2.10.6 Topographical effects on precipitation distribution

Precipitation amount may be increased in coastal areas by the difference in the convergence of air caused by increased friction over the land compared to that experienced over the sea. Such an effect is reflected in monthly rainfall totals over the coastal area of The Netherlands (see KNMI, 1972).

Differences between land and sea can cause fronts associated with mid-latitude depressions to become stationary just off-shore, which can lead to very large falls of precipitation in coastal regions, particularly if the land is hilly (see Bosart, 1975, 1981; Browning, 1983b). The sea or land breezes associated with air circulation caused by temperature differences between the land and the sea may organize convection such that precipitation amounts can be very much enhanced. These effects are evident in winter on the Great Lakes in North America (for example Passarelli and Braham, 1981).

2.11 Global atmospheric circulation

In this chapter we have discussed the way in which atmospheric water vapour may be condensed to form precipitation. The average time that a molecule of water is resident in the atmosphere as vapour or within a cloud is 10–12 days (Miller, 1977). During this period the molecule may travel a considerable distance within the atmospheric circulation before being returned to the surface of the Earth as precipitation. Indeed, it is the global circulation of the atmosphere as a whole which dictates the occurrence of the atmospheric systems outlined earlier, and hence the global precipitation distribution.

The climatological distribution of annual rainfall has been summarized by Houze (1981) and Sellers (1965). A more recent visualization is shown in Figure 2.14. The precipitation maxima in mid-latitudes are associated with depressions, showers and thunderstorms. In the tropics the rainfall maximum is associated with cloud clusters and tropical storms. All these features are present most of the time somewhere, as illustrated in the satellite image in Figure 2.15. The water which passes from the atmosphere to the surface of the Earth is replenished by the processes of evaporation and transpiration, which will be discussed in the next chapter.

Appendix 2.1 Growth of a raindrop

Assume T_a is the temperature of the air at a distance a from a raindrop, and T_s is the temperature at the surface of the raindrop. Then

$$T_s - T_a = \frac{D\lambda(\rho_a - \rho_s)}{Mk} \tag{2.30}$$

where λ is the latent heat of vaporization of water, M is the molecular weight of water, D is the coefficient of diffusion, ρ_a is the vapour density at a great distance from the raindrop, ρ_s is the vapour density at the surface of the raindrop, and k is the thermal

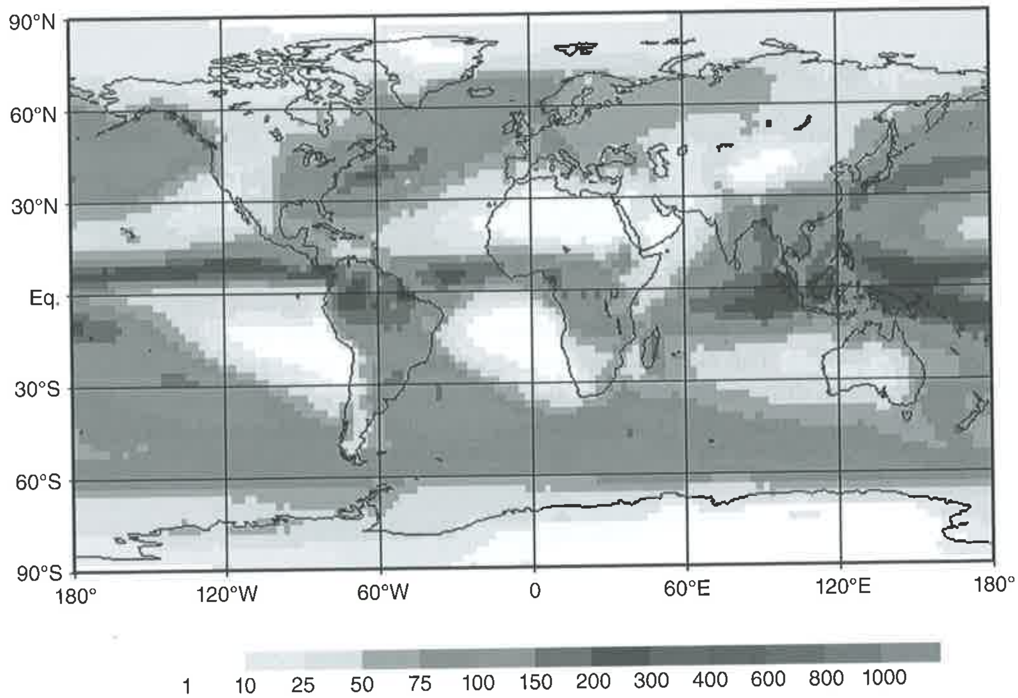

Figure 2.14 Globally averaged annual precipitation 1980–2004 Jan–Dec (mm per month) from the Global Precipitation Climatology Project (GPCP) Version 2 (source: GPCC Visualizer) (*see plate section for colour representation of this figure*)

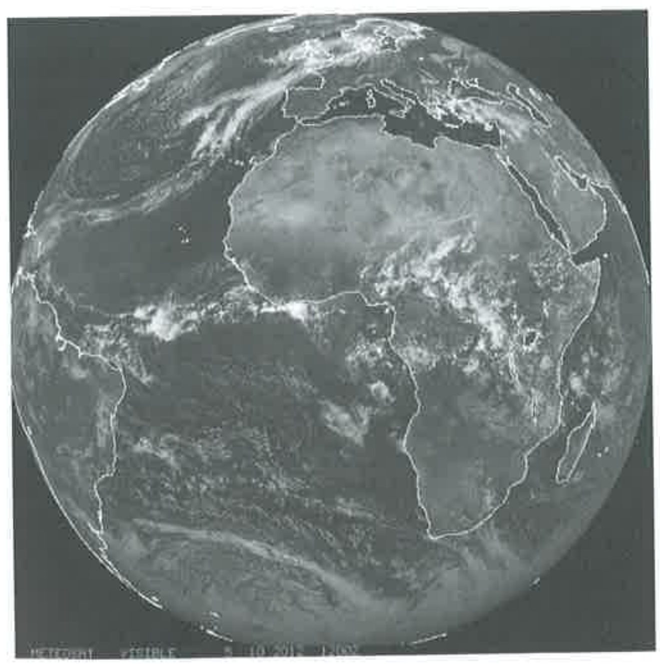

Figure 2.15 Satellite image in the visible wavelengths taken from visible data gathered by the European geostationary satellite MSG at 1200Z on 5 October 2012. Note the cloud associated with the ITCZ north of the equator. The cloud-free Sahara desert is clearly visible, as are cloud clusters in the tropics and mid-latitude depressions (courtesy of Eumetsat)

conductivity of the air. Taking $\Delta T = T_s - T_a$ and $\rho_s = \rho + (d\rho/dT)\Delta T$, where ρ is the saturation density at temperature T, we may derive (Best, 1957)

$$a_2^2 - a_1^2 = 2C(H-1)t \tag{2.31}$$

where $C = (D\rho Mk)/(Mk + D\lambda(d\rho/dT))$; $100H$ is the relative humidity of the environment; and t is the time for the raindrop to grow from radius a_1 to radius a_2. This formula does not apply when the raindrop is sufficiently small for the effect of dissolved salt to be appreciable. However the formula may be used for the evaporation of a droplet which is not completely saturated, i.e. $H < 1$.

Summary of key points in this chapter

1. The term 'precipitation' includes rain, drizzle, sleet, snow and hail.
2. The source of precipitation is water vapour, present in the atmosphere in varying amounts. When it is cooled, water condenses into cloud droplets.
3. The equation of state for a perfect gas relates pressure, volume and temperature using the molecular weight of the gas and the universal gas constant.
4. The first law of thermodynamics states that the energy of a system in a given state relative to a fixed normal state is equal to the algebraic sum of the mechanical equivalent of all the effects produced outside the system, when it passes in any way from the given state to the normal state.
5. Provided no condensation or evaporation occurs, atmospheric processes may be regarded as adiabatic, that is, as involving no exchange of heat between that part of the atmosphere under consideration and its surroundings.
6. The relationship between the equilibrium pressure and the temperature during a phase change is given by the Clausius–Clapeyron equation.
7. When condensation occurs in a sample of air and the resulting water falls out of the sample, then the mass of the sample changes and heat is lost. This is known as a pseudo-adiabatic process.
8. Moist air can become saturated and precipitation can occur as a result of movement of the air upwards as instability is released or convection occurs. Sloping convection, known also as baroclinic convection, is the basic mechanism for the development of large scale atmospheric weather systems.
9. The production of rain is known as the Wegener–Bergeron–Findeisen process. Cloud particles begin as cloud condensation nuclei above the 0 °C level. This is the cold rain process. Warm clouds, below the 0 °C level, produce rain via a warm rain process. If a cloud contains only water then rain is formed by coalescence, but if ice crystals exist then snow occurs by aggregation.
10. Horizontal scales of precipitation systems range from $10^{0.5}$ km to 10^3 km.