Plate 4.2 Spatio-Temporal Variations in Net Radiation 1984–1993

Introduction

Net radiation is the difference between total radiation falling upon and reflected by the Earth's surface (incoming and outgoing radiation):

$$N = G - R + L_{in} - L_{out} \text{ or } N = G \bullet (1 - \alpha) + L_{in} - L_{out}$$

where N = net radiation, G = global radiation, R = reflected short-wave radiation, α = albedo, L_{in} = incoming long-wave radiation and L_{out} = out-going long-wave radiation.

Components of the radiation

Global radiation is the portion of radiation which is generated directly or indirectly by the Sun. The temperature of the Sun's surface is 5700 °C, which means that solar radiation occurs practically only in the shorter wavelengths. Outside the Earth's atmosphere the energy flow rate is 1366.5 W/m^2. Since at any one time only half the Earth's surface receives sunlight and since the Earth is spherical, mean solar radiation is a quarter of the solar constant, i.e. 342 W/m^2 (fig. 1). Of this, 84 W/m^2 falls directly on to the Earth's surface. This portion is referred to as direct radiation. The remaining 258 W/m^2 reaches the atmosphere, which absorbs 98 W/m^2. The rest is reflected by clouds or diffused by molecules (Rayleigh scattering) or aerosols (Mie scattering). Thus 85 W/m^2 reaches the surface of the Earth as diffuse radiation and 75 W/m^2 is reflected back into space. The sum of direct radiation and diffuse radiation is referred to as global radiation at the Earth's surface. Mean global radiation across the planet is 169 W/m^2. Of this the Earth's surface absorbs 142 W/m^2 and 27 W/m^2 is reflected. The ratio between short-wave radiation reflected from the Earth's surface (R) and global radiation (G) is referred to as the albedo ($\alpha = R/G$). The mean albedo for the Earth's surface is around 0.16. Almost all short-wave radiation reflected from the Earth's surface goes back into space. In the overall «atmosphere–surface» system, the atmosphere absorbs 98 W/m^2 in the shorter wavelengths and 102 W/m^2 is reflected by the Earth back into space. Radiation absorbed by the Earth's surface serves to heat the atmosphere, the soil and the oceans.

Since mean temperatures on the Earth's surface vary from –60 °C to +40 °C, it emits radiation in the longer wavelengths. This long-wave outgoing radiation can be calculated using the Stefan-Boltzmann law: $L = \varepsilon \bullet \sigma \bullet T^4$; ε signifies total emissions from the Earth's surface and can be replaced by 1 with very good approximation, and σ is the Stefan-Boltzmann constant. Kirchhoff's law must also be taken into account. Since over a longer period of time the difference between surface and air temperature is negligible, the Earth's mean air temperature can be used (T; i.e. around 14 °C). This then gives a value for long-wave outgoing radiation of 385 W/m^2. If the whole of the Earth were under cloud all long-wave outgoing radiation would be absorbed by the atmosphere. If mean cloud cover around the Earth is 50 %, 168 W/m^2 is absorbed by the cloud-free atmosphere and 193 W/m^2 by the cloud cover. Thus 24 W/m^2 is reflected by the atmosphere directly back into space.

For its part, the cloud-free atmosphere constantly re-emits energy in the longer wavelengths: 111 W/m^2 back into space and 160 W/m^2 towards the Earth's surface. Under cloudy conditions the corresponding figures are 105 and 185 W/m^2. If one takes the total balance of long-wave radiation in the atmosphere it can be seen that there is an overall loss of 200 W/m^2. At the Earth's surface this loss can be quantified at 40 W/m^2.

The balance of short and long-wave radiation, i.e. net radiation, is +102 W/m^2 at the Earth's surface. In contrast, the atmosphere loses a total of 102 W/m^2. Thanks to the fact that the Earth's surface releases energy in the form of sensible and latent heat which is taken up by the atmosphere, the excess does not cause the Earth's surface to constantly heat up, and the loss does not cause the atmosphere to constantly cool down. Since the larger part of the Earth's surface is covered with water, latent heat flow (evaporation) is predominant with a value of 85 W/m^2, sensible heat flow being around 17 W/m^2. As net radiation at the Earth's surface governs sensible and latent heat flow, it is a crucial factor in relation to climatic phenomena and weather conditions. The total radiation balance of the Earth–atmosphere system is zero; the system is therefore balanced [1,3,4,5].

Basic principles of the methodology

All radiation parameters were calculated with a high resolution. Topography plays a decisive role with regard to net radiation, especially in the shorter wavelengths. Thus the floor of a narrow mountain valley receives less direct solar radiation than a plain, for example. For this reason topographical parameters such as altitude, gradient, aspect and horizon definition, which were obtained from the RIMINI digital terrain model with a raster mesh size of 250 m, are necessary to calculate net radiation. Global radiation was calculated on the basis of the global radiation values obtained by [7] from flat plains with no defined horizon. The albedo was calculated using land-use statistics for Switzerland, different types of use being summarised in six categories: water, forest, agriculture, built-up areas, rock and glaciers. The most important parameter, namely the number of days of snow cover, was taken from [2]. The temperature and humidity profiles needed for determining incoming long-wave radiation are based on the results of regular radiosonde observations made in Payerne. Incoming long-wave radiation was then calculated using the MODTRAN radiation transfer model. For outgoing long-wave radiation it was assumed that on a monthly average surface temperature equals air temperature. It was then possible to calculate outgoing long-wave radiation, after having determined air temperature for each point on the raster, using the Stefan-Boltzmann law. Details concerning methodology can be found in [8].

Spatial variability

Since the upper atmosphere contains fewer aerosols and is easier for the Sun to penetrate, global radiation rises slightly with increasing altitude. In contrast, the albedo depends very much on altitude. This is a consequence of the duration of snow cover. On average the albedo of dry snow is 0.71 and that of melting snow 0.58. Snow-free areas, however, always have an albedo of under 0.2. The number of days with snow cover, which varies according to altitude, therefore influences the albedo. Above 3000 m the snow cover is more or less permanent, so that at this altitude the albedo is practically constant with a value of 0.6. Thus values for the short-wave radiation balance decrease up to an altitude of 3000 m, although global radiation rises slightly with increasing altitude.

Long-wave radiation components fall with increasing altitude. Outgoing long-wave radiation depends solely on surface temperature, which also falls with increasing altitude. Incoming long-wave radiation is influenced by air temperature, the concentration of relevant gases and cloud cover. Effective emissions of the sky total 0.7 in lower areas with no cloud cover and fall to around 0.4 at 4000 m owing to the decreasing humidity content of the atmosphere. Under cloudy and especially foggy conditions emissivity is around 1. Since as a rule outgoing long-wave radiation is higher than incoming radiation, the long-wave radiation balance is negative and falls with increasing altitude.

The combination of the short and long-wave radiation balances results in a decrease in net radiation with increasing altitude. Above an altitude of around 3000 m mean annual net radiation is negative.

In areas with complex topography there is considerable spatial variation in net radiation (fig. 4).

Net radiation and evaporation

As already mentioned, net radiation is a decisive factor in relation to evaporation (fig. 3). In general real evaporation is around 60 to 80 % of the maximum potential evaporation calculated from net radiation. Evaporation for a given period of time (for example 1 month) can be calculated using the following formula:

$$\text{Maximum potential evaporation } [kg/m^2 \text{ or mm}] = \frac{\text{Net radiation } [W/m^2 \text{ or } J/s \bullet m^2] \bullet \text{time [s]}}{\text{Latent heat of water } [2.256 \bullet 10^6 \text{ J/kg}]}$$

Over areas of water real and maximum possible evaporation are practically identical.

In the high Alps the mean annual net radiation is negative. Although evaporation is also much lower at high altitudes than in the lowlands, it does not fall as low as zero. Despite the overall negative radiation balance there are times when net radiation is positive, when evaporation takes place.

Net radiation in Switzerland

Over Switzerland as a whole, mean annual net radiation is 44 W/m^2. The highest rates of over 60 W/m^2 occur in lower, south-facing areas of the Alps, in particular in the Rhone valley between Martigny and Visp, in the Rhine valley between Disentis and Chur, in the areas around Tiefen-castel and Zernez, as well as in the Bergell (Bregaglia) and the Puschlav (Poschiavo), on the south-facing slopes between Locarno and Bellinzona and in the Leventina valley. The lowest values occur on north-facing slopes in the high Valais Alps.

There is a clear seasonal pattern in net radiation. In January net radiation in Switzerland is negative (–20 W/m^2), while it is positive in the other months indicated: April 59 W/m^2, July 120 W/m^2 and October 19 W/m^2. It is surprising to note that values for the Tessin are generally not much higher than those for the Central Lowlands on the northern side of the Alps. On the one hand, the cloud cover south of the Alps is on average less [6] and the long-wave radiation balance therefore more negative than in the Central Lowlands; in addition global radiation is reduced through a higher horizon. On the other hand the air currents that flow from the Po plain into the Tessin, which carry more aerosols and therefore intensify scattering, could be a further factor contributing to a reduction in net radiation.

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