Atmospheric Correction of Ocean Color RS Observations

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Lecture Outlines (1)

1. Introduction
   - Brief history
   - Basic concept of ocean color measurements
   - Why need atmospheric correction
2. Radiometry and optical properties
   - Basic radiometric quantities
   - Apparent optical properties (AOPs)
   - Inherent optical properties (IOPs)
3. Optical properties of the atmosphere
   - Molecular absorption and scattering
   - Aerosol properties and models
     - Non- and weakly absorbing aerosols
     - Strongly absorbing aerosols (dust, smoke, etc.)
4. Radiative Transfer
   - Radiative Transfer Equation (RTE)
   - Successive-order-of-scattering method
   - Single-scattering approximation

Lecture Outlines (2)

5. Atmospheric Correction
   - Define reflectance and examine the various terms
   - Sea surface effects
   - Atmospheric diffuse transmittance
   - Normalized water-leaving radiance
   - Single-scattering approximation
   - Aerosol multiple-scattering effects
   - Open ocean cases: using NIR bands for atmospheric correction
   - Coastal and inland waters
     - Brief overviews of various approaches
     - The SWIR-based atmospheric correction
   - Examples from MODIS-Aqua, VIIRS, and GOCI measurements
6. Addressing the strongly-absorbing aerosol issue
   - The issue of the strongly-absorbing aerosols
   - Some approaches for dealing with absorbing aerosols
   - Examples of atmospheric correction for dust aerosols using MODIS-Aqua and CALIPSO data
7. Requirements for future ocean color satellite sensors
8. Some recent research results
9. Summary

Some Useful References

IOCCG Report #10 at http://www.ioccg.org/reports_ioccg.html
"Atmospheric Correction for Remotely-Sensed Ocean Colour Products".

Most reference citations from my presentation can be found from the report.

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- Van de Hulst, H. C. (1980), "Multiple Light Scattering," Academic Press, New York, 739pp.
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Introduction

1. Introduction
   - Brief history
   - Basic concept of ocean color measurements
   - Why need atmospheric correction

Brief History

- Color photographs of the oceans were obtained from spacecraft (Apollo, etc., 1960's).
- Clarke et al. (1970) showed the systematic measurements of radiance spectra of sea (ocean color) from aircraft, demonstrated the possibility of detecting the chlorophyll concentration within the ocean upper layers. They also showed atmospheric effects on the measured signals.
- Tyler & Smith (1970) performed field measurements of upward & downward irradiance spectra in different water bodies. Thereafter, there were many similar in situ water optical measurements.
- Interpretation of in situ reflectance measurements was given in the frame of *radiative transfer* by Gordon *et al*. (1975) and also in terms of optically-significant water substances by Morel and Prieur (1977) and Smith and Baker (1978).
- Gordon (1978; 1980) developed a single-scattering atmospheric correction algorithm for processing the NASA CZCS ocean color data, demonstrating the feasibility of satellite ocean color remote sensing.
- Advanced atmospheric correction algorithm (Gordon and Wang, 1994) has been developed for various more sophisticated ocean color satellite sensors, e.g., SeaWiFS, OCTS, MODIS, MERIS, VIIRS, etc., following the successful CZCS proof of concept mission.
- In recent years, atmospheric correction effort has been on dealing with more complex water properties in coastal and inland waters, as well as strongly-absorbing aerosols.

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Satellite Ocean Color Measurements

Satellite ocean color remote sensing is possible because:

- Ocean color (spectral radiance/reflectance) data---water-leaving radiance/reflectance spectra---can be related to water properties, e.g., chlorophyll-a concentration, CDOM, total suspended matter, water absorption/scattering (IOPs), etc. In other words, there exist various relationships between water optical property and water biological / biogeochemical et al. properties.
- It is possible to carry out **atmospheric correction** for deriving accurate water radiance/reflectance spectra data from satellite measurements.

Ocean Color Spectra

Satellite-Measured Reflectance (Radiance)

$$\rho_t(\lambda) = \underbrace{\rho_r(\lambda)}_{\text{Rayleigh}} + \underbrace{\rho_A(\lambda)}_{\text{Aerosols}} + \underbrace{\underbrace{t(\lambda)t_0(\lambda)}_{\text{Transmittance}}\underbrace{\left[\rho_w(\lambda)\right]_N}_{\text{Ocean}}}_{\text{Satellite-Measured Ocean Contribution}}$$

$$\rho(\lambda) = \pi L(\lambda)/\mu_0 F_0(\lambda)$$

$$\rho_{path}(\lambda) = \rho_r(\lambda) + \underbrace{\rho_A(\lambda)}_{\rho_a(\lambda)+\rho_{ra}(\lambda)}$$

Normalized water-leaving radiance:

$$\left[L_w(\lambda)\right]_N, nL_w(\lambda) \quad \mathrm{mW\,cm^{-2}\,\mu m^{-1}\,str^{-1}}$$

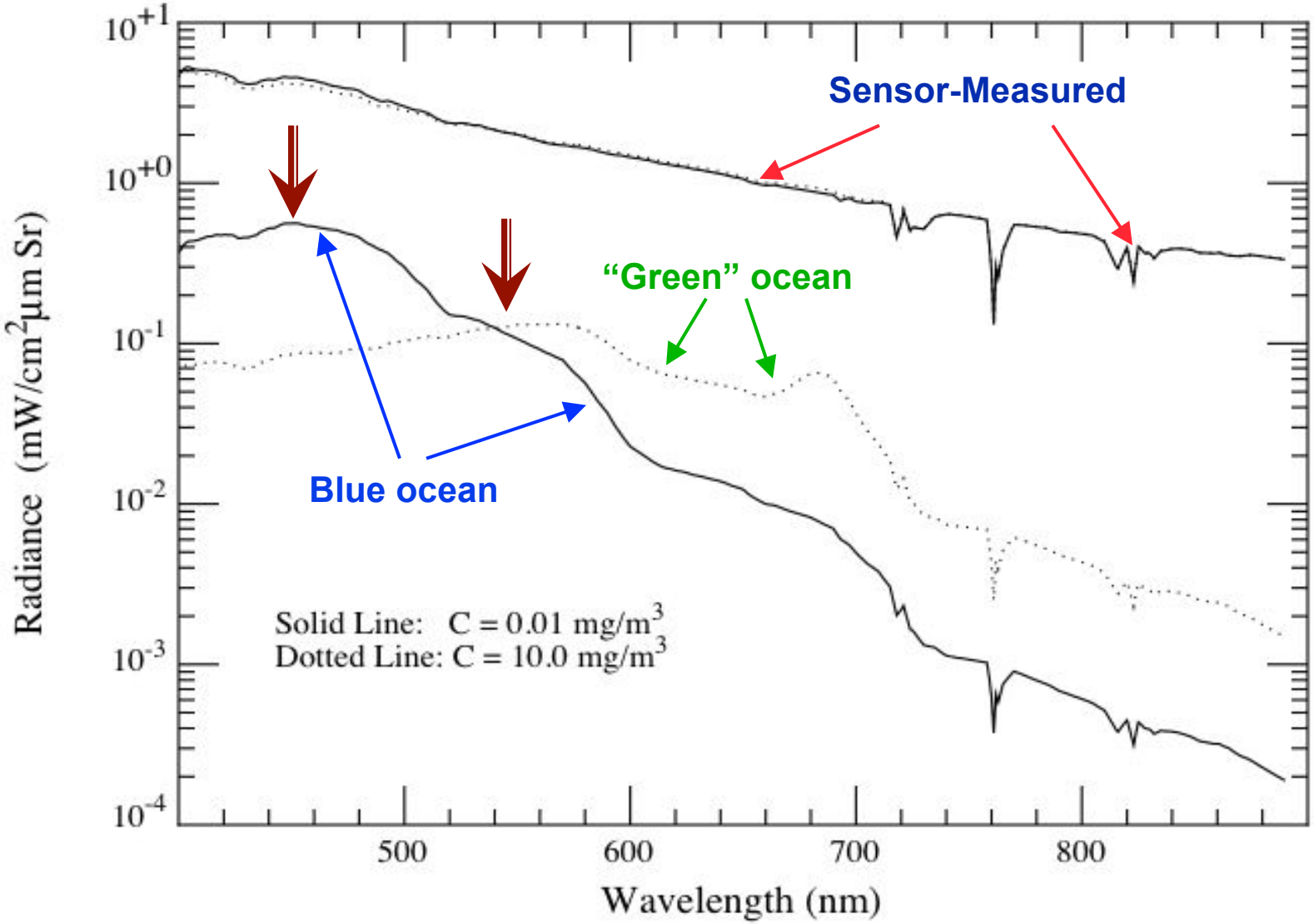
Normalized water-leaving reflectance:

$$\left[\rho_w(\lambda)\right]_N, \rho_{wN}(\lambda) \quad \text{unitless}$$

Remote sensing reflectance:

$$R_{rs}(\lambda) = \rho_{wN}(\lambda)/\pi \quad \mathrm{str^{-1}}$$

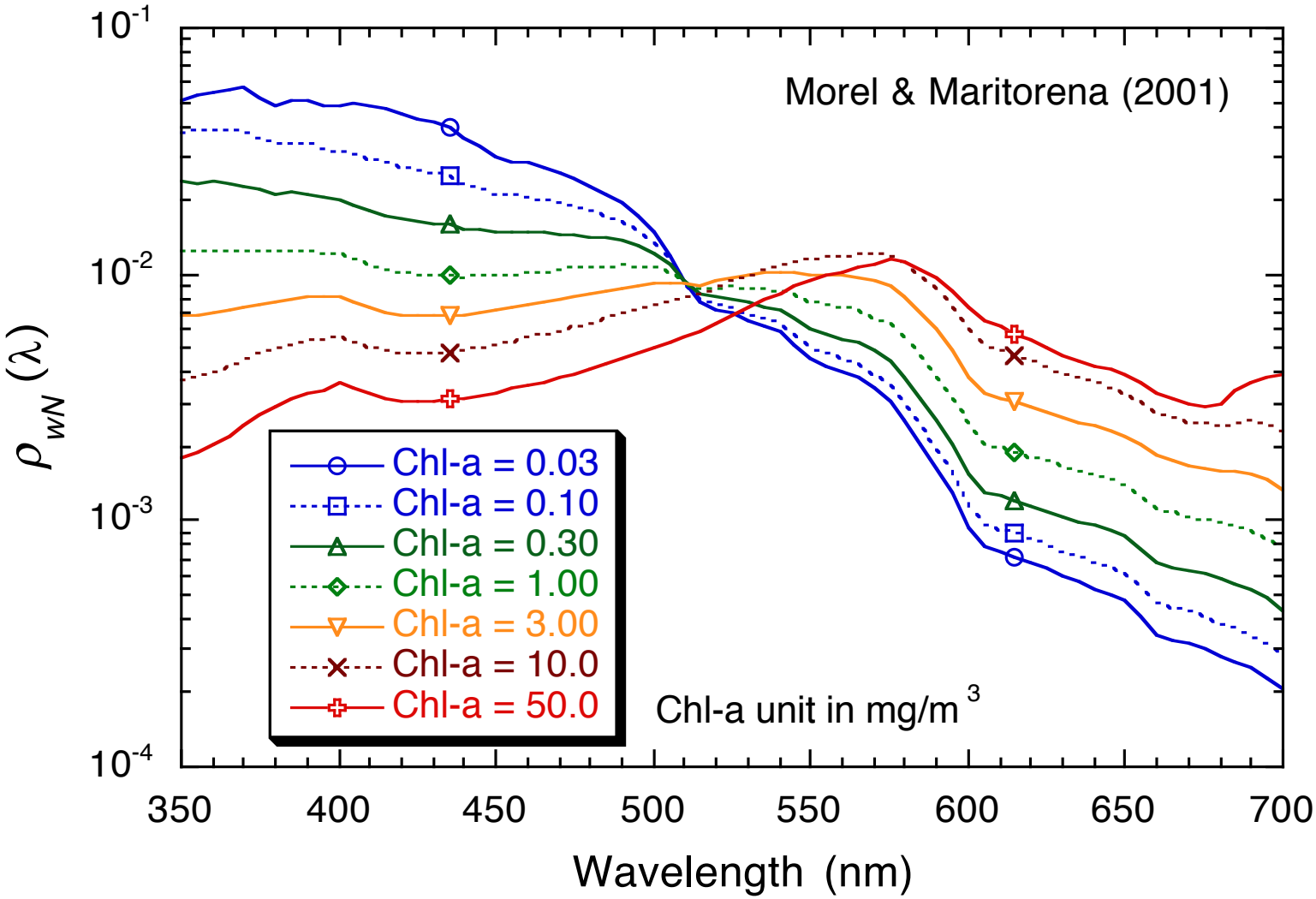
Ocean Color Remote Sensing



- Atmospheric Correction (removing >90% sensor-measured signals)
- Calibration (0.5% error in TOA >>>> 5% in surface)

From H. Gordon

Ocean Reflectance Contribution: Case-1 Waters



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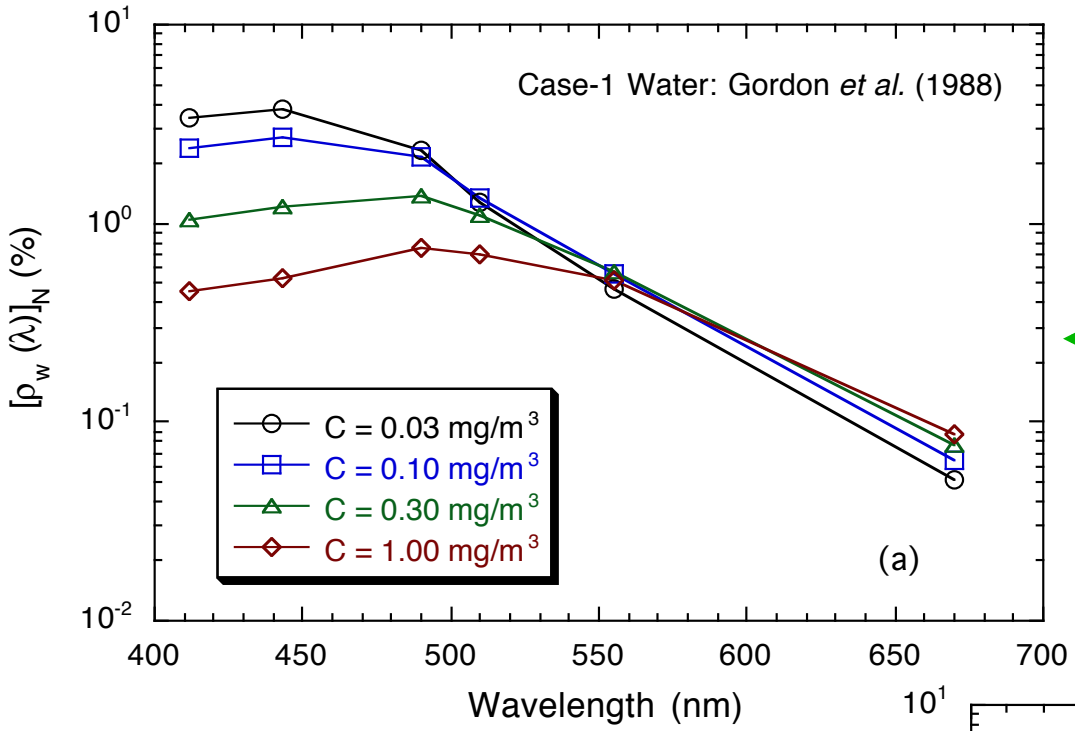
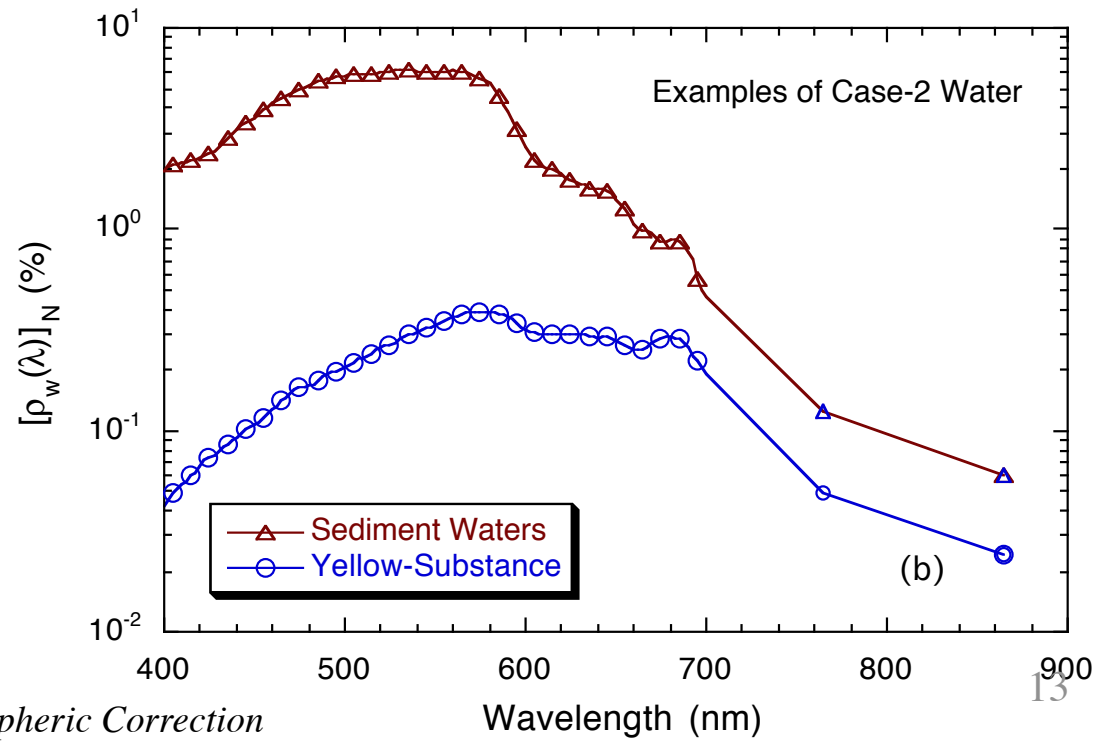


Ocean Contributions: Case-1 Waters

Ocean Contributions: Case-2 Waters



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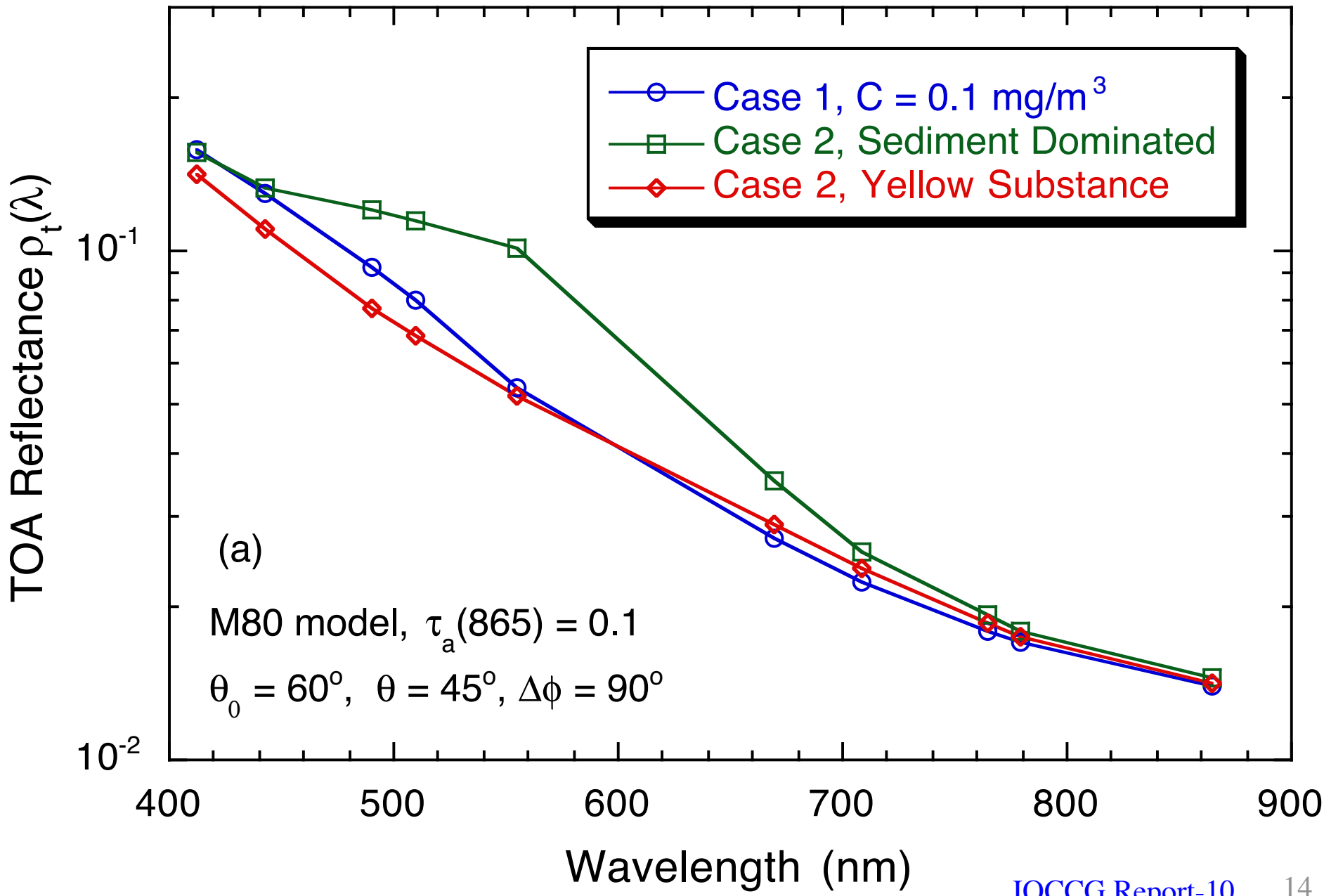
Satellite Sensor Measured TOA Reflectance Spectra



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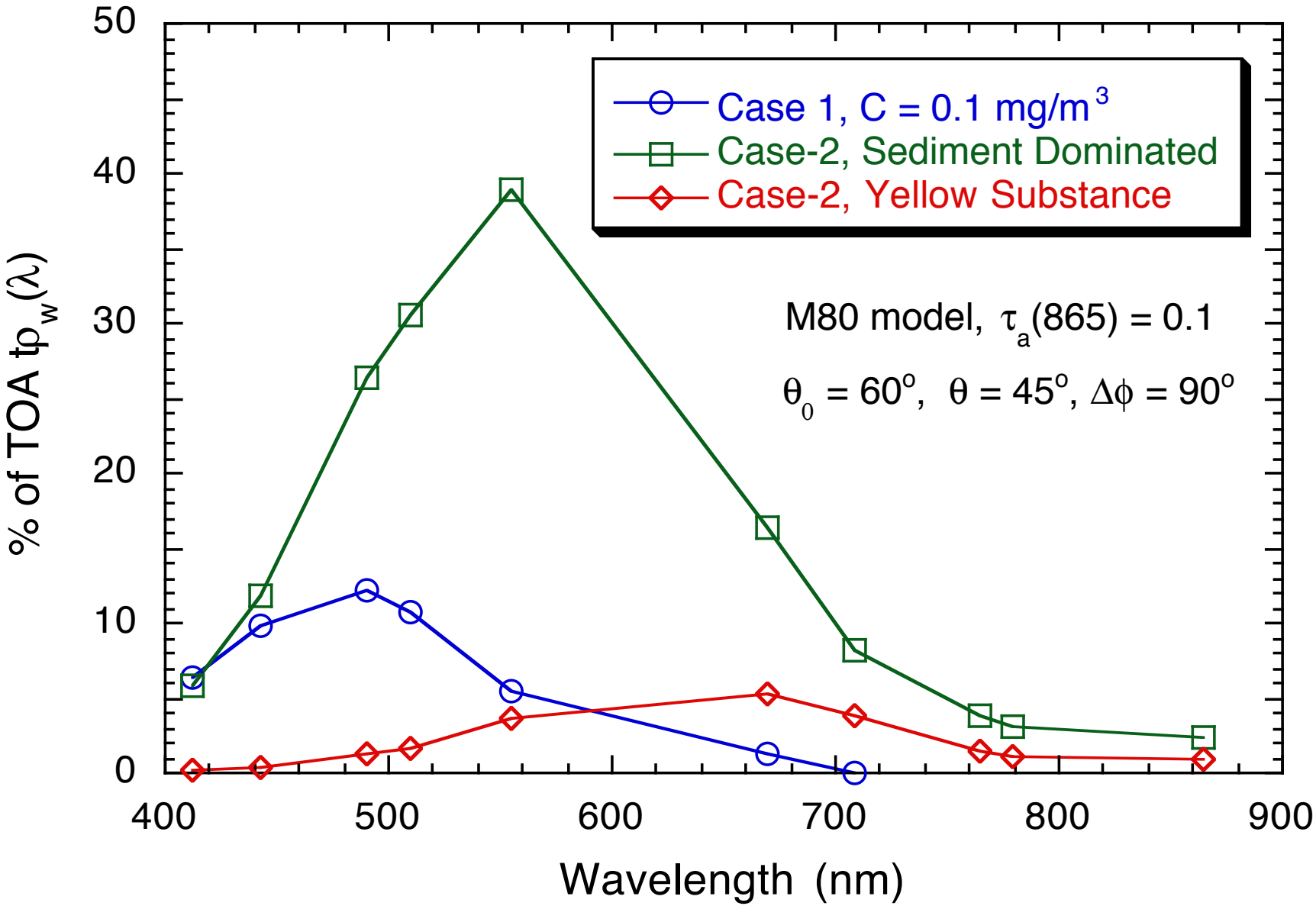
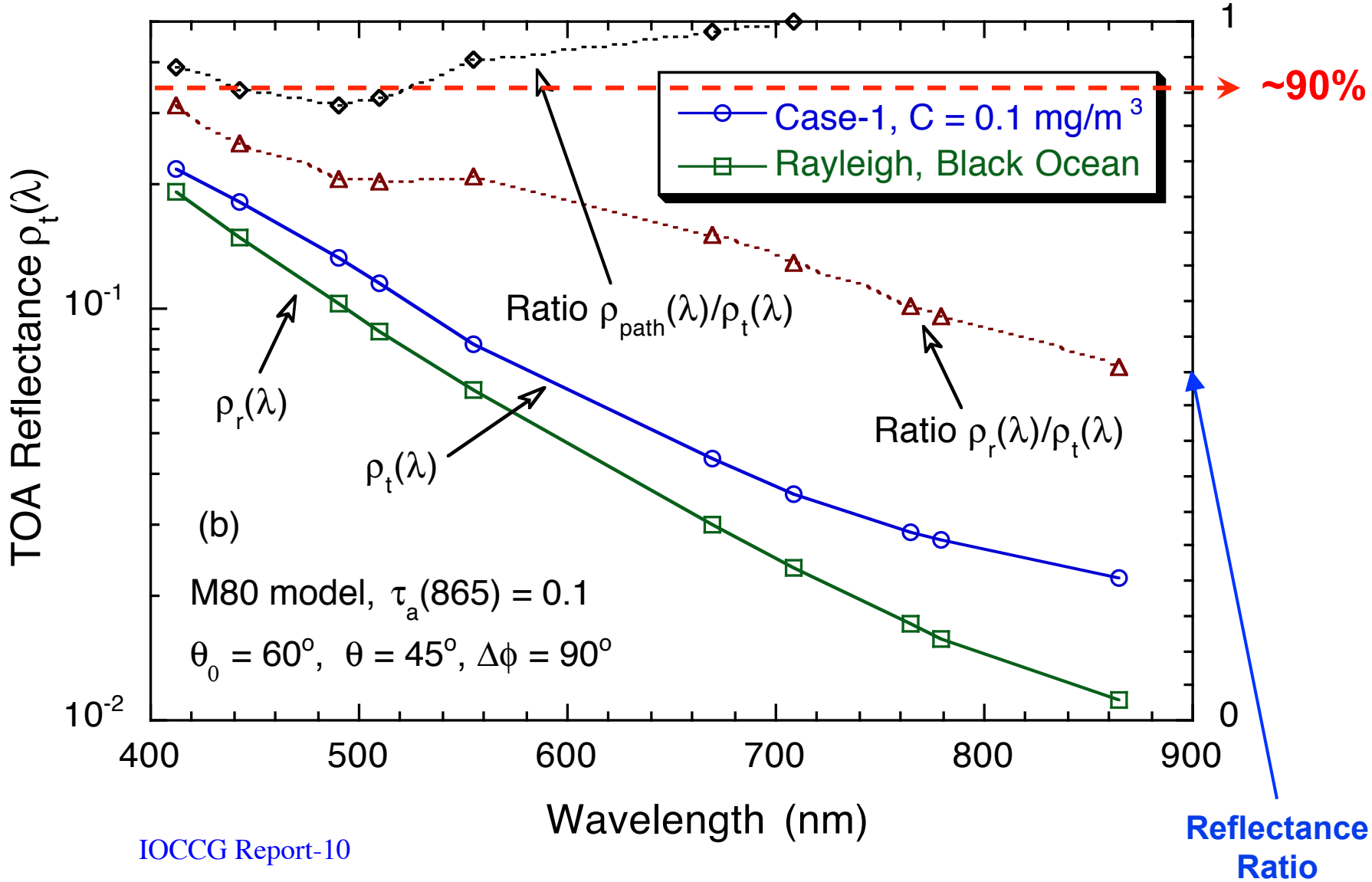
The TOA Ocean Contributions



Ocean TOA Radiance Contributions for Case-1 Waters



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For a typical **Case-1** (**open ocean**) water: Satellite-measured water-leaving reflectance contributions at the TOA are 6.4%, 9.8%, 12.1%, 10.8%, 5.5%, and 1.3% at wavelengths of 412, 443, 490, 510, 555, and 670 nm.

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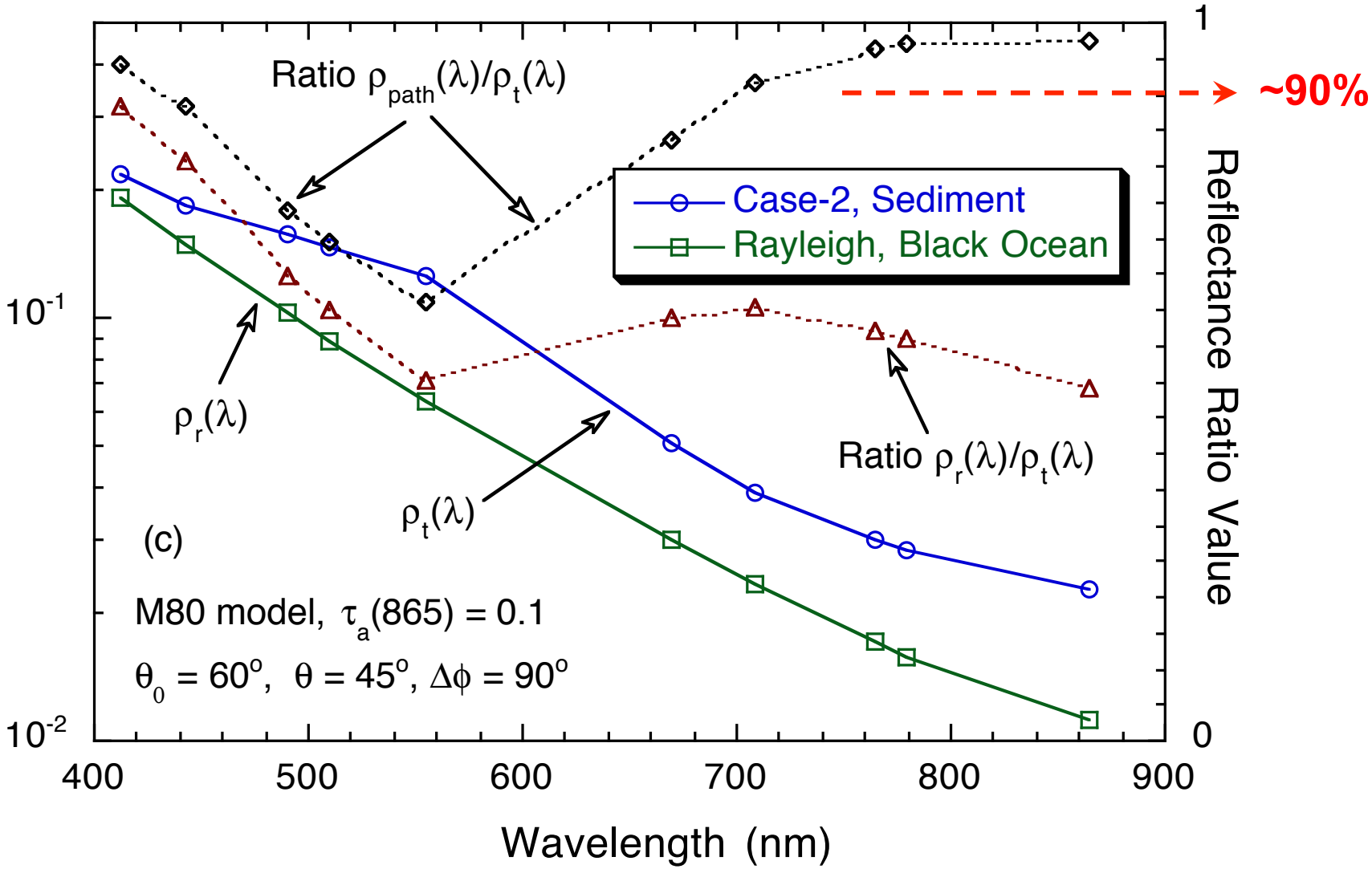
Ocean TOA Radiance Contributions for Sediment-type Case-2 Waters



For a typical **sediment-dominated** water:
Satellite-Measured water-leaving reflectance contributions at the TOA are 5.8%, 11.2%, 26.3%, 30.6%, 39.0%, 16.4%, **3.7%**, and **2.4%** at wavelengths of 412, 443, 490, 510, 555, 670, 765, and 865 nm.

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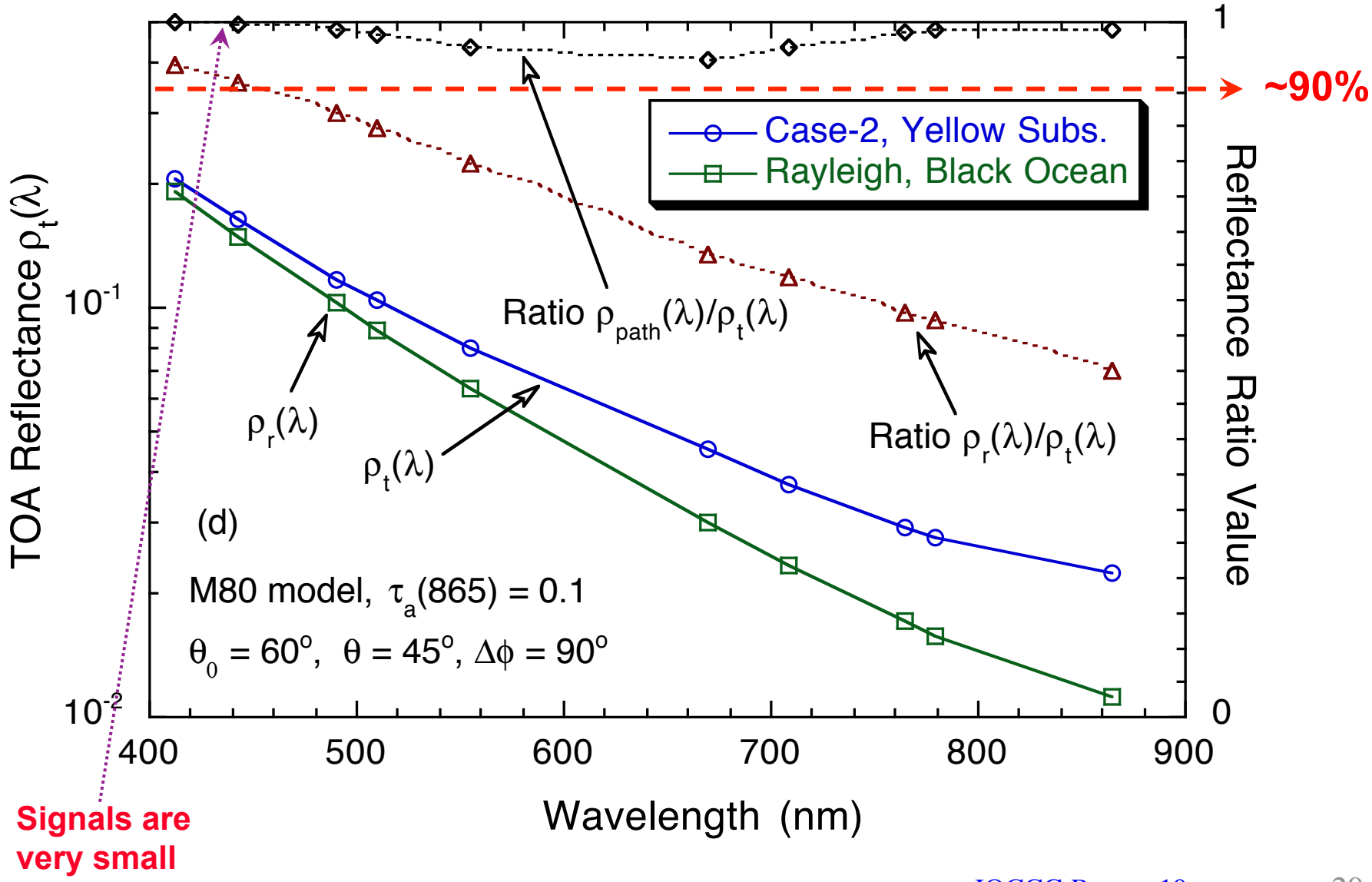
Ocean TOA Radiance Contributions for Yellow Substance-type Case-2 Waters



Signals are very small

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For a **yellow-substance-dominated** (CDOM) water: Satellite-Measured water-leaving reflectance contributions at the TOA are **0.2%, 0.4%,** 1.2%, 1.7%, 3.6%, 5.3%, 1.2%, and 1.0% at wavelengths of 412, 443, 490, 510, 555, 670, 765, and 865 nm.

It is very difficult to do atmospheric correction accurately at the blue bands!

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Atmospheric Correction

At satellite altitude
~90% of sensor-measured signal over ocean
comes from the **atmosphere & surface**!

Factor of 1:10 from TOA to the surface

- It is crucial to have accurate **atmospheric correction** and **sensor calibrations**.
- **0.5%** error in atmospheric correction or calibration corresponds to possible of **~5%** error in the derived ocean water-leaving radiance.
- We need **~0.1%** sensor calibration accuracy.

Ocean Color Remote Sensing: Derive the ocean water-leaving radiance spectra by accurately removing the atmospheric and surface effects.

Ocean properties: e.g., chlorophyll-a concentration, diffuse attenuation coefficient at 490 nm, TSM, etc., can be derived from the ocean water-leaving radiance spectra.

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Radiometry and Optical Properties

2. Radiometry and optical properties
   - Basic radiometric quantities
   - Apparent optical properties (AOPs)
   - Inherent optical properties (IOPs)

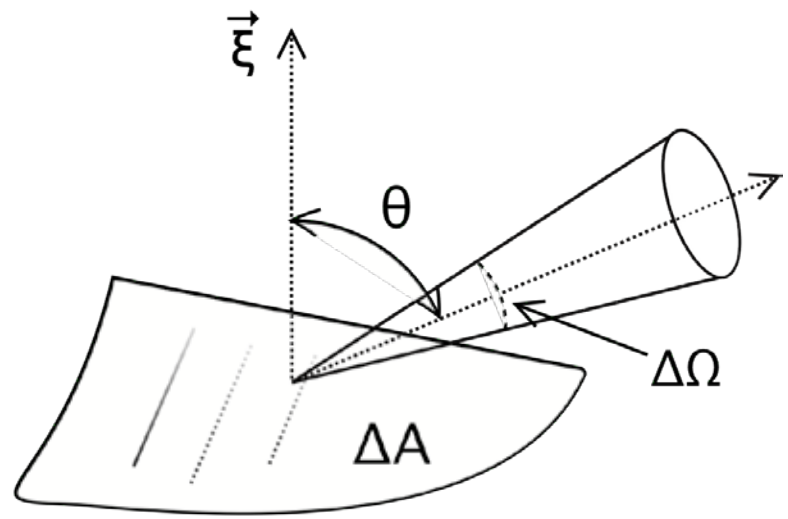
Apparent Optical Properties

Radiant Power:

$$P(\lambda) = N\frac{hc}{\lambda} / t = Energy / t \quad N - \# \text{ of photons}$$

Definition of Radiance *L*:

$$L(\lambda) = \frac{\Delta^2 P(\lambda)}{\cos\theta\,\Delta\Omega\,\Delta A}$$

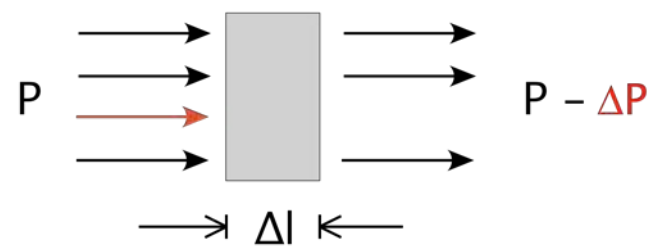


Radiance is Apparent Optical Property (**AOP**).
Ocean color satellite sensors measure Radiance!

Inherent Optical Properties (IOPs) (1)

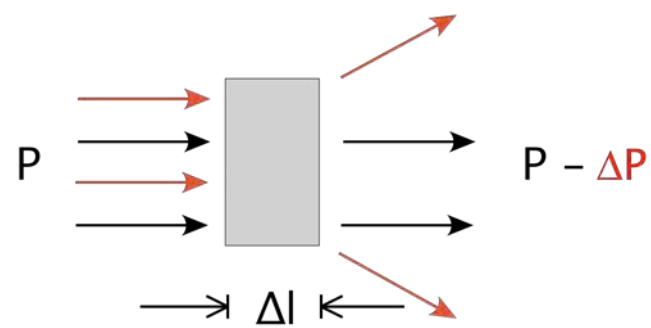
Absorption coefficient ***a***:

$$a(\lambda) = \frac{\Delta P(\lambda)}{P(\lambda)\Delta l}$$



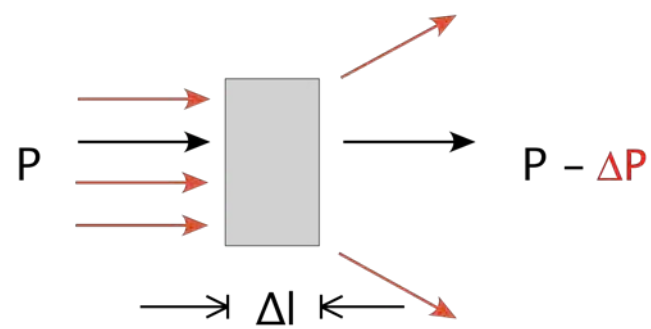
Scattering coefficient ***b***:

$$b(\lambda) = \frac{\Delta P(\lambda)}{P(\lambda)\Delta l}$$



Extinction coefficient ***c***:

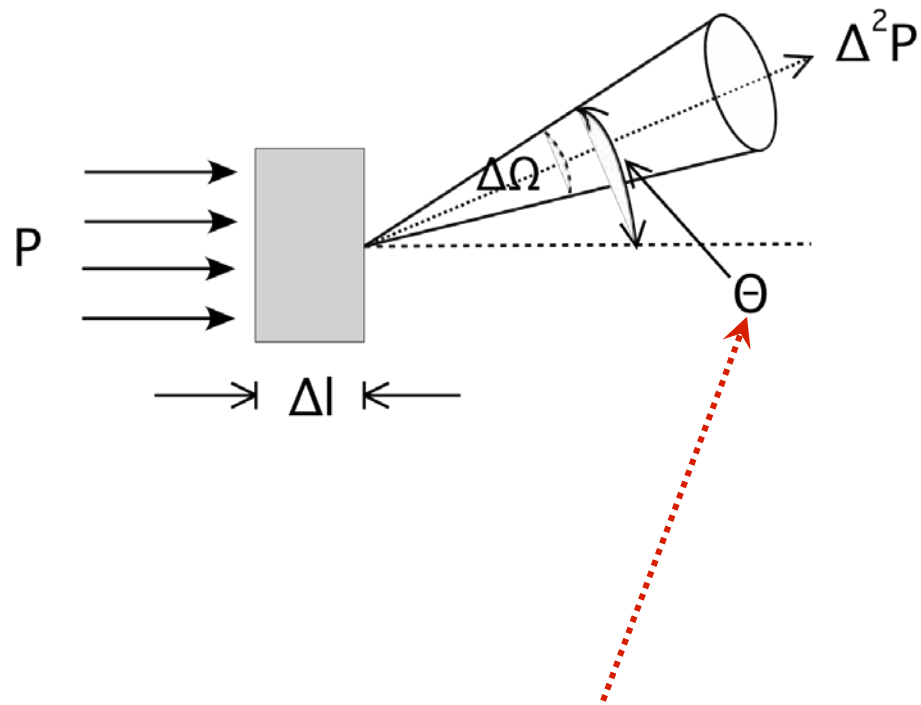
$$c(\lambda) = \frac{\Delta P(\lambda)}{P(\lambda)\Delta l} = a(\lambda) + b(\lambda)$$



Inherent Optical Properties (IOPs) (2)

Volume Scattering Function:

$$\beta(\Theta) = \frac{\Delta^2 P(\lambda)}{P(\lambda)\Delta\Omega\,\Delta l}$$



$$\int_{4\pi} \beta(\Theta) d\Omega = \frac{\Delta P(\lambda)}{P(\lambda)\Delta l} \equiv b(\lambda)$$

Definition of **Scattering Angle**

Scattering phase function ***P***:

$$P(\Theta) = \frac{\beta(\Theta)}{b}$$

Single-Scattering Albedo

The **single scattering albedo** is a dimensionless parameter which represents the percentage of photons being scattered in the single scattering case:

$$\omega_0(\lambda) = \frac{b(\lambda)}{c(\lambda)}$$

Optical Thickness (e.g., aerosol optical thickness (AOT), Rayleigh optical thickness, cloud optical thickness, etc.)

$$\tau(\lambda) = \int_0^Z c(z', \lambda)\, dz'$$

Optical Properties of the Atmosphere

3. Optical properties of the atmosphere
   - Molecular absorption and scattering
   - Aerosol properties and models
     - Non- and weakly absorbing aerosols
     - Strongly absorbing aerosols (dust, smoke, etc.)

Molecular (Rayleigh) Scattering

- The scattering of air molecules is purely electric dipole in character.
- The scattering phase function is symmetry, i.e., equal amounts of radiance are scattered into the forward and backward directions.
- The amount of scattered radiance varies nearly as the inverse fourth power of wavelength (blue sky).

Molecular (Rayleigh) Scattering

Rayleigh Phase Function: $P_r(\Theta) \propto 1 + \cos^2 \Theta$

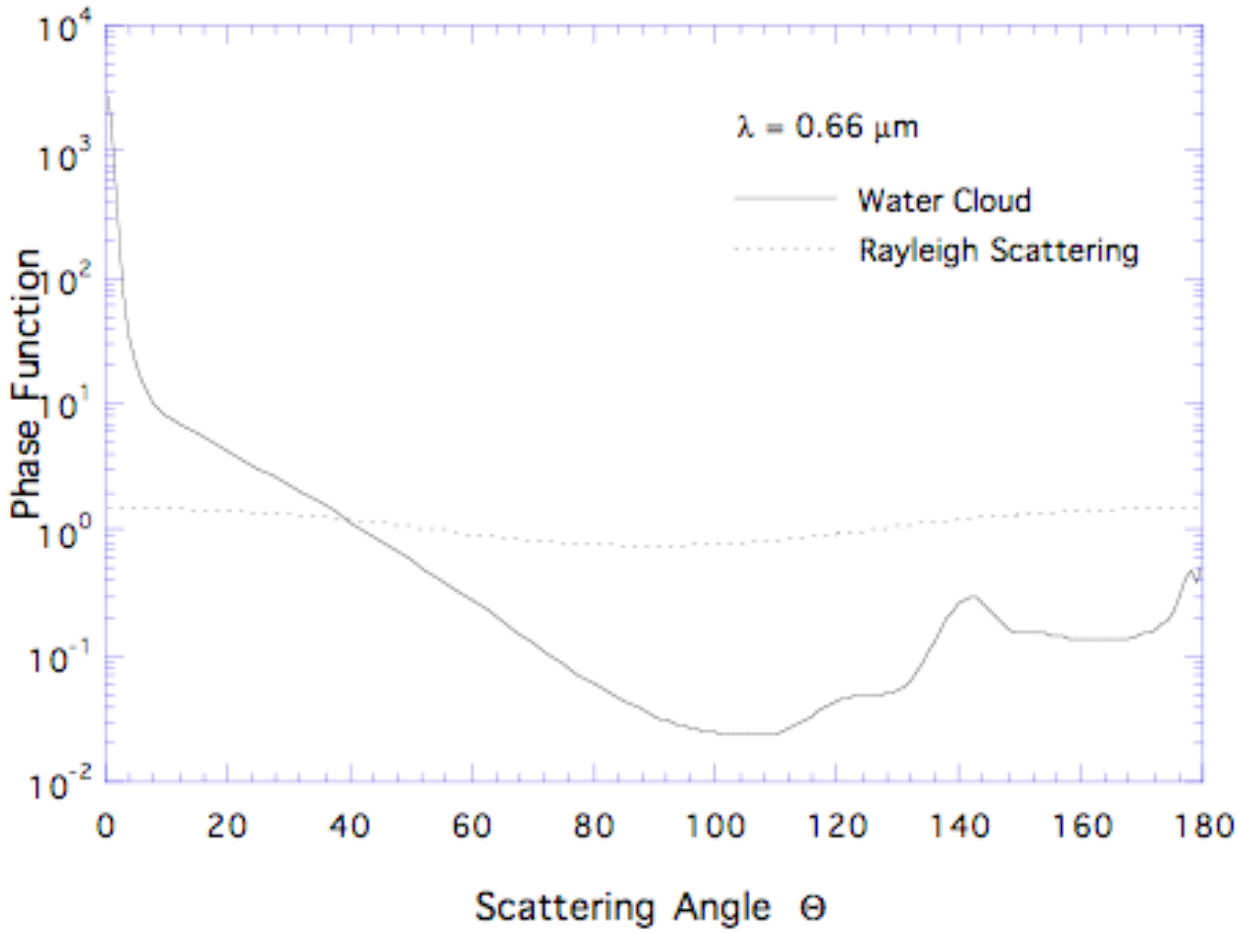
The Rayleigh optical thickness is (Hansen & Travis, 1974):

$$\tau_r(\lambda, P_0) = 0.008569\,\lambda^{-4}\left(1 + 0.0113\,\lambda^{-2} + 0.00013\,\lambda^{-4}\right)$$

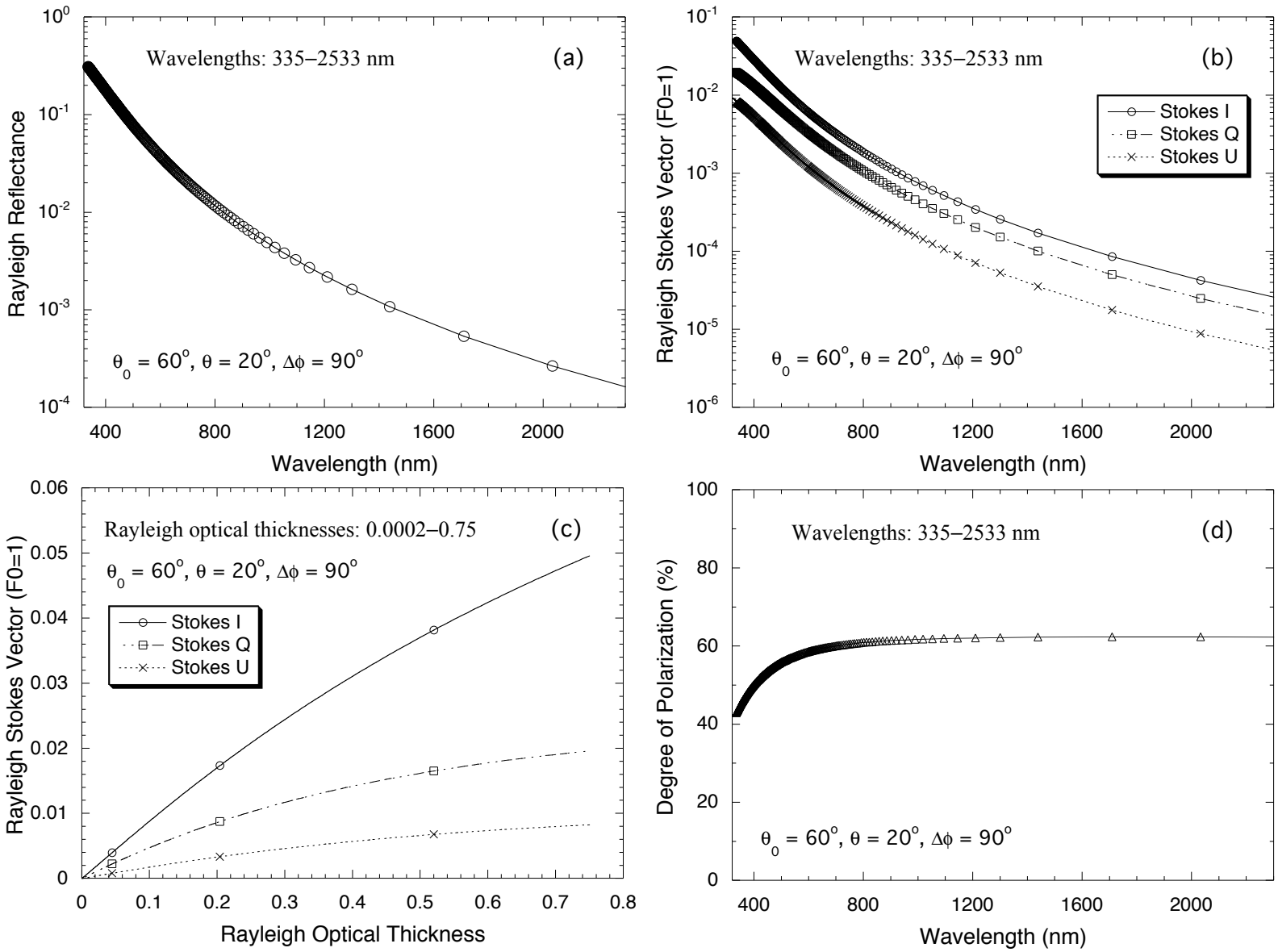
P_0 is the standard atmospheric pressure of 1013.25 hPa, and for a pressure P,

$$\tau_r(\lambda, P) = \tau_r(\lambda, P_0)\frac{P}{P_0}$$

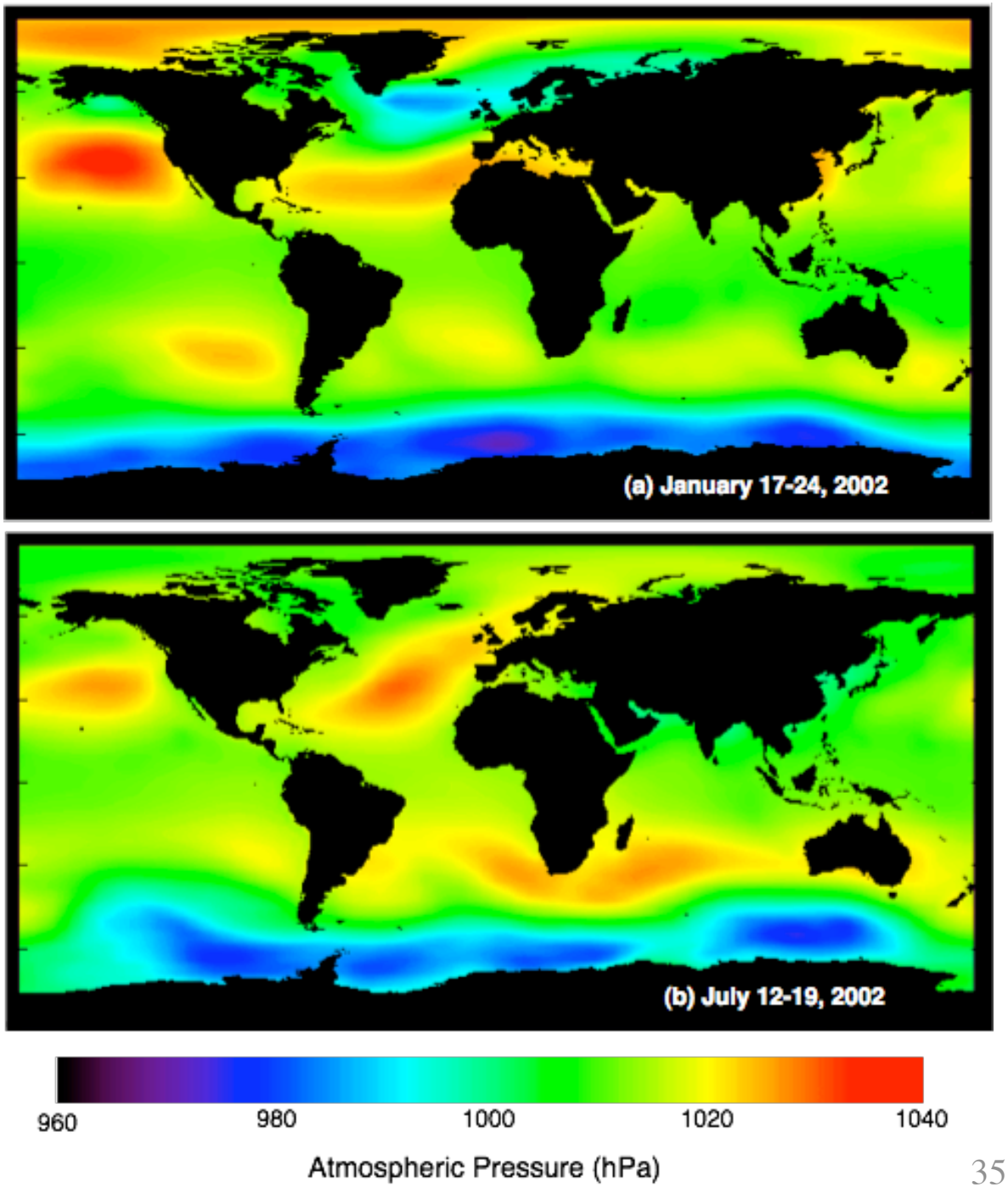
Rayleigh Phase Function



Rayleigh Scattering Radiance & Stokes Components



Atmospheric Pressure Variation



Usually, it is within ~±3%

Haze C Aerosol Model

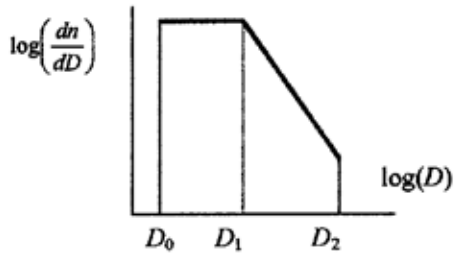
Aerosol Size Distributions

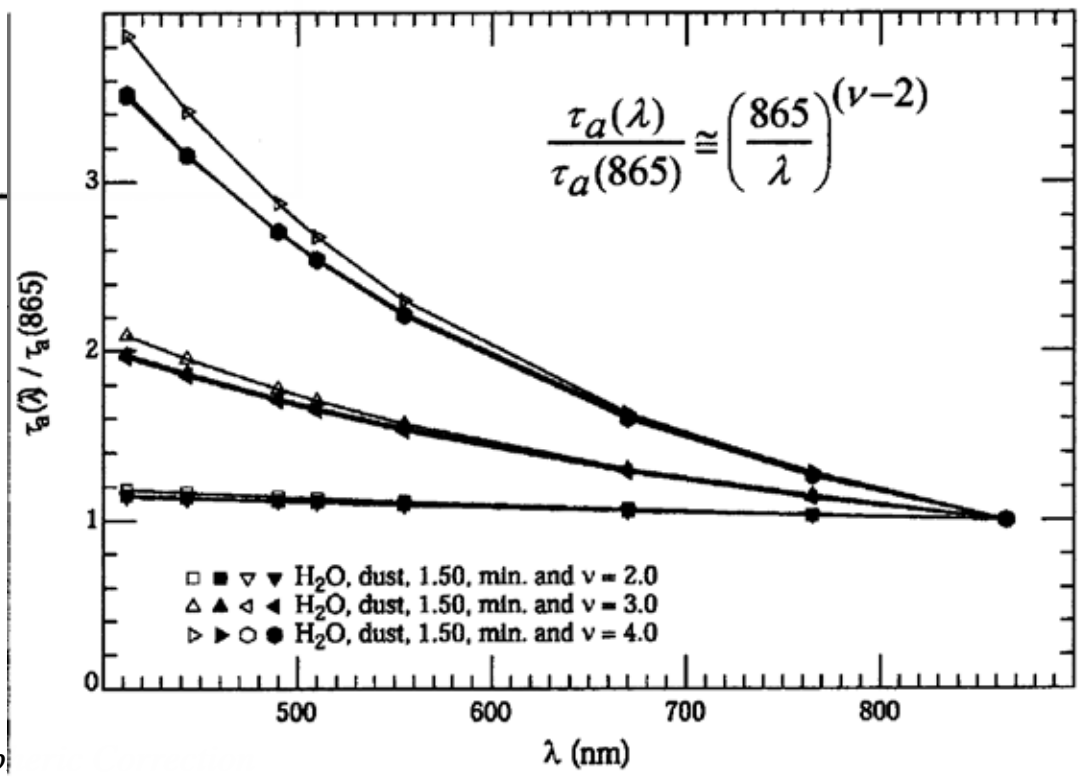
Size Distribution: $dn(D)$ = number/volume between D and $D + dD$

Total number: $N = \int_0^\infty \frac{dn}{dD} dD$

Haze C Models:

$$\frac{dn}{dD} = K, \qquad D_0 < D < D_1$$
$$= K\left(\frac{D_1}{D}\right)^{(\nu+1)}, \qquad D_1 < D < D_2$$
$$= 0, \qquad D > D_2$$



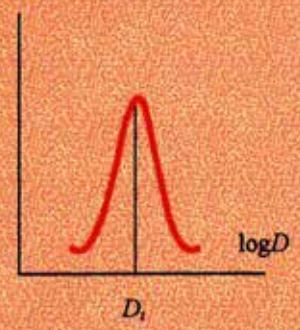


$$\frac{\tau_a(\lambda)}{\tau_a(865)} \cong \left(\frac{865}{\lambda}\right)^{(\nu-2)}$$

Shettle and Fenn Models:

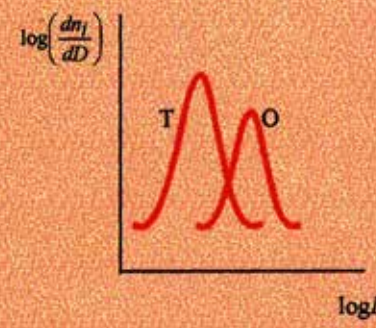
$$\frac{dn}{dD} = \sum_{i=1}^{2} \frac{dn_i}{dD}$$

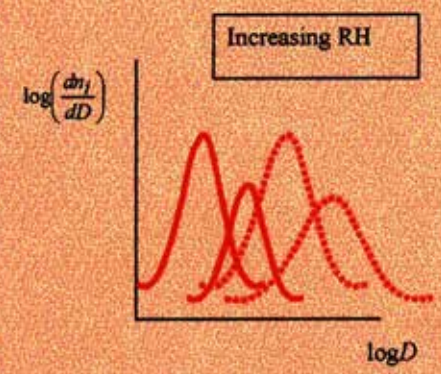
$$\frac{dn_i(D)}{dD} = \frac{N_i}{\log_e(10)\sqrt{2\pi}\sigma_i} \exp\left[-\frac{1}{2}\left(\frac{\log_{10}(D/D_i)}{\sigma_i}\right)^2\right]$$



Typical Shettle and Fenn Model

Two components. As RH increases, the components shift to larger size, and the "width" of the distribution changes.





Shettle and Fenn (1979) Aerosol Models

Shettle and Fenn particle properties

Small component:

- 70% water soluble
- 30% mineral

Large Component:

- Sea Salt

At 550 nm

Component	Size	m(RH=0)	m(RH=98%)	ω_0(RH=0)	ω_0(RH=98%)
Trop. (T)	Small	1.53-0.0066i	1.37-0.0012i	0.959	0.989
Oceanic (O)	Large	1.50	1.35	1	1

M: 99.0% T and 1% O by number

C: 99.5% T and 0.5% O by number

T: 100% T and 0% O by number

Aerosol Optical Property

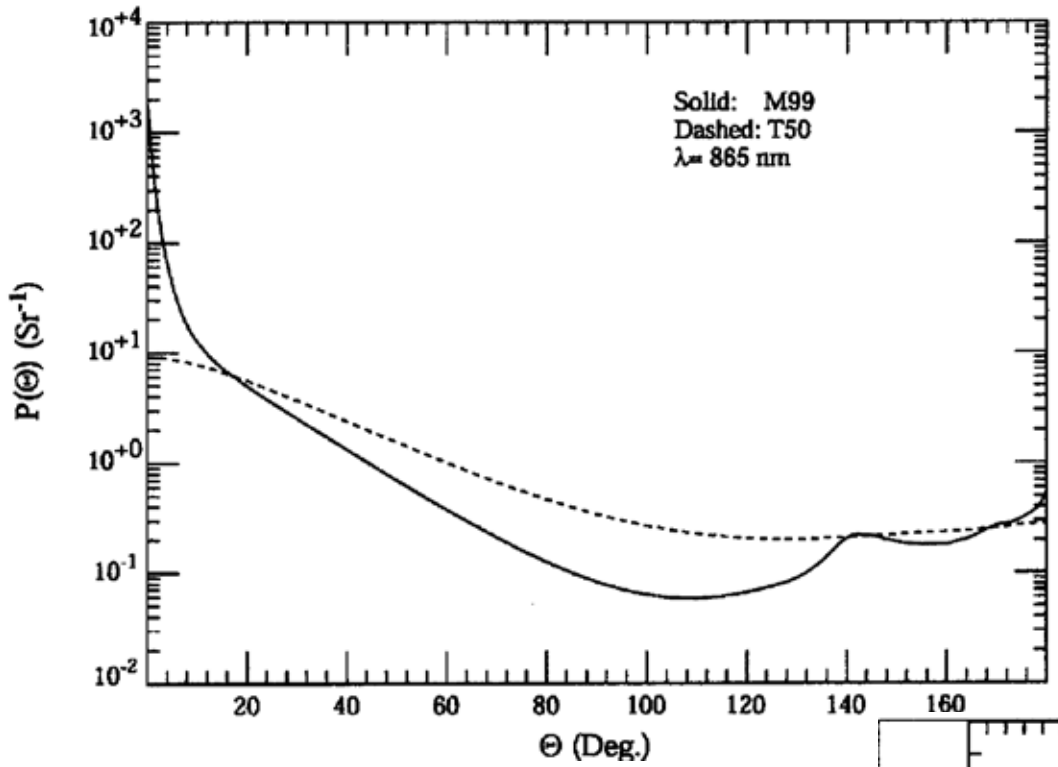
The complex refractive index of aerosol particles is:

$$m = m_r - i\, m_i,$$

m_r is the real refractive index, e.g., ~1.33 for water in the visible, and m_i the index related to particle absorption.

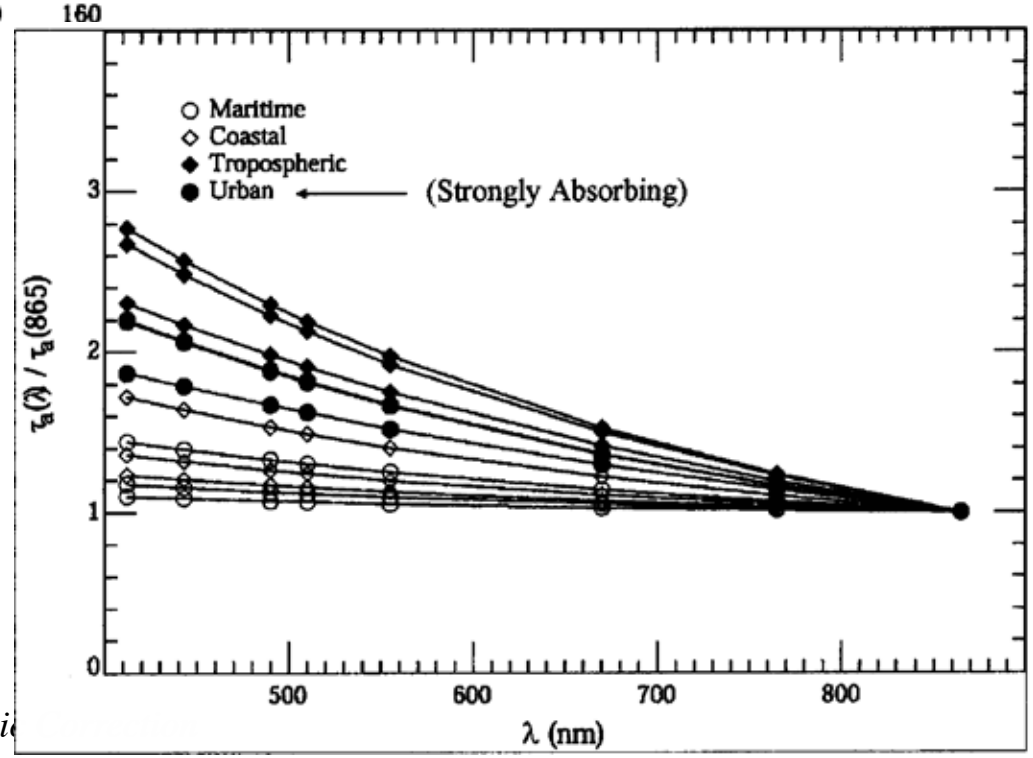
- Aerosol particles can be characterized with particle size distribution and particle refractive index.
- Given the size distribution, refractive index of the individual particles, and assuming that the particles are spherical, aerosol optical properties can be computed using the **Mie** theory. These properties are:
  - The scattering phase function $P(\Theta)$
  - The single scattering albedo ω_0
  - The extinction coefficient c

Shettle and Fenn (1979) Aerosol Models



Aerosol Phase Function

Aerosol Optical Thickness



The Ångström Exponent α (**particle size**)

$$\frac{\tau_a(\lambda)}{\tau_a(\lambda_0)} = \left(\frac{\lambda_0}{\lambda}\right)^{\alpha(\lambda,\lambda_0)}$$

$$\alpha(\lambda,\lambda_0) = \ln\left(\frac{\tau_a(\lambda)}{\tau_a(\lambda_0)}\right) \Big/ \ln\left(\frac{\lambda_0}{\lambda}\right)$$

Small aerosol particle size has large α value, while large particle has small α value.

The Asymmetry Parameter g (**phase function**)

$$g = \frac{1}{2}\int_0^{\pi} \cos\Theta P(\Theta)\sin\Theta\, d\Theta = \frac{1}{2}\int_{-1}^{1} \mu P(\mu)\, d\mu$$

The value of g ranges between -1 for entirely backscattered light to +1 for entirely forward-scattered light. Generally, $g = 0$ indicates scattering direction evenly distributed between forward and backward directions, e.g., **isotropic scattering**.

Aerosol Models

Shettle and Fenn (1979) aerosol models used (12 models):
Oceanic with relative humidity (RH) of 99%
Maritime with RH of 50%, 70%, 90%, and 99%
Coastal with RH of 50%, 70%, 90%, and 99%
Tropospheric with RH of 50%, 90%, and 99%

Other models from **AERONET** measurements (Ahmad et al., 2010)

Characteristics of the Aerosol Models

Aerosol Model	Single Scattering Albedo $\omega_a(865)$	Asymmetry Parameter g	Ångström Exponent $\alpha(510, 865)$
Oceanic†	1.0	0.724-0.840	-0.087~ -0.016
Maritime†	0.982-0.999	0.690-0.824	0.09-0.50
Coastal††	0.976-0.998	0.682-0.814	0.23-0.76
Tropospheric†	0.930-0.993	0.603-0.769	1.19-1.53
Urban†	0.603-0.942	0.634-0.778	0.85-1.14
Dust†††	0.836-0.994	0.662-0.763	0.29-0.36

† Shettle and Fenn (1979) aerosol models. ††Gordon and Wang (1994)
††† Shettle (1984) and Moulin et al. (2001).

Strongly-absorbing aerosols

Radiative Transfer

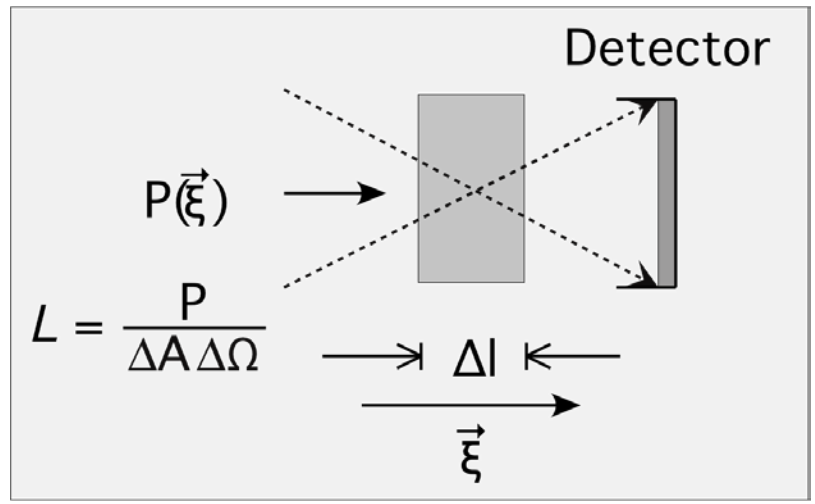
4. Radiative Transfer
   - Radiative Transfer Equation (RTE)
   - Successive-order-of-scattering method
   - Single-scattering approximation

The Radiative Transfer Equation (**RTE**)

The RTE is simply an accounting for the changes in radiance along a path.

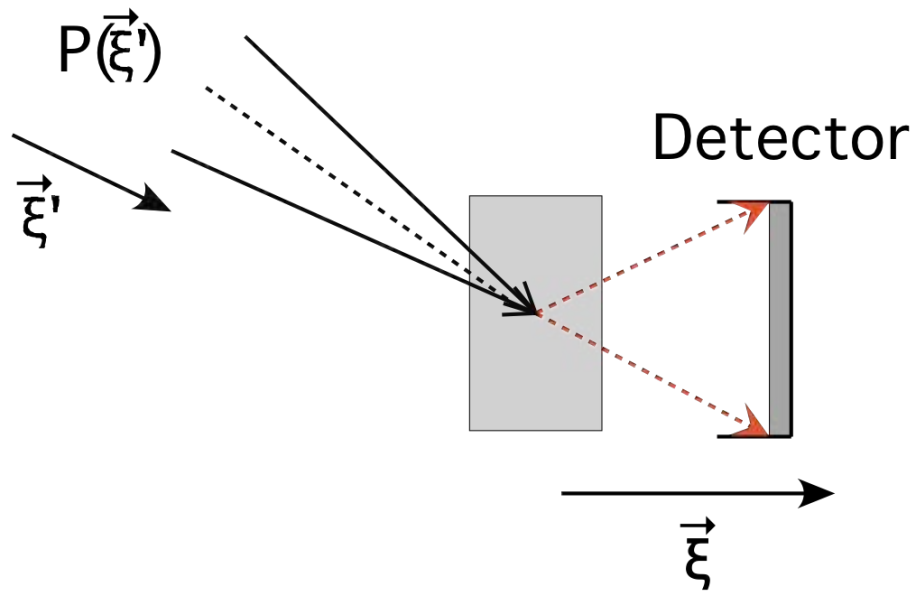
First, attenuation produces a loss:

$$L + \Delta L = \frac{P + \Delta P}{\Delta A \, \Delta \Omega} = \frac{P\left(1 - c\Delta l\right)}{\Delta A \, \Delta \Omega}$$



$$\frac{dL\left(\hat{\xi}\right)}{dl} = -c\,L\left(\hat{\xi}\right)$$

Second, scattering produces a gain:

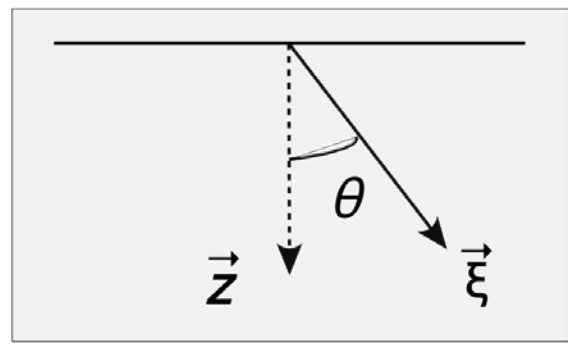


$$\frac{dL\left(\hat{\xi}\right)}{dl} = \int_{\Omega'} \beta\left(\hat{\xi}' \rightarrow \hat{\xi}\right) L\left(\hat{\xi}'\right) d\Omega\left(\hat{\xi}'\right)$$

The RTE:

$$\frac{dL\left(\hat{\xi}\right)}{dl} = -cL\left(\hat{\xi}\right) + \int_{\Omega'} \beta\left(\hat{\xi}' \to \hat{\xi}\right) L\left(\hat{\xi}'\right) d\Omega\left(\hat{\xi}'\right)$$

For the atmosphere-ocean system:



$$\cos\theta \frac{dL(z,\theta,\phi)}{dz} = -c(z)L(z,\theta,\phi) + \int_{\Omega'} \beta(z,\theta',\phi' \to \theta,\phi) L(z,\theta',\phi') d\Omega'$$

The RTE:

$$\cos\theta \frac{dL(z,\theta,\phi)}{dz} = -c(z)L(z,\theta,\phi) + \int_{\Omega'} \beta(z,\theta',\phi' \to \theta,\phi) L(z,\theta',\phi') d\Omega'$$

Can be written as:

$$\cos\theta \frac{dL(z,\theta,\phi)}{c(z)dz} = -L(z,\theta,\phi) + \omega_0(z) \int_{\Omega'} P(z,\theta',\phi' \to \theta,\phi) L(z,\theta',\phi') d\Omega'$$

$$\cos\theta \frac{dL(\tau,\theta,\phi)}{d\tau} = -L(\tau,\theta,\phi) + \omega_0(\tau) \int_{\Omega'} P(\tau,\theta',\phi' \to \theta,\phi) L(\tau,\theta',\phi') d\Omega'$$

where $\tau = \int_0^Z c(z')dz'$ is the optical thickness

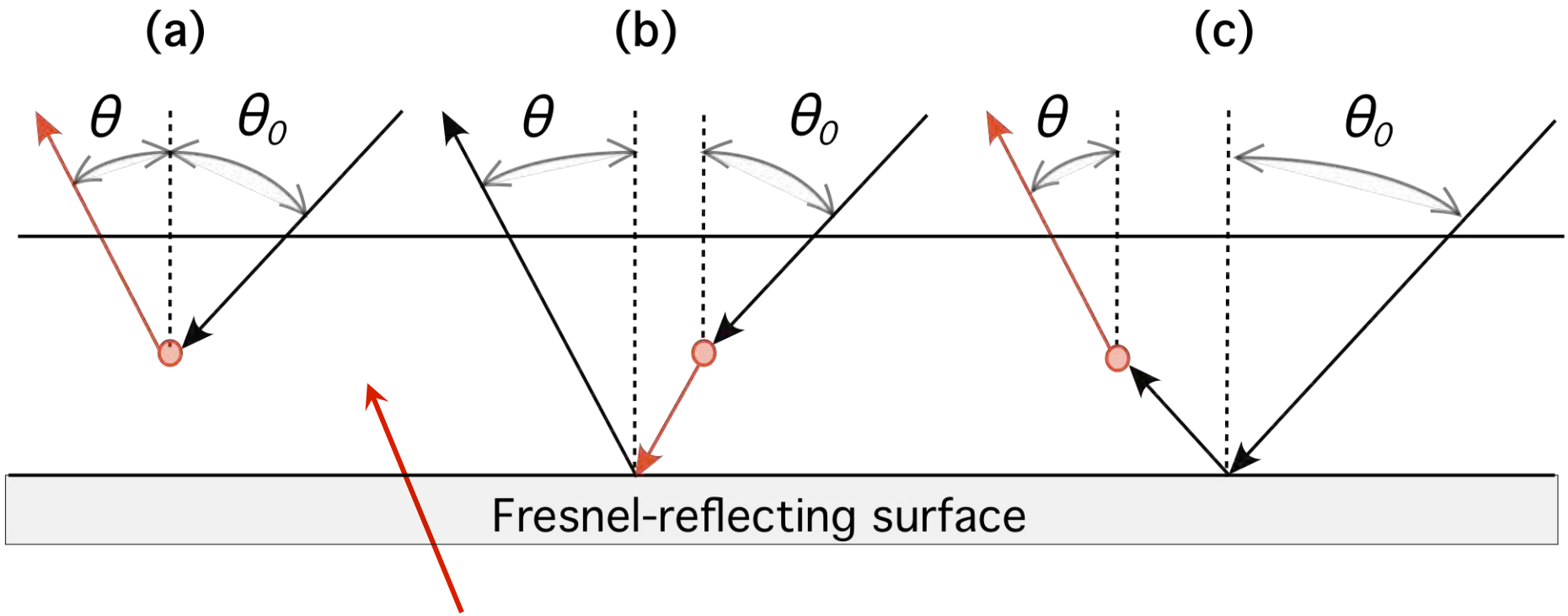
Successive-order-of-scattering method

Fourier Decomposition of the RTE

See section from Wang (1991).

Wang, M. (1991), "Atmospheric Correction of the Second Generation of Ocean Color Sensors," *Ph.D. Dissertation,* 135 pp., University of Miami, Coral Gables, Florida, USA.

Single-Scattering Approximation (1)



Atmosphere property:

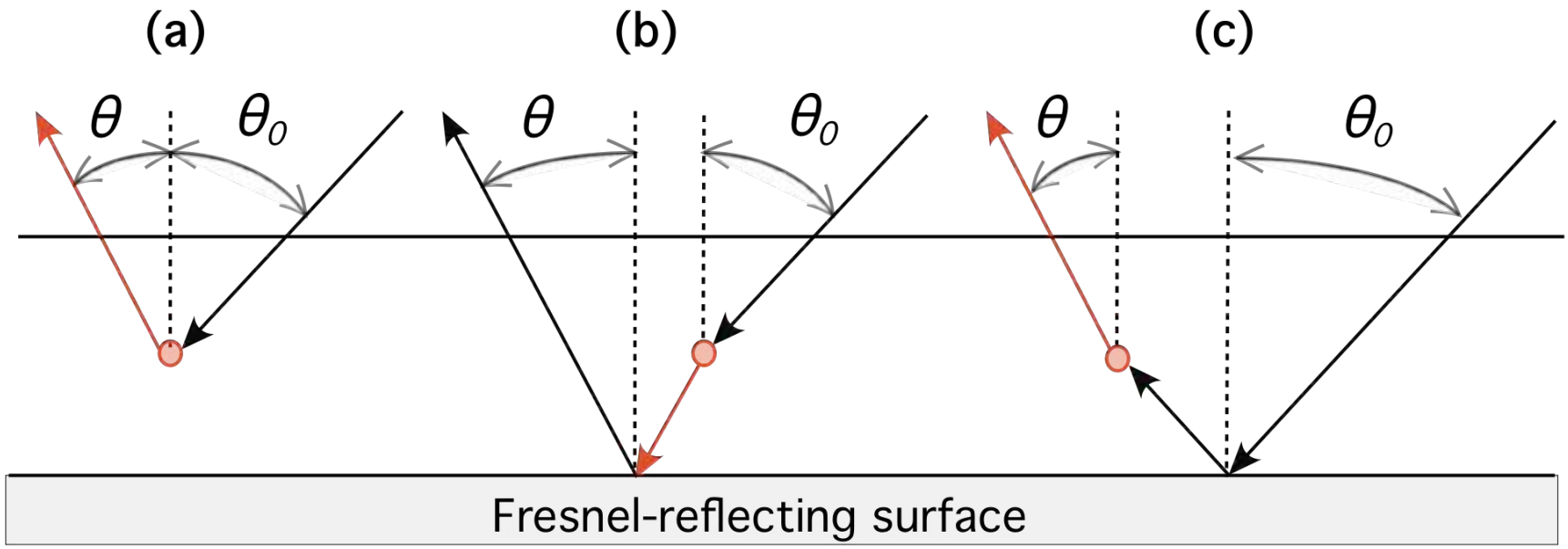
Optical thickness τ, $L \sim \tau$

Single-scattering albedo ω_0, $L \sim \omega_0$

Phase function $P(\Theta)$, $L \sim P(\Theta)$

Wang, M. (1991), "Atmospheric Correction of the Second Generation of Ocean Color Sensors," *Ph.D. Dissertation,* 135 pp., University of Miami, Coral Gables, Florida, USA.

Single-Scattering Approximation (2)



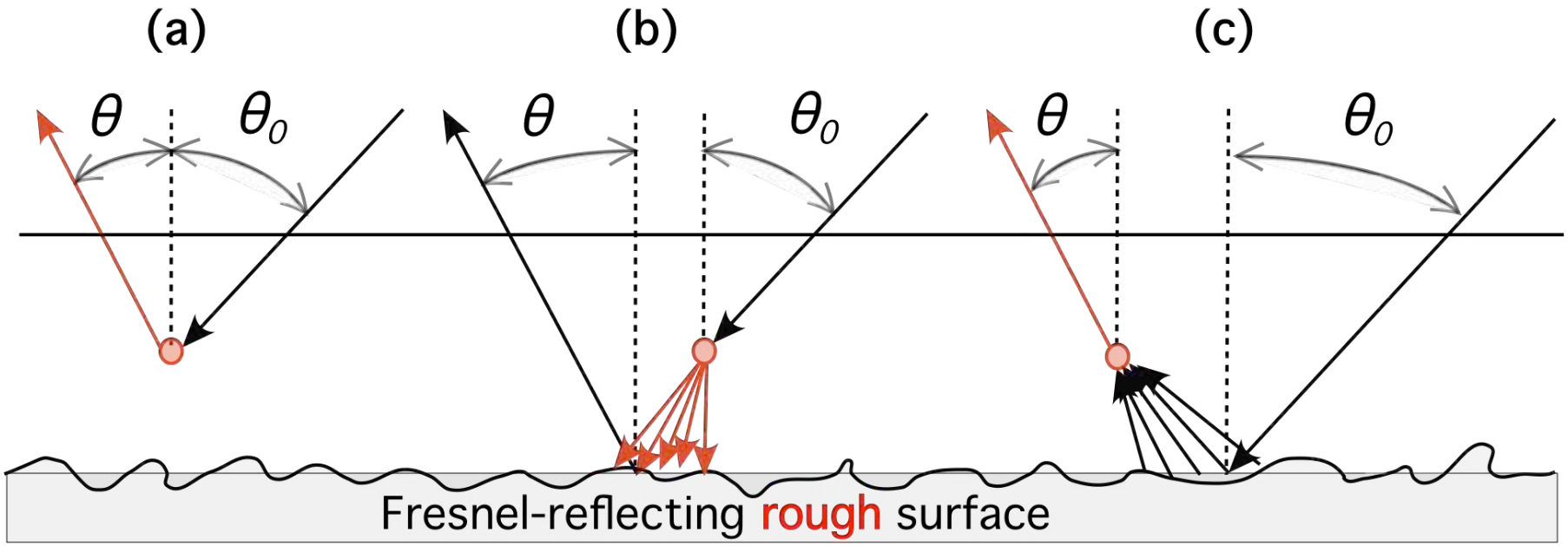
$$\rho_{as}(\lambda) = \frac{\omega_0(\lambda)\tau(\lambda)}{4\cos\theta\cos\theta_0}\left[P(\Theta_-,\lambda) + \left(r(\theta) + r(\theta_0)\right)P(\Theta_+,\lambda)\right]$$

$$\propto \omega_0(\lambda)\tau(\lambda)P(\Theta_-,\lambda)$$

(a) **(b)** **(c)**

Fresnel reflectivity of the surface

Single-Scattering Approximation (3)



See Single Scattering section from Wang (1991).

Wang, M. (1991), "Atmospheric Correction of the Second Generation of Ocean Color Sensors," *Ph.D. Dissertation,* 135 pp., University of Miami, Coral Gables, Florida, USA.

$$\rho_{as}(\lambda) = \frac{\omega_0(\lambda)\tau(\lambda)}{4\cos\theta\cos\theta_0}\left[P(\Theta_-,\lambda) + \left(r(\theta) + r(\theta_0)\right)P(\Theta_+,\lambda)\right]$$

$$\propto \omega_0(\lambda)\tau(\lambda)P(\Theta_-,\lambda)$$

Aerosol reflectance (radiance)

$$\propto \omega_a(\lambda)\tau_a(\lambda)P_a(\Theta_-,\lambda)$$

For non- and weakly absorbing aerosols, aerosol reflectance: $\propto \tau_a(\lambda)P_a(\Theta_-,\lambda)$

AOT can be routinely measured, but NOT Phase Function

Aerosol Single-Scattering Epsilon

$$\varepsilon(\lambda,\lambda_0) = \frac{\rho_{as}(\lambda)}{\rho_{as}(\lambda_0)}$$

$$= \frac{\omega_a(\lambda)\tau_a(\lambda)\left[P_a(\Theta_-,\lambda) + \left(r(\theta) + r(\theta_0)\right)P_a(\Theta_+,\lambda)\right]}{\omega_a(\lambda_0)\tau_a(\lambda_0)\left[P_a(\Theta_-,\lambda_0) + \left(r(\theta) + r(\theta_0)\right)P_a(\Theta_+,\lambda_0)\right]}$$

Spectral variation of aerosol reflectance

Characteristics of the Aerosol Models

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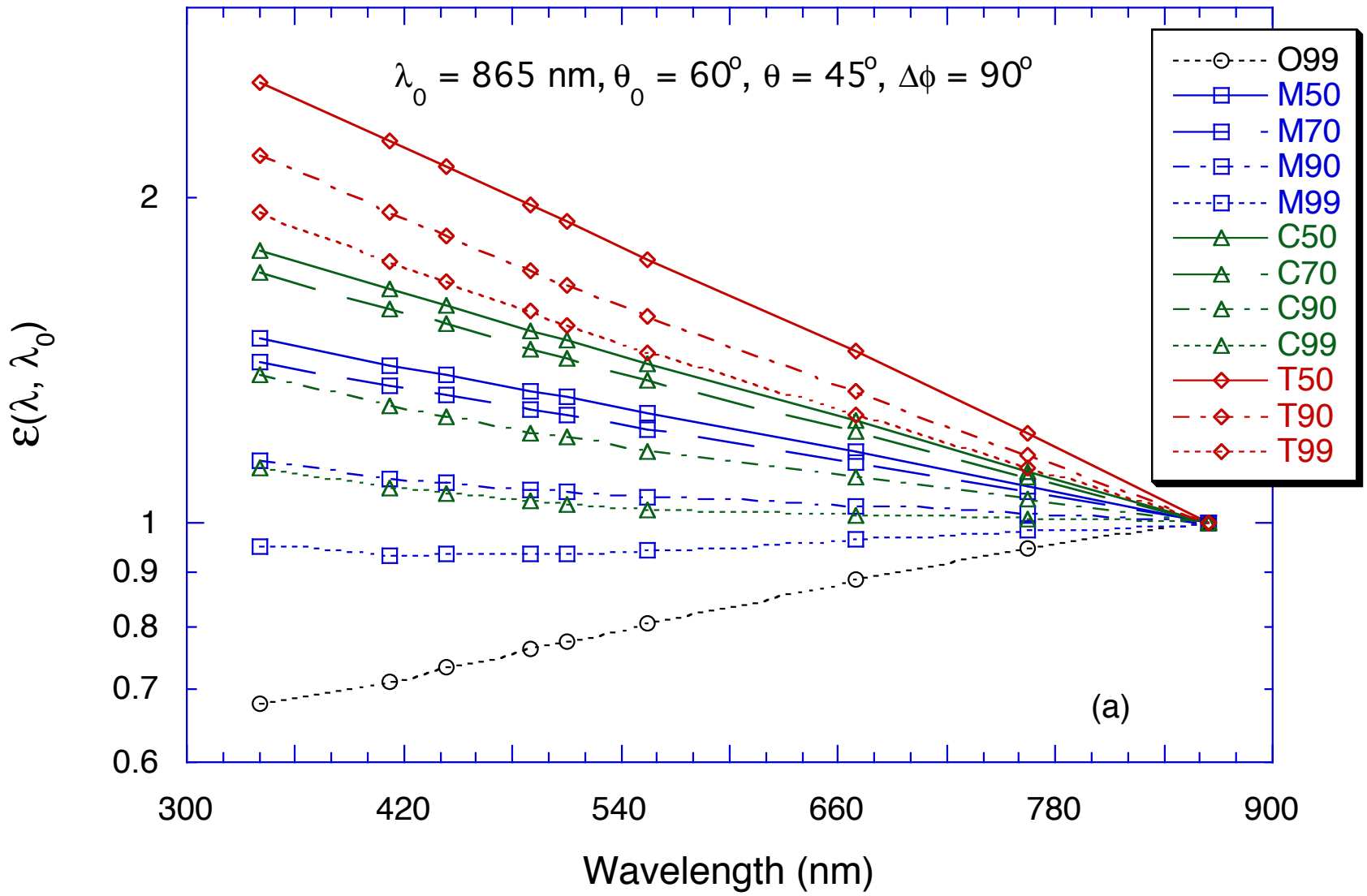
Strongly-absorbing aerosols

Non- and weakly-absorbing aerosols

Aerosol Single-Scattering Epsilon (λ_0 = 865 nm)



Epsilon and Aerosol Reflectance

$$\varepsilon(\lambda,\lambda_0) = \frac{\rho_{as}(\lambda)}{\rho_{as}(\lambda_0)} \approx \exp\{c(\lambda_0 - \lambda)\}$$

$\varepsilon(\lambda, \lambda_0)$ can be computed from two bands (e.g., NIR bands) $\rho_{as}(\lambda)$. Thus, $\rho_{as}(\lambda)$ values in other wavelengths (e.g., visible bands) can be estimated:

$$\rho_{as}(\lambda) = \varepsilon(\lambda,\lambda_0) \times \rho_{as}(\lambda_0)$$

A simple **atmospheric correction** (aerosol correction) scheme (Wang and Gordon, 1994)--Single-scattering approximation.

Questions?

Radiance L ~ ?