

Vertical Structure of the Atmosphere

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1 INTRODUCTION

This chapter reviews the vertical structure of the atmosphere from the surface to roughly 1000 km, as characterized mainly by temperature; the temperature is higher where solar radiation is absorbed. The focus in Section 2 is on four concentric atmospheric shells: the troposphere, stratosphere, mesosphere, and thermosphere; how temperature varies with altitude within each shell; and what physical processes control the temperature. Section 3 discusses standard atmospheres, idealized representations of average atmospheric conditions, particularly the vertical structure, suitable for engineering applications. The connection between standard atmospheres and aircraft altimetry is especially important for safe air travel and is specifically mentioned. Section 4 summarizes properties of the ionosphere, a layer containing both neutral and electrically charged particles extending from about 60-km altitude indefinitely upward. The charged

particles interact with the neutral particles and respond to the Earth's magnetic field.

A companion article, "Atmospheric Composition" (see Atmospheric Composition) covers (i) the definition of temperature on the molecular scale, (ii) elementary gas kinetics, which explains the behavior of the mixture of atmospheric gases at various altitudes, and (iii) the properties of each major atmospheric gas. The effects of weather on aircraft are covered in a second companion chapter on meteorology (see Meteorology). Space weather refers to physical processes occurring in the rarefied high atmosphere, where the incidence of high-energy particles and intense radiation is strongly affected by solar activity. The hazards posed by space weather to spacecraft and crew are discussed in Atmospheric Interactions with Spacecraft.

2 ATMOSPHERIC SHELLS DEFINED BY TEMPERATURE VARIATION WITH ALTITUDE

This section categorizes the atmosphere in terms of vertical temperature structure. Four main "shells," and the boundaries between them, are defined. In order of increasing altitude, they are: the *troposphere*, whose top boundary is called the *tropopause*; the *stratosphere*, whose top boundary is the *stratopause*; the *mesosphere*, whose top boundary is the *mesopause*; and the *thermosphere*. We consider each of these layers in turn and explain the temperature variation within them.

A *standard atmosphere* is an idealized representation of the atmosphere based upon long-term average conditions. Figure 1 shows the vertical temperature structure of the

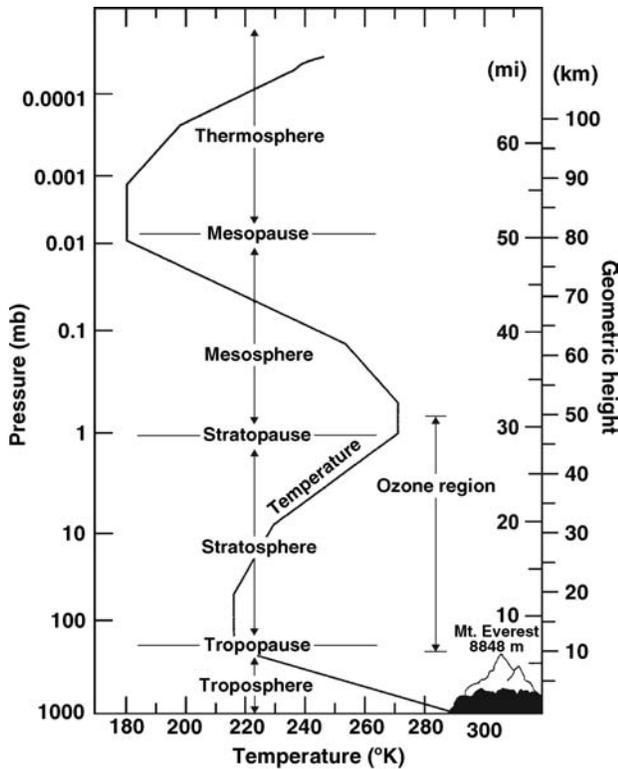


Figure 1. Vertical temperature structure of the atmosphere extending from the surface of the Earth to approximately 110-km altitude as given in the *US Standard Atmosphere* (1976). The principal layers defined by this temperature structure and the interfaces between them are labeled. As indicated, ozone is found principally in the stratosphere. Two vertical coordinates are given: pressure on the left in millibars (1 mb = 100 Pascal) and geometric altitude on the right (km). Adapted from Brasseur, Orland and Tyndall (1999) © Oxford University Press.

US Standard Atmosphere (1976), representative of mid-latitude conditions, to just beyond 100-km altitude, along with the atmospheric layers and interfaces. The temperature decreases with altitude in the troposphere, increases in the stratosphere, decreases again in the mesosphere, and then increases for good in the thermosphere, which extends far beyond the upper limit of Figure 1. More information on standard atmospheres follows in Section 3.

2.1 Troposphere

The troposphere is the lowest layer of the atmosphere, typically ranging from 9 to 17 km in thickness, in which temperature generally decreases with height. Clouds, precipitation, and storms inhabit the troposphere. This section explains why temperature decreases with altitude in the troposphere and how an equilibrium established between

radiative and convective processes determines the height of the tropopause.

Nearly half (47%) the amount of solar radiant energy incident at the top of the atmosphere is absorbed at the ground or in the oceans. Some of the absorbed energy is used to evaporate water, but much of it is involved in direct transfer of enthalpy from surface to atmosphere and infrared radiation upward from the surface. Convection explains why tropospheric temperature, on average, decreases with altitude. Tropospheric convection would still occur without water vapor but it would not be nearly so vigorous.

The decrease of temperature with height is called the *lapse rate*, expressed as $(-dT/dz)$. The atmospheric lapse rate seldom exceeds 9.8°C cooling per kilometer of altitude. This is called the *dry adiabatic lapse rate* Γ_d .

$$\Gamma_d = \frac{g}{c_p} \quad (1)$$

where g is the acceleration due to gravity ($\sim 9.8 \text{ m s}^{-2}$ at sea level) and c_p is the specific heat of air at constant pressure ($\sim 1005 \text{ J kg}^{-1} \text{ K}^{-1}$). Note that $dT/dz = -\Gamma_d$. The nomenclature arises because this is the rate at which a volume of cloud-free air, hypothetically isolated from its surroundings (no heat exchange), will cool if forced upward or warm if forced downward.

It is important to distinguish between a measured temperature profile (by an instrumented balloon or aircraft) and the parcel trajectory, the temperature change a volume of air would experience if forced upward or downward. If an air parcel is warmer than its surroundings at the same pressure, it is less dense (buoyant) and accelerates upward. Conversely, if it is cooler than its surroundings, it is more dense and accelerates downward.

Stability in the atmosphere is judged as follows. The atmosphere is considered *stable* if an air parcel, displaced upward or downward, experiences a restoring force and returns to its original position. It is considered *unstable* if a displacement results in a buoyancy force and acceleration away from its initial position. It is considered *neutral* if the initial displacement results in no force on the parcel.

The dry adiabatic lapse rate characterizes *neutral* stability because, if a volume of cloud-free air is displaced upward or downward under this condition, it continues on its way without acceleration. Rising or sinking parcels maintain the same temperature as their surroundings. Dry adiabatic lapse rates are common in the lowest few kilometers of the atmosphere on sunny summer afternoons. Thermals are common, and this is when soaring pilots like to fly.

The lapse rate is seldom greater than Γ_d except within centimeters of intensely heated ground. On the other hand, the lapse rate is often less than Γ_d . In fact, the temperature

sometimes increases with height. This is called an *inversion*. Inversions represent very stable stratification of the air because they strongly inhibit vertical motion. An air parcel rising through an inversion quickly finds itself cooler than its surroundings and sinks back toward its origin. The *tropopause* is marked by a semi-permanent stable layer, either an isothermal layer or an inversion, at altitudes of 9–17 km (higher in summer than in winter, and higher near the equator than near the poles). This stable layer invariably limits the vertical development of clouds.

Water vapor is the fuel for strong tropospheric convection. When water vapor condenses within a rising air parcel, heat is released. This boost of temperature within the air parcel can cause it to become warmer than its surroundings and remain buoyant, even as it rises many kilometers. When this happens, a tall cloud (often a thunderstorm) forms.

When the measured temperature and moisture profiles favor parcel buoyancy (due to condensation heating), the atmosphere is said to be *conditionally unstable*. Conditional instability occurs when the environmental lapse rate Γ (positive when temperature decreases with height) satisfies the following inequality:

$$\Gamma_m < \Gamma < \Gamma_d \quad (2)$$

where Γ_m is called the *moist adiabatic lapse rate*, the rate at which saturated (cloudy) air cools as it rises, and any condensate that forms does not fall out. The expression for Γ_m is considerably more complicated than that for Γ_d . It is derived in Bohren and Albrecht (1998, pp. 287–292). As noted earlier, the amount of vapor in saturated air (air on the verge of condensation) increases exponentially with temperature. In saturated ascent (for which Γ_m describes how the parcel temperature changes with altitude), the condensation rate and the heating within the parcel are both higher at higher temperatures because more moisture is present. In the lower troposphere, Γ_m is about 5°C km^{-1} . High in the troposphere, where the air is very cold and moisture is minimal, Γ_m approaches Γ_d .

In Figure 1, the standard tropospheric lapse rate is $6.5^\circ\text{C km}^{-1}$, less than Γ_d and within the normal range of Γ_m . We have already noted that the tropopause is highest over the equator and slopes downward toward the poles. Why should this be? The answer is that deep convection adjusts some very steep lapse rates that would occur if the troposphere were in pure radiative equilibrium.

Radiative equilibrium is said to occur if the upward and downward fluxes of radiation at each wavelength and at each atmospheric level are in balance. The simplest models of radiative equilibrium consider two broad wavelength intervals, which are very nearly separated: solar radiation up to

4- μm wavelength, and infrared radiation, emanating from the Earth's surface and atmosphere at wavelengths greater than 4 μm . The atmosphere is considered approximately transparent to solar radiation, both incident from the sun and reflected; thus, solar radiation hardly affects the radiative balance. Equilibrium thus requires equal upward and downward fluxes of only infrared radiation at each level. The calculations require knowledge of the absorbing and emitting properties of atmospheric gases.

The result of these simple models (Salby, 1996, pp. 233–240) is a nearly constant temperature profile above 15 km, then gradually steepening lapse rates at lower altitudes. The lapse rate for radiative equilibrium typically exceeds Γ_m by 6-km altitude and Γ_d by 5 km, becoming absolutely unstable below 5 km. Absolute instability cannot endure, and moist convection would quickly reduce the tropospheric lapse rate to a value below Γ_d . *Radiative-convective equilibrium* is the result. The point where the environmental temperature profile established by convection intersects at higher altitudes the nearly isothermal profile consistent with pure radiative equilibrium is effectively the tropopause.

Surface temperature is positively correlated with the height of the troposphere. In the tropics, where surface temperature and moisture supply are both higher than at most other latitudes, convection tends to be deeper and the troposphere higher. With increasing latitude, surface temperature lowers or, if not (e.g., in the subtropical deserts), moisture supply is usually limited. Convection, though sometimes more vigorous than in the tropics, is not as deep. In midlatitudes, seasonal swings in surface temperature bring corresponding swings in available moisture, frequency of convection, and tropopause height. During the polar winter, with little or no sunlight for weeks to months, surface inversions can become pronounced due to infrared radiation losses to the atmosphere and space. The tropopause can become ill defined under such conditions.

2.2 Stratosphere

The stratosphere extends from the tropopause to approximately 50-km altitude. Temperature increases throughout its depth, slowly near the bottom, more rapidly near the top. The stratosphere is therefore a very stable layer, not subject to convection. It is also dry because most of its water vapor enters through the tropopause, which, by virtue of its low temperature, is often called a “moisture trap.” Polar stratospheric clouds are occasionally observed, but only at exceptionally low temperatures. If fine particulate matter from an explosive volcanic eruption enters the stratosphere, the residence time is

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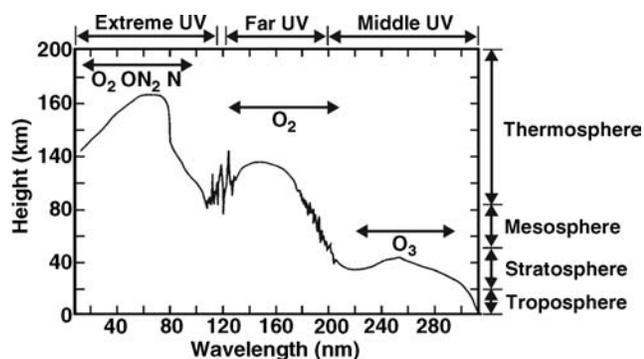


Figure 2. Altitude at which the intensity of ultraviolet (UV) radiation at each wavelength is reduced by absorption to $1/e$ of its value at the top of the atmosphere. Absorption results in both dissociation, the formation of atomic oxygen and nitrogen (O and N) from the corresponding molecules (O_2 and N_2), and ionization, stripping an electron from any of these species. Reproduced from Ondoh and Marubashi (2000) © Ohmsha, Ltd.

long, often more than a year, because there is no precipitation to wash the ash out.

The stratosphere contains 90% of atmospheric ozone (O_3), yet only a few molecules per million in the stratosphere are ozone molecules. The ozone mixing ratio is typically 3 parts per million by volume (ppmv) at 20-km altitude, rises to a maximum of 8–10 ppmv at 35 km, then decreases to 2 ppmv at the stratopause. Conditions in the stratosphere are dominated by radiative processes. The increase in temperature with altitude results from the absorption of solar ultraviolet radiation by ozone.

The absorption of solar ultraviolet (UV) radiation at stratospheric altitudes and above is fortuitous because unattenuated ultraviolet radiation at the Earth's surface would be harmful to plant and animal life alike. Figure 2 shows the altitude at which absorption lowers the intensity of ultraviolet radiation to $1/e$ (37%) of its value at the top of the atmosphere. Shorter wavelength radiation is generally absorbed at higher altitudes. UVA and UVB light, which most sun blocks screen effectively, occupy wavelengths of 400–320 nanometers (nm) and 320–280 nm, respectively. The principal gaseous absorbers at different altitudes are shown. As noted earlier, ozone is the principal absorber in the stratosphere.

2.3 Mesosphere

The mesosphere extends from the stratopause up to 80- to 90-km altitude. Molecular oxygen and nitrogen still comprise 99% of air, having virtually the same mixing ratios as at sea level. Tiny meteors burn up in the mesosphere regularly. High-speed collisions between meteorites and gas molecules generate enough heat to vaporize all but the

largest of them. The temperature decreases throughout the mesosphere, reaching a minimum of about 185 K at the mesopause. There is less UV absorption in the mesosphere than in the stratosphere, primarily because the concentration of ozone decreases with altitude. Ozone is still the most important absorber, followed by molecular oxygen. These are the two main sources of radiative warming. Carbon dioxide (CO_2) emits infrared radiation in the mesosphere, and is the major source of radiative cooling. The summer mesopause is the coldest region in the atmosphere. Temperatures as low as 170 K have been observed there. *Noctilucent clouds* of tiny ice particles occasionally form near the summer mesopause at high latitudes.

Radiative equilibrium alone does not explain the very low temperatures observed at the summer mesopause. In fact, despite the absence of sunlight at high latitudes of the winter hemisphere, the winter mesopause is not as cold. Large-scale upward motion and adiabatic cooling play a role in maintaining the low temperatures at the summer mesopause. Gravity waves are thought to be partly responsible for this; they influence, as well, the mean flow from west to east in the mesosphere. A gravity wave is a wave disturbance in which buoyancy acts as the restoring force on parcels of air displaced from hydrostatic equilibrium.

Most gravity waves emanate from the troposphere or lower stratosphere. Common sources are mountain ranges, vigorous convection, atmospheric fronts, and vertical shear of the horizontal wind. As they propagate upward from their source regions with a characteristic period of 1–2 h, gravity waves encounter less dense air and grow in amplitude, transporting energy and horizontal momentum vertically. Their energy per unit mass increases by one to two orders of magnitude as they propagate from troposphere to mesosphere. When the amplitude becomes sufficiently large, as often occurs in the mesosphere, the wave breaks, resulting in turbulence and energy dissipation. These breaking waves have a major effect on the mean mesospheric wind, causing local accelerations, referred to as *gravity wave drag*. They influence both the mean wind and the thermal structure at mesospheric altitudes (Fritts, 1995).

2.4 Thermosphere

The thermosphere extends upward from roughly 86 km and gradually merges with the cloud of neutral hydrogen that surrounds Earth. The temperature increases rapidly with altitude in the lower thermosphere up to about 120 km, then increases more and more slowly, approaching 1000 K by 700-km altitude. Because the temperature is so strongly affected by the absorption of ultraviolet radiation, it changes rapidly between

day and night. The temperature is also strongly affected by solar storms. The atmosphere is tenuous enough in the upper thermosphere that many satellites orbit the Earth at altitudes between 400 and 1000 km. Atmospheric density here is quite low but still sufficient to impose a weak frictional drag on spacecraft that affects orbital prediction. The aurora occurs at altitudes from 80 to 300 km. The ionosphere refers to the layer of free electrons and positively charged ions that overlaps the neutral thermosphere, and extends from the upper mesosphere to beyond 1000 km. It will be discussed in Section 4, but suffice it to say here that interactions among neutral gases and charged particles play a big role in the thermospheric structure.

The boundary between the homosphere and heterosphere (see Atmospheric Composition) lies near the bottom of the thermosphere, near 100-km altitude. Below this altitude, in the homosphere, eddy diffusion (bulk motion of large volumes of air) dominates atmospheric transport. Above this altitude, molecular diffusion controls the relative concentration of gases and the temperature profile. Absorption of extreme ultraviolet (EUV) at wavelengths less than 103 nm ionizes (strips electrons from) molecular and atomic oxygen and nitrogen. UV radiation also dissociates O₂ and N₂ into O and N so that, above 200-km altitude, the atomic species outnumber the molecular species. The number density of neutral particles in the thermosphere up through 1000 km is everywhere several orders of magnitude greater than the number density of electrons or positive ions.

According to Roble (2003), about one-third of the UV energy absorbed by the neutral gases of the thermosphere is used for local heating. The rest can be radiated away or transformed into chemical energy. Infrared cooling in the thermosphere is relatively weak, through emissions by CO₂, O, and nitric oxide (NO), but this is not enough to balance the UV heating, which is greatest in the high thermosphere because that is where the most energetic radiation is absorbed. Thus, the thermal stratification is controlled by downward molecular diffusion of heat to at least 120 km. The diffusion is particularly efficient above 500 km because the thermal diffusivity is so high; it is inversely proportional to the number density, which decreases by eight orders of magnitude between the mesopause and 1000-km altitude. That is why the temperature profile becomes isothermal high in the thermosphere.

3 STANDARD ATMOSPHERES

3.1 US Standard Atmosphere (1976)

Standard atmospheres are prescriptions of atmospheric properties, primarily as a function of altitude, useful for

engineering applications. One very widely used standard atmosphere is the *US Standard Atmosphere* (1976), already referenced several times. Table 1 is constructed from this source and summarizes the foregoing discussion of atmospheric shells. From the table, it is evident that pressure and density decrease by roughly an order of magnitude for each 15-km increase in altitude.

3.2 Other standard atmospheres

Other standard atmospheres have been developed in recent years. Among them are the International Civil Aviation Organization (ICAO) Standard Atmosphere (ICAO, 1993), which provides standard values of temperature, pressure, and density for levels up to 80 km and the underlying equations used to calculate them. This atmosphere contains no water vapor. It is used for calibration of pressure altimeters, evaluation of aircraft performance, and engineering design.

The International Standard Atmosphere (ISA) is published by the International Organization for Standardization (ISO, 1975). The ISA model for vertical profiles of pressure, density, temperature, and viscosity is based on average conditions at midlatitudes. The *US Standard Atmosphere* (1976) is identical to the ICAO Standard Atmosphere up to 32 km and to the International Standard Atmosphere up to 50 km. All are dry atmospheres.

The COSPAR International Reference Atmosphere (CIRA-86) (<http://badc.nerc.ac.uk/data/cira/>) is more detailed than other standard atmospheres in that it provides mean zonal wind (component of wind from the west) in addition to temperature, pressure and density, at altitudes up to 120 km, and at latitudes between 80°S and 80°N. At altitudes from 120 to 2000 km, CIRA-86 merges with the MSIS-86 standard atmosphere, mentioned in the next paragraph.

MSIS stands for Mass Spectrometer and Incoherent Scatter radar, the two primary sources of atmospheric data for development of MSIS standard atmospheres. MSIS-86 is an early version. The latest version, NRLMSISE-00, has been developed by the US Naval Research Laboratory (NRL) and has become a standard for international space research. It has been calibrated with actual satellite drag data. The *E* indicates that this standard atmosphere extends from the surface well into the exosphere; the 00 gives the year of release, 2000. The inputs to NRLMSISE-00 include year, day, time of day, latitude, longitude, altitude, local solar time, and information about recent solar activity and the magnetic index. The output is number density for primary atmospheric gases, total mass density, and temperature. For more information, go to <http://www.nrl.navy.mil/content.php?P=03REVIEW105>.

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Table 1. Temperature, pressure, and density as a function of geometric altitude from the *US Standard Atmosphere (1976)* for each of the main atmospheric layers. Dry air is assumed. Levels where the lapse rate changes are included, except in the thermosphere. A two-digit number preceded by the plus or minus sign indicates the power of ten by which the preceding number is to be multiplied.

Geometric altitude (m)	Temperature (K)	Pressure (hPa) ^a	Density (kg m ⁻³)
Troposphere – Constant lapse rate 0.65 K km ⁻¹ (cooling with altitude) from 0.0 to 11.0 km.			
0	288.150	1.01325 +03	1.2250 +00
1 000	281.651	8.9876 +02	1.1117 +00
2 000	275.154	7.9501 +02	1.0066 +00
4 000	262.166	6.1660 +02	8.1935 –01
6 000	249.187	4.7217 +02	6.6011 –01
8 000	236.215	3.5651 +02	5.2579 –01
10 000	223.252	2.6499 +02	4.1351 –01
Tropopause			
11 000	216.650	2.2632 +02	3.6480 –01
Stratosphere – Isothermal (216.65 K) from 11.0 to 20.0 km, then warming with altitude. Lapse rate –1.0 K km ⁻¹ from 20 to 32 km, then –2.8 K km ⁻¹ from 32.0 to 47.4 km.			
15 000	216.650	1.2111 +02	1.9476 –01
20 000	216.650	5.5293 +01	8.8910 –02
25 000	221.552	2.5492 +01	4.0084 –02
32 000	228.490	8.8906 +00	1.3555 –02
37 000	242.050	4.3324 +00	6.2355 –03
Stratopause			
47 400	270.650	1.1022 +00	1.4187 –03
Mesosphere – Isothermal (270.650 K) from 47.4 to 51.0 km, then cooling with altitude. Lapse rate 2.8 K km ⁻¹ from 51 to 71 km, then 2.0 K km ⁻¹ from 71 to 86 km.			
48 000	270.650	1.0229 +00	1.3167 –03
51 000	270.650	7.0458 –01	9.0690 –04
61 000	244.274	1.9157 –01	2.7321 –04
71 000	216.846	4.4795 –02	7.1966 –05
Mesopause			
86 000	186.87	3.7338 –03	6.958 –06
Thermosphere – Isothermal (186.87 K) from 86 to 92 km, then warming with altitude. Lapse rate decreases until it reaches –12 K km ⁻¹ at 120 km, then increases toward zero. Temperature asymptotically approaches 1000 K, and reaches that point at 815-km altitude.			
92 000	186.96	1.2887 –03	2.393 –06
100 000	195.08	3.2011 –04	5.604 –07
120 000	360.00	2.5382 –05	2.222 –08
150 000	634.39	4.5422 –06	2.076 –09
200 000	854.56	8.4736 –07	2.541 –10
300 000	976.01	8.7704 –08	1.916 –11
500 000	999.24	3.0236 –09	5.215 –13
700 000	999.97	3.1908 –10	3.070 –14
1 000 000	1000.00	7.5138 –11	3.561 –15

^a hPa = hectoPascal = 100 Pascals = 100 Newtons per square meter = 1 millibar.

3.3 Application of the standard atmosphere in aviation

One of the most important applications of the standard atmosphere is to keep aircraft vertically separated by a safe distance. Above what is called the *transition altitude* (18 000 ft in the United States, generally 3 000 to 14 000 ft elsewhere), aircraft under any kind of flight control fly on constant pressure surfaces. This is easy because their altimeters are really pressure sensors. An aircraft reporting a flight

level (z_{FL}) of 35 000 ft is flying on the pressure surface (p_{FL}) that corresponds to that altitude in the ICAO standard atmosphere. Note that z_{FL} is not the true altitude, but rather what the altitude would be in the standard atmosphere, given the pressure. At 10 000 m, the difference between z_{FL} and true altitude could be as much as 500 m. In practice, standard flight levels are separated by 1000 ft (305 m), except in China, where the separation is 300 m. Headings are based on ground-track magnetic directions, with eastbound defined as 0° through 179° (including N, NE, E, SE) and westbound

defined as 180° through 359° (including S, SW, W, NW). Eastbound and westbound aircraft occupy alternating flight levels for obvious reasons. The relation between p_{FL} and z_{FL} is given by

$$p_{FL} = p_0 \left(\frac{T_0 - \gamma z_{FL}}{T_0} \right)^{\frac{g}{\gamma R_d}}, \quad z_{FL} \leq 11\,000 \text{ m} \quad (3)$$

where p_{FL} and p_0 are expressed in hPa, $p_0 = 1013.25$ hPa (standard sea-level pressure), $T_0 = 288.15$ K (standard sea-level temperature), $\gamma = 0.0065 \text{ K m}^{-1}$ is the tropospheric lapse rate in the ICAO standard atmosphere g is the acceleration due to gravity, and R_d is the gas constant for dry air ($287 \text{ J kg}^{-1} \text{ K}^{-1}$). See <http://hurri.kean.edu/~yoh/calculations/standatm/StdAtm.html> for a derivation. Note that z_{FL} must be expressed in meters. For aircraft above a flight level of 11 km,

$$p_{FL} = p_1 \exp \left[-\frac{g}{R_d T_1} (z_{FL} - 11\,000) \right], \quad 11\,000 < z_{FL} \leq 20\,000 \text{ m} \quad (4)$$

where $p_1 = 226.32$ hPa is the pressure at the tropopause and $T_1 = 216.65$ K is the constant temperature of the lower stratosphere in the ICAO standard atmosphere (Stull, 2000, p. 13). Note that both equations (3) and (4) give the same answer when $z_{FL} = 11\,000$ m.

If an aircraft were to descend all the way to sea level, its altimeter would not read zero unless the sea-level pressure happened to be 1013.25 hPa. For the same reason, if an aircraft is landing at an airport whose runway elevation is z_{sta} , its altimeter will not read correctly when the wheels contact the runway unless the real atmosphere happens to match the ICAO standard atmosphere. For this reason, when flying below the transition altitude, pilots set their altimeter to the *altimeter setting* p_{as} , which is derived from the station pressure p_{sta} (true pressure as would be measured by a mercury barometer at the runway elevation) under the assumption of ICAO standard atmospheric conditions between the airport elevation and sea level. If in “level” flight, they will still be flying on surfaces of constant pressure below the transition altitude, but their altimeter will read the correct elevation when they land, even if the atmosphere both above and below the transition altitude is nonstandard. This is critical, especially when runway visibility is low. The altimeter setting is calculated from

$$p_{as} = \left[(p_{sta} - 0.3)^{\frac{R_d \gamma}{g}} + \frac{\gamma z_{sta}}{T_0} p_0^{\frac{R_d \gamma}{g}} \right]^{\frac{g}{R_d \gamma}} \quad (5)$$

This equation is closely related to equation (3) (see <http://hurri.kean.edu/~yoh/calculations/altimeter/>), and they

share the same constants. The pressure should be expressed in hPa because the constant 0.3 is in hPa. This constant accounts for the fact that the altimeter is mounted approximately 3 m above the runway (at a pressure about 0.3 hPa less than that on the runway) on many commercial aircraft. The altimeter setting is sometimes given in hPa and sometimes in inches of mercury (in Hg), depending on the country, and so a unit conversion may be appropriate after applying equation (5): 1 hPa = 0.02953 in Hg.

4 THE IONOSPHERE

4.1 Formation of the ionosphere

The ionosphere is a region of the upper atmosphere containing significant concentrations of ions (almost all carrying a positive charge) and free electrons. It extends from about 60-km altitude in the upper mesosphere to well beyond the 1000-km altitude considered in this chapter. The cloud of charged particles in the ionosphere is called a *plasma*, a mixture of nearly equal numbers of positively and negatively charged particles, in this case, within a medium of neutral gases, the thermosphere. Even where the number density of electrons (similar to the number density of positive ions) peaks, near 300-km altitude, the number density of neutral gases is still more than 100 times greater. Still, the plasma is highly conductive. The response of the plasma to electric and magnetic fields, and the interactions between charged and neutral particles strongly influence the properties of the ionosphere.

For this discussion, the following nomenclature is useful: x-rays correspond to the wavelength interval 0.01–1.0 nm; soft x-rays, 1.0 to 10 nm; extreme ultraviolet (EUV) 10 to 100 nm, and UV, 100–300 nm. The shorter the wavelength, the more energetic the radiation. Physicists denote the energy carried by photons as $h\nu$, where h is the Planck constant ($6.6261 \times 10^{-34} \text{ J s}$) and $\nu = c/\lambda$ is the frequency of the radiation. c is the speed of light ($2.998 \times 10^8 \text{ m s}^{-1}$), and λ is the wavelength in meters.

Radiation from the sun at UV and shorter wavelengths is energetic enough to strip electrons from neutral atmospheric gases, especially N_2 , O_2 , and O , thereby ionizing them. It also causes dissociation: splitting molecules into component parts without producing free electrons. At wavelengths shorter than about 35 nm, photons are so energetic that the electrons produced during ionization are “hot”; that is, they possess enough kinetic energy to cause ionization themselves during subsequent collisions with N_2 , O_2 , and O . The ionization and dissociation reactions listed in Table 2 occur throughout the ionosphere.

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Table 2. The most important ionization and dissociation reactions occurring in the ionosphere (Fuller-Rowell, 1993). The left column lists reactions caused by energetic photons. The right column lists reactions caused by hot electrons, many of which are produced by reactions in the left-hand column. The asterisk attached to the electron symbols on the right indicates that the electron must be energetic enough to cause ionization or dissociation during a collision. Following collisions, the resultant free electrons may or may not be hot.

Ionization by radiation	Ionization by collision
$O_2 + h\nu \rightarrow O_2^+ + e$	$O_2 + e^* \rightarrow O_2^+ + 2e$
$O_2 + h\nu \rightarrow O^+ + O + e$	$O_2 + e^* \rightarrow O^+ + O + 2e$
$O + h\nu \rightarrow O^+ + e$	$O + e^* \rightarrow O^+ + 2e$
$N_2 + h\nu \rightarrow N_2^+ + e$	$N_2 + e^* \rightarrow N_2^+ + 2e$
$N_2 + h\nu \rightarrow N^+ + N + e$	$N_2 + e^* \rightarrow N^+ + N + 2e$
Dissociation by radiation	Dissociation by collision
$O_2 + h\nu \rightarrow O + O$	$O_2 + e^* \rightarrow O + O + e$
$N_2 + h\nu \rightarrow N + N$	$N_2 + e^* \rightarrow N + N + e$

As short-wave radiation penetrates deeper into the atmosphere, the fraction absorbed increases, but so does the number density of atmospheric gases capable of absorbing it. The maximum electron/ion density occurs near 300 km. Below this level, photochemical equilibrium tends to prevail, whereby ion production is nearly balanced by recombination of the ionized species with electrons. The number density of particles and the high collision rate ensure this. Above 300 km, diffusive equilibrium prevails; as is the case with the neutral gases, the concentration of heavier ions decreases with altitude more rapidly than that of the lighter ions and much lighter electrons.

Before discussion of various regions of the ionosphere, it is instructive to examine Figure 3 (from Hobbs, 2000, p.

27), which summarizes much of the material presented here (and in Atmospheric Composition). At left is a global mean temperature profile, similar to that in Figure 1 except that it extends to 1500-km instead of 110-km altitude. Note the major difference in thermospheric temperatures depending on whether the sun is quiet or active. Large swings in temperature also occur between day and night. The middle part of the figure illustrates the gradual change in the mean molecular weight of dry air, constant at $28.9 \text{ kg kmol}^{-1}$ in the homosphere, where large-scale turbulence maintains the mixing ratios of atmospheric gases, but gradually decreasing in the heterosphere, where molecular diffusion predominates and the heavier gases undergo gravitational settling. In the exosphere, the number density of particles is so small that collisions are rare, and the lightest and fastest particles can escape the Earth's gravitational field. The square brackets refer to the number densities of the gases whose chemical symbols are inside. Thus, the number density for O becomes greater than that for O_2 at about 110-km altitude, and greater than that for N_2 near 180 km. At right, the ozone layer and various regions of the ionosphere are depicted along with representative electron densities. The vertical scale in kilometers is logarithmic.

4.2 Ionospheric regions

Several regions in the ionosphere are usually distinguished: the D-, E-, and F-regions. They are characterized by the photochemistry that occurs within them. The boundaries between them are not sharply defined.

The *D-region* extends from approximately 60- to 90-km altitude. X-rays are a source of ionization in the D-region;

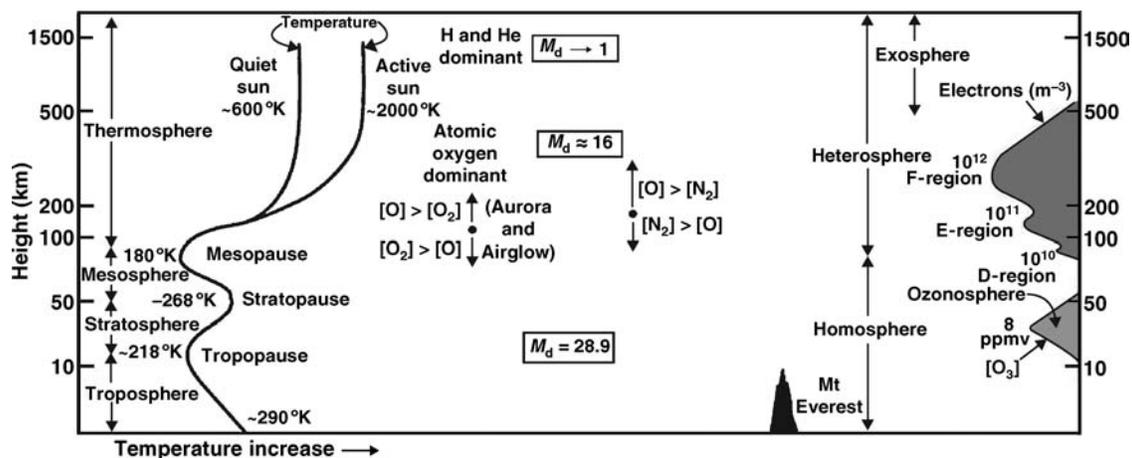
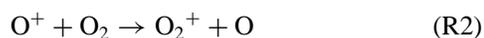


Figure 3. Left: representation of the vertical temperature structure in the atmosphere; center: variation of the mean molecular weight of dry air and a comparison of number densities of three atmospheric gases; right: the ionosphere and its relationship to the homosphere, heterosphere, and exosphere. Representative number densities (m^{-3}) of electrons are shown for three regions in the ionosphere. Reproduced from Hobbs (2000) © Cambridge University Press.

during a solar flare, the influx of x-rays can increase by two to three orders of magnitude, with a consequent large increase in D-region plasma density. At this altitude range, the shorter wavelengths of EUV radiation have been greatly depleted (Figure 2), but ionization of O₂, N₂, and nitric oxide NO occurs at wavelengths >112 nm. So-called Lyman-alpha radiation at 121.6 nm ionizes NO. Water cluster ions, both positive and negative, that incorporate various numbers of H₂O molecules form from the primary ionized species, NO⁺ and O₂⁺, and these dominate below about 85 km. The daytime ion number density approaches 10¹⁰ m⁻³, ten orders of magnitude less than the neutral gas number density. At night, ionization essentially ceases, the free electrons are quickly captured, and the D-region almost disappears. The most common manifestation of the D-region is the absorption of amplitude modulated (AM) radio waves during the day. After the sun sets, AM radio waves (frequencies 540–1700 kHz) can propagate to the higher-altitude F-region and bounce back to the surface, allowing long-range transmission of the signal, impossible by day.

The *E-region* extends from approximately 90- to 150-km altitude. Ion production rates are greatest for O₂⁺ and N₂⁺ in this altitude range, and less for O⁺, for which the maximum production rate occurs close to 180-km altitude. The ionizing radiation is principally Lyman-beta (102.5 nm), EUV radiation at less than 100 nm, and soft x-rays. NO⁺ is produced in two reactions facilitated by collisions, O⁺ + N₂ → NO⁺ + N and N₂⁺ + O → NO⁺ + N. After equilibrium is reached between production and loss (through recombination), NO⁺ and O₂⁺ become the dominant ions. Like the D-layer, the E-layer mostly disappears at night. Occasionally, a so-called *Sporadic E-layer*, a thin layer of high-density electrons will form in the same altitude range as the E-region. It is patchy and lasts from minutes to hours. It supports anomalous reflection of radio waves from 25 to 225 MHz.

The reactions that produce NO⁺, noted in the above paragraph, are two in a class of reactions called ion-molecule interchanges common in the E- and F-regions. Ions and neutral particles react through collisions to form new ions and neutral particles. The most important of these reactions are



The reactions most responsible for electron-ion recombination into neutral species (and the disappearance of the D- and E-regions at night) are



The *F-region* extends from approximately 150- to 400-km altitude. It makes possible short-wave radio communications over long distances. Atomic oxygen is the dominant neutral gas, followed by N₂ and O₂, but all three gases are ionized by EUV radiation, primarily at wavelengths <91.1 nm, which is at the short-wave end of the Lyman series.

During the daytime, the F-region divides into two regions, the F1 (150- to 250-km altitude) and the F2 (250- to 400-km altitude). Electron density increases steadily with altitude in the F1 region. Below about 180-km altitude, the ion-molecule interchange quickly consumes the O⁺ to make NO⁺ and O₂⁺ ions (reactions R1 and R2). The loss of these molecular ions occurs via recombination, and the rate of loss depends on the square of the electron (ion) density. The peak in the O⁺ production rate is near 180 km, and so during the day a bulge in the electron density profile appears near this altitude. Well above 180 km, the number densities of N₂ and O₂ are much less than that of O, so that the ion-molecule interchange reaction R3 occurs much more often than R1 and R2, leaving O⁺ as the dominant ion. The loss rate well above 180 km is proportional to the electron density, as opposed to the square of the density. At night, the F1 region virtually disappears because the electron loss rate is high and the ion production rate is almost nil.

The F2 region (250- to 400-km altitude) features a peak in electron density at about 300 km. If photochemical equilibrium prevailed above this altitude, the electron density would continue to increase, but plasma transport becomes important by means of either vertical diffusion or horizontal displacement by the neutral wind. The F2 region is in the heterosphere, where heavier particles seek hydrostatic equilibrium via diffusion. As a result, the number density of lighter particles decreases with altitude more slowly than that of heavier particles. This has an interesting effect on the ionospheric plasma, in that the heavier ions tend to settle downward while the much lighter electrons migrate upward. This generates an electrical field in the vertical, which tends to draw the electrons downward and the positive ions upward, an effect called *ambipolar diffusion*. The charges on these particles cause their number densities to decrease with altitude only half as rapidly as neutral species with the effect

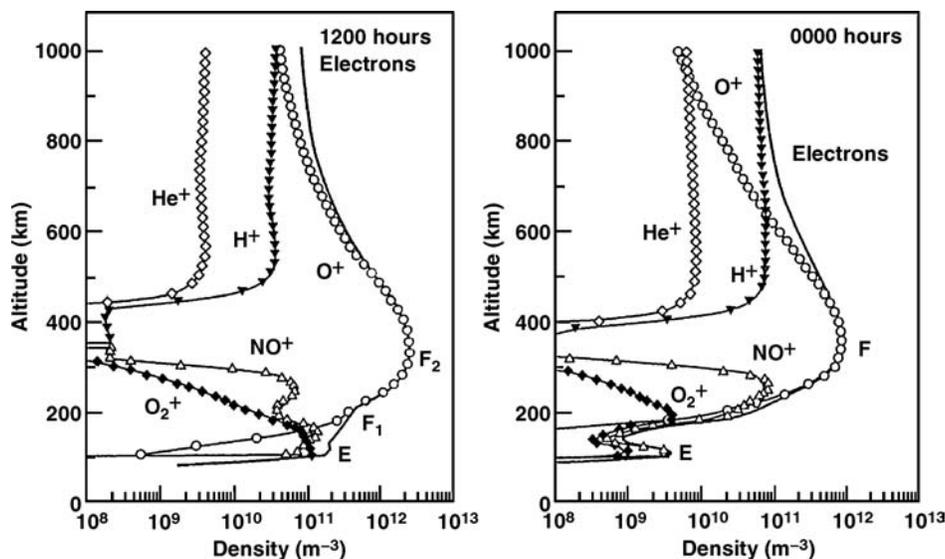


Figure 4. Vertical profile of ion and electron number density for solar noon (1200 hours, local time) and solar midnight (0000 hours). These conditions, typical of solar maximum, were extracted from the International Reference Ionosphere. The most important ions discussed in the text are labeled, as are the E, F1, and F2 regions. Reproduced from Ondoh and Marubashi (2000) © Ohmsha, Ltd.

that Maxwell-Boltzmann speed distributions remain valid for the plasma to much greater altitudes than for neutral particles.

The F2 region survives the night. With a collision frequency of less than one per second and a mean free path exceeding 1 km at 300 km, the electron density takes nearly 10 h to fall to 1/e of its sunset value, by which time the night is mostly over at low and midlatitudes. In addition, plasma diffuses downward at night into the F2 region from above.

The *topside ionosphere*, also called the *upper F-region*, is that region above the 300-km peak in electron density where O^+ is the dominant ion. It extends to above 1000-km altitude during the day but only to 600 km at night. Above that lies the *protonosphere*, where hydrogen and helium atomic ions (H^+ and He^+) dominate.

Figure 4 summarizes much of the preceding discussion. It displays a vertical profile of ion concentration, taken from the International Reference Ionosphere (IRI, see <http://modelweb.gsfc.nasa.gov/ionos/iri.html>). These examples, for solar noon and solar midnight, are typical during solar maximum. All ion (and electron) number densities are greater during daylight than at night, in the lower ionosphere by one or two orders of magnitude. The sum of ion number densities approximately equals the electron density at each altitude. Near 300 km, where electron density peaks, and for nearly 100 km above that, O^+ is the only ion of consequence. However, it is not the dominant ion in the E region nor in the D region just below.

The day-night difference in the thickness of the topside ionosphere is evident. Its upper boundary, where the number density of H^+ exceeds that of O^+ , is near 1000 km during the day and 600 km at night in this illustration.

It is important to note that ionospheric conditions are subject to very large and rapid changes because of solar heating, geomagnetic activity, and solar variability. A comprehensive reference on ionospheric physics is Schunk and Nagy (2000).

5 SUMMARY

A vertical profile of the long-term average temperature is a convenient way of characterizing different layers or shells of the Earth's atmosphere.

The troposphere is the lowest layer, extending from the surface to between 9 and 17 km, depending upon latitude and season. Tropospheric temperature normally decreases with altitude, but seldom more rapidly than about 10 K km^{-1} . Most of the time, the lapse rate is less than this; occasionally, however, the temperature increases with altitude – an inversion. Clouds and storms develop in the troposphere and can pose hazards to vehicles passing through them. The top of the troposphere, the tropopause, is marked by an inversion. The height of the tropopause is determined by radiative-convective equilibrium: in a dry atmosphere, radiative effects alone would lead to much steeper lapse rates than observed in the lower troposphere. Moist convection limits the lapse

rate. The vigor of convection is strongly correlated with the height of the tropopause.

The stratosphere extends from the tropopause to about 50-km altitude. Temperature increases slowly with altitude near the bottom but more rapidly near the top. The stratosphere is dry because most of the water vapor that reaches the stratosphere must come through the tropopause, which is very cold. The stratosphere is stably stratified, not subject to convection; consequently, any fine particulate matter that enters the stratosphere because of volcanic eruptions tends to stay there for many months because there is no precipitation to wash it out. The temperature variation in the stratosphere is mostly controlled by the absorption of ultraviolet radiation by ozone, whose concentration peaks near 35 km.

The mesosphere extends from 50 to between 80 and 90 km in altitude. The temperature decreases with altitude in the mesosphere, reaching a minimum of about 185 K at its upper boundary, the mesopause. Gravity waves and associated upward motion and adiabatic cooling help to maintain the very low temperatures observed in the upper mesosphere.

The thermosphere extends upward from the mesopause and gradually merges with outer space. Molecular absorption of energetic solar radiation heats the upper thermosphere to 1000 K. Molecular diffusion transfers heat downward to the lower thermosphere with the result that the temperature increases rapidly upward between 90 and 120 km, then more slowly, approaching 1000 K by 700-km altitude. Large day–night differences in temperature occur in the high thermosphere.

Standard atmospheres are prescriptions of atmospheric properties, primarily as a function of altitude, useful for engineering applications. One of the most prominent applications of the standard atmosphere is in aviation. Air traffic controllers separate aircraft in the vertical by having them fly on constant pressure surfaces. The “flight altitudes” are fictitious in that they are the altitude an aircraft would be, if flying in a standard atmosphere.

Overlapping the mesosphere and thermosphere is the ionosphere, where high-energy solar radiation dissociates atmospheric molecules and ionizes atoms and molecules. Positive ions and free electrons form a plasma in the upper atmosphere, which interacts with neutral particles and the Earth’s magnetic field. Three principal regions of the ionosphere are distinguished: the D-, E-, and F-regions. They affect radio wave propagation differently and respond differently to the day-night cycle of solar energy input. Ionospheric conditions are subject to very large and rapid changes because of solar heating, geomagnetic activity, and solar variability.

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