Lecture 7.

Multiple scattering as a source of radiation. Reflectance from surfaces. Remote sensing of ocean color.

1. Direct and diffuse radiation. Multiple scattering. Single scattering approximation.
2. Reflectance from surfaces.
3. Applications of passive remote sensing using extinction and scattering: Remote sensing of ocean color.

Required Reading:

S: 6.1, 6.3, 6.4, Petty: 11, Into to Chapter 13, 13.1

Additional reading

Remote sensing of ocean color: http://oceancolor.gsfc.nasa.gov/

Advanced reading

Yoder, J.A. and M.A. Kennelly. 2006. What have we learned about ocean variability from satellite ocean color imagers? *Oceanography,* 19(1), 152-171.

1. Direct and diffuse radiation. Multiple scattering.

The solar radiation field is traditionally considered as a sum of two distinctly different components: **direct** and **diffuse:** $I = I_{dir} + I_{dif}$

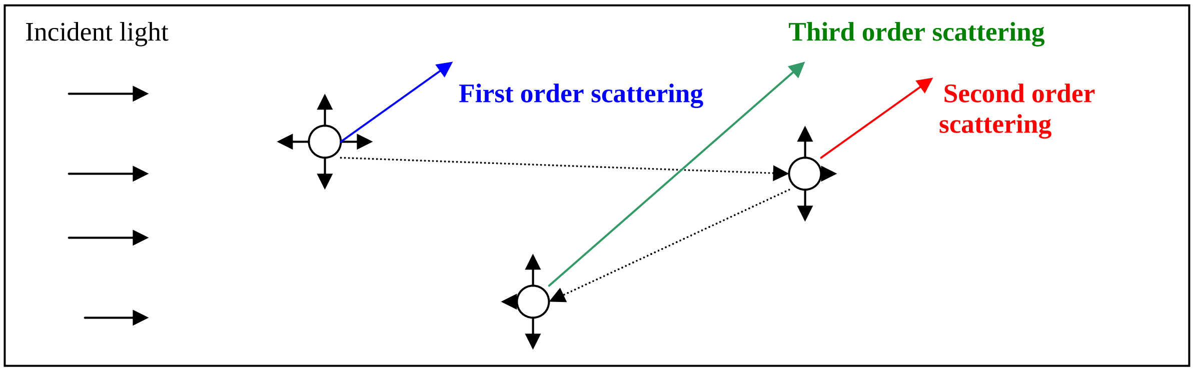
Direct radiation is a part of the radiation field that has survived the extinction passing through a layer with optical depth τ and it obeys the Beer-Bouguer-Lambert law (see Lecture 6):

$$I_{dir}^{\downarrow} = I_0 \exp(-\tau / \mu_0)$$

where $\boldsymbol{I_0}$ is the incident intensity at a given wavelength at the top of a layer and $\boldsymbol{\mu_0}$ is a cosine of the incident zenith angle $\boldsymbol{\theta_0}$ ($\boldsymbol{\mu_0} = \boldsymbol{cos}(\boldsymbol{\theta_0})$).

Diffuse radiation

Diffuse radiation arises from the light that undergoes one scattering event (**single scattering**) or many (**multiple scattering**).



For single scattering

$$\delta I_{\lambda}(\vec{\Omega}) = k_{s,\lambda} ds \frac{P_{\lambda}(\vec{\Omega},\vec{\Omega}')}{4\pi} I_{\lambda}(\vec{\Omega}')\delta\Omega' \qquad [7.1]$$

where $I_{\lambda}(\vec{\Omega}')$ is the incident intensity in the direction defined by a solid angle $\vec{\Omega}'(\mu',\varphi')$.

For **multiple scattering,** integrating over all directions:

$$dI_{\lambda}(\vec{\Omega}) = k_{s,\lambda} ds \int_{4\pi} \frac{P_{\lambda}(\vec{\Omega},\vec{\Omega}')}{4\pi} I_{\lambda}(\vec{\Omega}')d\Omega' \qquad [7.2]$$

NOTE: The above equation shows that the phase function redirects the incident intensity in the direction $\vec{\Omega}'(\mu',\varphi')$ to the direction $\vec{\Omega}(\mu,\varphi)$, and the integral accounts for all possible scattering events within the 4π solid angle.

According to the Beer-Bouguer-Lambert law, scattering radiance from path ***ds*** is

$$dI_{\lambda} = k_{e,\lambda} J_{\lambda} ds \qquad \text{(see Eq.[6.26b])}$$

thus (from Eq.[7.2]) **the scattering source function** is

$$J_{\lambda}(\vec{\Omega}) = \frac{\omega_{0,\lambda}}{4\pi} \int_{\vec{\Omega}'} I_{\lambda}(\vec{\Omega}') P_{\lambda}(\vec{\Omega},\vec{\Omega}') d\Omega' \qquad [7.3]$$

where $\omega_{0,\lambda} = \kappa_{s,\lambda} / \kappa_{e,\lambda}$ is the single scattering albedo.

The scattering source function:

1) has units of intensity

2) plays the role of the Planck function in thermal radiative transfer but the scattering source function is more complex

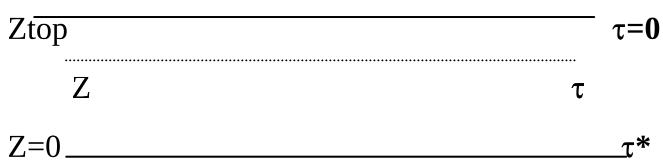
3) depends on the radiation intensity in the incident direction, $I_{\lambda}(\vec{\Omega}')$; fraction of radiation which is scattered, $\omega_{0,\lambda}$; and fraction scattered into the new direction $\frac{P(\vec{\Omega},\vec{\Omega}')}{4\pi}d\Omega'$

The monochromatic radiative transfer equation for a plane-parallel atmosphere:

expresses the net change in intensity due to extinction and scattering along path ***dz:***

dI = dI (extinction) + dI(scattering)

A plane-parallel atmosphere



Using ***ds = dz/cos(θ)***, the **radiative transfer equation** can be written as

$$\cos(\theta)\frac{dI_{\lambda}(z,\theta,\varphi)}{k_{e,\lambda}dz} = -I_{\lambda}(z,\theta,\varphi) + J_{\lambda}(z,\theta,\varphi) \qquad [7.4]$$

Introducing the optical depth measured from the outer boundary downward as

$$\tau_{\lambda}(z,0) = \int_{0}^{z} k_{e,\lambda}(z)dz \qquad [7.5]$$

and using $d\tau_{\lambda} = -k_{e,\lambda}(z)dz$ and ***μ = cos(θ),*** we have

$$\mu\frac{dI_{\lambda}(\tau;\mu;\varphi)}{d\tau} = I_{\lambda}(\tau;\mu,\varphi) - J_{\lambda}(\tau,\mu,\varphi) \qquad [7.6]$$

NOTE: Eq.[7.6] is called the Schwarzchild’s equation. This is a basic equation for the radiative transfer in a plane-parallel atmosphere.

Upward (or upwelling) intensity $I_\lambda^\uparrow$ is for $1 \geq \mu \geq 0$ (or $0 \leq \theta \leq \pi/2$);

Downward (or downwelling) intensity $I_\lambda^\downarrow$ is for $-1 \leq \mu \leq 0$ (or $\pi/2 \leq \theta \leq \pi$)

The **radiative transfer equation** [7.6] can be written for **upward and downward intensities:**

$$\mu \frac{dI_\lambda^\uparrow(\tau,\mu,\varphi)}{d\tau} = I_\lambda^\uparrow(\tau,\mu,\varphi) - J_\lambda^\uparrow(\tau,\mu,\varphi) \qquad [7.7a]$$

$$-\mu \frac{dI_\lambda^\downarrow(\tau,-\mu,\varphi)}{d\tau} = I_\lambda^\downarrow(\tau,-\mu,\varphi) - J_\lambda^\downarrow(\tau,-\mu,\varphi) \qquad [7.7b]$$

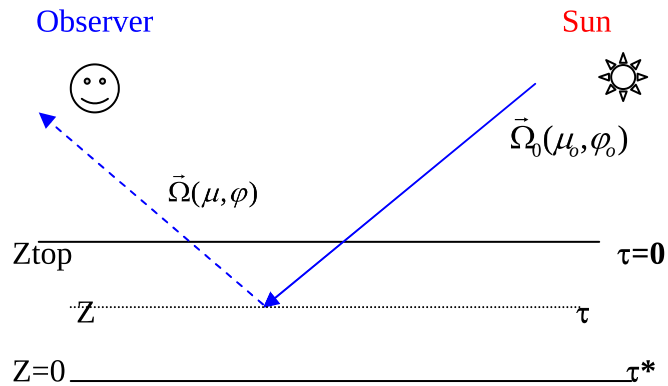
Solution of Eq.[7.7a] gives the upward intensity in the plane-parallel atmosphere:

$$I_\lambda^\uparrow(\tau,\mu,\varphi) = I_\lambda^\uparrow(\tau^*,\mu,\varphi)\exp\left(-\frac{\tau^* - \tau}{\mu}\right) + \frac{1}{\mu}\int_\tau^{\tau^*} \exp\left(-\frac{\tau' - \tau}{\mu}\right) J_\lambda^\uparrow(\tau',\mu,\varphi)\, d\tau' \qquad [7.8a]$$

Solution of Eq.[7.7b] gives the downward intensity in the plane-parallel atmosphere:

$$I_\lambda^\downarrow(\tau,-\mu,\varphi) = I_\lambda^\downarrow(0,-\mu,\varphi)\exp\left(-\frac{\tau}{\mu}\right) + \frac{1}{\mu}\int_0^{\tau} \exp\left(-\frac{\tau - \tau'}{\mu}\right) J_\lambda^\downarrow(\tau',-\mu,\varphi)\, d\tau' \qquad [7.8b]$$

- **<u>First-order scattering:</u>**



Direct solar radiation reaching the altitude z is

$$F^{\downarrow}(z) = F_0 \exp(-k_e(z_t - z)/\mu_0) = F_o \exp(-\tau/\mu_0)$$

where $\boldsymbol{F_0}$ is the solar constant at the top of the atmosphere.

Scattering of the direct beam is the source of diffuse radiation (see Eq.[7.1])

$$J(\tau,\mu,\varphi) = \frac{\omega_0}{4\pi} F_0 \exp(-\tau/\mu_0) P(\mu,\varphi,\mu_0,\varphi_0) \qquad [7.9]$$

Assuming no surface reflection (dark surface), the upwelling intensity at the level Z (or τ) can be found from Eq.[7.8a] as

$$I^{\uparrow}(\tau,\mu,\varphi) = \int_{\tau}^{\tau^*} J(\tau',\mu,\varphi) \exp[-(\tau'-\tau)/\mu] d\tau'/\mu \qquad [7.10]$$

Substituting in the source function

$$I^{\uparrow}(\tau,\mu,\varphi) = \frac{\omega_0}{4\pi} F_0 P(\Theta) \int_{\tau}^{\tau^*} \exp[-(\tau'-\tau)/\mu - \tau'/\mu_0] d\tau'/\mu \qquad [7.11]$$

An observer (i.e., a satellite detector) at Ztop (or τ= 0) measures

$$I^{\uparrow}(0,\mu,\varphi) = \frac{\omega_0}{4\pi} F_0 P(\Theta) \frac{\mu_0}{\mu+\mu_0} [1 - \exp(-(\frac{1}{\mu_o} + \frac{1}{\mu})\tau^*)] \qquad [7.12]$$

If **τ* <1 (called the single scattering approximation), Eq.[7.12] simplifies to**

$$\boxed{I^{\uparrow}(0,\mu,\varphi) = \frac{\omega_0}{4\pi} F_0 P(\Theta) \frac{\tau^*}{\mu}} \qquad [7.13]$$

NOTE: Eq.[7.13] provides the basic for the retrieval of aerosol optical depth from a satellite sensor. For instance, the AVHRR aerosol retrieval algorithm is based on the single-scattering approximation (see S: 6.4.2)

Radiative transfer equation with multiple scattering:

In the general case of multiple scattering

$$J(\tau,\mu,\varphi) = \frac{\omega_0}{4\pi} \int_0^{2\pi} \int_{-1}^{1} I(\tau,\mu',\varphi') P(\mu,\varphi,\mu',\varphi') d\mu' d\varphi' + \\ + \frac{\omega_0}{4\pi} F_0 P(\mu,\varphi,\mu_0,\varphi_0) \exp(-\tau/\mu_0) \qquad [7.14]$$

Using the source function for scattering, we can write the **radiative transfer equation for diffuse radiation** as an integro-differential equation:

$$\mu \frac{dI(\tau,\vec{\Omega})}{d\tau} = I(\tau,\vec{\Omega}) - \frac{\varpi_0}{4\pi}\int_{4\pi} I(\tau,\vec{\Omega})P(\vec{\Omega},\vec{\Omega}')d\Omega' - \frac{\omega_0}{4\pi}F_0 P(\vec{\Omega},\vec{\Omega}_0)\exp(-\tau/\mu_0) \qquad [7.15]$$

NOTE: To solve Eq.[7.15], one needs to know the scattering coefficient $\mathbf{k_{s,\lambda}}$, absorption coefficient $\mathbf{k_{a,\lambda}}$ and scattering phase function ***P*(μ, φ, μ',φ')** as a function of wavelength in each atmospheric layer.

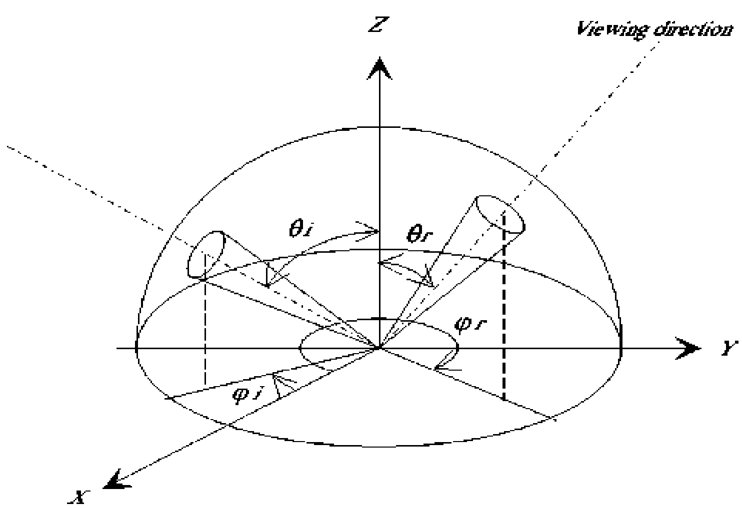
NOTE: Various approximate and "exact" (such as Discrete-ordinate, Adding-doubling, Monte-carlo, etc.) techniques have been developed to solve the radiative transfer eqaution for diffuse radiation. Each technique requires a sophisticated numerical code. There are a number of various munerical radiative trasnfer codes that are available to the scientific community.

2. Reflection from surfaces.

Bi-directional reflectance distribution function (BRDF) is introduced to characterize the angular dependence in the surface reflection and defined as the ratio of the reflected intensity (radiance) to the radiation flux (irradiance) in the incident beam:

$$R(\mu_r,\varphi_r,\mu_i,\varphi_i) = \frac{\pi dI^{\uparrow}(\mu_r,\varphi_r)}{I^{\downarrow}(\mu_i,\varphi_i)\mu_i d\Omega_i} \qquad [7.16]$$

where $\mu_i = \cos(\theta_i)$ and θ_i is the incident zenith angle, φ_i is the incident azimuthal angle, and $\mu_r = \cos(\theta_r)$ and θ_r is the viewing zenith angle, φ_r is the viewing azimuthal angle.



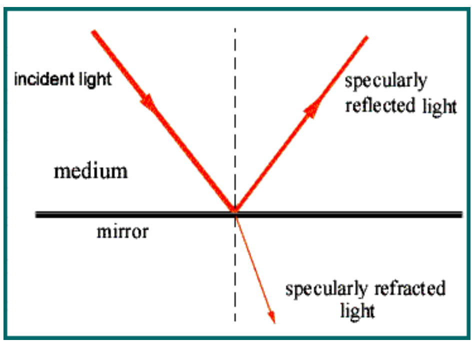
Two extreme types of the surface reflection:

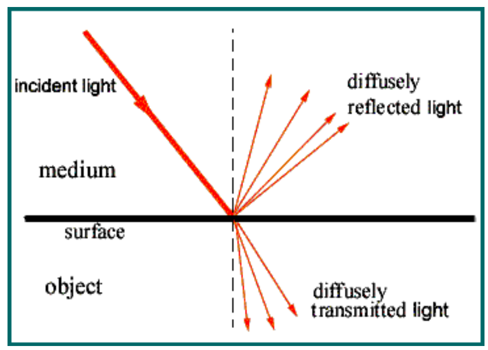
specular reflectance and **diffuse reflectance**.

Specular reflectance is the reflectance from a perfectly smooth surface (e.g., a mirror):

Angle of incidence =Angle of reflectance

- Reflection is generally **specular** when the "roughness" of the surface is smaller than the wavelength used. In the solar spectrum (about 0.4 to 2 μm), reflection is therefore specular on smooth surfaces such as polished metal, still water or mirrors.
- Practically all real surfaces are not smooth and the surface reflection depends on the incident angle and the angle of reflection. Reflectance from such surfaces is referred to as **diffuse reflectance**.

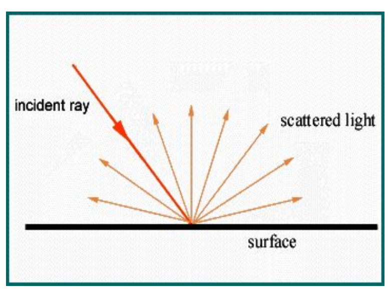




<u>Special case of diffuse reflection: Lambertian reflection.</u>

A surface called the **Lambert surface** if it obeys the **Lambert's Law** which states that the diffusely reflected light is isotropic and unpolarized (i.e., natural light) independently of the state of polarization and the angle of the incidence light.

✓ ***<u>Reflection from the Lambertian surface is isotropic:</u>***



$$R(\mu_r, \varphi_r, \mu_i, \varphi_i) = R_L \qquad [7.17]$$

where R_L is the **Lambert reflectance** (also called **surface albedo**).

✓ **In general, the surface reflectance is a function of wavelength**.

Examples of the surface albedo at ~ 550 nm:

fresh snow/ice =0.8-0.9;

desert=0.3,

soils=0.1-0.25;

ocean=0.05.

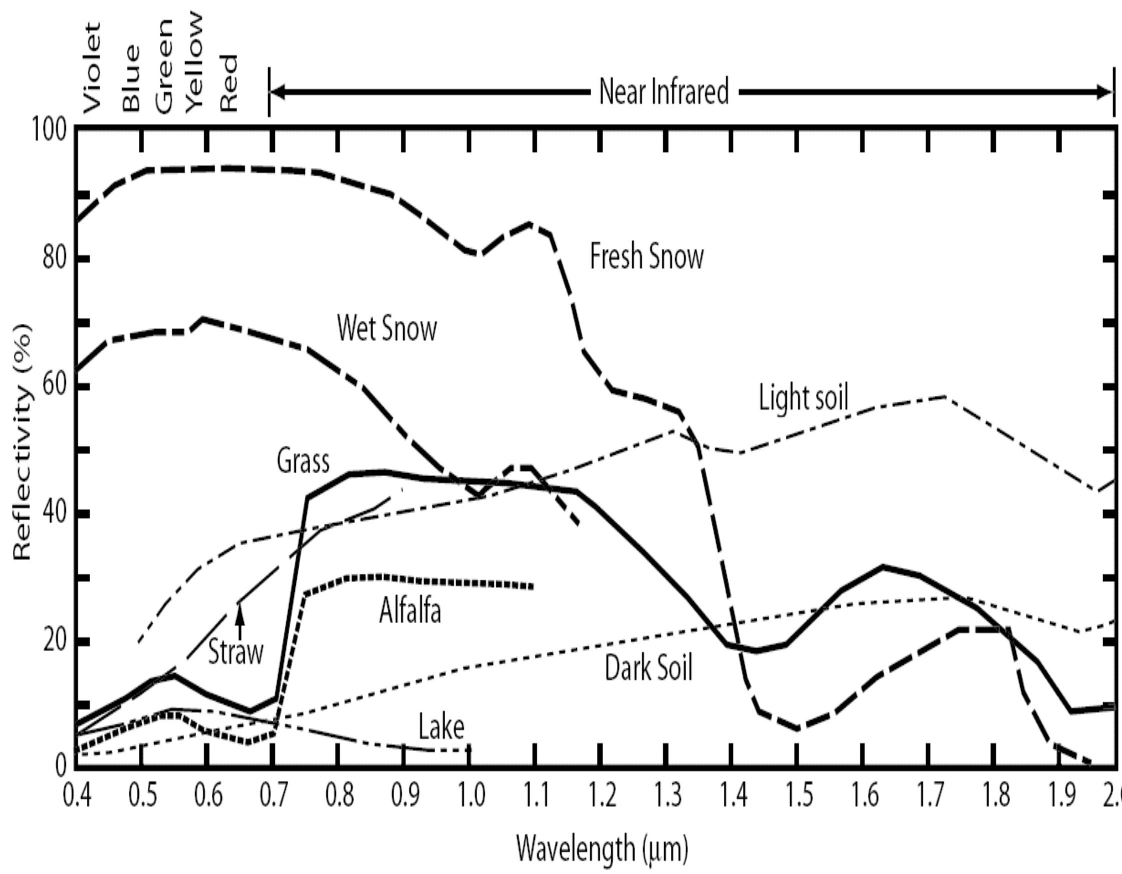


Figure 7.1 Representative spectral reflectances (albedo) of various surfaces.

NOTE: Each surface type has a specific spectral fingerprint that is the surface reflectance has a specific dependence on the wavelength. This plays a central role in the remote sensing of land and ocean surfaces.

The principle of interaction:

The resulting intensity emerging from the surface of the layer is a superposition of reflected and transmitted intensities.

NOTE: Eqs.[7.12]-[7.13] for the first order scattering were derived for non-reflecting surfaces (called black surfaces). The principle of interaction enables the incorporation of radiances reflected from the surfaces.

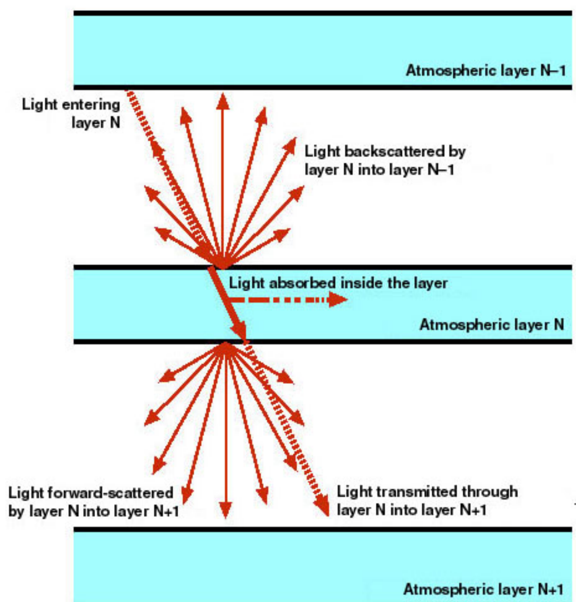
Consider an atmospheric layer that can reflect and transmit the incident radiation.

Reflection $R(\vec{\Omega},\vec{\Omega}')$ **and transmission** $T(\vec{\Omega},\vec{\Omega}')$ **functions of diffuse radiation** are defined as

$$I_{reflected}(\vec{\Omega},\vec{\Omega}') = R(\vec{\Omega},\vec{\Omega}')I(\vec{\Omega}')d\Omega' \qquad [7.18]$$

$$I_{transmitted}(\vec{\Omega},\vec{\Omega}') = T(\vec{\Omega},\vec{\Omega}')I(\vec{\Omega}')d\Omega' \qquad [7.19]$$

where $I_{\lambda}(\vec{\Omega}')$ is the incident intensity in the direction $\vec{\Omega}'(\mu',\phi')$.



If the atmopsheric layer illumitated by many sources of radiation from below and above with $I_{\lambda}(\vec{\Omega}'_k)$ of the *k*-th source below and $I_{\lambda}(\vec{\Omega}'_j)$ of the *j*-th source above, then the intensity emerging from the layer in the direction $\vec{\Omega}$ is

$$I(\vec{\Omega}) = \sum_j R(\vec{\Omega},\vec{\Omega}'_j)I(\vec{\Omega}'_j)d\Omega'_j + \sum_k T(\vec{\Omega},\vec{\Omega}'_k)I(\vec{\Omega}'_k)d\Omega'_k \qquad [7.20]$$

NOTE: See an example for the case of two layers in S 6.5.1.

3. Applications of passive remote sensing using extinction and scattering: Remote sensing of ocean color.

Remote sensing of ocean color provides information on the abundance of phytoplankton (chlorophyll) and the concentration of dissolved and particulate material in surface ocean waters.

Importance:

- biological productivity in the oceans (the oceans take up about 1/3 of CO_2, two major mechanism: solubility pump and biological pump, the latter is controlled by phytoplankton biomass)
- marine optical properties,
- the interaction of winds and currents with ocean biology,
- effects of human activities on the oceanic environment.

Ocean color is referred to the wavelength dependence of the water-leaving radiances at the ocean surface. Ocean color is the result of scattering and absorption by chlorophyll pigments, as well as dissolved and particulate matter in the surface ocean water.

Principles of ocean color retrievals:

Phytoplankton has a specific absorbing spectrum => its concentration can be retrieved if the spectral water-leaving radiances are measured.

Need for accurate atmospheric correction:

Water-leaving radiances can be as low as a few percent of the TOA (top-of-the-atmosphere) radiances => it is critical to quantify and correctly remove the contribution from the atmosphere to the TOA radiances.

SENSORS:

CZCS (Coastal Zone Color Scanner, flown on the NIMBUS-7 satellite): data for 1978 - 1986

SeaWiFS (Sea-viewing Wide Field-of-View Scanner, launched onboard Orbview-2 satellite): data from 1997

MODIS (Moderate Resolution Imaging Spectroradiometer), launched on Terra and Aqua satellites: data from 1999 for Terra.

Table 7.1 MODIS, SeaWiFS, and CZCS channels and their central wavelengths used for ocean color retrievals.

λ (nm)	MODIS	SeaWiFS	CZCS
412	+	+	-
443	+	+	+
490	+	+	-
530	+	+	+
550	+	+	+
670	+	+	+
681	+	-	-
750	+	+	-
865	+	+	-

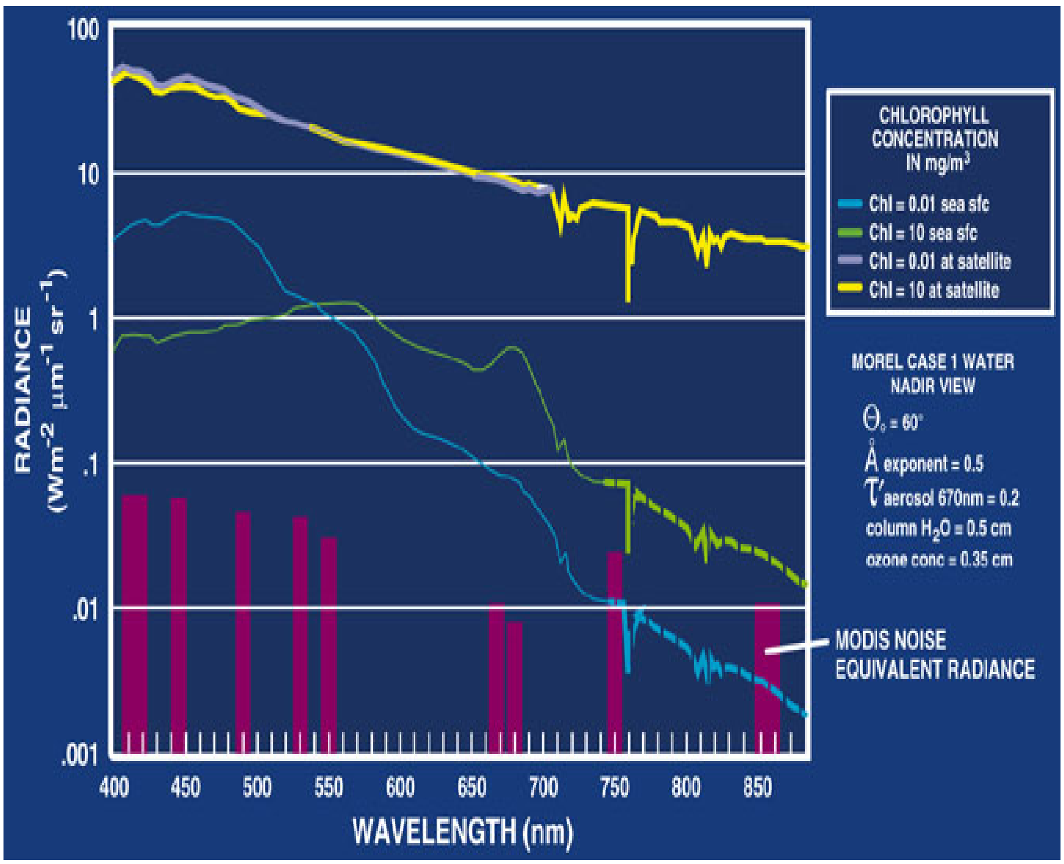


Figure 7.2 MODIS wavelength bands and the water leaving radiance in high and low chlorophyll waters without the atmospheric signal (lower curves) and with the atmospheric signal (upper curves).

Ocean color retrieval algorithm (retrieval of the chlorophyll concentration):

CZCS, SeaWiFS and MODIS algorithms use the **normalized water-leaving radiance** $[I_w]_N$ defined as

$$I_w(\lambda) = [I_w(\lambda)]_N T(\lambda) \quad [7.21]$$

where

$T(\lambda)$ is the diffuse transmittance;

$I_w(\lambda)$ is the radiance backscattered out of the water.

- The normalized water-leaving radiance is approximately the radiance that would exit the ocean in the absence of the atmosphere with the sun in the zenith.

Assuming the Lambertian surface, reflectance associated with the radiance $[I_w]_N$ can be defined as $$[R_w(\lambda)]_N = \frac{\pi}{F_0}[I_w(\lambda)]_N$$

and Eq. [7.21] becomes

$$R_w(\lambda) = [R_w(\lambda)]_N T(\lambda)$$

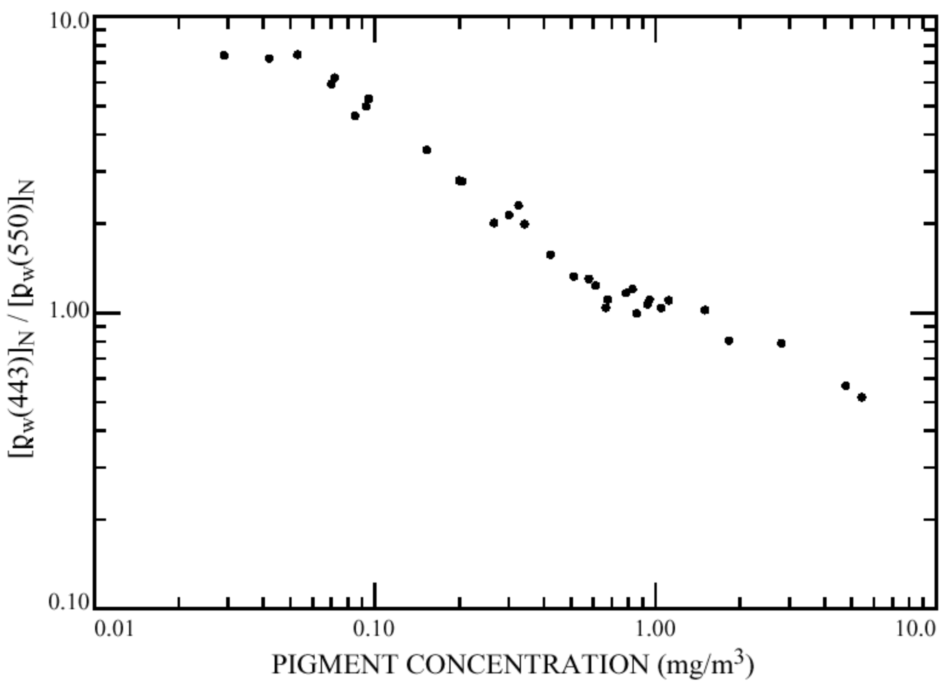


Figure 7.3 Normalized water-leaving reflectance ratio as a function of pigment (chrolophyll *a*) concentration (Gordon et al. 1988)

If the ratio $[R_w(443)]_N / [R_w(550)]_N$ is known, the pigment concentration C can be approximated as

$$\log_{10}(3.33C) = -1.2\log_{10} r + 0.5(\log_{10} r)^2 - 2.8(\log_{10} r)^3$$

where $r = 0.5[R_w(443)]_N / [R_w(550)]_N$

NOTE: Retrievals of ocean color are strongly affected by atmospheric conditions. No retrievals in the presence of clouds and heavy aerosol plumes (i.e., large optical depth).