Lecture 13

Applications of passive remote sensing using extinction and scattering: Ocean color. Atmospheric correction.

Objectives:

1. Principles of ocean color characterization. Ocean color sensors.
2. Ocean color retrieval algorithm (retrieval of the chlorophyll concentration).
3. Atmospheric correction

Required reading:

G: 6.3, 6.5.1

Additional/advanced reading:

Gordon H.R. Atmospheric correction of ocean color imagery in the Earth Observing System era. *J. Geophys. Res*., 102, 17081-17106, 1997.

Yan et al., Pitfalls in atmospheric correction of ocean color imagery: how should aerosols optical properties be computed? *Applied Optics* 41, 412-423, 2002.

1. Principles of ocean color characterization. Ocean color sensors.

Scientific Motivation

Remote sensing of ocean color provides information on the abundance of phytoplankton 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 (see Fig.13.1)
- effects of human activities on the oceanic environment.

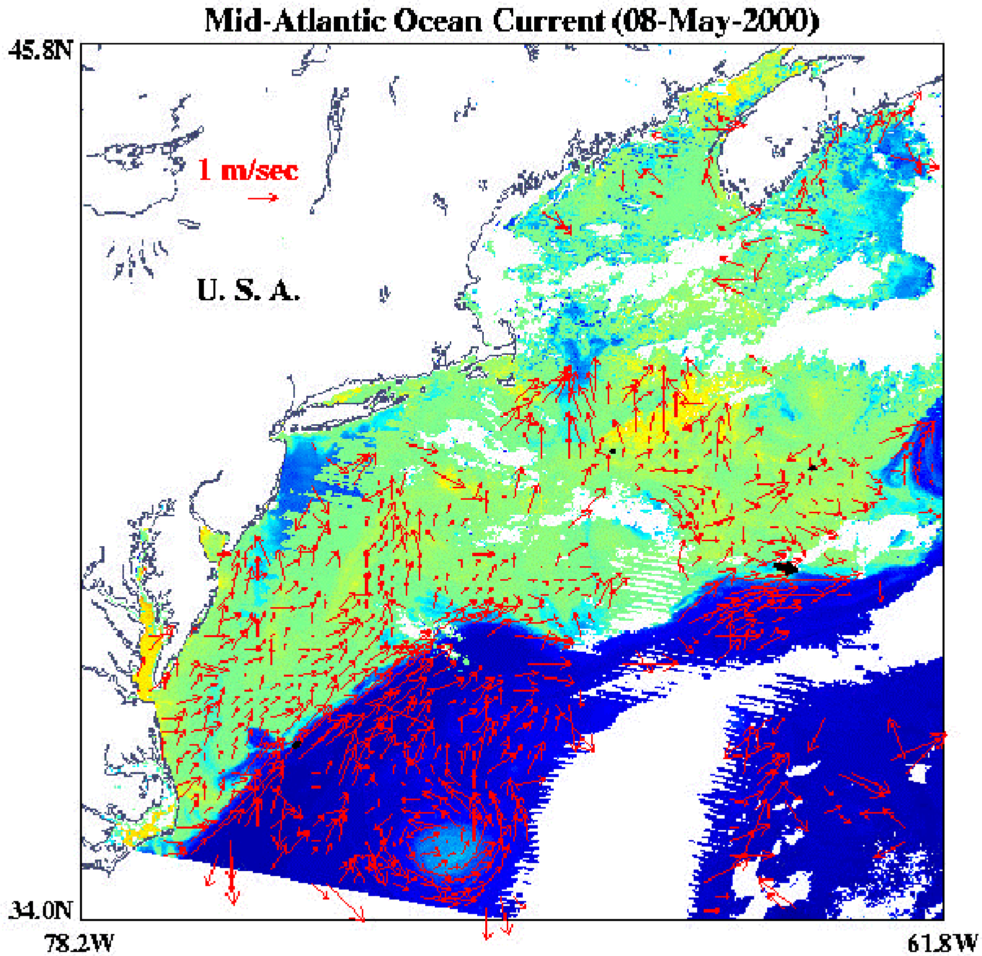


Figure 13.1 Surface layer drift (red arrows) derived from MODIS and SeaWiFS chlorophyll-a concentration data over the mid-North Atlantic Ocean (Liu et al, EOS, 83, 2002)

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.

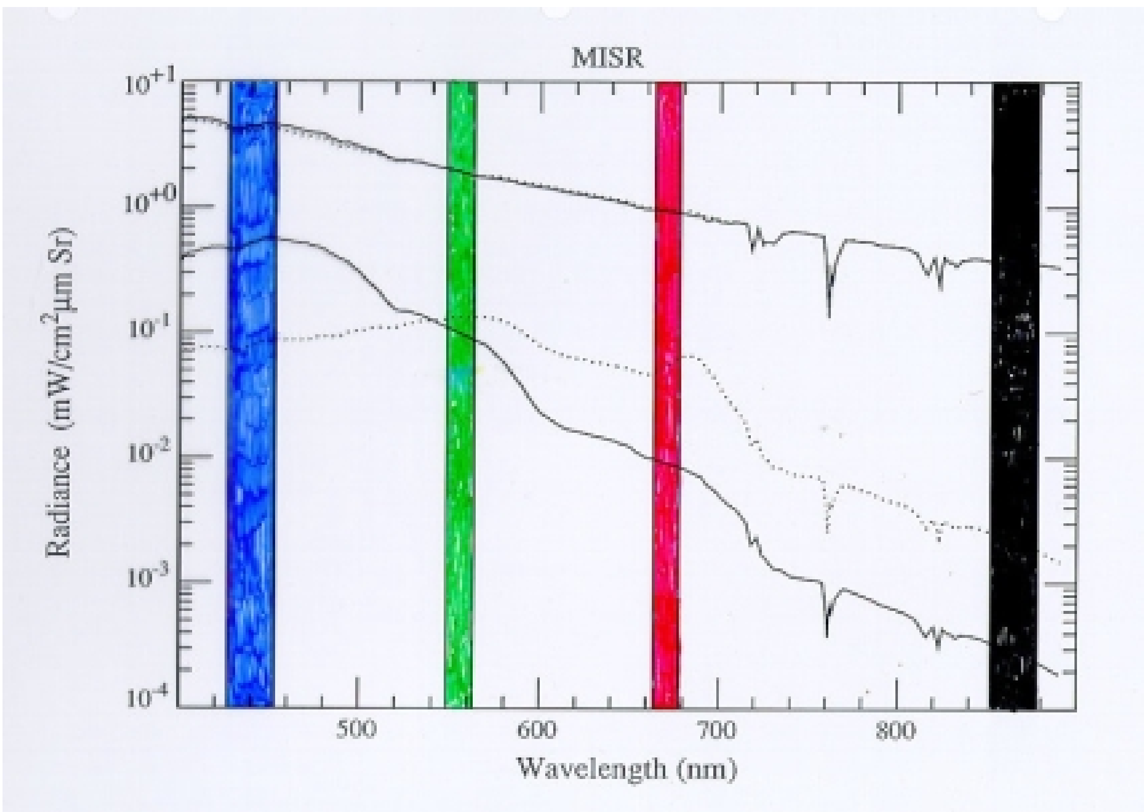


Figure 13.2 Four typically observed wavelength bands and the water leaving radiance in high (dotted) and low (solid) chlorophyll waters without the atmospheric signal (lower curves) and with the atmospheric signal (upper curves). The satellite observes the water leaving radiance with the signal due to the atmosphere (upper curves).

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 satellite):

data from 1999

Table 13.1 MODIS, SeaWiFS and CZCS channels for ocean color retrievals.

Band	λ (nm)	MODIS	SeaWiFS	CZCS
8	412	+	+	-
9	443	+	+	+
10	490	+	+	-
11	530	+	+	+
12	550	+	+	+
13	670	+	+	+
14	681	+	-	-
15	750	+	+	-
16	865	+	+	-

2. Ocean color retrieval algorithm (retrieval of the chlorophyll concentration).

CZCS, SeaWiFS and MODIS algorithms use the normalized water-leaving radiance $[\boldsymbol{I}_{\mathbf{w}}]_{\boldsymbol{N}}$ defined as

$$I_w(\lambda) = [I_w(\lambda)]_N \, T(\lambda) \qquad [13.1]$$

where

$\boldsymbol{T}$ **(λ)** is the diffuse transmittance (see Lecture 10);

$\boldsymbol{I}_{\mathbf{w}}$ **(λ)** 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 $[L_w]_N$ can be

defined as
$$[R_w(\lambda)]_N = \frac{\pi}{F_0}[I_w(\lambda)]_N$$

and Eq. [13.1] becomes

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

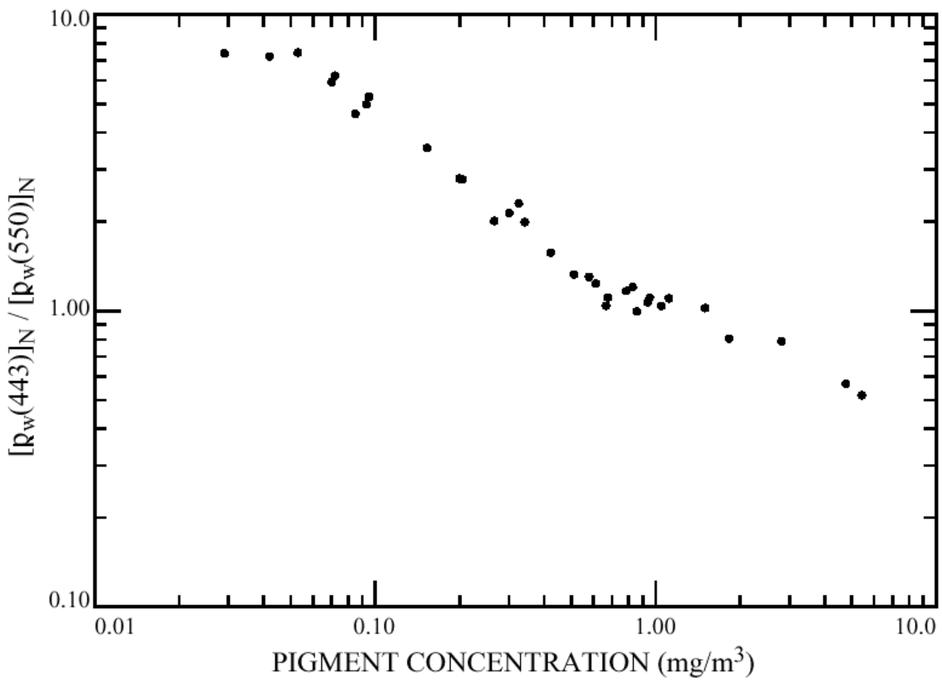


Figure 13.2 Normalized water-leaving reflectance ratio as a function of pigment concentration (Gordon et al. 1988)

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

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

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

$$\Downarrow$$

Need to correct for the atmospheric contribution to the TOA radiances measured by the sensor to retrieve the normalized water-leaving radiance.

3. Atmospheric correction.

The TOA radiance received by a sensor can be divided into the following components

$$I_{TOA}(\lambda) = I_{path}(\lambda) + T_{dir}(\lambda) I_g(\lambda) + T(\lambda) I_{wc}(\lambda) + T(\lambda) I_w(\lambda) \quad [13.2]$$

where

$I_{path}(\lambda)$ is the radiance generated along the optical path by scattering in the atmosphere and by specular reflection of atmospherically scattered light (skylight) from the ocean surface;

$I_g(\lambda)$ is the radiance arising from specular reflection of direct sunlight from the ocean surface (called sun glitter);

$I_{wc}(\lambda)$ is the radiance arising from sunlight and skylight reflecting from individual whitecaps on the sea surface;

$I_w(\lambda)$ is the desired water-leaving radiance (from the whitecap-free areas of the ocean surface);

$T_{dir}(\lambda)$ is the direct transmittance and

$$T_{dir}(\lambda) = \exp[-\{\tau_r(\lambda) + \tau_{o3}(\lambda) + \tau_a(\lambda)\} / \mu_{view}]$$

$\boldsymbol{\tau_r}$, $\boldsymbol{\tau_a}$, $\boldsymbol{\tau_{o3}}$ are the optical depth of Rayleigh scattering, aerosols, and ozone absorption, respectively;

Converting to reflectance, Eq.[13.2] becomes

$$R_{TOA}(\lambda) = R_{path}(\lambda) + T_{dir}(\lambda) R_g(\lambda) + T(\lambda) R_{wc}(\lambda) + T(\lambda) R_w(\lambda) \quad [13.3]$$

Thus, we need atmospheric correction algorithm to accurately estimate

$$R_{path}(\lambda); T_{dir}(\lambda) R_g(\lambda); T(\lambda) R_{wc}(\lambda); and\ T(\lambda)$$

$T_{dir}(\lambda)R_g(\lambda)$ this term is removed by applying a sun glitter mask.

$R_{wc}(\lambda)$ is estimated by introducing the whitecaps correction using the wind speed , W in m/s : $[R_{wc}]_N = 6.49 \times 10^{-7} W^{3.52}$

In general, $R_{path}(\lambda)$ can be decomposed into several components

$$R_{path}(\lambda) = R_r(\lambda) + R_a(\lambda) + R_{ra}(\lambda)$$

where

$R_r(\lambda)$ is the reflectance resulting from multiple scattering by air molecules (Rayleigh scattering) in the absence of aerosols;

$R_a(\lambda)$ is the reflectance resulting from multiple scattering by aerosols in the absence of the air;

$R_{ra}(\lambda)$ is the reflectance resulting from molecule-aerosol scattering (this term is zero in the single scattering case).

In the NIR region: $R_w(750) = 0$ and $R_w(865) = 0$

Thus, we need to solve the following problem: given the satellite measurement of the TOA radiances in the NIR, we predict the measured radiance that would be observed in the visible. The difference between the predicted and the measured radiance gives the water –leaving radiance transmitted to the top of the atmosphere.

<u>Let's consider the single-scattering approximation ($\tau < 1$)</u> (see Lecture 10, Eq.[10.2]):

For the single-scattering case $R_{path}(\lambda) = R_r(\lambda) + R_a(\lambda)$

$R_r(\lambda)$ can be calculated using the Rayleigh optical depth (for the surface pressure) and Rayleigh scattering phase function.

Thus, we can calculate $R_a(750)$ and $R_a(865)$ from satellite measurements at these channels and we can introduce

$$\delta(750,865) = \frac{R_a(750)}{R_a(865)} = \frac{\omega_a(750)\tau_a(750)P_a^*(750,\Omega,\Omega_0)}{\omega_a(865)\tau_a(865)P_a^*(865,\Omega,\Omega_o)}$$

where *Pa** is the aerosol scattering phase function modified to include Fresnel reflectance at the interface

NOTE: the algorithm approximates the atmosphere by a two-layer, plane-parallel model: upper layer has all air molecules, lower layer has all aerosols)

Thus, we need to develop the procedure for estimating $\delta(\lambda_i, 850)$ from measured $\delta(750, 850)$, where λ_i is 443&550 couple to retrieve the pigment concentration.

Strategy: pre-compute $\delta(\lambda_i, 850)$ for a set of **candidate aerosol models** (use Coastal, Maritime, Tropospheric and Urban aerosol models (based on Shettle and Fenn, 1979) for relative humidities 50, 70, 90, and 99%, 4{aerosol types} x4 {RH} =16)

SeaWiFS and MODIS algorithms use the same set of 16 models: each aerosol model has a prescribed log-normal size distribution and a set of refractive indices

Table 13.2 Example of Tropospheric and Urban Aerosol Models as a function of RH

RH	Tropospheric Aerosol Model r_0(μm)	Urban Aerosol Model: Two log-normal modes (μm) r_{01}	r_{02}
50%	0.02748	0.02563	0.4113
70%	0.02846	0.02911	0.4777
90%	0.03884	0.04187	0.7061
99%	0.05215	0.06847	1.4858

NOTE: recall the aerosol model used in MISR retrievals (see Table 8.1)

Remaining issues that are important in the atmospheric correction:

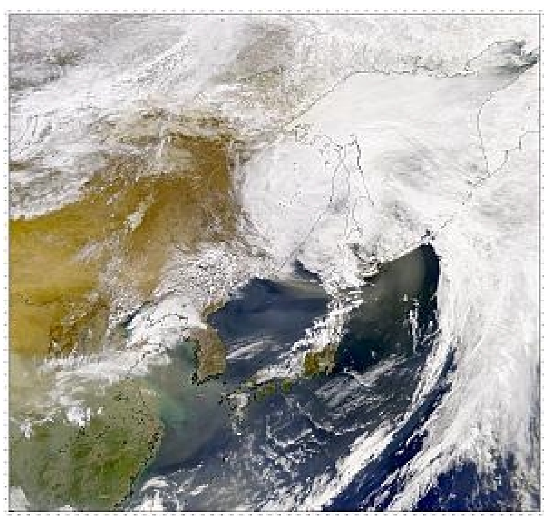
1) multiple scattering
2) effect of absorbing aerosols (need to know the vertical distribution)
3) the presence of stratospheric aerosols
4) the presence of cirrus clouds
5) polarization
6) BRDF and sea-surface roughness

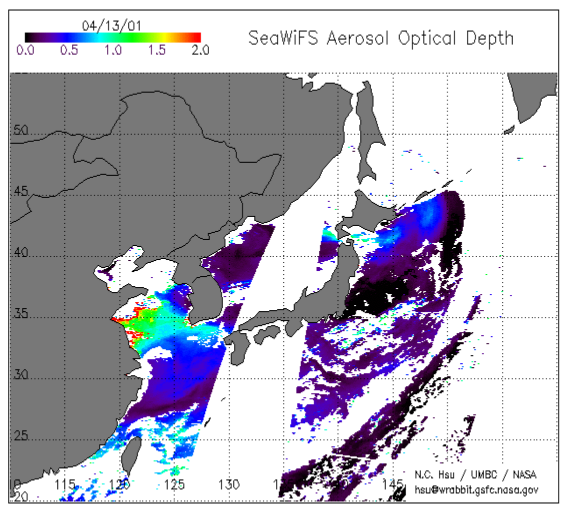
➢ SeaWiFS and MODIS aerosol retrieval algorithm

Aerosol retrieval uses the similar approach as the atmospheric correction algorithm but solves for aerosol optical depth.

NOTE: Aerosol retrieval often includes a larger set of candidate aerosol models

SeaWiFS image for April 13, 2001





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