Chapter 3 Radiation Balance Near the Surface

3.1 Radiation Laws and Definitions

The transfer of energy by rapid oscillations of electromagnetic fields is called radiative transfer or simply radiation. These oscillations may be considered as traveling waves characterized by their wavelength or wave frequency c_m/λ, where c_m is the wave speed in a given medium. All electromagnetic waves travel at the speed of light $c \cong 3 \times 10^8$ m s^{-1} in empty space and nearly the same speed in air ($c_m \cong c$). There is an enormous range or spectrum of electromagnetic wavelengths or frequencies. Here we are primarily interested in the approximate range 0.1–100 μm, in which significant contributions to the radiation balance of the atmosphere or the earth's surface occur. This represents only a tiny part of the entire electromagnetic wave spectrum. Of this, the visible light constitutes a very narrow range of wavelengths (0.40–0.76 μm).

The radiant flux density, or simply the radiative flux, is defined as the amount of radiant energy (integrated over all wavelengths) received at or emitted by a unit area of the surface per unit time. The SI unit of radiative flux is W m^{-2}, which is related to the CGS unit cal cm^{-2} min^{-1}, commonly used earlier in meteorology as 1 cal cm^{-2} min^{-1} $\cong$ 698 W m^{-2}.

3.1.1 Blackbody radiation laws

Any body having a temperature above absolute zero emits radiation. If a body at a given temperature emits the maximum possible radiation per unit area of its surface, per unit time, at all wavelengths, it is called a perfect radiator or 'blackbody.' The flux of radiation (R) emitted by such a body is given by the *Stefan–Boltzmann law*

$$R = \sigma T^4 \tag{3.1}$$

where σ is the Stefan–Boltzmann constant = 5.67×10^{-8} W m^{-2} K^{-4}, and T is the surface temperature of the body in absolute (K) units.

Planck's law expresses the radiant energy per unit wavelength emitted by a blackbody as a function of its surface temperature

$$R_\lambda = (2\pi h_p c^2/\lambda^5)[\exp(h_p c/b\lambda T) - 1]^{-1} \tag{3.2}$$

where h_p is Planck's constant $= 6.626 \times 10^{-34}$ J s, and b is the Boltzmann constant $= 1.381 \times 10^{-23}$ J K^{-1}. Note that the total radiative flux is given by

$$R = \int_0^\infty R_\lambda \, d\lambda \tag{3.3}$$

Equation (3.2) can be used to calculate and compare the spectra of blackbody radiation at various body surface temperatures. The wavelength at which R_λ is maximum turns out to be inversely proportional to the absolute temperature and is given by *Wien's law*

$$\lambda_{max} = 2897/T \tag{3.4}$$

when λ_{max} is expressed in micrometers.

From the above laws, it is clear that the radiative flux emitted by a blackbody varies in proportion to the fourth power of its surface temperature, while the wavelengths making the most contribution to the flux, especially λ_{max}, change inversely proportional to T.

3.1.2 Shortwave and longwave radiations

The spectrum of solar radiation received at the top of the atmosphere is well approximated by the spectrum of a blackbody having a surface temperature of about 6000 K. Thus, the sun may be considered as a blackbody with an equivalent surface temperature of about 6000 K and $\lambda_{max} \cong 0.48$ μm.

The observed spectra of terrestrial radiation near the surface, particularly in the absence of absorbing substances such as water vapor and carbon dioxide, can also be approximated by the equivalent blackbody radiation spectra given by Equation (3.2). However, the equivalent blackbody temperature (T_{eb}) is expected to be smaller than the actual surface temperature (T); the difference, $T - T_{eb}$, depends on the radiative characteristics of the surface, which will be discussed later.

Idealized or equivalent blackbody spectra of solar ($T_{eb} = 6000$ K) and terrestrial ($T_{eb} = 287$ K) radiations, both normalized with respect to their peak flux per unit wavelength, are compared in Figure 3.1. Note that almost all the solar energy flux is confined to the wavelength range 0.15–4.0 μm, while the

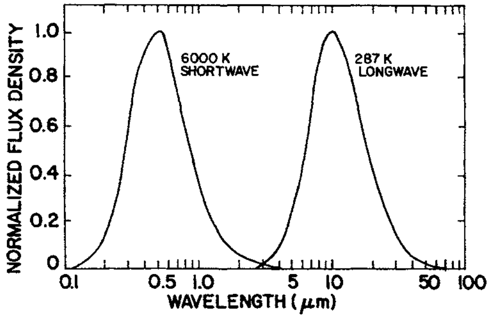

Figure 3.1 Calculated blackbody spectra of flux density of solar and terrestrial radiation, both normalized by their peak flux density. [From Rosenberg *et al.* (1983).]

terrestrial radiation is mostly confined to the range 3–100 μm. Thus, there is very little overlap between the two spectra, with their peak wavelengths (λ_{max}) separated by a factor of almost 20. In meteorology, the above two ranges of wavelengths are characterized as shortwave and longwave radiations, respectively.

3.1.3 Radiative properties of natural surfaces

Natural surfaces are not perfect radiators or blackbodies, but are, in general, gray bodies. They are generally characterized by several different radiative properties, which are defined as follows.

Emissivity is defined as the ratio of the energy flux emitted by the surface at a given wavelength and temperature to that emitted by a blackbody at the same wavelength and temperature. In general, the emissivity may depend on the wavelength and will be denoted by ε_λ. For a blackbody, $\varepsilon_\lambda = 1$ for all wavelengths.

Absorptivity is defined as the ratio of the amount of radiant energy absorbed by the surface material to the total amount of energy incident on the surface. In general, absorptivity is also dependent on wavelength and will be denoted by α_λ. A perfect radiator is also a perfect absorber of radiation, so that $\alpha_\lambda = 1$ for a blackbody.

Reflectivity is defined as the ratio of the amount of radiation reflected to the total amount incident upon the surface and will be denoted by r_λ.

Transmissivity is defined as the ratio of the radiation transmitted to the subsurface medium to the total amount incident upon the surface and will be denoted by t_λ.

It is clear from the above definitions that

$$\alpha_\lambda + r_\lambda + t_\lambda = 1 \tag{3.5}$$

so that absorptivity, reflectivity, and transmissivity must have values between zero and unity.

Kirchoff's law states that for a given wavelength, absorptivity of a material is equal to its emissivity, i.e.,

$$\alpha_\lambda = \varepsilon_\lambda \tag{3.6}$$

Natural surfaces are in general radiatively 'gray,' but in certain wavelength bands their emissivity or absorptivity may be close to unity. For example, in the so-called atmospheric window (8 to 14 μm wavelength band), water, wet soil, and vegetation have emissivities of 0.97–0.99.

In studying the radiation balance for the energy budget near the earth's surface, we are more interested in the radiative fluxes integrated over all the wavelengths of interest, rather than in their wavelength-dependent spectral decomposition. For this, an overall or integrated emissivity and an integrated reflectivity, pertaining to radiation in a certain range of wavelengths, are used to characterize the surface. In particular, the term 'albedo' is used to represent the integrated reflectivity of the surface for shortwave (0.15–4 μm) radiation, while the overall emissivity (ε) of the surface refers primarily to the longwave (3–100 μm) radiation. For natural surfaces, the emission of shortwave radiation is usually neglected, and the emission of longwave radiative flux is given by the modified Stefan–Boltzmann law

$$R_L = -\varepsilon\sigma T^4 \tag{3.7}$$

where the negative sign is introduced in accordance with our sign convention for radiative fluxes. Typical values of surface albedo (a) and emissivity (ε) for different types of surfaces are given in Table 3.1.

The albedo of water is quite sensitive to the sun's zenith angle and may approach unity when the sun is near the horizon (rising or setting sun). It is also dependent on the wave height or sea state. To a lesser extent, the solar zenith angle is also found to affect albedos of soil, snow, and ice surfaces. Although snow has a high albedo, deposition of dust and aerosols, including man-made pollutants such as soot, can significantly lower the albedo of snow. The albedo of a soil surface is strongly dependent on the wetness of the soil.

Infrared emissivities of natural surfaces do not vary over a wide range; most surfaces have emissivities greater than 0.9, while only in a few cases values

Table 3.1 Radiative properties of natural surfaces.

Surface type	Other specifications	Albedo (a)	Emissivity (ε)
Water	Small zenith angle	0.03–0.10	0.92–0.97
	Large zenith angle	0.10–1.00	0.92–0.97
Snow	Old	0.40–0.70	0.82–0.89
	Fresh	0.45–0.95	0.90–0.99
Ice	Sea	0.30–0.45	0.92–0.97
	Glacier	0.20–0.40	
Bare sand	Dry	0.35–0.45	0.84–0.90
	Wet	0.20–0.30	0.91–0.95
Bare soil	Dry clay	0.20–0.40	0.95
	Moist clay	0.10–0.20	0.97
	Wet fallow field	0.05–0.07	
Paved	Concrete	0.17–0.27	0.71–0.88
	Black gravel road	0.05–0.10	0.88–0.95
Grass	Long (1 m)	0.16	0.90
	Short (0.02 m)	0.26	0.95
Agricultural	Wheat, rice, etc.	0.18–0.25	0.90–0.99
	Orchards	0.15–0.20	0.90–0.95
Forests	Deciduous	0.10–0.20	0.97–0.98
	Coniferous	0.05–0.15	0.97–0.99

Compiled from Sellers (1965), Kondratyev (1969), and Oke (1987).

become less than 0.8. Green, lush vegetation and forests are characterized by the highest emissivities, approaching close to unity.

3.2 Shortwave Radiation

The ultimate source of all shortwave radiation received at or near the earth's surface is the sun. A large part of it comes directly from the sun, while other parts come in the forms of reflected radiation from the surface and clouds, and scattered radiation from atmospheric particulates or aerosols.

3.2.1 Solar radiation

As a measure of the intensity of solar radiation, the solar constant (S_0) is defined as the flux of solar radiation falling on a surface normal to the solar beam at the outer edge of the atmosphere, when the earth is at its mean distance from the sun. An accurate determination of its value has been sought by many investigators. The best estimate of $S_0 = 1368\ \mathrm{W\ m^{-2}}$ is based on a series of measurements from high-altitude platforms such as the Solar Maximum Satellite; other values proposed in the literature fall in the range 1350–1400 $\mathrm{W\ m^{-2}}$. There are speculations that the solar constant may vary with changes in the sun spot activity, which has a predominant cycle of about 11 years. There are also suggestions of very long-term variability of the solar constant which may have been partially responsible for long-term climatic fluctuations in the past.

The actual amount of solar radiation received at a horizontal surface per unit area over a specified time is called insolation. It depends strongly on the solar zenith angle γ and also on the ratio (d/d_m) of the actual distance to the mean distance of the earth from the sun. The combination of the so-called inverse-square law and Lambert's cosine law gives the flux density of solar radiation at the top of the atmosphere as

$$R_0 = S_0 \, (d_m/d)^2 \cos \gamma \tag{3.8}$$

Then, insolation for a specified period of time between t_1 and t_2 is given by

$$I_0 = \int_{t_1}^{t_2} R_0(t) \, dt \tag{3.9}$$

Thus, one can determine the daily insolation from Equation (3.9) by integrating the solar flux density with time over the daylight hours.

For a given calendar day/time and latitude, the solar zenith angle (γ) and the ratio d_m/d can be determined from standard astronomical formulas or tables, and the solar flux density and insolation at the top of the atmosphere can be evaluated from Equations (3.8) and (3.9). These are also given in Smithsonian Meteorological Tables.

The solar flux density (R_s) and insolation (I) received at the surface of the earth may be considerably smaller than their values at the top of the atmosphere because of the depletion of solar radiation in passing through the atmosphere. The largest effect is that of clouds, especially low stratus clouds. In the presence of scattered moving clouds, R_s becomes highly variable.

The second important factor responsible for the depletion of solar radiation is atmospheric turbidity, which refers to any condition of the atmosphere, excluding clouds, which reduces its transparency to shortwave radiation. The reduced transparency is primarily due to the presence of particulates, such as pollen, dust, smoke, and haze. Turbidity of the atmosphere in a given area may result from a combination of natural sources, such as wind erosion, forest fires, volcanic eruptions, sea spray, etc., and various man-made sources of aerosols. Particles in the path of a solar beam reflect part of the radiation and scatter the other part.

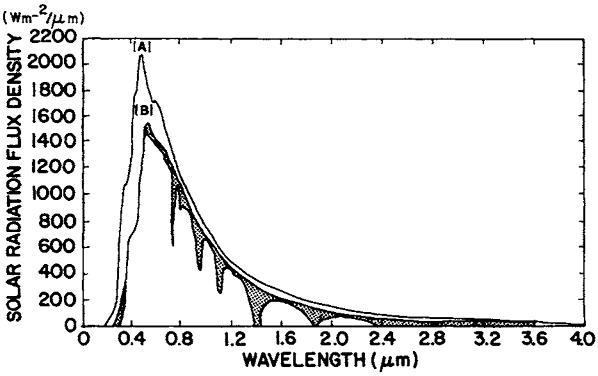

Figure 3.2 Observed flux density of solar radiation at the top of the atmosphere (curve A) and at sea level (curve B). The shaded areas represent absorption due to various gases in a clear atmosphere. [From Liou (1983.]

Large, solid particles reflect more than they scatter light, affecting all visible wavelengths equally. Therefore, the sky appears white in the presence of these particles. Even in an apparently clear atmosphere, air molecules and very small (submicrometer size) particles scatter the sun's rays. According to Rayleigh's scattering law, scattering varies inversely as the fourth power of the wavelength. Consequently, the sky appears blue because there is preferred scattering of blue light (lowest wavelengths of the visible spectrum) over other colors.

Even in a cloud-free nonturbid atmosphere, atmospheric gases, such as oxygen, ozone, carbon dioxide, water vapor, and nitrous oxides, absorb solar radiation in selected wavelength bands. For example, much of the ultraviolet radiation is absorbed by stratospheric ozone and oxygen, and water vapor and carbon dioxide have a number of absorption bands at wavelengths larger than 0.8 μm.

The combined effects of scattering and absorption of solar radiation in a clear atmosphere can be seen in Figure 3.2. Here, curve A represents the measured spectrum of solar radiation at the top of the atmosphere, curve B represents the measured spectrum at sea level, and the shaded areas represent the absorption of energy by various gases, primarily O_3, O_2, CO_2, and H_2O. The unshaded area between curve A and the outer envelope of the shaded area represents the depletion of solar radiation due to scattering.

3.2.2 Reflected radiation

A significant fraction (depending on the surface albedo) of the incoming shortwave radiation is reflected back by the surface. Shortwave reflectivities or albedos for the various natural and man-made surfaces are given in Table 3.1. Knowledge of surface albedos and of their possible changes (seasonal, as well as long term) due to man's activities, such as deforestation, agriculture, and urbanization, has been of considerable interest to climatologists. Satellites now permit systematic and frequent observations of surface albedos over large regions of the world. Any inadvertent or intentional changes in the local, regional, or global albedo may cause significant changes in the surface energy balance and hence in the micro- or macroclimate. For example, deliberate modification of albedo by large-scale surface covering has been proposed as a means of increasing precipitation in certain arid regions. It has also been suggested that large, frequent oil spills in the ice- and snow-covered arctic region could lead to significant changes in the regional energy balance and climate.

Snow is the most effective reflector (high albedo) of shortwave radiation. On the other hand, water is probably the poorest reflector (low albedo), while ice falls between snow and water. Since large areas of earth are covered by water, sea ice, and snow, significant changes in the snow and ice cover are likely to cause perceptible changes in the regional and, possibly, global albedo.

Most bare rock, sand, and soil surfaces reflect 10–45% of the incident shortwave radiation, the highest value being for desert sands. Albedos of most vegetative surfaces fall in the range 10–25%. Wetting of the surface by irrigation or precipitation considerably lowers the albedo, while snowfall increases it. Albedos of vegetative surfaces, being sensitive to the solar elevation angle, also show some diurnal variations, with their minimum values around noon and maximum values near sunrise and sunset (see Rosenberg *et al.*, 1983, Chapter 1).

Shortwave reflectivities of clouds vary over a wide range, depending on the cloud type, height, and size, as well as on the angle of incidence of radiation (Welch *et al.*, 1980). Thick stratus, stratocumulus, and nimbostratus clouds are good reflectors ($a = 0.6$–0.8, at normal incidence), large cumulus clouds are moderate reflectors ($a = 0.2$–0.5), and small cumuli are poor reflectors ($a < 0.2$). Only a small part of solar radiation may reach the ground in cloudy conditions. A significant part of the reflected radiation is likely to undergo multiple reflections between the surfaces and bases of clouds, thereby increasing the effective albedo of the surface.

3.2.3 Diffuse radiation

Diffuse or sky radiation is that portion of the solar radiation that reaches the earth's surface after having been scattered by molecules and suspended particulates in the atmosphere. In cloudy conditions it also includes the portion of the shortwave radiation which is reflected by the clouds. It is the incoming shortwave radiation in shade. Before sunrise and after sunset, all shortwave radiation is in diffuse form. The ratio of diffuse to the total incoming shortwave radiation varies diurnally, seasonally, and with latitude. In high latitudes, diffuse radiation is very important; even in midlatitudes it constitutes 30–40% of the total incoming solar radiation. Cloudiness considerably increases the ratio of diffuse to total solar radiation.

3.3 Longwave Radiation

The longwave radiation R_L received at or near the surface has two components: (1) outgoing radiation $R_{L\uparrow}$ from the surface, and (2) incoming radiation $R_{L\downarrow}$ from the atmosphere, including clouds. These are discussed separately in the following sections.

3.3.1 Terrestrial radiation

All natural surfaces radiate energy, depending on their emissivities and surface temperatures, whose flux density is given by Equation (3.7). As discussed earlier in Section 3.1, emissivities for most natural surfaces range from 0.9 to 1.0. Thus knowing the surface temperature and a crude estimate of emissivity, one can determine terrestrial radiation from Equation (3.7) to better than 10% accuracy. Difficulties arise, however, in measuring or even defining the surface temperature, especially for vegetative surfaces. In such cases, it may be more appropriate to determine the apparent surface temperature, or equivalent blackbody temperature T_{eb} from measurements of terrestrial radiation near the surface using Equation (3.1).

In passing through the atmosphere, a large part of the terrestrial radiation is absorbed by atmospheric gases, such as water vapor, carbon dioxide, nitrogen oxides, methane, and ozone. In particular, water vapor and CO_2 are primarily responsible for absorbing the terrestrial radiation and reducing its escape to the space (the so-called greenhouse effect).

3.3.2 Atmospheric radiation

From our discussion earlier, it is clear that the atmosphere absorbs much of the longwave terrestrial radiation and a significant part of the solar radiation. The atmospheric gases and aerosols which absorb energy also radiate energy, depending on the vertical distributions of their concentrations or mixing ratios and air temperature as functions of height. An important aspect of the atmospheric radiation is that absorption and emission of radiation by various gases occur in a series of discrete wavelengths or bands of wavelengths, rather than continuously across the spectrum, as shown in Figure 3.3. All layers of the atmosphere participate, to varying degrees, in absorption and emission of radiation, but the atmospheric boundary layer is most important in this, because the largest concentrations of water vapor, CO_2, and other gases occur in this layer.

Clouds, when present, are the major contributors to the incoming longwave radiation to the surface. They radiate like blackbodies ($\varepsilon \cong 1$) at their respective cloud base temperatures. However, some of the radiation is absorbed by water vapor, CO_2, and other absorbing gases before reaching the earth's surface.

Computation of incoming longwave radiation from the atmosphere is tedious and complicated, even when the distributions of water vapor, CO_2, cloudiness, and temperature are measured. It is preferable to measure $R_{L\downarrow}$ directly with an appropriate radiometer.

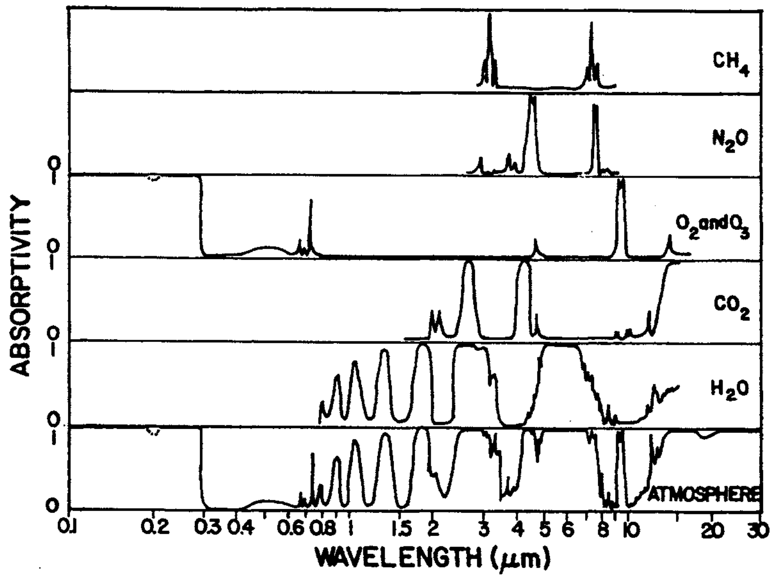

Figure 3.3 Absorption spectra of water vapor, carbon dioxide, oxygen and ozone, nitrogen oxide, methane and the atmosphere. [From Fleagle and Businger (1980).]

3.4 Radiation Balance Near the Surface

The net radiation flux R_N in Equations (2.1) and (2.2) is a result of radiation balance between shortwave (R_S) and longwave (R_L) radiations at or near the surface, which can be written as

$$R_N = R_S + R_L \tag{3.10}$$

Further, expressing shortwave and longwave radiation balance terms as

$$R_S = R_{S\downarrow} + R_{S\uparrow} \tag{3.11}$$

$$R_L = R_{L\downarrow} + R_{L\uparrow} \tag{3.12}$$

the overall radiation balance can also be written as

$$R_N = R_{S\downarrow} + R_{S\uparrow} + R_{L\downarrow} + R_{L\uparrow} \tag{3.13}$$

where the downward and upward arrows denote incoming and outgoing radiation components, respectively.

The incoming shortwave radiation ($R_{S\downarrow}$) consists of both the direct-beam solar radiation and the diffuse radiation. It is also called insolation at the ground and can be easily measured by a solarimeter. It has strong diurnal variation (almost sinusoidal) in the absence of fog and clouds.

The outgoing shortwave radiation ($R_{S\uparrow}$) is actually the fraction of $R_{S\uparrow}$ that is reflected by the surface, i.e.,

$$R_{S\uparrow} = -aR_{S\downarrow} \tag{3.14}$$

where a is the surface albedo. Thus, for a given surface, the net shortwave radiation $R_S = (1 - a)R_{S\downarrow}$ is essentially determined by insolation at the ground.

The incoming longwave radiation ($R_{L\downarrow}$) from the atmosphere, in the absence of clouds, depends primarily on the distributions of temperature, water vapor, and carbon dioxide. It does not show a significant diurnal variation. The outgoing terrestrial radiation ($R_{L\uparrow}$), being proportional to the fourth power of the surface temperature in absolute units, shows stronger diurnal variation, with its maximum value in the early afternoon and minimum value at dawn. The two components are usually of the same order of magnitude, so that the net longwave radiation (R_L) is generally a small quantity.

Under clear skies, $|R_L| \ll R_S$ during the bright daylight hours and an approximate radiation balance is given by

$$R_N \cong R_S = (1 - a)R_{S\downarrow} \tag{3.15}$$

which can be used to determine net radiation from simpler measurements or calculation of solar radiation at the surface. At nighttime, however, $R_{S\downarrow} = 0$, and the radiation balance becomes

$$R_N = R_L = R_{L\downarrow} + R_{L\uparrow} \tag{3.16}$$

Frequently at night $R_{L\downarrow} < -R_{L\uparrow}$, so that R_N or R_L is usually negative, implying radiative cooling of the surface.

Near the sunrise and sunset times, all the components of the radiation balance are of the same order of magnitude and Equation (3.10) or (3.13) will be more appropriate than the simplified Equation (3.15) or (3.16).

Example Problem 1

The following radiation measurements were made over a dry, bare field during a very calm and clear spring night:

Outgoing longwave radiation from the surface = 400 W m^{-2}.

Incoming longwave radiation from the atmosphere = 350 W m^{-2}.

(a) Calculate the equivalent blackbody surface temperature, as well as the actual surface temperature if the surface emissivity is 0.95.

(b) Estimate the ground heat flux, making appropriate assumptions about other fluxes.

Solution

(a) $R_{L\uparrow} = \sigma T_{eb}^4 = -400$ W m^{-2}; $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4}, so that $T_{eb} = (400/\sigma)^{1/4} \cong 289.8$ K, is the equivalent blackbody temperature.
Also, $R_{L\uparrow} = -\varepsilon\sigma T_S^4$, so that $T_S = (400/\varepsilon\sigma)^{1/4} \cong 293.6$ K, is the actual surface temperature.

(b) Using the energy budget equation (Equation 2.1), $H_G = R_N - (H + H_L)$.
Very calm night implies $H + H_L \cong 0$, so that
$H_G = R_N = R_{L\uparrow} + R_{L\downarrow}$, using Equation (3.16).
Thus, $H_G = -400 + 350 = -50$ W m^{-2}, is the ground heat flux.

3.5 Observations of Radiation Balance

For the purpose of illustration, some examples of measured radiation balance over different types of surfaces are provided. Figure 3.4 represents the various

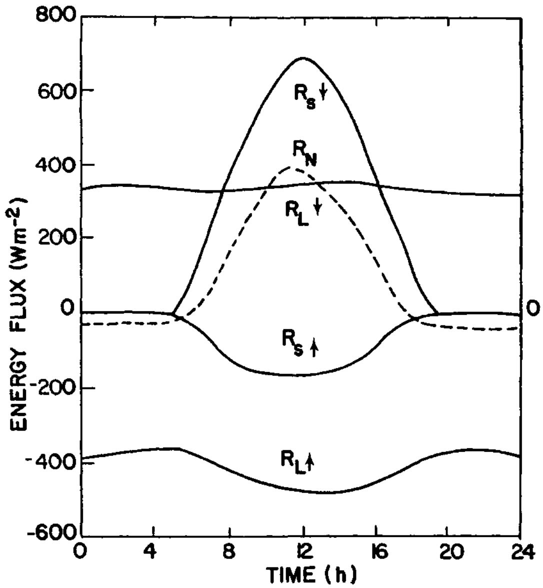

Figure 3.4 Observed radiation budget over a 0.2 m stand of native grass at Matador, Saskatchewan, on 30 July 1971. [From Oke (1987); after Ripley and Redmann (1976).]

components of Equation (3.10) for a clear August day in England over a thick stand of grass, when the diurnal variation of grass temperature was about 20°C. Note that in this case about one-fourth of the incident solar radiation was reflected back by the surface, while about three-fourths was absorbed. The net radiation is slightly less than R_S, even during the midday hours, because of the net loss due to longwave radiation. The diurnal variation of $R_{L\downarrow}$ is much less than that of $R_{L\uparrow}$ (28% variation in $R_{L\uparrow}$ is consistent with the observed diurnal range of 20°C in surface temperature).

Figure 3.5 shows the diurnal variation of radiation budget components over Lake Ontario on a clear day in August. Note that the measured shortwave radiation ($R_{S\downarrow}$) near the lake surface is only about two-thirds of the computed solar radiation (R_0) at the top of the atmosphere. Of this, 25–30% is in the form of diffuse-beam radiation (R_D) at midday and the rest as direct-beam solar radiation (not plotted). The outgoing shortwave radiation ($R_{S\uparrow}$) is relatively small, due to the low albedo ($a \cong 0.07$) of water. Both the incoming and outgoing longwave radiation components are relatively constant with time, due to small diurnal variations in the temperatures of the lake surface and the air above it. The net longwave radiation is a constant energy loss throughout the period of observations. The net radiation (R_N) is dominated by $R_{S\downarrow}$ during the day and is equal to R_L at night.

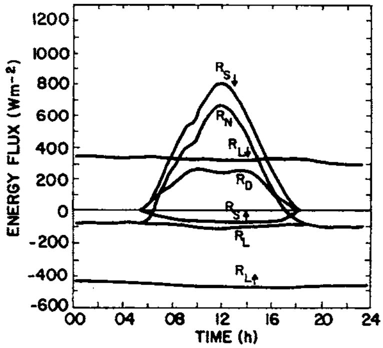

Figure 3.5 Observed radiation budget over Lake Ontario under clear skies on 28 August 1969. [From Oke (1987); after Davies *et al.* (1970).]

3.6 Radiative Flux Divergence

The concept of energy flux convergence or divergence and its relation to cooling or warming of a layer of the atmosphere of submedium has been explained in Chapter 1. Here, we discuss the significance of net radiative flux convergence or divergence to warming or cooling in the lowest layer of the atmosphere, namely, the PBL.

The rate of warming or cooling of a layer of air due to change of net radiation with height can be calculated from the principle of conservation of energy. Considering a thin layer between the levels z and $z + \Delta z$, where the net radiative fluxes are $R_N(z)$ and $R_N\,(z + \Delta z)$, as shown in Figure 3.6, we get

$$\rho c_p \Delta z(\partial T/\partial t)_R = R_N(z + \Delta z) - R_N(z) = (\partial R_N/\partial z)\Delta z$$

or

$$(\partial T/\partial t)_R = (1/\rho c_p)(\partial R_N/\partial z) \tag{3.17}$$

where $(\partial T/\partial t)_R$ is the rate of change of temperature due to radiation and $\partial R_N/\partial z$ represents the convergence or divergence of net radiation. Radiative flux convergence occurs when R_N increases with height $(\partial R_N/\partial z > 0)$ and divergence occurs when $\partial R_N/\partial z < 0$. The former leads to warming and the latter to cooling of the air. Warming or cooling of air due to radiative flux convergence or divergence is only a part of the total warming or cooling. Contributions of warm or cold air advection and turbulent exchange of sensible heat will be discussed later in Chapter 5.

In the daytime, during clear skies, net radiation is dominated by the net shortwave radiation (R_S), which usually does not vary with height in the PBL.

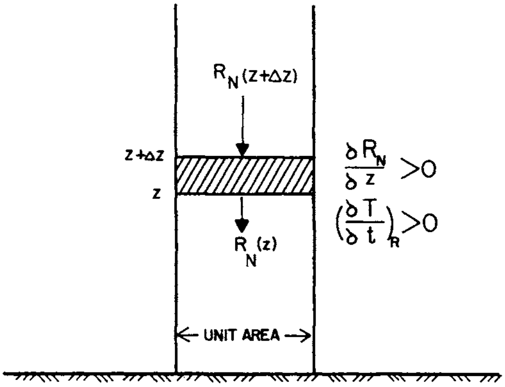

Figure 3.6 Schematic of radiative flux convergence or divergence in the lower atmosphere.

Therefore, changes in air temperature with time due to radiation are insignificant or negligible; the observed daytime warming of the PBL is largely due to the convergence of sensible heat flux. This may not be true, however, in the presence of fog and clouds in the PBL, when radiative flux convergence or divergence may also become important.

At night, the net radiation is entirely due to the net longwave radiation (R_L). Both the terrestrial and atmospheric radiations and, hence, R_L generally vary with height, because the concentrations of water vapor, CO_2, and other gases that absorb and emit longwave radiation vary strongly with height in the PBL. Frequently, there is significant radiative flux divergence (which implies that radiative heat loss increases with height) within the PBL, especially during clear and calm nights. A significant part of cooling in the PBL may be due to radiation, while the remainder is due to the divergence of sensible heat flux. In many theoretical studies of the nocturnal stable boundary layer (SBL), radiative flux divergence has erroneously been ignored; its importance in the determination of thermodynamic structure of the SBL has recently been pointed out (Garratt and Brost, 1981).

In the presence of strong radiative flux divergence or convergence, radiation measurements at some height above the surface may not be representative of their surface values. Corrections for the flux divergence need to be applied to such measurements.

Example Problem 2

The following measurements were made at night from a meteorological tower in the middle of a large farm:

Net radiation at the 5 m level = -75 W m^{-2}.

Net radiation at the 100 m level = -150 W m^{-2}.
Sensible heat flux at the surface = -45 W m^{-2}.
Planetary boundary layer height = 90 m.

Calculate the average rate of cooling in the PBL due to the divergence of (a) net radiation and (b) sensible heat flux.

Solution

(a) The rate of change of air temperature due to radiative flux divergence

$$\left(\frac{\partial T}{\partial t}\right)_R = \frac{1}{\rho c_p}\frac{\partial R_N}{\partial z} = \frac{1}{1200}\left(\frac{-150+75}{100-5}\right)$$

$$= -6.58 \times 10^{-4}\ \mathrm{K\ s^{-1}}$$

$$= -2.37\ \mathrm{K\ h^{-1}}$$

(b) From the definition of the PBL, the sensible heat flux at the top of the PBL (90 m) must be zero. The rate of change of air temperature due to sensible heat flux divergence is given by

$$\left(\frac{\partial T}{\partial t}\right)_H = -\frac{1}{\rho c_p}\frac{\partial H}{\partial z} = -\frac{1}{1200}\left(\frac{0+45}{90}\right)$$

$$= -4.17 \times 10^{-4}\ \mathrm{K\ s^{-1}}$$

$$= -1.50\ \mathrm{K\ h^{-1}}$$

Thus, the total rate of cooling in the PBL was 3.87 K h^{-1}.

3.7 Applications

The radiation balance at or near the earth's surface has the following applications:

- Determining climate near the ground.
- Determining radiative properties, such as albedo and emissivity of the surface.
- Determining net radiation, which is an important component of the surface energy budget.
- Defining and determining the apparent (equivalent blackbody) surface temperature for a complex surface.

- Parameterizing the surface heat fluxes to soil and air in terms of net radiation.
- Determining radiative cooling or warming of the PBL.

These and other applications are particularly important in determining the microclimates of various natural and urban areas.

Problems and Exercises

1. The following measurements were made over a short grass surface on a winter night when no evaporation or condensation occurred:
Outgoing longwave radiation from the surface = 365 W m^{-2}.
Incoming longwave radiation from the atmosphere = 295 W m^{-2}.
Ground heat flux from the soil = 45 W m^{-2}.
(a) Calculate the apparent (equivalent blackbody) temperature of the surface.
(b) Calculate the actual surface temperature if surface emissivity is 0.92.
(c) Estimate the sensible heat flux to or from air.

2.
(a) Estimate the combined sensible and latent heat fluxes from the surface to the atmosphere, given the following observations:
Incoming shortwave radiation = 800 W m^{-2}.
Heat flux to the submedium = 150 W m^{-2}.
Albedo of the surface = 0.35.
(b) What would be the result if the surface albedo were to drop to 0.07 after irrigation?

3. The following measurements or estimates were made of the radiative fluxes over a short grass surface during a clear sunny day:
Incoming shortwave radiation = 675 W m^{-2}.
Incoming longwave radiation = 390 W m^{-2}.
Ground surface temperature = 35°C.
Albedo of the surface = 0.20.
Emissivity of the surface = 0.92.
(a) From the radiation balance equation, calculate the net radiation at the surface.
(b) What would be the net radiation after the surface is thoroughly watered so that its albedo drops to 0.10 and its effective surface temperature reduces to 25°C?
(c) Qualitatively discuss the effect of watering on other energy fluxes to or from the surface.

4. Show that the variation of about 28% in terrestrial radiation in Figure 3.4 is consistent with the observed range of 10–30°C in surface temperatures.

5. Explain the nature and causes of depletion of the solar radiation in passing through the atmosphere.

6. Discuss the consequences of the absorption of longwave radiation by atmospheric gases and the so-called greenhouse effect.

7. Discuss the merits of the proposition that net radiation R_N can be deduced from measurements of solar radiation $R_{S\downarrow}$ during the daylight hours, using the empirical expression

$$R_N = AR_{S\downarrow} + B$$

where A and B are constants. On what factors are A and B expected to depend?

8.
(a) Discuss the importance and consequences of the radiative flux divergence at night above a grass surface.
(b) If the net longwave radiation fluxes at 1 and 10 m above the surface are -135 and -150 W m^{-2}, respectively, calculate the rate of cooling or warming in °C h^{-1} due to radiation alone.

9. The following measurements were made at night from a meteorological tower:

Net radiation at the 2 m level = -125 W m^{-2}.
Net radiation at the 100 m level = -165 W m^{-2}.
Sensible heat flux at the surface = -75 W m^{-2}.
Planetary boundary layer height = 80 m.

Calculate the average rate of cooling in the PBL due to the following:
(a) the radiative flux divergence;
(b) the sensible heat flux divergence.