Radiation and Heat

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Contents

Abstract

Within this chapter different aspects of radiation and heat are discussed for the tropics. On a global scale the amount of annual solar radiation reaching the Earth's surface is determined by atmospheric and astronomical factors and the average energy budget of the Earth could be explained by the global radiation budget. Within the tropics the net radiation varies clearly across the different climatic regions. In the humid tropics the amount of net radiation shows an almost uniform annual pattern with two maxima per annum. In contrast to this, the horizontal course of the isopleths in a radiation isopleths diagram for the marginal tropics still indicates a marked diurnal cycle. At the Earth's surface, the total incoming radiation is transformed into different heat fluxes which are either directed towards the ground or to the atmosphere. In the tropical rainforests, only 10 % of the incoming radiation reaches the ground, and only a weak flow of sensible and latent heat exists from the forest canopy towards the forest ground and vice versa.

Keywords

Tropics • Radiation and Heat • Global Radiation Budget • Heat Budget • Sensible and Latent Heat

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Global Radiation

About 99.9 % of all energy in the Earth's atmosphere is received from the sun and only a small fraction is produced from the Earth itself (e.g., by volcanism, by radioactive decay, or by the burning of organic materials). This energy is transported from the sun to the Earth in the form of electromagnetic radiation which is commonly known as solar radiation. Depending on its wavelength, it is referred to as shortwave radiation (wavelength is between 0.15 and 3.0 μm) or as longwave or infrared radiation (wavelengths between 3.0 and 100 μm). Temporal and spatial changes in the total amount of solar energy arriving on a horizontal surface at the top of the Earth's atmosphere are mainly influenced by the annual changing distance from the Earth to the sun and the inclination of the Earth's rotation axis. Over the course of the year, the average amount of insolation received at the top of the Earth's atmosphere at any point in time is about 1,367 W/m^2 (the so-called solar constant). The total sum of radiation received at any place on Earth depends on the duration and the intensity of insolation (the solar radiation which is received by a surface). The duration of insolation (which is specified by the length of day) is controlled by the Earth's rotation around its own axis. During the annual rotation of the Earth around the sun, the observable position of the sun moves south and northward of the equator, and the solar declination varies between 23°27′ N at the summer solstice in the northern hemisphere and 23°27′ S at the winter solstice. At the equinoxes, the sun is located vertical over the equator, and solar declination is zero. At the equator, all days of the year are 12 h and 7 min long. Astronomically, the sunshine duration would be 12 h precisely, but it takes 3.5 min for the upper half of the sun to disappear under the horizon at sunset and another 3.5 min before the lower half of the sun is above the horizon at sunrise. With increasing distance from the equator, the difference between the duration of the shortest and the longest day of the year increases by about 7 min per degree latitude. It is, for example, about 71 min at 10° and about 146 min at 20° latitude (Fig. 1) (Hay 1987; Lauer 1993; McGregor and Nieuwolt 1998; McArthur 2005). The intensity of insolation at the Earth's surface is a function of the geometry of the Earth's orbit around the sun and the scattering and absorption properties of the Earth's atmosphere. The Earth's orbit around the sun is elliptical, and the distance between the Earth and the sun varies from $1.47*10^8$ km (reached around the 3rd of January, perihelion) to $1.52*10^8$ km (reached around the 4th of July, aphelion) in the course of the year (Fig. 2). Thus, the amount of solar radiation received at the top of the atmosphere is about 7 % higher in January than in July. Nevertheless, the surplus of solar energy on the southern hemisphere in January is completely compensated by the global atmospheric circulation and the concentration of landmasses on the northern hemisphere. More important is the position of the sun in the sky above the horizon: a high position of the sun causes more intense insolation than a low position of the sun. One of the main reasons for this is that incoming rays from a high sun are spread over a smaller surface than rays from a low sun. Simultaneously, rays from a high sun have a shorter way through the atmosphere to the surface with lower absorption and scattering processes caused by aerosols and dust particles.

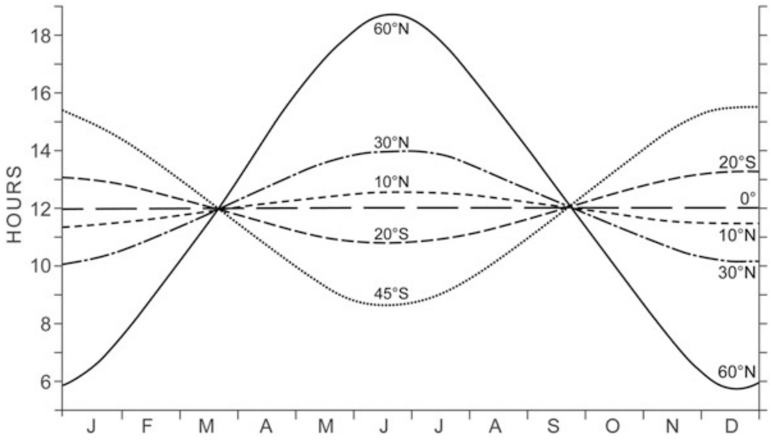

Fig. 1 Length of day during the course of the year at various latitudes (After McGregor and Nieuwolt 1998)

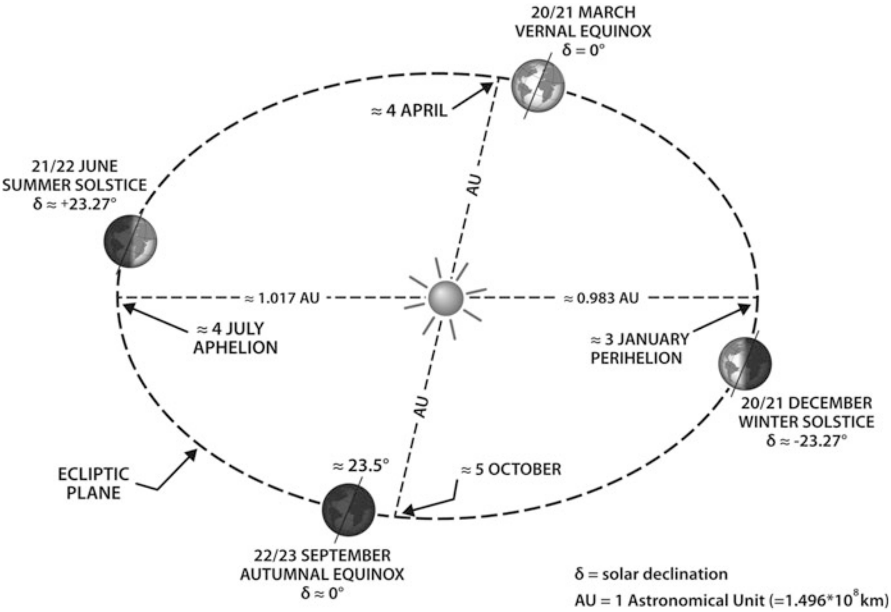

Fig. 2 Characteristics of the Earth's orbit around the sun (After McArthur 2005)

In addition, the albedo of a surface decreases with higher elevations of the sun. This effect is climatologically important because it is stronger over water and more than 66 % of the Earth's surface is occupied by oceans and seas (McGregor and Nieuwolt 1998). The amount of annual global solar radiation reaching the Earth's surface is determined by atmospheric and astronomical factors. On a global scale,

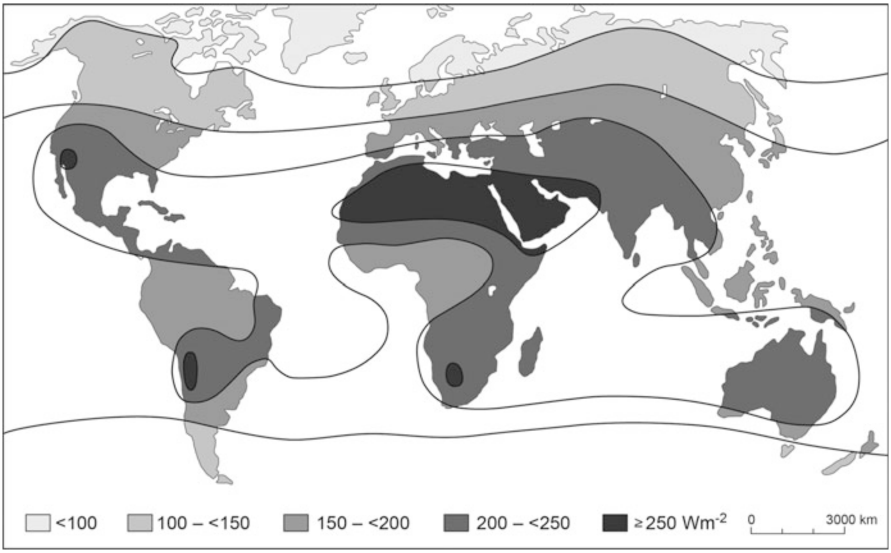

Fig. 3 Mean annual global solar radiation (After Budyko et al. 1962)

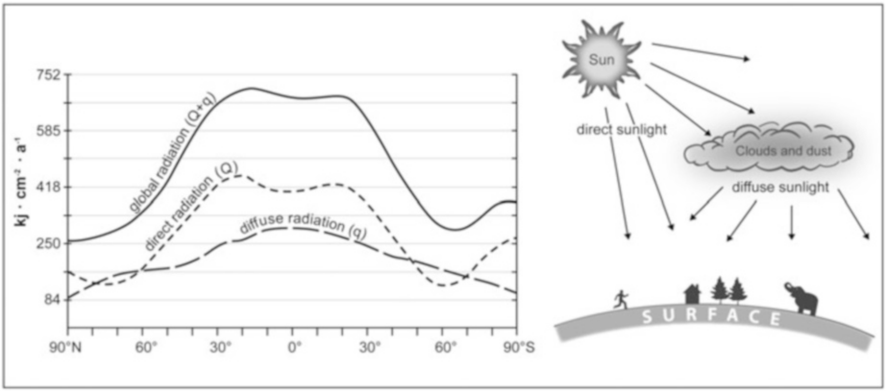

Fig. 4 Latitudinal profile of annual sum of global radiation (kJ2/year) and its components (After Sellers 1965)

the highest values of surface insolation appear between the tropics of Cancer and Capricorn. Here, areas of maximal global radiation are found over the tropical deserts and oceans where no cloud cover exists for most time of the year and the annual amount of direct radiation is maximal (Figs. 3 and 4). In comparison to the arid tropics and subtropics, the situation over the tropical rainforest areas is quite different. Here, annual sums of diffuse radiation are maximal whereas annual sums of direct radiation are reduced. This could be traced back to the high water vapor content in the atmosphere and the daily cloud formations which reduce the annual sum of direct radiation. Outside the tropics, annual means of solar radiation generally decrease towards the poles. This is mainly due to the geometry of the Earth's orbit around the sun, the inclination of the Earth axis, and the scattering and absorption properties of the Earth's atmosphere which are affecting the duration and intensity of insolation. On the average, the Earth's surface receives only 46 % of the solar radiation at the top of the atmosphere, and the tropical zone shows a clear surplus of absorbed radiation in comparison to the middle and higher latitudes (Fig. 4). However, distinct variations in the amount of energy received on the Earth's surface of the tropics could be observed. In the humid tropics, the ratio of direct (Q) to the diffuse radiation (q) is approximately 3:2. The large proportion of diffused radiation results from the frequent occurrence of clouds of great vertical thickness and a high content of water vapor and aerosols in the inner tropical atmosphere. In the desert areas of the marginal-tropical and subtropical arid regions, the ratio Q:q is approximately 4:1, which means that these zones receive slightly higher values of global radiation (Lauer 1993).

The Global Radiation Budget

The average energy budget of the Earth could be explained by Fig. 5. The incident solar radiation averaged over the globe is approximately 342 Wm^{-2}. It is calculated by the formula:

$$\text{Solar constant}^{*}\pi r^2/4\pi r^2 \tag{1}$$

where

r = radius of the Earth

$4\pi r^2$ = surface area of a sphere

This value is regarded as 100 % extraterrestrial radiation arriving at the outer edge of the atmosphere (Fig. 5). 3 % of the extraterrestrial radiation is absorbed by ozone mainly in the stratosphere, 17 % by water vapor and aerosols in the troposphere, and 3 % by water droplets in the clouds. Another 20 % was reflected back to space by clouds, 8 % by the Earth's surface, and 3 % by atmospheric scattering. Thus, in total, 31 % of the radiation energy is reflected back by the planetary albedo. The remaining 49 % reaches the Earth's surface as direct (Q = 28 %) or diffuse radiation (q = 21 %). As shown by Fig. 5, the pattern of outgoing longwave terrestrial radiation and heat fluxes is pretty different. At the Earth's surface, the absorbed insolation is temporarily stored in the form of sensible heat. However, the Earth's surface is not getting hotter in the long term, because the sensible heat energy is continually lost in the form of longwave terrestrial heat radiation into the overlaying atmosphere. In accordance with the mean surface temperature of approximately 288 K (15 °C), the Earth emits 114 % ground radiation. This is only possible since a high amount of the terrestrial radiation is reabsorbed and reradiated by the atmosphere. Only 12 % of the terrestrial radiation escapes directly

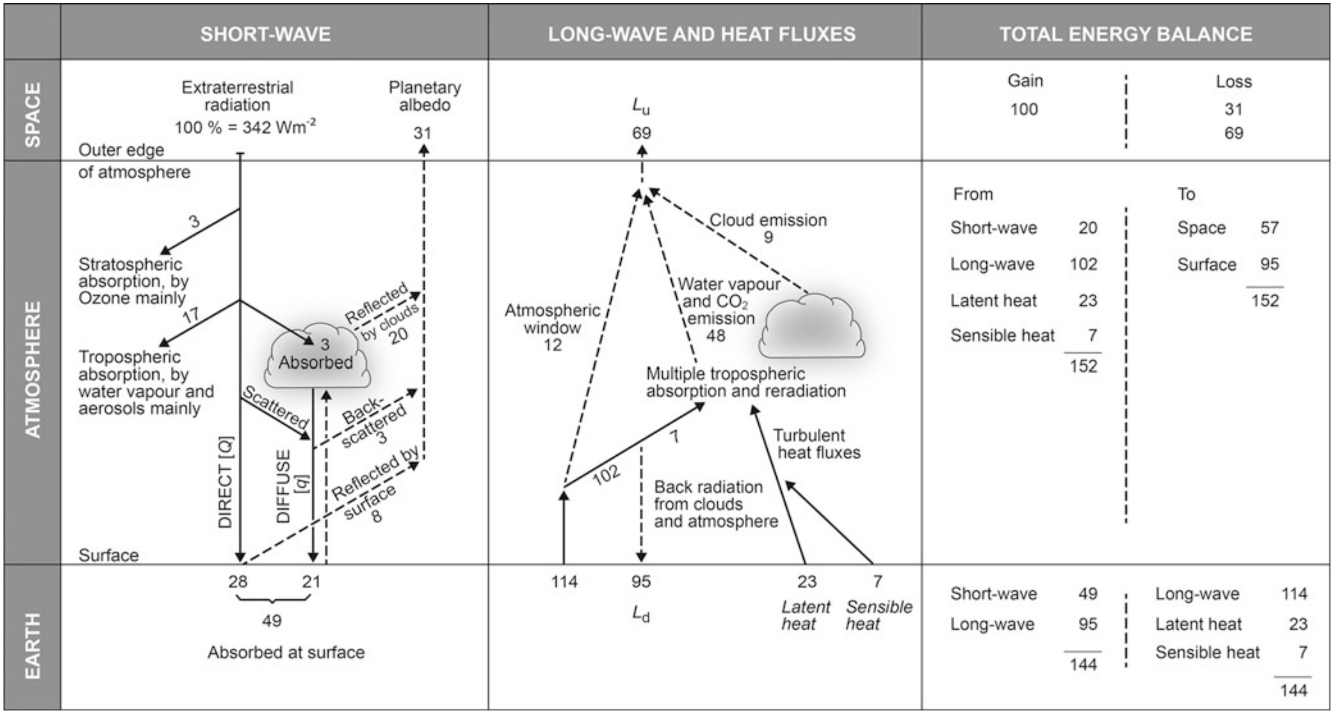

Fig. 5 The Earth's average energy budget. *Solid lines* indicate energy gains and *broken lines* energy losses by the atmosphere and surface in the *left panel* and the troposphere in the *middle panel* (After Barry and Chorley 2003)

to space through the atmospheric window, while another 7 % is absorbed and reradiated in the troposphere. These exchanges represent a time-averaged state for the whole globe, whereas the extraterrestrial radiation affects only the sunlit hemisphere. In return, no solar radiation is received by the nighttime hemisphere, but infrared exchanges continue during nighttime due to the accumulated heat in the ground. The atmosphere itself radiates 48 % from the emission by atmospheric water vapor and CO_2 and 9 % from cloud emission to space. Thus, 69 % (L_u) is lost to space all in all. Simultaneously, a great portion (95 %) of the emitted longwave terrestrial radiation from the surface is returned to the Earth's surface as back radiation. This concludes that $L_u + L_d = L_n$ is negative. Consequently, there is a difference in energy levels between the Earth's surface and the atmosphere and hence an impulse for the vertical transmission of sensible and latent heat. The energy transfers can be described by the formula:

$$R_n = (Q + q)^*(1 - a) + L_n \tag{2}$$

where:

R_n = net radiation
$(Q + q)$ = global solar radiation
a = albedo
L_n = net longwave radiation

At the surface, the net longwave radiation R_n is 30 %. This surplus is transferred to the atmosphere by sensible (7 %) and latent heat (23 %) and balances the negative energy budget of the atmosphere. However, numerous uncertainties are still to be resolved in these estimates. The surface shortwave and longwave radiation budgets show an uncertainty of approximately 20 Wm^{-2} and the turbulent heat fluxes of about 10 Wm^{-2} (Barry and Chorley 2003; Hay 1987; Lauer 1993; McGregor and Nieuwolt 1998; Mills 2005). As shown by Fig. 6, the net radiation R varies clearly across the different climatic regions of the tropics. Within the humid tropics (diagram for Yangambi, 0° 49′N 24° 29′E, 485 m a.s.l.), the amount of net radiation shows an almost uniform annual pattern with two maxima per annum. Both maxima depend on the course of the sun although they do not necessarily correspond with the highest position of the sun itself. During the sun's zenithal months, the maxima can be repressed by heavy cloud cover and adjusted to the following months with lower cloud amounts. During nighttime, the amount of outgoing net radiation is very low all over the year. This could be traced back to the relatively high amount of clouds and the high humidity which ensure the low variation of the nightly radiation loss during the whole year. Since the warming of the ground is reduced by the constant high soil moisture and by the high loss of energy for evaporation, the amount of longwave outgoing radiation from the hydrosphere and lithosphere is relatively small in the tropics. In contrast to this, the horizontal course of the isopleths in the diagram for the marginal tropics (diagram for Poona, 18° 32′N 73° 51′E, 555 m a.s.l.) still indicates a marked diurnal cycle. However, also a seasonal differentiation can be noticed. In the dry season, during the night, energy losses of −4 and −7 cal cm^{-2} h^{-1} are measured, while in

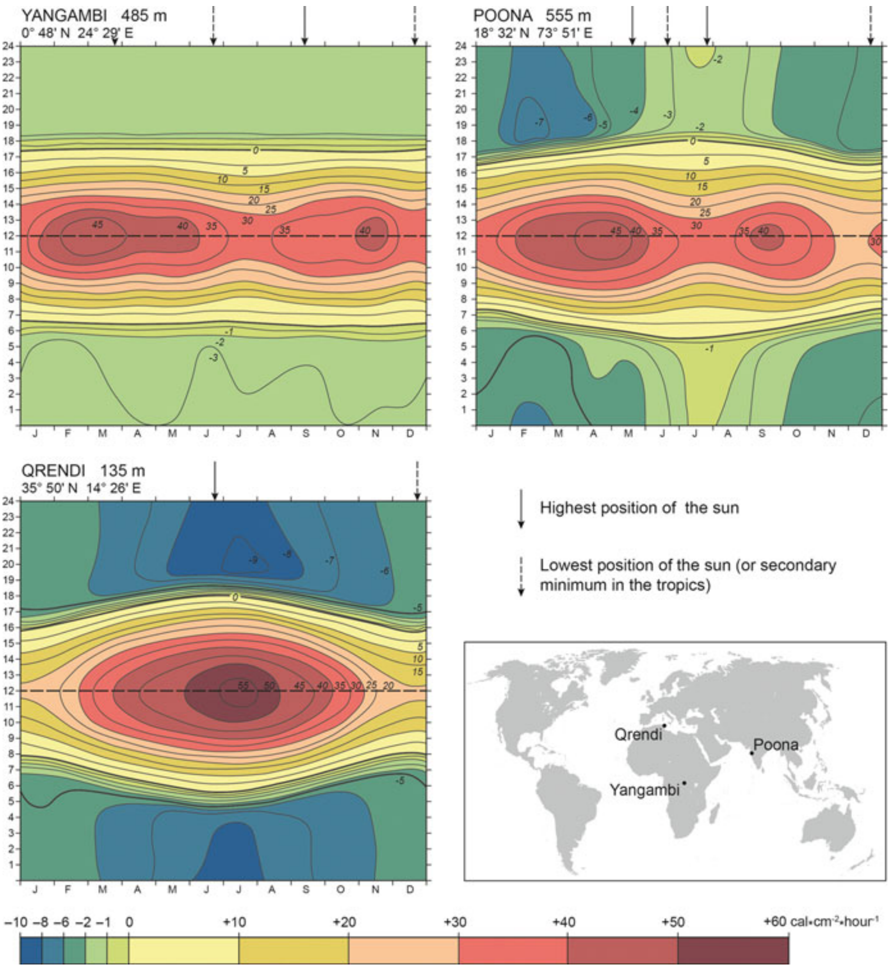

Fig. 6 Radiation isopleth diagrams of selected stations in the tropics (After Kessler 1973)

the rainy season, these value to -2 cal cm^{-2} h^{-1} due to the high water vapor content of the air. In addition to this, there is the increased amount of energy received at midday during the pre-monsoon time and the rainy season in the summer (45 and 35 cal cm^{-2} h^{-1}). In the subtropics, the annual course of net radiation is already more clearly defined (diagram for Qrendi, 35° 50′N 14° 26′E, 135 m a.s.l.). Here, only one radiation maximum in midsummer is observable and the amounts of summer radiation are distinctly higher than in the humid tropics, where the cloud cover limits the incoming radiation. The radiation balance about noon varies between +20 cal cm^{-2} h^{-1} in winter and +55 cal cm^{-2} h^{-1} in summer. This is also the time when the highest nocturnal radiation deficits are observed during summer with the lowest nocturnal cloud cover (Lauer 1993).

Heat Budget, Sensible and Latent Heat

At the Earth's surface, the total incoming radiation (=R_{sun} + R_{atm} + R_{env}) is transformed into different heat fluxes which are either directed towards the ground or to the atmosphere (Fig. 7). These processes are difficult to access within the humid tropics where the longwaved atmospheric back radiation is an important energy source. In the tropical rainforests, only 10 % of the incoming radiation

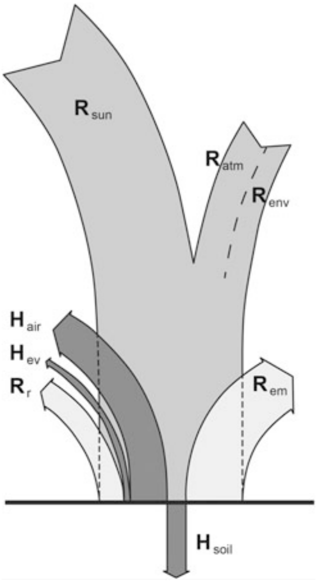

Fig. 7 Energy balance of a dry surface with strong solar radiation. R_{sun} = Q + q, R_{atm} = atmospheric longwave radiation, R_{env} = longwave radiation from the surroundings, R_{em} = amount of heat emitted by radiation, R_r = reflected solar radiation, H_{air} = heat transferred to the air through convection, H_{ev} = evaporation heat, H_{soil} = heat transferred to the soil (After Stoutjesdijk and Barkman 1992 and Richter 1997)

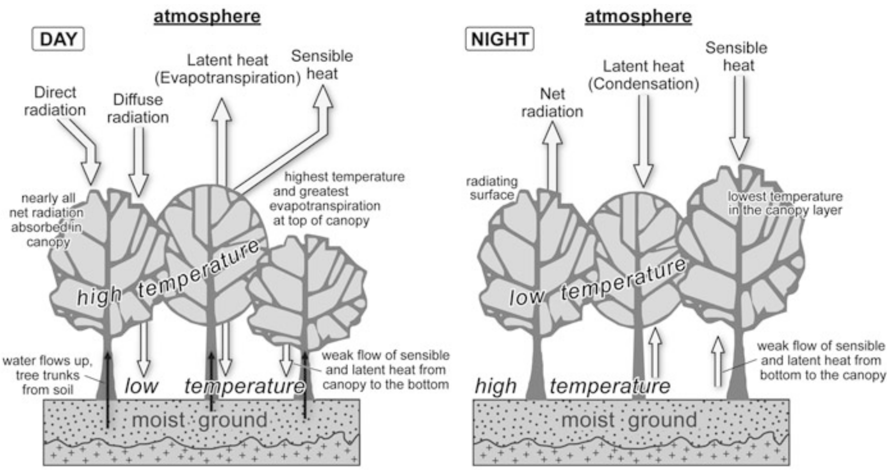

Fig. 8 Energy exchange within a forest during day and night (After Lauer 1989)

reaches the ground, and only a weak flow of sensible and latent heat exists from the forest canopy towards the forest ground and vice versa. Consequently, the transformation of incoming radiation mainly occurs in the forest canopy in the form of evaporation processes and sensible heat fluxes (Fig. 8). During the day, the received energy is primarily released into the atmosphere by evapotranspiration processes. While the air temperature in the forest stays relatively constant, the water vapor pressure increases clearly. This is characteristically for tropical rainforests where the continuous release of water vapor – as it is also the case over the sea – happens without the disturbance of the atmospheric layers. During nighttime, the cooling of the air leads to heated release in terms of condensation processes. Besides, sensible heat is absorbed from the atmosphere above the canopy as well as to a small extent from the tree base near the ground (Lauer 1993).

References

Barry R, Chorley R (2003) Atmosphere, weather and climate. Routledge, London

Budyko MF, Nayefimova NA, Aubenok LI, Strokhina LA (1962) The heat balance of the surface on the earth. Soviet Geogr 3(5):3–16

Hay J (1987) Solar radiation. In: Fairbridge (deceased) R, Oliver (deceased) J (ed) Climatology. Earth sciences series. Springer, Berlin/Heidelberg

Kessler A (1973) Zur Klimatologie der Strahlungsbilanz auf der Erdoberfläche. Erdkunde 17:1–10

Lauer W (1989) Climate and weather. In: Lieht H, Werger MJA (eds) Tropical rain forest ecosystem, Ecosystems of the world. Elsevier, Amsterdam, pp 7–53

Lauer W (1993) Climatology. In: Pancel L (ed) Tropical forestry handbook, vol 1. Springer, Berlin/Heidelberg

McArthur L (2005) Solar radiation. In: Oliver (deceased) J (ed) Encyclopedia of world climatology. Earth sciences series. Springer, Berlin/Heidelberg

McGregor GR, Nieuwolt S (1998) Tropical Climatology. Wiley, Chichester

Mills G (2005) Radiation climatology. In: Oliver (deceased) J (ed) Encyclopedia of world climatology. Earth sciences series. Springer, Berlin/Heidelberg

Richter M (1997) Allgemeine Pflanzengeographie Teil 1. Teubners Taschenbücher der Geographie, Stuttgart

Sellers WD (1965) Physical climatology. University of Chicago Press, Chicago

Stoutjesdijk PH, Barkman JJ (1992) Microclimate, vegetation and fauna. Opulus, Knivsta