

## Chapter 1

### Basic Properties of Radiation, Atmospheres, and Oceans

#### 1.1 Introduction

This chapter presents a brief overview of the spectra of the shortwave solar and long-wave terrestrial radiation fields and the basic structure of atmospheres and oceans. Some general properties of the emission spectra of the Sun and the Earth are described. Their broad features are shown to be understandable from a few basic radiative transfer principles. We introduce the four basic types of matter that interact with radiation: gaseous matter, aqueous matter, particles, and surfaces. The stratified vertical structure of the bulk properties of an atmosphere or ocean is shown to be a consequence of hydrostatic balance. The vertical temperature structure of the Earth's atmosphere is shown to result mainly from radiative processes. Optical paths in stratified media are described for a general line-of-sight direction. Radiative equilibrium, the greenhouse effect, feedbacks, and radiative forcing are introduced as examples of concepts to be dealt with in greater detail in Chapter 12.

The ocean's vertical temperature structure, and its variations with season, are discussed as resulting from solar heating, radiative cooling, latent heat exchange, and vertical mixing of water masses of different temperature and salinity. Its optical properties are briefly described, along with ocean color. Section 1.7 prepares the reader for the notation and units used consistently throughout the book.

#### 1.2 Parts of the Spectrum

In Table 1.1, we summarize the nomenclature attached to the various parts of the visible and infrared spectrum. The spectral variable is the *wavelength*  $\lambda$ . Here  $\lambda = c/\nu$ , where

Table 1.1. *Subregions of the spectrum.*

Subregion	Range	Solar variability	Comments
X rays	$\lambda < 10 \text{ nm}$	10–100%	Photoionizes all thermosphere species.
Extreme UV	$10 < \lambda < 100 \text{ nm}$	50%	Photoionizes O <sub>2</sub> and N <sub>2</sub> . Photodissociates O <sub>2</sub> .
Far UV	$100 < \lambda < 200 \text{ nm}$	7–80%	Dissociates O <sub>2</sub> . Discrete electronic excitation of atomic resonance lines.
Middle UV, or UV-C	$200 < \lambda < 280 \text{ nm}$	1–2%	Dissociates O <sub>3</sub> in intense Hartley bands. Potentially lethal to biosphere.
UV-B	$280 < \lambda < 320 \text{ nm}$	<1%	Some radiation reaches surface, depending on O <sub>3</sub> optical depth. Responsible for skin erythema.
UV-A	$320 < \lambda < 400 \text{ nm}$	<1%	Reaches surface. Benign to humans. Scattered by clouds, aerosols, and molecules.
Visible, or PAR <sup>a</sup>	$400 < \lambda < 700 \text{ nm}$	≤0.1%	Absorbed by ocean, land. Scattered by clouds, aerosols, and molecules. Primary energy source for biosphere and climate system.
Near IR	$0.7 < \lambda < 3.5 \text{ } \mu\text{m}$		Absorbed by O <sub>2</sub> , H <sub>2</sub> O, CO <sub>2</sub> in discrete vibrational bands.
Thermal IR	$3.5 < \lambda < 100 \text{ } \mu\text{m}$		Emitted and absorbed by surfaces and IR-active gases.

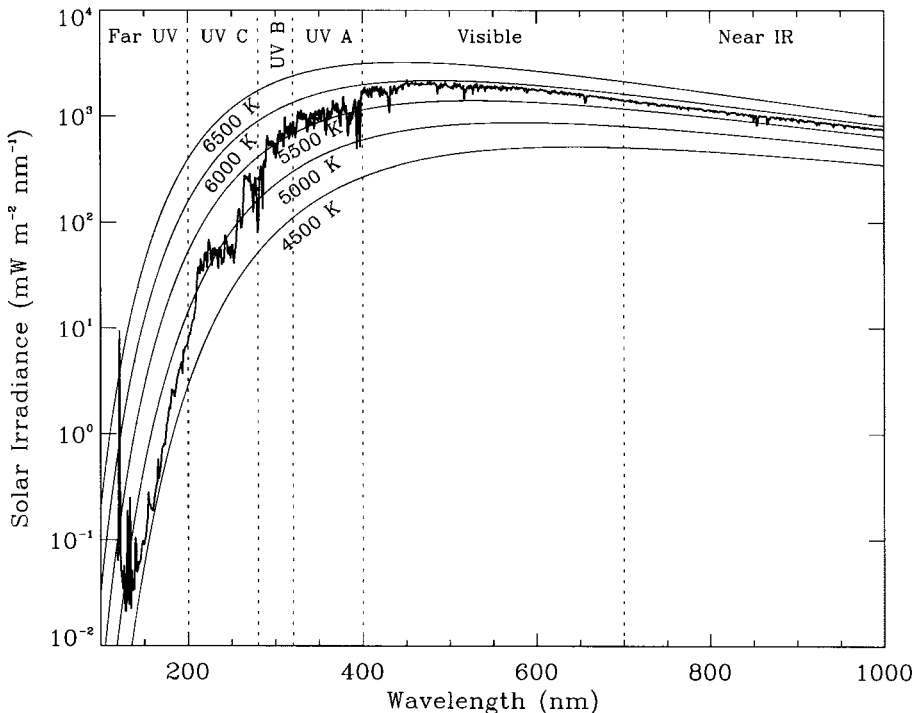
*Note:* <sup>a</sup>PAR stands for “photosynthetically active radiation.”

$c$  is the speed of light and  $\nu$  is the *frequency* ( $[\text{s}^{-1}]$  or  $[\text{Hz}]$ ). In the IR  $\lambda$  is usually expressed in *micrometers* (or more commonly *microns*, where  $1 \text{ } \mu\text{m} = 10^{-6} \text{ m}$ ). In the UV and visible spectral range,  $\lambda$  is expressed in *nanometers* ( $1 \text{ nm} = 10^{-9} \text{ m}$ ). A wavelength unit widely used in astrophysics and laboratory spectroscopy is the *Ångström* ( $1 \text{ Å} = 10^{-10} \text{ m}$ ). For completeness we list both X rays and the shorter-wavelength UV regions, even though they are not discussed in this book. The third column lists the known solar variability (in percent), defined as the maximum minus minimum divided by the minimum. We also provide brief comments on how radiation in each spectral subregion interacts with the Earth’s atmosphere. A common usage is to denote the solar part of the spectrum as *shortwave* radiation and the thermal infrared (IR) as *longwave* radiation. The latter is sometimes referred to as *terrestrial* radiation.

### 1.2.1 Extraterrestrial Solar Flux

In this section we consider some elementary aspects of solar radiation and the origin of its deviations from blackbody behavior. We will assume the reader is familiar with the concept of *absorption opacity*, or *optical depth*,  $\tau(\nu)$  at frequency  $\nu$ . The basic ideas are reviewed in Appendix G and covered more thoroughly in Chapter 2.

In Fig. 1.1 we show the measured *spectral flux*, or *irradiance*, of the Sun's radiative energy at a distance of one astronomical unit  $r_{\oplus}$  ( $r_{\oplus} = 1.5 \times 10^6$  km).<sup>1</sup> Integrated over all frequencies, this quantity is called the *solar constant*,  $S$  [ $\text{W} \cdot \text{m}^{-2}$ ]. These data were taken by a spectrometer on board an Earth-orbiting satellite, beyond the influences of the atmosphere.<sup>2</sup> The solar constant is not actually a constant but is slightly variable. For this reason, the modern term is the *total solar irradiance*, whose value<sup>3</sup> is about  $1368 \text{ W} \cdot \text{m}^{-2}$ . The total solar irradiance  $S$  represents the total instantaneous radiant energy falling normally on a unit surface located at the distance  $r_{\oplus}$  from the Sun. It is the basic forcing of the Earth's heat engine, and indeed for all planetary bodies that derive their energy primarily from the Sun. The quantity  $S(r_{\oplus}^2/r^2)$  is the total instantaneous radiant energy falling normally on a unit surface at the solar distance  $r$ .



**Figure 1.1** Extraterrestrial solar flux, or irradiance, measured by a spectrometer on board an Earth-orbiting satellite. The UV spectrum ( $119 < \lambda < 420$  nm) was measured by the SOLSTICE instrument on the *UARS* satellite (modified from a diagram provided by G. J. Rottmann, private communication, 1995). The vertical lines divide the various spectral subranges defined in Table 1.1. The smooth curves are calculated blackbody spectra for a number of emission temperatures.

Also shown in Fig. 1.1 are spectra of an ideal blackbody at several temperatures. Requiring that the total energy emitted be the same as a blackbody, one finds that the Sun's effective temperature is 5,778 K. If the radiating layers of the Sun had a uniform temperature at all depths, its spectrum would indeed match one of the theoretical blackbody curves exactly. The interesting deviations seen in the solar spectrum can be said to be a result of emission from a *nonisothermal atmosphere*. Radiative transfer lies at the heart of the explanation for this behavior.

We can explain the visible solar spectrum qualitatively by considering two characteristics of atmospheres – 1. their absorption opacity  $\tau(\nu)$  depends upon frequency and 2. their temperature varies with atmospheric depth – and one basic rule – that a radiating body emits its energy to space most efficiently at wavelengths where the opacity is approximately unity. This rule is explained in terms of the competing effects of absorption and emission. In spectral regions where the atmosphere is transparent ( $\tau(\nu) \ll 1$ ), it neither emits nor absorbs efficiently. In contrast, where it is opaque ( $\tau(\nu) \gg 1$ ), its radiative energy is prevented from exiting the medium, that is, it is reabsorbed by surrounding regions. At  $\tau(\nu) \approx 1$ , a balance is struck between these opposing influences.

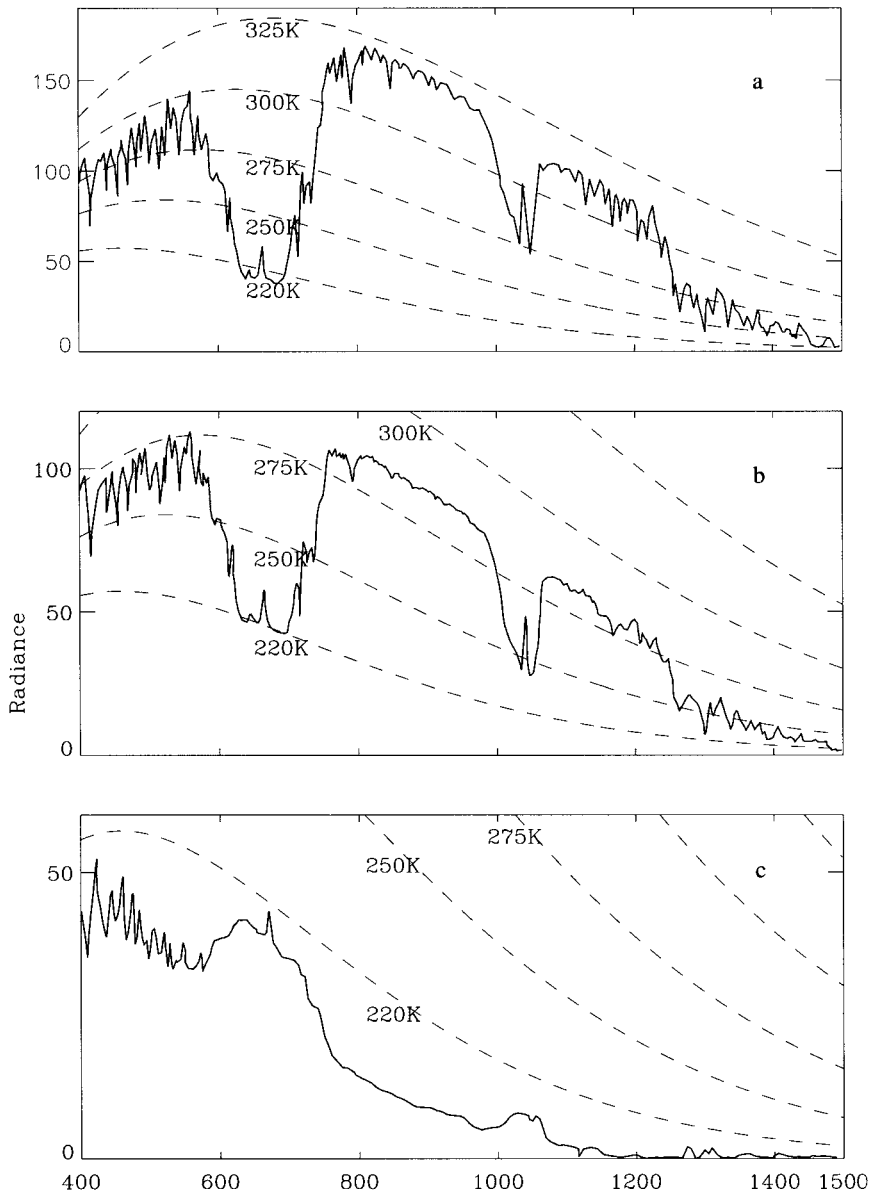
At visible wavelengths, the Sun's opacity is unity deep within the solar atmosphere in the relatively cool *photosphere*, where the temperature is  $\cong 5,780$  K. Regions as cool as 4,500 K are apparent at 140–180 nm (see Fig. 1.1). At shorter wavelengths the opacity increases, thereby raising the effective emission height into the higher-temperature *chromosphere*. The solar spectrum can be thought of as a “map” of the vertical temperature structure of the Sun. The map can be read provided one has knowledge of the dependence of opacity of the solar atmosphere on wavelength.

### 1.2.2 Terrestrial Infrared Flux

An understanding of radiative transfer is also essential for understanding the energy output of the Earth, defined to be the spectral region  $\lambda > 3.5 \mu\text{m}$ . Figure 1.2 shows the IR emission spectrum measured from a down-looking orbiting spacecraft, taken at three different geographic locations.<sup>4</sup> Also shown are blackbody curves for typical terrestrial temperatures. The spectral variable in this case is *wavenumber*  $\tilde{\nu} = 1/\lambda$ , commonly expressed in units of  $[\text{cm}^{-1}]$ . Again, as for the solar spectrum, the deviations are attributed to the nonisothermal character of the Earth's atmosphere. The spectral regions of minimum emission arise from the upper cold regions of the Earth's troposphere where the opacity of the overlying regions is  $\sim 1$ . Those of highest emission originate from the warm surface in transparent spectral regions (“windows”), with the exception of the Antarctic spectrum, where the surface is actually colder than the overlying atmosphere (see Fig. 1.2). In this somewhat anomalous situation, the lower-opacity region is one of higher radiative emission because of the greater rate of emission of the warm air. Again, the deviations from blackbody behavior can be understood qualitatively in terms of the temperature structure of the Earth's atmosphere and the variation with frequency of the IR absorption opacity.

## 1.2 Parts of the Spectrum

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**Figure 1.2** Thermal emission spectra of Earth measured by the IRIS Michelson interferometer instrument on the *Nimbus 4* spacecraft (see Endnote 4). Shown also are the radiances of blackbodies at several temperatures. (a) Sahara region; (b) Mediterranean; (c) Antarctic.

## 1.3 Radiative Interaction with Planetary Media

### 1.3.1 Feedback Processes

The properties of planetary media (chemical and dynamical) may themselves be affected by radiation, on all spatial scales. These changes may then further influence the way the media interact with radiation, etc. On the macrophysical (much greater than molecular) scale, we will mention two examples: 1. During daytime, solar radiation heats the Earth's surface and atmosphere. Often there results a fluid instability, which causes air to be set into convective motion, some air parcels moving upward, others downward. Upward air motion causes adiabatic cooling and, if the atmosphere is sufficiently moist, will lead to condensation and cloud formation. Clouds will alter the distribution of incoming sunlight and absorb and emit IR radiation, and thus affect the heating, etc. 2. A second example is that of ocean photosynthesis. The concentrations of light-absorbing phytoplankton determine the depth dependence of the radiation field, which itself governs the viability of such organisms.

If we had to concern ourselves with these “chicken-and-egg” problems of simultaneous mutual interactions of the medium and the radiation, this book would be very different and the subject much more difficult. Throughout, we will deal with the optical properties of planetary media as if they were specified *a priori*. This is a useful fiction for our purposes. Our neglect of these so-called feedback processes necessarily rules out a treatment of many interesting climate phenomena (cloud–radiation interactions, ocean photo-biologic feedback on climate change, etc.). Fortunately these feedback processes do not alter the fundamental description of the light–matter interaction. The speed of light is so high that the radiation field adjusts itself instantaneously to its environment. As a result radiative transfer is essentially a quasi-static phenomenon, and consequently its interaction with matter can usually be treated separately from other physical influences.

On the microphysical (molecular) scale, the presence of radiation can alter the basic optical properties of matter itself. Radiative heating leads to a redistribution of quantized states of excitation (for example the vibratory motion of molecules), which in turn alters the light interaction properties of the gas. In other words, the absorptive and emissive properties of a gas depend upon its temperature, which is itself affected by radiative heating. Again a fortunate circumstance usually allows us to decouple these two situations, so that the gas temperature may be considered to be an externally specified quantity, independent of the radiation field. This is contingent on the gas density being sufficiently high, so that *Kirchhoff's Law* is obeyed (§5.2.1). This condition is easily met for the lower portions of most planetary atmospheres and for the ocean.

### 1.3.2 Types of Matter that Affect Radiation

Pretending that they are independent of the radiation, we now focus on those aspects of oceans and atmospheres that are important in modifying the radiation field. For our

purposes, there are four forms of matter that can affect radiation:

*Gaseous matter:* Under local thermodynamic equilibrium conditions (§5.2.1), the density  $\rho$ , temperature  $T$ , and chemical composition are normally all that is required to determine the optical properties. Gas pressure  $p$  should also be included in this list, although it is not independent of  $\rho$  and  $T$ . Gas pressure, through its collisional effects on the quantized excited states of the molecules, affects absorption of light by altering the line strengths, as well as the line positions in frequency and their spectral width (§3.3.3). The quantities  $\rho$ ,  $T$ , and  $p$  are related to one another by an empirical “real-gas” formula, although it is almost always an adequate approximation to use the ideal gas law (see the following section).

*Aqueous matter:* As in gaseous media, density largely determines the optical properties of pure ocean water. Salinity, which is important for ocean dynamics, is unimportant for the optical properties. However, “pure sea water” hardly exists outside the laboratory. “Impurities” usually dominate the optical properties of natural bodies of water.

*Particles:* The atmospheric particulate population consists of suspended particles (aerosols) and condensed water (*hygrosols*). The latter is the generic term for water droplets and ice crystals, or combinations with dust. Airborne particles may be of biological origin or originate from pulverization of solid surfaces. Particles are frequently chemically or physically altered by the ambient medium, and these alterations can affect their optical properties. Particles with sizes comparable to the wavelength take on optical characteristics that can be quite different from their parent-solid bulk optical properties (§4.2). Oceanic particles consist of a large variety of dissolved and suspended organic and inorganic substances, such as the variously pigmented phytoplankton and organic yellow substances.<sup>5</sup>

*Solid and ocean surfaces:* The atmospheres of the terrestrial planets are all in contact with surfaces, which vary greatly in their visible-light reflectance and absorptance properties (§5.2). In many applications their strong continuous absorption in the IR allows them to be treated as thermally emitting blackbodies, an enormous theoretical simplification. Knowledge of the visible reflectance of underlying land and ocean surfaces is necessary for calculating the diffuse radiation field emergent from the atmosphere. In addition, the reflectance of the ocean bottom in shallow seas has an important influence on the diffuse radiation field in the ocean and on the radiation leaving the ocean surface.

## 1.4 Vertical Structure of Planetary Atmospheres

It is useful to describe those general aspects of similarity and dissimilarity of oceans and atmospheres. First, they are similar in that they are both *fluids*, that is, they readily flow under the influences of gravity and pressure differences. Also, they both obey the basic

equation of *hydrostatic equilibrium*. A fundamental difference is that atmospheres are highly *compressible*, whereas oceans are nearly *incompressible*. A quantitative difference arises from the fact that the average density of water ( $1 \times 10^3$  [kg · m<sup>-3</sup>]) is much higher than that of most planetary atmospheres. For the Earth's atmosphere on a clear day at sea level, a visible light photon can traverse unattenuated a horizontal path many hundreds of kilometers long. In the ocean, it penetrates at most a few hundred meters before being attenuated. Of course for sufficient depths in the atmospheres of Venus and of the giant outer planets, the atmospheric density can approach or even exceed that of water.

### 1.4.1 Hydrostatic and Ideal Gas Laws

In this section, we describe some important *bulk properties* of the atmosphere and ocean, described by its density, pressure, temperature, and index of refraction. As a result of gas being highly compressible, the *atmospheric density*,  $\rho$  [kg · m<sup>-3</sup>], the mass per unit volume, varies strongly with height,  $z$ . For both atmospheres and oceans in a state of rest, the pressure,  $p$ , must support the weight of the fluid above it. This is called a state of *hydrostatic equilibrium*. With increasing height in the atmosphere, the density decreases as the pressure decreases (*Boyle's Law*). With increasing depth in the ocean, this also holds true but the density change is slight.

Consider the atmospheric case. In differential form, the weight of the air (mass times the acceleration of gravity  $g$ ) in a small volume element  $dV$  is  $gdM$ , where  $dM$  is the mass of the air inside the volume. Now  $dM = \rho dV = \rho dAdz$ , and the net force exerted by the surrounding gas on the parcel is  $-dpdA$ . The differential  $dp$  is the change in pressure over the small height change  $dz$ . The minus sign comes from the fact that the pressure at  $z + dz$  is smaller than at  $z$ , and the upward buoyancy force must be positive. Equating the two forces,  $-dpdA = g\rho dAdz$ , we find upon cancellation of the  $dA$  term

$$dp = -g\rho dz. \quad (1.1)$$

For planetary atmospheres of moderate and low density, and specifically the Earth's atmosphere, the equation of state is closely approximated by the *ideal gas law*,

$$\rho = \frac{\bar{M}p}{RT} = \bar{M}n, \quad (1.2)$$

where  $\bar{M}$  is the *mean molecular mass*,  $R$  is the *gas constant per mole*, and  $n$  is the total *concentration* of molecules (number of molecules per unit volume). More detailed descriptions of each of the above quantities are given below. Substituting Eq. 1.2 into Eq. 1.1, we find

$$\frac{dp}{p} = -\frac{dz}{H}, \quad \text{where } H \equiv \frac{RT}{\bar{M}g}. \quad (1.3)$$

$H$  is called the *atmospheric scale height*.



Earth's atmosphere consists of a mixture of long-lived, permanent species, together with many other minor species molecules. The  $i$ th species contributes to the overall bulk density through its molecular mass  $M(i)$  and its concentration  $n_i$ . Air contains a "standard" mixture of 78% nitrogen ( $M(\text{N}_2) = 28$ ), 21% oxygen ( $M(\text{O}_2) = 32$ ), and about 1% argon ( $M(\text{Ar}) = 40$ ). The proportions of these constituents are essentially constant up to about 95 km. Above this height, the *homopause*,  $\bar{M}$ , decreases as the composition changes to lower-mass species, such as O, N, and He. This results from photodissociation of the heavier molecules, combined with gravitational separation tending to place the lighter elements at the highest levels.

Integrating Eq. 1.3 from  $z' = z_0$  to  $z' = z$ , and using the ideal gas law, Eq. 1.2, we obtain

*Three forms of the hydrostatic equation*

$$p(z) = p(z_0) \exp \left[ - \int_{z_0}^z dz' / H(z') \right], \quad (1.4a)$$

$$n(z) = n(z_0) \frac{T(z_0)}{T(z)} \exp \left[ - \int_{z_0}^z dz' / H(z') \right], \quad (1.4b)$$

$$\rho(z) = \rho(z_0) \frac{T(z_0)}{T(z)} \exp \left[ - \int_{z_0}^z dz' / H(z') \right]. \quad (1.4c)$$

It is clear that from a knowledge of surface pressure  $p(z_0)$  and the variation of scale height  $H(z)$  from  $z_0$  to  $z$ , Eqs. 1.4 allow us to determine the bulk gas properties at any height  $z$ .

The various quantities and their units [ ] are defined below:

$z, z_0$ : Height [m] above a reference level  $z_0$ .  $z_0$  is usually the surface,  $z_0 = 0$ . The more common atmospheric unit is [km].

$\rho$ : Gas density or mass per unit volume.  $\rho = 1.293$  [ $\text{kg} \cdot \text{m}^{-3}$ ] at  $0^\circ\text{C}$  and 1 bar, which is the condition of standard temperature and pressure (STP).

$p$ : Gas pressure [ $\text{N} \cdot \text{m}^{-2}$ ] at height  $z$ .  $p$  is the sum of the *partial pressures*  $\sum p_i$ . Pressure is expressed in bars, 1 [bar] = 101,325 [ $\text{N} \cdot \text{m}^{-2}$ ], or Pascals [Pa]. Another common unit is the *millibar* ([mbar]) or *hectoPascal*. 1 [mbar] = 1 [hPa] =  $10^{-3}$  [bar] =  $10^2$  [Pa].

$T$ : Atmospheric temperature in Kelvins [K]. In everyday use, degrees Celsius [C] is more frequent.  $T(\text{C}) = 273.2 + T(\text{K})$ .

$H$ : Atmospheric scale height [m], more commonly [km], equal to  $RT/\bar{M}g \equiv R_a T/g$ . Up to the homopause,  $H \approx 29.3 T$  ( $T$  in [K]) [m].

$R, R_a$ : Molar gas constant,  $8.3143 \times 10^3$  [ $\text{J} \cdot \text{K}^{-1} \cdot \text{kmol}^{-1}$ ], and specific gas constant =  $2.87 \times 10^2$  [ $\text{J} \cdot \text{K}^{-1} \cdot \text{kg}^{-1}$ ], respectively. The latter applies to air below the homopause. One kilogram-mole [kmol] is the quantity of matter in

a volume having a mass equal to its molecular weight. One kmol contains the number  $N_a$  (Avogadro's number) =  $6.022 \times 10^{26}$  molecules.

$n$ : Total gas concentration [ $\text{m}^{-3}$ ], equal to the sum of the individual gas concentrations  $\sum n_i$ .

$\bar{M}$ : Mean molecular mass per kilogram-mole [ $\text{kg} \cdot \text{kmol}^{-1}$ ]. For dry air,  $\bar{M} = 28.964 \text{ kg} \cdot \text{kmol}^{-1}$  up to the homopause.  $\bar{M} = \sum n_i M_i / n$ , where  $M_i$  are the molecular masses of the individual species.

$g$ : Acceleration due to gravity [ $\text{m} \cdot \text{s}^{-2}$ ].  $g(z) = g(z_0)(z_0 + R_\oplus)^2 / (z + R_\oplus)^2$ , where  $g(z_0) = 9.807 \text{ [m} \cdot \text{s}^{-2}]$  and  $R_\oplus$  is the Earth's mean radius, equal to 6,371 [km]. Strictly speaking,  $g$  also should include the small ( $\leq 1\%$ ) outward centripetal force due to the Earth's rotation, which varies with latitude.

The following approximation is often made to simplify the form of the hydrostatic equation. Assuming that  $g$ ,  $T$ , and  $\bar{M}$  are constant with height, we can integrate the argument of the exponential in Eqs. 1.4 to obtain

$$\frac{p(z)}{p(z_0)} \approx e^{-(z-z_0)/H}, \quad (1.5a)$$

$$\frac{n(z)}{n(z_0)} \approx e^{-(z-z_0)/H}, \quad (1.5b)$$

$$\frac{\rho(z)}{\rho(z_0)} \approx e^{-(z-z_0)/H}. \quad (1.5c)$$

The above equations show that  $H$  is an  $e$ -fold height for density.

Another important property of ideal gases is given by *Dalton's Law*, which states that the total pressure equals the sum of the *partial pressures*  $p_i$ , where  $i$  denotes the  $i$ th gas species. As long as the gases are well mixed, each separate species concentration  $n_i$  obeys the same hydrostatic equation. However, this does not apply to short-lived species (time scales short compared to a mixing time scale, typically  $\sim$ hours to a few days in the troposphere). Examples are ozone ( $\text{O}_3$ ), which is chemically destroyed or created, and water ( $\text{H}_2\text{O}$ ), which undergoes phase changes on short time scales. The hydrostatic equation does not apply to these species, and empirical or theoretically modeled determinations of species concentrations are then needed. This is one of the principal tasks of the subject of *aeronomy*.

The hydrostatic equation shows that atmospheric bulk density and pressure change rapidly with height with an  $e$ -folding height of 6 to 8 km. An important property of an atmosphere or ocean is its tendency to arrange itself into a vertically stratified and horizontally homogeneous medium. Quasi-horizontal motions tend to homogenize properties along constant-pressure (more correctly constant-entropy) surfaces. Horizontal variations do occur (after all, this is the origin of weather), particularly in temperature, but usually on spatial scales much greater than a scale height. However, all over the Earth, the pressure and density at sea level are nearly the same on a horizontal plane. Even during severe weather disturbances, the horizontal surface pressure