

Shuttle glow, due to atomic oxygen in the upper atmosphere, came as a rude surprise to astronomers and others. Seen here on STS-39 (right) to be roughly as bright as the aurora.



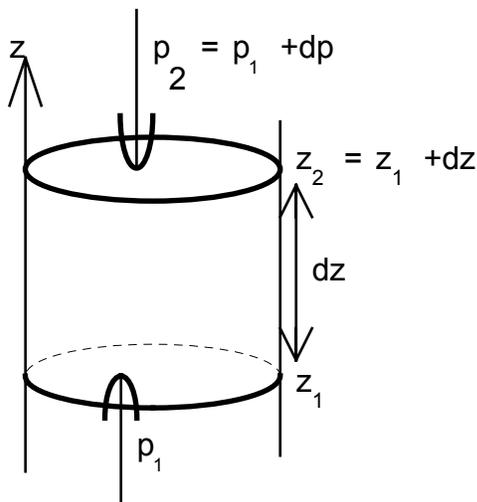
Chapter 6 The Atmosphere

A Introduction

The atmosphere of the earth is important for our purposes as a portion of the satellite (or rocket) environment, and as a source of material for the regions of space outside the sensible atmosphere. The region of interest for our purposes starts at an altitude of about 30 km and extends outward to several thousand kilometers. The neutral atmosphere overlaps the ionosphere, plasmasphere and the radiation belts. That portion of the upper atmosphere which is ionized is termed the ionosphere, extending from about 50 km altitude to ~1000 km altitude. The upper atmosphere is a dynamic system under the influence of gravity, radiation and a complex system of photochemical reactions.

B Law of Atmospheres: Pressure and Density

Our analytic approach to the atmosphere and ionosphere begins with an application of the scale height concept, originally presented in Chapter 1.1



Consider a gas column of unit cross-section as shown, and a coordinate system with z positive upward. If we now consider a slab of gas of thickness dz we can write down the condition that this volume of gas be in equilibrium:

$$\sum \text{Forces}_z = 0 \quad \Rightarrow$$

$$p_1 = p_2 + \rho g dz \quad (\text{Eqn. 6.1})$$

where

ρ = the mass density of the gas (kg/m^3),
 g = acceleration due to gravity,
 p_1 and p_2 are the pressures,

Figure 6.1

Noting also that $p_1 = p_2 + dp$

. We can combine the above equations into the form

$$p_2 - p_1 = dp = -\rho g dz \quad (\text{Eqn. 6.2})$$

which gives the relation between height and pressure change in differential form. Before we can integrate this equation we must express ρ as a function of p . The ideal gas law can be invoked now. The most useful form for our purposes is:

$$p = nkT \quad (\text{Eqn. 6.3})$$

where

$$\begin{aligned} p &= \text{pressure (N/m}^2\text{)} \\ n &= \text{number density (molecules/m}^3\text{)} \\ k &= \text{the Boltzmann constant} = 1.38 \times 10^{-23} \text{ (J / K)} \\ T &= \text{Absolute Temperature (K)} \end{aligned}$$

The mass density, ρ , is not the commonly used variable in atmospheric work, however. Instead, the mean molecular weight, μ , is used, along with the proton mass, m_p .

$$\rho = n m_p \mu \quad (\text{Eqn. 6.4})$$

$$\begin{aligned} dp &= -n m_p \mu g dz = -\frac{p}{kT} m_p \mu g dz \\ \frac{dp}{p} &= -\frac{\mu m_p g}{kT} dz \end{aligned} \quad (\text{Eqn. 6.5})$$

The quantity $\frac{kT}{\mu m_p g}$ appears to have the dimensions of a height, and is indeed is defined as the pressure scale height:

$$H_p = \frac{kT}{\mu m_p g} \quad (\text{Eqn. 6.6})$$

We can now integrate both sides from some reference level z_r (pressure p_r) to the final values z , and p .

$$\begin{aligned} \int_{p_r}^p \frac{dp}{p} &= -\int_{z_r}^z \frac{dz}{H_p} \Rightarrow \ln \frac{p}{p_r} = -\int_{z_r}^z \frac{dz}{H_p} \\ p(z) &= p_r e^{-\int_{z_r}^z \frac{dz}{H_p}} \end{aligned} \quad (\text{Eqn. 6.7})$$

This is the famous law of the atmospheres. If the scale height is constant with altitude, the integral is trivial:

$$\ln \frac{p}{p_r} = -\frac{z-z_r}{H_p} \Rightarrow p = p_r e^{-\frac{z-z_r}{H_p}} \quad (\text{Eqn. 6.8})$$

For a constant scale height atmosphere we shall now show that H_p is the thickness of this same atmosphere when compressed to a uniform pressure p_r . For an isothermal atmosphere the particle density is proportional to pressure because $p = nkT$, and we can write

$$dn(z) = n_r e^{-z/H_p} dz \quad (\text{Eqn. 6.9})$$

where n is the number density (#/volume) of molecules, and the total number in a 1 m² column is

$$\begin{aligned} n_T &= \int_0^\infty n_r e^{-z/H_p} dz = n_r H_p \int_0^\infty e^{-u} du \\ n_T &= -n_r H_p (0-1) = n_r H_p \end{aligned} \quad (\text{Eqn. 6.10})$$

which says that an atmosphere of constant particle density n_r and height H_p will have the same number of particles as the actual column has.

In the actual atmosphere the scale height is approximately constant up to about 120 km altitude at a value of

$$H_p \cong 8.4 \text{ km}$$

As we shall see later, the temperature begins to rise rapidly above 120 km. At the same time the mean molecular mass decreases as shown in Figure 6.2 below. The gravity also decreases slowly (Figure 6.3)

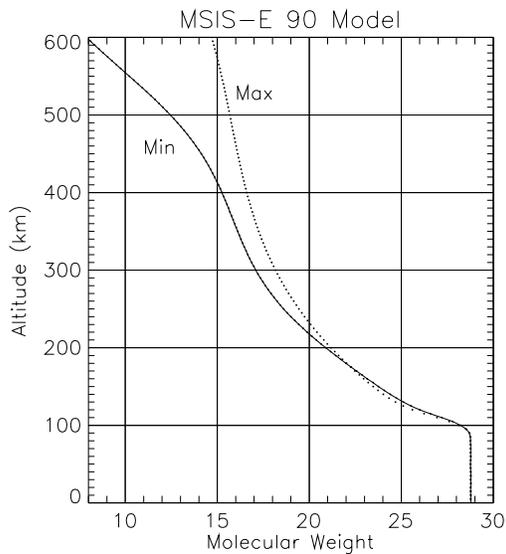


Figure 6.2 Mean molecular weight μ as a function of geometric altitude. MSIS-E Model 1990. Solar minimum (Min) taken 9/22/95; Solar maximum (Max) taken 9/22/89; both at 0° latitude, 20° longitude, local noon.

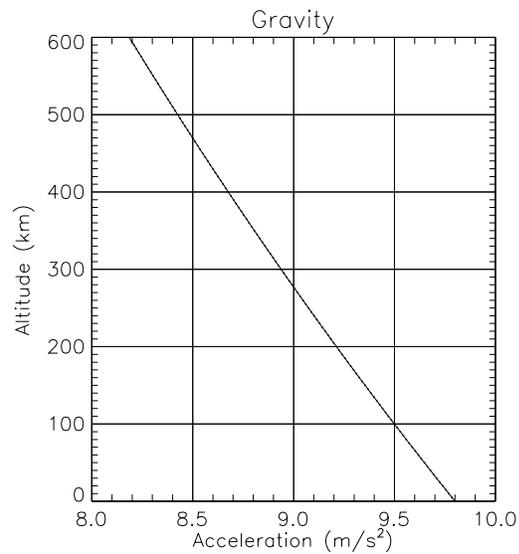


Figure 6.3 Acceleration due to gravity, g , as a function of geometric altitude. (Just a little physics.)

<http://nssdc.gsfc.nasa.gov/space/model/atmos/msis.html>

When these variations are combined with the previously discussed variation of T with altitude we obtain the variation of scale height H_p with altitude. Figure 6.4 shows how the pressure scale height varies with altitude. Since the primary dependence is on temperature, this figure closely resembles the atmospheric temperature profile. Once we have the scale height as a function of altitude we can get the particle density (and the pressure) by integrating the Law of Atmospheres from sea level to any height. Figures 6.5a and 6.5b show how the pressure and density behave for the (model) standard atmosphere. We see from these graphs that both the pressure and the particle density drop off at a slower rate above 100 km which corresponds to the larger scale height at these altitudes, as shown in Figure 6.4

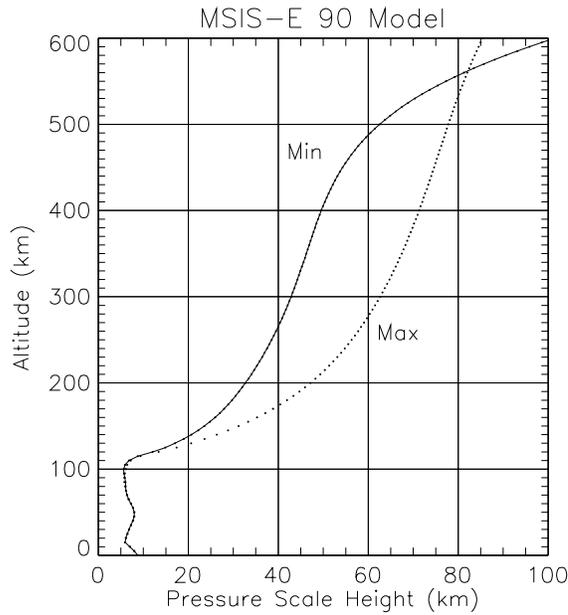


Figure 6.4 Pressure scale height, H_p as a function of geometric altitude. From the NASA MSIS-E-90 Model, implemented at the NASA/GSFC National Space Science Data Center. Calculations are shown for solar minimum (Min) and solar maximum (Max).

<http://nssdc.gsfc.nasa.gov/space/model/atmos/msis.html>

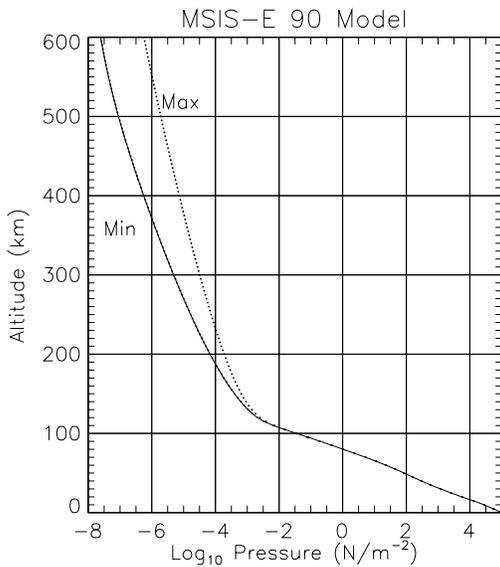


Figure 6.5a Pressure as a function of altitude in the standard atmosphere. Note the log scale on the horizontal axis.

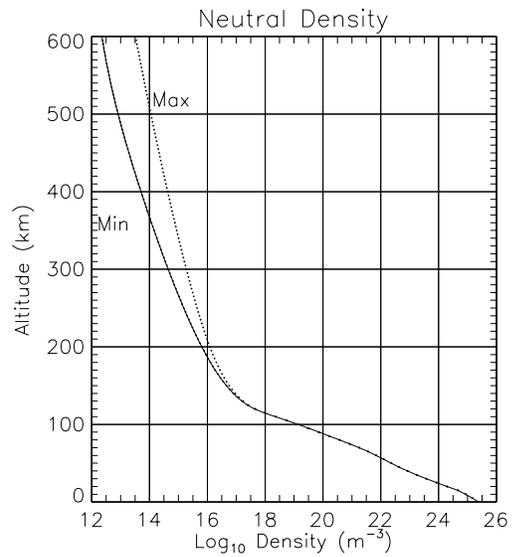


Figure 6.5b Number density as a function of geometric altitude, solar minimum (Min) and solar maximum (Max).

C Temperature of the Atmosphere

The atmosphere is divided up into distinct regions, primarily on the basis of how the temperature varies in different altitude ranges. The different regions are, in order of ascending altitude: the troposphere, the stratosphere, the mesosphere, and the thermosphere. Each of the four characteristic layers, is bounded by a 'pause', a transition region where the temperature profile has an inflection point (reversal in slope). (see Figure 6.6). The temperature gradient is called a 'lapse rate'.

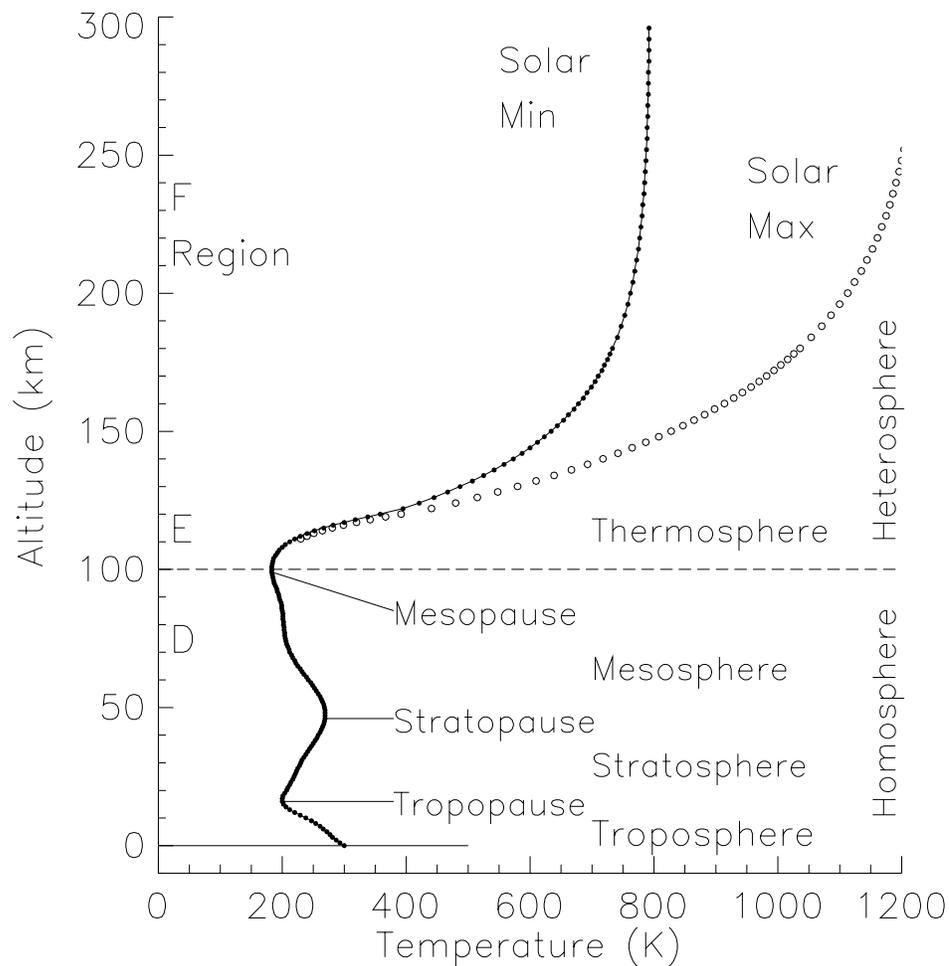


Figure 6.6 Variation in temperature with altitude. Figure created with MSIS-E 1990 model. Ionospheric regions (D, E, F, annotated along left hand axis).

The lowest region in the atmosphere is called the troposphere where clouds and weather originate and where turbulence is the dominant physical process. The temperature decreases at a fairly uniform rate of $-6.5^{\circ}\text{C}/\text{km}$ for about 15 km and reaches a minimum of approximately -60°C (for mid latitudes) at the tropopause. If we calculate a convective temperature lapse rate for a dry atmosphere under adiabatic conditions we get a value of about $-9^{\circ}\text{C}/\text{km}$. The difference is mostly due to the presence of water vapor which modifies the purely convective cooling in two ways (a) The condensation of water vapor to form clouds gives up energy which heats the air and (b) water molecules are excellent absorbers (and radiators) of IR radiation emitted by the surface of the planet, a process which leads to further heating of the air (the so-called Greenhouse Effect). The net result of these processes is to decrease the temperature lapse rate to the observed value of $-6.5^{\circ}\text{C}/\text{km}$. The tropopause is where the temperature gradient vanishes. It occurs at an altitude of 6 km to 18 km (higher and colder over the equator). It is the domain of high winds and the highest observed cirrus clouds.

The next region is the stratosphere which is about 25 km thick at mid latitude (thicker over the poles and thinner over the equator). The temperature in the stratosphere increases with increasing altitude as a result of the absorption of solar ultraviolet radiation near the base of the stratosphere. This effect is particularly pronounced in the ozone layer where a concentration of only a few parts per million of O_3 leads to very pronounced absorption of wavelengths in the 200-300 nm region and subsequent heating of the atmospheric gases. The temperature increase terminates at the stratopause which occurs at an altitude of about 50 km and is followed by another region of declining temperature, the so-called mesosphere. The temperature in this region is mostly determined by convection, the warmer stratosphere now acting in much the same way as the planetary surface acts with respect to the troposphere. This cooling trend terminates at the mesopause where the coldest temperatures of the atmosphere (about 180°K) are observed.

Above about 80 km we enter the thermosphere where strong positive temperature gradients exist, caused by a combination of facts: (a) the density of the gases is becoming quite low (b) there are few if any triatomic molecules present (c) solar radiation of short wavelength ($\lambda < 175\text{ nm}$) heats the atmosphere from above (d) exothermic reactions between ions and electrons occur.

Since the heating now occurs from above rather than below there is no convection and the temperature increases although we are now entering a region where collisions are too infrequent to establish thermodynamic equilibrium. At extreme altitudes (300 - 400 km) the temperature rise stops as we enter the exosphere where the temperature becomes essentially independent of altitude.

The high temperatures found in the upper reaches of our atmosphere do not imply that there is a large heat reservoir there, since particle densities are very low at these altitudes. Instead these high temperature (approaching $1000 - 1500^{\circ}\text{K}$) merely reflect the inefficiency of the heat removal processes. Only about 10^{-6} of the solar energy supplied to the earth is absorbed in the thermosphere, the rest being absorbed in the lower atmosphere and the ground.

There also exists considerable variability of this terminal temperature both as function of solar activity (900° to 1700°) and on a daily basis. These variations result from the low heat capacity of the upper atmosphere so that fluctuations in the UV radiation are rapidly reflected in temperature changes. Geomagnetic storms can also result in large temperature fluctuations of the upper atmosphere probably due to increased particle precipitation.

D Composition of the Atmosphere

If we divide the atmosphere according to composition there are three characteristic spheres:

(a) Homosphere: the region of essentially uniform composition in the sense that the mean molecular weight remains constant as we proceed from the surface upward. Even as we traverse the region of maximum ozone concentration we find that the mean molecular weight remains essentially constant.

(b) Heterosphere: At about 100 km we enter the region of significantly varying composition. The "molecular weight" diminishes from 29 at about 90 km to about 16 at 500 km. Above the level of oxygen dissociation, nitrogen begins to dissociate and diffusive separation sets in.

(c) Exosphere is the region in which the escape of molecules from the atmosphere becomes significant. The lower limit of the exosphere is not clearly defined since the escape depends on the mass of the atoms involved, but values of 500 km to 1000 km are typical.

Table 6.1. Normal Composition of Clean, Dry Atmospheric Air near Sea Level^a

Constituent gas and formula	Content, percent by volume	Molecular weight ^b
Nitrogen (N ₂)	78.084	28.0134
Oxygen (O ₂)	20.9476	31.9988
Argon (Ar)	0.934	39.948
Carbon Dioxide (CO ₂)	0.0314 ξ	44.00995
Neon (Ne)	0.001818	20.183
Helium (He)	0.000524	4.0026
Krypton (Kr)	0.000114	83.80
Xenon (Xe)	0.0000087	131.30
Hydrogen (H ₂)	0.00005	2.01594
Methane (CH ₄)	0.0002 ξ	16.04303
Nitrous oxide (N ₂ O)	0.00005	44.0128
Ozone (O ₃)	Summer: 0 to 0.000007 ξ Winter: 0 to 0.000002 ξ	47.9982 7
Sulfur Dioxide (SO ₂)	0 to 0.0001 ξ	64.0628
Nitrogen Dioxide (NO ₂)	0 to 0.000002 ξ	46.0055
Ammonia (NH ₃)	0 to trace	17.0306
Carbon Monoxide (CO)	0 to trace	28.01055
Iodine (I ₂)	0 to trace	253.8088

^aAdapted from Dubin, Sissenwine, and Wexler [1].

^bOn basis of carbon-12 isotope scale for which C¹² = 12.

ξ The content of these gases may under go significant variations from time to time or from place to place *relative* to the normal indicated for those gases.

The composition of the neutral atmosphere changes with increasing altitude as a result of chemical (including photochemical) reactions. The region of the atmosphere in which chemical reactions are prevalent is called the chemosphere and extends from roughly 20 km to 110 km, although some reactions occur both above and below this region. Table 6.1 shows the normal composition of the atmosphere near sea level.

Apart from the variable water vapor content the composition of the atmosphere is approximately constant up to an altitude of 30 km. At this altitude the concentration of ozone (O_3) begins to increase and because of its very high absorption coefficient for ultraviolet radiation it plays a dominant role in the physical processes of the atmosphere.

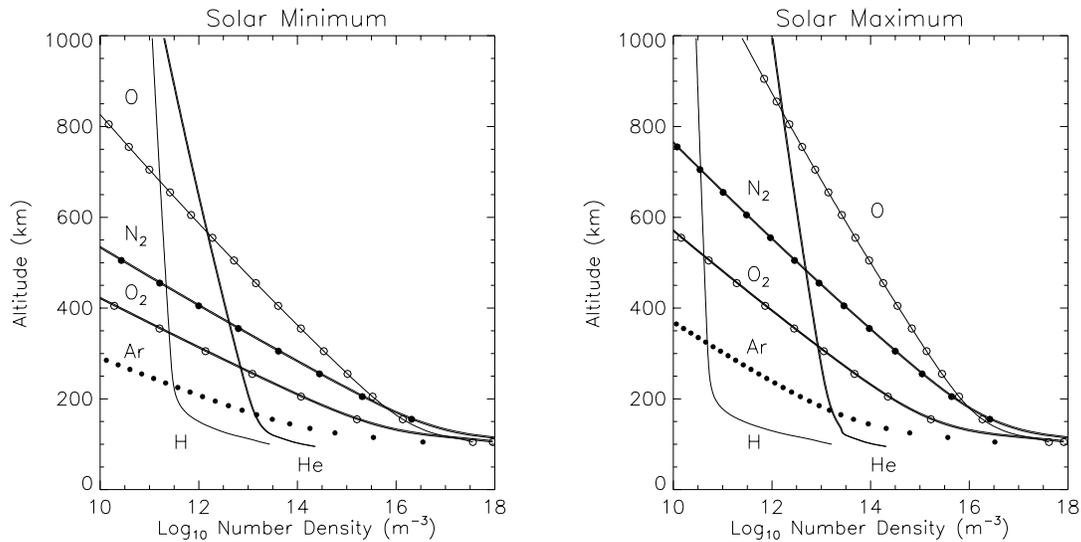


Figure 6.7 MSIS-E 1990 model. Relative concentrations of atmospheric constituents.

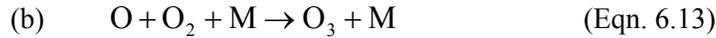
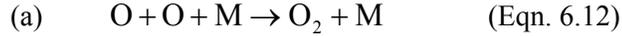
As the altitude increases further the ozone concentration decreases and at about 100 km atomic oxygen becomes an important constituent. Nitrogen remains mostly in its diatomic form and hence the region between 100 to 1000 km is a mixture of molecular oxygen, atomic oxygen, and molecular nitrogen with atomic oxygen increasing in relative concentration as the altitude increases. Between 1000 km and 2000 km helium appears to be the most important constituent and above 2000 km the atmosphere consists essentially of hydrogen.

The major photochemical process occurring in the atmosphere is the dissociation of molecular oxygen



which can occur at wavelengths below 250 nm and represents a very efficient absorption mechanism for the far ultraviolet portion of the solar spectrum.

One important loss mechanism for atomic oxygen is the three body recombination reaction which can take two forms:



where the third body M takes up some of the energy and momentum of chemical bond formation. Since these two reactions are but two of literally dozens which enter in the establishment of the steady state it is obvious that a quantitative determination of actual equilibrium concentrations of say O_3 as a function of altitude and time is a very complicated process involving reaction rates for many processes. Nevertheless, we can roughly see how the ozone layer represents a steady state determined by production of O_3 :



The two major loss mechanisms are:

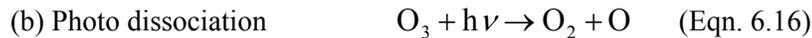


Photo dissociation of ozone occurs very efficiently for photons in the 200 to 300 nm wavelengths region. Although the ozone concentration never exceeds a few parts per million even in the so-called "ozone layer" it nevertheless represents a most effective filter for potentially harmful uv radiation. At 255 nm where the ozone absorption curve peaks the absorption coefficient is about 276 cm^{-1} , the ozone layer has a thickness of about 0.25 cm (at $p = 760 \text{ mm Hg}$ and $T = 18^\circ \text{ C}$).

Thus the vertical attenuation is $I = I_0 e^{-0.25 \times 276}$ or $\frac{I}{I_0} = e^{-69} \approx 10^{-30}$

which is a pretty good attenuation considering how few ozone molecules there are in this so called ozone layer.

E Transmission of Electromagnetic Waves Through the Atmosphere

As shown in Figure 6.8 the atmosphere is largely opaque to EM waves but there are several important "windows" through which radiation of certain wavelengths does reach the surface of the earth. Practically all our knowledge of the external universe comes to us through these "windows". Furthermore the photon energy of the radiation passing through the "optical window" is just right to sustain biological processes. If that window were to shift in either direction by an appreciable amount life as we know it would not be possible. We shall consider 8 regions of the spectrum (Figure 6.8) and briefly discuss the various processes in the atmosphere which do (or don't) absorb the incident radiation.

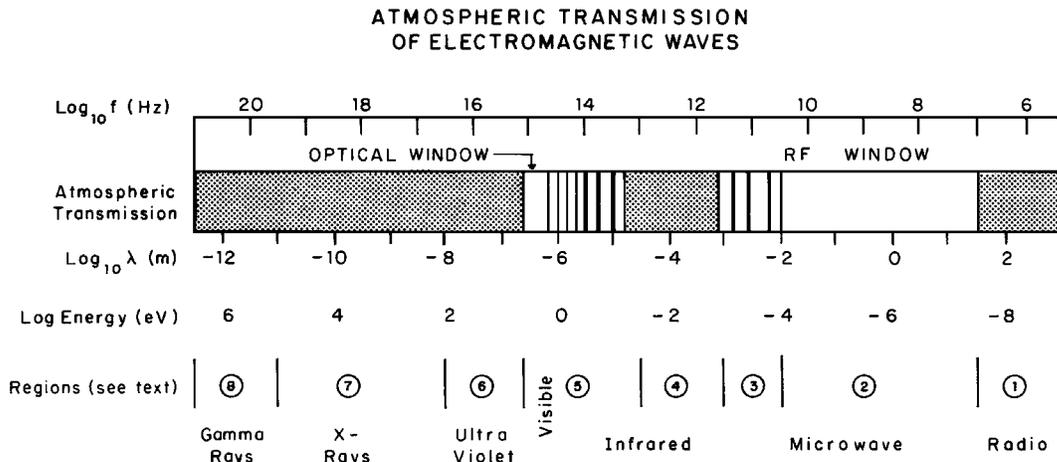


Figure 6.8 Atmospheric transmission of electromagnetic waves

1) RF Region:

Strong interaction (not necessarily absorption) up to 5 or 10 MHz. Any frequency below the plasma frequency f_p ($f_p = 9\sqrt{n_e}$) will not propagate, hence the atmosphere is opaque (see chapter 7)

2) RF Window:

10 MHz - 20 GHz.

No effective absorption process occur, hence the atmosphere is transparent. Frequency region of radio astronomy.

3) Microwave Absorption Region:

20 - 300 GHz ($\lambda = 1.5$ cm to $\lambda = 1$ mm). Absorption by molecular rotations. H_2O and O_2 have major rotational modes at 22 and 60 GHz respectively. Absorption in this region is intermittent with respect to wavelength.

4) Infrared Absorption Region:

$\lambda = 1000$ to 30 μm ($f = 3 \times 10^{11}$ to 10^{13} Hz). Absorption by vibrational transitions in H_2O , CO_2 and to some extent O_3 . Rotational levels spread vibrational levels into bands. The Black Body spectrum of the earth is 4 - 60 μm peaking at 15 μm . Thus the long wavelength part of the earth radiation is completely absorbed by the atmosphere.

5) Optical Window: $\lambda = 30 \mu\text{m}$ to $.3 \mu\text{m}$.

In the IR portion of the window ($30\mu\text{m}$ to $.7\mu\text{m}$) there is intermittent absorption by all molecular species present (H_2O , CO_2 , N_2O , CH_4 , O_3 , etc.) leading to strongly frequency dependent transmission. In the visible portion of the window ($0.7 \mu\text{m}$ to $0.4 \mu\text{m}$) there is very little absorption.

6) UV and Soft X Ray Region: $\lambda = 300$ to 10 nm . ($3000 \text{ \AA} - 100 \text{ \AA}$)

Near UV Region: Absorption by photo dissociation of O_3 . At shorter wavelengths electronic transitions in O_2 . Below 100 nm absorption by electronic transitions in N_2 and O .

7) X-Ray Region: $\lambda = 10$ to $.01 \text{ nm}$. ($100 \text{ \AA} - 0.1 \text{ \AA}$)

Absorption by excitation of inner electrons in both atomic and molecular species (mostly N_2 and O_2).

8) Gamma Ray Region: $\lambda < .01 \text{ nm}$.

Absorption by Photoelectric Effect and Compton Scattering in all species present in atmosphere.

Some perspective on the absorption and transmission windows can be obtained by means of the curve shown in Figure 6.9. This figure gives the depth of penetration for various wavelengths into the atmosphere from above. The atoms and molecules responsible for the absorption are identified, and in particular the fate of Lyman α and Lyman β .

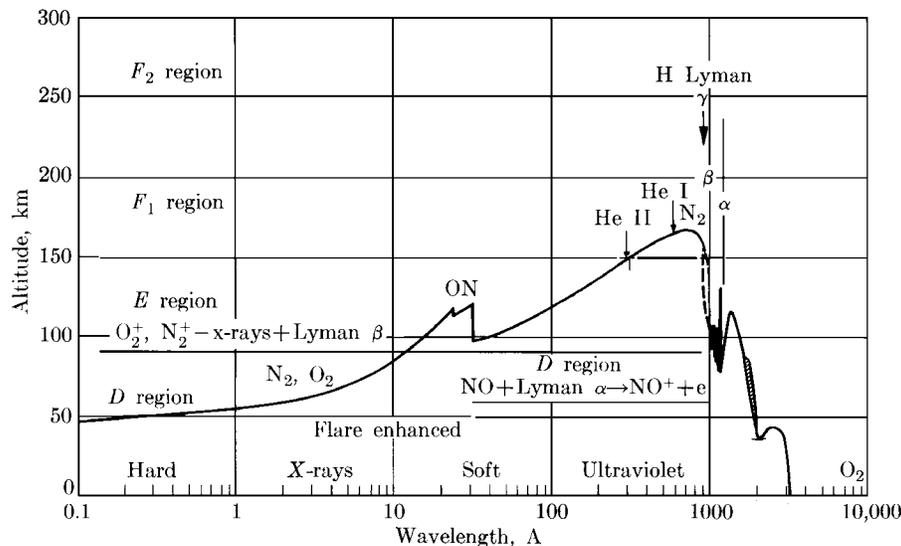


Figure 6.9 The depth to which $1/e$ of the incident solar intensity penetrates, for 0.1 to $10,000 \text{ \AA}$. Copied from "Atomic and Space Physics", p 131, by Alex Green and Philip Wyatt, 1965. ($1 \text{ nm} = 10 \text{ \AA}$)

Originally: R. Tousey, "Ultraviolet Spectroscopy of the Sun", Chapter 1, Space Astrophysics, W. Liller ed., 1961

Finally perspective on the visible and IR portions of the electromagnetic spectrum is given in Figure 6.10, which shows characteristic black body spectra for the sun and earth, along with the absorption percentage on the ground, and at 11 km altitude. Note the considerably diminished effect of H₂O at 11 km altitude.

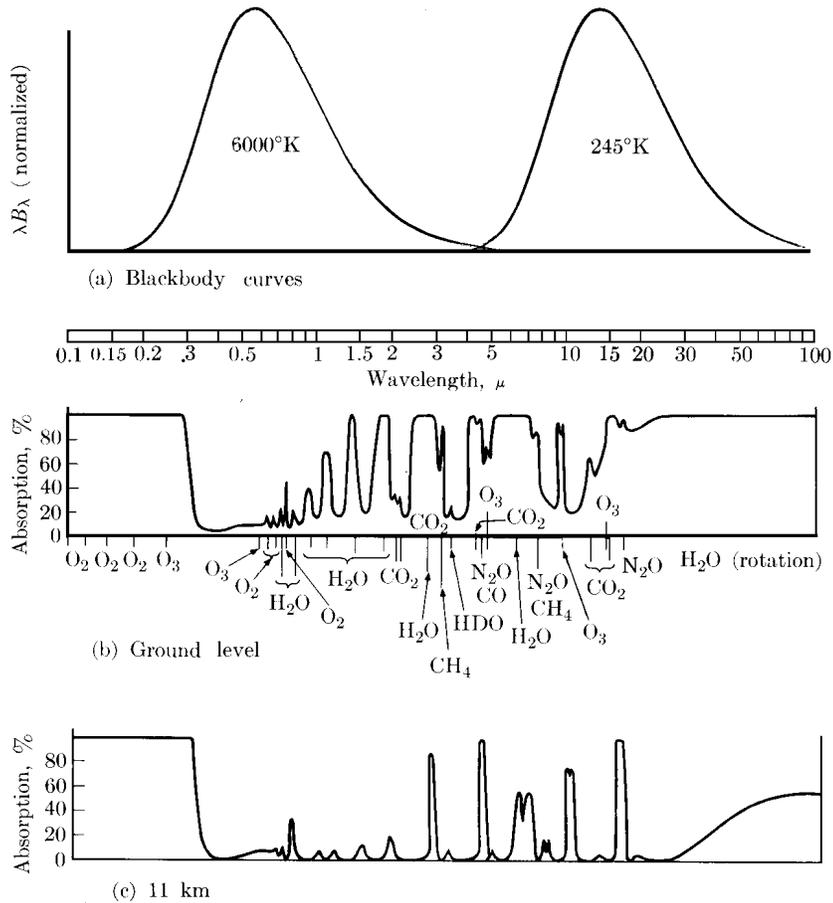


Figure 6.10. Atmospheric absorptions. (a) Blackbody curves for 6000 K and 245 K. (b) Atmospheric gaseous absorption spectrum for a solar beam reaching ground level. (c) The same for a beam reaching the temperate tropopause. The axes are chosen so that areas are (a) proportional to radiant energy. Integrated over the earth's surface and over all solid angles the solar and terrestrial fluxes are equal; consequently, the two blackbody curves are drawn with equal areas beneath them. An absorption continuum has been drawn beneath bands in (b). This partly hypothetical because it is difficult to distinguish from the scattering continuum, particularly in the visible and near infrared spectrum. Conditions are typical of midlatitudes and for a solar elevation of 40° or diffuse terrestrial radiation. Copied from "Atomic and Space Physics", p 130, by Alex Green and Philip Wyatt, 1965. Originally from: R. M. Goody, Atmospheric Radiation I. Theoretical Basis, Clarendon Press, Oxford, 1964.

F Problems

1. Estimate the scale height above 200 km altitude, using the information in Figure 6.5b. How does this compare to the values in Figure 6.4? Now calculate the scale height for N_2 using a temperature of 1000 K (ignore the altitude dependence of g).
2. If the temperature of the upper atmosphere doubled (from 1000 to 2000 K), how would the number density at 700 km altitude change (give a number). In order to do this you should use the scale height calculated in problem 1, assume the atmosphere is isothermal, and that the density at 200 km does not change. (This type of temperature increase is the source of some of the changes seen in Figure 6.7)
3. A) If the temperature of human skin is 98.6 F, at what wavelength does the blackbody radiation from skin peak? Which IR window is best for night time surveillance of humans?
B) If the average temperature of the stacks on a surface ship is 750 K, at what wavelength does blackbody radiation from the stacks peak? Which IR window is best for night time surveillance of ships?