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RADIATION (SOLAR)

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Introduction

The Sun as an average star is a typical main-sequence dwarf of spectral class G-2. Its radius is 6.960×10^8 m. The mean distance between the Sun and the Earth is 1.496×10^{11} m and is known as the astronomical unit (AU). Solar radiation is the electromagnetic radiation emitted by the Sun. Almost all known physical and biological cycles in the Earth system are driven by the solar radiation reaching the Earth. Solar radiation is also the cause of climate change that is truly exterior to the Earth system.

Solar Spectrum and Solar Constant

The distribution of solar radiation as a function of the wavelength is called the solar spectrum, which consists of a continuous emission with some superimposed line structures. The Sun's total radiation output is approximately equivalent to that of a blackbody at 5776 K. The solar radiation in the visible and infrared spectrum fits closely with the blackbody emission at this temperature. However, the ultraviolet (UV) region ($<0.4\,\mu$m) of solar radiation deviates greatly from the visible and infrared regions in terms of the equivalent blackbody temperature of the Sun. In the interval 0.1–0.4 μm, the equivalent blackbody temperature of the sun is generally less than 5776 K with a minimum of about 4500 K at about 0.16 μm. The deviations seen in the solar spectrum are a result of emission from the nonisothermal solar atmosphere.

The solar constant is the amount of solar radiation received outside the Earth's atmosphere on a surface normal to the incident radiation per unit time and per unit area at the Earth's mean distance from the Sun. The solar constant is an important value for the studies of global energy balance and climate. Reliable measurements of solar constant can be made only from space and a more than 20-year record has been obtained based on overlapping satellite observations. The analysis of satellite data suggests a solar constant of 1366 W m^{-2} with a measurement uncertainty of ± 3 W m^{-2}. Of the radiant energy emitted from the Sun, approximately 50% lies in the infrared region ($>0.7\,\mu$m), about 40% in the visible region (0.4–0.7 μm), and about 10% in the UV region ($<0.4\,\mu$m).

The solar constant is not in fact perfectly constant, but varies in relation to the solar activities. Beyond the very slow evolution of the Sun, a well-known solar activity is the sunspots, which are relatively dark regions on the surface of the Sun. The periodic change in the number of sunspots is referred to as the sunspot cycle, and takes about 11 years, the so-called 11-year cycle. The cycle of sunspot maxima having the same magnetic polarity is referred to as the 22-year cycle. The Sun also rotates on its axis once in about 27 days. Satellite observations suggest that the solar cycle variation of the solar constant is on the order of about 0.1%, which might be too small to directly cause more than barely detectable changes in the tropospheric climate. However, some indirect evidence indicates that the changes in solar constant related to sunspot activity may have been significantly larger over the last several centuries. Furthermore, solar variability is much larger (in relative terms) in the UV region, and induces considerable changes in the chemical composition, temperature, and circulation of the stratosphere, as well as in the higher reaches of the upper atmosphere.

Distribution of Solar Insolation at the Top of the Atmosphere

The solar insolation is the actual amount of solar radiation incident upon a unit horizontal surface over a specified period of time for a given locality. It depends strongly on the solar zenith angle θ_0 and also on the ratio (d/d_m) of the actual distance to the mean distance of the Earth from the Sun. The solar irradiance at the top of the atmosphere may be expressed by

$$F = S\left(\frac{d_m}{d}\right)^2 \cos\theta_0 \qquad [1]$$

where S is the solar constant. The solar zenith angle depends on the latitude, day of year, and time of day, and is given by

$$\cos\theta_0 = \sin\lambda\sin\delta + \cos\lambda\cos\delta\cos h \qquad [2]$$

where λ is the latitude, δ the solar declination, and h the hour angle. The hour angle is zero at solar noon and increases by 15° for every hour. The solar zenith angle is 90° at sunset and sunrise. Then the solar insolation for a specified period of time between t_1 and t_2 is given by

$$Q = \int_{t_1}^{t_2} F(t)\,dt \qquad [3]$$

The daily insolation at the top of the atmosphere can be determined by integration of eqn [3] over a day. For a given day of the year, the solar declination and the ratio d/d_m can be determined from standard astronomical formulas. Under present astronomical conditions, the solar declination varies from 23° 27′ on 21 June to −23° 27′ on 22 December, while $(d_m/d)^2$ ranges from 1.0343 on 3 January to 0.9674 on 5 July. The values of the daily insolation at the top of the atmosphere are presented in **Figure 1** as a function of latitude and time of year.

It is clear that the distribution of solar insolation at the top of the atmosphere depends on the Earth's elliptical orbit around the Sun through δ and d/d_m.

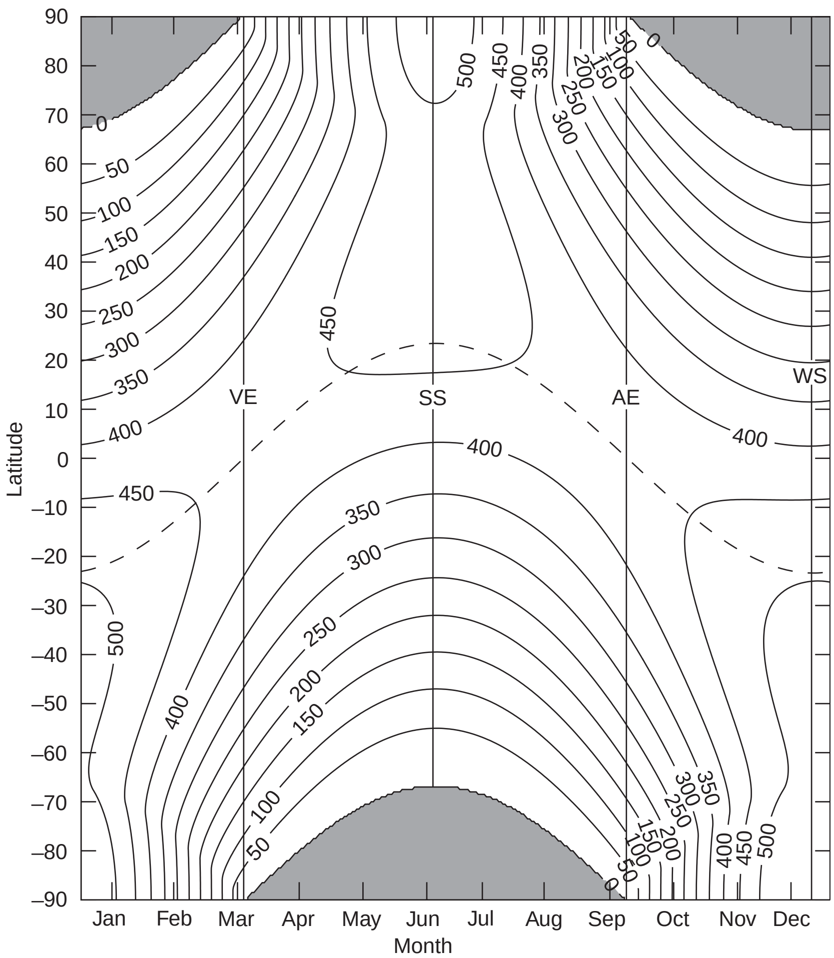

Figure 1 Daily solar insolation (86 400 J m^{-2}) at the top of the atmosphere as a function of latitude and day of year using a solar constant of 1366 W m^{-2}. The shaded areas denote zero insolation. The positions of vernal equinox (VE), summer solstice (SS), autumnal equinox (AE), and winter solstice (WS) are indicated with solid vertical lines. Solar declination is shown with a dashed line. (Adapted with permission from Liou KN (2002) *An Introduction to Atmospheric Radiation*. San Diego, CA: Academic Press.)

Because of the gravitational attraction between the Earth and other planets, the orbital parameters including the eccentricity of the orbit, the tilt of the angle, and the longitude of the perihelion vary with characteristic periods of about 100 000, 41 000, and 21 000 years, respectively. The variations of these orbital parameters of the Earth may constitute a cause for climate changes such as those experienced during the Pleistocene ice ages.

Scattering and Absorption of Solar Radiation in the Earth–Atmosphere System

Solar radiation entering the Earth's atmosphere is absorbed and scattered by atmospheric gases, aerosols, clouds, and the Earth's surface. The absorbed radiation is added directly to the heat budget, whereas the scattered radiation is partly returned to space and partly continues its path through the Earth–atmosphere system where it is subject to further scattering and absorption. The fraction of the incident solar radiation that is reflected and backscattered to space is called the albedo. We might speak of the albedo of the entire Earth or of individual surfaces with reference either to monochromatic radiation or to the total incident solar radiation. In this last sense, the albedo of the Earth as a whole is about 0.31.

Effects of Atmospheric Gases on Solar Radiation

The scattering of solar radiation by air molecules can be described by a theory developed by Rayleigh, who showed that the amount of scattering is inversely proportional to the fourth power of the wavelength, when the sizes of particles are much smaller than the wavelength of the incident radiation. We see blue sky because atmospheric molecules scatter solar radiation much more in the blue than in the red part of the spectrum. In fact, the sky is made visible through the scattering process. On the other hand, sunsets and sunrises appear reddish because the blue light in the direct light is removed by scattering during the long path through the atmosphere, leaving the remaining reddish colors of the spectrum.

Atmospheric gases also absorb solar radiation in selected wavelength bands. The UV radiation with wavelengths shorter than 0.3 μm is lethal to the biosphere. The UV radiation in the interval 0.2–0.3 μm is mainly absorbed by O_3 in the stratosphere. The small amount of radiation with wavelengths shorter than 0.2 μm is absorbed at higher levels by O_2, N_2, O, and N. The photochemical processes due to absorption of solar UV radiation involving various forms of oxygen are critical in determining the amount of ozone in the stratosphere. The absorption spectrum of O_2 between 0.2 and 0.26 μm is weak, but of significance in the formation of ozone.

In the troposphere, the absorption of solar radiation occurs in the visible and near-infrared regions, owing primarily to H_2O, CO_2, O_2, and O_3. The absorption in the visible, however, is very weak. **Figure 2** shows the depletion of solar radiation in a clear atmosphere. The top curve is the solar spectrum at the top of the Earth's atmosphere and the lower curve represents the spectrum at sea level; the shaded area gives the combined effects of scattering and absorption of solar radiation by atmospheric gases. It is evident that the depletion of solar radiation is dominated by ozone absorption in the UV, Rayleigh scattering in both UV and visible, and water vapor absorption in the near infrared.

Effects of Aerosols on Solar Radiation

Aerosols are suspensions of liquid and solid particles in the atmosphere, excluding clouds and precipitation. The aerosol particle sizes range from 10^{-4} to 10 μm, falling under the following broad categories: sulfates, black carbon, organic carbon, dust, and sea salt. Aerosol concentrations and compositions vary significantly with time and location. Visibility measurements reflect the aerosol concentration at ground level. The visual range can vary from a few meters to 200 km, depending on the proximity to sources, the strength of the sources, and atmospheric conditions.

Aerosols scatter and absorb solar radiation. Sulfate aerosols scatter primarily solar radiation and cause cooling of the Earth–atmosphere system. The increase in the reflected solar radiation at the top of the atmosphere due to such nonabsorbing aerosols is nearly identical to the reduction in solar radiation at the surface. Carbonaceous aerosols (black carbon and organics) absorb and scatter solar radiation. The presence of black carbon aerosols results in the absorption of solar radiation, which reduces the solar radiation reaching the surface. At the same time, these aerosols absorb the upward solar radiation reflected from below and reduce the solar radiation reflected to space. Therefore, the effect of black carbon aerosols opposes the cooling effect of other aerosols at the top of the atmosphere, whereas at the surface all aerosols reduce solar radiation. The changes arising from the aerosol scattering and absorption of solar radiation are referred to as their direct radiative forcing. Aerosols can also modify solar radiation through their role in cloud condensation and as ice nuclei, an effect known as aerosol indirect radiative forcing.

Aerosol particles in the atmosphere are produced both in nature and by people. A global aerosol optical

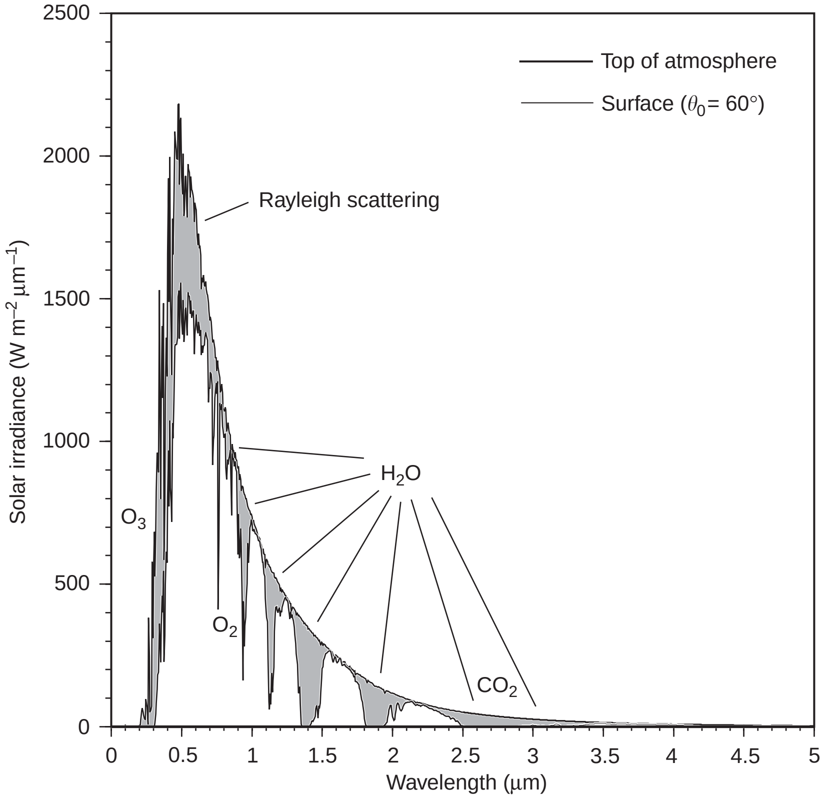

Figure 2 Solar spectrum at the top of the atmosphere and at the surface for a solar zenith angle of 60° in a clear atmosphere. Absorption and scattering regions are indicated. (Adapted with permission from Liou KN (2002) *An Introduction to Atmospheric Radiation*. San Diego, CA: Academic Press.)

depth of about 0.12 is suggested. These aerosols increase the reflected solar radiation at the top of the atmosphere by about $3\,W\,m^{-2}$ globally. Anthropogenic sources contribute significantly to the global aerosol optical depth. Global anthropogenic emissions of sulfates, organics, and black carbon even exceed natural sources. Such a large perturbation of the global aerosol loading due to human activities may significantly modify regional and global climates.

Effects of Clouds on Solar Radiation

Clouds regularly cover about 65% of the Earth, and occur in various types. Some, such as cirrus in the tropics and stratus near the coastal areas and in the Arctic are climatologically persistent. Like aerosols, clouds show substantial spatial and temporal variations.

Clouds are the most important regulator of solar radiation. By reflecting incoming solar radiation back to space, they cool the Earth–atmosphere system – the so-called cloud albedo effect. Clouds also absorb solar radiation in the near-infrared region. The cooling of the Earth–atmosphere system by the cloud albedo effect occurs primarily at the surface. The solar albedo of clouds depends substantially on cloud type and cloud form, as well as the solar zenith angle.

The most straightforward and simple diagnostic measure of the impact of clouds on solar radiation is the short-wave cloud forcing, which is defined as the difference of the net solar irradiances at the top of the atmosphere between all-sky and cloudless conditions. Here the net irradiance is the incoming solar radiation minus the reflected radiation. Satellite measurements suggest that the global short-wave cloud forcing is about $-45\,W\,m^{-2}$. Short-wave cloud forcings are maximized (about $-120\,W\,m^{-2}$) in the summer hemisphere at about latitude 60° where solar input is large and low clouds are abundant, with a secondary maximum in tropics. Note that the magnitude of short-wave cloud forcing is about ten times as large as those for a CO_2 doubling. Hence small changes in the cloud-radiative forcing fields can play a significant role as a climate feedback mechanism.

Solar Radiation at the Earth's Surface

The radiation coming directly from the Sun received at the Earth's surface is called direct solar radiation. The amount of scattered radiation coming from all other directions is called diffuse solar radiation. The sum of both components as received on a horizontal surface is called global solar radiation.

A significant fraction of the incoming solar radiation is reflected back by the surface. The surface albedo, defined as the ratio of the reflected over the incoming radiation, depends on the nature of the surface, solar zenith angle, and wavelength. For a water surface the albedo is about 0.06, whereas for snow the albedo is about 0.6–0.8. The albedo of bare sea ice is about 0.4–0.6. Since large areas of earth are covered by water, snow and sea ice, changes in the snow and sea ice cover can have a significant impact on the global albedo.

Bare land surfaces have typical surface albedo of 0.1–0.35, with the highest value for the desert sand. Albedos of most vegetation surfaces fall in the range 0.1–0.25. The albedo for green vegetation depends greatly on wavelength, reflecting strongly in the near infrared but absorbing in the ultraviolet and visible regions.

Annual Global Mean Energy Budget of Solar Radiation

The energy budget of solar radiation can be derived by combining observations and modeling studies, which show the combined effects of atmospheric gases, aerosols, clouds, and surfaces. Under the annual global mean condition, the incident solar radiation at the top of the atmosphere is $342\,\mathrm{W\,m^{-2}}$. Of this incident solar radiation, $67\,\mathrm{W\,m^{-2}}$ is absorbed during passage through the atmosphere. A total of $107\,\mathrm{W\,m^{-2}}$ is reflected back to space: $30\,\mathrm{W\,m^{-2}}$ from the surface and $77\,\mathrm{W\,m^{-2}}$ from clouds and aerosols and atmosphere. The remaining $168\,\mathrm{W\,m^{-2}}$ is absorbed at the Earth's surface. It is noted that while the incoming and reflected solar irradiances at the top of the atmosphere are constrained by satellite observations, uncertainties may exist for the partitioning of the absorbed solar radiation between the atmosphere and the surface on the global scale.

See also

Energy Balance Model, Surface. **Radiative Transfer:** Absorption and Thermal Emission; Non-local Thermodynamic Equilibrium; Scattering. **Solar Terrestrial Interactions**. **Solar Winds**.

Further Reading

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RADIATIVE TRANSFER

Contents

Absorption and Thermal Emission

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Solar and Thermal Radiation

In the final analysis, all physical or biological processes that take place on Earth, owe their existence to the absorption of radiation from the Sun. In the absence of this radiation the Earth would be cold and lifeless, at a temperature close to that of the cosmos, 2.73 K.

Solar radiation corresponds approximately to a black body emission temperature of 5780 K. Solar