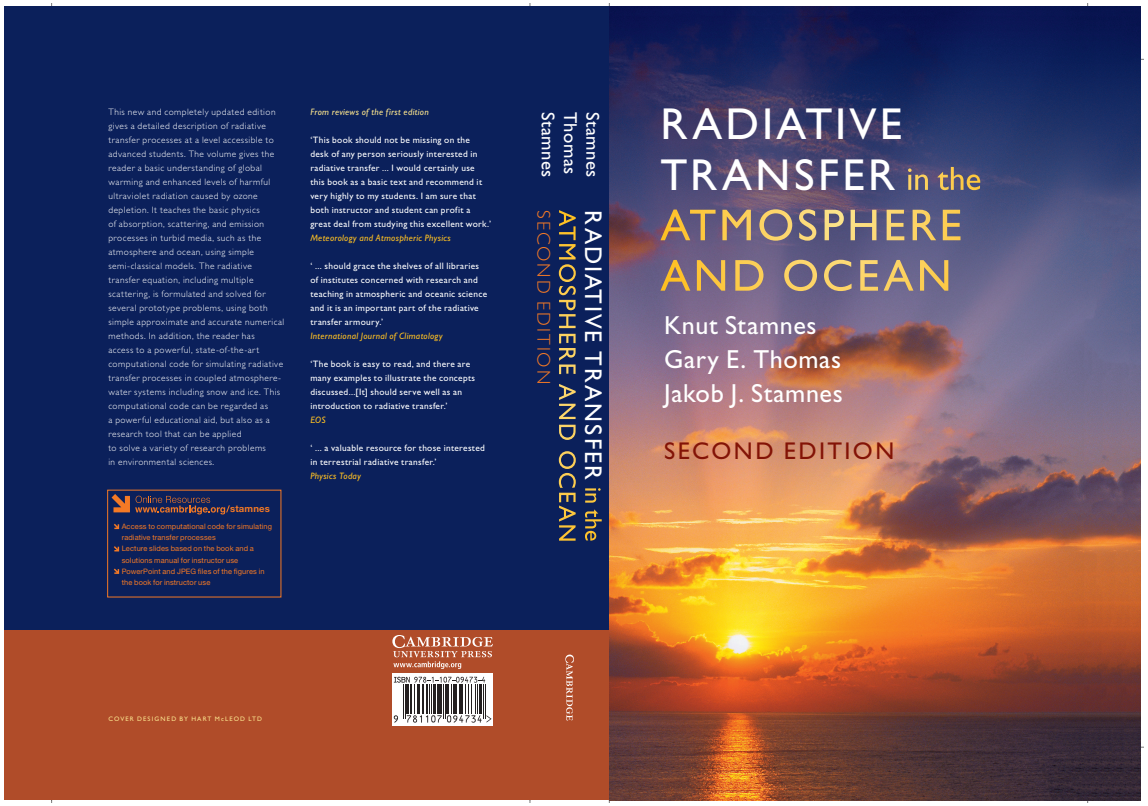
# Lecture *Notes:*
# Basic Atmosphere/Ocean Properties




**Based on Chapter 1 in K. Stamnes, G. E. Thomas, and J. J. Stamnes, Radiative Transfer in the Atmosphere and Ocean, Cambridge University Press, 2017.**

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# Basic Properties of Radiation, Atmospheres and Oceans (1)

## Shortwave Solar Spectrum and Longwave Terrestrial Spectrum:

- The solar spectrum consists of shortwave radiation: Ultraviolet radiation (290 nm $< \lambda <$ 400 nm), visible light (400 nm $< \lambda <$ 700 nm), and near-infrared (near-IR) radiation (700 nm $< \lambda <$ 3,500 nm).
- The spectrum emitted by the Earth, also called the **terrestrial** radiation, consists of longwave, thermal infrared radiation with wavelengths $\lambda >$ 3,500 nm.
- Interaction Processes:
  - **Photoionization** is caused by X-rays and extreme UV radiation
  - **Photodissociation** is caused by extreme UV (10 nm $< \lambda <$ 100 nm), far UV (100 nm $< \lambda <$ 200 nm), and middle UV or UV-C (200 nm $< \lambda <$ 280 nm) radiation
  - Absorption of UV-B (280 nm $< \lambda <$ 320 nm) radiation by $O_3$ in the Earth's ozone layer
  - **Scattering** of UV-A (320 nm $< \lambda <$ 400 nm) and visible radiation by clouds and aerosols
  - **Absorption** of near-IR radiation (0.7 $\mu$m $< \lambda <$ 3.5 $\mu$m) by trace gases like $O_3$, $H_2O$ and $CO_2$
  - **Absorption and Emission** of thermal IR radiation (3.5 $\mu$m $< \lambda <$ 100 $\mu$m) by surfaces and IR-active gases.

# Basic Properties of Radiation, Atmospheres and Oceans (2)

Table 1: Sub-regions of the Spectrum. PAR stands for "Photosynthetically Active Radiation".

| Sub-region | 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 active gases. |

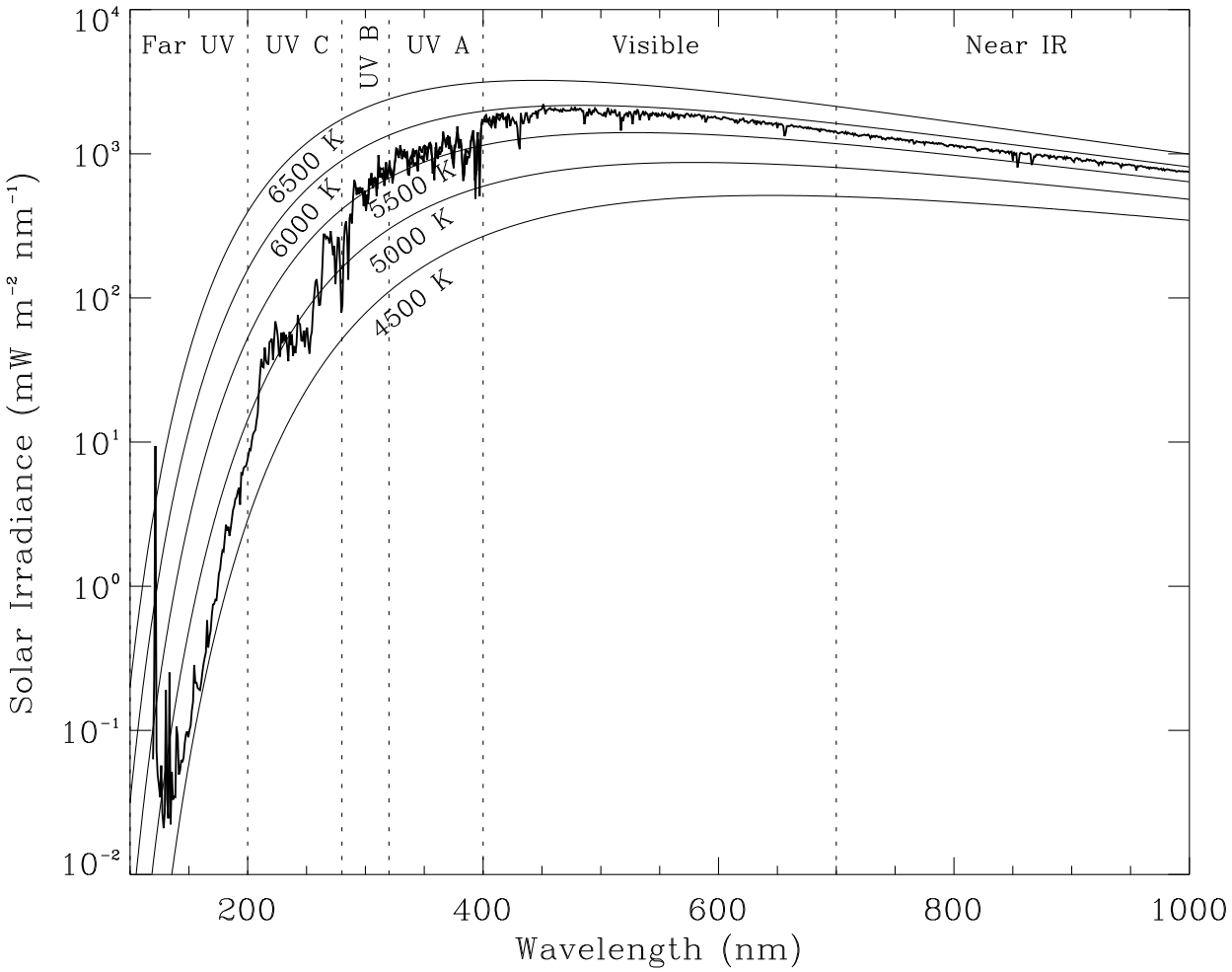


Figure 1: **Extraterrestrial solar irradiance, measured by a spectrometer on board an earth-orbiting satellite. The UV spectrum ($119 < \lambda < 420$ nm) was measured by the SOLSTICE instrument on the UARS satellite (modified from a diagram provided by G. J. Rottmann, private communication, 1995). The vertical lines divide the various spectral sub-ranges defined in Table 1. The smooth curves are calculated blackbody spectra (see Chapter 3) for a number of emission temperatures.**

## Extraterestrial Solar Irradiance

- **Absorption opacity** or **optical depth** $\tau(\nu)$ plays a major role.
- The measured extraterrestrial solar spectral irradiance $S(\nu)$ is shown in Fig. 1 and compared to the radiation emitted by a blackbody for several temperatures. The measurements were taken at a distance of one astronomical unit $r_\oplus$ ($r_\oplus = 1.5 \times 10^8$ km).
- By integrating $S(\nu)$ over all frequencies we obtain the solar constant $S_0 = \int d\nu\, S(\nu)$ [ W·m$^{-2}$].
- This so-called solar constant is not actually a constant, but slightly variable. It is called the *total solar irradiance*, which is about $S_0 = 1368$ W·m$^{-2}$.*
- $S_0$ represents the total instantaneous radiant energy falling on a unit surface located at a distance $r_\oplus$ ($r_\oplus = 1.4960 \times 10^{11}$ m) from the Sun.
- By assuming that the total solar energy emitted by the Sun is the same as that from a blackbody, we find that the effective temperature of the Sun is 5778 K.



*The most current published value is ($S_0 = 1360.8 \pm 0.5$) W · m$^{-2}$, Kopp and Lean (2011).

## Extraterestrial Solar Irradiance (cont.)

- If the radiating layers of the Sun had the same temperature at all depths, its spectrum would be exactly the same as that from a theoretical blackbody.
- Deviations from a blackbody curve are due to emissions from a nonisothermal atmosphere. Radiative transfer is required to explain this behavior.
- The visible solar spectrum can be qualitatively explained by considering **two characteristics** of atmospheres and **one basic rule** obeyed by radiating objects. The **two characteristics** are that:
  - the optical depth or opacity varies with frequency, i.e., $\tau = \tau(\nu)$,
  - the temperature varies with depth.

## Extraterestrial Solar Irradiance (cont.)

- The **basic rule** is that a radiating object **emits energy to space most efficiently** at wavelengths where **the opacity is approximately unity**. This rule can be explained as follows:
  - In spectral regions where the atmosphere is transparent so that $\tau(\nu) << 1$, it neither absorbs nor emits.
  - In spectral regions where the atmosphere is opaque so that $\tau(\nu) >> 1$, its radiative energy is prevented from leaving the atmosphere, because it is reabsorbed by the surrounding medium.
  - When $\tau(\nu) \approx 1$, a balance is struck between these opposing tendencies.
- At visible wavelengths, the Sun's opacity is unity deep within the solar atmosphere in the relatively cool photosphere, where the temperature is approximately 5780 K (Fig. 1).

## Earth's Energy Balance (1)

Figure 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.†
- 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**.



†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).

# Basic Properties of Radiation, Atmospheres and Oceans (5c)

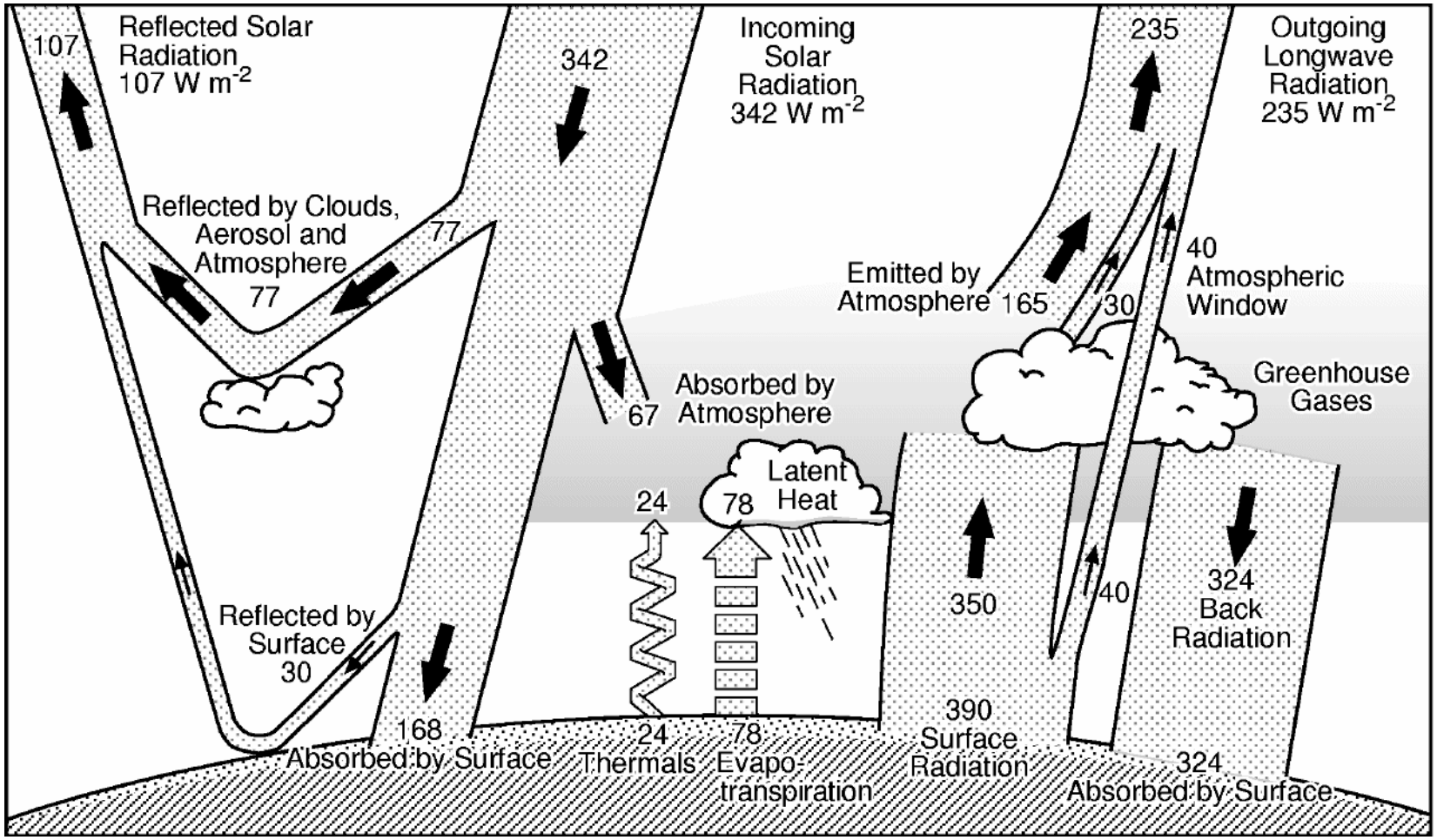


Figure 2: **Earth's energy budget based on a $S_0$ value of 1368 $\mathrm{W \cdot m^{-2}}$ [adapted from Trenberth et al. (1997). American Meteorological Society. Used with permission.]. An updated version of this illustration with revised budget numbers based on a $S_0$ value of 1365.2 $\mathrm{W \cdot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

## Earth's Energy Balance (2)

- The well-documented increase in greenhouse gases ($CO_2$ and others), above what is believed to be the natural level existing in the pre–industrial era, 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 until a new radiative equilibrium (see §4.3.1) 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 the energy received.

The amount of warming depends crucially on how the entire Earth climate system responds. 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.

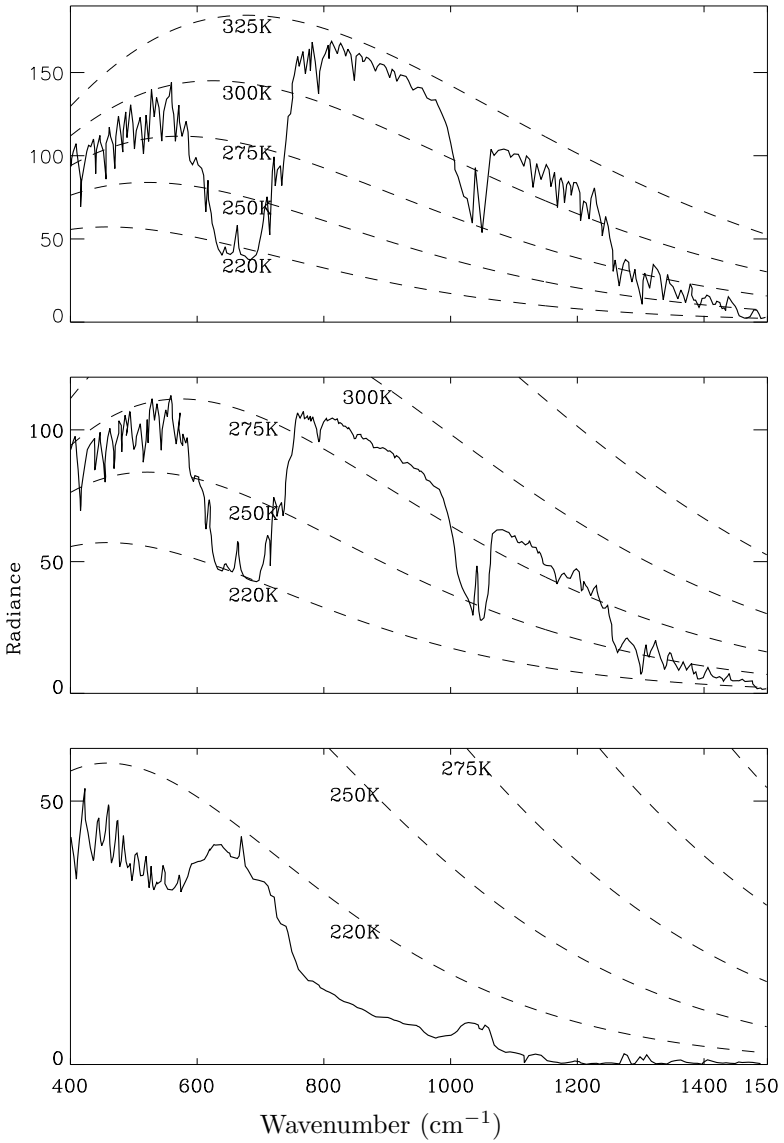


Figure 3: **Thermal emission spectra of the Earth measured by the IRIS Michelson interferometer in the *Nimbus 4* spacecraft. Also shown are radiances of blackbodies at several temperatures. Top: Sahara region; Middle: Mediterranean; Bottom: Antarctic.**

## Terrestrial IR radiation

- An understanding of radiative transfer is essential also for understanding the emission from the Earth at wavelengths $\lambda > 3.5$ $\mu$m.
- Figure 3 shows the IR emission spectrum measured from a down-looking satellite instrument at three different locations: over Sahara, the Mediterranean, and the Antarctic. Note that on the horizontal axis the frequency is given in wavenumber units [$\mathrm{cm}^{-1}$].
- Deviations from blackbody behavior are due to the **nonisothermal** nature of the Earth's atmosphere.
- The spectral regions of minimum emission arise from the upper cold regions of the Earth's atmosphere, where the opacity of the overlying regions is approximately unity.
- Highest emission arises from the warm surface in the transparent spectral regions ("windows"). The Antarctic represents an exception, because the surface there is actually colder than the overlying atmosphere.

## Radiative Interaction with Planetary Media – Feedback Processes

- We consider two examples of feedback processes.
  1. During daytime, solar radiation warms the Earth's surface and atmosphere – air is set in convective motion. Upward air motion causes adiabatic cooling and, if the atmosphere is sufficiently moist, will lead to condensation and cloud formation. Clouds will reflect and absorb incoming sunlight, and scatter, absorb and emit IR radiation, and thus affect the warming.
  2. The photosynthesis in the ocean determines the concentration of light-absorbing phytoplankton, which in turn determines the light penetration depth.
- We will ignore such feedback processes, because the speed of light is so high that the radiation field adjusts itself instantaneously to the environment. As a result radiative transfer is essentially a quasi-static phenomenon, and consequently its interaction with matter can be treated separately from other influences.
- We ignore feedback processes also on the microphysical scale. For example, we assume that the temperature of a gas is given independent of the light intensity. This situation holds if the density is sufficiently high that Kirchhoff's Law[‡] is obeyed.



[‡]Applied to a planetary surface, Kirchhoff's Law states that the energy absorbed equals the energy emitted: "a good absorber is a good emitter".

## Types of Matter that Affect Radiation

- **Gaseous matter.** In local thermodynamic equilibrium (LTE)[§] the optical properties are determined by density $\rho$, temperature $T$, and chemical composition. In addition comes gas pressure $p$, although it depends on $\rho$ and $T$, either through empirical formulas or through the ideal gas law.
- **Aqueous matter.** The optical properties of liquids are largely determined by the density of pure water. Salinity, which is important for ocean dynamics, is unimportant for the optical properties, but it is worth noting that "pure water" hardly exists outside the laboratory. "Impurities" such as dissolved organic matter and particles usually dominate the optical properties of natural water.
- **Particles.** Particulate matter in the atmosphere consists of suspended particles (**aerosols**) and condensed water (**hygrosols**). Hygrosols is the generic term for **water droplets** or **ice crystals**, whereas aerosols is the generic term for all other **suspended particles** in the atmosphere. **Marine particles** consist of a large variety of **suspended organic and inorganic substances**. **Air bubbles** in the water can also influence its optical properties.



[§]See §4.3.1 for a definition of LTE.

## Vertical Structure of Planetary Atmospheres

- Atmospheres are highly *compressible.*
- Oceans are nearly *incompressible.*
- Oceans have much greater density than most planetary atmospheres. (Deep inside the atmosphere of Venus the atmospheric density can approach and even exceed that of water.)
- On a clear day at sea level, a visible light photon can propagate without attenuation many hundreds of kilometers along a horizontal path, but in the ocean it penetrates at most a few hundred meters before being attenuated.
- **Hydrostatic and Ideal Gas Laws.**
  - For both oceans and atmospheres **hydrostatic equilibrium** prevails: **pressure $p$ must support the weight of the fluid above it**.
  - Boyle's lov: With increasing height in the atmosphere the density decreases as the pressure decreases. With increasing depth in the ocean, this situation also prevails, but the change in density is slight.

## Atmospheric case

The weight of a small volume element $dV$ is $gdm$, where $g$ is the acceleration due to gravity and $dm$ is the mass of air inside $dV$.

For a vertical cylindrical volume element with area $dA$ and height $dz$, we have:

$$dm = \rho dV = \rho dA dz$$

Net force on the volume element $dV$ from the surrounding gas is $-dpdA$, where the minus sign indicates that the pressure on the upper end of the cylinder located at height $z + dz$ is less than the pressure on the lower end at height $z$, i.e. $dp < 0$.

The weight of the cylinder must be balanced by the pressure force. Thus: $gdm = g\rho dV = g\rho dA dz = -dpdA$ or:

$$dp = -g\rho dz. \tag{1}$$

For planetary atmospheres of moderate and low density:

$$\rho = \frac{\bar{M}p}{RT} = \bar{M}n \quad \Longleftarrow \quad p = nRT \quad \Longleftarrow \quad \text{ideal gas law.} \tag{2}$$

Here $\bar{M}$ 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).

Substituting Eq. 2 into Eq. 1, we find:

$$\frac{dp}{p} = -\frac{dz}{H} \quad \text{where} \quad H \equiv \frac{RT}{\bar{M}g}. \tag{3}$$

$H$ is called the **atmospheric scale height**. Below about 95 km (the *homopause*) in the Earth's atmosphere, $H$ is approximately 29.3 $T$ [m], where the temperature $T$ is given in units of Kelvin [K].

- Below the *homopause*, air contains a "standard" mixture of 78% nitrogen ($M(\mathrm{N_2}) = 28$), 21% oxygen ($M(\mathrm{O_2}) = 32$), and about 1% argon ($M(\mathrm{Ar}) = 40$).
- Above the *homopause*, $\bar{M}$ decreases as the composition changes to lower-mass species, such as O, N, and He.
- This change in composition results from photodissociation of the heavier molecules, combined with gravitational separation tending to place the lighter elements on top.

Integrating Eq. 3 from $z' = z_0$ to $z' = z$, and using Eq. 2, we find:

Three forms of the *hydrostatic equation*

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

$$n(z) = n(z_0)\frac{T(z_0)}{T(z)}\exp\left[-\int_{z_0}^{z} dz'/H(z')\right] \quad \text{(b)} \qquad (4)$$

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

Knowing the surface pressure $p(z_0)$ and the variation of scale height $H(z)$ from $z_0$ to $z$, we may use Eqs. 4 to determine the bulk gas properties at any height $z$.
If $g$, $T$, and $\bar{M}$ are constant with height, we can integrate Eqs. 4 to obtain:

$$\frac{p(z)}{p(z_0)} \propto e^{-(z-z_0)/H} \text{ (a)}; \quad \frac{n(z)}{n(z_0)} \propto e^{-(z-z_0)/H} \text{ (b)}; \quad \frac{\rho(z)}{\rho(z_0)} \propto e^{-(z-z_0)/H} \text{ (c)}. \qquad (5)$$

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

# Basic Properties of Radiation, Atmospheres and Oceans (13)

- **Dalton's Law** for an ideal gas: 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*. BUT:
- the hydrostatic equation **does not** apply to ozone ($O_3$) and water ($H_2O$), because both are subject to rapid changes. **Ozone** is formed and destroyed by **chemical reactions**, and **water** undergoes **phase transitions**.
- An important property of atmospheres and oceans is that they have a tendency to form a **vertically stratified**, but horizontally homogeneous medium. Often we therefore assume that the atmosphere or a body of water is vertically stratified locally.
- Important exceptions are clouds and aerosol layers in the atmosphere and algal blooms in the ocean.
- If the assumption of vertical stratification is fulfilled, then we talk about a **plane parallel medium** without variations in the horizontal.

# Basic Properties of Radiation, Atmospheres and Oceans (14)

- The temperature in the Earth's atmosphere decreases with height in the **troposphere**, and has a minimum at the **tropopause** ($\sim$ 10 km);
- Above the tropopause the temperature increases due to absorption of UV light by ozone in the **stratosphere** (10–50 km), but decreases again in the mesosphere (above 50 km) due to decreasing ozone concentration and increased IR cooling.
- At the **mesopause** (85 km) a new temperature minimum occurs ("coldest place on Earth").
- The uppermost layer of the atmosphere is called the **thermosphere**: very warm due to heating by photoionization (energetic UV radiation and X-rays).

What physical processes give rise to the observed temperature structure?

- Upward transport of heat (daytime) tends to expand the air near the surface. Since pressure tends to remain constant, the air density decreases (Eq. 2).
- Lower density $\Longrightarrow$ air is more buoyant than the overlying cooler air: a convective instability may occur: air is set into small-scale turbulent motion. The rising air cools by expansion, and displaces cooler air.

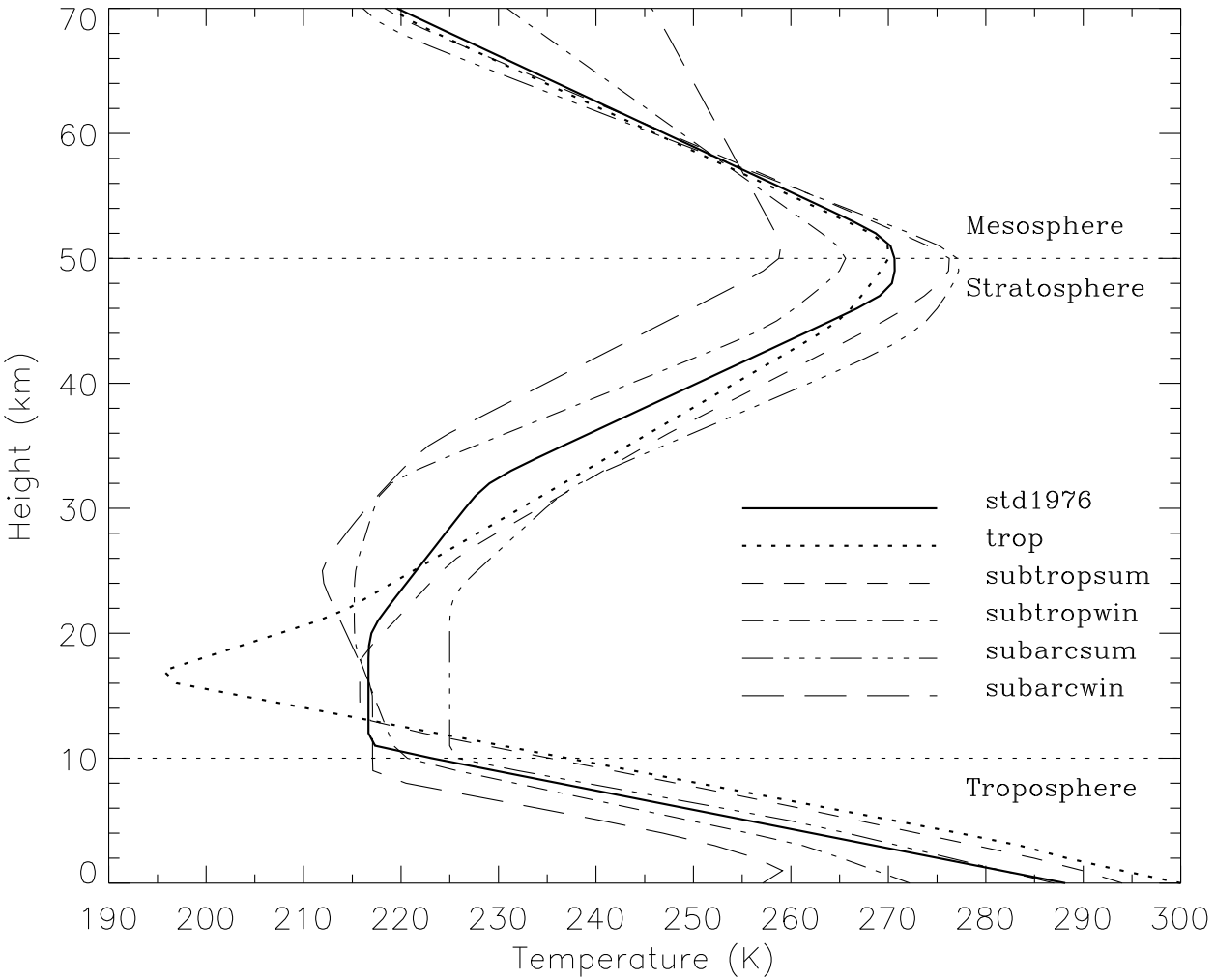


Figure 4: **Standard empirical model temperature profiles for various locations and seasons. The 1976 Standard atmosphere is appropriate for the global mean. The Tropical atmosphere is valid for latitudes less than 30°; the Subtropical model for 30–45°; the Subarctic atmosphere for 45–60°; and the Arctic for 60–90°.**

- Higher density $\Longrightarrow$ the upper cooler air sinks and compressively heats.

If allowed to approach equilibrium, a temperature gradient is set up. In the absence of any other heating/cooling processes (i.e., it is an **adiabatic process**):

$$\left(\frac{\partial T}{\partial z}\right)_{\rm ad} = -\frac{g}{c_{\rm p}} \quad \text{(dry atmosphere)} \tag{6}$$

where $c_{\rm p}$ is the specific heat of (dry) air = 1006 [J$\cdot$ kg$^{-1}\cdot$ K$^{-1}$]. This temperature gradient is called the *dry adiabatic lapse rate*; for Earth its value is approximately $-9.8$ K$\cdot$ km$^{-1}$.

- For moist air, condensation and release of latent heat causes the gradient to be considerably smaller (in magnitude) than $-g/c_p$. In fact, $\partial T/\partial z$ can be as small as $-3$ K$\cdot$ km$^{-1}$, depending upon the moisture content of the air.
- The tendency towards an **adiabatic lapse rate** for an atmosphere heated from below is a **fundamental property** of planetary atmospheres. Figure 4 shows a nearly-linear temperature lapse rate of $-6.5$ K$\cdot$ km$^{-1}$ in the troposphere.
- The explanation of the *tropopause*, and the increasing temperature in the *stratosphere*, lies in radiative processes, solar UV absorption by the ozone layer.

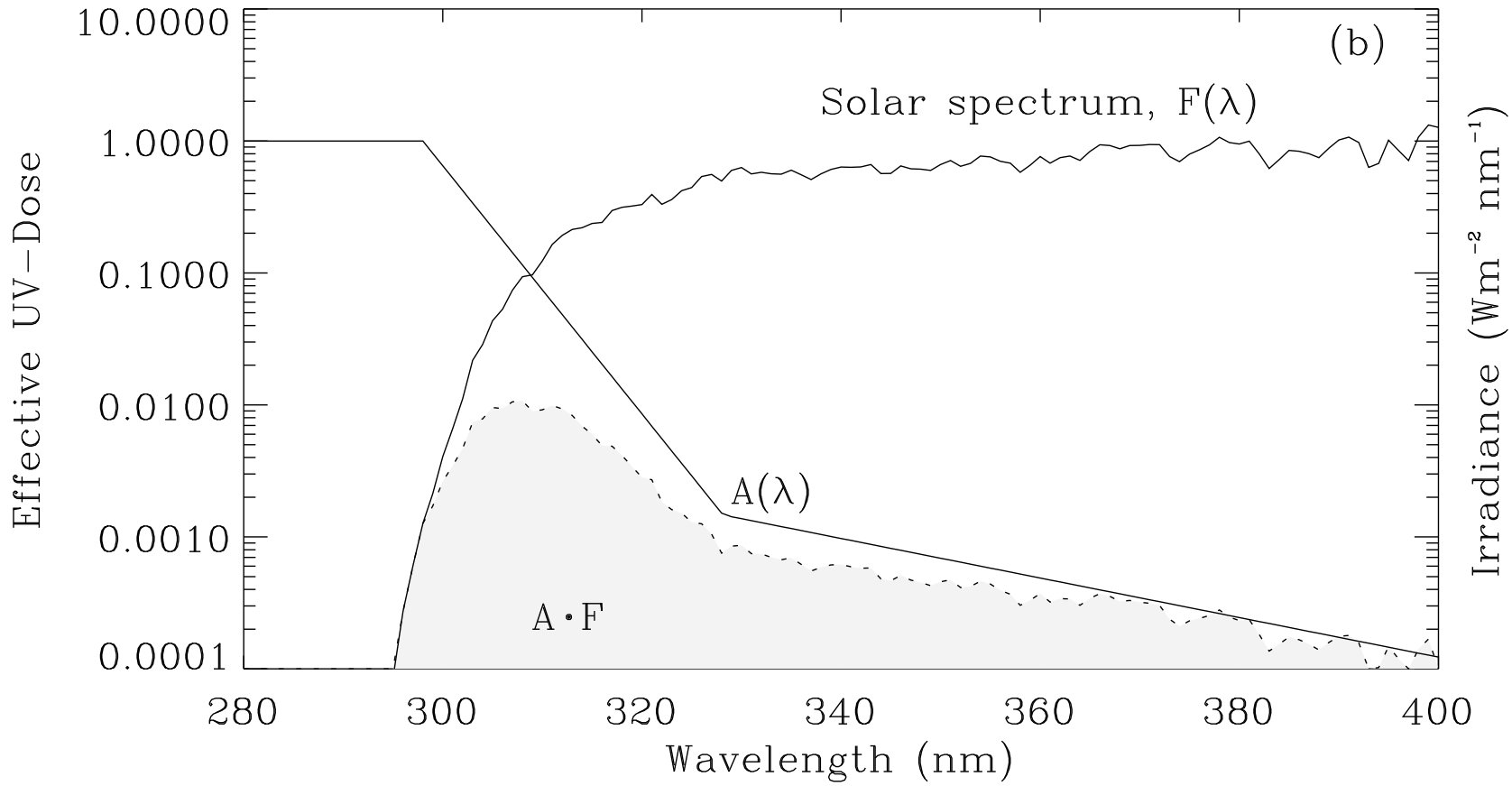


Figure 5: **Illustration of chemical and biological effects of solar radiation. The UV exposure is determined by the product of a biological "action spectrum", $A(\lambda)$, and the solar irradiance, $F_\lambda$. The instantaneous dose rate is the area under the product curve as indicated.**

## Minor Species in the Atmosphere

- There are dozens of chemical species in the Earth's atmosphere: The ones of greatest interest are those which interact with UV and IR radiation, the poly-atomic **radiatively significant** gases.
- The principal gases $N_2$, $O_2$ and Ar are generally unimportant for visible and IR absorption.
- The study of atmospheric chemistry has importance in problems of air pollution, climate, and the health of the biosphere.
- Shortwave radiation plays a crucial role in determining the concentrations of **photochemically active** species through the process of **photolysis** in which molecules are split up into smaller 'fragments' (atoms and molecules).
- One of these species, the ozone ($O_3$) molecule, is of paramount importance in providing life a protective shield to biologically damaging ultraviolet radiation.
- The response of biological systems to UV radiation is illustrated in Fig. 5.

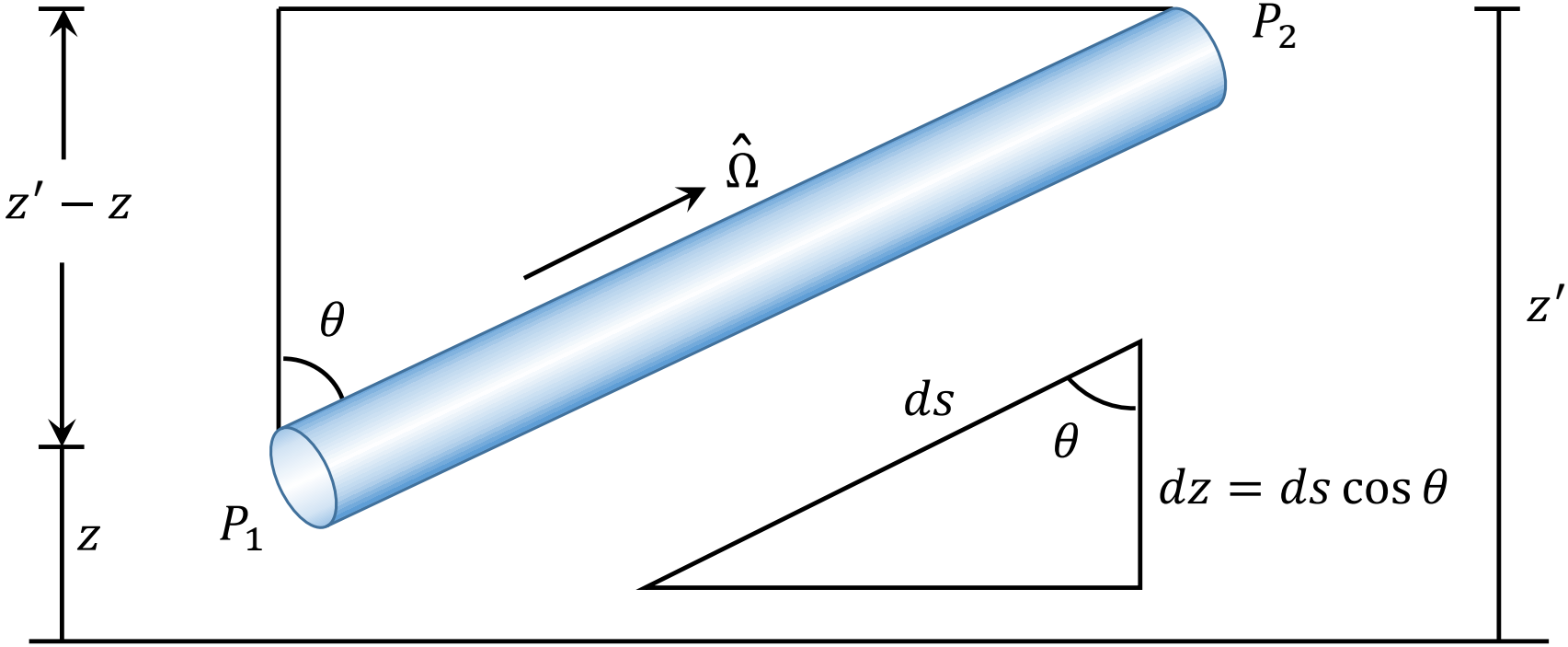


Figure 6: **Geometry of the slant-column number. The right-circular cylinder of unit geometrical cross section contains $\mathcal{N}$ molecules (number per m$^2$).**

## Optical Line-of-sight Columns

A line-of-sight integration over the $i$th species density, given by $\rho_i(z)$ is required to:

- evaluate the line-of-sight **slant column mass**, $\mathcal{M}_i$ between points $P_1$ and $P_2$, along the direction of propagation $\hat{\Omega}$.

Assuming constant temperature and gravity, we integrate along the line-of-sight distance variable $s$ to obtain

$$\mathcal{M}_i(1,2) \;=\; \int_1^2 ds\rho_i(s) \;= \sec\theta\left[\int_{z_1}^{\infty} dz\rho_i(z) - \int_{z_2}^{\infty} dz\rho_i(z)\right]. \tag{7}$$

Referring to Fig. 6, we have transformed from the variable $ds$ to $dz$, $ds = dz\sec\theta$, where $\theta$ is the polar angle $(0 \leq \theta < \pi/2)$ made by the vector $\hat{\Omega}$ with the vertical, $\cos\theta = |\hat{z}\cdot\hat{\Omega}|$. $\hat{z}$ is a unit vector in the positive $z$-direction. Using Eq. 5c $(\frac{\rho(z)}{\rho(z_0)} \propto e^{-(z-z_0)/H})$, we find:

$$\mathcal{M}_i(1,2) = \rho_i(z_0)H_i\sec\theta(e^{-z_1/H_i} - e^{-z_2/H_i}) = [\rho_i(z_1) - \rho_i(z_2)]H_i\sec\theta. \tag{8}$$

## Optical Line-of-sight Columns (cont.)

We find that the species slant column mass between two atmospheric points separated by the heights $z_1$ and $z_2$, and whose direction is $\theta$ from the normal, is the difference of the two quantities $\mathcal{M}_i(1,2) = \mathcal{M}_i(z_1,\theta) - \mathcal{M}_i(z_2,\theta)$, where:

$$\mathcal{M}_i(z,\theta) = \rho_i(z) H_i \sec\theta = \frac{p_i(z)}{g}\sec\theta \tag{9}$$

is the slant column mass between the height $z$ and $z \to \infty$ along the direction $\hat{\Omega}(\theta)$. In the second result we have used the ideal gas law ($p_i = n_i RT$, $\rho_i = M_i n_i = M_i \frac{p_i}{RT}$, and the definition of $H_i = \frac{RT}{M_i g}$). A common notation for **vertical column mass** is $u$.

- The optical depth is obtained by multiplying $\mathcal{M}_i$ by a mass absorption coefficient $\alpha_m$. This multiplication assumes $\alpha_m$ does not vary along the path – otherwise it must be inside the integration over path length.
- We can visualize $\mathcal{M}_i(1,2)$ as being the species atmospheric mass contained within a geometric cylinder of unit cross-section extending from points $P_1$ to $P_2$.

## Optical Line-of-sight Columns (cont.)

Equations 7–9 are invalid for nearly horizontal paths when $\theta$ is near 90°. For $\theta > 82°$, we must consider the spherical nature of the atmosphere. In fact:

- $\mathcal{M}_i$ does *not* diverge as $\theta \to \pi/2$, as predicted by Eq. 9, but reaches a finite value, depending upon the radius of the planet.

We define the **slant column number** $\mathcal{N}_i(1,2)$ as the **number** of molecules (of composition $i$ each with molecular mass $M_i$) in a cylinder of unit cross section:

$$\mathcal{N}_i(1,2) = \int_1^2 ds\; n_i = \mathcal{M}_i(1,2)/M_i = \mathcal{N}_i(z_2,\theta) - \mathcal{N}_i(z_1,\theta). \qquad (10)$$

- $\mathcal{N}_i$ has the unit [$\mathrm{m}^{-2}$ or $\mathrm{cm}^{-2}$], but is more commonly expressed in terms of the length in [cm] of a column filled with the species of interest, and compressed to standard temperature and pressure (273.16 K and 1013.25 mb).
- To convert from [$\mathrm{cm}^{-2}$] to [cm], one divides $\mathcal{N}_i$ by **Lochmidt's number**, the air density at STP, $n_\mathrm{L} = 2.687 \times 10^{19}\ \mathrm{cm}^{-3}$.

# Basic Properties of Radiation, Atmospheres and Oceans (20)

- Well-mixed species (those with long photochemical lifetimes) such as $N_2$, $O_2$, and $CO_2$ may be described by Eqs. 7–9 up to the homopause. BUT:
- most chemical species are photochemically produced and lost over time scales short compared with a mixing time scale (see Fig. 7), implying that we *cannot* use Eqs. 7–9 for the column number, or column mass.
- Nevertheless, the dominant height variation of many species, for example $N_2O$ and $CH_4$, is at least quasi-exponential.

The **mixing ratio** describes deviations from the hydrostatic case:

- the **mass mixing ratio**, $w_m^i(z) = \rho_i(z)/\rho(z)$; or
  the **volume mixing ratio**, $w^i(z) = n_i(z)/n(z)$,
  $\rho(z)$ = total density and $n(z)$ = total concentration of the ambient atmosphere.
- Typically $w^i$ is specified in parts per million by volume [ppmv], or parts per billion by volume [ppbv]. $w^i$ is also sometimes called the *molar fraction* of species $i$. $w_m^i$ is usually specified in $[g \cdot g^{-1}]$ (grams per gram).
- Figure 7 displays the volume mixing ratios of various species versus height.

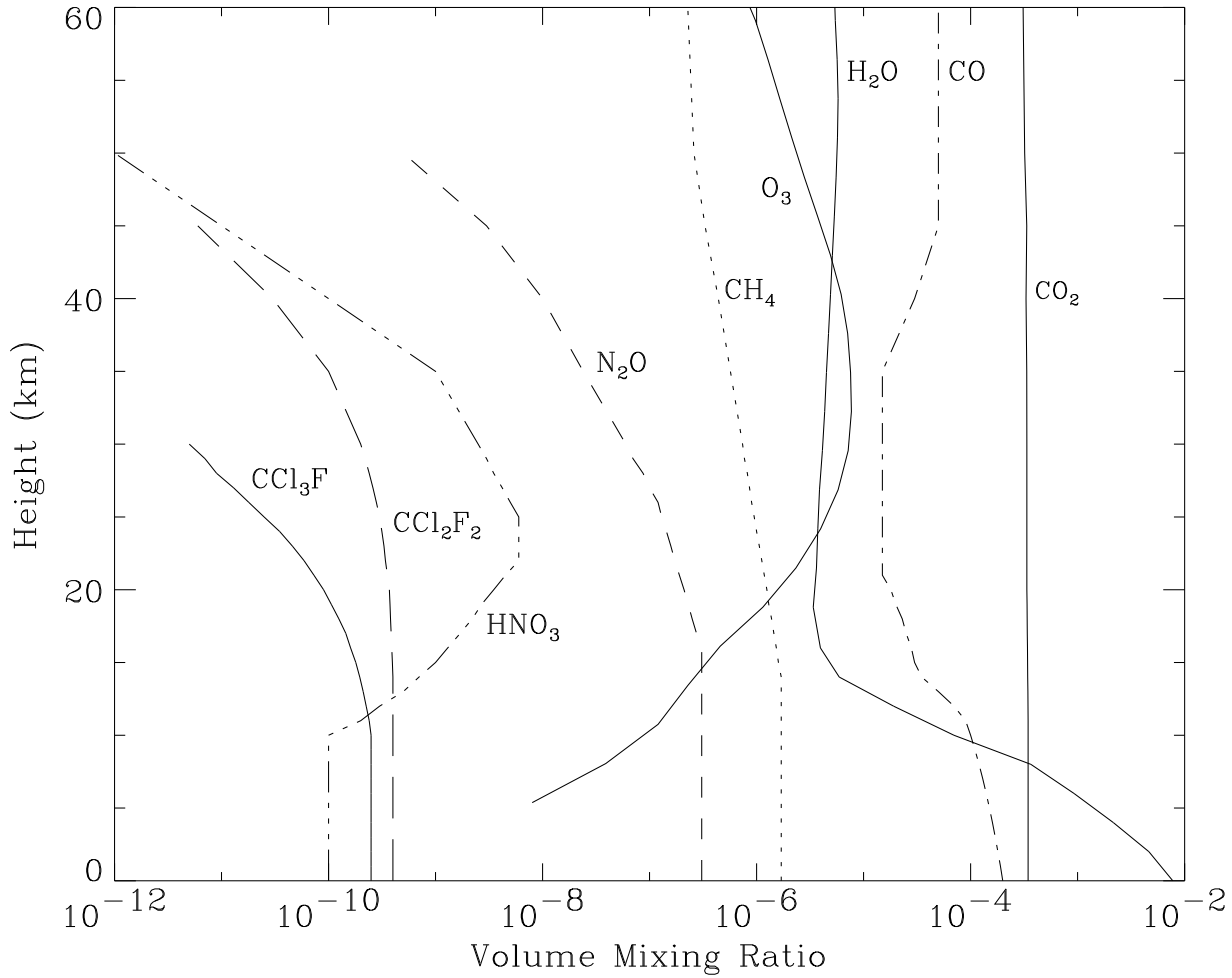


Figure 7: **Radiatively-significant atmospheric constituent height profiles of volume mixing ratio. The ozone and water profiles (above 25 km) are taken from globally-averaged data from the HALOE experiment on board the UARS satellite. The remaining profiles are taken from standard atmosphere compilations and from balloon measurements.**

# Radiative Equilibrium and the Thermal Structure of Atmospheres (21)

- **The Earth (any planet) is heated by absorption of solar radiation, and cooled by emission of thermal IR radiation.**
- Atmospheric temperature plays a key role in the energy balance through its influence on the emission process, because according to *Kirchhoff's Law*:
- Matter emits radiation at a rate that depends both on its absorptive properties and its absolute temperature $T$: **A planet 'sheds' its energy solely by this emission process (radiating to space).**
- At any one time or location the **rates** at which the planet receives energy from the Sun and radiates energy to space **do not balance**:
- The temperature of a rotating planet with an axial tilt will undergo both diurnal and seasonal cycles. However, when **averaged over the entire Earth's surface and over an orbital period**:
- the **mean temperature is expected to remain constant**, in the absence of internal or external changes.

# Radiative Equilibrium and the Thermal Structure of Atmospheres (22)

- This *negative feedback* of the planetary temperature (through Kirchoff's Law) acts like a giant thermostat, continually adjusting the rate of energy loss to compensate for excess heating or cooling.
- The situation in which thermal emission is balanced locally at all wavelengths by heating due to solar radiation is called *local radiative equilibrium*. This situation seldom occurs in the Earth's atmosphere.
- Earth's climate has been remarkably stable over geologic time. Life's continuing existence for 3.8 billion years indicates a relatively stable climate.
- Current concerns are focused on the potential for extreme climate changes that could place great stress on the biosphere: Changes in rainfall patterns or ocean currents due to natural causes or human influences could influence agricultural and ocean productivity and thereby the world's food supply.
- Other planets face different "climate problems". The surface temperature of Venus is $T_s = 750$ K, which requires a powerful greenhouse mechanism.
- Absorption of solar radiation by the dust on Mars leads to warming and temperature inversions.

# Radiative Equilibrium and the Thermal Structure of Atmospheres (23)

- The average column-integrated energy for the entire climate system is

$$\bar{E} \equiv E/4\pi R^2$$

where $E$ is the total thermal energy for the entire system, $R$ the Earth's radius, and $4\pi R^2$ its total surface area.

- According to the First Law of Thermodynamics, the time rate of change of the globally averaged column energy $\bar{E}$ is given by ($\bar{\rho}$ is the spherical albedo):

$$\frac{\partial \bar{E}}{\partial t} \equiv \bar{N} = (1-\bar{\rho})\bar{F}^{\rm s} - \bar{F}_{\rm TOA} \tag{11}$$

where $\bar{N}$ is the mean **radiative forcing**, i.e. the net irradiance at the TOA.

  - $\bar{F}^{\rm s}$ is the mean solar irradiance falling on the planet,
  - $(1-\bar{\rho})\bar{F}^{\rm s}$ is the fraction of the mean solar irradiance **absorbed** by the planet,
  - $\bar{\rho}\bar{F}^{\rm s}$ is the fraction of the mean solar irradiance **reflected** by the planet, and
  - $\bar{F}_{\rm TOA}$ is the **mean thermal IR emission** at the top of the atmosphere (TOA).

## Radiative Equilibrium and the Thermal Structure of Atmospheres (24)

- If we characterize the mean outgoing IR radiation by an effective temperature $T_e$, we obtain

$$\bar{F}_{\rm TOA} = \sigma_{\rm B} T_{\rm e}^4$$

  where $\sigma_{\rm B} = 5.67 \times 10^{-8}$ [W·m$^{-2}$·K$^{-4}$] is the Stefan-Boltzmann constant.

- By averaging over one or several years, we find a time- and space-averaged radiative forcing, which is close to zero, i.e. $\langle \bar{N} \rangle \approx 0$.

- When $\langle \bar{N} \rangle = 0$, we have **planetary radiative equilibrium**, which implies:

$$\bar{N} = (1 - \bar{\rho})\bar{F}^{\rm s} - \bar{F}_{\rm TOA} = (1 - \bar{\rho})\bar{F}^{\rm s} - \sigma_{\rm B} T_{\rm e}^4 = 0.$$

Planetary radiative equilibrium therefore means that the effective temperature of the planet is given by:

$$T_{\rm e} = \left[ \frac{(1 - \bar{\rho})\bar{F}^{\rm s}}{\sigma_{\rm B}} \right]^{1/4} . \tag{12}$$

# Radiative Equilibrium and the Thermal Structure of Atmospheres (24)

- The spherical albedo of Earth is $\bar{\rho} = 0.3$, and $\bar{F}^{\rm s}$ is one-quarter of the solar constant, *i.e.* $\bar{F}^{\rm s} = S_0/4 = 342$ [W·m$^{-2}$]. (The factor of 1/4 is the ratio of the total surface area to the disk area of the Earth intercepting the radiation.) Equation 12 yields $T_{\rm e} = 255$ K $= -18$°C.
- This effective temperature of $T_{\rm e} = -18$°C is considerably lower than the Earth's mean surface temperature $T_{\rm s} = +15$°C, due to the greenhouse effect.
- $T_{\rm s} > T_{\rm e}$ because the atmosphere is relatively transparent to visible solar radiation, but opaque to thermal IR radiation.
- A portion of the IR radiation emitted by the surface and the atmosphere is absorbed by atmospheric greenhouse gases, and emission of this absorbed radiation takes place in all directions.
- Some of the re-emitted radiation eventually escapes to space, but a significant fraction is emitted in the downward direction and contributes to the warming of the surface. Thus, the **greenhouse effect** is:
- The additional warming of the Earth by the downward IR "sky emission".
- **The surface warming due to absorption of diffuse sky and cloud radiation is almost twice the direct solar warming.**

# Climate Change: Radiative Forcing and Feedbacks (25)

- The ocean plays a major role in the energy exchange between the Sun and the Earth; we consider the combined atmosphere-ocean as the climate "system".
- To relate the radiative forcing to the surface temperature, we separate the **outgoing irradiance** into two terms: the **dominant term due to surface emission**, and a second term due to **emission from the atmosphere**.
- If we neglect the contribution from the atmosphere and take the surface to be black, we can approximate the outgoing irradiance by

$$F_{\mathrm{TOA}} \approx \mathcal{T}\sigma_{\mathrm{B}}T_{\mathrm{s}}^4$$

  where $\mathcal{T}$ is the diffuse transmittance, i.e. the fraction of the radiation from the surface that survives passage through the atmosphere.
- Correcting for the neglected atmospheric contribution, we may write

$$F_{\mathrm{TOA}}(T_{\mathrm{s}}) = \mathcal{T}_{\mathrm{eff}}\sigma_{\mathrm{B}}T_{\mathrm{s}}^4 \qquad (0 < \mathcal{T} < \mathcal{T}_{\mathrm{eff}} < 1), \tag{13}$$

  which shows that
  **the radiative forcing depends on surface temperature**.

## Climate Change: Radiative Forcing and Feedbacks (cont.) (26)

Consider a perturbation to radiative forcing: $N(T_s) \longrightarrow N(T_s) + \Delta N$. Assume:

1. perturbed climate system reaches a new equilibrium state;
2. change is small $\Longrightarrow$ total derivative of $N$ is the sum of the radiative forcing (the "cause") and the atmospheric response (the "effect"):

$$\overbrace{\Delta N}^{forcing} + \overbrace{\frac{\partial N}{\partial T_s}\Delta T_s}^{response} = 0 \quad \Longrightarrow \quad \Delta T_s^d = \alpha \Delta N; \quad \alpha \equiv -(\partial N/\partial T_s)^{-1}. \tag{14}$$

Use of the equation $N = (1-\bar{\rho})\bar{F}^s - \bar{F}_{\rm TOA}$ yields:

$$\alpha = \Big[\frac{\partial \bar{F}_{\rm TOA}}{\partial T_s} - \frac{\partial[(1-\bar{\rho})\bar{F}]}{\partial T_s}\Big]^{-1} \quad \Longleftarrow \textbf{climate sensitivity}. \tag{15}$$

- $\Delta T_s^d =$ **direct** temperature response: change in $T_s$ due to change $\Delta N$ in forcing.

- Note: $N =$ net irradiance at TOA: incident irradiance minus outgoing irradiance.

# Example 1: Climate Response to a $CO_2$ doubling (27)

- The radiative forcing from a doubling of $CO_2$ may be calculated from a detailed spectral radiative transfer model. Its value is $\Delta N \sim 4\ \mathrm{W}\cdot\mathrm{m}^{-2}$.
- The temperature response due solely to the (negative) feedback of the change of surface irradiance with temperature, or in other words, due to the change in the $F_{\mathrm{TOA}}$ term, we find by first computing the climate sensitivity:

$$\alpha = [\frac{\partial F_{\mathrm{TOA}}}{\partial T_{\mathrm{s}}}] = [\frac{\partial(\sigma_{\mathrm{B}} T_{\mathrm{s}}^4 \mathcal{T}_{\mathrm{eff}})}{\partial T_{\mathrm{s}}}]^{-1},$$

  which yields ($F_{\mathrm{TOA}}(T_{\mathrm{s}}) = \mathcal{T}_{\mathrm{eff}}\sigma_{\mathrm{B}} T_{\mathrm{s}}^4$):

$$\alpha = \left[4\sigma_{\mathrm{B}} T_{\mathrm{s}}^3 \mathcal{T}_{\mathrm{eff}}\right]^{-1} = \frac{T_{\mathrm{s}}}{4F_{\mathrm{TOA}}} = \frac{288}{(4\times 240)} = 0.3\ \mathrm{K}/(\mathrm{W}\cdot\mathrm{m}^{-2}).$$

- Thus, the direct response becomes:

$$\Delta T_s^d = \alpha \Delta N = 0.3\ \mathrm{K}/(\mathrm{W}\cdot\mathrm{m}^{-2}) \times 4\ (\mathrm{W}\cdot\mathrm{m}^{-2}) = 1.2\ \mathrm{K}.$$

## Example 2: Climate Response to a Change in the Solar Constant (28)

- If the radiation from the Sun were to decrease by 1%, and no feedback mechanisms were operative except the negative feedback due to reduced thermal emission, what is the change in surface temperature?
- Since we ignore the change in albedo, the climate sensitivity is given by:

$$\alpha = [\frac{\partial F_{\rm TOA}}{\partial T_{\rm s}}] = [\frac{\partial(\sigma_{\rm B} T_{\rm s}^4 \mathcal{T}_{\rm eff})}{\partial T_{\rm s}}]^{-1} = \frac{T_{\rm s}}{4F_{\rm TOA}}.$$

- From $N = (1-\bar{\rho})\bar{F}^{\rm s} - \bar{F}_{\rm TOA}$, the change in the radiative forcing becomes:

$$\Delta N = (1-\bar{\rho})\Delta F^{\rm s} \qquad \Longrightarrow \qquad \Delta T_{\rm s}^d = \alpha \Delta N = \alpha(1-\bar{\rho})\Delta F^{\rm s}.$$

Since the unperturbed system is in thermal equilibrium, $F_{\rm TOA} = (1-\bar{\rho})F^{\rm s}$:

$$\Delta T_{\rm s}^d = \alpha(1-\bar{\rho})\Delta F^{\rm s} = \frac{T_{\rm s}}{4F_{\rm TOA}}(1-\bar{\rho})\Delta F^{\rm s} = (\frac{T_{\rm s}}{4})\frac{\Delta F^{\rm s}}{F^{\rm s}} = \frac{288}{4}(-0.01) = -0.72 \text{ K}.$$

## Climate Change: Radiative Forcing and Feedbacks (cont.) (29)

- Let us now consider the indirect effects on the surface temperature caused by temperature-dependent processes.
- A **positive feedback** tends to **amplify** the temperature response. For example, increased temperature tends to increase evaporation and thus raise the humidity. Since water vapor is a greenhouse gas, increased IR opacity causes a further increase in temperature.
- In contrast, a **negative feedback** tends to **dampen** the temperature response. For example, an increase in low-level cloudiness could result from a warmer climate, and more clouds cause a higher albedo that tends to cool the Earth.
- To illustrate this concept we now consider a parameter $Q$ (for example albedo) that depends upon the surface temperature. The direct radiative forcing is augmented by an additional term ($\bar{N} = (1 - \bar{\rho})\bar{F}^{\rm S} - F_{\rm TOA} = 0$):

$$(\frac{\partial N}{\partial Q})(\frac{\partial Q}{\partial T_{\rm s}})\Delta T_{\rm s},$$

so that the total change in the radiative forcing becomes ($\Delta N$ is the original (direct) radiative forcing):

$$\Delta N + (\frac{\partial N}{\partial Q})(\frac{\partial Q}{\partial T_{\rm s}})\Delta T_{\rm s} = (\frac{\partial F_{\rm TOA}}{\partial T_{\rm s}})\Delta T_{\rm s}.$$

## Climate Change: Radiative Forcing and Feedbacks (cont.) (30)

- If $N$ depends on several parameters $Q_1$, $Q_2$,$\cdots$, each of which depends on $T_{\rm s}$, then the total change in the radiative forcing becomes:

$$\Delta N + \Big(\frac{\partial N}{\partial Q_1}\frac{\partial Q_1}{\partial T_{\rm s}} + \frac{\partial N}{\partial Q_2}\frac{\partial Q_2}{\partial T_{\rm s}} + \cdots\Big)\Delta T_{\rm s}.$$

- Per definition this change is equal to the change in the outgoing irradiance at the top of the atmosphere, i.e.

$$\Delta N + (\frac{\partial N}{\partial Q_1}\frac{\partial Q_1}{\partial T_{\rm s}} + \frac{\partial N}{\partial Q_2}\frac{\partial Q_2}{\partial T_{\rm s}} + \cdots)\Delta T_{\rm s} = \frac{\partial F_{\rm TOA}}{\partial T_{\rm s}}\Delta T_{\rm s} \qquad \Longrightarrow \qquad \Delta T_{\rm s} = \alpha \Delta N$$

$$\alpha = [\frac{\partial F_{\rm TOA}}{\partial T_{\rm s}} - \sum_i \frac{\partial N}{\partial Q_i}\,\frac{\partial Q_i}{\partial T_{\rm s}}]^{-1} \qquad \Longleftarrow \textbf{climate sensitivity.} \qquad (16)$$

From $F_{\rm TOA} = \mathcal{T}_{\rm eff}\sigma_{\rm B}T_{\rm s}^4$, we obtain: $\frac{\partial F_{\rm TOA}}{\partial T_{\rm s}} = \frac{4F_{\rm TOA}}{T_{\rm s}}$, so that:

$$\alpha = [\frac{4F_{\rm TOA}}{T_{\rm s}} - \sum_i \frac{\partial N}{\partial Q_i}\,\frac{\partial Q_i}{\partial T_{\rm s}}]^{-1}, \qquad (17)$$

$$\Delta T_{\rm s} = \frac{\Delta N}{\frac{4F_{\rm TOA}}{T_{\rm s}}(1 - \Sigma_i\,\lambda_i)}; \qquad \lambda_i = \frac{T_{\rm s}}{4F_{\rm TOA}}\,\frac{\partial N}{\partial Q_i}\,\frac{\partial Q_i}{\partial T_{\rm s}}.$$

Since the factor $\Delta N/(4F_{\rm TOA}/T_{\rm s}) = \Delta N/\frac{\partial F_{\rm TOA}}{\partial T_{\rm s}} = \Delta T_{\rm s}^{d}$, we may write:

$$\Delta T_{\rm s} = \alpha \Delta N = \frac{1}{(1-\Sigma_i\,\lambda_i)}\underbrace{[\frac{T_{\rm s}\,\Delta N}{4F_{\rm TOA}}]}_{\Delta T_{\rm s}^{d}} = f\Delta T_{\rm s}^{d} \tag{18}$$

where the *gain* of the climate system is:

$$f = [1-\sum_i \lambda_i]^{-1} \text{ where } \lambda_i \equiv \frac{T_{\rm s}}{4F_{\rm TOA}}\,\frac{\partial N}{\partial Q_i}\,\frac{\partial Q_i}{\partial T_{\rm s}}. \tag{19}$$

$f = 1$ in the absence of feedbacks, $f > 1$ when the net feedbacks are positive ($\Sigma\,\lambda_i > 0$), and $f < 1$ if the net feedbacks are negative ($\Sigma\,\lambda_i < 0$).

- Equation (18) tells us that the temperature response is independent of the nature of the forcing mechanism, which means that two different greenhouse gases have the same effect on climate if their individual forcings, $\Delta N_i$ are the same.

# Climate Change: Radiative Forcing and Feedbacks (cont.) (32)

- There are two important exceptions from this rule:
  - The radiative forcing caused by stratospheric ozone has an altitude dependence that is different from that of well-mixed tropospheric gases.
  - Regionally concentrated sulfate aerosols act in the global average in a different way than the well-mixed long-lived greenhouse gases.
- Figure 8a shows the radiative forcing caused by several greenhouse gases for different time epochs.
- Figure 8b shows estimates of the globally averaged radiative forcing due to changes in greenhouse gases and aerosols from pre–industrial times to the present, and changes in solar variability from 1850 to the present. A significant indirect effect is due to reduction in stratospheric ozone.
- The oceans play a very important role for the time-dependent response, since they absorb solar radiation and emit thermal IR radiation. It is also important to note that whereas the time scale for changes in the atmosphere are of the order one year, the time scale in the oceans is many years.

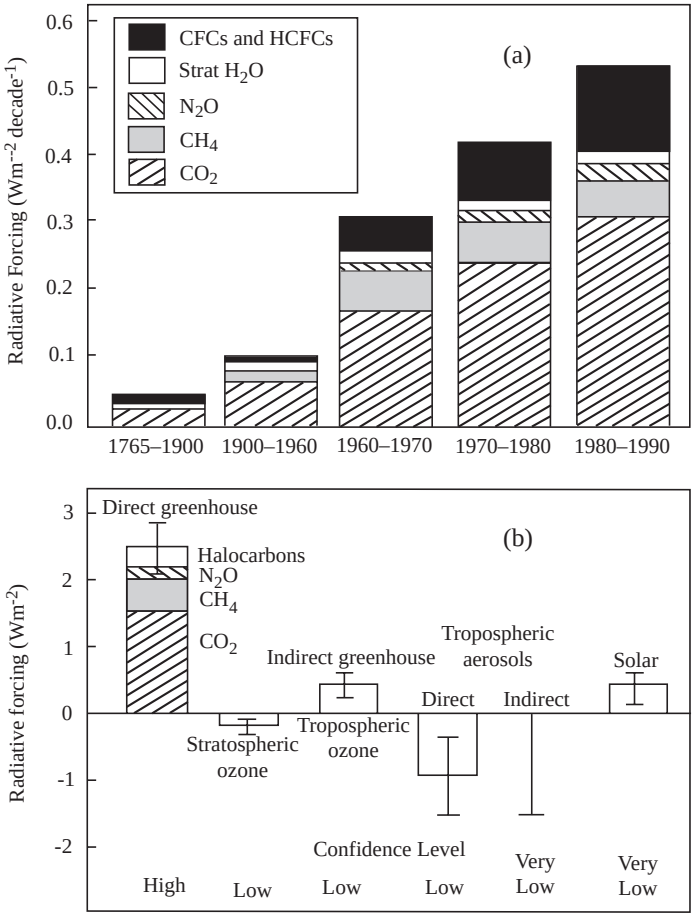


Figure 8: **(a) The degree of radiative forcing produced by selected greenhouse gases in five different epochs. Until about 1960, nearly all the forcing was due to $CO_2$; today the other greenhouse gases combined nearly equal the $CO_2$ forcing. (b) Estimates of the globally averaged radiative forcing due to changes in greenhouse gases and aerosols from pre-industrial times to 1990 and changes in solar variability from 1850 to 1990. The height of the bar indicates a mid-range estimate of the forcing. The lines show the possible range of values. An indication of relative confidence levels in the estimates is given below each bar.**

# Density Structure of the Ocean (33)

- The hydrostatic equation is the same in an aquatic medium like the ocean as in a gaseous medium like the atmosphere, but the density $\rho$ is *not* proportional to $p/T$. The pressure at depth $h$ in the ocean is given by:

$$p(h) = p_a + \int_0^h dh' \rho(h') g \approx p_a + \bar{\rho} g h \tag{20}$$

where $p_a$ is the atmospheric pressure at the ocean surface, $\bar{\rho}$ is the depth-averaged value of the density, and $\bar{\rho} g h$ is the weight of the overlying column of water at depth $h$.

**Integration Along a Slant Path**

- If the line of sight makes an angle $\theta$ with the upward vertical direction (see Fig. 6), we have:

$$\mathcal{M}(1,2) = \bar{\rho} h_2 \sec\theta - \bar{\rho} h_1 \sec\theta. \tag{21}$$

  - Note that this result is the same as for the atmosphere, except that the scale height is replaced with the ocean depth $h$.
  - Knowledge of the slant-path water mass is usually sufficient for shortwave radiative transfer in very pure waters, in which only molecular (Rayleigh) scattering needs to be considered. The reason is that water molecules scatter light in much the same way as air molecules as we shall see in Chapter 3.
  - However, in most cases, water impurities play a crucial role for absorption and scattering of light. These impurities consist of dissolved organic substances, suspended mineral particles, and organic matter.
  - Ocean radiative energy budget is important because of the intimate coupling of the ocean (covering nearly three fourths of the total world surface area) and the atmosphere.

# The Mixed Layer and the Deep Ocean (34)

- The ocean's vertical structure can be crudely subdivided into two regions:
  - An **upper mixed layer**, typically 50 to 200 m in depth, and
  - The **deep ocean** having an average depth of 4 km.
- The mixed layer has a nearly uniform temperature and salinity. It is maintained by turbulent transport caused by mechanical stirring by wind stress.
- The transition from the upper mixed layer to the colder, denser bottom water is marked by an abrupt *decrease in temperature*, the **thermocline**, and an abrupt *increase in salinity*, the **pycnocline**.
- The ocean receives fresh (less dense) water through precipitation, and melting of sea ice. On the other hand, evaporation tends to leave more saline water behind.
- Figure 9 shows depth profiles of temperature at low, mid and high latitudes. The temperature generally decreases with depth, while the density increases.
- In the polar oceans, and at lower latitudes during winter, the thermocline and the pycnocline may disappear, so that the water temperature and salinity becomes nearly constant with depth.
- The ocean's energy source is the absorption of sunlight (direct and diffuse) within a shallow layer 1 to 10 m thick. The supply of solar energy depends on season.
- Figure 10 shows the annual variation of the diurnally averaged mixed-layer temperature at a northern hemisphere site (50 °N, 145 °W in the eastern North Pacific). The increase in temperature between March and August is due to solar heating that creates a lighter and hence more buoyant upper layer, which is resistant to mixing. The shallower the mixed layer, the higher its temperature.

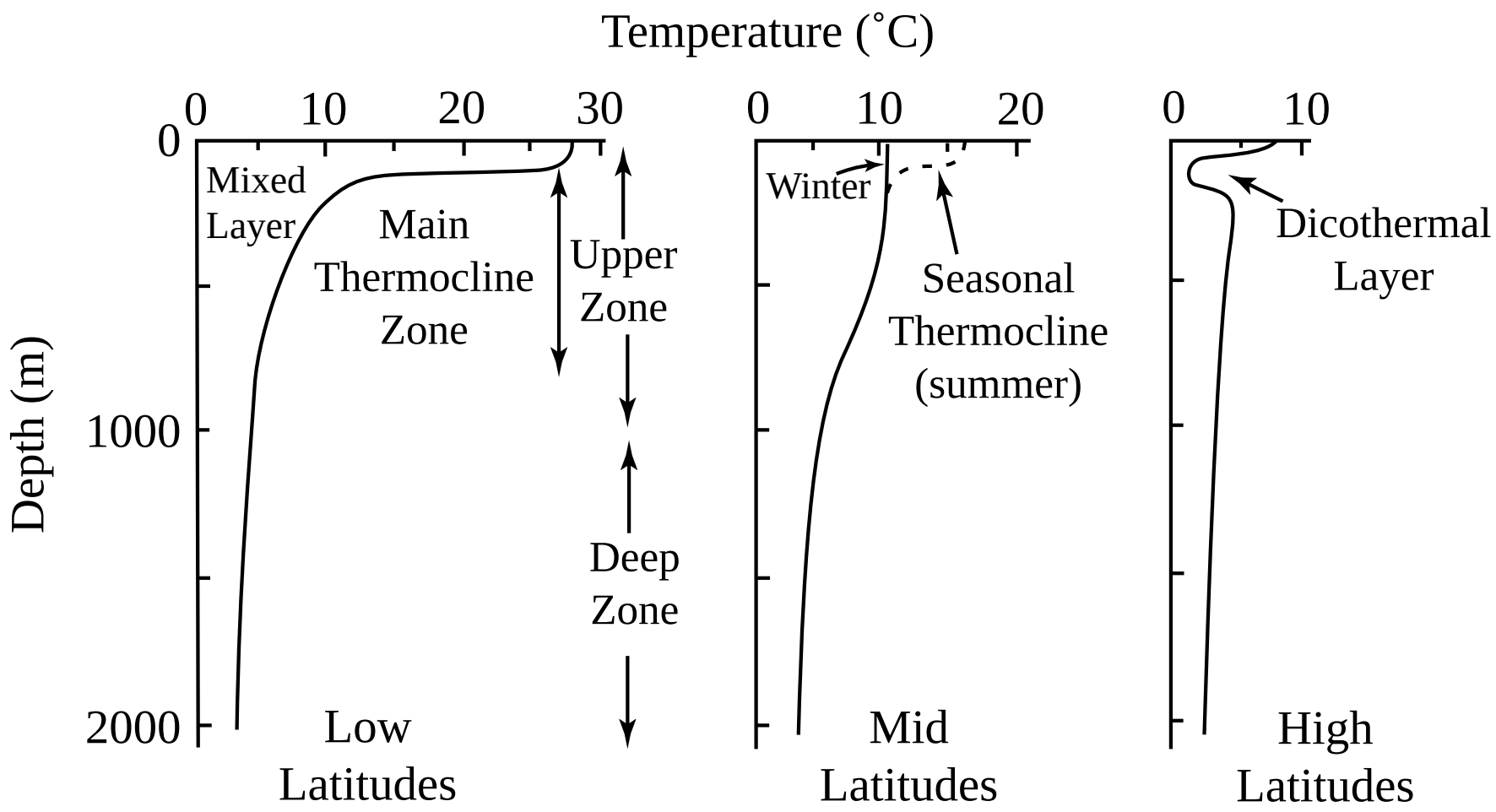


Figure 9: **Typical mean temperature/depth profiles for the open ocean. The *dicothermal layer* refers to a layer of cold water that often occurs in northern high latitudes between 50 and 100 m. Stability in this layer is maintained by an increase in salinity with depth.**

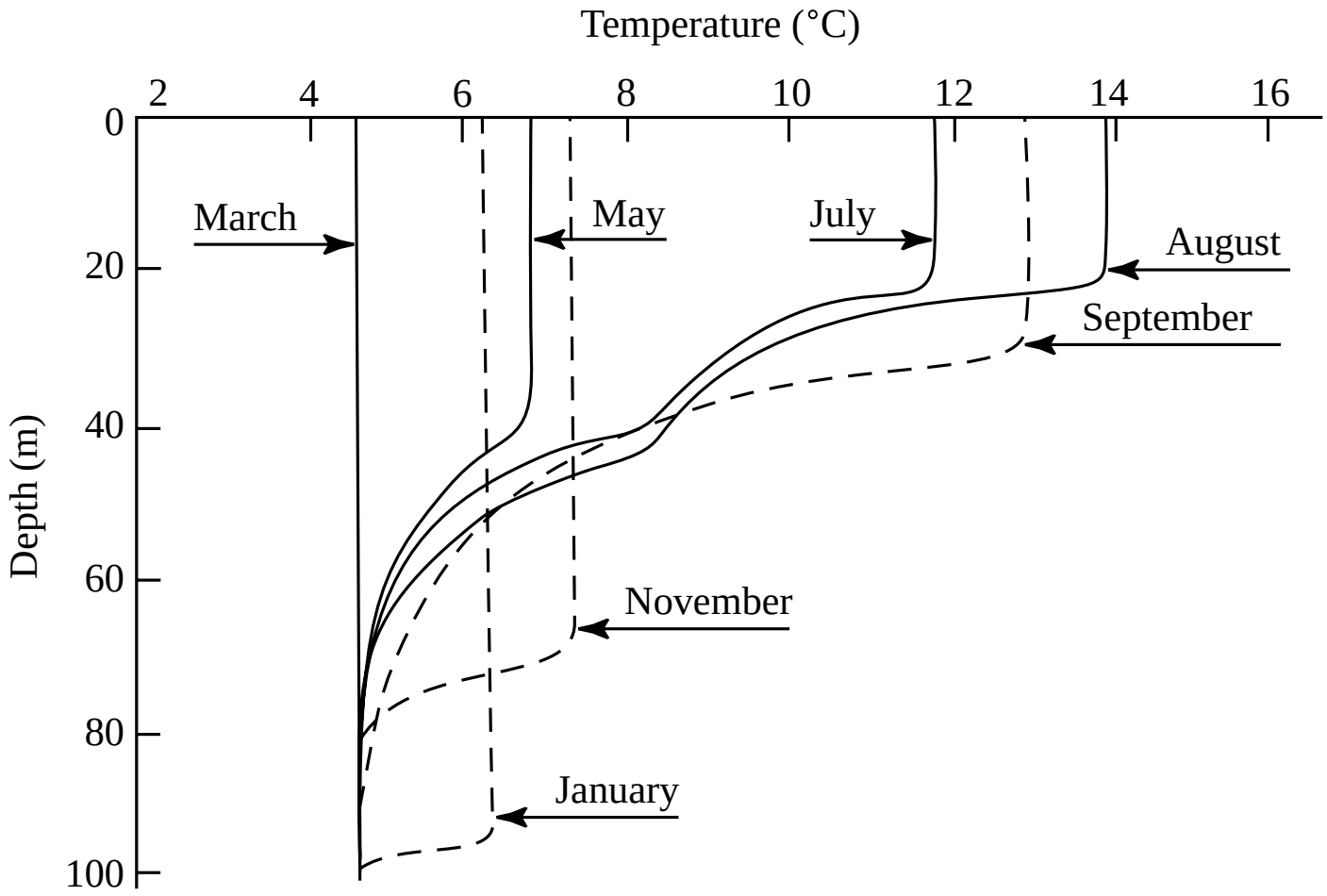


Figure 10: **Growth and decay of the seasonal thermocline at 50°N, 145°W in the eastern North Pacific.**

# Seasonal Variations of Optical Properties (35)

- The depth of the mixed layer varies with season, and with the degree of mechanical stirring induced by winds. For example, a storm of several-days duration can lower the thermocline by 30-50 m, while cooling the surface by 1-2 K. Figure 10 shows that the mixed layer is deepest when the solar heating is a minimum during winter. Then the ocean emits more thermal infrared energy than it receives from the Sun, and the deepening of the mixed layer occurs because the colder, denser (and thus less stable) water can be mixed to greater depths than during summer.
- The mixed layers' annual temperature cycle, although driven by solar heating at the surface, is controlled more by turbulent heat transport than by radiative transfer.
- The deep ocean shows virtually no seasonal variation. Its temperature decreases slowly with depth, approaching 0°C at **abyssal** depths (4,000–6,000 m).
- Downwelling occurs primarily in the polar regions. The return of colder water to the surface by upwelling occurs over a much greater fraction of the world ocean than that of downwelling. The time scale of this overturning process is of the order of several centuries.

# Sea-Surface Temperature (36)

- Water is very opaque in the IR, implying that downward transport of IR radiation through water is negligible. The surface sheds its energy upward by radiation and by exchange of latent heat with the overlying atmosphere via evaporation and precipitation. If the water is calm, the daytime "skin temperature" may be appreciably different from that of water just below the surface, because even a thin layer of water takes up a considerable amount of energy.
- The sea-surface temperature measured by infrared sensors always refers to this skin layer.
- Because the ocean has a high IR absorptance, it must in accordance with Kirchhoff's law also be an efficient emitter of thermal radiation.
- Measurements show that as the ocean temperature increases, the **net cooling** actually **decreases**, due to the strong dependence of atmospheric radiation on atmospheric water vapor content.
- The **excess of evaporation over condensation** associated with the warmer ocean temperature, leads to **higher absolute humidity**, and
- the **increased warming of the ocean from downwelling atmospheric radiation more than compensates for the increased emission from the surface**.

# Ocean Spectral Reflectance and Opacity (37)

- At **normal incidence** the reflectance of pure water is $(\frac{m-1}{m+1})^2 = (\frac{0.33}{2.33})^2 = 0.02$[¶]: The ocean absorbs 98% of the incoming solar radiation. But, ocean particles that scatter radiation and a solar elevation that varies with time of the day as well as season, lead to a visible reflectance that is closer to 7%.
- Figure 11 shows that the ocean is quite transparent in the visible. Thus, incoming radiative energy can reach up to 100 m depth. This transparency "window" between 0.4 and 0.6 $\mu$m, which coincides with a corresponding atmospheric window, lies at the peak in the solar spectrum.
- Water is opaque in the UV: Before a UV-protective ozone layer formed (about one billion years ago), life could exist in the ocean, protected from damaging UV radiation, but bathed in visible radiation required for photosynthesis.
- The ocean color ranges from deep blue to green or greenish-yellow. A deep indigo color is characteristic of tropical waters, where biology is scarce.
- At higher latitudes the color changes from **green-blue to green** in the polar regions.



[¶]Here $m$ is the refractive index assumed to be 1.0 for air and 1.33 for water.

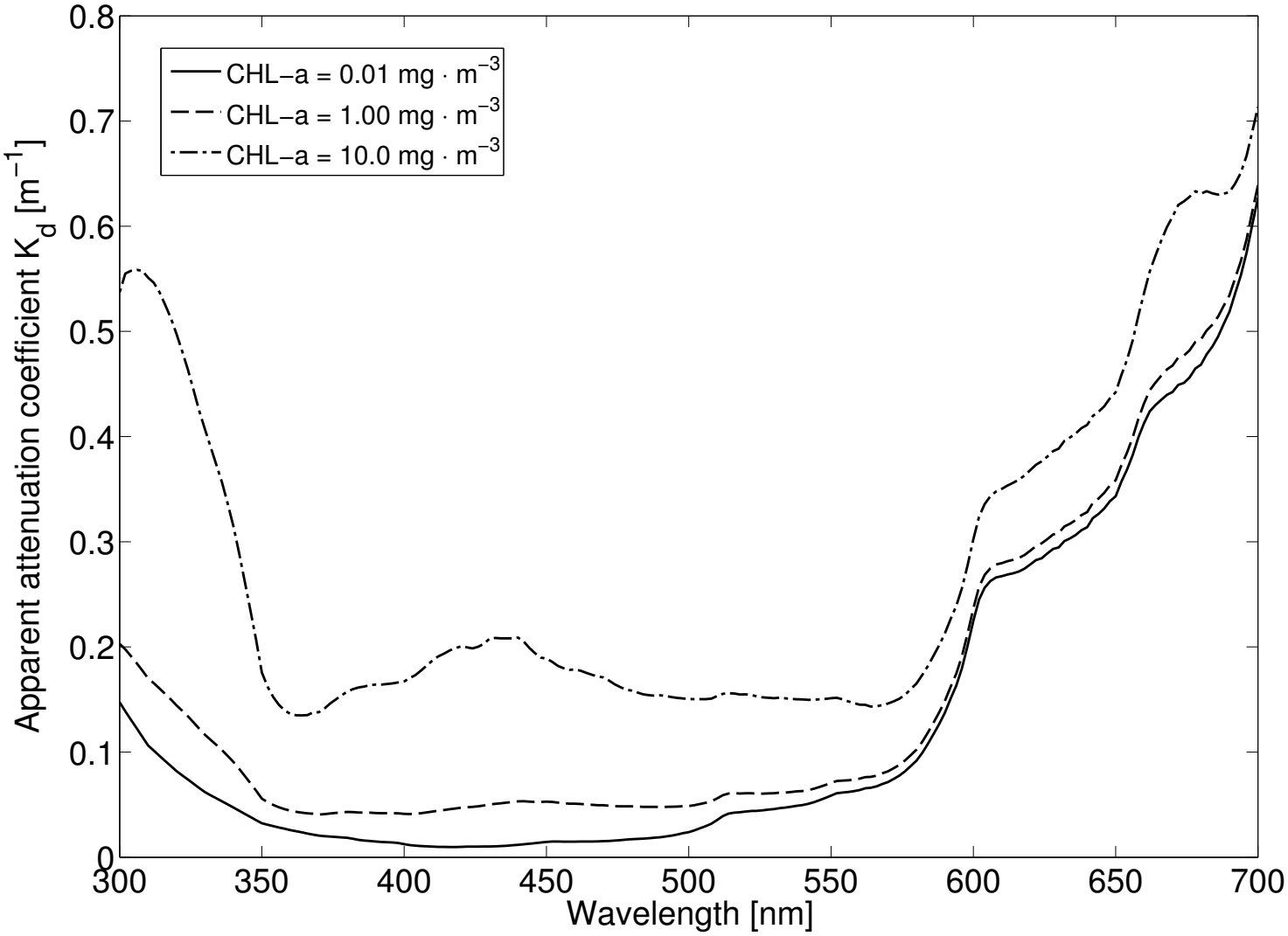


Figure 11: **Spectral dependence of the apparent attenuation coefficient $K_d$ (see Eq. 10.3) of the ocean for three different chlorophyll concentrations: Chl-$a$ = 0.1 $\mathrm{mg}\cdot\mathrm{m}^{-3}$, Chl-$a$ = 1.0 $\mathrm{mg}\cdot\mathrm{m}^{-3}$, and Chl-$a$ = 10.0 $\mathrm{mg}\cdot\mathrm{m}^{-3}$. The computations were done with the AccuRT radiative transfer tool described in §10.5.**

# Ocean Spectral Reflectance and Opacity (cont.) (38)

- The blue color of open-ocean waters containing very little particulate matter is due to both absorption and scattering processes. At a few meters depth, the red and yellow light will be selectively removed by absorption (Fig. 11). Thus, only blue light remains to be scattered by the water molecules.
- This scattering process is proportional to the inverse fourth power of the light wavelength, and thus favors light at the shorter (blue) wavelengths. The short-wavelength light that is scattered in the upward direction and eventually leaves the ocean surface gives a deep-blue color to the ocean.
- In **more fertile waters, green** phytoplankton and yellow substance will absorb blue light, and the color will shift toward the **green**. Certain colors, such as the "red tide" are characteristic of particular phytoplankton blooms.
- Satellite remote sensing of ocean color is a valuable enterprise to fisheries and for making surveys of ocean fertility on a global scale.
- The upward radiance at the top of the atmosphere (TOA) is used to infer abundances of optically significant constituents in the water by remote sensing.

# Ocean Spectral Reflectance and Opacity (cont.) (30)

Figure 12 shows a simulation of the upward (nadir) radiance just below the ocean surface (Subsurface) as well as at the TOA.

- The gray curve pertains to a chlorophyll concentration of Chl-$a = 0.1$ mg $\cdot$ m$^{-3}$, and the black curve to Chl-$a = 10$ mg $\cdot$ m$^{-3}$.
- Note the significant change in Subsurface color with increasing chlorophyll concentration. But:
- **there is only a slight change at the TOA, showing that TOA spectra are dominated by light from atmospheric scattering.**
- A radiative transfer model for a coupled atmosphere-ocean system was used to carry out the simulations shown in Fig. 12 (see §10.5).

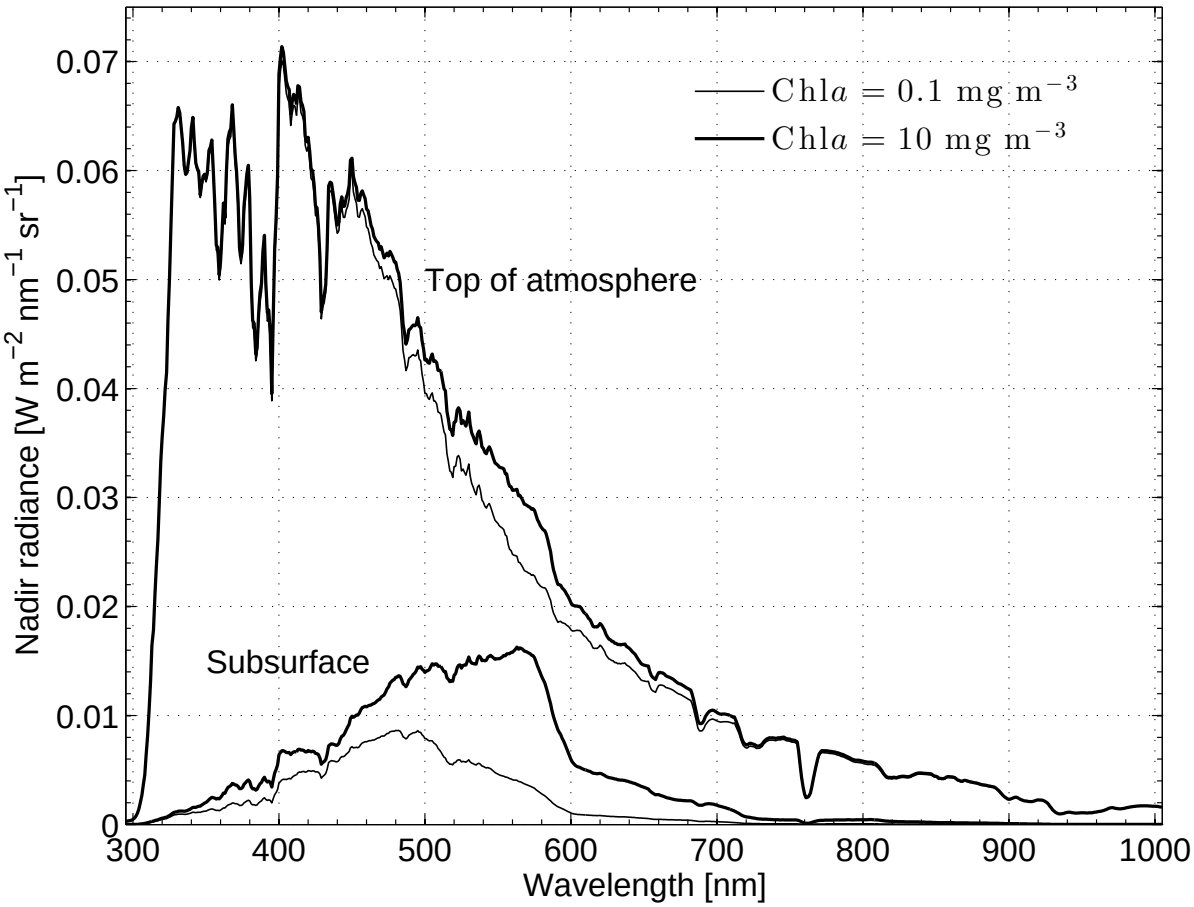


Figure 12: Ocean color dependence on biological activity. Gray curve: Chl-$a$ = 0.1 mg $\cdot$ m$^{-3}$. Black curve: Chl-$a$ = 10 mg $\cdot$ m$^{-3}$. Solar zenith angle: 45°. Spectral resolution: 5 nm. The computations were done with the AccuRT radiative transfer tool described in §10.5.