GEOB 300

Reading Package

Lecture 7-8

Monteith, J.L. and M.H. Unsworth, 1990: Principles of Environmental Physics, Arnold, London, Pages 50-54.

TERRESTRIAL RADIATION

Most natural surfaces can be treated as 'full' radiators which emit 'terrestrial' or long-wave radiation in contrast to the solar or short-wave radiation emitted by the sun. At a surface temperature of 288 K, the energy per unit wavelength of terrestrial radiation reaches a maximum at 2897/288 or 10 μm (Fig. 3.5) and arbitrary limits of 3 and 100 μm are usually taken to define the long-wave spectrum.

Most of the radiation emitted by the earth's surface is absorbed in specific wavebands by atmospheric gases, mainly water vapour and carbon dioxide. These gases have an emission spectrum similar to their absorption spectrum (Kirchhoff's principle, p. 24) and Fig. 4.9 shows the approximate spectral distribution of the downward flux of atmospheric radiation at the earth's surface. Part of the radiation emitted by the atmosphere is lost to outer space; to satisfy the First Law of Thermodynamics for the earth as a planet, the average annual loss of energy must be equal to the average net gain from solar radiation.

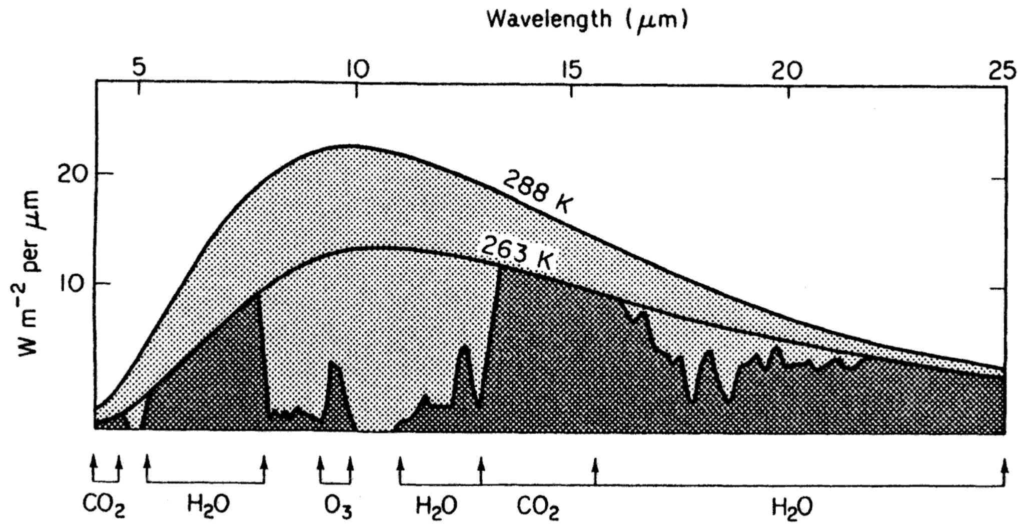

Fig. 4.9 Spectral distribution of long wave radiation for black bodies at 288 K and 263 K. Dark grey areas show the emission from atmospheric gases at 263 K. The light grey area therefore shows the net loss of radiation from a surface at 288 K to a cloudless atmosphere at a uniform temperature of 263 K (after Gates, 1980).

Analysis of the exchange and transfer of long-wave radiation throughout the atmosphere is one of the main problems of physical meteorology but micrometeorologists are concerned primarily with the simpler problem of measuring or estimating fluxes at the surface. The upward flux from a surface L_u can be measured with a radiometer or from a knowledge of the surface temperature and emissivity. The downward flux from the atmosphere L_d can also be measured radiometrically, or calculated.

Cloudless skies

It is easy to demonstrate with a radiometer that the radiance of a cloudless sky in the long-wave spectrum (or the effective radiative temperature) is least at the zenith and greatest near the horizon. This variation is a direct consequence of the increase in the pathlength of water vapour and carbon dioxide, the main emitting gases. In general, more than half the radiant flux received at the ground from the atmosphere comes from gases in the lowest 100 m and roughly 90% from the lowest kilometre. The magnitude of the flux is therefore strongly determined by temperature gradients near the ground.

It is convenient to define the apparent emissivity of the atmosphere ε_a as the flux density of downward radiation divided by full radiation at screen temperature T_a, i.e.

$$\mathbf{L}_d = \varepsilon_a \sigma T_a^4 \tag{4.11}$$

Similarly, the apparent emissivity at a zenith angle ψ or $\varepsilon_a(\psi)$ can be taken as the flux density of downward radiation at ψ divided by σT_a^4. Many measurements show that the dependence of $\varepsilon_a(\psi)$ on ψ over short periods can be expressed as

$$\varepsilon_a(\psi) = a + b \ln(u \sec \psi) \tag{4.12}$$

where u is precipitable water (corrected for the pressure dependence of radiative emission) and a and b are constants, changing with the vertical gradient of temperature and with the distribution of aerosol (Unsworth and Monteith, 1975). Integration of this equation over a hemisphere using equation (3.24) gives the effective (hemispherical) emissivity as

$$\varepsilon_a = a + b(\ln u + 0.5) \tag{4.13}$$

Comparing equations (4.12) and (4.13) shows that the hemispherical emissivity is identical to the emissivity at a representative angle ψ' such that

$$\ln(u \sec \psi') = \ln u + \ln \sec \psi' = \ln u + 0.5$$

It follows that $\ln \sec \psi' = 0.5$, giving $\psi' = 52.5°$ irrespective of the values of a and b.

Because of the difficulty of choosing appropriate values of a and b *a priori*, $\mathbf{L}_d$ is often estimated from empirical formulae as a function of temperature and/or vapour pressure at screen height. One of the most simple formulae is

$$\mathbf{L}_d = c + d\sigma T_a^4 \tag{4.14}$$

and for measurements in the English Midlands (Unsworth and Monteith, 1975) which covered a temperature range from −6 to 26°C, $c = -119 \pm 16$ W m^{-2} and $d = 1.06 \pm 0.04$. The uncertainty of a single estimate of $\mathbf{L}_d$ was ± 30 W m^{-2}. Measurements in Australia (Swinbank, 1963) gave similar values of c and d but with much less scatter.

Using a linear approximation to the dependence of full radiation on temperature above a base of 283 K allows equation (4.14) to be written in the form

$$L_d = 213 + 5.5T_a \tag{4.15}$$

where T_a is now in °C. Outward long-wave radiation assumed to be σT_a^4 is given by a similar approximation as

$$L_u = 320 + 5.2T_a \tag{4.16}$$

The net loss of long-wave radiation is therefore

$$L_u - L_d = 107 - 0.3T_a \tag{4.17}$$

implying that 100 W m^{-2} is a good average figure for the net loss to a clear sky (see Fig. 4.10).

An expression for the effective radiative temperature of the atmosphere T can be obtained by writing

$$L_d = 320 + 5.2T_b = 213 + 5.5T_a$$

so that

$$T_b = (5.5T_a - 107)/5.2 = (T_a - 21) + 0.06T_a$$

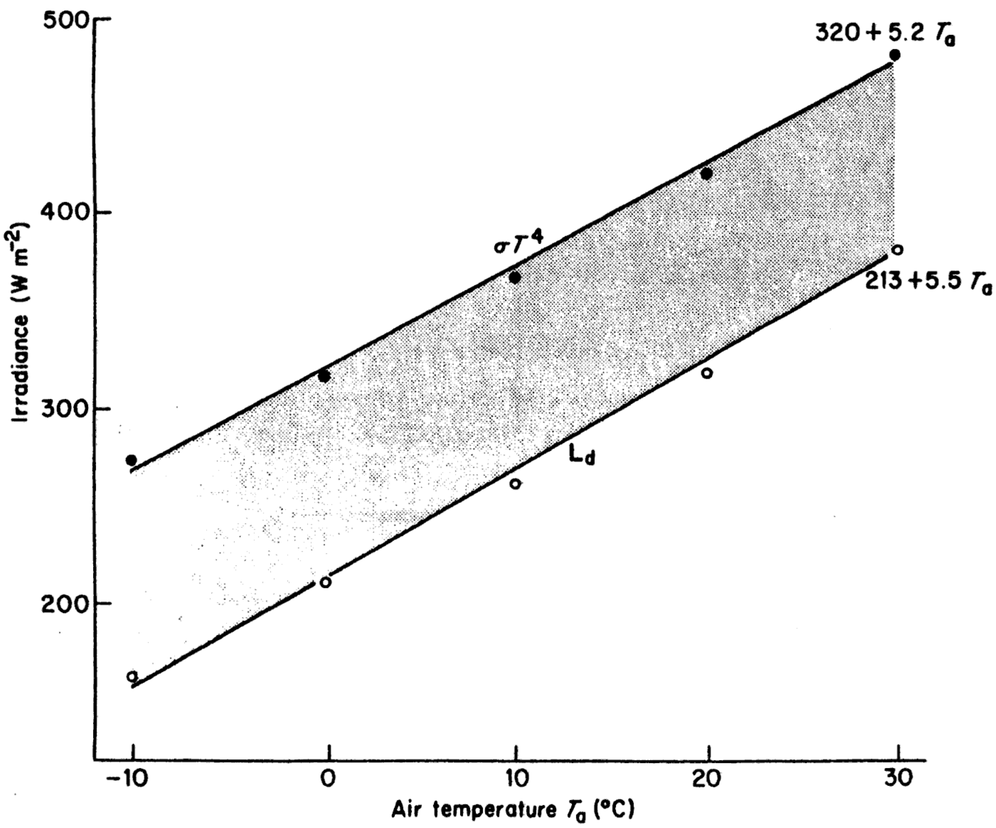

Fig. 4.10 Black body radiation at T_a (full circles) and long wave radiation from clear sky (open circles) from equation (4.14). Straight lines are approximations from equations (4.16) and (4.15) respectively.

This relation shows that when T_a is between 0 and 20°C, the mean effective radiative temperature of a cloudless sky is usually about 21 to 19°C below the mean screen temperature.

Much more complex formulae have been derived to fit observations of L_d under cloudless skies and the most accurate of these incorporate vapour pressure at screen height as well as temperature in order to allow for differences in precipitable water (Hatfield, 1983). However, these formulae have little additional merit for climatological work where the main uncertainty lies in the influence of cloud—the next topic for discussion.

Cloudy skies

Clouds dense enough to cast a shadow on the ground emit like full radiators at the temperature of the water droplets or ice crystals from which they are formed. The presence of cloud increases the flux of atmospheric radiation received at the surface because the radiation from water vapour and carbon dioxide in the lower atmosphere is supplemented by emission from clouds in the waveband which the gaseous emission lacks, i.e. from 8 to 13 μm (see Fig. 4.9). Because most atmospheric radiation originates below the base of clouds, the gaseous component of the downward flux can be treated as if the sky was cloudless with an apparent emissivity ε_a. From Kirchhoff's principle, the transmission of radiation beneath the layer beneath cloud base is $1-\varepsilon_a$ and if the base temperature is T_c the downward radiation from an overcast sky will be

$$\begin{aligned} L_d &= \varepsilon_a \sigma T_a^4 + (1-\varepsilon_a)\sigma T_c^4 \\ &= \sigma T_a^4\{1-(1-\varepsilon_a)4\delta T/T_a\} \end{aligned} \tag{4.18}$$

using the linear approximation equation (3.20) with $\delta T = T_a - T_c$.

The analysis of a series of measurements near Oxford (Unsworth and Monteith, 1975) gave an annual mean of $\delta T = 11$ K with a seasonal variation of ±2 K, figures consistent with a mean cloud base at 1 km, higher in summer than in winter. Taking 283 K as a mean value of T_a gives $4\delta T/T_a = 0.16$, so that the emissivity of an overcast sky at this site is

$$\varepsilon_a(1) = L_d/\sigma T_a^4 = 1-0.16\{1-\varepsilon_a\} = 0.84+0.16\varepsilon_a \tag{4.19}$$

For a sky covered with a fraction c of cloud, interpolation gives

$$\begin{aligned} \varepsilon_a(c) &= c\varepsilon_a(1) + (1-c)\varepsilon_a \\ &= (1-0.84c)\varepsilon_a + 0.84c \end{aligned} \tag{4.20}$$

The main limitation to this formula lies in the choice of appropriate values for cloud and for δT which depend on base height and therefore on cloud type.

It is important to remember that the formulae presented in this section are statistical correlations of radiative fluxes with weather variables at particular sites and do not describe direct functional relationships. For prediction, they are most accurate under average conditions, e.g. when the air temperature does not increase or decrease rapidly with height near the surface and when the air is not unusually dry or humid. They are therefore appropriate for climatological studies of radiation balance but are often not accurate enough for micrometeorological analyses over periods of a few hours. In particular, the equations cannot be used to investigate the diurnal variation of L_d. At some sites, the amplitude of L_d in clear weather is much smaller than the amplitude of L_u, behaviour to be expected if changes of atmospheric temperatures were governed by and followed changes of surface temperature. At other sites, L_d appears to change more than L_u but this difference is hard to explain in terms of atmospheric behaviour and it may be a consequence of small calibration errors in radiometers.