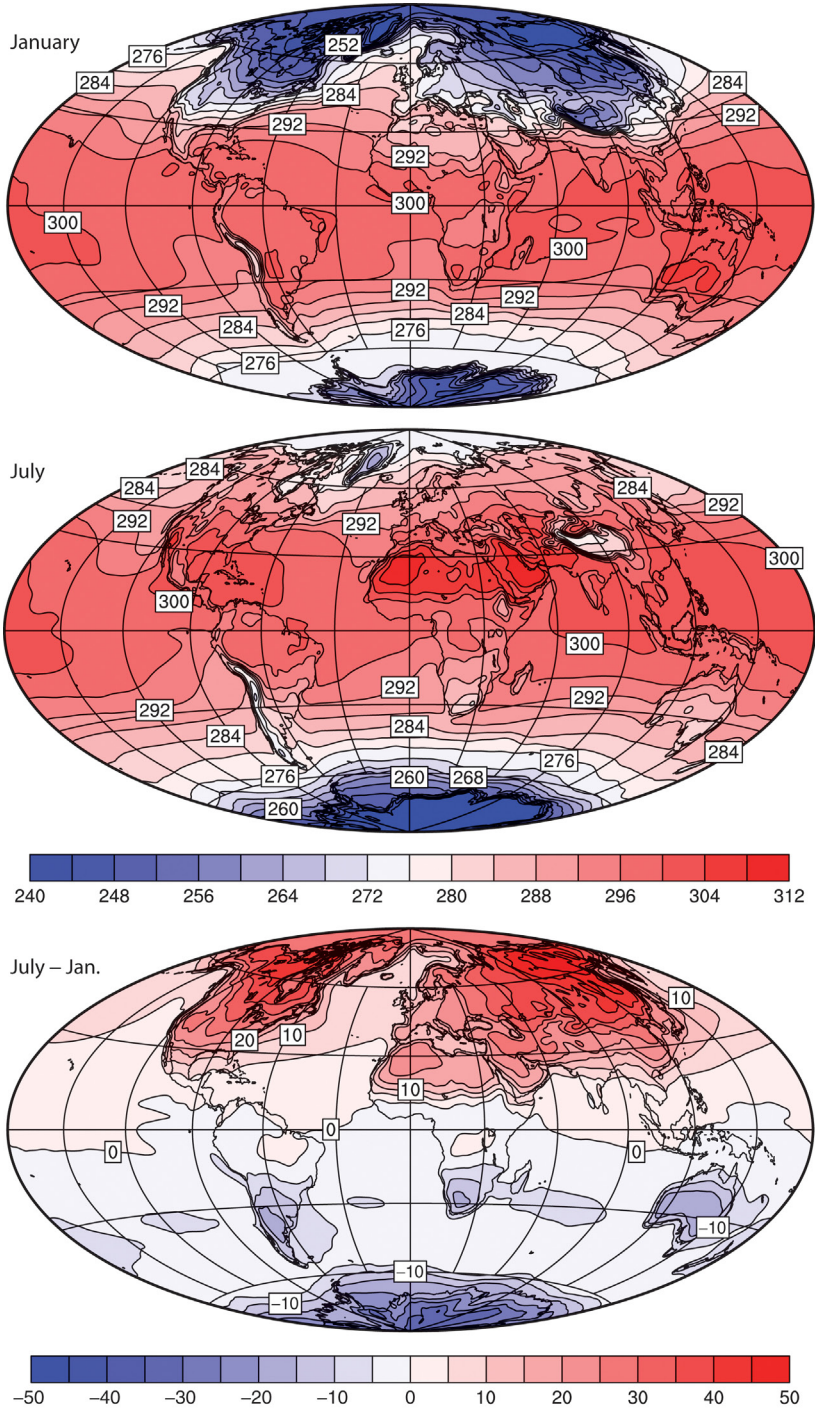


FIGURE 1.6 Global map of the January and July surface temperature and July minus January. Data from ERA-Interim reanalysis.

TABLE 1.1 Composition of the Atmosphere

Constituent	Chemical formula	Molecular weight (¹² C = 12)	Fraction by volume in dry air	Total mass (g)
Total atmosphere		28.97		5.136×10^{21}
Dry air		28.96	100.0 %	5.119×10^{21}
Nitrogen	N ₂	28.01	78.08 %	3.87×10^{21}
Oxygen	O ₂	31.99	20.95 %	1.185×10^{21}
Argon	Ar	39.95	0.934 %	6.59×10^{19}
Water vapor	H ₂ O	18.02	Variable	1.7×10^{19}
Carbon dioxide	CO ₂	44.0	391 ppmv*	$\sim 2.76 \times 10^{18}$
Neon	Ne	20.18	18.18 ppmv	6.48×10^{16}
Krypton	Kr	83.8	1.14 ppmv	1.69×10^{16}
Helium	He	4.00	5.24 ppmv	3.71×10^{15}
Methane	CH ₄	16.04	1.8 ppmv*	$\sim 4.9 \times 10^{15}$
Xenon	Xe	131.3	87 ppbv	2.02×10^{15}
Ozone	O ₃	47.99	Variable	$\sim 3.3 \times 10^{15}$
Nitrous oxide	N ₂ O	44.01	324 ppbv*	$\sim 2.3 \times 10^{15}$
Carbon monoxide	CO	28.0	120 ppbv	$\sim 5.9 \times 10^{15}$
Hydrogen	H ₂	2.02	500 ppbv	$\sim 1.8 \times 10^{14}$
Ammonia	NH ₃	17.0	100 ppbv	$\sim 3.0 \times 10^{13}$
Nitrogen dioxide	NO ₂	46.0	1 ppbv	$\sim 8.1 \times 10^{12}$
Sulfur dioxide	SO ₂	64.1	200 pptv	$\sim 2.3 \times 10^{12}$
Hydrogen sulfide	H ₂ S	34.1	200 pptv	$\sim 1.2 \times 10^{12}$
CFC-12	CCl ₂ F ₂	120.9	528 pptv*	$\sim 1.0 \times 10^{13}$
CFC-11	CCl ₃ F	137.4	238 pptv*	$\sim 6.8 \times 10^{12}$

* Values of trace constituents valid in 2011 (ppmv = 10^{-6} , ppbv = 10^{-9} , pptv = 10^{-12}).

importance for surface temperature, followed by methane, nitrous oxide, and a host of other minor species (Table 1.1).

1.4 HYDROSTATIC BALANCE

The atmosphere is composed of gases held close to the surface of the planet by gravity. The vertical forces acting on the atmosphere at rest are gravity, which pulls the air molecules toward the center of the planet, and the pressure gradient force, which tries to push the atmosphere out into

space. These forces are in balance to a very good approximation, and by equating the pressure gradient force and the gravity force one obtains the *hydrostatic balance*. Since force is mass times acceleration, we may express the vertical force balance per unit mass as an equation between the downward acceleration of gravity, g , and the upward acceleration that would be caused by the increase of pressure toward the ground, if gravity were not present to oppose it.

$$g = -\frac{1}{\rho} \frac{dp}{dz} \quad (1.2)$$

For an ideal gas, pressure (p), density (ρ), and temperature (T) are related by the formula

$$p = \rho RT \quad (1.3)$$

where R is the gas constant. After some rearrangement, (1.2) and (1.3) yield

$$\frac{dp}{p} = -\frac{dz}{H} \quad (1.4)$$

where

$$H = \frac{RT}{g} = \text{scale height}. \quad (1.5)$$

If the atmosphere is *isothermal*, with temperature ~ 260 K, then the temperature and scale height are constant and the hydrostatic equation may be integrated from the surface, where $p = p_s = 1.01325 \times 10^5$ Pa, to an arbitrary height, z , yielding an expression for the distribution of pressure with height.

$$p = p_s e^{-z/H} \quad (1.6)$$

The pressure thus decreases exponentially away from the surface, declining by a factor of $e^{-1} = (2.71828)^{-1} = 0.368$ every scale height. The scale height for the mean temperature of Earth's atmosphere is about 7.6 km. [Figure 1.7](#) shows the distribution of atmospheric pressure with altitude. The pressure is largest at the surface and decreases rapidly with altitude in accord with the exponential decline given by (1.6). We can rearrange (1.2) to read

$$dm \equiv \rho dz = \frac{-dp}{g} \quad (1.7)$$

The mass between two altitudes, dm , is related to the pressure change between those two levels. Because of hydrostatic balance, the total mass of the atmosphere may be related to the global mean surface pressure.

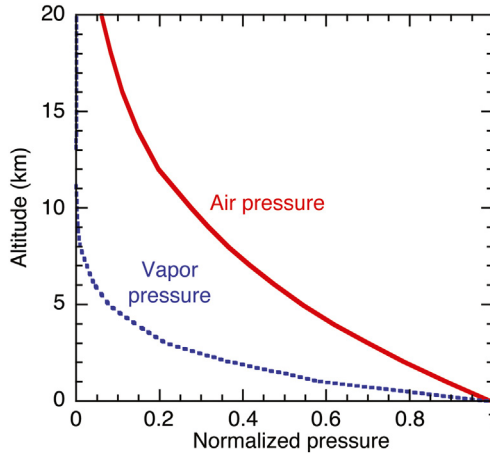


FIGURE 1.7 Vertical distributions of air pressure and partial pressure of water vapor as functions of altitude for globally and annually averaged conditions. Values have been normalized by dividing by the surface values of 1013.25 and 17.5 hPa, respectively.

$$\text{Atmospheric mass} = \frac{p_s}{g} = 1.03 \times 10^4 \text{ kg m}^{-2} \quad (1.8)$$

The vertical column above every square meter of Earth's surface contains about 10,000 kg of air.

Because the surface climate is of primary interest, and because the mass of the atmosphere is confined to within a few scale heights of the surface, or several tens of kilometers, it is the lower atmosphere that is of most importance for climate. For this reason, most of this book will be devoted to processes taking place in the troposphere, at the surface, or in the ocean. The stratosphere has some important effects on climate, however, and these will be described where appropriate.

1.5 ATMOSPHERIC HUMIDITY

Atmospheric humidity is the amount of water vapor carried in the air. It can be measured as vapor pressure, mixing ratio or specific humidity. Specific humidity is the ratio of vapor mass to total air mass, whereas mixing ratio is the ratio of the mass of vapor to the mass of dry air. The atmosphere must carry away the water evaporated from the surface and supply water to areas of rainfall. Water that flows from the land to the oceans in rivers was brought to the land areas by transport in the atmosphere as vapor. Atmospheric water vapor is also the most important greenhouse gas in the atmosphere. Water vapor condenses to form clouds,

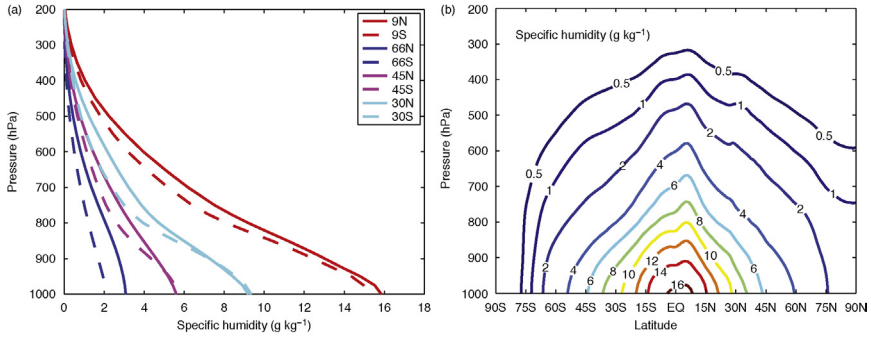


FIGURE 1.8 Profiles of specific humidity as a function of pressure, for annual-mean conditions as (a) Line plots and (b) Contour plot. Data from ERA-Interim.

which can release rainfall and are also extremely important in both reflecting solar radiation and reducing the infrared radiation emitted by Earth.

The partial pressure of water vapor in the atmosphere decreases very rapidly with altitude (Fig. 1.7). The partial pressure of water vapor decreases to half of its surface value by 2 km above the surface and to less than 10% of its surface value at 5 km. Atmospheric water vapor also decreases rapidly with latitude (Fig. 1.8). The amount of water vapor in the atmosphere at the equator is nearly 10 times that at the poles.

The rapid upward and poleward decline in water vapor abundance in the atmosphere is associated with the strong temperature dependence of the saturation vapor pressure. The vapor pressure in equilibrium with a wet surface increases very rapidly with temperature. The temperature dependence of saturation pressure of water vapor over a water surface is governed by the *Clausius–Clapeyron relationship*.

$$\frac{de_s}{dT} = \frac{L}{T(\alpha_v - \alpha_l)} \quad (1.9)$$

In (1.9), e_s is the saturation vapor pressure above a flat liquid surface, L is the latent heat of vaporization, T is the temperature in K, and α represents the specific volume of the vapor α_v and liquid α_l forms of water. The Clausius–Clapeyron relation can be manipulated to express the fractional change of saturation vapor pressure $\Delta e_s / e_s$, and thereby the specific humidity at saturation, q^* , to the fractional change of temperature. The specific humidity is related to the water vapor pressure approximately as $q \approx 0.622 \frac{e}{p}$.

$$\frac{\Delta q^*}{q^*} = \frac{\Delta e_s}{e_s} \approx \left(\frac{L}{R_v T} \right) \frac{\Delta T}{T} = r \frac{\Delta T}{T} \quad (1.10)$$

where R_v is the gas constant for water vapor.

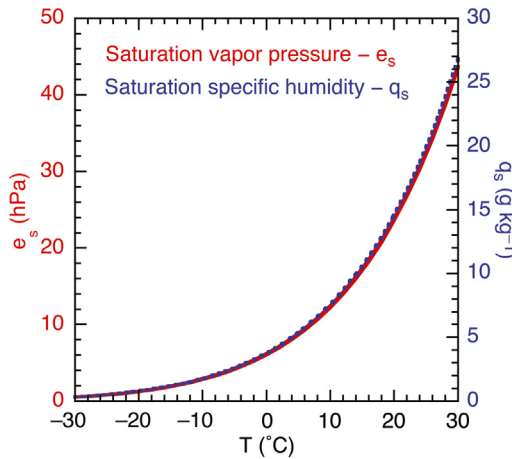


FIGURE 1.9 Saturation vapor pressure and specific humidity as functions of temperature at standard pressure.

For terrestrial conditions, $T \sim 260$ K, and the factor r is approximately 20. This means that a 1% change in temperature of about 3 K will result in a 20% change in the saturation vapor pressure, or about 7% for 1 K. If the relative humidity (the ratio of the actual specific humidity to the saturation specific humidity) remains fixed, then the actual water vapor in the atmosphere will increase by 7% for every 1 K temperature increase. This rapid exponential increase of saturation pressure with temperature can be seen more explicitly if we consider the approximate solution to (1.9) valid near standard pressure and temperature of 1013.25 hectoPascals (hPa) and 273 K.

$$e_s \cong 6.11 \cdot \exp \left\{ \frac{L}{R_v} \left(\frac{1}{273} - \frac{1}{T} \right) \right\} \quad (1.11)$$

The exponential dependence of the saturation vapor pressure on temperature expressed by (1.11) is shown in Fig. 1.9.

1.6 ATMOSPHERIC THERMODYNAMICS, VERTICAL STABILITY AND LAPSE RATE

Conservation of energy is a central constraint on climate, and when combined with the hydrostatic relationship, it determines much about the vertical structure of the atmosphere, including the lapse rate.

1.6.1 First Law of Thermodynamics

The first law of thermodynamics states that energy is conserved, so that for a unit mass of gas, the applied heat δQ is equal to the sum of the change in internal energy dU and the work done dW . If we assume that the external work done by air is only that associated with volume changes, $dW = p d\alpha$, and we use the definition of the specific heat at constant volume $c_v = (dU / dT)_v$, we obtain a useful form of the first law of thermodynamics.

$$\delta Q = c_v dT + p d\alpha \quad (1.12)$$

Here p is pressure, dT is the change in temperature, and $d\alpha$ is the change in specific volume, α . Using the ideal gas law,

$$p\alpha = RT \quad (1.13)$$

we may write that,

$$p d\alpha = R dT - \alpha dp \quad (1.14)$$

Using (1.14), and

$$R = c_p - c_v \quad (1.15)$$

(1.12) becomes,

$$\delta Q = c_p dT - \alpha dp \quad (1.16)$$

or using (1.13) again,

$$\delta Q = c_p dT - \frac{RT}{p} dp \quad (1.17)$$

1.6.2 Potential Temperature

An adiabatic process is one for which no heat is added or taken away so, $\delta Q = 0$, and (1.17) can be rearranged to read,

$$\frac{dT}{T} - \frac{R}{c_p} \frac{dp}{p} = d \ln T - \frac{R}{c_p} d \ln p = d \ln \left(T p^{-R/c_p} \right) = 0 \quad (1.18)$$

so that for a parcel of gas that undergoes an adiabatic process,

$$T p^{-R/c_p} = \text{constant} \quad (1.19)$$

If we define Θ to be the temperature at some reference pressure p_o , which is usually taken to be 1000 hPa, then the potential temperature Θ is the temperature a parcel of air would have if it were brought adiabatically to the reference pressure.

$$\Theta = T \left(\frac{p_0}{p} \right)^{R/c_p} \quad (1.20)$$

1.6.3 Static Stability and the Adiabatic Lapse Rate

The potential temperature is useful because it remains constant as a parcel undergoes an adiabatic change of pressure. The vertical gradient of potential temperature determines the dry static stability of the atmosphere. If the potential temperature increases with height, then parcels raised adiabatically from their initial height will always be colder and thus denser than their environment and will sink back to their original pressure. If the potential temperature decreases with height, then parcels raised up will be warmer than their environment and will be accelerated upward by buoyancy; therefore, if the potential temperature decreases with height, the temperature profile is unstable.

$$\begin{aligned} \frac{d\Theta}{dz} > 0 & \quad \text{Stable} \\ \frac{d\Theta}{dz} < 0 & \quad \text{Unstable} \end{aligned} \quad (1.21)$$

The rate at which temperature changes as a parcel moves up or down in the atmosphere without heating can be derived by using the hydrostatic equation,

$$\alpha dp = -g dz \quad (1.22)$$

in (1.16), and setting $\delta Q = 0$, to give

$$c_p dT + g dz = 0 \quad (1.23)$$

or,

$$-\left(\frac{\partial T}{\partial z} \right)_{\text{adiabatic}} = \frac{g}{c_p} = \Gamma_d = 9.8 \text{ K km}^{-1} \quad (1.24)$$

1.6.4 Moist Processes and Equivalent Potential Temperature

When moisture is present in air and an air parcel is raised adiabatically, the parcel can become supersaturated such that the water vapor condenses and latent heat is released. One can incorporate the latent heat release as heating in the first law of thermodynamics by writing the heat release in terms of the change in saturation water vapor mixing ratio, dq^* . The saturated adiabatic lapse rate can then be derived (Wallace and Hobbs; 2006).

$$\Gamma_s = \frac{\Gamma_d}{1 + \frac{L}{c_p} \frac{dq^*}{dT}} \quad (1.25)$$

The saturated adiabatic lapse rate is generally less than the dry adiabatic lapse rate, and becomes smaller as the temperature rises. As a saturated parcel rises, water condenses, latent heat is released and the parcel cools more slowly with increasing altitude than an unsaturated or dry parcel.

Another useful quantity is the equivalent potential temperature, which is the temperature that would be obtained by a moist air parcel if it were first raised moist-adiabatically until all of its water condensed out, and then brought adiabatically back to a reference surface pressure.

$$\Theta_e = \Theta \exp \left(\frac{Lq^*}{c_p T} \right) \quad (1.26)$$

Equivalent potential temperature incorporates the moist static energy of air parcels in a hydrostatic atmosphere. The moist static energy includes the sensible, potential, and latent energy per unit mass (1.27).

$$\text{Moist static energy} = c_p T + gz + Lq \quad (1.27)$$

Potential temperature includes the dry static energy, from which the latent energy in (1.27) is excluded. If the equivalent potential temperature decreases with height, then the air parcel is only conditionally unstable. It is unstable only if it becomes saturated. If the equivalent potential temperature increases with height, then the parcel is absolutely stable.

The important difference between dry adiabatic ascent and moist ascent can be illustrated by plotting the dry and moist adiabats for a few representative cases. The adiabats are the temperature profiles that would be experienced by parcels as they are raised upward adiabatically from the surface. Examples of some dry and saturated adiabats are plotted in Fig. 1.10. Because of the release of latent heat, the temperatures of moist adiabats decrease less rapidly with height in the lower troposphere, but become parallel to the dry adiabats at low temperatures where latent heating is nearly zero, because the saturation vapor pressure is very small. This difference is particularly evident at high temperatures, where the saturated parcel starts out at the surface with much more latent energy and therefore its temperature drops less rapidly with altitude. Because the curvature of the saturated adiabats increases with temperature, when the temperature of the saturated parcel at the surface is increased, its temperature when it arrives at any layer higher in the troposphere is increased by a larger amount than the surface temperature increase. For example, parcels started from the surface at 20 and 30°C have temperatures of −45.6 and −15.5°C when they reach 10 km altitude. The difference of 30°C at 10 km

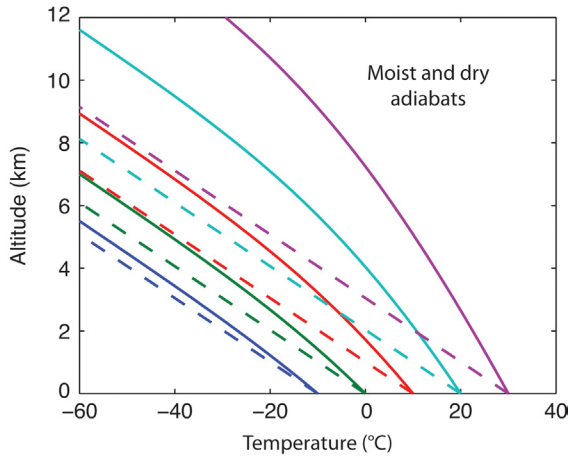


FIGURE 1.10 Moist and dry adiabatic temperature profiles. Parcels of air start at 1000 hPa either saturated with water vapor (solid) or completely dry (dashed), and are raised adiabatically while conserving moist static energy. Starting temperatures of -10 , 0 , 10 , 20 and 30°C are shown.

is three times their difference at the surface. This basic mechanism would indicate that significant changes in lapse rate and dry static stability should be expected when the climate changes, especially in the tropics where moist convection strongly controls the temperature profile.

1.7 THE WORLD OCEAN

The atmosphere contains a tiny fraction of the total water in the climate system, about one part in 10^5 . Most of the surface water of Earth is contained in the oceans and in ice sheets (Table 1.2). Earth contains

TABLE 1.2 Water on Earth

Water reservoir	Depth if spread over the entire surface of Earth (m)	Total (%)
Oceans	2650	97
Icecaps and glaciers	60	2.2
Groundwater	20	0.7
Lakes and streams	0.35	0.013
Soil moisture	0.12	0.013
Atmosphere	0.025	0.0009
Total	2730	100

about $1.35 \times 10^9 \text{ km}^3$ of water, of which about 97% is seawater. Since all the oceans are connected to some degree, we can think of them collectively as the world ocean. The world ocean is a key element of the physical climate system. Ocean covers about 71% of Earth's surface to an average depth of 3730 m. The ocean has tremendous capability to store and release heat and chemicals on time scales of seasons to centuries. Ocean currents move heat poleward to cool the tropics and warm the extratropics. The world ocean is the reservoir of water that supplies atmospheric water vapor for rain and snowfall over land. The ocean plays a key role in determining the composition of the atmosphere through the exchange of gases and particles across the air–sea interface. The ocean removes carbon dioxide from the atmosphere and produces molecular oxygen, and participates in other key geochemical cycles that regulate the surface environment of Earth.

Temperature in the ocean generally decreases with depth from a temperature very near that of the surface air temperature to a value near the freezing point of water in the deep ocean (Fig. 1.11). A thin, well-mixed surface layer is stirred by winds and waves so efficiently that its temperature and salinity are almost independent of depth. Most of the temperature change occurs in the *thermocline*, a region of rapid temperature change with depth in the first kilometer or so of the ocean. Below the thermocline is a deep layer of almost uniform temperature. In middle and high latitudes, the mixed layer is thin in summer and deep in winter (e.g., 45°N in Fig. 1.11).

Salinity of seawater is defined as the number of grams of dissolved salts in a kilogram of seawater. Salinity in the open ocean ranges from about 33 g kg^{-1} to 38 g kg^{-1} . In seawater with a salinity of 35 g kg^{-1} , about 30 g kg^{-1} are composed of sodium and chloride (Table 1.3). Salinity is an important

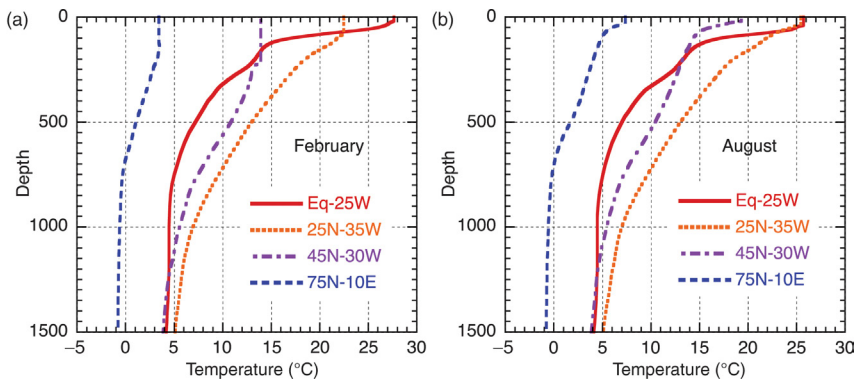
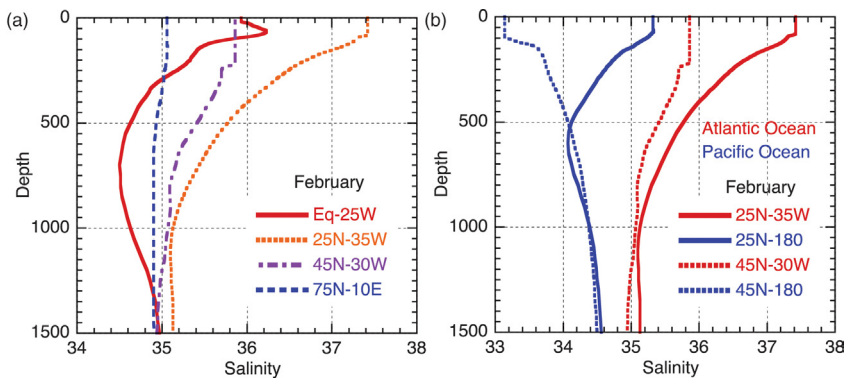


FIGURE 1.11 Annual-mean ocean potential temperature profiles for various latitudes and as a function of depth in meters for (a) February and (b) August. MIMOC data.

TABLE 1.3 Concentrations of the Major Components of Sea Water with a Salinity of 35‰

Component	Grams per kilogram
Chloride	19.4
Sodium	10.8
Sulfur	0.91
Magnesium	1.29
Calcium	0.41
Potassium	0.39
Bicarbonate	0.14
Bromide	0.067
Strontium	0.008
Boron	0.004
Fluoride	0.001

contributor to variations in the density of seawater at all latitudes and is the most important factor in high latitudes and in the deep ocean, where the temperature is close to the freezing point of water. Variations in the density of seawater drive the deep-ocean circulation, which is critical for heat storage and transport and for the recirculation of nutrients necessary for life. Salinity of the global ocean varies systematically with latitude in the upper layers of the ocean (Fig. 1.12). In subtropical latitudes (10–30°), the surface salinity is large because evaporation exceeds precipitation and leaves the seawater enriched in salt. In middle and high

**FIGURE 1.12** Profiles of salinity during February for various locations. MIMOC data.

latitudes, precipitation of freshwater exceeds evaporation, and so surface salinities are quite low. Near the equator, a thin layer of fresh water from precipitation sits atop more saline water below. In the deep ocean, salinity variations are much smaller than near the surface, because the sources and sinks of freshwater are at the surface and the deep water comes from a few areas in high latitudes. The Atlantic is much saltier than the Pacific at nearly all latitudes and for this reason the formation of cold, salty water that can sink to the bottom of the ocean is much more prevalent in the Atlantic than the Pacific (see Chapter 7).

1.8 THE CRYOSPHERE

All of the ice near the surface of Earth is called the *cryosphere*. About 2% of the water on Earth is frozen, and this frozen water constitutes about 80% of the freshwater. Most of the mass of ice is contained in the great ice sheets of Antarctica (89%) and Greenland (8.6%) (Table 1.4). For climate, it is often not the mass of ice that is of primary importance, but rather the surface area that is covered by ice of any depth. This is because surface ice of any depth generally is a much more effective reflector of solar radiation than the underlying surface. Also, sea ice is a good insulator and allows air temperature to be very different from that of the seawater under just a few meters of sea ice. Currently, year-round (perennial) ice covers about 11% of the land area and 7% of the world ocean. During some seasons, the amount of land covered by seasonal snow cover exceeds the surface area covered by perennial ice cover. The surface areas covered by ice sheets, seasonal snow, and sea ice are comparable. Ice sheets cover about $16 \times 10^6 \text{ km}^2$, seasonal snow about $50 \times 10^6 \text{ km}^2$, and sea ice up to $23 \times 10^6 \text{ km}^2$.

1.9 THE LAND SURFACE

Although the land surface covers only 29% of Earth, the climate over the land surface is extremely important to us because humans are land-dwelling creatures. Cereal grains are the world's most important food source, and supply about half of the world's calories and much of the protein. About 80% of the animal protein consumed by humans comes from meat, eggs, and dairy products, and only 20% from seafood.

Over the land surface, temperature and soil moisture are key determinants of natural vegetation and the agricultural potential of a given area. Vegetation, snow cover, and soil conditions also affect the local and

TABLE 1.4 Estimated Global Inventory of Land and Sea Ice

			Area (km ²)	Volume (km ³)	Total ice mass (%)
Land ice	Antarctic ice sheet		13.9×10^6	30.1×10^6	89.3
	Greenland ice sheet		1.7×10^6	2.6×10^6	8.6
	Mountain glaciers		0.5×10^6	0.3×10^6	0.76
	Permafrost	Continuous	8×10^6	(Ice content) $0.2\text{--}0.5 \times 10^6$	0.95
		Discontinuous	17×10^6		
	Seasonal snow (avg. max)	Eurasia	30×10^6	$2\text{--}3 \times 10^3$	
		America	17×10^6		
Sea ice	Southern Ocean	Max	18×10^6	2×10^4	
		Min	3×10^6	6×10^3	
	Arctic Ocean	Max	15×10^6	4×10^4	
		Min	8×10^6	2×10^4	

The volume of water in the ground that annually freezes and thaws at the surface of permafrost (active layer), and in regions without permafrost but with subfreezing winter temperatures is not included in this table.

After Untersteiner (1984); printed with permission from Cambridge University Press.

global climate, so that local climate and land surface conditions participate in a two-way relationship. Land topography plays an important role in modifying regional climates, and weathering of rocks on land is a key component of the carbon cycle that controls the carbon dioxide content of the atmosphere on millennial time scales.

The arrangement of land and ocean areas on Earth plays a role in determining global climate. The arrangement of land and ocean varies on time scales of millions of years as the continents drift about. At the present time, about 68% of Earth's land area is in the Northern Hemisphere (Fig. 1.13). The fact that the Northern Hemisphere has most of the land area causes significant differences in the climates of the Northern and Southern Hemispheres, and plays an important role in climate change. The Northern Hemisphere has much more dramatic east–west variations in continental elevation, especially in middle latitudes where the Himalaya and Rocky Mountains are prominent features (Fig. 1.14). The topography of the land surface and the arrangement and orientation of mountain ranges are key determinants of climate.

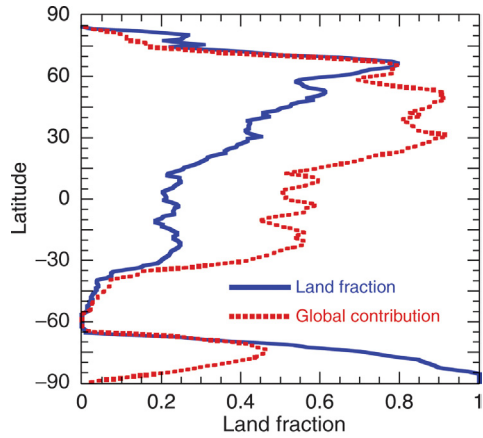


FIGURE 1.13 Fraction of surface area covered by land as a function of latitude (solid line) and contribution of each latitude belt to the global land surface area (dashed line).

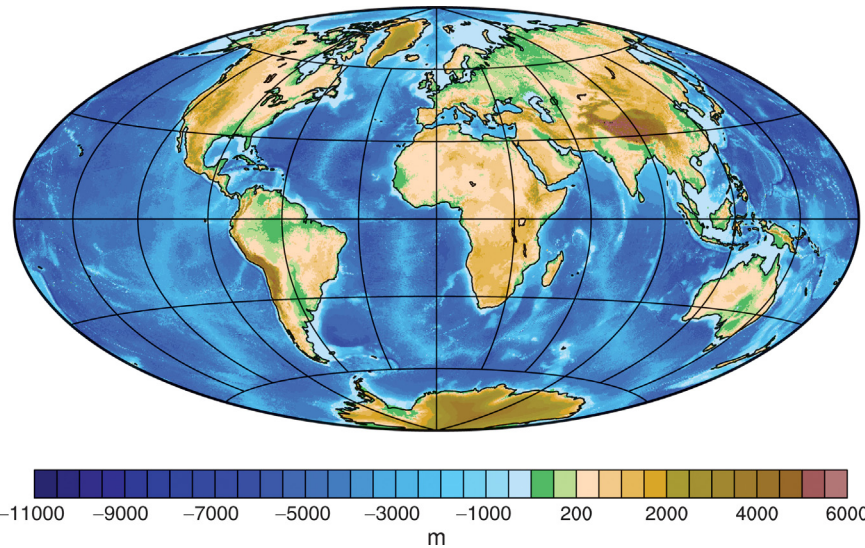


FIGURE 1.14 Color contour plot of the topography of Earth relative to sea level. Scale is in meters.

EXERCISES

1. Give several reasons why the amplitude of the annual variation of surface temperature is greatest in Siberia (Fig. 1.6).
 2. If you are standing atop Mount Everest at 8848 m, about what fraction of mass of the atmosphere is below you? (Use eq. (1.6).)
 3. An airplane is flying at 10,000 m above the surface. What is the pressure outside the airplane in hectoPascals? What is the temperature in degree Celsius? Use global averages.
 4. If the atmosphere warmed up by 5°C, would the atmospheric pressure at 5 km above sea level increase or decrease, and by approximately how much? (Use eq. (1.6).)
 5. Compute the difference of saturation vapor pressure between 0°C and 30°C. Compare the results you get with eqs (1.10) and (1.11).
 6. Explain why the North Polar temperature inversion is present in winter but not in summer?
 7. Why do you think the salinity at 45°N–180°E is so much less than the salinity at 25°N–180°E (Fig. 1.11)?
-