

B. Daily Radiation on a Horizontal Surface

On a given day, let G_{0n} be the extraterrestrial irradiance on a surface normal to the rays from the sun,

$$G_{0n} = G_{SC} \left(\frac{r_0}{r} \right)^2 = G_{SC} \varepsilon_j \quad (1)$$

$$r_0 = D_{SE} = 1 \text{ AU} = 1.496 \cdot 10^8 \text{ km}$$

ε_j the *eccentricity correction factor* of the earth's orbit,

$$\varepsilon = (r_0/r)^2 = 1.00011 + 0.034221 \cos \theta_j + 0.00128 \sin \theta_j + 0.000719 \cos 2 \theta_j + 0.000077 \sin 2 \theta_j$$

$\theta_j = 2\pi (j-1)/365$ is the day angle

and j is the day number of the year (Julian day).

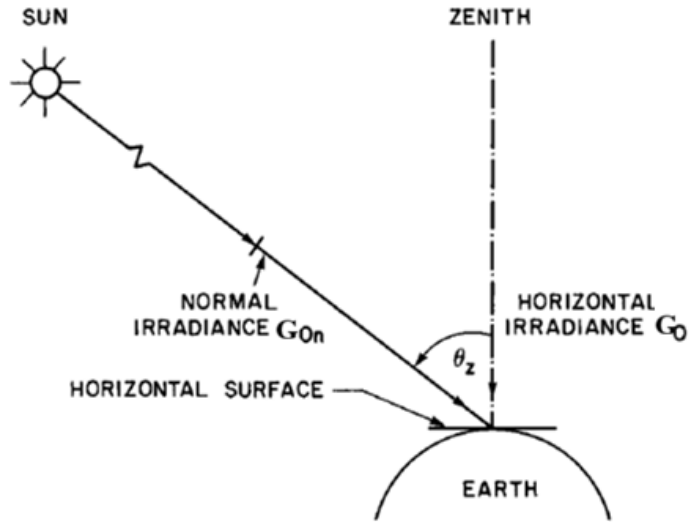


Figure 1. Relationship between the direct normal and the horizontal irradiance (Iqbal).

The **irradiance** on a horizontal surface can be written

$$G_0 = G_{0n} \cos \theta_z \quad (2)$$

$$G_0 = G_{SC} \varepsilon_j (\sin \phi \sin \delta + \cos \phi \cos \delta \cos \omega) \quad (3)$$

The units are W/m^2 .

The **irradiation** (amount of energy) dH_0 during a short period of time dt will be

$$dH_0 = G_{SC} \varepsilon_j \cos \theta_z dt \quad (4)$$

The time dt is in hours, and $\cos \theta_z$ contains the solar hour angle ω in degrees or radians. The time in hours can be converted to the hour angle as follows:

Ω = rotational speed of the earth around its polar axis

$$\Omega = \frac{2\pi \text{ rad}}{24 \text{ h}} = \frac{d\omega}{dt}$$

$$dt = \frac{12}{\pi} d\omega \quad (5)$$

$$dH_0 = \frac{12}{\pi} G_{SC} \varepsilon_j (\sin \phi \sin \delta + \cos \phi \cos \delta \cos \omega) d\omega \quad (6)$$

The irradiation during a day, from sunrise (sr) to sunset (ss), is given by

$$H_0 = \int_{sr}^{ss} G_0 dt \quad (7)$$

$$= 2 \int_0^{ss} G_0 dt$$

Assuming that ε_j remains constant for one day and after converting the time dt to the hour angle, we obtain

$$H_{0,d} = \frac{24}{\pi} G_{SC} \varepsilon_j \int_0^{\omega_s} (\sin \phi \sin \delta + \cos \phi \cos \delta \cos \omega) d\omega \quad (8)$$

$$H_{0,d} = \frac{24}{\pi} G_{SC} \varepsilon_j ((\omega_s \pi / 180)(\sin \phi \sin \delta) + (\cos \phi \cos \delta \sin \omega_s)) \quad (9)$$

Where ω_s is in degrees.

The above expression, $H_{0,d}$, can be written in terms of the sunrise hour angle, which is determined by setting the solar zenith angle equal to 90° :

$$\cos \omega_s = -\tan \phi \tan \delta \quad (10)$$

$$\cos \phi \cos \delta = -\sin \phi \sin \delta / \cos \omega_s \quad (11)$$

$$H_{0,d} = \frac{24}{\pi} G_{SC} \varepsilon_j \sin \phi \sin \delta \left((\omega_s \pi / 180) - (\tan \omega_s) \right) \quad (12)$$

Or,

$$H_{0,d} = \frac{24}{\pi} G_{SC} \varepsilon_j \cos \phi \cos \delta \left((\sin \omega_s) - \left(\frac{\omega_s \pi}{180} \right) \cos \omega_s \right) \quad (13)$$

Equations (12) and (13) are not valid at either $\phi = 0^\circ$ or $\phi = 90^\circ$.

Example: Compute the extraterrestrial daily irradiation on 15 February of a horizontal surface at Melbourne ($37^\circ 49'$ S).

15 February $j = 46$ $\phi = -37.816^\circ$

Solar constant $G_{SC} = 1367 \text{ W/m}^2 \quad \times 3.6 \quad G_{SC} = 4921 \text{ kJ/m}^2 \text{h}$

$\varepsilon_j = 1.02555$

$\delta_j = -12.952^\circ$

$\omega_s = 100.282^\circ$

From Eq (9),

$$H_0 = \frac{24}{\pi} G_{SC} \varepsilon_j \left((\omega_s \pi / 180) (\sin(-37.8) \sin(-12.9)) + (\cos(-37.8) \cos(-12.9) \sin \omega_s) \right)$$

```
# Exe 2
> J<-46    ##15 february

> correction_spencer <- function(J) {
+   1.00011 +
+   0.034221 * cos(2*pi*(J-1)/365) +
+   0.00128 * sin(2*pi*(J-1)/365) +
+   0.000719 * cos(4*pi*(J-1)/365) +
+   0.000077 * sin(4*pi*(J-1)/365)
+ }
```

```

> eccentricity<-correction_spencer(J)
> eccentricity
[1] 1.025555

> AngleDeclin <- function(x)
+   return((0.006918 -
+           0.399912*cos(2*pi*(x-1)/365) +
+           0.070257*sin(2*pi*(x-1)/365) -
+           0.006758*cos(4*pi*(x-1)/365) +
+           0.000907*sin(4*pi*(x-1)/365) +
+           0.00148*sin(6*pi*(x-1)/365) -
+           0.002697*cos(6*pi*(x-1)/365))*180/pi)
> delta<-AngleDeclin(J)*pi/180
> delta
[1] -0.226061
> delta*180/pi
[1] -12.95234

> G_SC<-1367 #in W/m2
> I_SC<-G_SC*3.6
> I_SC
[1] 4921.2

> Lat_deg <- -37 ##Melbourne
> Lat_min <- -49
> Lat <- Lat_deg + Lat_min/60
> Lat
[1] -37.81667

+   Unghi_Orar_rasarit_S_oriz<-function(x,y)
+   return<-acos(-tan(x)*tan(y))

> Omega_s<-Unghi_Orar_rasarit_S_oriz(Lat*pi/180,delta)
> Omega_s
[1] 1.750266
> Omega_s*180/pi
[1] 100.2828

> H0<-(I_SC*eccentricity*24/pi)*(Omega_s*sin(delta)*sin(Lat*pi/180)+cos(delta)*cos(Lat*pi/180)*sin(Omega_s))
> H0 ##in kJ/m2 day
[1] 38480.7
> H0/1000 ##in MJ/m2 day
[1] 38.4807
> H0/3600 ## in kWh/m2 intr-o zi !!!!
[1] 10.68908

```

10.68868 kWh/m2 in a day!!!

A special case of the extraterrestrial daily radiation on a horizontal surface is at the equator, $\phi = 0$, $\omega_s = \pi/2$.

$$H_0 = \frac{24}{\pi} G_{SC} \varepsilon_j \left((\omega_s \pi / 180) (\sin 0 \sin \delta) + (\cos 0 \cos \delta \sin \pi / 2) \right)$$

$$H_0 = \frac{24}{\pi} G_{SC} \varepsilon_j \cos \delta \quad (14)$$

Chapter 5 A CLOUDLESS-SKY ATMOSPHERE AND ITS OPTICS

Solar radiation emanating from the Sun is attenuated, before reaching the ground, by the earth's atmosphere, which can be classified into two broad types: (1) atmosphere without clouds and (2) atmosphere with clouds. In this chapter, constituents of a cloudless atmosphere are described.

5.1 The Earth's Atmosphere

Since the early 20th century, standard atmosphere models have been developed to describe the vertical structure of the Earth's atmosphere. The U.S. Standard Atmosphere adopted in 1976 provides a reference model for atmospheric properties. Figure 1 (adapted from Salby) illustrates the vertical profiles of atmospheric pressure, density, and temperature up to an altitude of 100 km.

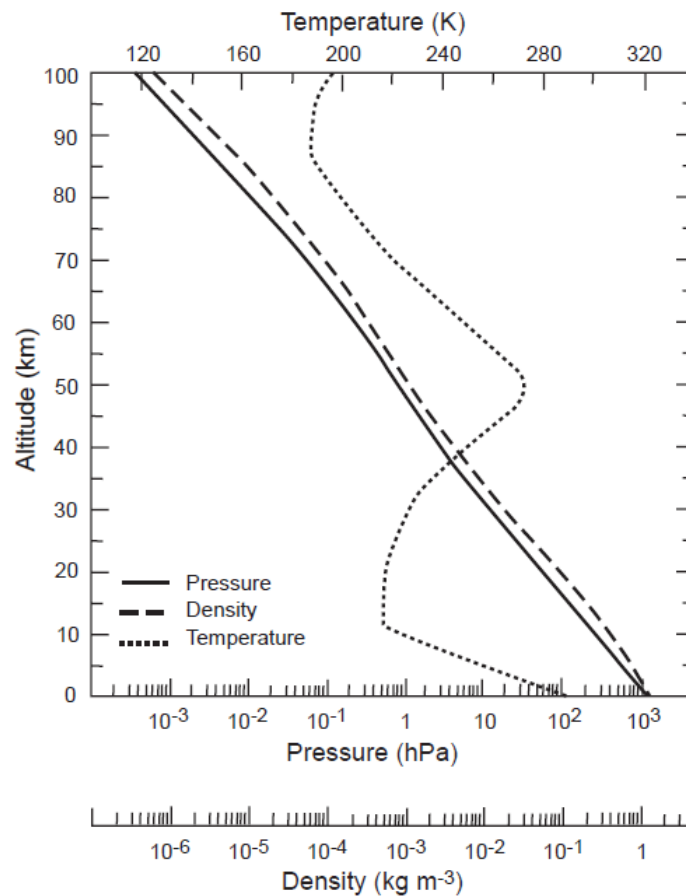


Figure 1. Pressure (solid), density (dashed), and temperature (dotted), as function of altitude.

The earth's atmosphere is divided into several concentric spheres: troposphere, stratosphere, mesosphere, thermosphere, etc., and important temperature variations exist in each of these. However, density and pressure decrease continuously.

The Earth's atmosphere is a mixture of gases composed primarily of molecular nitrogen (N_2), which accounts for approximately 78% of its volume, and molecular oxygen (O_2), which accounts for about 21% (Table 1.1). The remaining fraction consists mainly of argon and other trace constituents, including water vapor (H_2O), carbon dioxide (CO_2), and ozone (O_3).

Although present in much smaller concentrations, certain trace constituents, particularly water vapor, carbon dioxide, and ozone, play an important role in the Earth's energy balance because they absorb radiation at specific wavelengths.

Table 1.1. Atmospheric Composition. Constituents are listed with volume mixing ratios representative of the Troposphere or Stratosphere, how the latter are distributed vertically, and controlling processes

Constituent	Tropospheric Mixing Ratio	Vertical Distribution (Mixing Ratio)	Controlling Processes
N_2	.7808	Homogeneous	Vertical Mixing
O_2	.2095	Homogeneous	Vertical Mixing
* H_2O	≤ 0.030	Decreases sharply in Troposphere Increases in Stratosphere Highly Variable	Evaporation, Condensation, Transport Production by CH_4 Oxidation
Ar	.0093	Homogeneous	Vertical Mixing
* CO_2	380 ppmv	Homogeneous	Vertical Mixing Production by Surface and Anthropogenic Processes
* O_3	10 ppmv [§]	Increases sharply in Stratosphere Highly Variable	Photochemical Production in Stratosphere; secondarily through pollution in troposphere Destruction at Surface Transport
* CH_4	1.8 ppmv	Homogeneous in Troposphere Decreases in Middle Atmosphere	Production by Surface Processes Oxidation Produces H_2O
* N_2O	320 ppbv	Homogeneous in Troposphere Decreases in Middle Atmosphere	Production by Surface and Anthropogenic Processes Dissociation in Middle Atmosphere Produces NO Transport
* CO	70 ppbv	Decreases in Troposphere Increases in Stratosphere	Production Anthropogenically and by Oxidation of CH_4 Transport
NO	0.1 ppbv [§]	Increases Vertically	Production by Dissociation of N_2O Catalytic Destruction of O_3
*CFC-11	0.2 ppbv	Homogeneous in Troposphere	Industrial Production
*CFC-12	0.5 ppbv	Decreases in Stratosphere	Mixing in Troposphere
*HFC-134A	30 ppt		Photo-dissociation in Stratosphere

* Radiatively active

[§] Stratospheric value

ppmv înseamnă parts per million by volume — părți per milion în volum.

ppbv înseamnă parts per billion by volume — părți per miliard în volum.

De exemplu, dacă concentrația de CO₂ este de 380 ppmv, înseamnă că există aproximativ 380 de volume de CO₂ la un milion de volume de aer.

There are two groups of gases, one with nearly permanent concentrations (nitrogen, oxygen, argon, carbon dioxide) and another with variable concentrations (water vapor, ozone, sulfur dioxide, nitrogen dioxide). The atmosphere also contains various kinds of aerosols, clouds and precipitations, which are highly variable in space and time.

Water vapor is an important radiatively active atmospheric constituent. It absorbs incoming solar radiation, particularly in the near-infrared region, and absorbs and emits terrestrial infrared radiation, thereby influencing both atmospheric transmittance and the Earth's energy balance.

Water vapor (H₂O) is a highly variable atmospheric constituent, both spatially and temporally. Most atmospheric water vapor is concentrated in the troposphere, where its abundance depends on temperature, humidity, altitude, and meteorological conditions. Its concentration generally decreases with altitude. Only a small amount of water vapor enters the stratosphere

All molecules of air deplete solar energy by *scattering*, which takes place at all wavelengths, and which therefore is called a continuum process. Air molecules, though, *absorb* solar radiation only at selective wavelengths.

The most important of the dry air absorbers are *ozone, carbon dioxide, oxygen, oxides of nitrogen*, nitrogen, and hydrocarbon combinations.

In the upper atmosphere, *ozone* is created mainly by ultraviolet solar radiation. On the ground, it is formed through decomposition of nitrogen oxide that enters the atmosphere from factory smoke and forest fires, for example. The amount of total ozone *l* in a vertical column of air is given in units of atmosphere centimeters (atm cm). This is the height of gaseous ozone if all the ozone in a vertical column of unit area were brought to normal temperature and surface pressure (NTP). In engineering literature, this thickness is usually expressed in cm (NTP). Around the equator, total ozone averages about 0.24 cm (NTP), and the amount increases with latitude: in the polar regions, total ozone may be as much as 0.46 cm (NTP). At high latitudes, there are distinct seasonal variations: in each hemisphere, it is maximum in the spring and minimum in the fall (in the sense of seasons in each hemisphere). Table 2 gives the seasonal variation of ozone content for the northern and the southern hemispheres.

Seasonal Variation of Atmospheric Ozone [in cm (NTP)]^a

Latitude	Month											
	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
90° N	0.33	0.39	0.46	0.42	0.39	0.34	0.32	0.30	0.27	0.26	0.28	0.30
80° N	0.34	0.40	0.46	0.43	0.40	0.36	0.33	0.30	0.28	0.27	0.29	0.31
70° N	0.34	0.40	0.45	0.42	0.40	0.36	0.34	0.31	0.29	0.28	0.29	0.31
60° N	0.33	0.39	0.42	0.40	0.39	0.36	0.34	0.32	0.30	0.38	0.30	0.31
50° N	0.32	0.36	0.38	0.38	0.37	0.35	0.33	0.31	0.30	0.28	0.29	0.30
40° N	0.30	0.32	0.33	0.34	0.34	0.33	0.31	0.30	0.28	0.27	0.28	0.29
30° N	0.27	0.28	0.29	0.30	0.30	0.30	0.29	0.28	0.27	0.26	0.26	0.27
20° N	0.24	0.26	0.26	0.27	0.28	0.27	0.26	0.26	0.26	0.25	0.25	0.25
10° N	0.23	0.24	0.24	0.25	0.26	0.25	0.25	0.24	0.24	0.23	0.23	0.23
0°	0.22	0.22	0.23	0.23	0.24	0.24	0.24	0.23	0.23	0.22	0.22	0.22
10° S	0.23	0.24	0.24	0.24	0.24	0.24	0.24	0.24	0.24	0.24	0.24	0.23
20° S	0.24	0.25	0.24	0.25	0.25	0.25	0.25	0.26	0.26	0.26	0.26	0.25
30° S	0.27	0.28	0.26	0.27	0.28	0.28	0.29	0.31	0.32	0.32	0.29	0.29
40° S	0.30	0.29	0.28	0.29	0.31	0.33	0.35	0.37	0.38	0.37	0.34	0.32
50° S	0.31	0.30	0.29	0.30	0.32	0.36	0.39	0.40	0.40	0.39	0.37	0.35
60° S	0.32	0.31	0.30	0.30	0.33	0.38	0.41	0.42	0.42	0.40	0.39	0.35
70° S	0.32	0.31	0.31	0.29	0.34	0.39	0.43	0.45	0.43	0.40	0.38	0.34
80° S	0.31	0.31	0.31	0.28	0.35	0.40	0.44	0.46	0.42	0.38	0.36	0.32
90° S	0.31	0.30	0.30	0.27	0.34	0.38	0.43	0.45	0.41	0.37	0.34	0.31

Table 2. The seasonal variation of ozone content for the northern and the southern hemispheres.

The vertical distribution of ozone varies with latitude and season. It is mainly concentrated between a 10- and 35-km altitude. The profile given in Fig. 2 has been considered to represent average conditions at midlatitudes. This profile results in 0.35 cm (NTP) ozone.

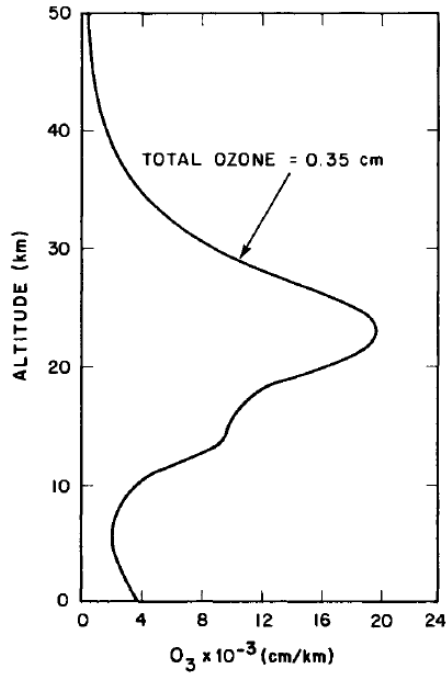


Figure 2. Variation of ozone concentration with altitude.

Dobson Unit (DU)

The **Dobson unit (DU)** is used to express the total amount of ozone in a vertical column of the atmosphere, from the Earth's surface to the top of the atmosphere. It is defined as the thickness of the layer of pure ozone that would result if all the ozone in this atmospheric column were brought to the Earth's surface at a temperature of 273 K and a pressure of $(1.013 \times 10^5 \text{ Pa})$. The thickness is expressed in hundredths of a millimeter.

One Dobson unit corresponds to approximately (2.69×10^{16}) ozone molecules per square centimeter. A typical global value of total column ozone is about 300 DU, equivalent to a layer of pure ozone approximately 3 mm thick under the specified reference conditions.

The unit is named after **G. M. B. Dobson**, who developed the Dobson spectrophotometer in the 1920s. This instrument measures solar ultraviolet radiation at selected wavelengths, some strongly absorbed by ozone and others less affected by its absorption, allowing the total ozone column to be determined.

Water vapor

Precipitable water is the total amount of water vapor contained in a vertical atmospheric column extending from the Earth's surface to the top of the atmosphere. It is expressed as the height of the liquid water layer that would form if all the water vapor in the column condensed. Precipitable water is commonly measured in centimeters (cm). For example, a precipitable water value of 2 cm means that the water vapor in the atmospheric column would produce a 2-cm layer of liquid water upon complete condensation.

Precipitable water is an important parameter in solar radiation models because water vapor absorbs solar radiation in several spectral bands, particularly in the near-infrared region.

$$1 \text{ cm of precipitable water} = 1 \frac{\text{g}}{\text{cm}^2} = 10 \text{ kg/m}^2$$

About one-half the precipitable water is concentrated in the first 2 km above sea level; beyond 12 km, precipitable water is practically nonexistent.

The amount of precipitable water can be highly variable. It varies with season. An extremely dry atmosphere may contain as little as 1 mm of precipitable water and a humid atmosphere may contain more than 40 mm.

The overall mean monthly values are given in Table:

Table 5.4.3
*Mean Monthly Values of Precipitable Water (cm) for the United States,
Based on 1946–1956 Data^a*

Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
0.98	0.94	1.05	1.37	1.89	2.49	2.99	2.90	2.28	1.72	1.20	1.11

^a From Reitan [7].

It can be seen from this table that the precipitable water during the summer months in the United States is about three times as high as during the winter months.

Precipitable water can be estimated from surface meteorological measurements, such as water vapor pressure or dew-point temperature, using empirical relationships.

For example,

Cole 1976:

$$w' = 0.125 \exp(0.296 p_v^{1/2} - 0.803) \quad (1)$$

where w' is in centimeters and p_v (the vapor pressure at ground level) in millibars.

A correlation valid for United States (**Smith** 1966):

$$w' = \exp(0.07074 t_d + y) \quad (2)$$

Here, w' is in centimeters, t_d is the station dew-point temperature in degrees Celsius, and $y = -0.0229$ from April to June and 0.02023 for the remaining months.

Won 1977:

$$w' = 0.1 \exp(2.2572 + 0.05454 t_d) \quad (3)$$

The precipitable water as obtained from Eqs. (1)-(3) applies to prevailing station pressure and temperature. However, the attenuation equations (discussed in the following chapters) often require this quantity to be reduced to a datum of 1013.25 mbars pressure and 273 K temperature.

Paltridge and Platt 1976 suggest the following formula for reduction of w' to the datum conditions:

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

where w is the reduced precipitable water in centimeters (normalizata la presiunea si temperatura standard), p is station pressure in millibars, and T is the surface temperature in degrees Kelvin.

Leckner (1978) has presented the following formula, which expresses precipitable water in terms of relative humidity:

$$w = 0.493 \phi_r p_s / T \quad (5)$$

where ϕ_r is relative humidity, T is ambient temperature in degrees Kelvin, and p_s is the partial pressure of water vapor in saturated air and is given by the following semiempirical equation:

$$p_s = \exp(26.23 - 5416/T) \quad (6)$$

In this chapter, although we are interested in conditions under cloudless skies, data on precipitable water are usually given as the mean of all days.

Aerosols

2.7 PARTICULATE MATTER (AEROSOLS)

Particles in the atmosphere arise from natural sources, such as windborne dust, seaspray, and volcanoes, and from anthropogenic activities, such as combustion of fuels. Whereas an *aerosol* is technically defined

TABLE 2.19 Terminology Relating to Atmospheric Particles

Aerosols, aerocolloids, aerodisperse systems	Tiny particles dispersed in gases
Dusts	Suspensions of solid particles produced by mechanical disintegration of material such as crushing, grinding, and blasting; $D_p > 1 \mu\text{m}$
Fog	A term loosely applied to visible aerosols in which the dispersed phase is liquid; usually, a dispersion of water or ice, close to the ground
Fume	The solid particles generated by condensation from the vapor state, generally after volatilization from melted substances, and often accompanied by a chemical reaction such as oxidation; often the material involved is noxious; $D_p < 1 \mu\text{m}$
Hazes	An aerosol that impedes vision and may consist of a combination of water droplets, pollutants, and dust; $D_p < 1 \mu\text{m}$
Mists	Liquid, usually water in the form of particles suspended in the atmosphere at or near the surface of Earth; small water droplets floating or falling, approaching the form of rain, and sometimes distinguished from fog as being more transparent or as having particles perceptibly moving downward; $D_p > 1 \mu\text{m}$
Particle	An aerosol particle may consist of a single continuous unit of solid or liquid containing many molecules held together by intermolecular forces and primarily larger than molecular dimensions ($> 0.001 \mu\text{m}$); a particle may also consist of two or more such unit structures held together by interparticle adhesive forces such that it behaves as a single unit in suspension or on deposit
Smog	A term derived from smoke and fog, applied to extensive contamination by aerosols; now sometimes used loosely for any contamination of the air
Smoke	Small gasborne particles resulting from incomplete combustion, consisting predominantly of carbon and other combustible materials, and present in sufficient quantity to be observable independently of the presence of other particles; $D_p \geq 0.01 \mu\text{m}$
Soot	Agglomerations of particles of carbon impregnated with "tar," formed in the incomplete combustion of carbonaceous material

Suspensions of liquid and solid particles are relevant to radiative as well as chemical processes in the atmosphere. Aerosol particles range in size from 10^{-3} to $10^2 \mu\text{m}$. They promote cloud formation. Aerosol particles serve as condensation nuclei for water droplets and ice crystals, which do not form readily in their absence. These small particulates are produced naturally (e.g., as dust, sea salt, and volcanic debris). They are also produced anthropogenically through combustion and industrial processes. An important source of aerosol is *gas-to-particle conversion*. It occurs through chemical reactions involving, among other precursors, the gaseous emission of sulfur dioxide (SO_2). Sulfur dioxide is produced by industry and naturally by volcanos.

Aerosol concentrations are high in urban areas and near industrial complexes (Fig. 3). The concentration of anthropogenic aerosols is much higher than that of marine aerosols, while aerosol concentrations during desert dust storms can be even higher.

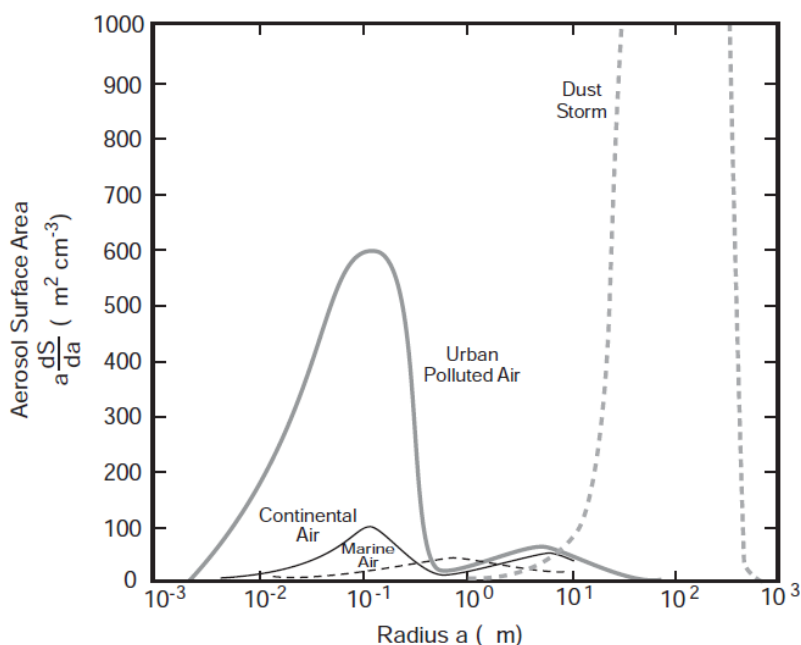


Figure 3. Size spectrum of aerosol surface area as a function of particle radius (equal area under the curves represent equal number densities of aerosol particles) [2].

In addition to their role in cloud formation, aerosol particles interact with radiation: They scatter solar radiation at ultraviolet and visible wavelengths. They absorb IR radiation that is emitted by the Earth's surface and atmosphere. These radiative properties involve aerosols in the global energy balance, where they have a role analogous to greenhouse gases.

Sporadic increases of aerosol produced by major volcanic eruptions have been linked to changes in thermal, optical, and chemical properties of the atmosphere. Figure 4a plots the change of global-mean temperature in the lower troposphere following the eruption of Mt. Pinatubo in 1991.

Pinatubo introduced volcanic gas and debris well into the stratosphere. Its enhancement of aerosol there remained long enough to modify transfers of radiative energy, in particular, to diminish solar radiation that reaches the Earth's surface. Global-mean temperature decreased following the eruption, attaining a maximum depression of ~ 0.5 K one and a half years later. The depression remained for another 2 years, while global mean temperature gradually returned to its unperturbed value. Accompanying the depression of temperature was a depression of the column abundance of water vapor (Fig. 4b). Precipitable water vapor decreased in much the same fashion, attaining a maximum depression at about the same time of ~ 0.5 mm.

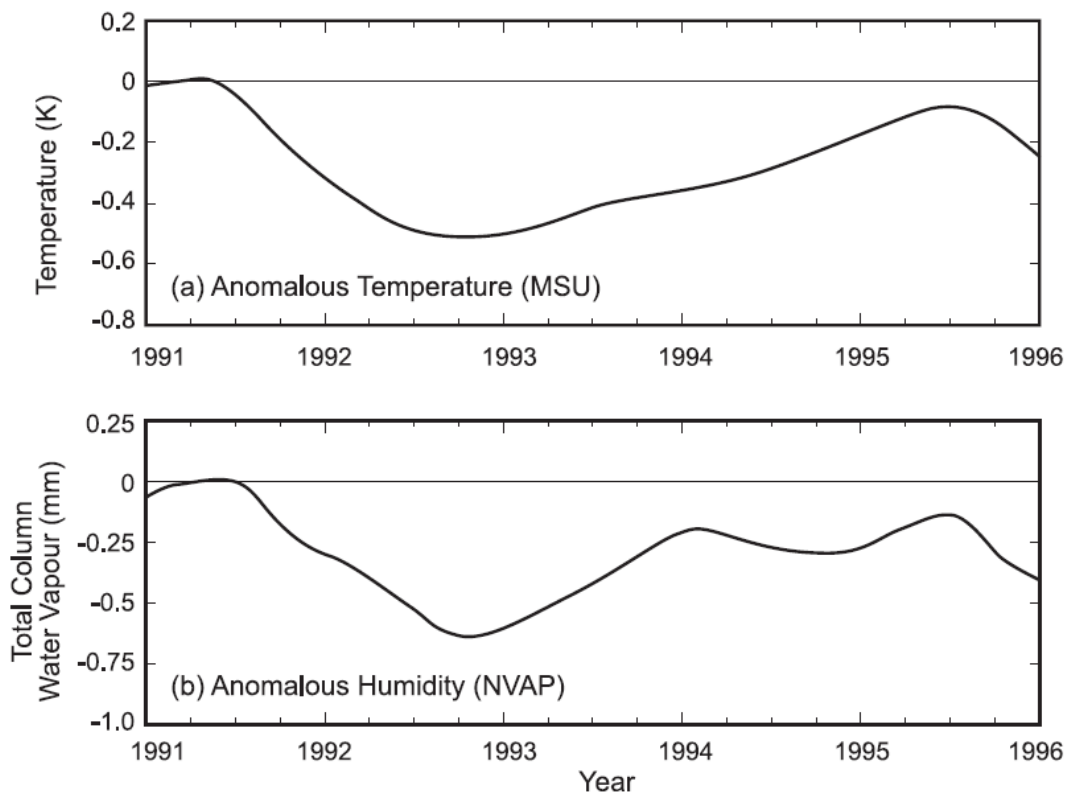


Figure 1.27 (a) Anomalous temperature observed by the Microwave Sounding Unit (MSU) and (b) anomalous column abundance of water vapor from the NASA Water Vapor Project (NVAP) following the eruption of Pinatubo. *Source:* Soden et al. (2002).

Figure 4. (a) Anomalous temperature observed and (b) anomalous column abundance of water vapor following the eruption of Pinatubo.

References

- [1] Iqbal M (1983) An Introduction to Solar Radiation, ACADEMIC PRESS.
- [2] Murry L. Salby, Physics of the Atmosphere and Climate, Cambridge University Press, 2012.
- [3] John H. Seinfeld, Spyros N. Pandis, Atmospheric Chemistry Physics From Air Pollution to Climate Change, Third Edition, John Wiley & Sons, Inc 2016.