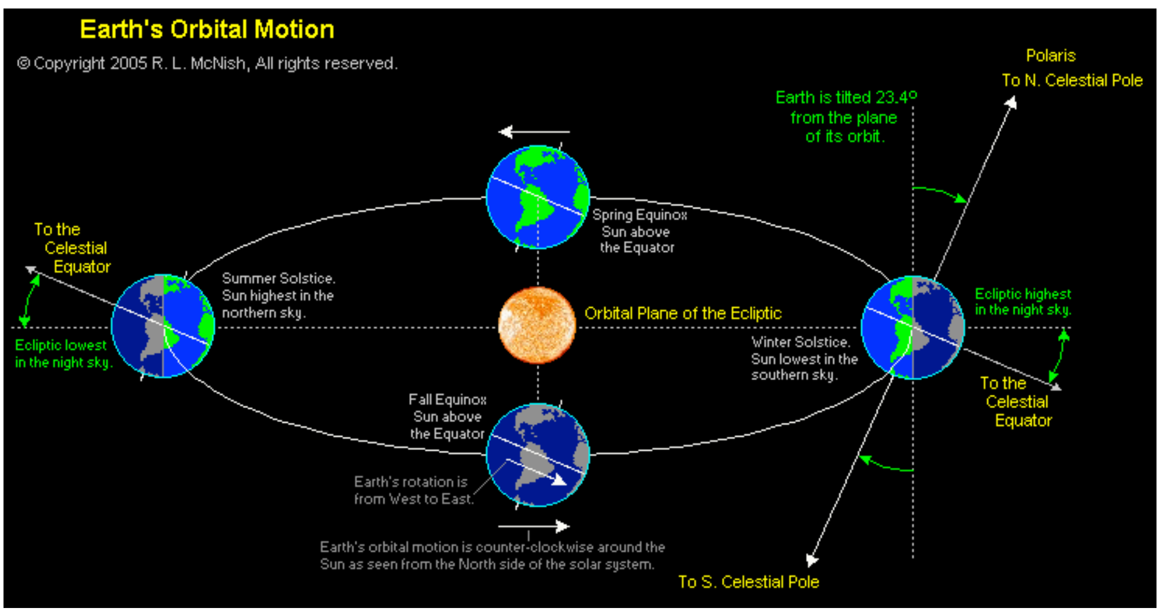
1. Basic Earth and atmospheric properties

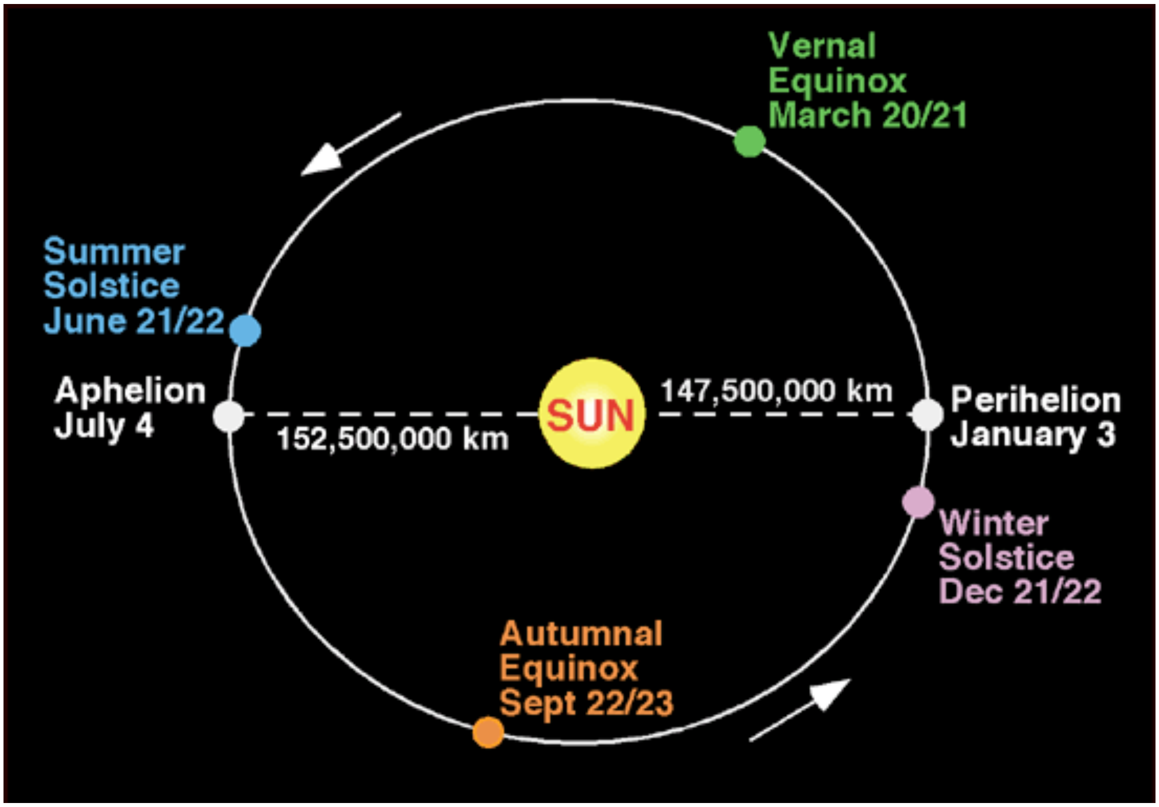
1.1 Earth's orbit and the seasons

The Earth orbits the Sun in a slightly elliptical orbit (eccentricity = 0.0167), with a semi-major axis of 1.496×10^8 km (1 *astronomical unit*, au). It is closest to the Sun (the *perihelion*) on January 3 and farthest from the Sun (the *aphelion*) on July 4. Its orbital period (the *sidereal* orbit, equal to the period relative to the distant "fixed" stars is 365.256 days.

The *ecliptic* is the plane defined by the orbit of the Earth about the Sun. It is the 23.45° tilt of the Earth's rotation axis with respect to the ecliptic (its *obliquity*) that is primarily responsible for the Earth's seasons: The winter solstice is the point in the Earth's orbit when the North Pole is most directly pointed away from the Sun (putting the Sun lowest in the northern sky). At the summer solstice, the North Pole is most directly pointed toward the Sun, and the Sun is highest in the northern sky. At the vernal and autumnal equinoxes, the Sun is above the Equator.



http://www.aapscience7.net/Chapter%20Work/radec_earth_orbit.gif, from Bowie Jr. High School, Odessa, TX.



http://earthguide.ucsd.edu/images/bas/sun/helions.gif, Geosciences Research Division, Scripps Institution of Oceanography

Also, see EPS238-2014/refdata/solar/solar.dat

1.2 Hydrostatic equilibrium

The Earth's atmosphere is, and other planetary atmospheres are, normally well-described as being close to a state of *hydrostatic equilibrium*, wherein the relationship of pressure, P, altitude, z, and temperature, T, is given by

$$dP = -g\rho dz, \quad \rho = \frac{\bar{m}P}{RT}. \tag{1.1}$$

Then

$$\frac{dP}{P} = d\ln P = -\frac{dzg\bar{m}}{RT} \equiv -\frac{dz}{H}. \tag{1.2}$$

g is the gravitational acceleration corresponding to the altitude (it varies little over the distance corresponding to the bulk of the atmosphere, except for the gas giant planets), ρ the gas density, $\bar{m}$ the mean molecular weight of the gas (28.9644 grams per mole, for normal air), R the ideal gas constant (82.057361 cm^3 atmosphere mole^{-1} ^{o}K^{-1}), and H is the local pressure *scale height*, defined to describe the change of pressure with altitude. A typical range of scale heights in the lower and middle Earth's atmosphere is 7-8 km. Thus, hydrostatic equilibrium can be conveniently used to relate pressure and temperature (or to determine the overall atmospheric structure, given two of the three variables z, P, and T). Density scale heights may also be defined. Scale heights are sometimes used to describe variations of the vertical profiles for different gases, which can vary with gas, sometimes dramatically, as well as with height. In this case, they are not determined by the above simple thermodynamic relationship, since photochemical activity may strongly affect distributions with respect to altitude and solar zenith angle.

1.3 Adiabatic lapse rate

The decrease in atmospheric temperature from purely thermodynamic considerations can be derived by considering the work done by air as it expands under conditions where the heat content, q, does not change, $dq = 0$, that is, adiabatic expansion.

$$\frac{dT}{dz} = \frac{-g}{c_p}; \; -10^{\circ} K / km, \tag{1.3}$$

where c_p is the heat capacity at constant pressure (See *Houghton* **1.4**, for the full derivation).

1.4 Albedo and spectral reflectance

The albedo of a scene of a planetary body is the ratio of the outgoing to the incoming flux of total solar radiation. For Earth, the global average is about 0.3, representing a weighted average of bright scenes (particularly those containing clouds, ice, and snow), and the normally darker ocean and land scenes.

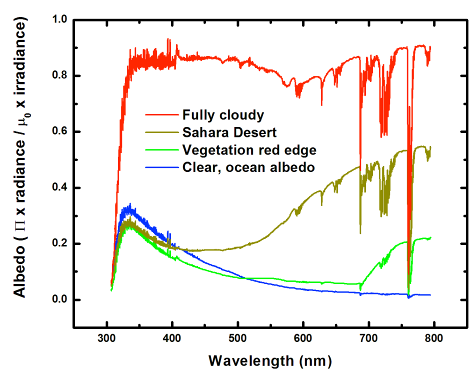


Figure 1.3. Reflectance for several prototypical scenes in the near-ultraviolet, visible, and near infrared from the European Space Agency's Global Ozone Monitoring Experiment. μ_0 is the cosine of the solar zenith angle.

The spectral reflectance of Earth scenes, illustrated for several important cases in **Figure 1.3**, demonstrates major contributions to the albedo. As the solar radiation peaks in the visible (400-700 nm, Chapter 4), this wavelength range dominates the albedo. Fully-cloudy scenes, and snow/ice covered scenes are bright and nearly white (pure white is reflectance = 1). Clear ocean and land scenes are substantially darker and largely blue in color, due to the wavelength-dependent *Rayleigh scattering* (scattering by air molecules, described in Chapter 10). Vegetation-covered scenes also include substantial absorption in the visible by chlorophyll (with a minimum absorption seen in the figure near 550 nm) and a substantial contribution to reflectance in the near infrared from the internal structure of leaves. Desert scenes are, as expected, quite bright in the visible and infrared wavelength regions. The sharp molecular absorptions and detailed scattering features will be discussed in later chapters in more detail.

1.5 Basic Earth's atmospheric structure and variability - *troposphere - tropopause - stratosphere - stratopause - mesosphere – thermosphere*

Figure 1.4 shows the approximate shape and range of variability of the height (and pressure) versus temperature profile for the Earth's atmosphere. Note particular the higher, and colder, tropopause at summer and low latitudes versus the lower, and warmer, tropopause at high latitude spring (except in the polar vortex). See *Houghton*, **Figure 5.1** for more detail.

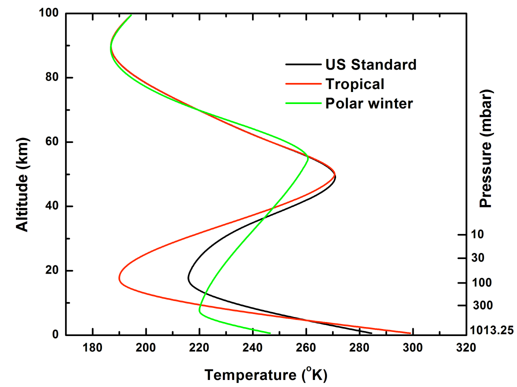


Figure 1.4 Basic structure and variability

The lowest layer of the Earth's atmosphere is the *troposphere*. The temperature decreases with altitude, due to the thermodynamic lapse until the decrease is balanced by radiative heating in the stratosphere (particularly absorption of UV radiation by ozone). This point is the tropopause, which normally located between 8-18 km, dependent on location and season. Temperature then rises in the stratosphere, due to absorption by O_3 and, especially higher in the stratosphere, by O_2, until the *stratopause* (about 50 km), the boundary between the stratosphere and mesosphere. Temperature decreases through thermodynamic lapse in the mesosphere until the *mesopause* (about 82 km), its boundary with the thermosphere, where temperature rises again, now due to absorption of shorter-wavelength UV by O_2.

Appendix A gives examples of *z*, *P*, and *T* for tropical, midlatitude summer and winter, subarctic summer and winter, and the U.S. 1976 Standard (*U.S. Standard Atmosphere, 1976*). For future reference, Appendix B contains typical atmospheric profiles of constituent mixing ratios, adapted from *Brasseur and Solomon* (Brasseur and Solomon contains good discussions of stratospheric gas distributions in particular, by chemical families).

1.6 Overview of atmospheric composition.

TBD

References

Brasseur, G.P., and S. Solomon, Aeronomy of the Middle Atmosphere: Chemistry and Physics of the Stratosphere and Mesosphere, 3rd Edition, Springer, Dordrecht, ISBN: 1402032846, 2005

Houghton, J.T., The Physics of Atmospheres, 3rd Edition, Cambridge University Press, New York, ISBN: 0521804566Houghton, 2002.

U.S. Standard Atmosphere, 1976, National Oceanic and Atmospheric Administration, 1976.

Problems (assigned January 30, due February 11)

1.1 Note the use of the ideal gas law to determine ρ. How much would the use of the van der Waals correction for non-ideal behavior change the P,T relationship in the Earth's atmosphere?

1.2 Use the data in us**76.dat** to determine the atmospheric scale height from the ground to the stratopause. Plot it against altitude and also against temperature.