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Solar and Atmospheric Radiation

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Abstract – An account is given of the solar and atmospheric radiation fluxes received at the earth's surface written in a form designed to be of use to workers interested in the practical application of solar energy. Notes on the measurement of these fluxes, calculations of the motion of the sun, a discussion of the effects of the atmosphere and clouds on solar radiation, and the estimation of atmospheric radiation from routine surface meteorological data are included.

Introduction

A clear understanding of the radiation falling on an exposed solar collector is essential for anyone who is concerned with the design, installation, or use of solar energy equipment, for this radiation is the basic energy input to the system. Here I shall describe the essential features of this radiation at an intermediate technical level in sufficient detail for use in practical engineering work. I assume that the reader already has a knowledge of the basic physics required to follow the text, but I shall explain carefully the concepts peculiar to this field. The subject matter of this article is based in part on lectures that I give in a course on solar energy at the Asian Institute of Technology.

Classification of Radiation Fluxes [1]

There are two kinds of radiation falling on a solar collector that are important in solar energy work, namely solar radiation and atmospheric radiation.

Solar radiation is the thermal electromagnetic radiation emitted by the hot incandescent surface of the sun in the ultra-violet, visible, and near infra-red regions of the spectrum.

Atmospheric radiation is the thermal electromagnetic radiation emitted by the earth's atmosphere in the far infra-red region of the spectrum.

Because of the relationship between the wavelengths of visible light and infra-red radiation, solar radiation is often called *short-wave radiation,* and atmospheric radiation is often called *long-wave radiation.* Figure 1 shows the regions of the electromagnetic spectrum occupied by the major portions of short-wave and long-wave radiation. It is of great practical importance that these two regions do not overlap.

In addition to solar radiation and atmospheric radiation, a collector may receive radiation from objects on the ground if its collecting area is exposed to them. This will contain long-wave radiation emitted by the object itself, and short-wave radiation if sunlight is reflected from it. Such radiation will, however, be ignored in this article since it is not usually important in practice.

The solar radiation reaching the surface of the earth may be divided into two components: *direct solar radiation* coming directly from the sun's disk, and *diffuse solar radiation* coming from the sky with the exception of the sun's disk. It is the direct solar

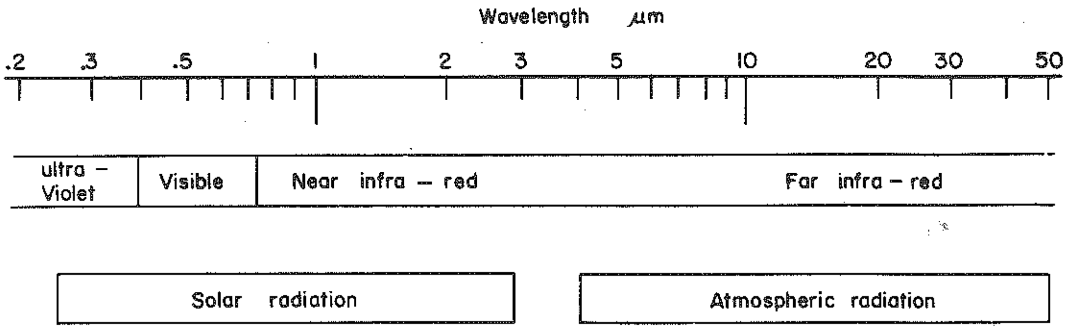


Fig. 1. Solar and atmospheric radiation in the electromagnetic spectrum.

radiation that throws sharp shadows and can be focussed by optical systems. Diffuse solar radiation throws no sharp shadows and cannot be focussed. These two kinds of solar radiation are shown in Figure 2.

Radiation fluxes are measured in terms of the quantity of energy flowing through unit area in unit time. The standard measure of direct solar radiation is the flux from the sun's disk incident on a surface perpendicular to the solar beam. In bright sunlight its value is about 0.9 kW/m^2. The standard measure of diffuse solar radiation is the diffuse short-wave flux incident on a horizontal surface facing upwards. Its value depends on weather conditions. Under a clear sky it is typically 0.1 kW/m^2, but under a cloudy sky it may vary from 0.3 kW/m^2 to 0.6 kW/m^2.

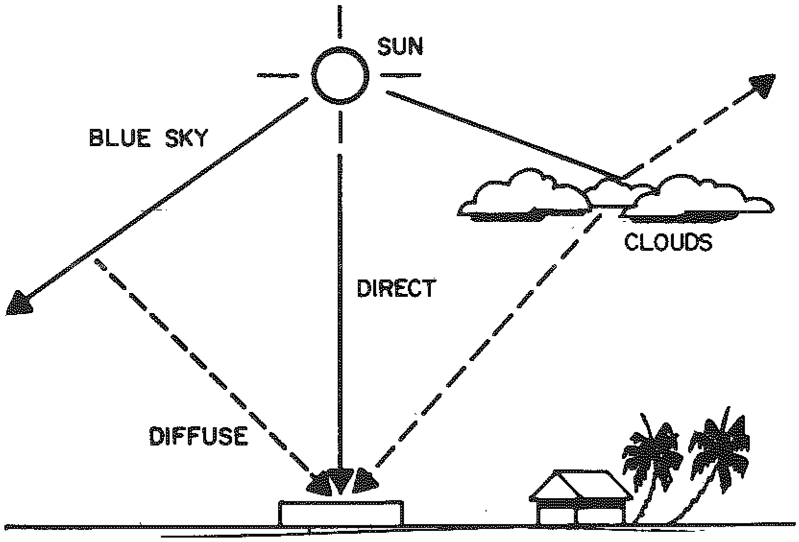


Fig. 2. Direct and diffuse solar radiation.

The sum of the direct and diffuse solar radiation fluxes incident on a horizontal surface facing upwards is called *global solar radiation.* If I is the direct solar radiation flux falling on a surface perpendicular to the solar beam, θ is the angle of incidence of the solar beam on a horizontal surface, and D is the diffuse solar radiation falling on the horizontal surface, then the global solar radiation K is given by

$$K = I \cos\theta + D.$$

In the case of atmospheric radiation all the radiation is diffuse, and so there is only one standard measure of it, namely the long-wave radiation flux incident on a horizonal surface facing upwards. This is called *downward atmospheric radiation.*

The sum of the short-wave and long-wave radiation fluxes incident on a horizontal surface facing upwards is called the *total downward radiation.* If L is the downward atmospheric radiation, then the total downward radiation Q is given by

$$Q = K + L.$$

Measurement of Solar Radiation [1]

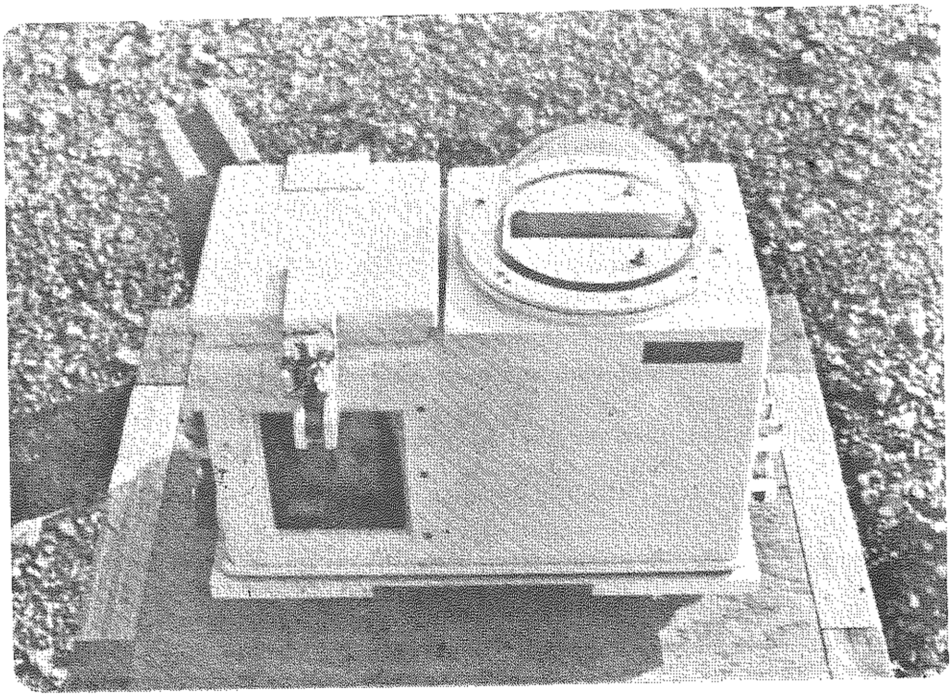
An instrument for measuring radiation is called a *radiometer.* It is usually desirable for radiometers to respond equally to equal amounts of energy at all wavelengths over the wavelength range of the radiation to be measured. Most radiometers therefore work by using a thermopile to measure the temperature rise of a sensitive element whose receiving surface is painted dull black. Instruments for measuring solar radiation that use a photovoltaic cell as the sensitive element have a non-uniform spectral response.

A *pyrheliometer* is an instrument for measuring direct solar radiation at normal incidence. It is so designed that it measures only the radiation from the sun's disk (which has an apparent diameter of ½°) and from a narrow annulus of sky of diameter about 5° around the disk.

A *pyranometer* is an instrument for measuring solar radiation from the solid angle 2π onto a plane surface. When mounted horizontally facing upwards it measures global solar radiation. If it is provided with a shade that prevents direct solar radiation from reaching the receiver, it measures diffuse solar radiation.

These radiometers have to be calibrated periodically against a standard. An accuracy of about 3% is then obtainable in good instruments.

Figure 3 shows a Robitzsch type solar radiation recorder [2]. The receiving surface consists of a strip of bimetal painted dull black mounted between two similar strips of bimetal shielded from solar radiation by white painted covers. A clear plastic hemisphere protects the receiver from wind and rain. The bimetallic strips are connected to a lever mechanism whose movements are recorded on a rotating chart. The movements of the pen arm are proportional to the difference in temperature between the black and the shielded bimetallic strips. The height of the graph traced on the chart is thus proportional to the intensity of the radiation and is independent of the ambient temperature. The response of the instrument is slow (5 to 10 min), and there is friction between the pen and the chart. Consequently it is not suitable for instantaneous measurements of radiation, but it can be used to obtain daily totals of global solar radiation with an accuracy of about 10%. This



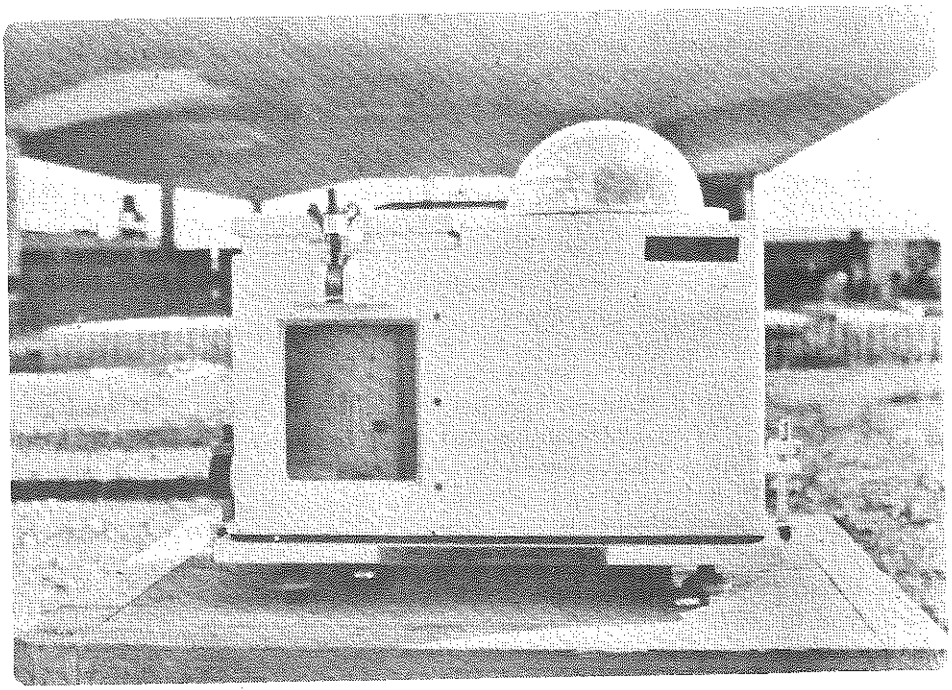


Fig. 3. Solar radiation recorder.

instrument is less expensive and easier to use than the other radiometers because it does not require elaborate electrical accessory equipment.

Apparent Position of the Sun

The solar radiation flux falling on a collector is determined by the position of the sun in the sky, and by the state of the atmosphere. The position of the sun as seen from a particular site on the earth can be calculated exactly from the time, the date, and the location of the site.

The ellipticity of the earth's orbit round the sun and the inclination of the earth's polar axis to the plane of the orbit make the times between successive transits of the sun across the meridian at any one site slightly different from day to day. The length of the day by the sun therefore varies. Clocks, however, indicate mean solar time, and the legnth of the day by the clock is constant.

Greenwich Mean Time (GMT) is the mean time on the meridian through Greenwich, London, starting at midnight.

Zone Mean Time (ZMT) is the mean time on the nearest standard meridian. Standard meridians are multiples of 15° east or west of Greenwich, and each ZMT is a whole number of hours ahead or behind GMT. In Thailand, for example, the standard meridian is 105°E, and ZMT is 7 hours ahead of GMT. Clocks are set to ZMT over the whole country.

Local Mean Time (LMT) is the mean time on the meridian of a particular site. To obtain LMT add to, or subtract from, ZMT 4 minutes for each degree of longitude the site is east, or west, respectively of the standard meridian. The longitude of the Asian Institute of Technology is 100.62°E. The Institute is therefore 4.38° west of the standard meridian, and LMT is 17.5 min behind ZMT.

Apparent Solar Time (AST) is such that the sun crosses the meridian of the site at exactly 12.00 noon every day. It is a consequence of the variation in the length of the day by the sun that AST is sometimes ahead of LMT and sometimes behind LMT. The difference

$$E = \text{AST} - \text{LMT}$$

is called the *equation of time*. Table 1 shows the extreme values of E during the year. Care is needed when referring to data on the equation of time because some authors use the negative of the definition given above.

The *hour angle* H of the sun is the angle between the plane of the meridian and the plane containing the sun and the earth's axis, measured towards the west. Thus at a time t hours after apparent solar noon

$$H = 15t \text{ degrees.}$$

The *declination* D of the sun is the angular distance, on the celestial sphere, of the sun north of the celestial equator. Table 2 shows the extreme values of D during the year.

Table 1. Extreme values of the equation of time.

Date	E (min)
11 February	-14.30
15 April	0
14 May	+ 3.73
14 June	0
26 July	- 6.42
1 September	0
3 November	+16.40
25 December	0

Table 2. Extreme values of the declination of the sun.

Date	D (deg)
20 March	0
21 June	+23.45
23 September	0
22 December	-23.45

The position of the sun as seen from a point on the earth's surface is often more conveniently expressed in terms of the *altitude* A, which is the angle between the direction of the sun and the horizontal plane, and the *azimuth* Z, which is the angle between the south point on the horizon and the foot of the perpendicular drawn from the sun to the horizon measured towards the west.

The angles H, D, A and Z are all shown in Figure 4. The *latitude* L of the site in question is also shown as the angle between the direction of the zenith and the plane of the celestial equator. It can be shown by trigonometry that A and Z can be calculated from L, D and H by means of the following formulae:

$$\sin A = \cos L \cos D \cos H + \sin L \sin D,$$

$$\sin Z = \cos D \sin H / \cos A,$$

$$\cos Z = (\sin L \cos D \cos H - \cos L \sin D)/\cos A.$$

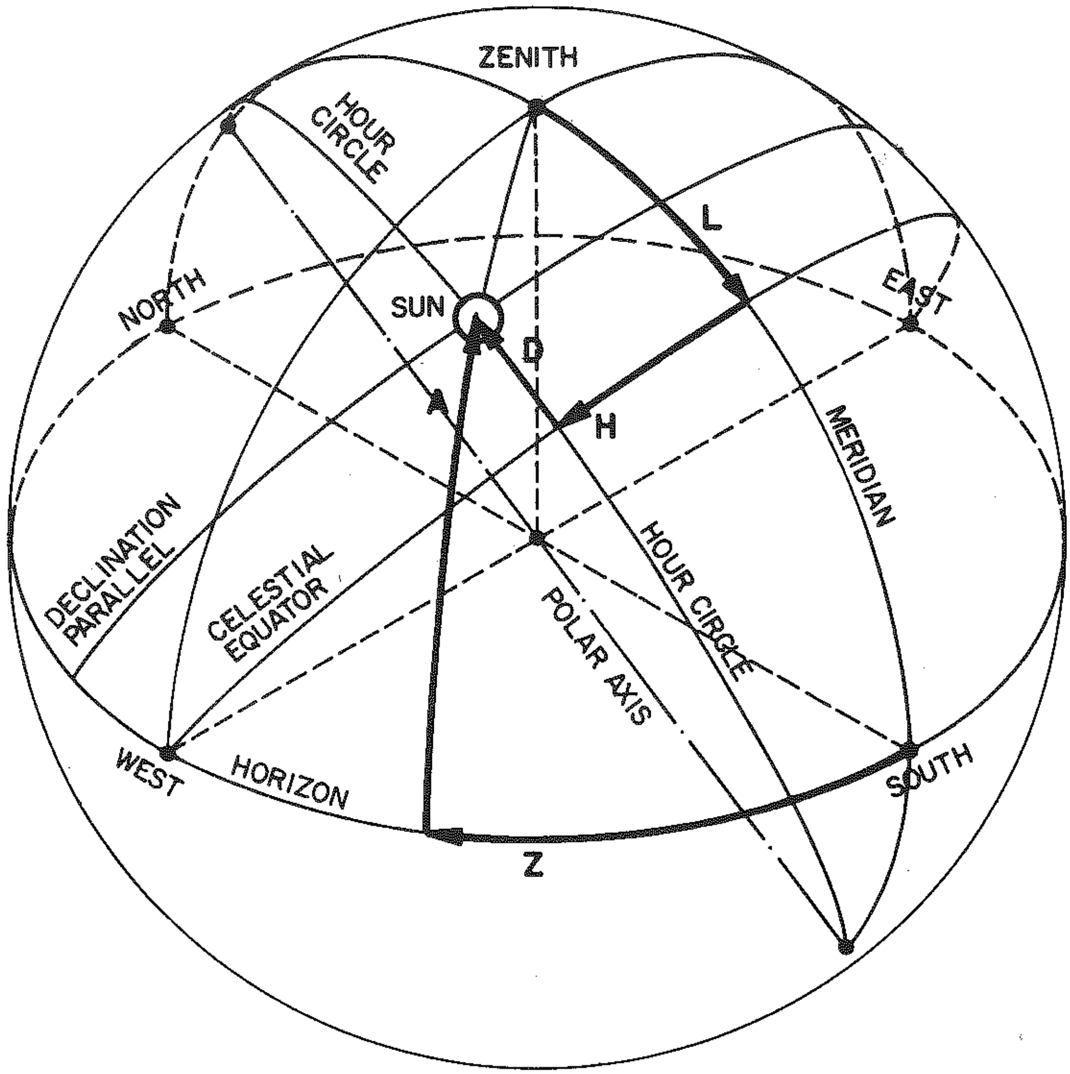


Fig. 4. The celestial sphere.

Figure 5 shows, as an example, how the sun moves across the sky during the day as seen from the Asian Institute of Technology, which has a latitude of 14.08°N.

The values of A calculated from the above formulae are slightly in error due to the refraction of the solar beam by the atmosphere. As a result the apparent altitude is slightly greater than the calculated altitude. For practical purposes the correction is negligible, being only 0.6° when the sun is on the horizon, and less than 0.1° for altitudes greater than 10°.

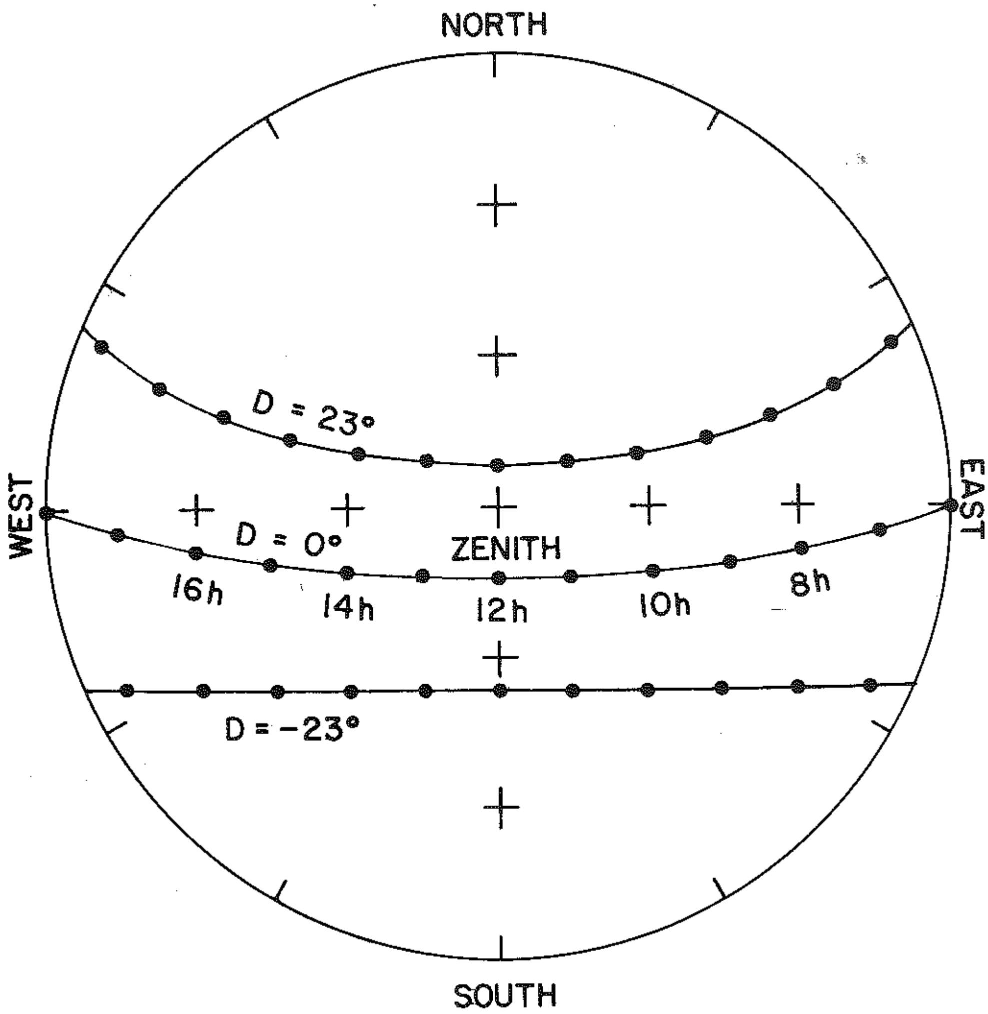


Fig. 5. Apparent motion of the sun in latitude 14° N for three values of the solar declination.

Incidence on an Inclined Plane

The flux of direct solar radiation incident on a flat-plate collector depends on the angle of incidence. This angle can be calculated from the position of the sun in the sky and the orientation of the plane of the collector.

Let A_s and Z_s be the altitude and azimuth of the sun, and let the orientation of the collector be specified by the altitude A_p and the azimuth Z_p of the normal to the collector plane. Then the angle of incidence θ of direct radiation on the collector is given by

$$\cos\theta = \sin A_s \sin A_p + \cos A_s \cos A_p \cos (Z_s - Z_p).$$

If I is the direct solar radiation flux on a plane normal to the solar beam, then $I \cos \theta$ is the flux incident on a flat-plate collector with orientation given by A_p and Z_p.

The diffuse solar radiation flux on an inclined plane, from a uniformly bright sky is $\frac{1}{2}D(1 + \sin A_p)$, where D is the diffuse solar radiation on a horizontal plane. However, if the sky is not uniformly bright (as when the sun is shining) departures from this amount may be expected, and for low values of A_p reflexion of solar radiation from the ground may be important.

Solar Radiation Outside the Atmosphere [3]

The upper graph in Figure 6 shows the spectrum of the radiation emitted by the sun in the wavelength range 0.2 μm to 3 μm. It is close to that of a black body at a temperature of 5900 K. About 8% of the energy is in the ultra-violet region, 44% is in the visible region, and 48% is in the infra-red (see also Figure 1).

The *solar constant* I_o is the direct solar radiation flux outside the earth's atmosphere when the earth is at its mean distance from the sun. Its value is

$$I_o = 1.35 \pm 0.02 \text{ kW/m}^2.$$

Variations in the distance of the earth from the sun due to the ellipticity of the earth's orbit cause the actual intensity of solar radiation outside the atmosphere to depart from I_o by a few percent. Table 3 gives the extreme values of the relative intensity I/I_o, where I is the actual direct solar radiation flux outside the atmosphere.

Effects of the Atmosphere and the Earth

The processes affecting solar radiation that are of importance in solar energy work are scattering, absorption, and reflexion.

The *scattering* of solar radiation is mainly by air molecules, water vapour molecules, water droplets, and dust. This process returns to space about 6% of the incident radiation, and about 20% reaches the earth's surface as diffuse solar radiation.

Air molecules scatter sunlight with an intensity proportional to λ^{-4}, where λ is the wavelength of the radiation. This is called Rayleigh scattering, and it applies to

Table 3. Extreme values of I/I_o.

Date	I/I_o
4 January	1.035
4 April	1.000
5 July	0.967
5 October	1.000

particles of radius less than one tenth of the wavelength. The wavelength effect can be seen in the blue colour of a clear sky, and in the setting sun, which appears red because mainly blue light has been scattered out of the direct beam. Scattering from large particles of radius greater than 25λ is independent of the wavelength. As a result, sunlight scattered from the water droplets in mist, or the dust particles in haze, is white.

The *absorption* of solar radiation is mainly by molecules of ozone and water vapour. Absorption by ozone takes place in the upper atmosphere at elevations above 40 km. It occurs mainly in the ultra-violet region of the spectrum where it is so intense that very little solar radiation of wavelengths less than 0.3 μm reaches the earth's surface. About 3% of the solar radiation is absorbed in this way.

At low levels about 14% of the solar radiation is absorbed by water vapour, mainly in the infra-red. Clouds absorb very little solar radiation, which explains why they do not evaporate in strong sunlight. Their effect on solar radiation is through scattering and reflexion.

The effect of absorption by carbon dioxide on solar radiation is slight, though it is important in atmospheric radiation.

The lower graph in Figure 6 shows qualitatively the spectrum of solar radiation received at the surface of the earth under a clear sky. Due to the scattering process, the peaks of the graph follow a curve below that of the radiation emitted by the sun. The

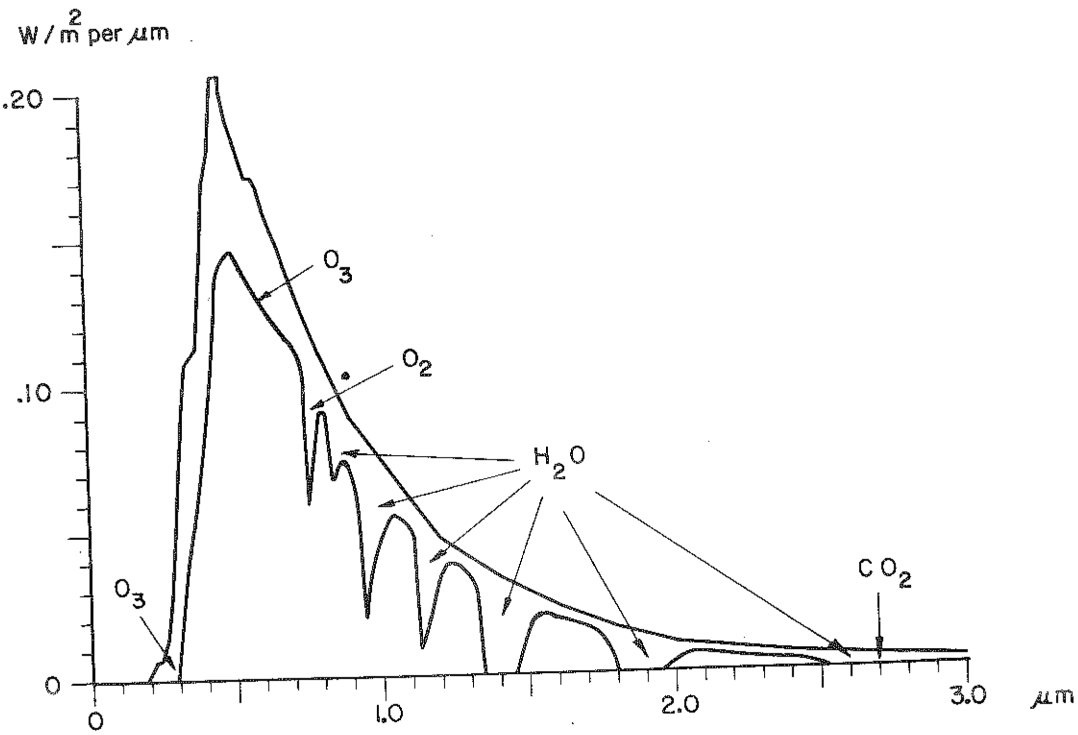


Fig. 6. The solar spectrum.

troughs mark the regions of molecular absorption by the various gases indicated. Variations in the humidity and turbidity of the atmosphere, and in the altitude of the sun, produce large variations in this spectrum.

The *reflexion* of solar radiation depends on the nature of the reflecting surface. The fraction of the intensity of solar radiation that is reflected is called the *albedo* of the surface. The total albedo, which includes all wavelengths, is closely related to visible albedo. Table 4 gives some typical values for the sun overhead. When the sun is low in the sky the albedo of water surfaces is much greater than that tabulated.

Solar Radiation under Clear Skies

Useful calculations of the solar radiation flux at the earth's surface under clear skies in various atmospheric conditions have been published by Schüepp [4]. A simplified presentation of Schüepp's results will be given in this section with emphasis on data applicable in tropical Asia.

The main parameters affecting the intensity of solar radiation are the altitude A of the sun, the water vapour content w of atmosphere expressed in centimetres of precipitable water, and Schüepp's turbidity coefficient B, which is zero in a dust-free atmosphere and increases in value as the air becomes more turbid.

Exact determinations of the water vapour content of the atmosphere require the use of upper air data, but if they are not available approximate estimates can be made with the help of the formula

$$w = 0.18\, e,$$

where e is the surface vapour pressure in millibars [5]. In a tropical wet and dry climate w typically varies from 2 cm in the dry season to 5 cm or more in the wet season.

Direct determinations of the turbidity coefficient B require measurements of direct solar radiation in different ranges of the spectrum using a pyrheliometer with coloured filters. Measurements in India [6], and indirect estimates for Thailand [5], show that in a tropical wet and dry climate B typically varies from near zero during the wet season to

Table 4. Albedo of surfaces with the sun overhead.

Surface	Albedo
vegetation	0.2
light coloured soil	0.3
dark coloured soil	0.1
water	0.1
clouds	0.5-0.9

about 0.1 during the dry season. If, however, there is smoke in the air B can be greater. Inland in Thailand the equation

$$B = 0.25 - 0.017\ V,$$

where V is the mean visibility in kilometres, appears to provide estimates of the mean value of B to within $\pm$ 0.02.

Schüepp's values of the direct solar radiation I for several values of A, w and B are given in Table 5. Small corrections given by Schüepp for variations in the ozone content of the atmosphere, and for variations in the surface air pressure, are ignored. However, it should be observed that the tabulated values are for application at sea level, and would underestimate the solar radiation at elevated mountain sites. A correction for variations in the earth's distance from the sun should be made (see Table 3).

Table 5. Direct solar radiation I, kW/m^2;

w	A	B		
(cm)	(deg)	0	0.1	0.2
2	90	1.047	0.879	0.768
	30	0.879	0.684	0.524
	20	0.810	0.530	0.384
	14	0.712	0.426	0.265
	10	0.628	0.314	0.181
5	90	0.977	0.838	0.698
	30	0.838	0.628	0.475
	20	0.740	0.489	0.349
	14	0.656	0.384	0.244
	10	0.572	0.286	0.161

Diffuse solar radiation is determined mainly by the solar altitude A, the turbidity B, and the albedo of the ground around the site. Table 6 gives Schüepp's values of the diffuse solar radiation D for several values of A and B when the albedo of the ground is 0.25. For albedos 0.1, 0.2 and 0.3 multiply the tabulated values of D by the correction factors 0.90, 0.96 and 1.04 respectively.

Table 6. Diffuse solar radiation D, kW/m^2, for albedo 0.25.

A (deg)	B = 0	B = 0.1	B = 0.2
90	0.063	0.133	0.202
30	0.045	0.087	0.119
20	0.034	0.063	0.084
14	0.024	0.047	0.059
10	0.018	0.034	0.039

Schüepp also gives calculated daily totals of global solar radiation under clear skies in low latitudes. A few representative values are shown in Table 7 for water vapour content $w = 2$ cm and turbidity $B = 0$. For water vapour content $w = 5$ cm multiply the tabulated values by the correction factor 0.93, and for turbidity coefficients $B = 0.1$ and 0.2 apply the correction factors 0.90 and 0.84 respectively.

Table 7. Daily totals of global solar radiation, MJ/m^2, for $w = 2$ cm and $B = 0$.

Latitude (deg)	Date: 15 March	15 June	15 September	15 December
25 N	25.1	30.9	26.4	16.5
15 N	27.3	29.3	27.9	21.2
5 N	28.5	26.8	28.3	25.1
5 S	28.8	23.6	27.9	28.5
15 S	28.2	19.9	26.6	31.2

Effects of Cloud

The effects of cloud on solar radiation received at the earth's surface are complex, and a detailed discussion of them is beyond the scope of this article. However, a few general remarks can usefully be made.

If there is cloud between the sun and the point of observation, then the direct solar radiation is weakened or eliminated. Diffuse solar radiation, on the other hand, may be greater or less in the presence of cloud than its value under a clear sky, depending on the type and amount of the cloud. Thin layers of cloud and scattered clouds reflecting sunlight increase the diffuse solar radiation, while thick layers reduce it.

Global solar radiation is usually reduced by cloud, but if the sun is shining in a clear part of the sky and there are brightly illuminated clouds nearby, then the global solar radiation may be greater than it would have been if the sky had been completely clear.

The geographical, seasonal, and diurnal variations of solar radiation at the earth's surface are controlled as much by the incidence of cloud as they are by the movement of the sun. Consequently, the study of these variations is closely related to the study of climate, and many data are required to build up and adequate description of them. There exists in the literature a good world survey of solar radiation [7], and detailed reports have been published for many individual countries. For tropical Asia, studies of the solar radiation climate in India [8] and Thailand [9] may be mentioned.

In Thailand, the highest mean values of daily global solar radiation are about 20 MJ/m^2 per day, and are widespread in spring. The lowest values are below 15 MJ/m^2 per day in restricted localities with heavy rainfall in the autumn. Diffuse radiation averages 8.4 MJ/m^2 per day.

The frequency distribution of daily totals of global solar radiation has a peak near 20 MJ/m^2 per day at Bangkok during the dry season in winter and spring. It is skewed towards low values with the important result that about 60% of days have daily global solar radiation greater than the mean value. During the wet season in summer and autumn the distribution is more dispersed. Histograms of these distributions are shown in Figure 7.

A study of the diurnal variation of global solar radiation in Thailand shows that the mean midday radiation fluxes range from 0.81 kW/m^2 in spring to 0.58 kW/m^2 in autumn. On the average the radiation received in the afternoon is slightly less than that received in the morning.

Atmospheric Radiation

In the absence of cloud the transfer of long-wave infra-red radiation through the atmosphere is governed mainly by the molecular emissions and absorptions of water vapour, carbon dioxide, and ozone; aerosols may also take part in the process. Near the earth's surface the effects of water vapour are dominant. Consequently, most of the downward atmospheric radiation comes from the moist lower layers below heights of one or two kilometres. In the tropics a flux of 400 W/m^2 is typical.

Figure 8 shows the specrum of atmospheric radiation for a temperature of 27°C. The region from about 8 μm to 12 μm is called the *atmospheric window* where, except for the ozone band at 9.6 μm, the emission and absorption of radiation is weak.

The existence of the atmospheric window causes the intensity of radiation emitted by the air at a particular temperature to be less than the intensity of blackbody radiation at the same temperature. This allows objects on the ground to radiate energy through the

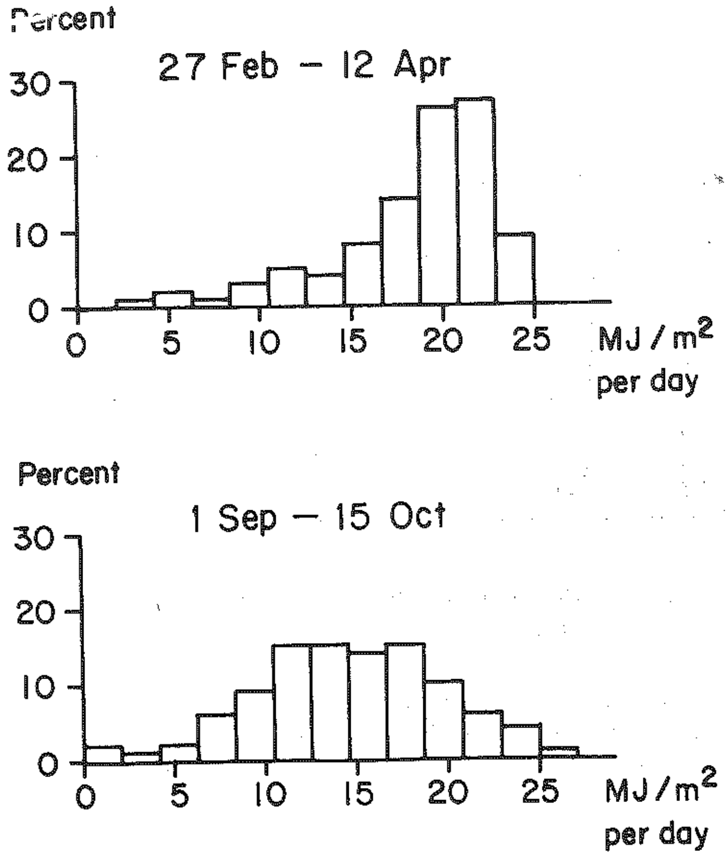


Fig. 7. Distributions of daily global solar radiation at Bangkok.

atmosphere into space, a process that is most noticeable in radiative cooling under a clear sky at night.

Measurement of Atmospheric Radiation

The measurement of atmospheric radiation is difficult because the instrument used is at a temperature close to that of the radiation itself. All radiometers measure the exchange of energy between the receiving surface and its immediate surroundings. In pyranometers the exchange of long wave radiation is mainly with the glass cover over the receiving surface and is negligible compared with the effect of solar radiation.

A *pyrradiometer* is an instrument for measuring the total radiation (both short-wave and long-wave) from the solid angle 2π onto a plane. When mounted facing upwards it measures the total downward radiation. In such an instrument the receiving surface must exchange radiation freely with the sky above. Among the instruments with this property some have the receiving surface protected by a thin polyethylene cover, which is transparent to longwave radiation; others have an unshielded receiving surface ventilated

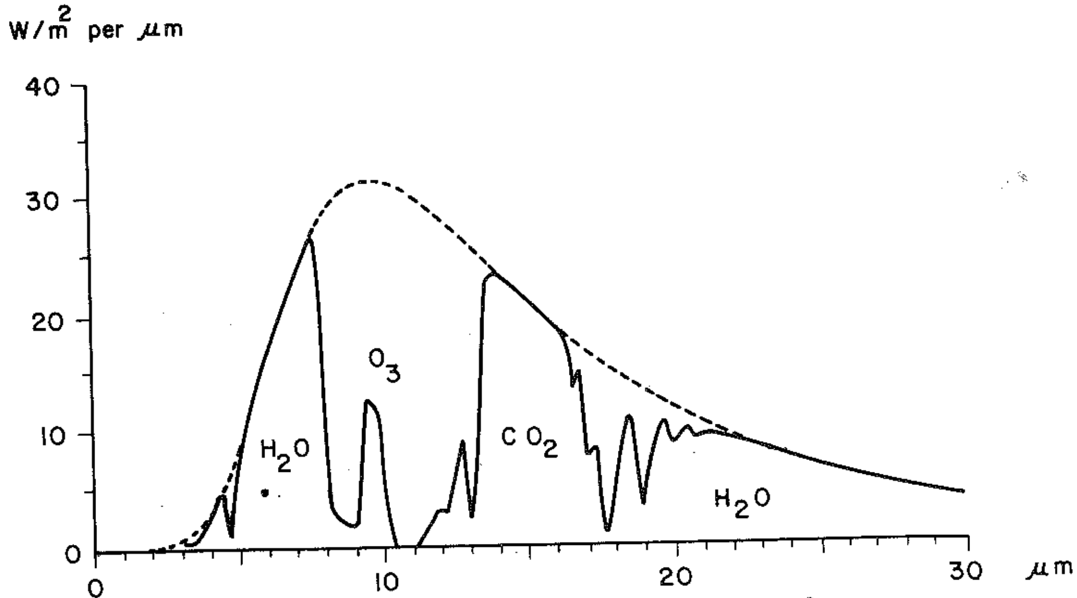


Fig. 8. Spectrum of atmospheric radiation at 27° C. Dotted curve shows blackbody spectrum.

by a jet of air to stabilize the heat exchange by convection. If measurements of atmospheric radiation only are required, the observations must be made at night since there is no way of removing the unwanted solar radiation component.

Calculation of Atmospheric Radiation

Measurements of atmospheric radiation are rare, especially in developing countries. However, its intensity can be calculated from routine meteorological data with an accuracy comparable with that of the measurements. A simple method of calculation will be outlined in this section. A comprehensive treatment of the subject is available elsewhere [10].

The atmospheric radiation flux on a horizontal surface under clear skies at night can be estimated from the surface air temperature with a probable error less than 5 W/m² with the help of an empirical relation due to Idso and Jackson [11] that fits experimental data and appears to be valid for all latitudes and seasons. The Idso-Jackson formula is

$$L_o/\sigma T^4 = 1 - 0.261 \exp\{-0.000777\,(273 - T)^2\}$$

where L_o is the downward atmospheric radiation flux from a clear sky, σ is the Stefan-Boltzmann constant 5.67×10^{-8} W/m²K⁴; and T is the absolute screen-level air temperature in kelvins. Table 8 gives values of the black-body radiation flux σT^4 and the downward atmospheric radiation flux L_o calculated from the Idso-Jackson formula for various surface air temperatures.

The effective sky temperature T_e is defined to be the temperature of blackbody radiation having the same flux as the downward atmospheric radiation. It is always less than the surface air temperature because the atmospheric radiation flux is always less than the blackbody radiation flux at the same temperature. Values of T_e are given in Table 8. They are of interest because they indicate the coldest temperatures that can be reached by objects radiating heat to the sky under the prevailing conditions.*

Table 8 Values of σT^4, L_o and T_e for various surface air temperatures

Temperature (°C)	σT^4 (W/m^2)	L_o (W/m^2)	T_e (°C)
0	315.6	233.3	-19.9
5	339.4	252.5	-14.8
10	364.5	276.4	- 8.9
15	390.9	305.2	- 2.3
20	418.7	338.6	4.8
25	448.0	376.1	12.2
30	478.9	416.8	19.7
35	511.2	459.7	26.9
40	545.2	504.2	33.9
45	580.9	549.5	40.6

In using the Idso-Jackson formula to estimate the downward atmospheric radiation two corrections are necessary. The first is a diurnal correction for variations in the air temperature lapse rate near the surface. During the afternoon the high temperature at the surface, and the resulting steep lapse rate in the bottom layer of the atmosphere, causes the Idso-Jackson formula to overestimate L_o, which contains substantial contributions from the cooler air several hundred metres above. The correction required is about $-20\ W/m^2$. Likewise, at dawn after a clear night there is a ground level temperature inversion produced by nocturnal radiative cooling of the surface, and the opposite process causes the Idso-Jackson formula to underestimate L_o by about $15\ W/m^2$. Between these times during the forenoon and in the evening the Idso-Jackson formula gives good estimates of L_o.

The second correction necessary in using the Idso-Jackson formula is for the effect of thermal radiation from clouds, which always increases the downward atmospheric radiation. A simple formula for the downward atmospheric radiation L in the presence of cloud is

$$L = L_o + (\sigma T^4 - L_o)\, kn,$$

where L_o is the radiation in the absence of cloud, k is a factor that depends on the cloud type and height, and n is the cloud amount on the scale $n = 0$ for a clear sky and $n = 1$ for an overcast sky. Suitable values of k for low, medium, and high clouds in a tropical climate are 0.86, 0.50 and 0.17 respectively. In colder climates k should have slightly greater values than these because the air contains less water vapour and therefore has less of a screening action on the radiation between the clouds and the surface.

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