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 longwave 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 which interact with radiation: gaseous, aqueous, particles, and surfaces. The stratified vertical structure of the bulk properties of an atmosphere or ocean are 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 8.

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. The last section prepares the reader for the notation and units used consistently throughout the book. Finally in the last section, we describe the conventions used for the various symbols which may depart from standard usage.

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* λ. Here $\lambda = c/\nu$ where c is the speed of light and ν is the *frequency* [s^{-1}] or [Hz]. In the infrared (IR), λ is usually expressed in *micrometers* (where 1 μm $= 10^{-6}$ m).

Table 1.1 *Subregions of the spectrum.*

Subregion	Range	Solar variability	Comments
X-rays	$\lambda < 10$ nm	10–100%	Photoionizes all thermosphere species.
Extreme UV	$10 < \lambda < 100$ nm	50%	Photoionizes O_2 and N_2. Photodissociates O_2.
Far UV	$100 < \lambda < 200$ nm	7–80%	Dissociates O_2. Discrete electronic excitation of atomic resonance lines.
Middle UV, or UV-C	$200 < \lambda < 280$ nm	1–2%	Dissociates O_3 in intense Hartley bands. Potentially lethal to biosphere.
UV-B	$280 < \lambda < 320$ nm	$< 1\%$	Some radiation reaches surface, depending on O_3 optical depth. Damaging to biosphere. Responsible for skin erythema.
UV-A	$320 < \lambda < 400$ nm	$< 1\%$	Reaches surface. Benign to humans. Scattered by clouds, aerosols, and molecules.
Visible, or PAR	$400 < \lambda < 700$ nm	$\leq 0.1\%$	Absorbed by ocean, land. Scattered by clouds, aerosols, and molecules. Primary energy source for biosphere and climate system.
Near IR	$0.7 < \lambda < 3.5\ \mu$m	–	Absorbed by O_2, H_2O, CO_2 in discrete vibrational bands.
Thermal IR	$3.5 < \lambda < 100\ \mu$m		Emitted and absorbed by surfaces and IR significant gases.

Note: PAR stands for photosynthetically active radiation.

In the ultraviolet (UV) and visible spectral ranges, λ is expressed in *nanometers* ($1\ \text{nm} = 10^{-9}$ m). A wavelength unit widely used in astrophysics and laboratory spectroscopy is the *Ångström* ($1\ \text{Å} = 10^{-10}$ m). For completeness we list both X-rays and the shorter-wavelength UV regions, even though they are not discussed in this book. A column lists the known solar variability, defined as the maximum minus minimum divided by the minimum in percentages. 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 IR as *longwave* radiation. The latter is sometimes referred to as *terrestrial* radiation.

1.2.1 Extraterrestrial Solar Irradiance

In this section, we consider some elementary aspects of solar radiation and the origin of its deviations from blackbody behavior. We will assume that the reader

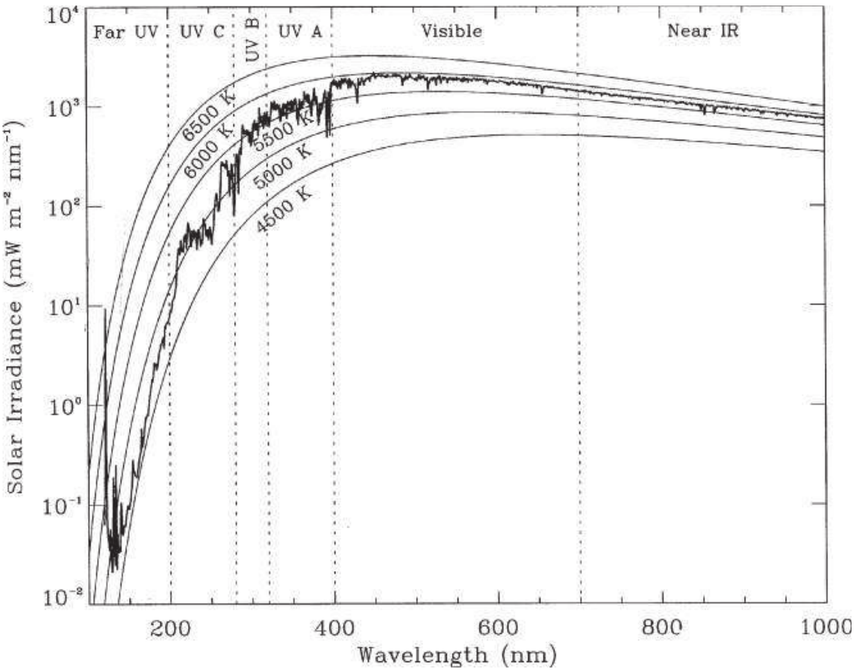

Figure 1.1 Extraterrestrial solar irradiance measured by a spectrometer on board an Earth-orbiting satellite. The UV spectrum ($119 < \lambda < 420$ nm) was measured by the SOLSTICE instrument on the UARS satellite (modified from a diagram provided by G. J. Rottmann, private communication, 1995). The 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.

is familiar with the concept of *absorption opacity*, or *optical depth*, $\tau(\nu)$ at frequency ν. The basic ideas are reviewed in Appendix E and covered more thoroughly in Chapter 2.

In Fig. 1.1, we show the measured *spectral irradiance* of the Sun's radiative energy at the mean distance between the Sun and the Earth, known as one astronomical unit $r_\oplus$ ($r_\oplus = 1.4960 \times 10^{11}$ m).[1] Integrated over all frequencies, this quantity is called the *solar constant* [$\mathrm{W \cdot m^{-2}}$]. These data were taken by a spectrometer onboard an Earth-orbiting satellite, beyond the influences of the atmosphere.[2] The solar constant is not actually a constant, but slightly variable. For this reason, the modern term is the *total solar irradiance*, S_0, whose most current published value[3] is $(1360.8 \pm 0.5)\ \mathrm{W \cdot m^{-2}}$. S_0 represents the total instantaneous radiant energy

[1] The visible spectrum is taken from a variety of sources. See Albritton et al. (1985), Nicolet (1989), and rredc.nrel.gov/solar/spectra/am1.5/ASTMG173/ASTMG173.html (see §10.3.3).
[2] See Rottman et al. (1993), lasp.colorado.edu/home/sorce/data/tsi-data/, and Kopp and Lean (2011).
[3] Kopp and Lean (2011). This value supersedes the value 1365.2 $\mathrm{W \cdot m^{-2}}$ used by Trenberth et al. (2009).

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 $S_0(r_\oplus^2/r^2)$ is the total instantaneous radiant energy falling normally on a unit surface at the solar distance r.

Also shown in Fig. 1.1 are spectra of an ideal blackbody at several temperatures. As the total energy emitted must be the same as that of a blackbody, one finds that the Sun's effective temperature is 5778 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 and one basic rule: (1) their absorption opacity $\tau(\nu)$ depends upon frequency and (2) their temperature varies with atmospheric depth. The basic rule is 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. On the other hand, 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 $\cong$ 5780 K. Regions as cool as 4500 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.

The bulk of the Earth's atmosphere (99% by mass) consists of molecular nitrogen and oxygen, in the form of radiatively inactive homonuclear, diatomic molecules. Trace amounts of polyatomic molecules are responsible for atmospheric absorption and emission of radiation in several hundred thousands of individual spectral lines arising from rotational and vibrational transitions. Water vapor, carbon dioxide, and ozone are the main absorbers (and emitters) contributing to warming and cooling of the atmosphere and underlying surface. These gases warm our planet by absorbing radiation emitted by the surface – without them, the Earth would be some 33°C colder than at present and in a state of permanent glaciation. Hence, the so-called greenhouse effect is very important for life itself. This effect also explains the

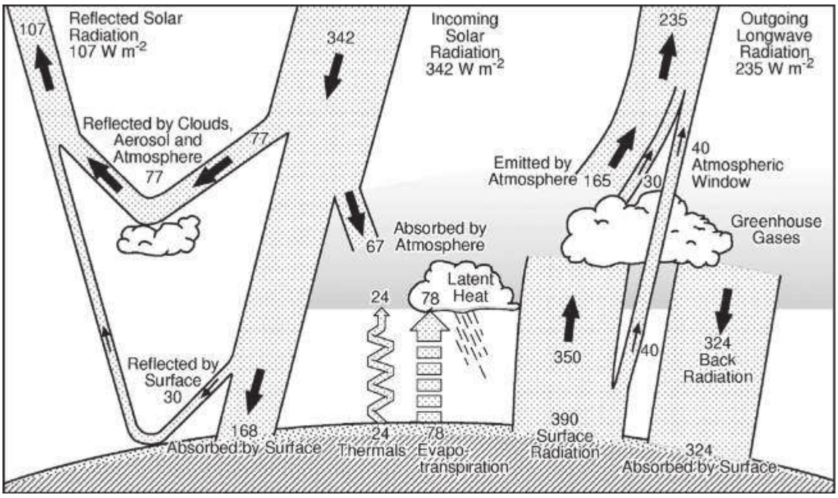

Figure 1.2 Earth's energy budget based on a S_0 value of 1368 W $\cdot$ m^{-2} (adapted from Kiehl and Trenberth, 1997). American Meteorological Society. Used with permission). An updated version of this illustration with revised budget numbers based on a S_0 value of 1365.2 W $\cdot$ m^{-2} is available (Trenberth et al., 2009).

high surface temperature of Venus and may have played a key role in maintaining temperatures high enough in an early primitive atmosphere of Mars to sustain running water and possibly even primitive life. Other trace gases make smaller contributions to warming/cooling of the Earth's atmosphere and surface. Some have natural origins, while others are partially (such as methane) or wholly (such as the chlorofluorocarbons) anthropogenic.

Figure 1.2 is a schematic diagram of the significant components of the Earth's energy balance. Of the incoming solar irradiance (342 W $\cdot$ m^{-2} averaged over the entire planet), 31% is reflected to space.[4] The absorbed solar energy (235 W $\cdot$ m^{-2}) is balanced by an equal amount radiated to space in the IR. Within the atmosphere, the land surface, and the ocean's mixed layer, the transformation of radiative energy into chemical, thermal, and kinetic energy drives the "engine" of weather and climate. Perturbations of this complex system can arise internally. Examples of internal forcing would be a change in atmospheric chemical composition or distribution of land masses. External forcing of the climate can arise from a change in the Sun's output, and by changes in the Earth's orbit.

The well-documented increase in CO_2 abundance, above what is believed to be the natural level existing in the preindustrial era, has been a matter of considerable

[4] Albedo values derived from satellite data are uncertain and range from about 28% to 34% depending on data source and estimation method (Trenberth et al., 2009).

concern. The reason for this concern is simply that the enhanced levels of CO_2 (and other already existing greenhouse gases or the release of new ones) absorb and trap terrestrial radiation that would otherwise escape to space. This situation causes an imbalance between the energy received and emitted by the planet. If the planet receives more energy from the Sun than it is able to emit to space, then by increasing its temperature it will increase the energy emitted (by the *Stefan–Boltzmann Law*) until a new radiative equilibrium between the Sun and the Earth is established. Hence, this additional trapping of terrestrial radiation by the enhanced levels of greenhouse gases is expected to lead to a warming so as to make the net energy emitted by the planet equal to that received.

The amount of warming depends crucially on how the entire Earth climate system, including the atmosphere, the land, the ocean, the cryosphere (snow and ice), and living things (the biosphere), responds to this warming. For example, could the Earth partly compensate for this extra heat source by increasing its albedo? Increase in low cloudiness in response to warming (which is expected to enhance evaporation) may lead to increased reflection of solar energy and thus offset the warming. But more high clouds (*cirrus*) could on the other hand lead to additional trapping of terrestrial radiation and therefore an amplification of the warming.

1.2.2 Terrestrial Infrared Irradiance

An understanding of radiative transfer is also essential for understanding the energy output of the Earth, defined to be the energy emitted in the spectral region where $\lambda > 3.5\ \mu$m. Figure 1.3 shows the IR emission spectrum measured by a high-resolution interferometer from a down-looking orbiting spacecraft, taken at three different geographic locations. Also shown are blackbody curves for typical terrestrial temperatures. The spectral variable in Fig. 1.3 is *wavenumber* $\tilde{\nu} = 1/\lambda$, commonly expressed in units of [cm^{-1}]. Again, as for the solar spectrum, the deviations from blackbody curves 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 ~ 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. 1.3). 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.

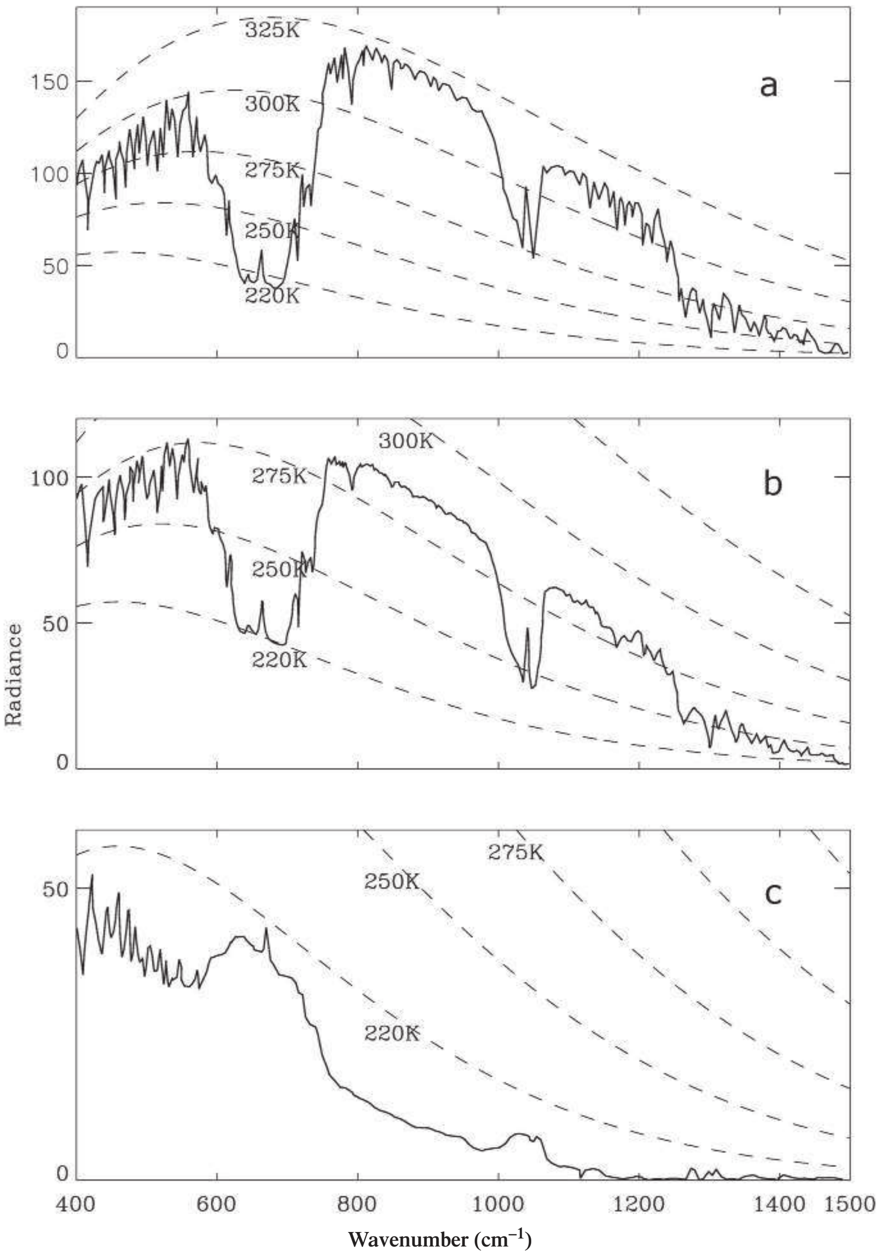

Figure 1.3 Thermal emission spectra of Earth measured by the IRIS Michelson interferometer instrument on the Nimbus 4 spacecraft (Hanel and Conrath, 1970). Shown also are the radiances of blackbodies at several temperatures: (a) Sahara region; (b) Mediterranean; (c) Antarctic.

The effect of windows is clearly seen in Fig. 1.3. In the high-transparency regions, the Earth's surface emission is evident. The contribution of the upwelling atmospheric radiation occurs within the opaque bands, at an effective temperature lower than that of the surface. The emitted radiance is reduced in the regions of high opacity, because the radiation received by the satellite instrument is emitted from the upper colder atmospheric regions, where the lines are optically thin. Notice that in the case of the Antarctic, where the surface is colder than the atmosphere, more radiation is emitted from the warmer atmosphere in the vicinity of the bands than from the surface (in the windows).

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. 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. (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.

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 internal vibrational energy 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 circumstance is contingent on the gas density being sufficiently high, so that *Kirchhoff's Law* is obeyed (§5.2.1). This condition is met for the lower portions of most planetary atmospheres and for the ocean.

1.3.2 Types of Matter Which Affect Radiation

Pretending that they are independent of the radiation, we now focus on those aspects of oceans and atmospheres which are important in modifying the radiation field. For our purposes, there are four forms of matter which can affect radiation:

Gaseous matter: Under local thermodynamic equilibrium conditions (§5.2.1), the density ρ, 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 ρ 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). ρ, 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 §1.4).

Aqueous matter: Similar to 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 which can be quite different from their parent-solid bulk optical properties (§4.2). Oceanic particles consist of a large variety of suspended organic and inorganic substances, such as the variously pigmented phytoplankton and mineral compounds.[5] Particles that are small enough to pass through a standard filter are referred to as "dissolved" and include the organic yellow substances.[6]

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

[5] For a good discussion of the exchange of energy and optically significant constituents between the ocean and the atmosphere see, Bréviére et al. (2015).

[6] Yellow substances are a large class of dissolved organic material derived mainly from the remains and metabolic products of marine plants and animals; see Jerlov (1968).

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 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 (1×10^3 [$\mathrm{kg \cdot m^{-3}}$]) is much higher than that of most planetary atmospheres. For the Earth's atmosphere on a clear day at sea level, visible light 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, at 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 Equilibrium and Ideal Gas Laws

In this section, we describe some important *bulk properties* of the atmosphere and ocean, in terms of their density, pressure, temperature, and index of refraction. As a result of gas being highly compressible, the *atmospheric density*, ρ [$\mathrm{kg \cdot m^{-3}}$], 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 situation 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, hydrostatic equilibrium 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 $dM = \rho dV = \rho dA dz$, 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, $-dpdA = g\rho dA dz$, we find

$$dp = -g\rho dz. \tag{1.1}$$