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Direct beam solar irradiance and illuminance

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ABSTRACT. In solving many practical problems, related to buildings etc, correct assessment of direct beam solar irradiance and/or illuminance is a necessary prerequisite. For this purpose computational methods are usually adopted. Since 1940, when Parry Moon first collected and compared the necessary information and put them in usable form for engineering applications, several studies on different aspects of the subject have been made leading to new information. In the present paper more recent and reliable information on extra-terrestrial solar spectral irradiance, molecular and aerosol scattering of spectral energy, and the absorption of radiation by ozone, water vapour and carbon dioxide etc present in the terrestrial atmosphere, have been used to compute the spectral transmissivities and spectral energy on the earth's surface for different conditions and subsequently integrated values of direct beam solar irradiance and illuminance normal to sun rays. The results, presented in the form of simple usable curves, have been compared with other available data and some measured values at Roorkee. It is believed that these will be useful from practical application point of view.

1. Introduction

In many practical problems such as those related with buildings more often than not computational methods are adopted for estimating incident solar irradiance and illuminance on building facades under clear (cloudless) sky conditions. For this purpose the variation in the intensity of direct beam solar irradiance or illuminance with different conditions on the earth's surface, in addition to their diffuse components, should be known with certain degree of preciseness. The estimation of direct solar irradiance which plays an important role in the energy balance of buildings, has received the attention of a number of workers. Moon (1940) was first to collate and compare the necessary information and propose standard curves for sunlight at sea level for use in engineering calculations. His work has received wide acceptance in the field of building science. Moon's fundamental procedure has been extended by several other workers (Threlkeld & Jordan 1957; Rao & Seshadri 1961; Curtis & Lawrence 1972; Cole 1976). These authors have generally taken Moon's equations and data on atmospheric scattering and absorption but used contemporary extra-terrestrial spectral irradiance and presented the final results in somewhat different fashion. The data on direct solar radiation given in such reference guides as ASHRAE (1960) and IHVE (1970) are based on Moon's work.

However, since 1940 studies on scattering and absorption of incoming solar radiation by the constituents of the earth's atmosphere as also on extra-terrestrial spectral irradiance, have led to more reliable information on input data which should be used for the computation of direct beam solar irradiance and illuminance on the earth's surface. Besides this, it seems appropriate to provide these quantities over a wider range of atmospheric conditions to make results more comprehensive. The present work is an attempt in this direction. The final results are presented as curves of direct beam solar irradiance and illuminance varying with solar altitude for different values of Angstrom's turbidity coefficient and precipitable water. Angstrom's turbidity coefficient has been used to account for the scattering due to aerosols, *i.e.,* all the solid and liquid particles present in the atmosphere, *in lieu* of the separate formulae for dust and water vapour scattering as done by earlier workers. The results so obtained are compared with the existing data, some of which are based on empirical expressions (Spencer 1965; Chrocicki 1972). Comparisons are also made with some measurements carried out at Roorkee.

2. Extra-terrestrial solar spectral irradiance and solar constant

The electromagnetic spectrum emitted by the sun extends from about X-rays of wavelengths 0.1 nm to radiowaves of 100 metres. However, about 99.9 per cent of the sun's energy is found to lie in the range from 0.2μ to 9.0μ. As recommended by Thekaekara (1973) and also adopted by C. I. E. (1972) as current International Standard, this range has been considered in this paper. The upper curve of Fig. 2 shows Thekaekara's distribution of the solar spectral irradiance (normal to sun rays) for mean sun-earth distance. According to this curve the value of the solar constant (integrated spectral irradiance) is 135.3 mW/cm^2 or 1.94 cal/cm^2/min.

3. Atmospheric attenuation of solar radiation due to scattering

Two types of scattering take place in the earth's atmosphere. When the radii of the particles are much smaller in comparison with the wavelength, *i.e.,* roughly $\leqslant 0.1\ \lambda\ \mu$, Rayleigh scattering is applicable. This holds good in the case of air molecules. The spectral radiation intensity after depletion due to air molecules of the earth's atmosphere according to Bouger-Lambert law is given by :

$$I_{a\lambda} = I_{0\lambda} \exp \left[-\sigma_{a\lambda} m_h \ (\theta). \ H \right]$$
$$= I_{0\lambda} \ \tau_{a\lambda}^{m_h(\theta)} \qquad (1)$$

where $I_{o\lambda}$ and $I_{a\lambda}$ are the spectral energy at the wavelength λ outside the earth's atmosphere and at the earth's surface, respectively, $\sigma_{a\lambda}$ is the volume scattering coefficient per unit length for Rayleigh particles (air molecules), $H = 7991$ metres is the vertical height of the imaginary homogeneous atmosphere at normal temperature and pressure (NTP), which is equivalent to the actual atmosphere, $\tau_{a\lambda}$ is the zenith spectral transmissivity due to Rayleigh scattering of the entire terrestrial atmosphere and $m_h \ (\theta) = m_r \ (\theta) \ P_h/P_0$, is the absolute actual airmass. $m_r \ (\theta) \approx$ cosec θ is the relative airmass and P_h and P_0 are respectively the atmospheric pressures at height h and mean sea level (MSL). In this paper more accurate values of relative airmass given by Bemporad (List 1958) which take into account atmospheric refraction and earth's curvature etc, have been used.

The Rayleigh volume scattering coefficient at a particular wavelength λ may be written as :

$$\sigma_{a\lambda} = \frac{32\pi^3 \ (n\lambda - 1)^2}{3\lambda^4 \ N} \left[\frac{6 + 3 \ \rho_{n\lambda}}{6 - 7 \ \rho_{n\lambda}} \right] (2)$$

where N is the number density (number of molecules/cm^3), n_λ the relative index of refraction of the medium and $\rho_{n\lambda}$ the depolarization factor. The volume scattering coefficients given by Penndorf (1957) have been used in this paper. There is good reason to believe that these values are more accurate.

The large solid and liquid particles suspended in the atmosphere whose radii are much larger, varying from about 0.04 to 10 μ, are called aerosols. In the case of aerosol particles more complicated scattering theory developed by Mie (1908) is applicable. For determining the total extinction of solar radiation due to atmospheric haze, according to Mie's theory, the density and size distribution of the aerosols and vertical extent of their distribution in the atmosphere must be precisely known alongwith scattering cross-section over the entire range of size parameters. Nevertheless, these parameters are inadequately known. In view of this Angstrom's analytical formula (1964) has been used here for calculating the extinction due to aerosols as it is simple and accurate enough for practical computations. In the Moon's paper as also in other papers based on his work separate formulae for dust and water vapour scattering, based on some measurements by Fowle, have been given. However, the data on the representative values of dust (number of particles/cm^3), which is essentially a dust index rather than absolute concentration, are rarely available. On the other hand global distribution of Angstrom's turbidity coefficient has been determined (*see* Appendix). According to Angstrom's formula the scattering coefficient $\sigma_{\beta\lambda}$ for aerosols is given by

$$\sigma_{\beta\lambda} = \beta\lambda^{-\alpha} \qquad (3)$$

where β is Angstrom's turbidity coefficient and α the wavelength exponent. In the present calculations α has been taken constant and equal to 1.3. The spectral solar intensity at the earth's surface due to aerosol scattering, $I_{\beta\lambda}$, is given by :

$$I_{\beta\lambda} = I_{0\lambda} \exp \left[-\sigma_{\beta\lambda}. \ m_r \ (\theta) \right]$$
$$= I_{0\lambda} \ \tau_{\beta\lambda}^{m_r(\theta)} \qquad (4)$$

Here $\tau_{\beta\lambda}$ is the zenith spectral transmissivity of the atmosphere due to aerosol scattering.

4. Atmospheric absorption of solar radiation

Whereas atmospheric scattering is a continuous function of wavelength, the atmospheric absorption is generally selective in nature. The main absorbing components of the atmosphere whose contribution is of significance to the total depletion in the solar beam are water vapour, carbon dioxide, oxygen and ozone. Influence of ozone absorption on the incoming energy is confined mostly to so called Hartley bands (0.20 μ to 0.30 μ), Huggins bands (0.30 μ to 0.36 μ) and Chappuis bands (0.44 μ to 0.76 μ). In the present calculations the most widely used data on ozone absorption given by Vigroux (1953) have been utilized. Variation in the amount of ozone is found to cause only negligible change in the quantity of incoming solar radiation. Therefore, calculations are included only for the average value of ozone content of the atmosphere, *i.e.*, $z=3.4$ mm of ozone. The spectral intensity due to absorption by ozone, $I_{z\lambda}$, is given by :

$$I_{z\lambda} = I_{0\lambda} \exp \left(-0.34. \ m_r \ (\theta) \ \sigma_{z\lambda} \right)$$
$$= I_{0\lambda} \ \tau_{z\lambda}^{m_r(\theta)} \qquad (5)$$

where $\tau_{z\lambda}$ is the zenith spectral transmissivity for average ozone content of the atmosphere.

Water vapour and carbon dioxide are mainly responsible for the absorption of radiant energy in the near infrared region of the solar spectrum. Absorption bands by precipitable water vapour occur at wavelengths 0.7, 0.8, 0.9, 1.1, 1.4, 1.9, 2.7, 3.2 and 6.3μ and those by carbon dioxide at wavelengths 1.6, 2.0, 2.7 and 4.3 μ. Beyond 6.3 band of H_2O there is an almost transparent region. Though sophisticated methods are available for the calculation of band transmittance in the

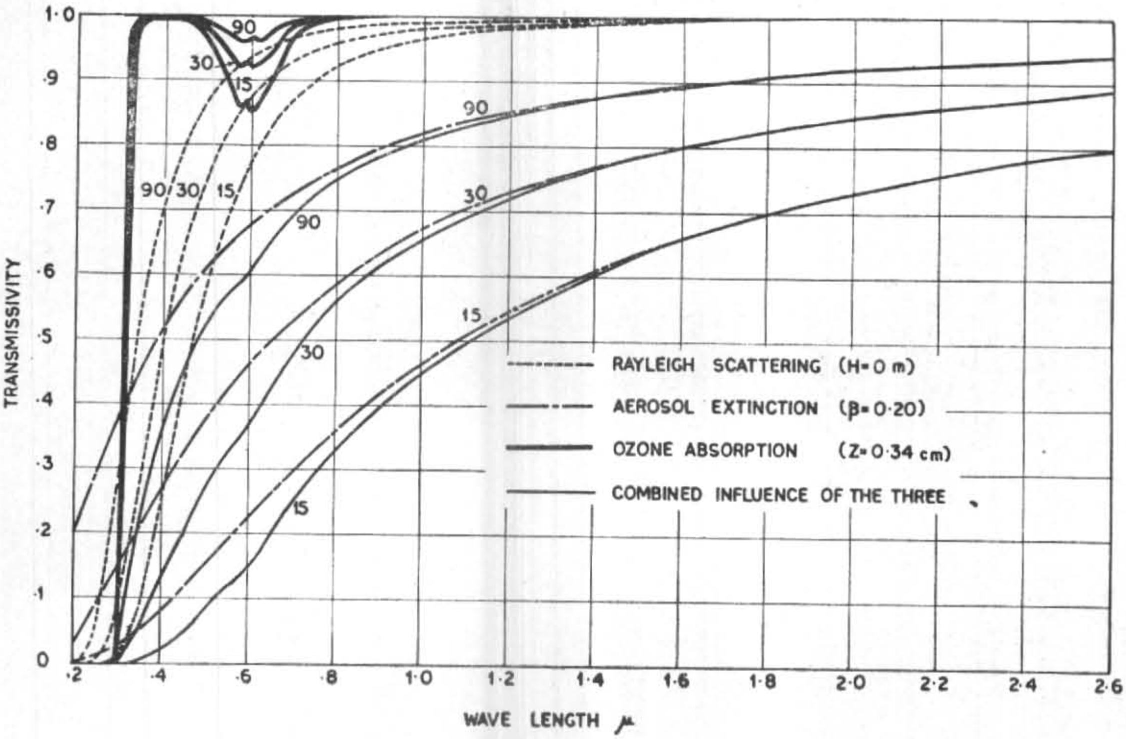

Fig. 1. Typical spectral transmissivities at solar altitudes of 15, 30 and 90 degrees

far infrared region beyond 6.3 μH_2O band (Elsasser 1942; Goody 1962) no such method has been successfully applied in the near infrared region. Yamamoto (1962) has given absorptivities over H_2O, CO_2 and O_2 bands based on the experimental work of some other workers (Howard *et al.* 1955; Fowle 1917; Langley and Abbot 1900; Gates 1960). These absorptivity values have been used in this paper.

5. Direct beam solar irradiance at the earth's surface

To compute the direct beam irradiance at the earth's surface the spectral transmissivities were computed with an IBM 1620 computer at ten airmasses corresponding to zenith distances of 0, 15, 30, 45, 60, 75, 80, 83 and 85 degrees and for different atmospheric conditions. The input data used in the calculations is given in Table 1. Validity of Bouger-Lambert law was assumed to account for the combined influence of scattering and absorption. A sample variation of spectral transmissivities upto 2.6 μ for Rayleigh particles, aerosols and ozone for three solar altitudes of 15, 30 and 90 degrees are plotted on Fig. 1. Overall transmissivities resulting from the combined influence of the three constituents are also shown on this figure. Spectral transmissivities due to Rayleigh scattering become almost equal to unity at wavelengths above about 1.6 μ at all airmasses. The aerosol transmissivities are lowest and the variation becomes slow only above about 3.0 μ. Below about 0.3μ the transmission due to ozone absorption is practically zero owing to very strong absorption in the Hartley bands and above it, in Huggins and Chappuis bands, the absorption is weak and limited almost to a narrow wavelength region around 0.6μ. The sea level spectral energy distributions, for the conditions of Fig. 1, are plotted on Fig. 2.

The integrated direct beam solar irradiance, $I_N(\theta, \beta, w, z)$, at the earth's surface can be written as :

$$I_N(\theta,\beta,w,z) = \int_{0.20}^{9.0} I_{0\lambda} \exp\left[-\left\{\sigma_{a\lambda}.\, H.\, m_h(\theta) + \left(\beta\lambda^{-1.3} + 3.4\sigma_{z\lambda} + w\sigma_{\omega\lambda} + \ldots\right) m_r(\theta)\right\}\right] d\lambda \qquad (6)$$

$$= \int_{0.20}^{9.0} I_{0\lambda}.\, \tau_{a\lambda}^{m_h(\theta)} \left(\tau_{z\lambda}\, \tau_{z\lambda} \cdot \tau_{\omega\lambda} \ldots\right)^{m_r(\theta)} d\lambda \qquad (7)$$

where ω, $\sigma_{\omega\lambda}$ and $\tau_{z\lambda}$ are respectively the amount absorption coefficient and transmissivity of atmospheric precipitable water. The terms representing absorptions due to CO_2 and O_2 can also be added in the above expressions as indicated

TABLE 1

Basic data used in the computations

λ	$I_{0\lambda}$	$\sigma_{a\lambda}$	$\sigma_{z\lambda}$	$V(\lambda)$	λ	$I_{0\lambda}$	$\sigma_{a\lambda}$	$\sigma_{z\lambda}$	$V(\lambda)$
0.20	1.07	954.2×10^{-8}	8.61	—	0.57	171.2	106.0×10^{-9}	0.114	0.952
0.22	5.75	603.1	14.7	—	0.58	171.5	987.0×10^{-10}	0.116	0.870
0.23	6.67	493.4	122.0	—	0.59	170.0	921.0	0.108	0.757
0.24	6.30	406.1	216.0	—	0.60	166.6	860.0	0.124	0.631
0.25	7.04	338.0	299.0	—	0.62	160.2	753.1	0.105	0.381
0.26	13.0	284.2	295.0	—	0.64	154.4	662.0	0.091	0.175
0.27	23.2	240.8	205.0	—	0.66	148.6	584.4	0.064	0.061
0.28	22.2	205.5	104.0	—	0.68	142.7	517.8	0.035	0.017
0.29	48.2	176.5	35.7	—	0.70	136.9	460.5	0.022	0.0041
0.30	51.4	152.5	10.34	—	0.72	131.4	410.9	0.015	0.0011
0.31	68.9	132.6	2.74	—	0.75	123.5	348.4	0.010	0.0001
0.32	83.0	115.8	0.894	—	0.80	110.9	268.4	—	—
0.33	105.9	101.6	0.129	—	0.90	89.1	167.0	—	—
0.34	107.4	895.9×10^{-9}	0.064	—	1.00	74.8	109.2	—	—
0.35	109.3	792.9	0.007	—	1.2	48.5	525.0×10^{-11}	—	—
0.36	106.8	704.5	0.002	—	1.4	33.7	282.8	—	—
0.37	118.1	628.2	—	—	1.6	24.5	165.5	—	—
0.38	112.0	562.0	—	—	1.8	15.9	103.3	—	—
0.39	109.8	504.3	—	0·0001	2.0	10.3	677.0×10^{-12}	—	—
0.40	142.9	454.0	—	0.0004	2.2	7.9	277.0	—	—
0.41	175.1	409.8	—	0.0012	2.4	6.2	277.0	—	—
0.42	174.7	370.8	—	0.0040	2.6	4.8	277.0	—	—
0.43	163.9	336.5	—	0.0116	2.8	3.9	277.0	—	—
0.44	181.0	306.0	—	0.023	3.0	3.1	133.5	—	—
0.45	200.6	278.9	—	0.038	3.2	2.26	720.4×10^{-13}	—	—
0.46	206.6	254.8	—	0.060	3.4	1.66	720.4	—	—
0.47	203.3	232.3	—	0.091	3.6	1.35	720.4	—	—
0.48	207.4	214.0	0.016	0.139	3.8	1.11	720.4	—	—
0.49	195.0	196.6	0.019	0.208	4.0	0.95	422.2	..	—
0.50	194.2	181.0	0.030	0.323	4.5	0.59	263.6	—	—
0.51	188.2	166.9	0.039	0.503	5.0	0.38	172.9	—	—
0.52	183.3	154.2	0.047	0.710	6.0	0.18	833.7×10^{-14}	—	—
0.53	184.2	142.6	0.063	0.862	7.0	0.10	425.1	—	—
0.54	178.3	132.1	0.071	0.954	8.0	0.06	263.7	—	—
0.55	172.5	122.6	0.084	0.995	9.0	0.04	164.7	—	—
0.56	169.5	113.9	0.097	0.995					

λ=Wavelength (μ)

$I_{0\lambda}$=Extra-terrestrial solar spectral irradiance averaged over small bandwidth centred at wavelength λ (mW/cm^2/μ)

$\sigma_{a\lambda}$=Rayleigh scattering coefficient at wavelength λ (cm^{-1})

$\sigma_{z\lambda}$=Ozone absorption coefficient at wavelength λ (cm^{-1}).

$V(\lambda)$=Relative spectral luminosity of the human eye for photopic vision at wavelength λ.

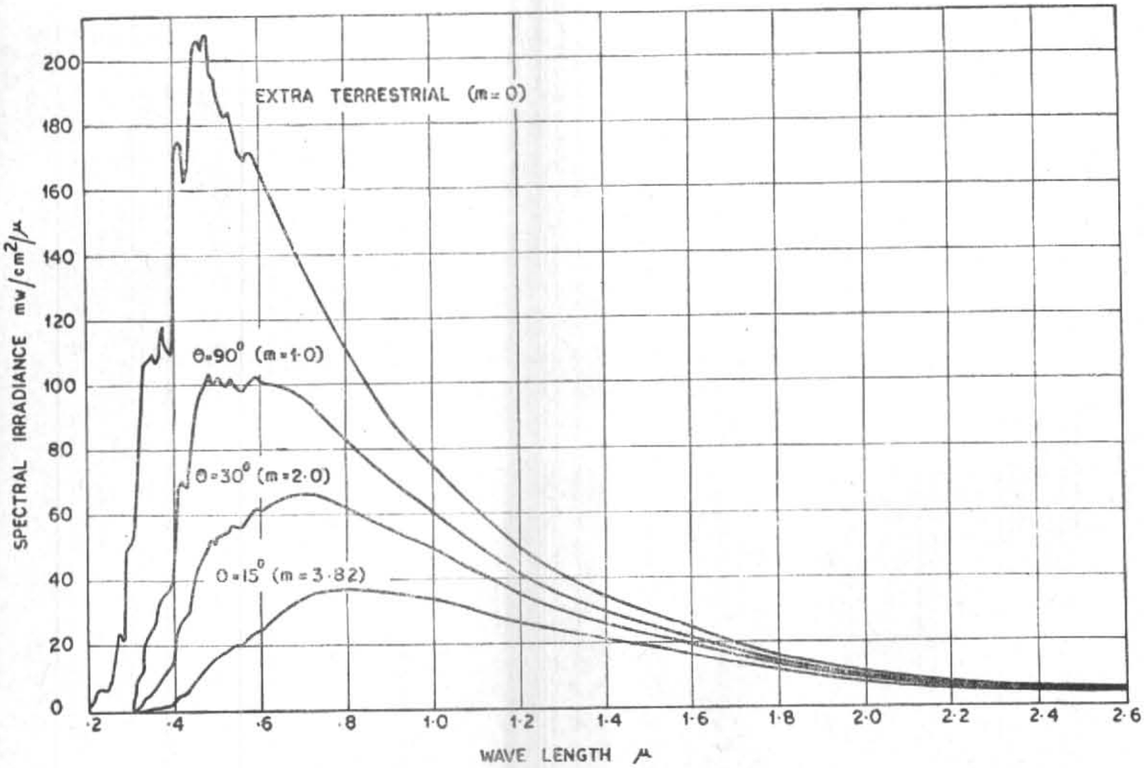

Fig. 2. Spectral irradiance outside the terrestrial atmosphere and at the earth's surface at solar altitudes of 15, 30 and 90 degrees for $H=0$, m=0.20 and $z=0.34$ cm

by dots. As it is rather difficult to introduce spectral absorption coefficients due to H_2O, CO_2 and O_2 in the Eqns. (6) and (7), their influence was taken into account by using the total absorptivities, A_{wco}, over all their bands given by Yamamoto as stated earlier. The direct beam solar irradiance was, therefore, calculated as below :

$$I_N'(\theta, \beta, z) = \int_{0\cdot 20}^{9\cdot 0} I_{0\lambda}.\ \tau_{a\lambda}^{m_h(\theta)} . \left(\tau_{\beta\lambda}.\ \tau_{z\lambda} \right)^{m_r(\theta)} d\lambda \qquad (8)$$

$$I_N(\theta, \beta, w, z) = I_N'(\theta, \beta, z) \times (1 - A_{wao}) \quad (9)$$

The integral in Eq. (8) was evaluated by computer using Simpson's rule. The accuracy of the integration method was checked by integrating the extra-terrestrial spectral irradiance which gave almost exactly the value of solar constant.

Fig. 3(a) shows the direct beam irradiance plotted against solar altitude for different atmospheric conditions at sea level. Curves are plotted for Angstrom's turbidity coefficients 0.05, 0.10, 0.20 and 0.30 and precipitable water of 0.5, 1, 3, 5 and 7 cm. For intermediate values of these parameters the interpolated values are found to be quite close to actually calculated values. The curves for pure Rayleigh atmosphere is also shown on the figure. From Fig. 3(a) it can be seen that the influence of variation in aerosol content (change in turbidity coefficient) on incoming direct radiation is most predominant. The influence of variation in water vapour content is also appreciable. To investigate the effect of station height on incoming radiation, calculations were carried out for other elevations. The radiation curves for 5000 m are shown in Fig. 3(b). The change in radiation is found to be of the order of 10 m cal/cm²/min for every 1000 m change in elevation. Substantial increase in incident solar radiation at higher elevations may be attributed to decreased concentration of aerosols and precipitable water at these heights as has also been indicated by Threlkeld and Jordan.

6. Direct beam solar illuminance

The human eye can perceive radiant energy in the wavelength region from 0.38μ to 0.78μ only, which is known as visible region of the solar spectrum. The same amount of energy at different wavelengths in the visible region gives different amounts of light depending upon the spectral luminosity function at that wavelength. The values of the relative spectral luminosity function $V(\lambda)$ for photopic vision (Table 1) are those given in Smithsonian Institution Meteorological Tables (List 1958) and used in the present calculations. The direct beam illuminance was calculated from the formula :

$$E_N(\theta, \beta, z) = K_{\max} \int_{0\cdot 38}^{0\cdot 78} V(\lambda)\ I_{0\lambda}\ \tau_{a\lambda}^{m_h(\theta)} \left(\tau_{\beta\lambda}.\ \tau_{z\lambda} \right)^{m_r(\theta)} d\lambda \quad (10)$$

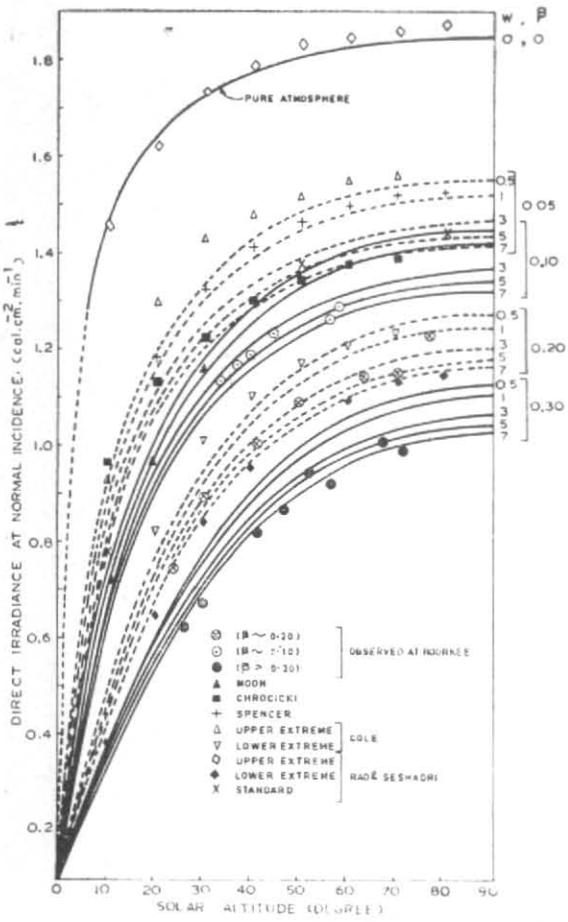

Fig. 3(a). Curves for direct solar irradiance at normal incidence for $H=0$ m, $z=0.34$ cm and different values of Angstrom's turbidity coefficient and precipitable water

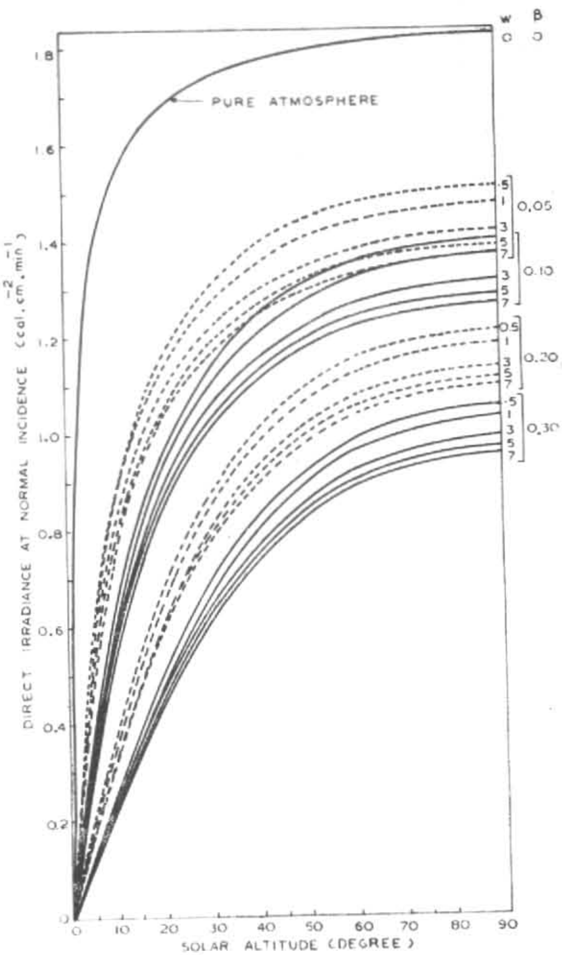

Fig. 3(b). Curves for direct solar irradiance at normal incidence for $H=5000$ m, other conditions being same as in Fig. 3(a)

where, $K_{max} = 680$ lumens/watt ($\approx$ 475K lux per cal cm^{-2} min^{-1}). The absorption due to water vapour etc can be neglected in this case since in the visible region it is very small and occurs at wavelengths where the value of relative luminosity function itself is too low. Hence the values of $E_N(\theta, \beta, z)$ are not significantly affected by water vapour absorption. Fig. 4 shows the variation of direct beam illuminance curves plotted against solar altitude for the same turbidity coefficients and station heights as the irradiance curves. As can be seen the influence of Angstrom's turbidity coefficients is much more pronounced in this case also as compared to changes in station heights above m.s.l. The influence of precipitible water vapour and/or station height on irradiance and illuminance becomes less pronounced with increasing values of turbidity coefficient.

7. Comparison with other data

The values of direct irradiance taken from Moon, Rao and Seshadri, Cole, Spencer and Chrocicki and some values measured at Roorkee (Lat. 29° 51′ N, 274 m above m.s.l.) are plotted in Fig. 3(a). Moon's values which correspond to d (dust) $=300$ particles/cm³, $w=2$ cm and $z=2.8$ mm, mostly lie on the curve for $\beta=0.10$ and $w=1$ cm. The upper extreme values of Cole ($d=200$ particles/cm³, $w=0.5$ cm) are nearer to the upper extreme curve, corresponding to $\beta=0.05$ and $w=0.5$ cm. The upper extreme value of Rao and Seshadri corresponding to $d=100$ particles/cm² and $w=0$ cm are much larger. These are not likely to be obtained near sea level (Chandra 1977). Lower extreme values of Cole ($d=800$ particles/cm³, $w=7$ cm) and Rao and Seshadri ($d=800$ particles/cm³, $w=8$ cm) lie near the curves for $\beta=0.20$ which represent fairly high atmospheric turbidity conditions. The curves for $\beta=0.30$ may represent only extremely hazy conditions as are often obtained over northern India in pre-monsoon summer months when the sky is laden with large sized dust particles. The Spencer's values of irradiance based on observed data at Melbourne correspond to less turbid conditions whereas the data of Chrocicki, based on measurements at eighteen stations

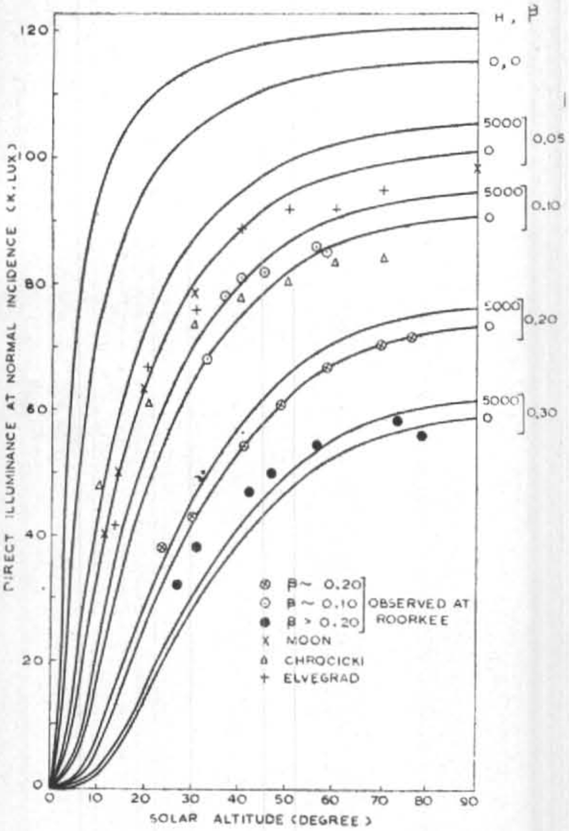

Fig. 4. Curves for direct solar illuminance at normal incidence for H=20 and 5000 m, z=0.34 and different values of Angstroms turbidity coefficient

in Europe, correspond to relatively higher turbidity. The values of direct illuminance obtained from Moon (1940), Chrocicki (1972) and Elvegrad & Sjostedt (1940) are shown in Fig. 4 for comparison with the present illuminance curves. In tropics where the atmosphere is more turbid the illuminance values are less.

An attempt was also made to compare the present results with some measured values of direct beam irradiance and illuminance. The direct irradiance was measured with a Linke and Fuessner actinometer and the direct illuminance was obtained through computations from measured direct and diffuse illumination on a horizontal surface from the formula $E_N=(E_{TH}-E_{dH})/\sin\theta$. The turbidity coefficient was estimated from actinometric measurements of spectral solar radiation using the Schott red standard meteorological filter RG2. The measured values of irradiance and illuminance are plotted for three typical days. On these days the values of turbidity coefficients were found to be of the order of 0.10, 0.20 and >0.20, respectively. The measured values of irradiance and illuminance are quite close to corresponding calculated values. The comparison seems to be good.

8. Conclusions

The direct irradiance and illuminance curves presented in this paper have been computed with more recent and reliable data on extra-terrestrial spectral irradiance and atmospheric scattering and absorption and are given over varied range of atmospheric conditions. There is a close agreement between the calculated and experimental results. If the turbidity conditions and precipitable water are known or estimated approximately (*see* Appendix) for any place, the relevant quantities can be read or intrapolated from these curves. For planes other than normal to sun's rays the well known trigonometric formulae may be used.

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Appendix

For the estimation of irradiance/illuminance from the curves given in this paper, a knowledge of Agnstrom's turbidity coefficient is required. If for a particular place or region it is not known, the standard values of turbidity coefficient B of Angstrom-Schuepp ($\beta = B/1.05$ for $\alpha = 1.3$ according to Mani & Chacko 1973) given by Robinson (1966) may be used. For India β has been given for a number of stations by Mani and Chacko (1973). The annual variation of β for some stations of the world is given by Angstrom (1961). Yamamot o *et al.* (1968) have also given hemispherical distribution of turbidity coefficient.

Water vapour content of the atmosphere is generally available from publications of the National Meteorological Departments. If it is not available for a particular station the values given by Robinson (1966) may be utilized. Mean precipitable water vapour in the atmosphere over India in different months is given by Mokashi (1968).