Chapter 6 ***SOLAR SPECTRAL RADIATION UNDER CLOUDLESS SKIES***

Absorption of Direct Solar Radiation by Gases

Unlike scattering by air molecules and aerosols, which varies continuously with wavelength, the absorption of solar radiation by atmospheric gases is a selective process.

Gases such as O_2, O_3, CO_2, CO, and water vapor absorb radiation at specific wavelengths, depending on the energy transitions within their molecules. These gases are called molecular absorbers and are particularly active in the near- and far-infrared regions. In contrast, atomic oxygen and nitrogen absorb mainly in the ultraviolet region.

When radiation interacts with a particle, part of its energy may be scattered and part absorbed. In the latter case, the particle acts as an energy sink. For gases, absorption occurs only when the radiation energy matches the energy difference between two molecular states. Absorption at individual wavelengths is called selective absorption, while absorption over a group of closely spaced wavelengths is known as band absorption.

Absorbers in the Ultraviolet and Visible Spectrum

Diatomic oxygen (O_2), nitrogen (N_2), atomic oxygen (O), nitrogen (N), and ozone (O_3) are the five principal absorbers in the ultraviolet and visible spectrum.

Atomic ***oxygen*** and ***nitrogen*** absorb x rays and other short-wave radiation continuously up to about $0.085\ \mu m$. As these two gases are found high in the atmosphere, no radiation of wavelength less than $0.085\ \mu m$ passes through to the lower atmosphere.

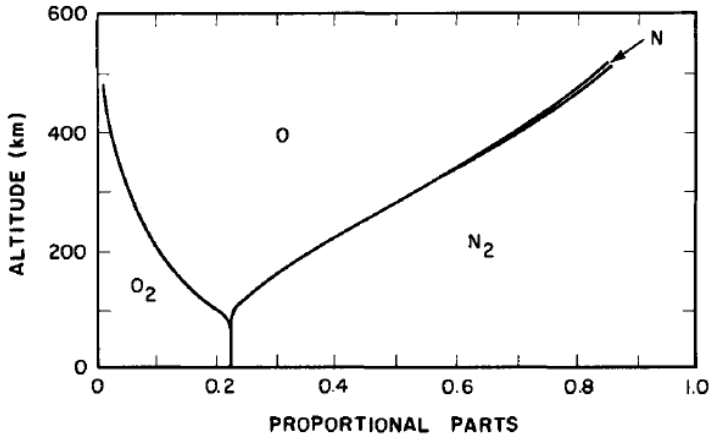


Figure 8. Variation of atmospheric oxygen and nitrogen with altitude (Iqbal).

Diatomic oxygen and nitrogen absorb solar radiation in a number of overlapping bands in wavelengths below $0.20\,\mu m$. Since the upper (beyond 90 km) and lower atmospheres are composed primarily of these four gases (atomic and molecular oxygen and nitrogen), no radiation below $0.20\,\mu m$ reaches the surface of the earth. This is fortunate, because this far-ultraviolet radiation degrades many materials and colors. It is also harmful to human beings.

Diatomic oxygen also has three weak absorption bands in the visible spectrum centered at 0.63, 0.69, and 0.76 μm. However, in this region there is some overlap by ozone bands. ***Ozone*** exhibits a number of absorption bands beyond $0.20\,\mu m$ in the ultraviolet, visible, and near infrared. Ozone has a ***strong absorption band*** from 0.2 to $0.3\,\mu m$, weaker bands from 0.3 to $0.35\,\mu m$, and stronger bands again in the visible region from 0.45 to $0.77\,\mu m$. Ozone also has absorption bands below $0.20\,\mu m$ wavelengths. However, since all radiation below $0.20\,\mu m$ is absorbed by O, N, O_2 , and N_2 before it reaches the stratosphere where ozone exists, we need not be concerned with this range as far as the calculation of irradiance on the earth is concerned.

Absorbers in the Infrared Spectrum

Several gases absorb solar electromagnetic radiation in the infrared wavelengths. Of these gases, the important absorbers are H_2O , CO_2 , O_3 , N_2O , CO, O_2 , CH_4 , and N_2.

The ***water-vapor*** bands are at 0.72, 0.82, 0.94, 1.1, 1.38, 1.87, 2.7, 3.2 and 6.3 μm;

those of ***carbon dioxide*** are at 1.45, 1.6, 2.0, 2.7, 4.3, 4.8, and 5.2 μm;

and those of ***oxygen*** are at 0.69 and 0.76 μm.

Since we are concerned with the solar spectrum at $\lambda < 4\,\mu m$, we can ignore the H_2O and CO_2 bands at higher wavelengths, and the higher wavelength bands of other gases. The concentration of N_2O , CO and CH_4 in the atmosphere is very small, and their total effect on the solar radiation arriving on the earth is minimal.

Transmittances of Molecular Absorbers

To compute ***direct spectral irradiance at the Earth's surface***, we need to determine the spectral transmittance of each atmospheric molecular absorber.

Unlike scattering, molecular absorption varies strongly with wavelength, exhibiting narrow absorption lines and bands. Its accurate calculation requires information about the frequency, intensity, shape, and width of the spectral lines.

Lambert's law is commonly used as an approximation to calculate molecular transmittance. A difficulty is the highly variable spectral structure of molecular absorption. Available absorption

data are often smoothed or averaged and may not correspond exactly to the wavelengths used in extraterrestrial irradiance data. This introduces some uncertainty when calculating spectral transmittance.

For practical calculations, atmospheric molecular absorbers are usually considered in three groups:

1. Ozone treated separately because its atmospheric distribution is relatively well characterized.
2. Other dry-air molecules, such as carbon dioxide, and carbon monoxide, are often approximated as uniformly mixed in the atmosphere.
3. Water vapor, whose transmittance is considered separately.

Transmittance of Ozone

Ozone absorbs solar radiation in the ultraviolet, visible, and infrared regions. Its infrared absorption bands occur at wavelengths longer than 4 μm and are therefore of limited interest in the present context.

For the ***ultraviolet*** and ***visible*** bands, the absorption coefficients reported by Vigroux (1953) are commonly used. These coefficients are expressed in a form suitable for applying Lambert's law.

$$G_{n\lambda} = G_{0n\lambda} exp(-k_\lambda m) \tag{1}$$

For ozone, the transmittance can therefore be written as:

$$\tau_{o\lambda} = \exp(-k_{o\lambda}\, l\, m_r) \tag{2}$$

where $k_{o\lambda}$, is the ozone absorption coefficient, l is the amount of ozone in cm (NTP), and m_r is the relative optical mass for ozone. The attenuation coefficients $k_{o\lambda}$, are given in the next Table.

Spectral Absorption Coefficient $k_{o\lambda}$ for Ozone[a]

λ (μm)	$k_{o\lambda}$ (cm^{-1})	λ (μm)	$k_{o\lambda}$ (cm^{-1})	λ (μm)	$k_{o\lambda}$ (cm^{-1})
0.290	38.000	0.485	0.017	0.595	0.120
0.295	20.000	0.490	0.021	0.600	0.125
0.300	10.000	0.495	0.025	0.605	0.130
0.305	4.800	0.500	0.030	0.610	0.120
0.310	2.700	0.505	0.035	0.620	0.105
0.315	1.350	0.510	0.040	0.630	0.090
0.320	0.800	0.515	0.045	0.640	0.079
0.325	0.380	0.520	0.048	0.650	0.067
0.330	0.160	0.525	0.057	0.660	0.057
0.335	0.075	0.530	0.063	0.670	0.048
0.340	0.040	0.535	0.070	0.680	0.036
0.345	0.019	0.540	0.075	0.690	0.028
0.350	0.007	0.545	0.080	0.700	0.023
0.355	0.000	0.550	0.085	0.710	0.018
0.445	0.003	0.555	0.095	0.720	0.014
0.450	0.003	0.560	0.103	0.730	0.011
0.455	0.004	0.565	0.110	0.740	0.010
0.460	0.006	0.570	0.120	0.750	0.009
0.465	0.008	0.575	0.122	0.760	0.007
0.470	0.009	0.580	0.120	0.770	0.004
0.475	0.012	0.585	0.118	0.780	0.000
0.480	0.014	0.590	0.115	0.790	0.000

These coefficients are valid at 18°C. Although ozone is mainly concentrated in the stratosphere, where temperatures can reach approximately −40°C, the difference between the absorption coefficients at these two temperatures is less than 3%.

The ozone coefficients are listed for wavelengths longer than 0.29 μm because practically all radiation at shorter wavelengths is absorbed by the ozone layer. It is evident from Table that attenuation is very strong in the far-ultraviolet region. It diminishes sharply with wavelength until, at $\lambda \approx 0.35\ \mu m$, ozone becomes almost transparent to solar radiation. The extinction coefficient increases again in the visible region and reaches a maximum value at $\lambda = 0.605\ \mu m$ (in fact precisely at 6038 Ä). Figure 9 contains the plots of $\tau_{o\lambda}$ for various values of lm. These plots show that the parameter lm has a very strong influence on the spectral transmittance at wavelengths shorter than $0.33\ \mu m$. This effect reappears in the visible range.

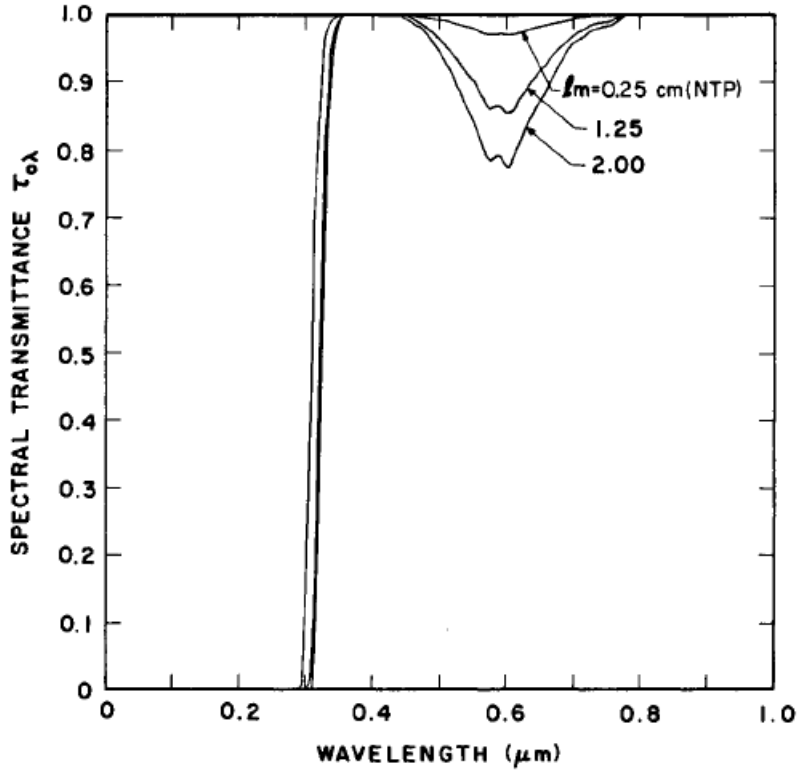


Figure 9. Ozone spectral transmittance (Iqbal).

Transmittances of Uniformly Mixed Gases and Water Vapor

Besides ozone, the main atmospheric absorbers considered here are the remaining dry-air gases and water vapor.

Early measurements of water-vapor absorption bands were performed by Fowle (1915). Although widely used, these data were limited by the instrumentation available at the time and did not account for pressure and temperature effects. Later, Howard et al. (1956) carried out more accurate laboratory measurements of water-vapor and carbon-dioxide absorption under different pressures and optical path lengths.

Gates and Harrop measured atmospheric spectral absorptance under clear-sky conditions in Colorado, including the contributions of water vapor, carbon dioxide, and minor absorbers. Their absorption coefficients were subsequently used by Thekaekara to calculate direct spectral irradiance. However, the mismatch between the wavelengths of these measurements and those of extraterrestrial irradiance data introduced some uncertainty.

To improve atmospheric transmittance calculations, the Air Force Cambridge Research Laboratories (AFCRL) developed more rigorous computational methods. Based on this work, McClatchey et al. developed the LOWTRAN computer code. Leckner subsequently used these results to propose a simplified and accurate algorithm for calculating direct spectral irradiance, suitable for engineering applications.

In this approach, spectral transmittances are provided for 144 wavelength intervals, corresponding to the WRC extraterrestrial spectrum. The absorption due to CO_2, O_2, and other dry-air gases is treated collectively under the term uniformly mixed gases. The spectral transmittance due to uniformly mixed gases is expressed as:

$$\tau_{g\lambda} = \exp\left(\frac{-1.41 k_{g\lambda} m_a}{(1+118.93 k_{g\lambda} m_a)^{0.45}}\right) \tag{3}$$

The corresponding spectral transmittance due to water-vapor absorption is:

$$\tau_{w\lambda} = \exp\left(\frac{-0.2385\, k_{wa\lambda \mathrm{w}}\, m_r}{(1+20.07 k_{wa\lambda \mathrm{w}}\, m_r)^{0.45}}\right) \tag{4}$$

The attenuation coefficients, $k_{g\lambda}$ and $k_{wa\lambda}$, appearing in these equations are listed in the following tables.

Spectral Absorption Coefficients of Uniformly Mixed Gases[a]

λ (μm)	$k_{g\lambda}$	λ (μm)	$k_{g\lambda}$	λ (μm)	$k_{g\lambda}$
0.76	0.300E + 01	1.75	0.100E − 04	2.80	0.150E + 03
0.77	0.210E + 00	1.80	0.100E − 04	2.90	0.130E + 00
		1.85	0.145E − 03	3.00	0.950E − 02
1.25	0.730E − 02	1.90	0.710E − 02	3.10	0.100E − 02
1.30	0.400E − 03	1.95	0.200E + 01	3.20	0.800E + 00
1.35	0.110E − 03	2.00	0.300E + 01	3.30	0.190E + 01
1.40	0.100E − 04	2.10	0.240E + 00	3.40	0.130E + 01
1.45	0.640E − 01	2.20	0.380E − 03	3.50	0.750E − 01
1.50	0.630E − 03	2.30	0.110E − 02	3.60	0.100E − 01
1.55	0.100E − 01	2.40	0.170E − 03	3.70	0.195E − 02
1.60	0.640E − 01	2.50	0.140E − 03	3.80	0.400E − 02
1.65	0.145E − 02	2.60	0.660E − 03	3.90	0.290E + 00
1.70	0.100E − 04	2.70	0.100E + 03	4.00	0.250E − 01

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Spectral Absorption Coefficients of Water Vapor[a]

λ (μm)	$k_{wa\lambda}$ (cm^{-1})	λ (μm)	$k_{wa\lambda}$ (cm^{-1})	λ (μm)	$k_{wa\lambda}$ (cm^{-1})
0.69	0.160E − 01	0.93	0.270E + 02	1.85	0.220E + 04
0.70	0.240E − 01	0.94	0.380E + 02	1.90	0.140E + 04
0.71	0.125E − 01	0.95	0.410E + 02	1.95	0.160E + 03
0.72	0.100E + 01	0.96	0.260E + 02	2.00	0.290E + 01
0.73	0.870E + 00	0.97	0.310E + 01	2.10	0.220E + 00
0.74	0.610E − 01	0.98	0.148E + 01	2.20	0.330E + 00
0.75	0.100E − 02	0.99	0.125E + 00	2.30	0.590E + 00
0.76	0.100E − 04	1.00	0.250E − 02	2.40	0.203E + 02
0.77	0.100E − 04	1.05	0.100E − 04	2.50	0.310E + 03
0.78	0.600E − 03	1.10	0.320E + 01	2.60	0.150E + 05
0.79	0.175E − 01	1.15	0.230E + 02	2.70	0.220E + 05
0.80	0.360E − 01	1.20	0.160E − 01	2.80	0.800E + 04
0.81	0.330E + 00	1.25	0.180E − 03	2.90	0.650E + 03
0.82	0.153E + 01	1.30	0.290E + 01	3.00	0.240E + 03
0.83	0.660E + 00	1.35	0.200E + 03	3.10	0.230E + 03
0.84	0.155E + 00	1.40	0.110E + 04	3.20	0.100E + 03
0.85	0.300E − 02	1.45	0.150E + 03	3.30	0.120E + 03
0.86	0.100E − 04	1.50	0.150E + 02	3.40	0.195E + 02
0.87	0.100E − 04	1.55	0.170E − 02	3.50	0.360E + 01
0.88	0.260E − 02	1.60	0.100E − 04	3.60	0.310E + 01
0.89	0.630E − 01	1.65	0.100E − 01	3.70	0.250E + 01
0.90	0.210E + 01	1.70	0.510E + 00	3.80	0.140E + 01
0.91	0.160E + 01	1.75	0.400E + 01	3.90	0.170E + 00
0.92	0.125E + 01	1.80	0.130E + 03	4.00	0.450E − 02

[a] Reprinted with permission from Leckner [6], Copyright 1978, Pergamon Press, Ltd.

Virtually all radiation is absorbed at $\lambda > 4.0\ \mu m$. Furthermore, only 1% of extraterrestrial solar irradiance is contained in wavelengths greater than $4.0\ \mu m$; consequently, these tables are limited to this wavelength.

In Eqs. (3) and (4), m_a and m_r are the optical masses for uniformly mixed gases and water vapor, respectively, and w is precipitable water vapor in centimeters.

The spectral transmittances due to uniformly mixed gases and water vapor are shown in Figs. 10 and 11. These figures demonstrate very low transmittance values in some of the infrared bands. In these figures an optical mass of unity is used. At higher optical masses, the transmittance will further reduce considerably.

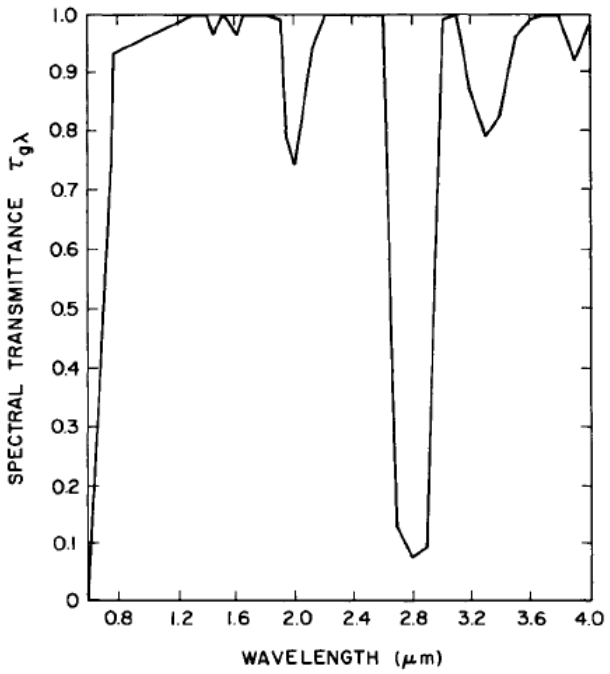


Figure 10. Spectral transmittance of the uniformly mixed gases (Iqbal), $m = 1$.

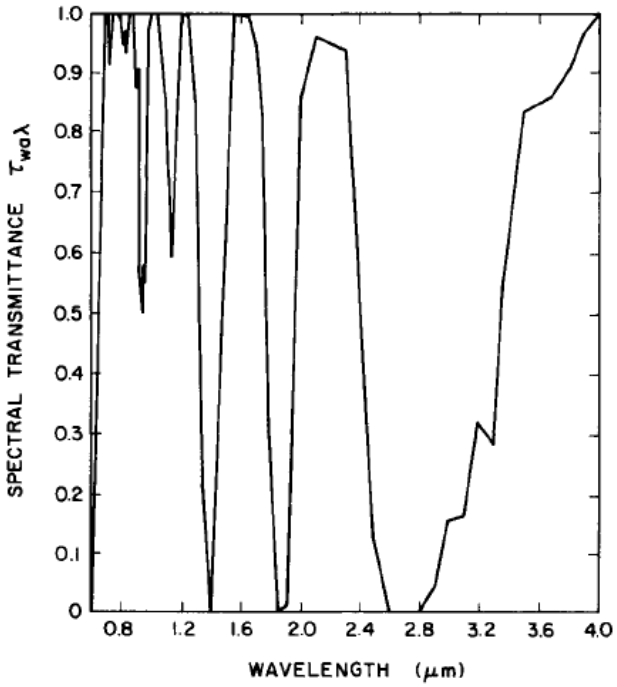


Figure 11. Spectral transmittance of water vapor (Iqbal), $m = 1, w = 2\ cm$.

At certain wavelengths, approximately 1.2, 1.6, 2.2, and 3.8 μm, atmospheric transmittance is close to unity. Consequently, most of the incoming solar radiation within these spectral regions reaches the Earth's surface with little attenuation. These relatively narrow spectral intervals are known as atmospheric windows, where the atmosphere is nearly transparent to radiation.

Atmospheric windows also allow part of the thermal radiation emitted by the Earth's surface to escape into space. This transmission of terrestrial infrared radiation plays an important role in the Earth's energy balance and thermal regulation.

The total transmittance due to the molecular absorbers can now be written as follows:

$$\tau_{ma\lambda} = \tau_{o\lambda}\, \tau_{g\lambda}\, \tau_{wa\lambda} \tag{5}$$

Example: Spectral transmittances evaluated at λ = 0.7 and 2.0 μm due to molecular absorbers under cloudless-sky conditions and assuming zenith angle 60°, precipitable water vapor 2 cm, and ozone layer thickness 0.3 cm (NTP).

$\theta_z = 60°$ and $m = 2.0$

a) $\lambda = 0.7\ \mu m$

$$k_{o\lambda} = 0.023\ cm^{-1}$$

$$k_{g\lambda} = 0 \quad \rightarrow \quad \tau_{g\lambda} = 1$$

$$k_{wa\lambda} = 0.024\ cm^{-1}$$

Transmittance due to absorption by ozone:

$$\tau_{o\lambda} = \exp(-k_{o\lambda}\, l\, m_r\,) = \exp(-0.023\ \times 0.3\ \times 2.0) = 0.986$$

Transmittance due to absorption by water vapor:

$$\tau_{w\lambda} = \exp\left(\frac{-0.2385\ k_{wa\lambda \mathrm{w}}\, m_r}{(1+20.07 k_{wa\lambda \mathrm{w}}\, m_r)^{0.45}}\right) = \exp\left(\frac{-0.2385\times 0.024\times 2\times 2}{(1+20.07\times 0.024\times 2\times 2\,)^{0.45}}\right) = 0.985$$

Total transmittance due to molecular absorbers:

$$\tau_{ma\lambda} = \tau_{O\lambda}\, \tau_{g\lambda}\, \tau_{wa\lambda} = 0.986\ \times 1\ \times 0.985 = 0.972$$

```
tau_o<-function(k,l,m)
+     return<-exp(-k*l*m)
>   TauO<-tau_o(0.023,0.3,2.0)
>   TauO
[1] 0.9862948
>   tau_w<-function(k,w,m)
+     return<-exp((-0.2385*k*w*m)/(1+20.07*k*w*m)^0.45)
>   TauW<-tau_w(0.024,2,2)
>   TauW
[1] 0.9859775
>   Tau_abs<-TauO*TauW
>   Tau_abs
[1] 0.9724645
```

b)

$\lambda = 2.0\ \mu m$

$$k_{o\lambda} = 0 \quad \rightarrow \quad \tau_{o\lambda} = 1$$

$$k_{g\lambda} = 3$$

$$k_{wa\lambda} = 2.9\ cm^{-1}$$

$$\tau_{g\lambda} = \exp\left(\frac{-1.41 k_{g\lambda} m_a}{(1+118.93 k_{g\lambda} m_a)^{0.45}}\right) = \exp\left(\frac{-1.41\times3\times2}{(1+118.93\times3\times2)^{0.45}}\right) = 0.644$$

$$\tau_{w\lambda} = exp\left(\frac{-0.2385\ k_{wa\lambda} \mathrm{w}\ m_r}{(1+20.07 k_{wa\lambda} \mathrm{w}\ m_r)^{0.45}}\right) = exp\left(\frac{-0.2385\ 2.9\times2\times\ 2}{(1+20.07\times2.9\times2\times\ 2)^{0.45}}\right) = 0.788$$

$$\tau_{ma\lambda} = \tau_{O\lambda}\, \tau_{g\lambda}\, \tau_{wa\lambda} = 1 \times 0.644 \times 0.788 = 0.507$$

```
tau_g<-function(k,m)
+     return<-exp((-1.41*k*m)/(1+118.3*k*m)^0.45)
>   TauG<-tau_g(3,2)
>   TauG
[1] 0.6436243
>   TauW<-tau_w(2.9,2,2)
>   TauW
[1] 0.7884706
>   Tau_abs<-TauW*TauG
>   Tau_abs
[1] 0.5074789
```

At 2.0 μm, $\tau_{ma\lambda}$ is about half of its value at 0.7 μm .

Direct Spectral Irradiance on the Ground

Monochromatic distribution of a direct solar beam can now be computed as a function of a few variables, including optical mass and a wide variety of atmospheric parameters, for example, water-vapor content, ozone layer thickness, and turbidity parameters. With equations

$$\tau_{c\lambda} = \tau_{r\lambda}\, \tau_{a\lambda} \quad \text{and} \quad \tau_{ma\lambda} = \tau_{O\lambda}\, \tau_{g\lambda}\, \tau_{wa\lambda} \tag{6}$$

combined, the monochromatic transmittance due to the combined effects of continuum attenuation and molecular absorption is as follows:

$$\tau_\lambda = \tau_{c\lambda}\ \tau_{ma\lambda} = \tau_{r\lambda}\, \tau_{a\lambda}\, \tau_{O\lambda}\, \tau_{g\lambda}\, \tau_{wa\lambda} \tag{7}$$

$\tau_{r\lambda}$ is the spectral transmittance of the direct beam due to molecular scattering

$\tau_{a\lambda}$ is the spectral transmittance of the direct beam due to scattering and absorption by aerosols

$\tau_{O\lambda}$ is the spectral transmittance of the direct beam due to absorption by the ozone layer

$\tau_{g\lambda}$ is the spectral transmittance of the direct beam due to absorption by uniformly mixed gases such as CO_2 and O_2

$\tau_{wa\lambda}$ is the spectral transmittance of the direct beam due to absorption by water vapor

To obtain an overall view, all these individual transmittances have been plotted simultaneously in Fig. 12.

It is apparent from this figure that ***infrared absorption*** is due mainly to water vapor and uniformly mixed gases.

In the ***ultraviolet and visible*** region, it is essentially ozone absorption, Rayleigh scattering, and aerosols that control attenuation of the direct beam.

The transmittance by aerosols is minimum at the short wavelengths and increases slowly as the wavelength increases. However, the rate of attenuation by aerosols will change if turbidity parameters are altered.

Furthermore, in this diagram the transmittance curves due to Rayleigh and aerosols are drawn only beyond that of the ozone absorption curve.

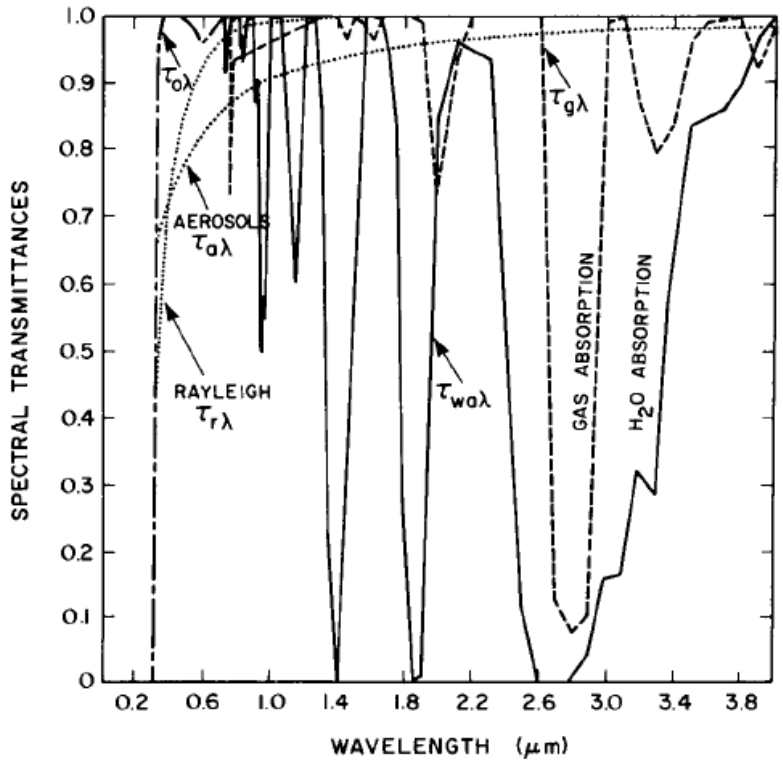
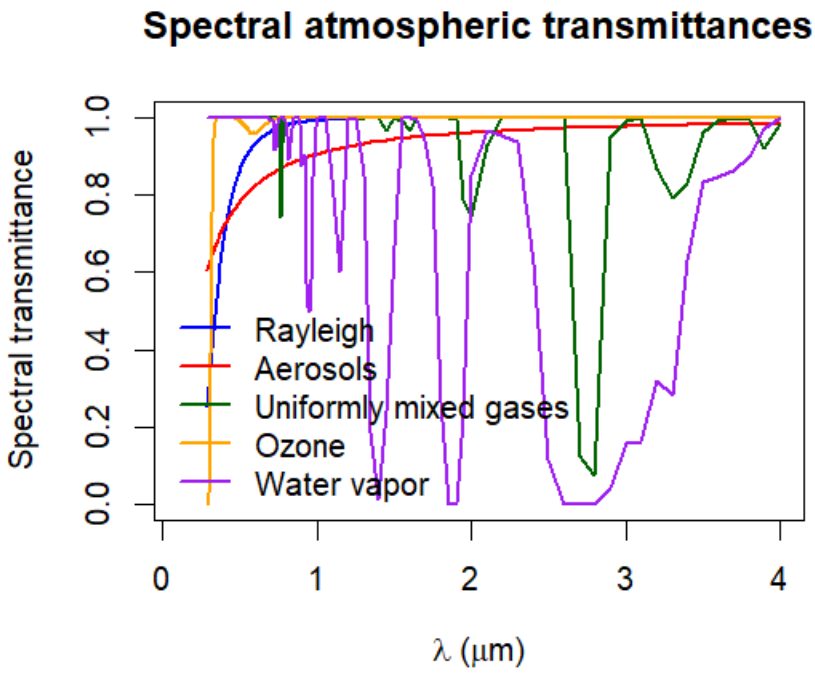


Figure 12. Spectral transmittance of the various atmospheric constituents. $m = 1$, $l_{O3} = 0.35\ cm$, $w = 2cm$, $\alpha = 1.3$ and $\beta = 0.1$ (Iqbal).




```
##direct spectral transmitance
>   lo3<-0.35
>   oam<-1
>   wv<-2
>   alfa<-1.3
>   beta<-0.1

>   # Transmitanta Rayleigh
>   tau_r <- function(lambda, m) {
+     exp(-0.008735 * m * lambda^(-4.08))
+   }
>   # Transmitanta aerosolilor
>   tau_a <- function(lambda, alpha, b, m) {
+     exp(-m * b * lambda^(-alpha))
+   }
>   ##transmitanta mixed gases
>   tau_g<-function(k,m)
+     return(exp((-1.41*k*m)/(1+118.3*k*m)^0.45))
>   ##transmitanta ozonului
>   tau_o<-function(k,l,m)
+     return(exp(-k*l*m))
>   tau_w<-function(k,w,m)
+     return(exp((-0.2385*k*w*m)/(1+20.07*k*w*m)^0.45))

>   ## Citirea coeficientilor spectrali Leckner
>   ## Iradianta spectrala extraterestra trebui X 10 pt W/m2 micro m
>   coef <- read.table("ETR_CoefAbs_Leckner_Davos_Smart2.txt",
+                      header = TRUE)
>   ## Lungimea de unda
>   lambda <- coef$wave

>   ## Calculul transmitantelor
>   Tr <- tau_r(lambda, oam)
>   Ta <- tau_a(lambda, alfa, beta, oam)
```

```
>   Tg <- tau_g(coef$Kg, oam)
>   To <- tau_o(coef$Ko, lo3, oam)
>   Tw <- tau_w(coef$Kw, wv, oam)

>   ## Grafic
>   plot(lambda, Tr, type = "l", lwd = 2, col = "blue",
+        xlim = c(0.1, 4), ylim = c(0, 1),
+        xlab = expression(lambda~"("*mu*"m)"),
+        ylab = "Spectral transmittance",
+        main = "Spectral atmospheric transmittances")
>   lines(lambda, Ta, col = "red", lwd = 2)
>   lines(lambda, Tg, col = "darkgreen", lwd = 2)
>   lines(lambda, To, col = "orange", lwd = 2)
>   lines(lambda, Tw, col = "purple", lwd = 2)
>   legend("bottomleft",
+          legend = c("Rayleigh", "Aerosols",
+                     "Uniformly mixed gases", "Ozone",
+                     "Water vapor"),
+          col = c("blue", "red", "darkgreen",
+                  "orange", "purple"),
+          lwd = 2, bty = "n")
```

Direct spectral irradiance at any wavelength on a surface normal to the sun's rays and at mean sun-earth distance is as follows:

$$G_{n\lambda} = G_{0n\lambda}\,\tau_{\lambda} \tag{8}$$

and on a horizontal surface,

$$G_{n\lambda} = G_{0n\lambda}\,\tau_{\lambda}\cos\theta_z \tag{9}$$

The extraterrestrial spectrum is modified by atmospheric constituents before it reaches the ground. We shall use the extraterrestrial spectrum from the World Radiation Center only. It seems appropriate to show an example of differences between the results obtained by using the WRC and the NASA spectra. Such an example for a ***typical atmosphere*** $\beta = 0.1$, $\alpha = 1.3$, $l_{O3} = 0.35\ cm$, w = 2 cm, and $\theta_z = 60^{\circ}$ is given in Fig. 13. The WRC spectrum results in higher amounts of irradiance, particularly in the visible portion. For this reason, the WRC spectrum will be used throughout this chapter. The spectrally integrated (called broadband) values of the direct irradiances are 687 W/m^2 for the WRC and 672 W/m^2 for the NASA spectrum. (2% difference)

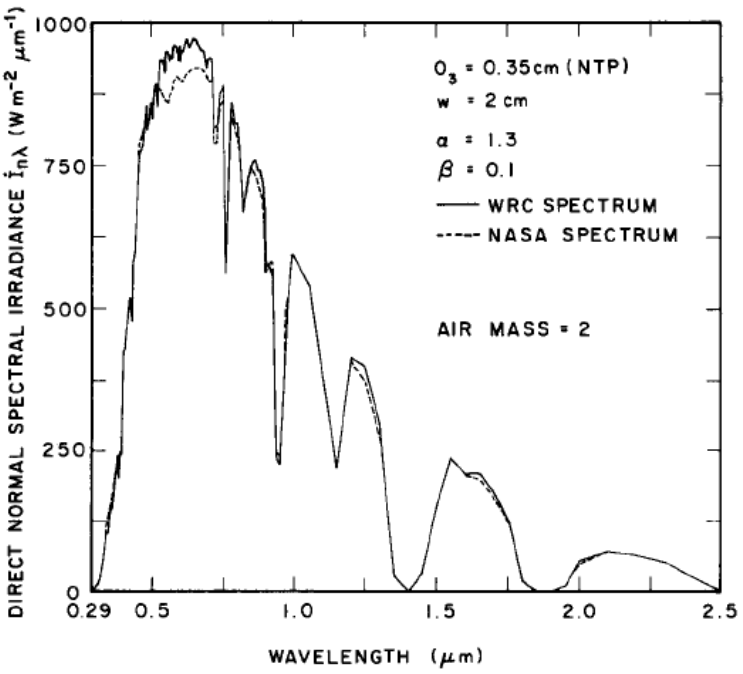
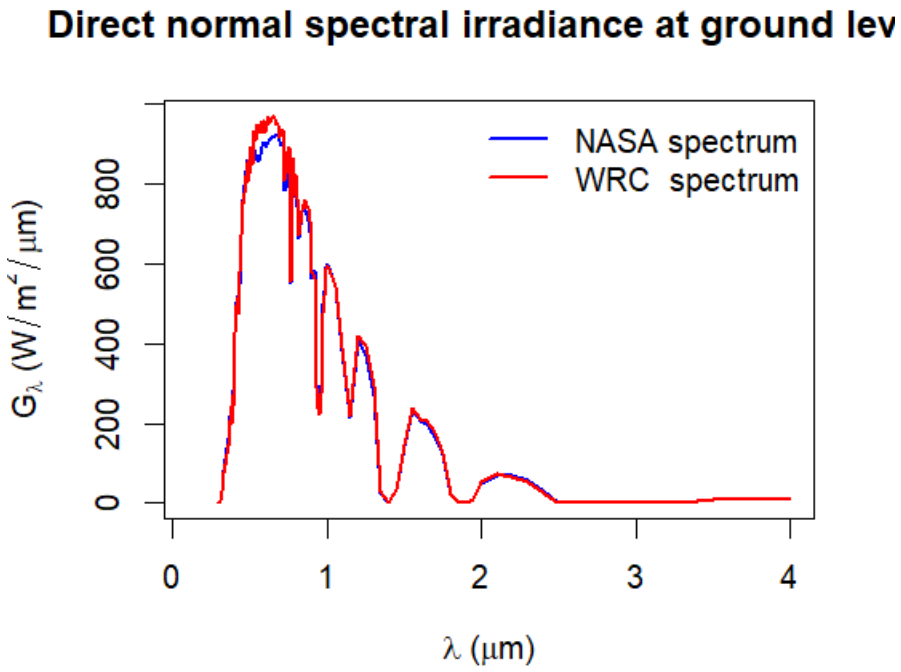


Figure 13. Comparison of the direct normal spectral irradiance using WRF spectrum and the NASA spectrum, $m = 2$, $l_{O3} = 0.35\ cm$, $w = 2cm$, $\alpha = 1.3$ and $\beta = 0.1$ (Iqbal).




```
###spectru la nivelul solului, NASA si WRC
>   ## Parametrii atmosferici
```

```
>   lo3 <- 0.35
>   oam <- 2
>   wv <- 2
>   alfa <- 1.3
>   beta <- 0.1

>   ## Lungimea de unda
>   lambda <- coef$wave

>   ## Calculul celor cinci transmitante
>   Tr <- tau_r(lambda, oam)
>   Ta <- tau_a(lambda, alfa, beta, oam)
>   Tg <- tau_g(coef$Kg, oam)
>   To <- tau_o(coef$Ko, lo3, oam)
>   Tw <- tau_w(coef$Kw, wv, oam)

>   ## Transmitanta atmosferica totala
>   Ttot <- Tr * Ta * Tg * To * Tw

>   ## Iradianta spectrala directa normala la sol

>   ## folosind spectrul extraterestru NASA
>   DNI_NASA <- 10*coef$Gextw * Ttot

>   ## folosind spectrul extraterestru WRC
>   DNI_WRC <- 10*coef$Gext_D * Ttot

>   ## Reprezentarea grafica
>   plot(lambda, DNI_NASA,
+        type = "l", lwd = 2, col = "blue",
+        xlim = c(0.1, 4),
+        ylim = c(0, max(DNI_NASA, DNI_WRC, na.rm = TRUE)),
+        xlab = expression(lambda~"("*mu*"m)"),
+        ylab = expression(G[lambda]~"("*W/m^2/mu*m*")"),
+        main = "Direct normal spectral irradiance at ground level")
>   lines(lambda, DNI_WRC,
+         col = "red", lwd = 2)
>   legend("topright",
+          legend = c("NASA spectrum",
+                     "WRC  spectrum"),
+          col = c("blue", "red"),
+          lwd = 2, bty = "n")
```

In Fig. 14 we identify the various molecular absorption bands. In this diagram we consider a clean atmosphere $(\beta = 0)$ with $l_{O3} = 0.35\,cm$ (NTP), w = 2 cm. The first curve under the extraterrestrial line represents depletion of the direct beam by Rayleigh scattering only. The next curve below represents the actual amount of direct solar flux reaching the ground. The amount of energy absorbed by the various gases is shown by the darkened areas. The size of the shaded areas (except those for O_2 and CO_2) depends on the amount of O_3 and H_20 present in the atmosphere. Effect of the zenith angle (air mass) applies to all areas.

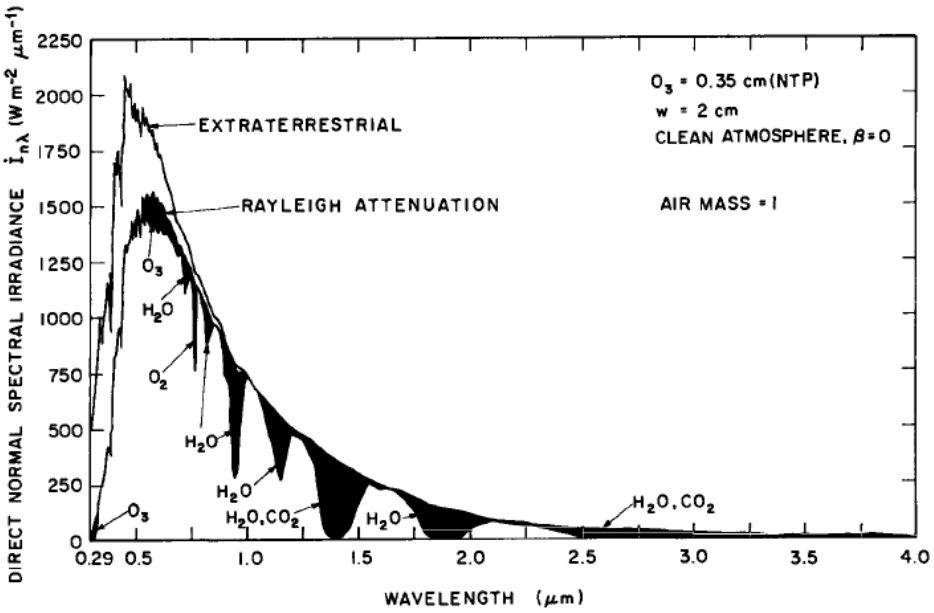


Figure 14. Identification of various molecular absorbers

The effect of a variation in ozone only is shown on Fig. 15. This diagram represents clean dry atmosphere (β = w = 0), and l_{O3} = 0, 0.2, and 0.4 cm. Attenuation by ozone is confined to the ultraviolet and the visible spectrum and is almost linear with amount of ozone.

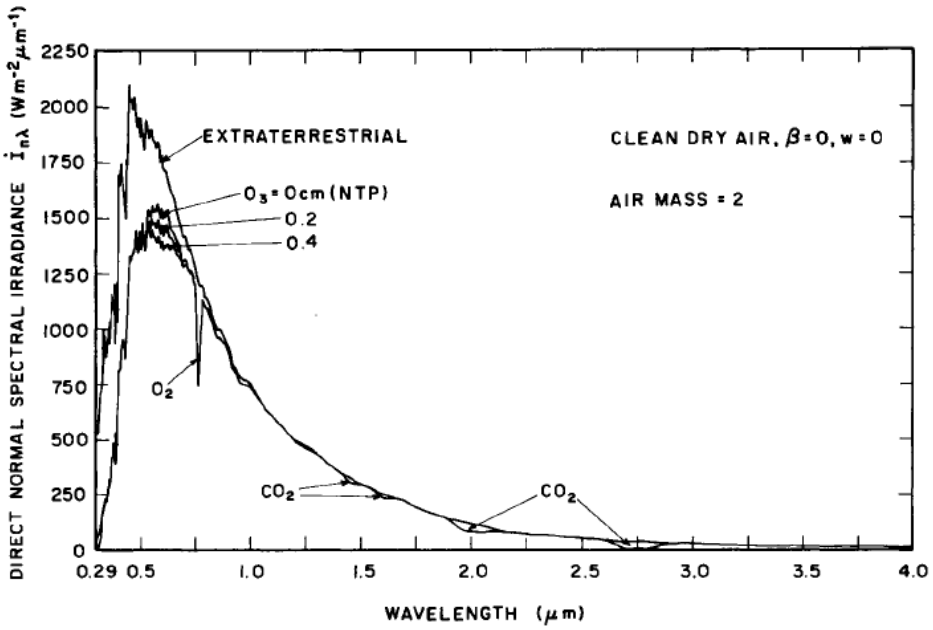


Figure 15. Direct normal spectral irradiance under varying amounts of ozone

Depletion of the extraterrestrial beam by a varying amount of water vapor is demonstrated in Fig. 16. Again, this diagram is for a clean atmosphere. Ozone is assumed to be zero, and absorption by oxygen and carbon dioxide is also neglected. As such, this diagram represents Rayleigh scattering and absorption by water vapor only. Under the extraterrestrial curve, the three irradiance curves are for w = 0, 0.4, and 4 cm of precipitable water. It is evident from this diagram that the rate of attenuation diminishes as the amount of precipitable water increases. The first few millimeters deplete as much energy as the last few centimeters of precipitable water. Furthermore, even a small amount of moisture in the atmosphere shuts off all radiation beyond approximately 2.5 μm. In this example, spectrally integrated irradiances are 1142, 1025, and 934 W/m^2 at w = 0, 0.4, and 4.0 cm. The first 4 mm of water vapor depletes by 10% the energy reaching ground, and it takes a further 3.6 cm of water vapor to deplete another 10% of the energy.

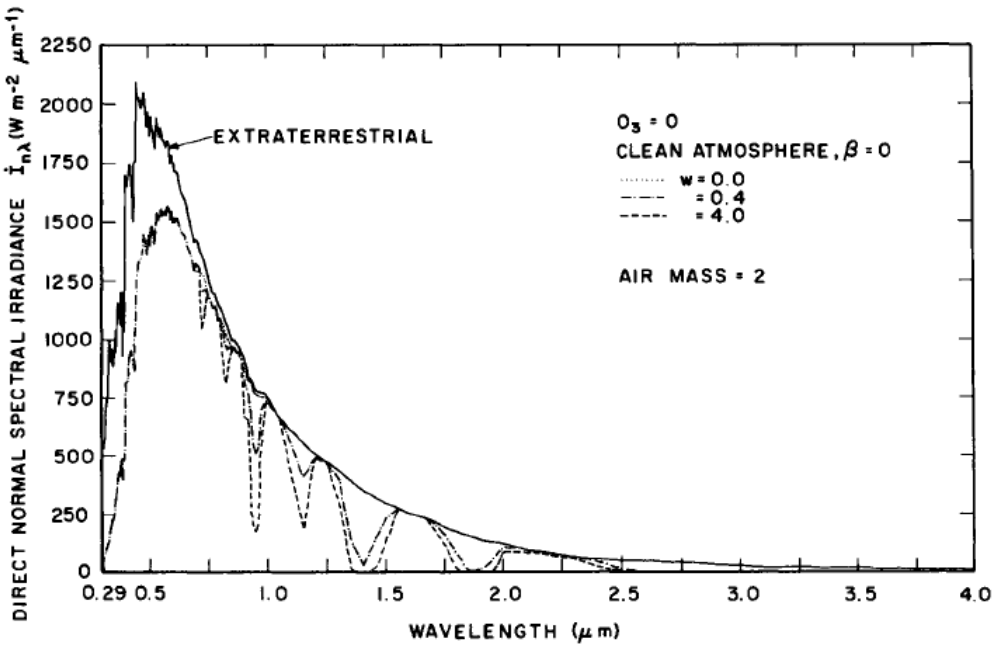


Figure 16. Direct normal spectral irradiance under varying amounts of water vapor.

For a typical atmosphere [β = 0.1, α = 1.3, l_{O3} = 0.35 cm, and w = 2 cm], the effect of the air mass is shown in Fig. 17. As the air mass increases, the ultraviolet and visible spectrum undergoes much stronger depletion (by absorption and scattering) than the infrared portion. For this reason, the sun appears red when it is near the horizon.

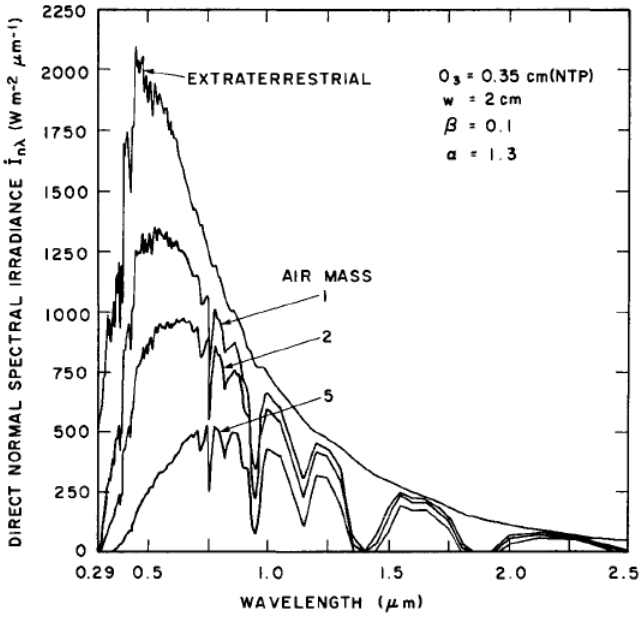


Figure 17. Direct normal spectral irradiance under different air masses.

The effect of variation in the turbidity parameter β (representing the amount of aerosols present), and of the Angstrom exponent α (representing average particle size), is shown in Figs. 18 and 19, respectively. The turbidity parameter β has a strong influence on the spectral flux reaching the ground. An increase in β diminishes irradiance. The wavelength exponent α, has only a minor influence on spectral irradiance. Irradiance at $\lambda > 1\ \mu m$ remains unchanged by a change in α. By keeping total amount of aerosols constant (β = const), smaller values of α (representing larger particle size) yield more irradiance.

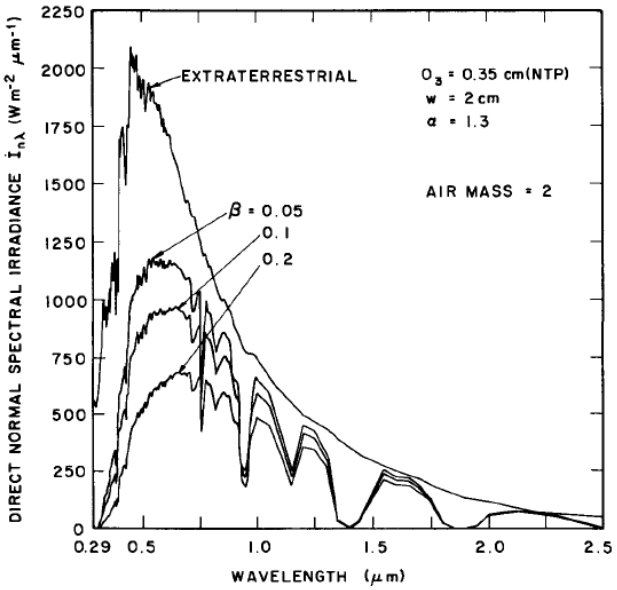


Figure 18. Direct normal spectral irradiance under different aerosol amounts.

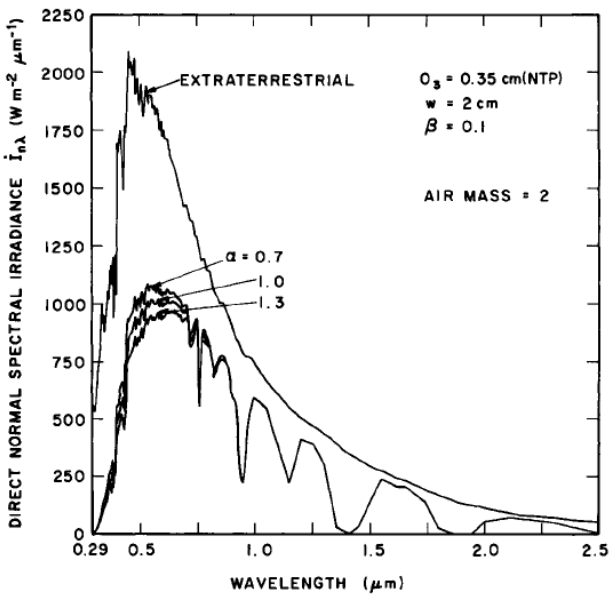


Figure 19. Direct normal spectral irradiance under varying values of α .

In the formulation

$$\tau_\lambda = \tau_{c\lambda}\ \tau_{ma\lambda} = \tau_{r\lambda}\,\tau_{a\lambda}\,\tau_{O\lambda}\,\tau_{g\lambda}\,\tau_{wa\lambda} \tag{10}$$

it is understood that at a single point in the vertical direction, attenuation processes occur simultaneously. However, processes dominant in the short wave such as ***ozone absorption*** first occur high in the atmosphere. The remaining energy in the direct beam then begins to be attenuated by ***Rayleigh scattering***, which occurs predominantly in the short wave and in the upper layers of the atmosphere. Further down toward the ground, the direct beam is attenuated by water vapor, uniformly mixed gases, and aerosols, all of which occur essentially near the earth's surface. A general picture of the depletion of energy within various bandwidths and at various heights above sea level is shown in Table.

Attenuation of Solar Radiation by the Earth's Atmosphere[a]

Pressure (mbars)	Altitude (km)	Wavelength regions (μm) 0.12–0.20	0.20–0.29	0.29–0.32	0.32–0.35	0.35–0.55	0.55–0.9	0.9–2.5	2.5–7	7–20	Altitude (km)	
0.2	60	O_2 absorbs almost completely	Attenuation by scattering increases markedly toward shorter wavelengths								Above 60	110 km
7.5	33		(0.20–0.21 μm absorption by O_2) Absorption by O_3 appreciable	O_3 absorption not important					Energy small	Energy very small	60–33	Chemosphere
227	11		No radiation penetrates below about 11 km	O_3 absorption attenuates more than loss by scattering	O_3 absorption significantly attenuates radiation	Irradiation diminished mostly by scattering by permanent gases in atmosphere	H_2O responsible for major absorption; CO_2 absorbs slightly at 2 μm; water vapor (or ice crystals) is found up to about 21 km		No significant penetration below 2 km except in "windows" at approximately 3.8 and 4.9 μm	Strong O_3 absorption at 9.6 μm; strong CO_2 absorption 12–17 μm	33–11	20 km Stratosphere 11 km
795	2				Highly-variable dust, haze (H_2O and smoke) responsible for attenuation in regions 0.32–0.7 μm		Energy transmitted with small loss down to 2 km	Energy penetrates to sea level only through "windows" at approximately 1.2, 1.6, and 2.2 μm		Energy transmited with moderate loss; many absorption bands due to atmospheric gases	11–2	Troposphere
1013			Sea level	Appreciable penetration through "clear" atmosphere to sea level	Penetration through "clear" atmosphere to sea level about 40%	Dust may rise to more than 4 km					2–sea level	
				About 7%	About 30%							0 km

[a] From "Handbook of Geophysics and Space Environments" by J. N. Howard, J. I. F. King, and P. R. Gast. Copyright © 1961 Macmillan Publishing Co., Inc.

Example 1: The direct spectral irradiance on a horizontal surface on 15 June under the conditions of $\lambda = 0.7$ and $2.0\ \mu m$ under cloudless-sky conditions and assuming zenith angle 60°, $\beta = 0.1$, and $\alpha = 1.3$, precipitable water vapor 2 cm, and ozone layer thickness 0.3 cm.

The direct normal spectral irradiance on an individual day is given by

$$G_{n\lambda} = G_{0n\lambda}\, \varepsilon_j \prod_{i=1}^{i=j} \tau_{i\lambda} \qquad (11)$$

$$Where\ \varepsilon_j = 0.9684\ for\ 15\ June, J = 166.$$

a)$\lambda = 0.7\ \mu m$

Din spectrul WRC_Davos $G_{0n\lambda} = 1427.5\ Wm^{-2}\mu m^{-1}$

$$k_{o\lambda} = 0.023\ cm^{-1}$$

$$k_{g\lambda} = 0 \quad \rightarrow \quad \tau_{g\lambda} = 1$$

$$k_{wa\lambda} = 0.024\ cm^{-1}$$

$$G_{n\lambda} = 1427.5 \times 0.9684 \times \tau_{c\lambda} \times \tau_{ma\lambda}$$

$$= 1427.5 \times 0.9684 \times 0.675 \times 0.972 =$$

$907.55\ Wm^{-2}\mu m^{-1}$

Beam irradiance on a horizontal surface is

$$G_{b\lambda} = G_{n\lambda} \cos\theta_z = 907.55 \times 0.5 = 453.77 W m^{-2} \mu m^{-1}$$

```
J<-166 ##15 June
>   alfa<-1.3
>   beta<-0.1
>   lo<-0.3
>   w<-2
>   Corectie_Spencer_excentr<-function(x)
+     return<-(1.00011+0.034221*cos(2*pi*(x-1)/365)+0.00128*sin(2*pi*(x-1)/36
5)+0.000719*cos(2*2*pi*(x-1)/365)+0.000077*sin(2*2*pi*(x-1)/365))
>   ej<-Corectie_Spencer_excentr(J)
>   ej
[1] 0.9683586
>   la<-0.7 ##micro metri
>   G0n_07<-1427.5  ##din WRC
>   TauR<-tau_r(la,2)
>   TauR
[1] 0.9278665
>   TauA<-tau_a(la,alfa,beta,2)
>   TauA
[1] 0.7276159
>   TauA*TauR
[1] 0.6751304
>   TauO<-tau_o(0.023,lo,2.0)
>   TauO
[1] 0.9862948
>   TauG<-tau_g(0,2)
>   TauG
[1] 1
>   TauW<-tau_w(0.024,w,2)
>   TauW
[1] 0.9859775
>   TauO*TauG*TauW
[1] 0.9724645
>   Gn_07<-G0n_07*ej*TauA*TauR*TauO*TauG*TauW
>   #Gn_07<-G0n_07*ej*tau_r(0.7,2)*tau_a(0.7,1.3,0.1,2)*tau_o(0.023,0.3,2.0)*
tau_w(0.024,2,2)*tau_g(0,2)
>   Gn_07
[1] 907.5567
>   Gn_07*0.5
[1] 453.7783
```

b) $\lambda = 2.0\ \mu m$

$G_{0n\lambda} = 118.5\ W m^{-2} \mu m^{-1}$

$k_{o\lambda} = 0 \rightarrow \tau_{o\lambda} = 1$

$k_{g\lambda} = 3$

$k_{wa\lambda} = 2.9\ cm^{-1}$

$$G_{n\lambda} = 118.5 \times 0.9684 \times \tau_{c\lambda} \times \tau_{ma\lambda} = 53.63 Wm^{-2}\mu m^{-1}$$

$$G_{b\lambda} = I_{n\lambda} \cos\theta_z = 53.69 \times 0.5 = 26.81 Wm^{-2}\mu m^{-1}$$

```
la<-2.0 ##micro metri
>   G0n_2<-118.5 ##din spectru WRC
>   TauR<-tau_r(la,2)
>   TauR
[1] 0.9989676
>   TauA<-tau_a(la,alfa,beta,2)
>   TauA
[1] 0.921986
>   TauO<-tau_o(0.0,lo,2.0)
>   TauO
[1] 1
>   TauG<-tau_g(3,2)
>   TauG
[1] 0.6436243
>   TauW<-tau_w(2.9,w,2)
>   TauW
[1] 0.7884706
>   Gn_2<-G0n_2*ej*TauA*TauR*TauO*TauG*TauW
>   #Gn_2<-G0n_2*ej*tau_r(2.0,2)*tau_a(2.0,1.3,0.1,2)*tau_o(0.0,0.3,2.0)*tau_
w(2.9,2,2)*tau_g(3,2)
>   Gn_2
[1] 53.635
>   Gn_2*0.5
[1] 26.8175
```

Example 2: The direct spectral irradiance on a horizontal surface on 15 June under the conditions of λ = 0.7 and 2.0 μm under cloudless-sky conditions and assuming zenith angle 60°, β = 0.1, and α = 0.7, precipitable water vapor 2 cm, and ozone layer thickness 0.3 cm.

a)$\lambda = 0.7\ \mu m$

Din spectrul WRC_Davos $G_{0n\lambda}$ =1427.5 $Wm^{-2}\mu m^{-1}$

$\tau_{r\lambda}$ =0.927 $\tau_{a\lambda}$ =0.773 $\tau_{o\lambda} = 0.986$ $\tau_{g\lambda} = 1$ $\tau_{w\lambda}$ =0.985

$$G_{n\lambda} = 964.89\ Wm^{-2}\mu m^{-1}$$

$$G_{b\lambda} = G_{n\lambda} \cos\theta_z = 964.89 \times 0.5 = 482.44 Wm^{-2}\mu m^{-1}$$

With α changing from 1.3 to 0.7, there is a slight increase in $G_{n\lambda}$.

```
alfa<-0.7
>   la<-0.7 ##micro metri
>   G0n_07<-1427.5  ##din WRC
>   TauR<-tau_r(la,2)
>   TauR
[1] 0.9278665
>   TauA<-tau_a(la,alfa,beta,2)
>   TauA
```

```
[1] 0.773584
>   TauO<-tau_o(0.023,lo,2.0)
>   TauO
[1] 0.9862948
>   TauG<-tau_g(0,2)
>   TauG
[1] 1
>   TauW<-tau_w(0.024,w,2)
>   TauW
[1] 0.9859775
>   Gn_07<-GOn_07*ej*TauA*TauR*TauO*TauG*TauW
>   #Gn_07<-GOn_07*ej*tau_r(0.7,2)*tau_a(0.7,1.3,0.1,2)*tau_o(0.023,0.3,2.0)*
tau_w(0.024,2,2)*tau_g(0,2)
>   Gn_07
[1] 964.8928
>   Gn_07*0.5
[1] 482.4464
```

a) $\lambda = 2.0\ \mu m$

$G_{0n\lambda} = 118.5\ Wm^{-2}\mu m^{-1}$

$\tau_{r\lambda} = 0.998 \quad \tau_{a\lambda} = 0.884 \quad \tau_{o\lambda} = 1 \quad \tau_{g\lambda} = 0.643 \quad \tau_{w\lambda} = 0.788$

$G_{n\lambda} = 51.43\ Wm^{-2}\mu m^{-1} \quad G_{b\lambda} = 25.71\ Wm^{-2}\mu m^{-1}$

With α changing from 1.3 to 0.7, there is a slight decrease $G_{n\lambda}$.

```
la<-2.0 ##micro metri
>   GOn_2<-118.5 ##din spectru WRC
>   TauR<-tau_r(la,2)
>   TauR
[1] 0.9989676
>   TauA<-tau_a(la,alfa,beta,2)
>   TauA
[1] 0.8841625
>   TauO<-tau_o(0.0,lo,2.0)
>   TauO
[1] 1
>   TauG<-tau_g(3,2)
>   TauG
[1] 0.6436243
>   TauW<-tau_w(2.9,w,2)
>   TauW
[1] 0.7884706
>   Gn_2<-GOn_2*ej*TauA*TauR*TauO*TauG*TauW
>   #Gn_2<-GOn_2*ej*tau_r(2.0,2)*tau_a(2.0,1.3,0.1,2)*tau_o(0.0,0.3,2.0)*tau_
w(2.9,2,2)*tau_g(3,2)
>   Gn_2
[1] 51.43467
>   Gn_2*0.5
[1] 25.71734
```

EXAMPLE .3. The turbidity parameter β determined on 15 June, when the sun is 30° above the horizon, at a location where a single-wavelength sun photometer reads 600 W /m^2 μm^{-1} at $\lambda = 1\ \mu m$; the estimated amount of precipitable water in the atmosphere is 1.0 cm and ozone 0.30 cm; ambient temperature is 25°C and pressure is 1025 mbars.

$\theta_z = 60^{\circ}$,

$$m_r = [\cos\theta_z + 0.15(93.885 - \theta_z)^{-1.253}]^{-1}$$

$m_r = 1.9927$,

$$m_a = m_r(p/1013.25)$$

$m_a = 2.0158$

For water vapor we need to carry out pressure and temperature correction

$$w = w' \left(\frac{p}{1013.25}\right)^{3/4} \left(\frac{273}{T}\right)^{1/2}$$

w'=1 cm, $w = 0.965$

$G_{0n\lambda} = 743.99$

$k_{o,1} = 0,\quad k_{g,1} = 0,\quad k_{w,1} = 0.0025 \quad \rightarrow \quad \tau_{g\lambda} = 1 \quad \tau_{o\lambda} = 1 \quad \tau_{w\lambda} \cong 1$

τ_r =0.982553

$G_{n\lambda}$ =600= $743.99 \times 0.9684 \times 0.982553 \times \tau_a$

$$\tau_{a\lambda} = \exp(-\beta\lambda^{-\alpha} m_a) \qquad \tau_{a\lambda} = \exp(-\beta m_a)$$

$\tau_a = 0.847565$ β= 0.08207788

```
 teta_z<-60
>   mr<-function(x)
+     return<-(cos(x*pi/180)+0.15*(93.885-x)^(-1.253))^(-1)
>   m_r<-mr(teta_z)
>   m_r
[1] 1.992764
```

```
>   ma<-function(m,p)
+     return<-m*(p/1013.25)

>   pres<-1025
>   m_a<-ma(m_r,pres)
>   m_a
[1] 2.015873

>   w_prec<-1 ##cm
>   Temp<-25+273.15

>   water<-function(p,T)
+     return<-w_prec*(p/1013.25)^0.75*(273/T)^0.5
>   w<-water(pres,Temp)
>   w
[1] 0.9652045

>   G0n_1<-743.99  ##WRC

>   TauR<-tau_r(1.0,m_a)
>   TauR
[1] 0.9825455
>   TauW<-tau_w(0.0025,w,m_a)
>   TauW
[1] 0.9988881

>   Gn_1<-600

>   TauAerosol<-Gn_1/(G0n_1*TauR*TauW)
>   TauAerosol
[1] 0.8217026

>   betaa<- -log(TauAerosol)/m_a
>   betaa
[1] 0.09741524
```