4

Atmospheric moisture budget

Learning objectives

When you have read this chapter you will:

- Be familiar with the major atmospheric components of the hydrological cycle,
- Know the main controls of evaporation and condensation,
- Be aware of the spatial and temporal characteristics of moisture in the atmosphere, evaporation and precipitation,
- Know the different forms of precipitation and typical statistical characteristics,
- Know the major geographical and altitudinal patterns of precipitation and their basic causes,
- Understand the nature and characteristics of droughts.

This chapter considers the role of water in its various phases (solid, liquid and vapour) in the climate system and the transfers (or cycling) of water between the major reservoirs – the oceans, the land surface and the atmosphere. We discuss measures of humidity, large-scale moisture transport, moisture balance, evaporation and condensation.

A THE GLOBAL HYDROLOGICAL CYCLE

The global hydrosphere consists of a series of reservoirs interconnected by water cycling in various phases. These reservoirs are the oceans; ice sheets and glaciers; terrestrial water (rivers, soil moisture, lakes and ground water); the biosphere (water in plants and animals); and the atmosphere. The oceans, with a mean depth of 3.8 km and covering 71 per cent of the earth's surface, hold 97 per cent of *all* the earth's water (23.4×10^6 km^3). Approximately 70 per cent of the total *fresh* water is locked up in ice sheets and glaciers, while almost all of the remainder is ground water. It is an astonishing fact that rivers and lakes hold only 0.3 per cent of all fresh water and the atmosphere a mere 0.04 per cent (Figure 4.1). The average residence time of water within these reservoirs varies from hundreds or thousands of years for the oceans and polar ice to only about ten days for the atmosphere. Water cycling involves evaporation, the transport of water vapour in the atmosphere, condensation, precipitation and terrestrial runoff. The equations of the water budget for the atmosphere and for the surface are respectively:

$$\Delta Q = E - P + D_Q$$

and

$$\Delta S = P - E - r$$

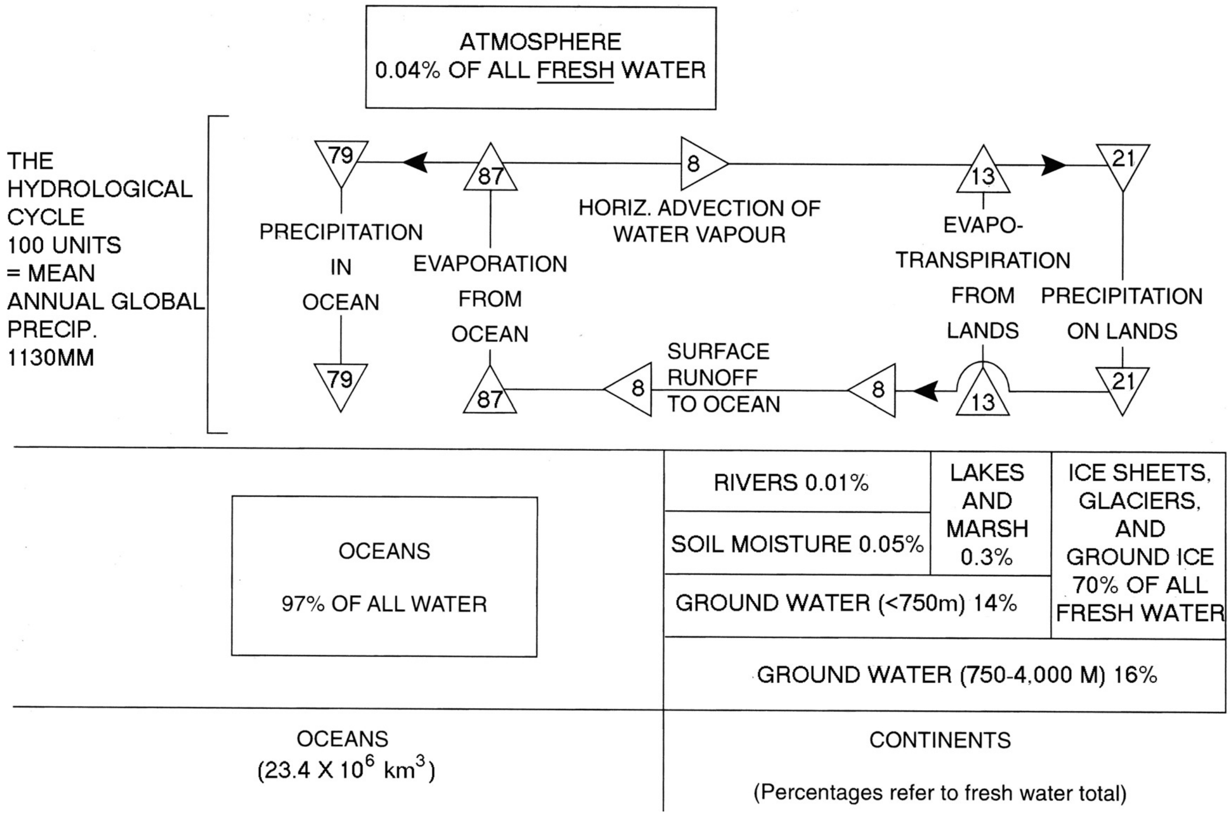

Figure 4.1 The hydrological cycle and water storage of the globe. The exchanges in the cycle are referred to 100 units, which equal the mean annual global precipitation of 1130 mm. The percentage storage figures for atmospheric and continental water are percentages of all *fresh* water. The saline ocean waters make up 97 per cent of *all* water. The horizontal advection of water vapour indicates the *net* transfer.

Source: From More (1967) updated after Korzun (1978).

where ΔQ is the time change of moisture in an atmospheric column, E = evaporation, P = precipitation, D_Q = moisture divergence out of the column, ΔS = surface storage of water, and r = runoff. For short-term processes, the water balance of the atmosphere may be assumed to be in equilibrium; however, over periods of tens of years, global warming may increase its water storage capacity.

Because of its large heat capacity, the global occurrence and transport of water is closely linked to global energy. Atmospheric water vapour is responsible for the bulk of total global energy lost into space by infra-red radiation. Over 75 per cent of the energy input from the surface into the atmosphere is a result of the liberation of latent heat by condensation and, principally, the production of rainfall.

The average storage of water vapour in the atmosphere (Table 4.1), termed the precipitable water content (about 25 mm), is sufficient for only ten days' supply of rainfall over the earth as a whole. However, intense (horizontal) influx of moisture into the air over a given region makes possible short-term rainfall totals greatly in excess of 30 mm. The phenomenal record total of 1870 mm fell on the island of Réunion, off Madagascar, during twenty-four hours in March 1952, and much greater intensities have been observed over shorter periods (see E.2a, this chapter).

Table 4.1 Mean water content of the atmosphere (in mm of rainfall equivalent).

	Northern hemisphere	*Southern hemisphere*	*World*
January	19	25	22
July	34	20	27

Source: After Sutcliffe (1956).

B HUMIDITY

1 Moisture content

Atmospheric moisture comprises water vapour, and water droplets and ice crystals in clouds. Moisture content is determined by local evaporation, air temperature and the horizontal atmospheric transport of moisture. Cloud water, on average, amounts to only 4 per cent of atmospheric moisture. The moisture content of the atmosphere can be expressed in several ways, apart from the vapour pressure (p. 24), depending on which aspect the user wishes to emphasize. The total mass of water in a given volume of air (i.e. the density of the water vapour) is one such measure. This is termed the *absolute humidity* (r_w) and is measured in grams per cubic metre (g m^{-3}). Volumetric measurements are seldom used in meteorology and more convenient is the *mass mixing ratio* (x). This is the mass of water vapour in grams per kilogram of dry air. For most practical purposes, the *specific humidity* (q) is identical, being the mass of vapour per kilogram of air, including its moisture.

More than 50 per cent of atmospheric moisture content is below 850 mb (approximately 1450 m) and more than 90 per cent below 500 mb (5575 m). Figure 4.2 illustrates typical vertical distributions in spring in middle latitudes. It is also apparent that the seasonal effect is most marked in the lowest 3000 m (i.e. below 700 mb). Air temperature sets an upper limit to water vapour pressure – the saturation value (i.e. 100 per cent relative humidity); consequently we may expect the distribution of mean vapour content to reflect this control. In January, minimum values of 1–2 mm (equivalent depth of water) occur in northern continental interiors and high latitudes, with secondary minima of 5–10 mm in tropical desert areas, where there is subsiding air (Figure 4.3). Maximum vapour contents of 50–60 mm are over southern Asia during the summer monsoon and over equatorial latitudes of Africa and South America.

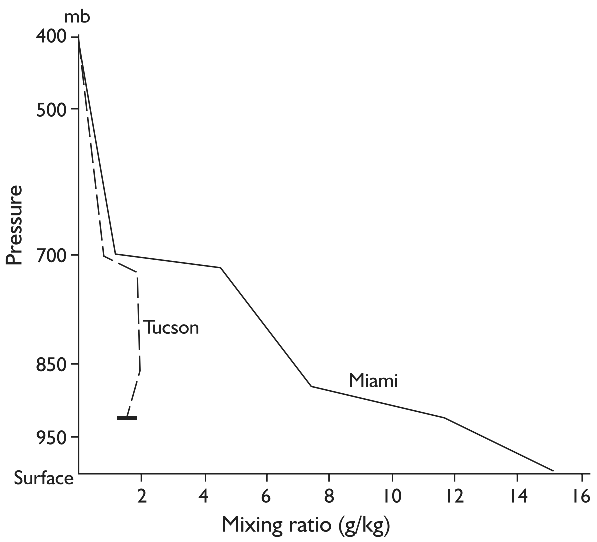

Figure 4.2 The vertical variation of atmospheric vapour content (g/kg) at Tucson, AZ and Miami, FL at 12 UTC on 27 March 2002.

Another important measure is *relative humidity* (r), which expresses the actual moisture content of a sample of air as a percentage of that contained in the same volume of saturated air at the same temperature. The relative humidity is defined with reference to the mixing ratio, but it can be determined approximately in several ways:

$$r = \frac{x}{x_s} \times 100 < \frac{q}{q_s} \times 100 < \frac{e}{e_s} \times 100$$

where the subscript s refers to the respective saturation values at the same temperature; e denotes vapour pressure.

A further index of humidity is the dew-point temperature. This is the temperature at which saturation occurs if air is cooled at constant pressure without addition or removal of vapour. When the air temperature and dew point are equal the relative humidity is 100 per cent, and it is evident that relative humidity can also be determined from:

$$\frac{e_s \text{ at dew-point}}{e_s \text{ at air temperature}} \times 100$$

The relative humidity of a parcel of air will change if either its temperature or its mixing ratio is changed. In general, the relative humidity varies inversely with temperature during the day, tending to be lower in the early afternoon and higher at night.

Atmospheric moisture can be measured by at least five types of instrument. For routine measurements the *wet-bulb thermometer* is installed in a louvred instrument shelter (Stevenson screen). The bulb of the standard thermometer is wrapped in muslin, which is kept moist by a wick from a reservoir of pure water. The

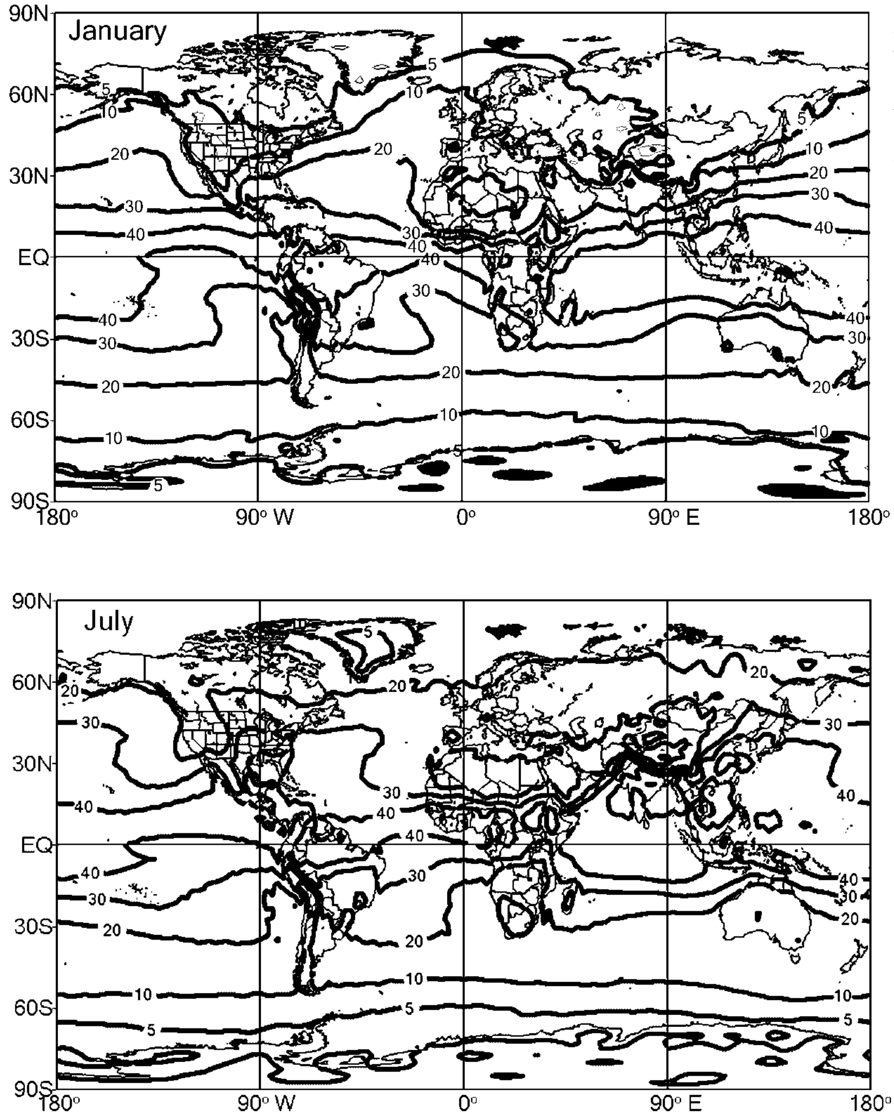

Figure 4.3 Mean atmospheric water vapour content in January and July 1970 to 1999, in mm of precipitable water.

Source: Climate Diagnostics Center, CIRES-NOAA, Boulder, CO.

evaporative cooling of this wet bulb gives a reading that can be used in conjunction with a simultaneous dry-bulb temperature reading to calculate the dew-point temperature. A similar portable device – the aspirated *psychrometer* – uses a forced flow of air at a fixed rate over the dry and wet bulbs. A sophisticated instrument for determining the dew-point, based on a different principle, is the *dew-point hygrometer*. This detects when condensation first occurs on a cooled surface. Two other types of instrument are used to determine relative humidity. The *hygrograph* utilizes the expansion/contraction of a bundle of human hair, in response to humidity, to register relative humidity continuously by a mechanical coupling to a pen arm marking on a rotating drum. It has an accuracy of ± 5 to 10 per cent. For upper air measurements, a *lithium chloride* element detects changes in electrical resistance to vapour pressure differences. Relative humidity changes are accurate within ± 3 per cent.

2 Moisture transport

The atmosphere transports moisture horizontally as well as vertically. Figure 4.1 shows a net transport from oceans to land areas. Moisture must also be transported meridionally (south–north) in order to maintain the required moisture balance at a given latitude (i.e. evaporation – precipitation = net horizontal transport of moisture into an air column). Comparison of annual average precipitation and evaporation totals

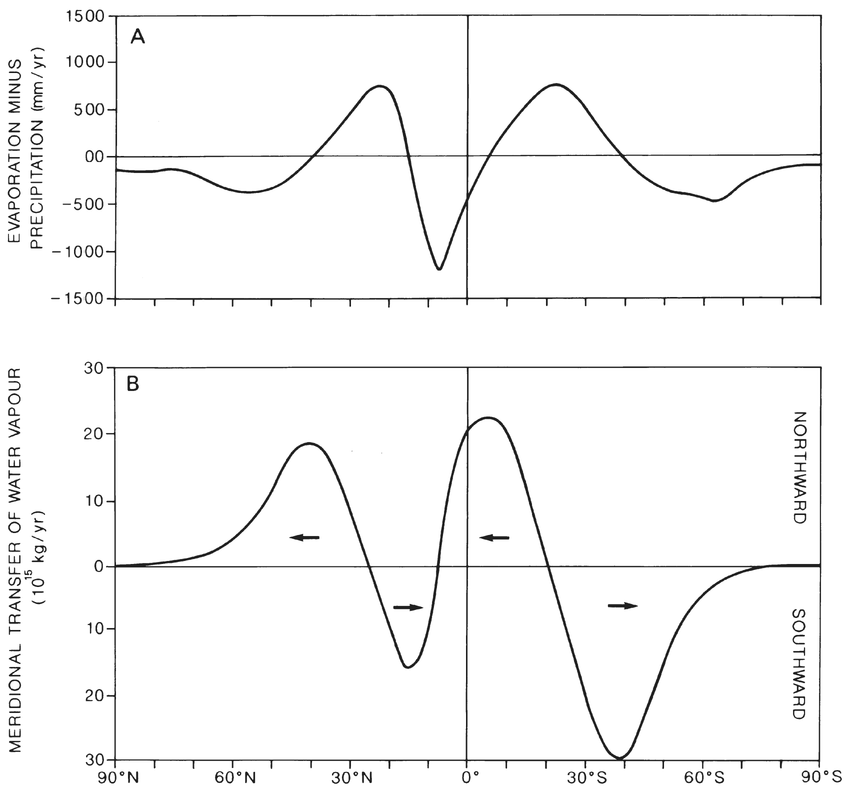

Figure 4.4 Meridional aspects of global moisture. (A) Estimates of annual evaporation minus precipitation (in mm) as a function of latitude; (B) Annual meridional transfer of water vapour (in 10^{15} kg).

Source: (A) After J. Dodd, from Browning (1993). By permission of NERC. (B) From Sellers (1965).

for latitude zones shows that in low and middle latitudes $P > E$, whereas in the subtropics $P < E$ (Figure 4.4A). These regional imbalances are maintained by net moisture transport into (convergence) and out of (divergence) the respective zones (D_Q, where divergence is positive):

$$E - P = D_Q$$

A prominent feature is the equatorward transport into low latitudes and the poleward transport in middle latitudes (Figure 4B). Atmospheric moisture is transported by the global westerly wind systems of middle latitudes towards higher latitudes and by the easterly trade wind systems towards the equatorial region (see Chapter 7). There is also significant exchange of moisture between the hemispheres. During June to August there is a moisture transport northward across the equator of 18.8×10^8 kg s^{-1}; during December to February the southward transport is 13.6×10^8 kg s^{-1}. The net annual south to north transport is 3.2×10^8 kg s^{-1}, giving an annual excess of net precipitation in the northern hemisphere of 39 mm. This is returned by terrestrial runoff into the oceans.

It is important to stress that local evaporation is, in general, not the major source of local precipitation. For example, 32 per cent of the summer season precipitation over the Mississippi River basin and between 25 and 35 per cent of that over the Amazon basin is of 'local' origin, the remainder being transported into these basins by moisture advection. Even when moisture is available in the atmosphere over a region, only a small portion of it is usually precipitated. This depends on the efficiency of the condensation and precipitation mechanisms, both microphysical and large scale.

Using atmospheric sounding data on winds and moisture content, global patterns of average water vapour flux divergence (i.e. $E - P > 0$) or convergence (i.e. $E - P < 0$) can be determined. The distribution of atmospheric moisture 'sources' (i.e. $P < E$) and 'sinks' (i.e. $P > E$) form an important basis for understanding global climates. Strong divergence (outflow) of moisture occurs over the northern Indian Ocean in summer, providing moisture for the monsoon. Subtropical divergence zones are associated with the high-pressure areas. The oceanic subtropical highs are evaporation sources; divergence over land may reflect underground water supply or may be artefacts of sparse data.

C EVAPORATION

Evaporation (including transpiration from plants) provides the moisture input into the atmosphere; the oceans provide 87 per cent and the continents 13 per cent.

The highest annual values (1500 mm), averaged zonally around the globe, occur over the tropical oceans, associated with trade wind belts, and over equatorial land areas in response to high solar radiation receipts and luxuriant vegetation growth (Figure 4.5A). The larger oceanic evaporative losses in winter, for each hemisphere (Figure 4.5B), represent the effect of outflows of cold continental air over warm ocean currents in the western North Pacific and North Atlantic (Figure 4.6) and stronger trade winds in the cold season of the southern hemisphere.

Evaporation requires an energy source at a surface that is supplied with moisture; the vapour pressure in the air must be below the saturated value (e_s); and air motion removes the moisture transferred into the surface layer of air. As illustrated in Figure 2.14, the saturation vapour pressure increases with temperature. The change in state from liquid to vapour requires energy to be expended in overcoming the intermolecular attractions of the water particles. This energy is often acquired by the removal of heat from the immediate surroundings, causing an apparent heat loss (*latent heat*), as discussed on p. 55, and a consequent drop in temperature. The latent heat of vaporization needed to evaporate 1 kg of water at 0°C is 2.5×10^6 J. Conversely, condensation releases this heat, and the temperature of an airmass in which condensation is occurring is increased as the water vapour reverts to the liquid state.

The diurnal range of temperature can be moderated by humid air, when evaporation takes place during the day and condensation at night. The relationship of saturation vapour pressure to temperature (Figure 2.14) means that evaporation processes limit low latitude ocean surface temperature (i.e. where evaporation is at a maximum) to values of about 30°C. This plays an important role in regulating the temperature of ocean surfaces and overlying air in the tropics.

The rate of evaporation depends on a number of factors, the two most important of which are the difference between the saturation vapour pressure at the water surface and the vapour pressure of the air, and the existence of a continual supply of energy to the surface. Wind velocity also affects the evaporation rate, because

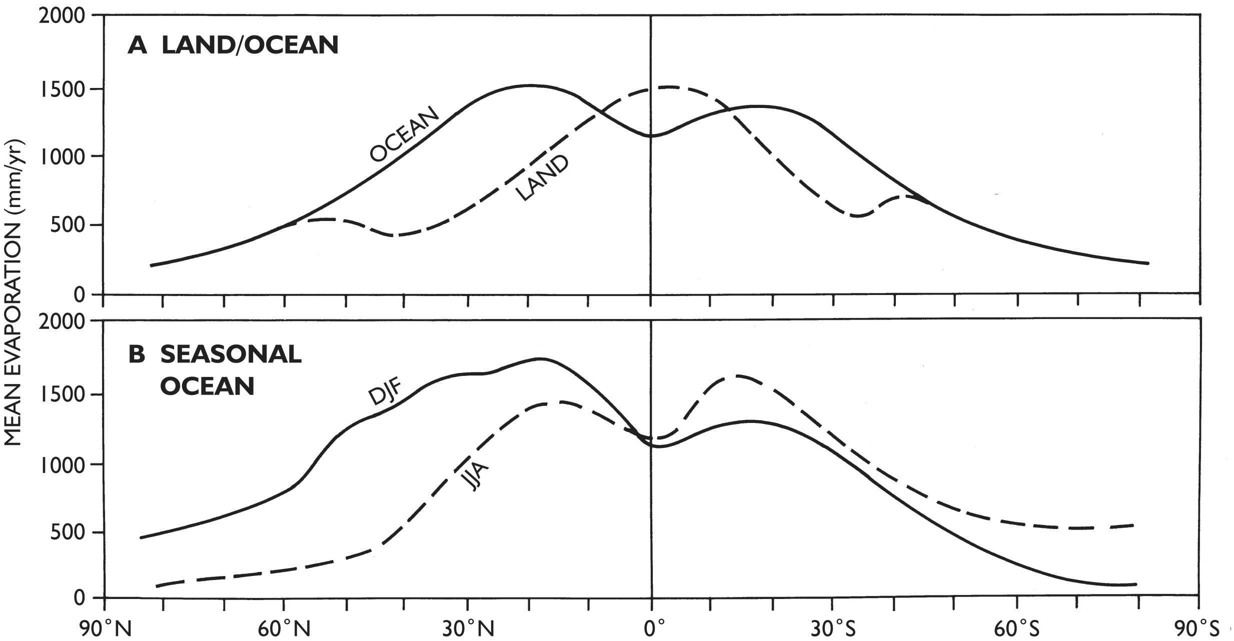

Figure 4.5 Zonal distribution of mean evaporation (mm/year): (A) annually for the ocean and land surfaces, and (B) over the oceans for December to February and June to August.

Sources: After Peixoto and Oort (1983). From *Variations in the Global Water Budget*, ed. A. Street-Perrott, M. Beran and R. Ratcliffe (1983), Fig. 22. Copyright © D. Reidel, Dordrecht, by kind permission of Kluwer Academic Publishers. Also partly from Sellers (1965).

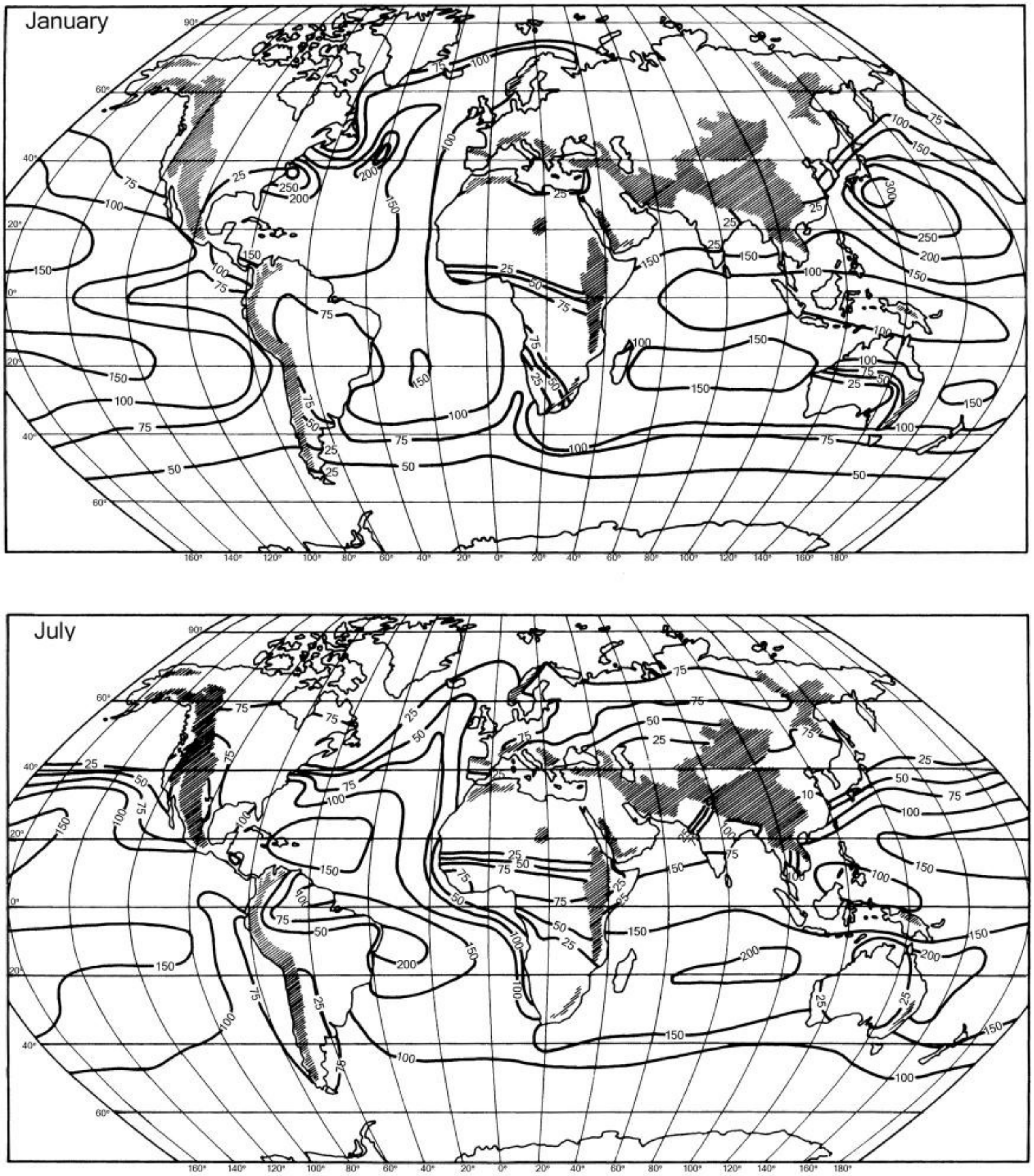

Figure 4.6 Mean evaporation (mm) for January and July.

Source: After M.I. Budyko, *Heat Budget Atlas of the Earth* (1958).

the wind is generally associated with the advection of unsaturated air, which will absorb the available moisture.

Water loss from plant surfaces, chiefly leaves, is a complex process termed *transpiration*. It occurs when the vapour pressure in the leaf cells is greater than the atmospheric vapour pressure. It is vital as a life function in that it causes a rise of plant nutrients from the soil and cools the leaves. The cells of the plant roots can exert an osmotic tension of up to about 15 atmospheres upon the water films between the adjacent soil particles. As these soil water films shrink, however, the tension within them increases. If the tension of the soil films exceeds the osmotic root tension, the continuity of the plant's water uptake is broken and wilting occurs. Transpiration is controlled by the atmospheric factors that determine evaporation as well as by plant factors such as the stage of plant growth, leaf area and leaf temperature, and also by the amount of soil moisture (see Chapter 12C). It occurs mainly during the day, when the *stomata* (small pores in the leaves), through which transpiration takes place, are open. This opening is determined primarily by light intensity. Transpiration naturally varies greatly with season, and during the winter months in mid-latitudes conifers lose only 10 to 18 per cent of their total annual transpiration losses and deciduous trees less than 4 per cent.

In practice, it is difficult to separate water evaporated from the soil, *intercepted moisture* remaining on vegetation surfaces after precipitation and subsequently evaporated, and transpiration. For this reason, evaporation, or the compound term *evapotranspiration*, may be used to refer to the total loss. Over land, annual evaporation is 52 per cent due to transpiration, 28 per cent soil evaporation and 20 per cent interception.

Evapotranspiration losses from natural surfaces cannot be measured directly. There are, however, various indirect methods of assessment, as well as theoretical formulae. One method of estimation is based on the moisture balance equation at the surface:

$$\mathrm{P} - \mathrm{E} = \mathrm{r} + \Delta\mathrm{S}$$

This can be applied to a gauged river catchment, where precipitation and runoff are measured, or to a block of soil. In the latter case we measure the percolation through an enclosed block of soil with a vegetation cover (usually grass) and record the rainfall upon it. The block, termed a *lysimeter*, is weighed regularly so that weight changes unaccounted for by rainfall or runoff can be ascribed to evapotranspiration losses, provided the grass is kept short! The technique allows the determination of daily evapotranspiration amounts. If the soil block is regularly 'irrigated' so that the vegetation cover is always yielding the maximum possible evapotranspiration, the water loss is called the *potential evapotranspiration* (or PE). More generally, PE may be defined as the water loss corresponding to the available energy. Potential evapotranspiration forms the basis for the climate classification developed by C. W. Thornthwaite (see Appendix 1).

In regions where snow cover is long-lasting, evaporation/sublimation from the snowpack can be estimated by lysimeters sunk into the snow that are weighed regularly.

A meteorological solution to the calculation of evaporation uses sensitive instruments to measure the net effect of eddies of air transporting moisture upward and downward near the surface. In this 'eddy correlation' technique, the vertical component of wind and the atmospheric moisture content are measured simultaneously at the same level (say, 1.5 m) every few seconds. The product of each pair of measurements is then averaged over some time interval to determine the evaporation (or condensation). This method requires delicate rapid-response instruments, so it cannot be used in very windy conditions.

Theoretical methods for determining evaporation rates have followed two lines of approach. The first relates average monthly evaporation (E) from large water bodies to the mean wind speed (u) and the mean vapour pressure difference between the water surface and the air ($e_w - e_d$) in the form:

$$E = K_u(e_w - e_d)$$

where K is an empirical constant. This is termed the *aerodynamic approach* because it takes account of the factors responsible for removing vapour from the water surface. The second method is based on the energy budget. The *net balance* of solar and terrestrial radiation at the surface (R_n) is used for evaporation (E) and the transfer of heat to the atmosphere (H). A small proportion also heats the soil by day, but since nearly all of this is lost at night it can be disregarded. Thus:

$$R_n = LE + H$$

where L is the latent heat of evaporation (2.5×10^6 J kg^{-1}). R_n can be measured with a net radiometer and the ratio $H/LE = ß$, referred to as *Bowen's ratio*, can be estimated from measurements of temperature and vapour content at two levels near the surface. ß ranges from <0.1 for water to ≥ 10 for a desert surface. The use of this ratio assumes that the vertical transfers of heat and water vapour by turbulence take place with equal efficiency. Evaporation is then determined from an expression of the form:

$$E = \frac{R_n}{L(1 + b)}$$

The most satisfactory climatological method devised so far combines the energy budget and aerodynamic approaches. In this way, H.L. Penman succeeded in expressing evaporation losses in terms of four meteorological elements that are measured regularly, at least in Europe and North America. These are net radiation (or an estimate based on duration of sunshine), mean air temperature, mean air humidity and mean wind speed (which limit the losses of heat and vapour from the surface).

The relative roles of these factors are illustrated by the global pattern of evaporation (see Figure 4.6). Losses decrease sharply in high latitudes, where there is little available energy. In middle and lower latitudes there are appreciable differences between land and sea. Rates are naturally high over the oceans in view of the unlimited availability of water, and on a seasonal basis the maximum rates occur in January over the western Pacific and Atlantic, where cold continental air blows across warm ocean currents. On an annual basis, maximum oceanic losses occur about 15 to 20°N and 10 to 20°S, in the belts of the constant trade winds (see Figures 4.5B and 4.6). The highest annual losses, estimated to be about 2000 mm, are in the western Pacific and central Indian Ocean near 15°S (cf. Figure 3.30); 2460 MJ m^{-2} yr^{-1} (78 W m^{-2} over the year) are equivalent to an evaporation of 900 mm of water. There is a subsidiary equatorial minimum over the oceans, as a result of the lower wind speeds in the doldrum belt and the proximity of the vapour pressure in the air to its saturation value. The land maximum occurs more or less at the equator due to the relatively high solar radiation receipts and the large transpiration losses from the luxuriant vegetation of this region. The secondary maximum over land in mid-latitudes is related to the strong prevailing westerly winds.

The annual evaporation over Britain, calculated by Penman's formula, ranges from about 380 mm in Scotland to 500 mm in parts of south and southeast England. Since this loss is concentrated in the period May to September, there may be seasonal water deficits of 120 to 150 mm in these parts of the country necessitating considerable use of irrigation water by farmers. The annual moisture budget can also be determined approximately by a bookkeeping method devised by C.E. Thornthwaite, where potential evapotranspiration

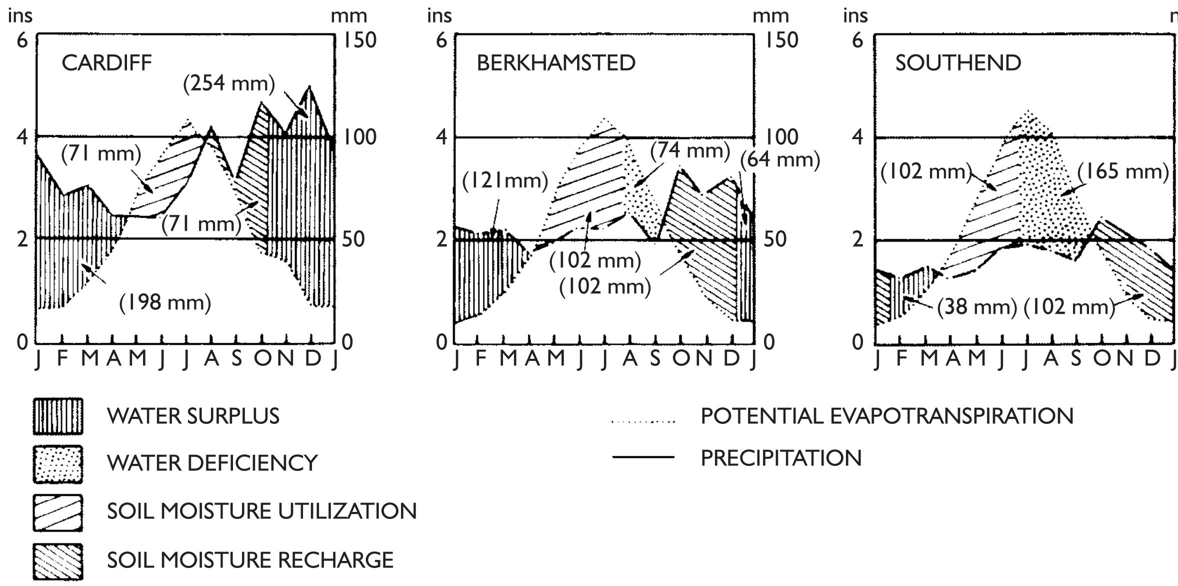

Figure 4.7 The average annual moisture budget for stations in western, central and eastern Britain determined by Thornthwaite's method. When potential evaporation exceeds precipitation soil moisture is used; at Berkhamsted in central England and Southend on the east coast, this is depleted by July to August. Autumn precipitation excess over potential evaporation goes into replenishing the soil moisture until field capacity is reached.

Source: From Howe (1956). Reprinted from *Weather*, by permission of the Royal Meteorological Society. Crown copyright ©.

is estimated from mean temperature. Figure 4.7 illustrates this for stations in western, central and eastern Britain (compare Figure 10.22). In the winter months there is an excess of precipitation over evaporation; this goes to recharging the soil moisture, and further surplus runs off. In summer, when evaporation exceeds precipitation, soil moisture is used initially to maintain evaporation at the potential value, but when this store is depleted there is a water deficiency, as shown in Figure 4.7 for Southend.

In the United States, monthly moisture conditions are commonly evaluated on the basis of the Palmer Drought Severity Index (PDSI). This is determined from accumulated weighted differences between actual precipitation and the calculated amount required for evapotranspiration, soil recharge and runoff. Accordingly, it takes account of the persistence effects of drought situations. The PDSI ranges from ≥4 (extremely moist) to ≤–4 (extreme drought). Figure 4.8 indicates an oscillation between drought and unusually moist conditions in the continental USA during the period October 1992 to August 1993.

D CONDENSATION

Condensation is the direct cause of all the various forms of precipitation. It occurs as a result of changes in air volume, temperature, pressure or humidity. Four mechanisms may lead to condensation: (1) the air is cooled to dew-point but its volume remains constant; (2) the volume of the air is increased without addition of heat; this cooling occurs because adiabatic expansion causes energy to be consumed through work (see Chapter 5); (3) a joint change of temperature and volume reduces the moisture-holding capacity of the air below its existing moisture content; or (4) evaporation adds moisture to the air. The key to understanding condensation lies in the fine balance that exists between these variables. Whenever the balance between one or more of these variables is disturbed beyond a certain limit, condensation may result.

The most common circumstances favouring condensation are those producing a drop in air temperature; namely contact cooling, radiative cooling, mixing of airmasses of different temperatures and dynamic cooling of the atmosphere. Contact cooling occurs

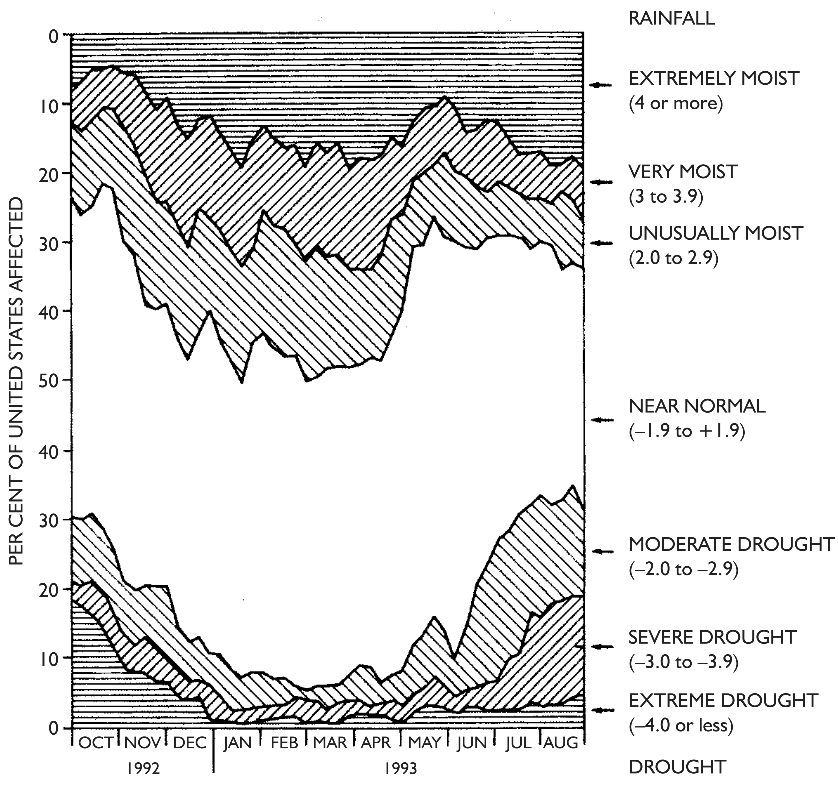

Figure 4.8 Percentage of the continental USA affected by wet spells or drought, based on the Palmer Index (see scale on right), during the period October 1992 to August 1993.

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