THE ROLE OF SOLAR RADIATION IN DETERMINING THE STRUCTURE OF THE EARTH'S ATMOSPHERE

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1. OVERVIEW ON THE ORIGINS OF THE ATMOSPHERE

Although every planet in our solar system with the exception of Mercury, has an atmosphere, the earth's atmosphere is probably the one that is composed almost entirely of nitrogen and oxygen with a trace of water vapor. Mars and Venus have atmospheres composed mostly of carbon dioxide, while Jupiter and Saturn have atmospheres that are comprised mostly of hydrogen, helium, methane and ammonia.

Amid much speculation, on the origins of the earth's atmosphere, one deduction seems fairly certain: when the earth formed approximately 4.5 billion years ago, it was much too hot to retain any atmosphere that it may potentially have had. This is because high temperatures imply high kinetic energy that result in high molecular speeds. The force that keeps these moving gas molecules from escaping from a planet is gravitation which depends on the masses of the attracted bodies. Earth is massive and cool enough at the present time to retain the thin envelope of gases that form its atmosphere.

At the beginning of the existence, its original atmosphere was most probably composed chiefly of a mixture of methane (CH_4) and ammonia (NH_3), together with other gases such as compounds of chlorine (hydrogen chloride HCl), fluorine (hydrogen fluoride HF), and sulfur -(H_2S). It needs to be noted that methane and ammonia are even now important constituents of Jupiter, Saturn, Uranus and Neptune. It is almost certain that the early earth atmosphere was almost devoid of free oxygen. According to one theory, the earth's present atmosphere did not evolve until much of the original atmosphere had been driven off, due to high gaseous molecular speeds, and the earth had started to cool. This new atmosphere was formed when gases that had been dissolved in molten rock bubbled out to the surface. These first gases were primarily steam at first together with some carbon dioxide and nitrogen, which are the principal gaseous emissions from active volcanoes.

As cooling continued, the water vapor condensed to form the great oceans. This liquid mass of water gradually absorbed almost of the atmosphere's carbon dioxide, thus leaving nitrogen as the predominant gas. The oxygen in the atmosphere is understood to have appeared only after the appearance of primitive plant life approximately 800 million years ago. This plant life acted on the carbon dioxide through photosynthesis to form oxygen. An alternative theory for the production of oxygen in the atmosphere, centers on the photolysis of water vapor and the subsequent escape of hydrogen from the earth's atmosphere. The present mixture of essentially nitrogen and oxygen has therefore probably been in existence since sometime after the advent of photosynthesis.

2. SPECTRAL DISTRIBUTION OF SOLAR RADIATION WITHIN THE ATMOSPHERE

Solar radiation reaches the top of the atmosphere at a mean rate of 1.9 calories cm^{-2} min^{-1}. On entering the atmosphere, the radiation is modified in intensity and polarization. Intensity changes are introduced by absorption by atmospheric gases and particulates and by absorption and reflection by the underlying terrestrial and water surfaces. Polarization of solar radiation results from scattering by molecules, water droplets, dust, and other aerosol particles that are present in the atmosphere.

The spectral distribution of solar radiation incident at the top of the earth's atmosphere is compared to that emitted from a black body at a temperature of 5800^{o}K in the upper two curves of Figure 1. The significant discrepancies that occur in the ultra-violet region are primarily due to the electronic transitions that occur in the outer layer gases surrounding the sun (Coulson 1975). Beyond the regions of these transitions, the two curves exhibit a similar shape and magnitude. The various processes in the earth's atmosphere operate to change spectral distribution as solar radiation traverses the atmosphere. This modified distribution is sown in the lower of the two curves in **Figure 1**. The main absorption is produced by water vapor which is apparent in the strong bands in the infrared region, and by high altitude atmospheric ozone (O_3). The latter is effective in limiting radiation that reaches the surface in appreciable quantities, to wavelengths greater than approximate 0.3µm. relatively small amounts of energy are absorbed by ozone and oxygen (O_2) in the 0.6 µm to 0.7µm range. Scattering of radiation, which is particularly important at the shorter wavelengths, is responsible for the decrease in surface radiation indicated by the bands in the visible and near ultra-violet spectral regions.

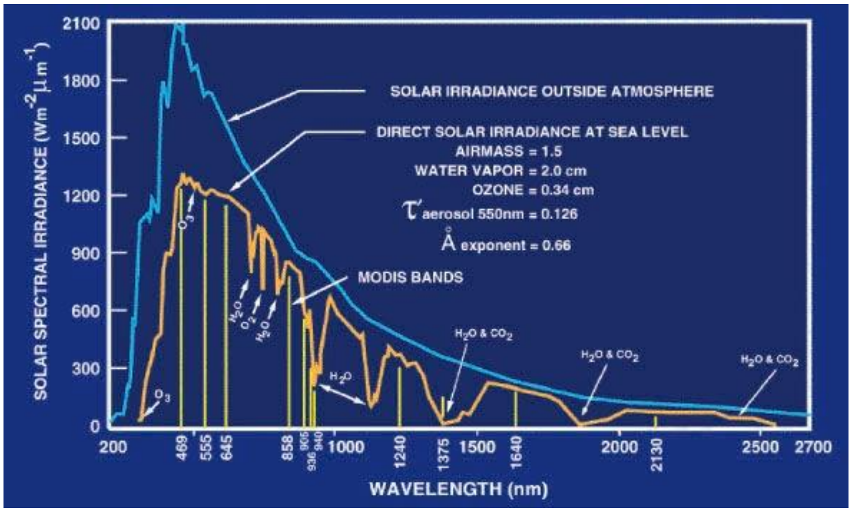


Figure1- Source: http://objectivistindividualist.blogspot.com/2013/02/infrared-absorbing-gases-and-earths.html

It is this absorption and scattering of solar radiation that contributes to the modification of the atmosphere as discussed in the following sections.

3. COMPOSITION AND STRUCTURE OF THE EARTH'S ATMOSPHERE

THE HOMOSPHERE AND HETEROSPHERE

The earth's atmosphere consists of a mixture of various gases that surround the planet up to heights many miles. Held to the earth by the gravitational force, this spherical annulus of gaseous mass is most dense at the terrestrial surface and decreases in density rapidly in the vertical direction (normal to the surface). Although 97% of the mass of the atmosphere lies within 29km (18mi) of the earth's surface, the upper limit off the atmosphere cannot be drawn sharply because the density of gas molecules decreases into the near-emptiness of interplanetary space (Strahler).

Upwards from the earth's surface to an altitude of approximately 100-120km, the chemical composition of the atmosphere is highly uniform. Consequently, the term homosphere has been applied to this lower uniform layer. In contrast the overlying heterosphere which is non-uniform can be characterized as an arrangement of spherical shells.

The homosphere consists of:

i. A mixture of gases referred to collectively as 'pure' dry air

ii. Water vapor
iii. Aerosols – a combination of minute particles is suspension.

The first two components are true gases composed of discrete molecules, whereas dust particles consist of solid particles much larger than molecules but tiny enough to mix freely with the gases and stay aloft indefinitely.

Of the gases (dry air) in the atmosphere, nitrogen (N_2) at 78% and oxygen ($O_{2)}$ at 21% make up 99% of the gases by volume. Other gases including argon (A), carbon dioxide ($CO_{2)}$, neon (Ne) , helium (he), krypton (Kr), xenon (Xe), hydrogen (H_2), methane (CH_4) and nitrous oxide (N_2O) make of the remaining volume (Tables 1 & 2).

Gas	Volume (%)
Nitrogen	78.084
Oxygen	20.9476
Argon	0.934
Carbon Dioxide	0.0314
Neon	0.00182
Helium	0.000524
Krypton	0.000114
Xenon	0.000087
Hydrogen	0.00005
Methane	0.00002

Table 1. Non-varying gases of the Troposphere

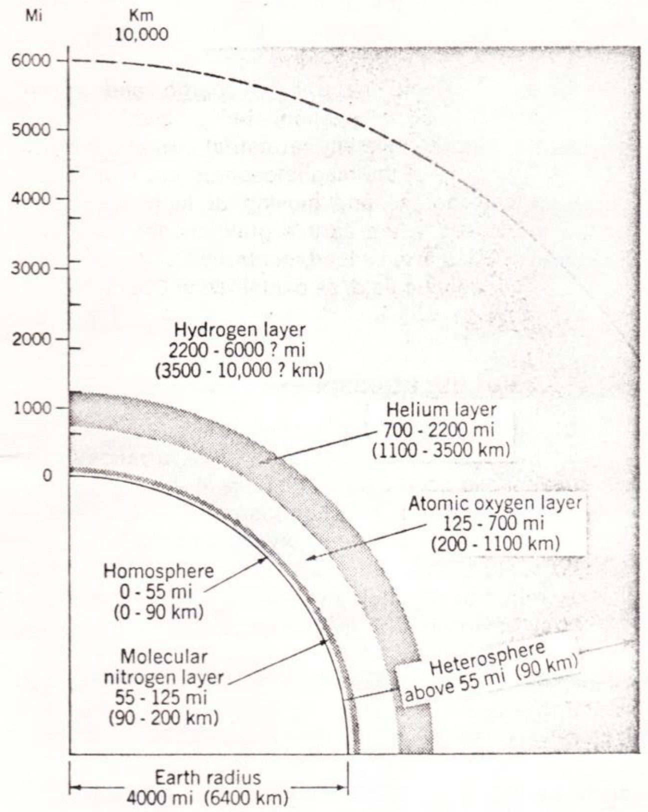
The second major component of the homosphere is water vapor, the gaseous state of water. In this state, water (H_2O) molecules have the same freedom of movement as molecules of nitrogen, oxygen and the trace gases. Therefore the water molecules mix completely within the homosphere. Dispersion of gases due to diffusion becomes more important at altitudes above 90km, where the effects atmospheric motions are minimized.

Aerosols in the lower atmosphere consist of particles of dust from the land and ocean surfaces. Smoke from grass and forest fires is an important source, as are the winds blowing over arid regions which introduce mineral particles into suspension, often up to a few kilometers above the surface of the earth. Active volcanoes contribute dust particles, whose travel via global atmospheric circulation channels, can be traced thousands of miles away from the source. Dust is also added by meteorites that vaporize upon entering the earth's upper atmosphere, creating countless solid particles in the process.

Especially important in the formation of clouds and precipitation are the tiny salt crystals that result from the evaporation of ocean spray droplets swept up in turbulent winds blowing over the crests of breaking waves.

The heterosphere, encountered at altitudes greater than 100kms consists of four layers, each of distinctive composition (Figure 2). Lowermost is the molecular nitrogen layer consisting dominantly of molecules of N_2 and extending vertically to approximately 200km. Above this height lies the atomic oxygen layer consisting dominantly of oxygen atoms (O). In the region from 1100km to 3500km, lies the helium layer composed of atomic helium (He). Beyond this region hydrogen atoms (H) form the hydrogen layer.

FIGURE 2 Homosphere and heterosphere. (Based on data of R. Jastrow, NASA, and M. Nicolet.)



It needs to be noted that the four layers described here have transitional boundary zones rather than sharply define=d surfaces of separation. The arrangement of these predominant gases is driven by their molecular and atomic weights. Molecular nitrogen the heaviest is lowest in the four layers and atomic hydrogen, the lightest, forms the outermost layer. Furthermore, it should be noted that at the extremely high altitudes of the heterosphere, the density of the gas molecules and atoms is extremely low (order of 10^{-11} g/cm^3).

The homosphere and heterosphere comprise the first order of subdivision into layers and these in turn are subject to further subdivisions. For the most part, the layers are not all sharply defined and their limits arbitrarily established based on a number of physical and chemical attributes. As the various physical and chemicals properties are arranged in altitude zones, the system of vertical layering and the names applied to the layers depend on the class of properties selected.

METHODS OF DEFINING THE VERTICAL STRUCTURE OF THE ATMOSPHERE

One method or basis for describing the structure of the atmosphere as it changes with altitude is on the basis of density and pressure variations. These properties change relatively smoothly with height without distinct discontinuities in their values that could be used for distinguishing layers.

Another basis for subdividing the atmosphere into layers is according to the variation of temperature that is observed with increasing altitude. Because rather rapid temperature changes are encountered at various levels within the atmosphere, this property has provided the basis for one of the principal schemes of naming the atmospheric layers.

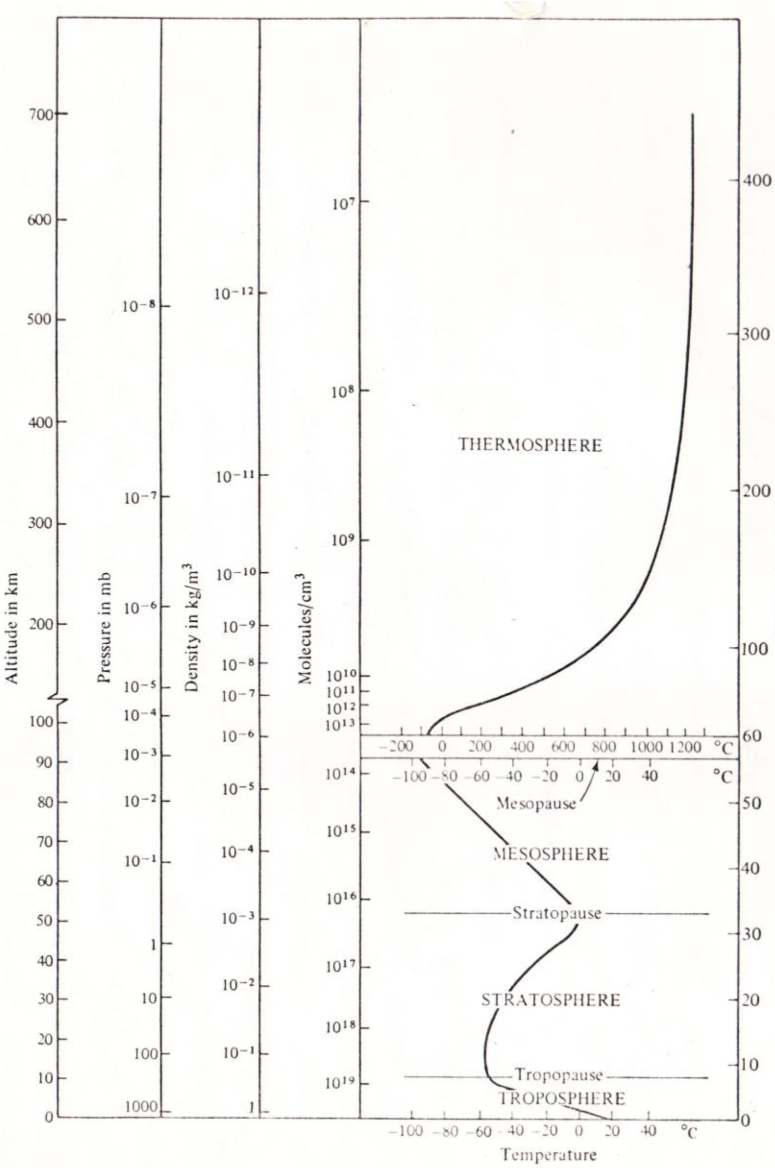
Yet another, and very useful, system of defining atmospheric layers is based upon the chemical and electrical properties of the atmosphere. As an example, we find one layer rough between the heights of 20km and 130km in which chemical reactions are induced by the direct action of solar radiation. Above this is a zone in which solar radiation produces multiple electrical phenomena. Still higher in the atmosphere, we find a zone in which molecules from the atmosphere can escape into 'outer' space.

The sun is the principal source of energy available to the earth and its atmosphere. Only through the absorption of solar radiation and emission of radiation from the earth and the atmosphere into space, can our planet receive and loose energy. The solar received by the earth-ocean-atmosphere system is converted into internal energy which in turn is converted to potential energy, latent heat and kinetic energy. Thus atmospheric and oceanic motions are derived from solar radiation coupled with the earth's rotational forces.

PRESSURE, DENSITY AND TEMPERATURE VARIATIONS IN THE TROPOSPHERE

The upper atmosphere derives most its energy directly by absorption through various atmospheric gases although the energy so absorbed represents only a few percent of the total solar energy reaching the earth's atmosphere. The troposphere, in contrast, obtains its energy indirectly from re-radiation, convection and conduction from the earth's surface and the atmosphere (Kriester, 1968).

One can infer from this that solar radiation does not have a direct influence in determining the vertical structure of the lower atmosphere or troposphere. However, visible and near infrared radiation is of prime importance in this region of the atmosphere for reasons that will become apparent later.



(a)

FIG. 3 Vertical distribution of atmospheric properties and phenomena. (Miller + Thompson, 1970. p 4-5)

In studying the structure of the troposphere we take into consideration the vertical distribution of temperature (T), pressure (p) and density (ρ). These physical properties of the atmosphere are related to each other as is evident in the equation of state and hydrostatic equation:

$$p\alpha = R'T \ (1)$$

$$\frac{\partial p}{\partial z} = -\rho \mathrm{g} \ \ (2)$$

If the geopotential is defined as

$$\Phi = \int_0^z \frac{Fg}{m} = \int_0^z g \, dz \ \ (3)$$

Then

$$d\Phi = gdz$$

⇒ The hydrostatic equation (2) can be written as:

$$\frac{\partial p}{\partial \Phi} = -\rho \ \ (4)$$

Substituting the equation of state (1) into (4) gives:

$$\frac{\partial p}{\partial \Phi} = -\frac{p}{R_m T} \ \ (5)$$

For the isothermal case equation (5) can be integrated to give the vertical pressure gradient in the form known as the barometric equation:

$$p = p_0 \exp(-\Phi R'T) \ \ (6)$$

Similarly the vertical density distribution is found to be of the form:

$$\rho = \rho_0 \exp(-\Phi R'T) \ \ (7)$$

This means that both pressure (p) and density (ρ) decrease exponentially with height.

The troposphere is the region where most of the weather phenomena are concentrated. This means that there is a high degree of turbulence in existence here. This turbulence leads to much mixing of the air hence it is also known as the homosphere. Within the homosphere the mean molecular weight is taken as constant. Given that the weather systems responsible to this turbulent mixing derive their energy primarily from solar radiation (direct and indirect), we see that the homogenous nature of the troposphere is indirectly related to solar radiation.

THE CHEMOSPHERE AND 'OZONOSPHERE' AS REGIONS IN THE STRATOSPHERE

The region above the troposphere, known as the stratosphere extends from about 15km to approximately 50km. The stratosphere is one of the regions where solar radiation has a direct and significant effect on the vertical structure of the earth's atmosphere.

There are two major regions within the stratosphere: the chemosphere and ozonosphere. However as the ozonosphere is the result of chemical reactions in the chemosphere, the two are usually regarded as one. This shell (spherical annulus) is known for the important chemical reactions produced by the direct effects of solar radiation.

In the region known as the ozone layer, primarily concentrated in the altitude range of 20km to 35km, the absorption of 0.2μm to 0.3μm ultraviolet rays produces an important chemical effect upon the atmospheric oxygen. This effect is evident to a lesser at altitudes as low as 10km above the earth's surface.

The effect is the result of a photochemical reaction that leads to the dissociation of molecular oxygen by the absorption of a photon. That is, and an oxygen molecule (O_2) will dissociate completely into atomic oxygen (O) when exposed to ultraviolet radiation below ~0.24μm.

The reaction is expressed by the following equation:

$$O_2 + h\upsilon \longrightarrow O + O \quad (8)$$

Where $h\upsilon$ is the energy of the photon with a frequency of υ. The minimum energy required for this reaction corresponds to a wavelength of 0.2424μm.

Following the formation of atomic oxygen, a series of further reactions may occur, of which the most important are:

$$O + O + M \longrightarrow O_2 + M \quad (9)$$

$$O_2 + O + M \longrightarrow O_3 + M \quad (10)$$

$$O_3 + O + M \longrightarrow 2\,O_2 \quad (11)$$

The resultant O_2 in (11) is in an excited state. The third body M in equations (9) and (10) may be any particle that is capable of absorbing the extra energy and momentum that is released by the reaction.

Equation (10) represents the process that is primarily responsible for the production of ozone (O_3) in the atmosphere. This is an exothermic process so energy is released and absorbed by the third body M. This ozone is decomposed again by the reaction (11) and by the additional important reaction:

$$O_3 + h\upsilon \longrightarrow O_2 + O \quad (12)$$

This reaction is also an exothermic process and this is the main source of heating that is observed in this region of the atmosphere **(Fig 4).** The production and destruction of ozone occurs simultaneously and a level of equilibrium is maintained.

The ozone layer shows seasonal changes in altitude and degree of concentration as depicted in Fig 5. In general the layer is at its lowest height in winter and at its highest in summer. It has also been found that the center of concentration of the ozone layer varies in height with latitude, being much higher in low latitudes than it is the high latitudes **(Fig 5).**

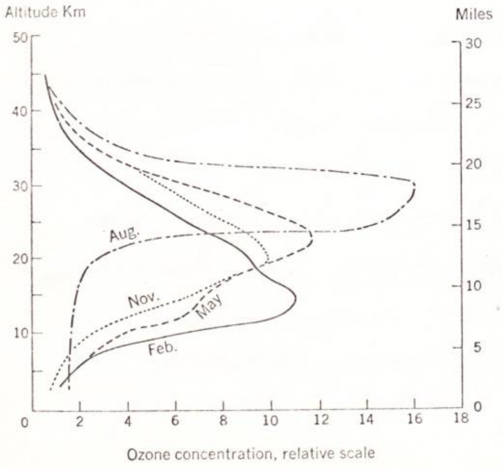


FIGURE 4 This graph of the concentration of ozone in the atmosphere above Flagstaff, Arizona, shows that seasonal changes occur in the level of greatest concentration. [After *Handbook of Geophysics* (1960), New York, Macmillan, p. 8-5.]

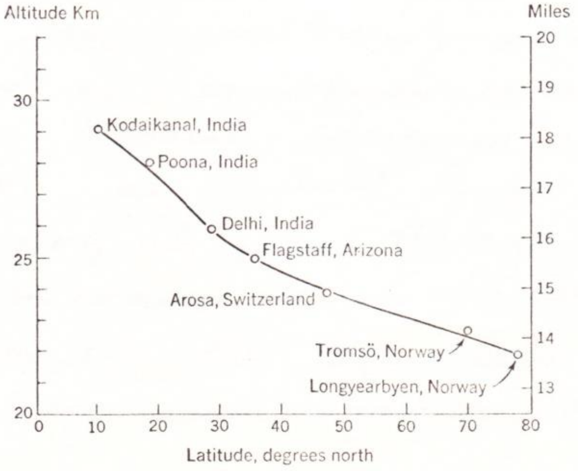


FIGURE 5 When the average elevation of greatest concentration of ozone is plotted against latitude, there results a marked decrease in level as latitude increases. [After *Handbook of Geophysics* (1960), New York, Macmillan, p. 8-6.]

As stated by Strahler (1963), “ozone would be a deadly poison to life forms if present in large concentrations, but fortunately it is almost completely absent in the troposphere. At the same time ozone is essential for the existence of life on earth, as it absorbs the shorter wavelength UV radiation that would otherwise destroy all exposed bacteria and severely burn animal tissues". The ozone layer also absorbs much of the longer wavelength UV radiation and some of the visible and infrared wavelengths as well. This absorption and the resulting exothermic reactions heat up the atmosphere causing the temperature maximum of the stratosphere at an altitude of approximately 50km.

THE MESOSPHERE, IONOSPHERE AND THERMOSPHERE

The mesosphere occupies the volume from 50km to 85km. This is a region of decreasing temperature, where the temperature is seen to decrease from 270^{o}K at the stratopause, down to 180^{o}K at the mesopause. The decrease is temperature is brought about because of an insufficient supply of the third body M in equation (10): $O_2 + O + M \longrightarrow O_3 + M$.

This results in an insufficient conversion of oxygen to ozone, hence there is a decrease in the temperature with height as both amounts of oxygen and M present, is reduced. There are varying amounts of N_2, O_2, CO_2, O_3 He, NO, and NO_2 present in the mesosphere. The upper portion of the mesosphere forms the base of what is known as the ionosphere. In the ionosphere molecules are converted and existed primarily in their ionic form.

Above the mesosphere lies the thermosphere, which also includes the ionosphere. This is a region of constantly increasing temperature and does not have an upper limit, though the limit has been quoted as 500km by some researchers. The temperature increase with height results from the absorption of radiation down to 0.2μm and the dissociation of O_2 to produce ions and heat: $O_2 + hv \longrightarrow O + O$. There is little production or conversion of oxygen to ozone.

It is in thermosphere that the bulk of the ionosphere is concentrated. As stated earlier this is the region where molecules are converted to atoms and ions, the majority of which is concentrated above 100km.

The ionosphere is formed when molecules of primarily oxygen and nitrogen in the upper atmosphere absorb the highly energetic gamma rays ($10^{-6} – 10^{-4}\mu$m), X rays ($10^{-4} – 10^{-4}\mu$m), and ultraviolet rays ($10^{-1} – 0.4\mu$m) of the solar spectrum. In doing so, each impacted molecule or atom loses an electron, and becomes a positively charged molecule or atom, known as an ion. Typical process that occur are:

F2 Layer: $O + hv \longrightarrow O^{+} + e^{-}$ (14)

E1 Layer: $O_2 + hv \longrightarrow O_2^{+} + e^{-}$ (15)

$$\text{E2 Layer: } ON_2 + h\upsilon \longrightarrow N_2^{+} + e^{-} \quad (16)$$

The ionization process begins at an altitude of 100km and is effective down to approximately 50km. The atmosphere becomes increasingly rarefied upward, and the intensity of the shortwave radiation decreases downward. This results in a region from about 80 to 400km above the earth’s surface where the concentration of positive ions and negative electrons is most dense. It is this region that forms the core of the ionosphere.

Electrons that ejected from the molecules of nitrogen and oxygen are free to ‘travel’ as an electric current. Thus we can think of the ionosphere as a series of electrically conducting layers. This is of prime importance in radio wave propagation.

Because the formation of ions in the ionosphere requires solar radiation we should expect the greatest density of ions to be present on the illuminated side of the earth and the lowest density on the shadow or dark side of the planet. **Figure 6,** shows diagrammatically the way in which the ions form into a number of dense layers over the sunlit hemisphere.

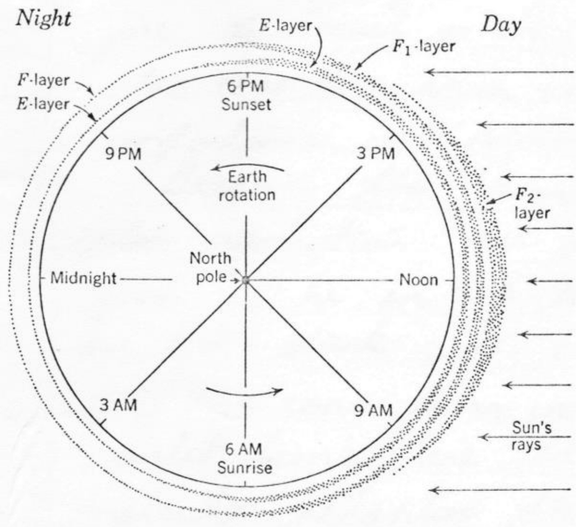


FIGURE 6 A schematic representation of the ionospheric layers under the sun's rays and in hours of darkness. [After B. F. Howell, Jr., (1959), *Introduction to Geophysics*, New York, McGraw-Hill, p. 351.]

During the hours of darkness, the ionized molecules recapture free electrons; thus ionization rapidly disappears and only one or two thin layers remain over the darkened hemisphere.

In general, “The region from the mesopause to the lower ionosphere is one in which the relative amounts of the various atmospheric constituents change considerably. This has several important consequences, such as the change in the mean molecular weight M, and also affects considerably the radiation field. In addition to changes from molecular to atomic states following dissociation by solar radiation and changes from the various consequential chemical reactions, the composition of the atmosphere is also affected above 100km by diffuse separation. Above this altitude, atomic oxygen become progressively more dominant until about 800km, above which the lighter gases helium and hydrogen become the main constituents." (Fedele and Zancala, 1968).

Most of the major features up to 200km can be explained in terms of the chemistry of an oxygen-nitrogen atmosphere. As noted earlier the sun's ultraviolet radiation produces considerable dissociation of oxygen and also the minor constituents above 80km. Re-combination occurs following three-body collisions and photochemical equilibrium is approached. At some altitudes this is rapid and at others it is quite slow, so that the resulting distribution of the various constituents is dependent on atmospheric motions, mixing, diffusion as well as the photo-chemical processes. In addition, a number of chemical reactions between neutral and ionized products of the photochemistry also occurs.

"As a result of all of these effects, both theoretical considerations and actual measurements indicate that the distribution of the various elements is neither the same in in different seasons nor diurnally" (Fedele and Zancala, 1968).

4. CONCLUSION

In conclusion, from the discussion above, a direct effect of the earth's gravity is the influence that it has in determining the vertical pressure and density distributions of the atmosphere.

More importantly, other than the fact that solar radiation is the primary driving mechanism for all major atmospheric phenomena, one key observation stands out, that is, the effect of solar radiation on the constituents of the atmosphere is responsible for the defining the many aspects of the vertical structure of the earth's atmosphere. This observation holds true irrespective of the manner in which one describes the atmosphere, be it on a thermally stratified basis, chemical stratification or electrical stratification, because each of these methods of regionalizing the atmosphere either depends on or influences the other.

5. REFERENCES

Charles R. Anderson, Physics *Infrared-Absorbing Gases and the Earth's Surface Temperature: A Relatively Simple Baseline Evaluation of the Physics*' (February 18, 2013)
http://objectivistindividualist.blogspot.com/2013/02/infrared-absorbing-gases-and-earths.html

Coulson, K.L. *Solar and Terrestrial Radiation*, Academic Press, New York, 1975

Fleagle, R.G. and Businger, J.A. *An Introduction to Atmospheric Physics*, Academic Press, New York, 1963

Kondrayev, K.Y. (Ed.), *Radiation Characteristics of Atmosphere and the Erath's Surface*, Amerind Publishing – New Delhi, 1973

Nicolet, M., *Density of the Heterosphere Related to Temperature*, SAO Special Report #75 (1961)

Rawer, K. (Ed.), *Winds and Turbulence in the Stratosphere, Mesosphere and Ionosphere*, North-Holland Publishing, Amsterdam, 1968

Pettersen, S., *Introduction to Meteorology*, McGraw-Hill, New York, 1971