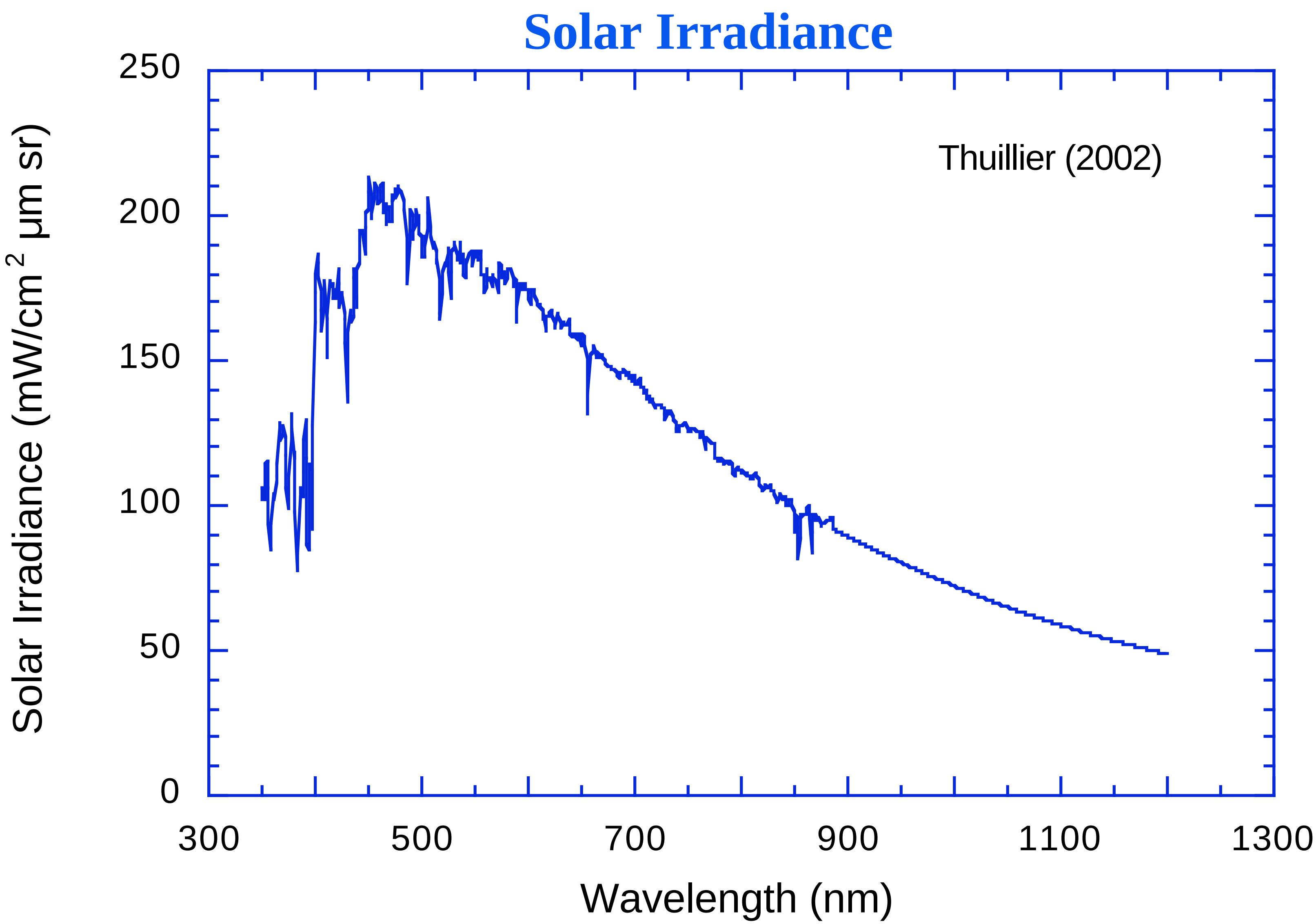
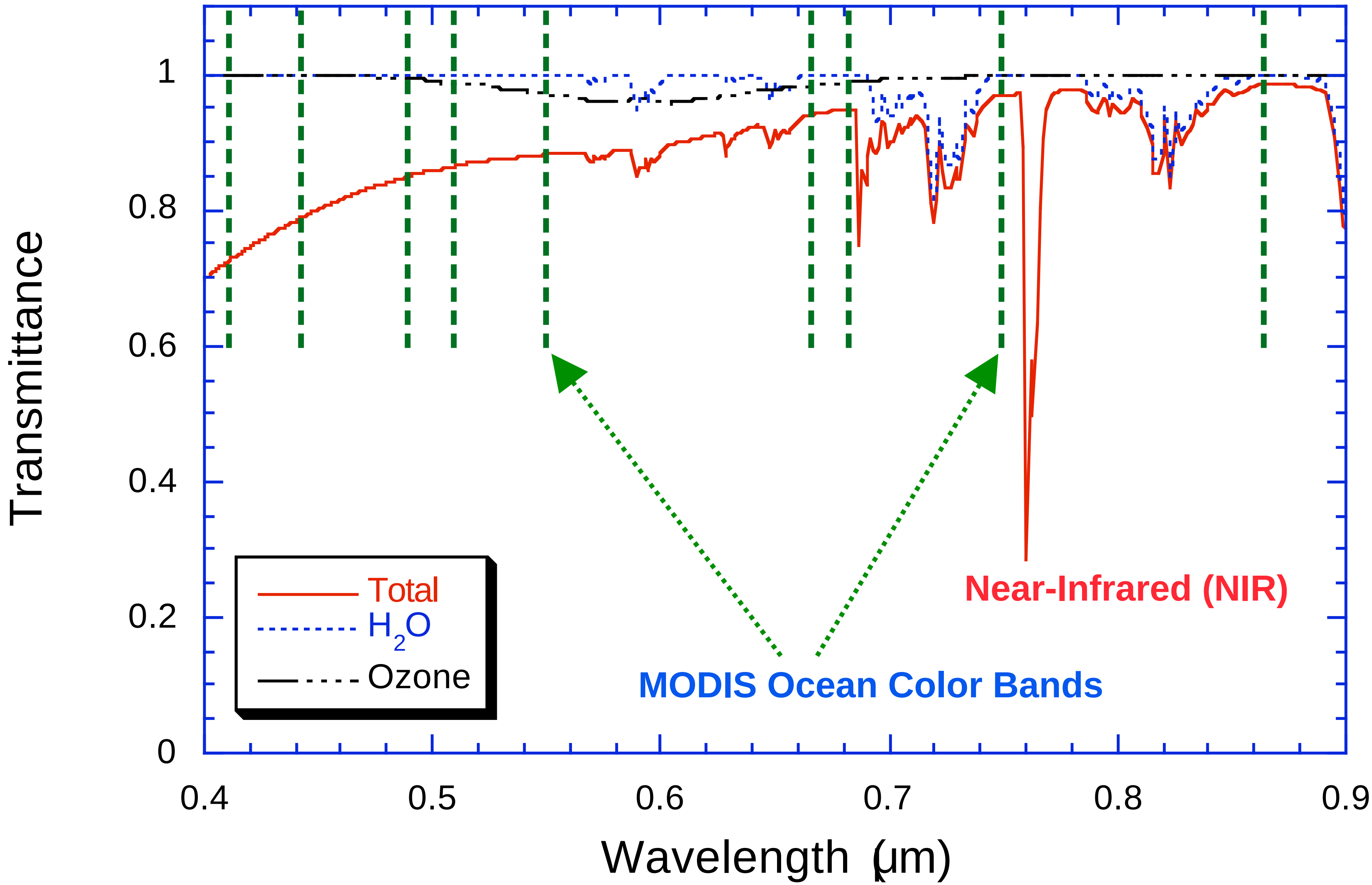
Atmospheric Correction

5. Atmospheric correction
   - Define reflectance and examine 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 overview of various approaches
     - The SWIR-based atmospheric correction
   - Examples from MODIS-Aqua measurements



Passive Remote Sensing: Sensor-measured signals are all originated from the sun!

Atmospheric Windows (VIS-NIR)



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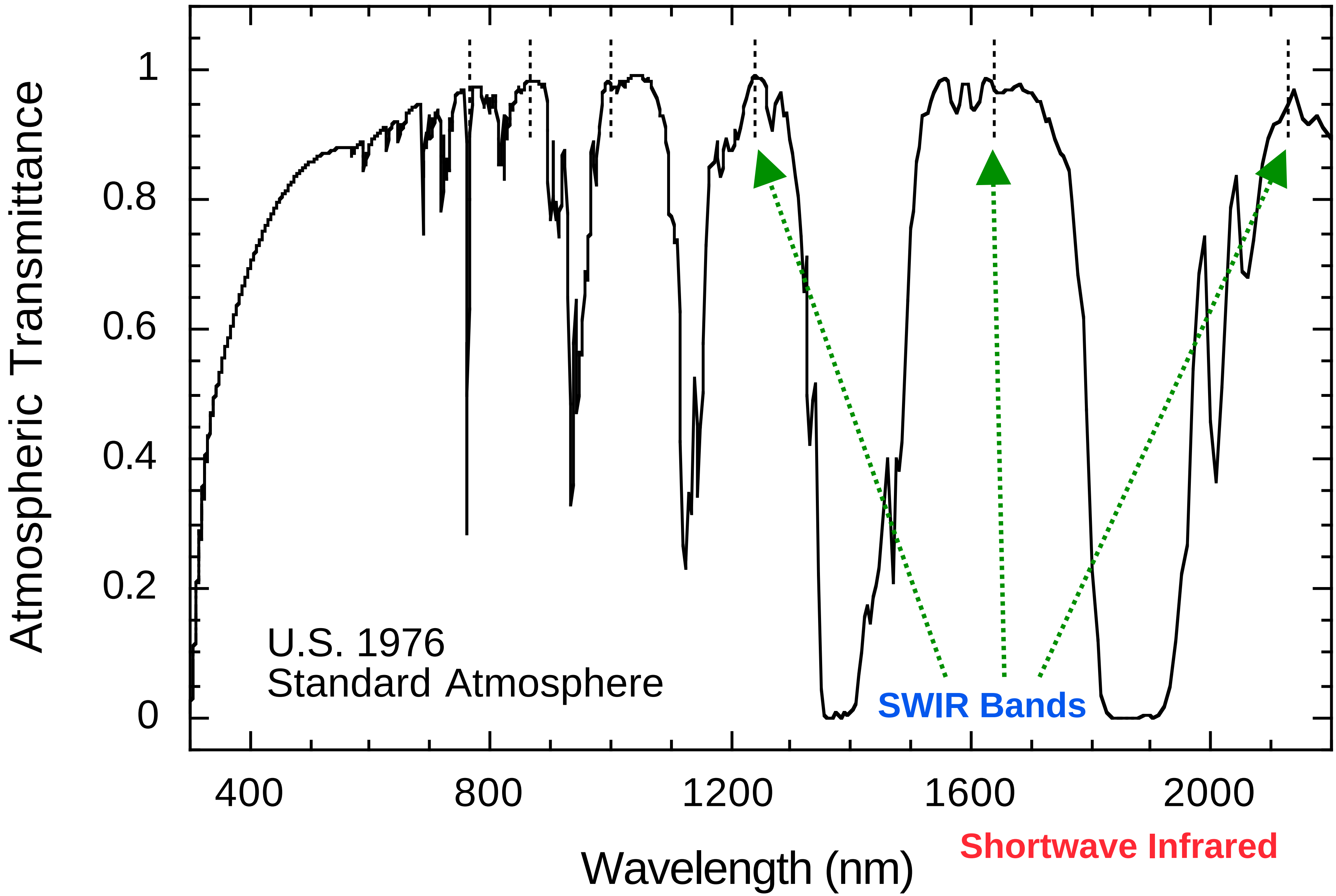
Atmospheric Windows



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The Ocean Color and Other Useful Spectral Bands for VIIRS, MODIS, and SeaWiFS

VIIRS		MODIS		SeaWiFS
Ocean Bands (nm)	Other Bands (nm)	Ocean Bands (nm)	Other Bands (nm)	Ocean Band (nm)
410		412	645	412
443		443	859	443
486		488	469	490
—		531	555	510
551	*SWIR Bands*	551	*SWIR Bands*	555
671	1238	667	1240	670
745	1610	748	1640	765
862	2250	869	2130	865

VIIRS has similar SWIR bands as **MODIS**

Satellite-Measured Reflectance (Radiance)

$$\rho_t(\lambda) = \underbrace{\rho_r(\lambda)}_{\text{Rayleigh}} + \underbrace{\rho_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{\rho_{wc}(\lambda)}_{\text{Whitecap}} + T(\lambda)\underbrace{\rho_g(\lambda)}_{\text{Sun Glint}}$$

$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Transmittance}}\underbrace{\left[\rho_w(\lambda)\right]_N}_{\text{Ocean}}$$

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

$$\rho_A(\lambda) = \rho_a(\lambda) + \underbrace{\rho_{ra}(\lambda)}_{\text{Rayleigh-aerosol}}$$

$$\rho_t(\lambda) = \underbrace{\rho_r(\lambda)}_{\text{Rayleigh}} + \underbrace{\rho_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{\rho_{wc}(\lambda)}_{\text{Whitecap}} + T(\lambda)\underbrace{\rho_g(\lambda)}_{\text{Sun Glint}}$$

$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Transmittance}}\underbrace{\left[\rho_w(\lambda)\right]_N}_{\text{Ocean}}$$

Atmospheric correction: from sensor-measured $\boldsymbol{\rho_t(\lambda)}$ spectra to derive $[\boldsymbol{\rho_w(\lambda)}]_N$ spectra.

It is for correction of reflectance (radiance) ~ $\omega_0\boldsymbol{(\lambda)}\tau\boldsymbol{(\lambda)}P(\Theta,\lambda)$

Algorithms for Various Ocean Color Sensors

(Routine Global Ocean Color Data Processing)

- **Gordon and Wang** (1994) for **SeaWiFS** and **MODIS** ocean color products.
- **Fukushima** et al. (1998) for **OCTS** and **GLI** ocean color products.
- **Antoine and Morel** (1999) for **MERIS** ocean color products.
- **Deschamps** et al. (1999) for **POLDER** ocean color products.

IOCCG, Atmospheric Correction for Remotely-Sensed Ocean-Color Products, Wang, M. (ed.), *Reports of International Ocean-Color Coordinating Group*, No. 10, IOCCG, Dartmouth, Canada, 2010.

Atmospheric Correction Algorithm

MODIS and SeaWiFS algorithm (Gordon and Wang 1994)

$$\rho_t = \rho_r + \rho_A + t\rho_{wc} + T\rho_g + t\rho_w, \quad \rho = \pi L / \mu_0 F_0$$

- **ρ_w is the desired quantity in ocean color remote sensing.**
- $T\rho_g$ is the sun glint contribution—avoided/masked/corrected.
- $T\rho_{wc}$ is the whitecap reflectance—computed from wind speed.
- ρ_r is the scattering from molecules—computed using the Rayleigh lookup tables (vector RTE, wind speed, atmospheric pressure dependents).
- $\rho_A = \rho_a + \rho_{ra}$ is the aerosol and Rayleigh-aerosol contributions —estimated using **aerosol models**.
- For Case-1 waters at the open ocean, ρ_w is usually negligible at 750 & 865 nm. ρ_A can be estimated using these two NIR bands. Ocean is usually not black at NIR for the coastal regions.

Gordon, H. R. and M. Wang, "Retrieval of water-leaving radiance and aerosol optical thickness over the oceans with SeaWiFS: A preliminary algorithm," *Appl. Opt.*, **33**, 443-452, 1994.

$$\rho_t(\lambda) = \underbrace{\rho_r(\lambda)}_{\text{Rayleigh}} + \underbrace{\rho_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{\rho_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\ \underbrace{\rho_g(\lambda)}_{\text{Sun Glint}}$$

$$+\ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}}\ \underbrace{\left[\rho_w(\lambda)\right]_N}_{\text{Ocean}}$$

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\ \underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

$$+\ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}}\ \underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

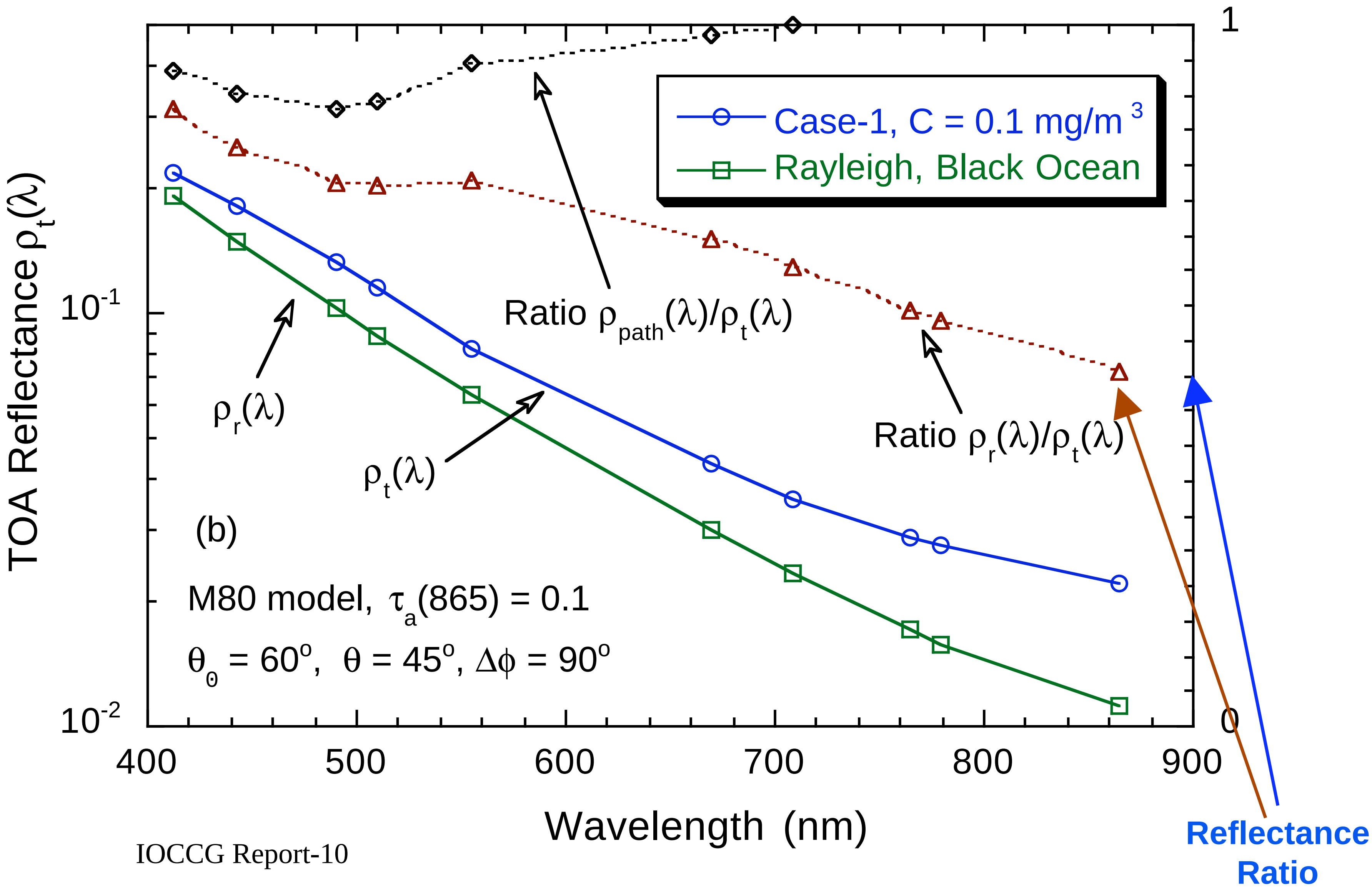
We look at these terms…..

Rayleigh Scattering Radiance

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}} \underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}} \underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

Ocean TOA Radiance Contributions for Case-1 Waters



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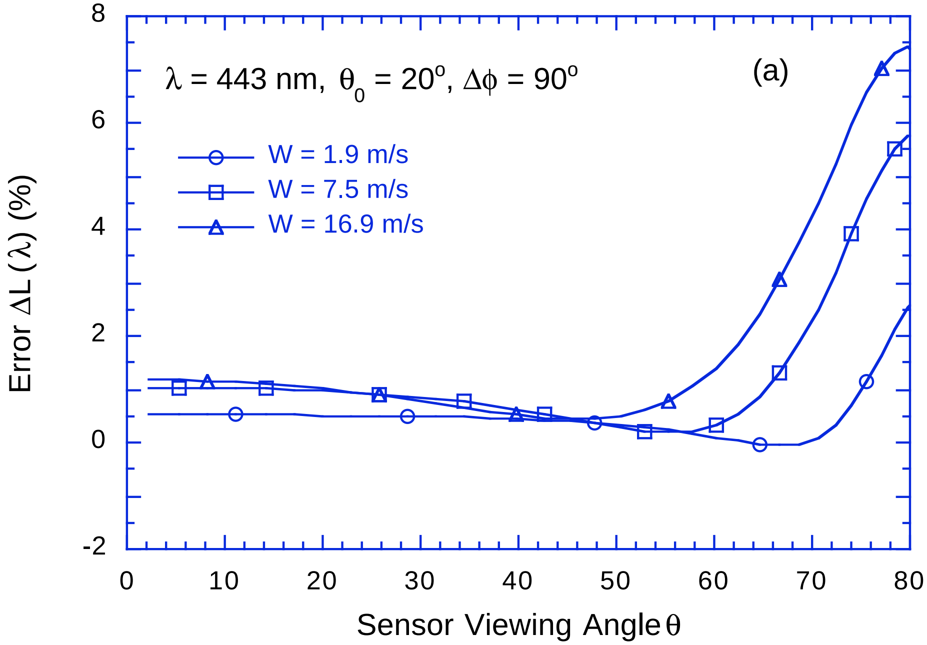
Rayleigh Lookup Tables

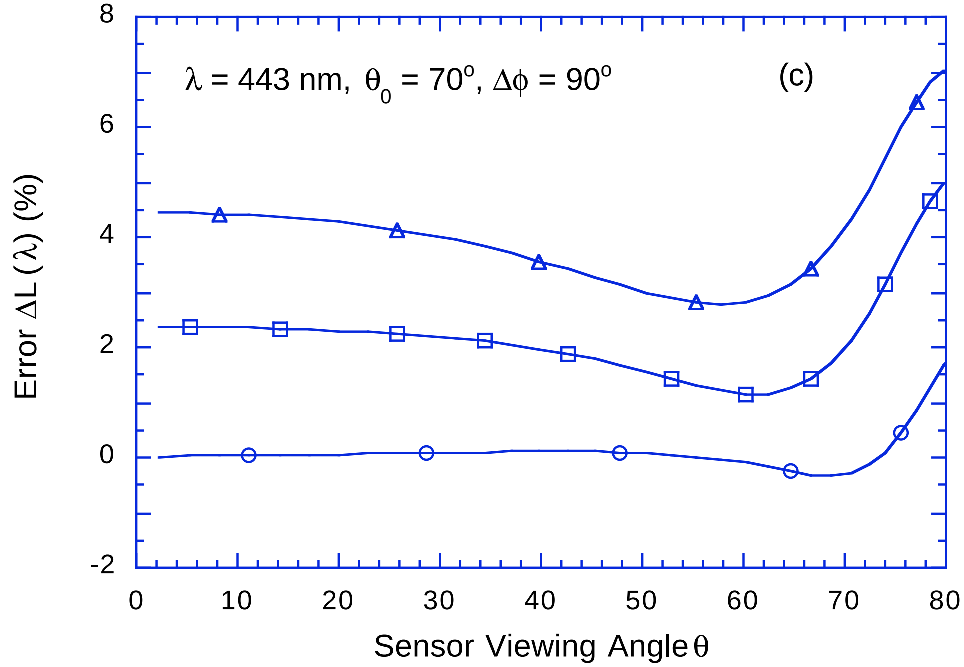
The Rayleigh lookup tables were generated using the method developed by Gordon and Wang (1992) for the various surface wind speeds. Cox and Munk (1954) model was used for sea surface roughness.

- **Vector** (polarization) RTE used for SeaWiFS/MODIS spectral bands.
- Solar-zenith angles 0^{o}-80^{o} (2^{o}), sensor-zenith angles 0^{o}-84^{o} (~2^{o}).
- They are stored as Fourier coefficients for computing any azimuth angles.
- **8 wind speeds**: 0, 1.9, 4.2, 7.5, 11.7, 16.9, 22.9, and 30.0 m/s.
- Improved atmospheric pressure correction algorithm for Rayleigh radiance computation.
- Need ancillary data of wind speed and atmospheric pressure.

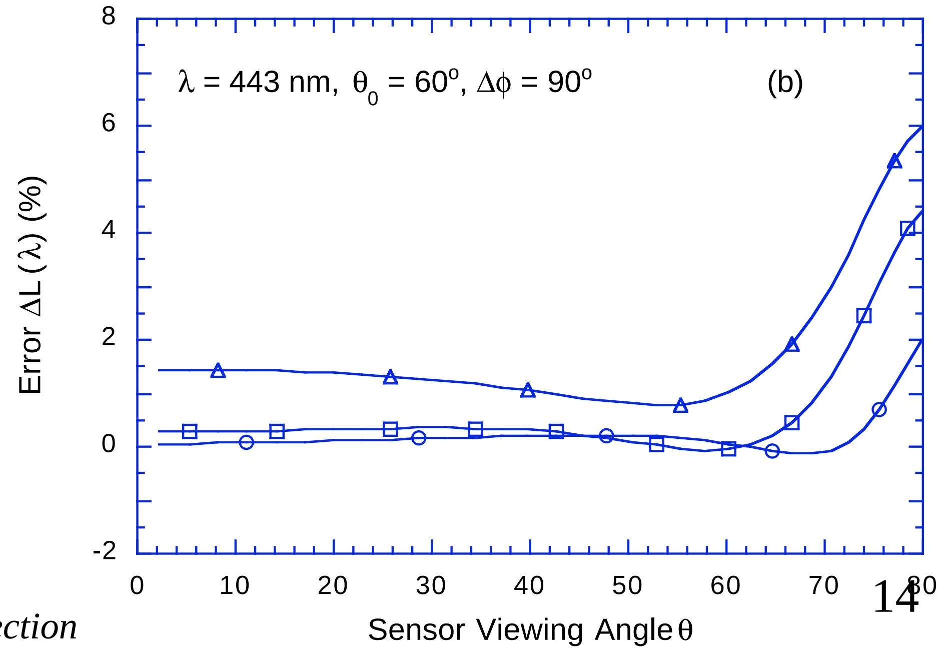
Rayleigh scattering radiance can be computed accurately.

Wind Effects on Rayleigh Radiance





$$\text{Error } \Delta L(\lambda) = L_r(\lambda, W\!=\!0) - L_r(\lambda, W\!=\!0)$$



Gordon and Wang (1992) and Wang (2002)

Atmospheric pressure effects

For atmosphere pressure changed from P_0 to $P_0 + \Delta P$, Gordon et al. (1988) found through trial and error that the Rayleigh radiance can be estimated as:

$$L_r\left[\tau_r\left(\lambda, P_0 + \Delta P\right)\right] = L_r\left[\tau_r\left(\lambda, P_0\right)\right] \frac{1 - \exp\left[-\tau_r\left(\lambda, P_0 + \Delta P\right)/\cos\theta\right]}{1 - \exp\left[-\tau_r\left(\lambda, P_0\right)/\cos\theta\right]}$$

$$\tau_r\left(\lambda, P_0 + \Delta P\right) = \tau_r\left(\lambda, P_0\right) \frac{P_0 + \Delta P}{P_0}$$

Where is θ sensor-zenith angle and $P_0 = 1013.25$ hPa.

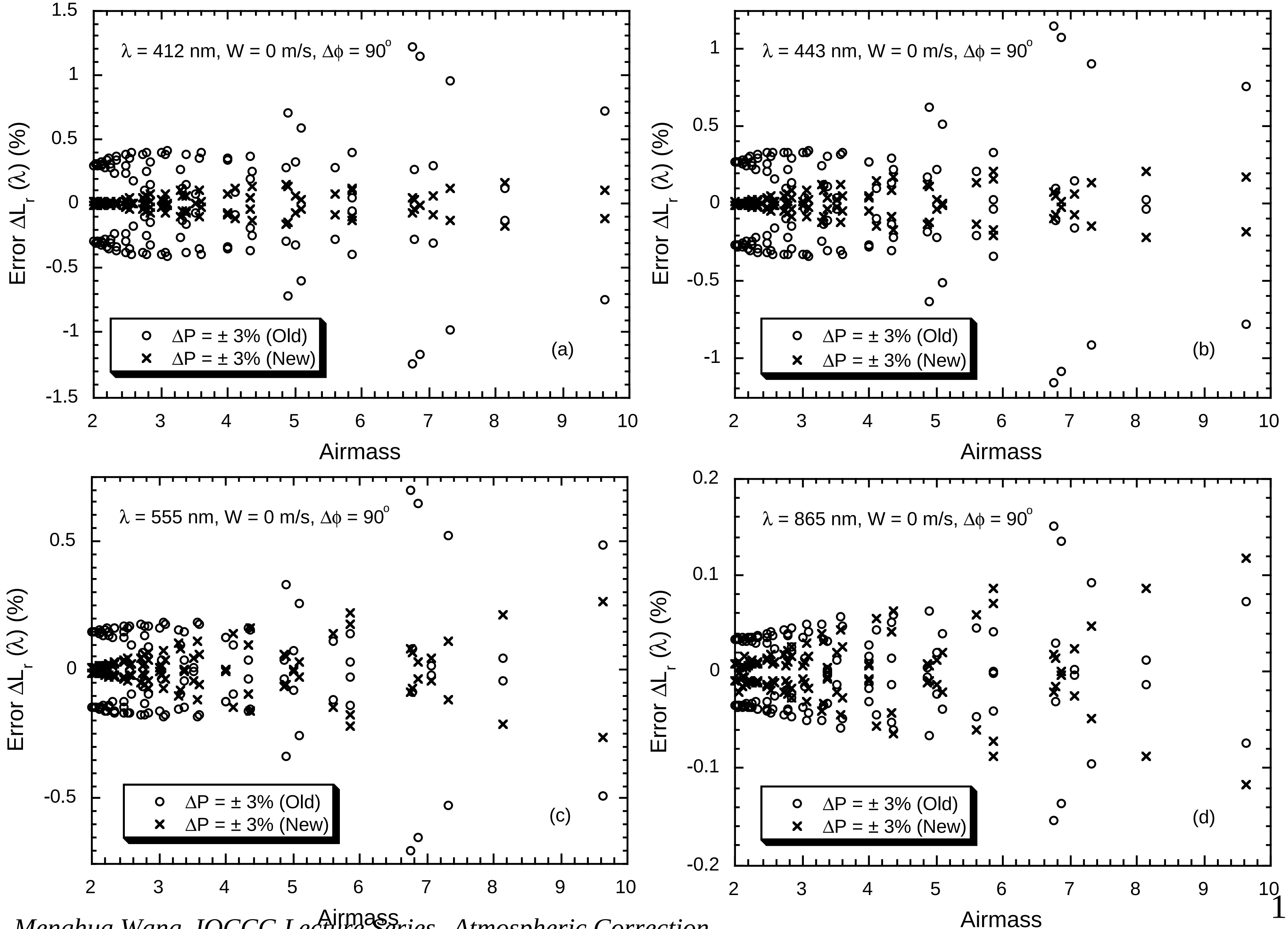
Wang (2005) improved this computation with assumption of

$$L_r\left[\tau_r\left(\lambda, P_0 + \Delta P\right)\right] = L_r\left[\tau_r\left(\lambda, P_0\right)\right] \frac{1 - \exp\left[-C\left(\lambda, M\right)\tau_r\left(\lambda, P_0 + \Delta P\right)M\right]}{1 - \exp\left[-C\left(\lambda, M\right)\tau_r\left(\lambda, P_0\right)M\right]}$$

$M = 1/\cos\theta_0 + 1/\cos\theta$ is air mass value.

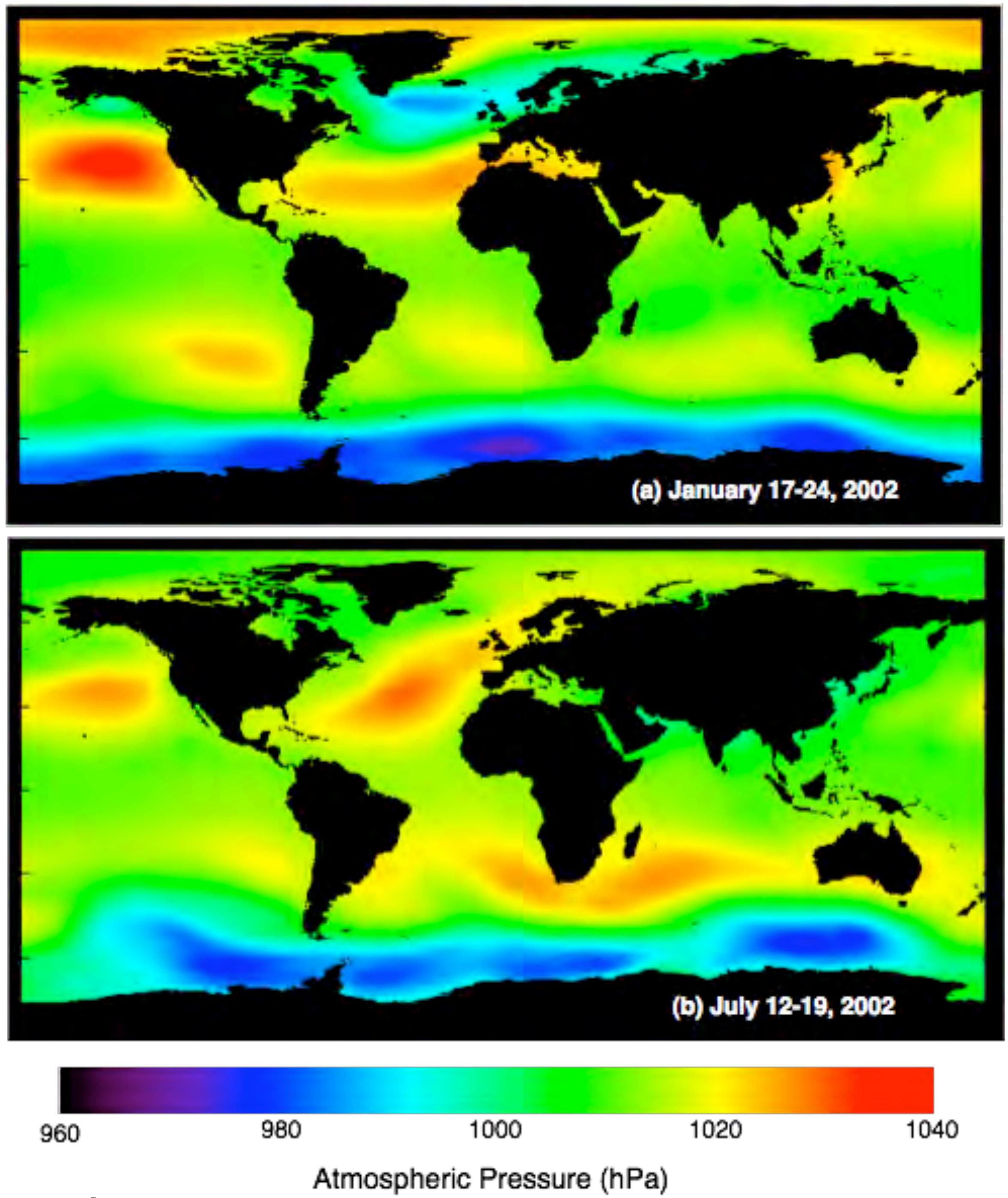
$$C\left(\lambda, M\right) = a\left(\lambda\right) + b\left(\lambda\right)\log_e\left(M\right), \text{ and } \quad \begin{aligned} a\left(\lambda\right) &= -0.6543 + 1.608\,\tau_r\left(\lambda\right) \\ b\left(\lambda\right) &= 0.8192 - 1.2541\,\tau_r\left(\lambda\right) \end{aligned}$$

Effects of Atmospheric Pressure Variation on Rayleigh Radiance

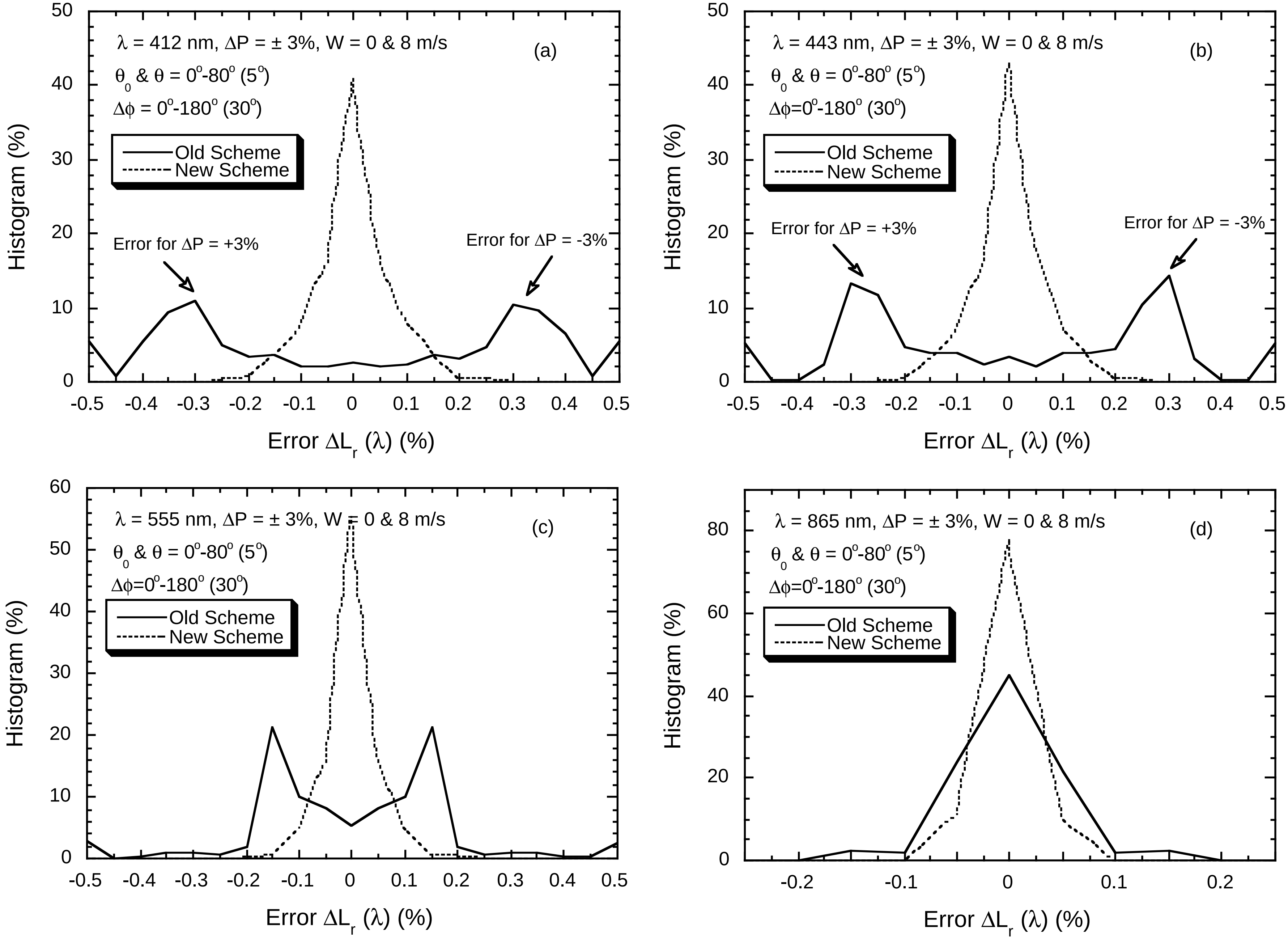


Atmospheric Pressure Variation

Usually, it is within ~±3%



Effects of Atmospheric Pressure Variation on Rayleigh Radiance



Ocean Whitecap Contribution

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\ \underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

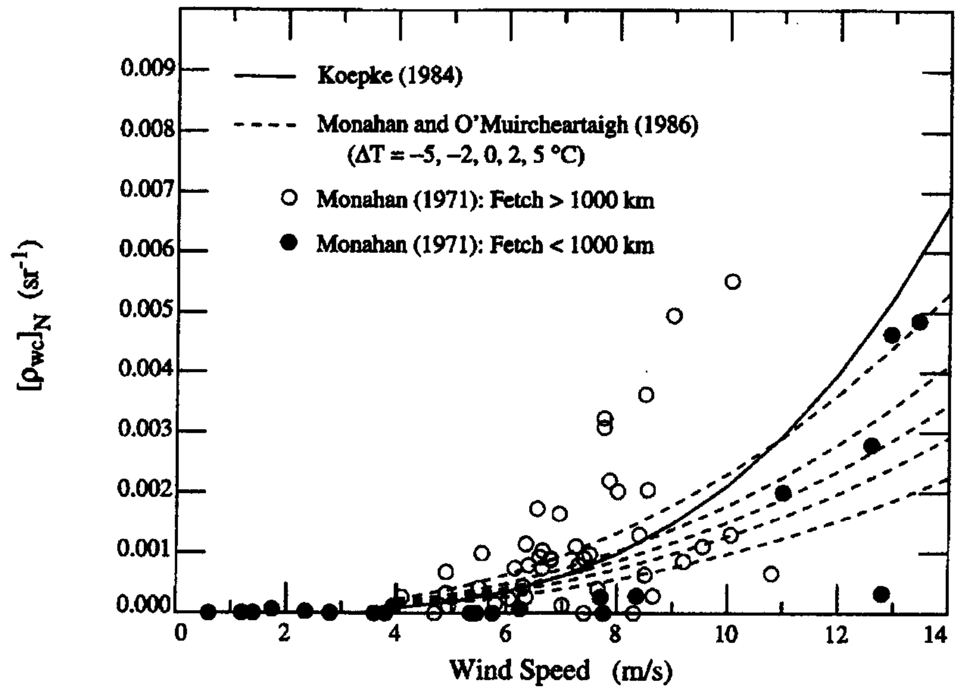
$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}}\ \underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

Ocean Whitecap Contribution

Following Koepke(1984), Gordon and Wang (1994b) proposed ocean whitecap reflectance to be

$$\left[\rho_{wc}\right]_N = 6.5\times 10^{-7} W^{3.52}, \text{ the TOA contribution is } tt_0\left[\rho_{wc}\right]_N$$

Frouin et al. (1996) and *Moore et al.* (1998) found that contrary to the previous measurements, the whitecap reflectance is spectrally dependent. The reflectance contributions are significantly smaller at the near-infrared than in the visible, because of the stronger ocean water absorption at the longer wavelengths that reduces the reflected photons from the submerged bubbles. *Moore et al.* (2000) also found much less whitecap reflectance contributions.



- SeaWiFS/MODIS use Frouin et al. (1996) spectral variation with ~1/4 of the Koepke (1984) reflectance, and kept maximum value at *W*=8 m/s.

Sun Glint Contribution

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}} \underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

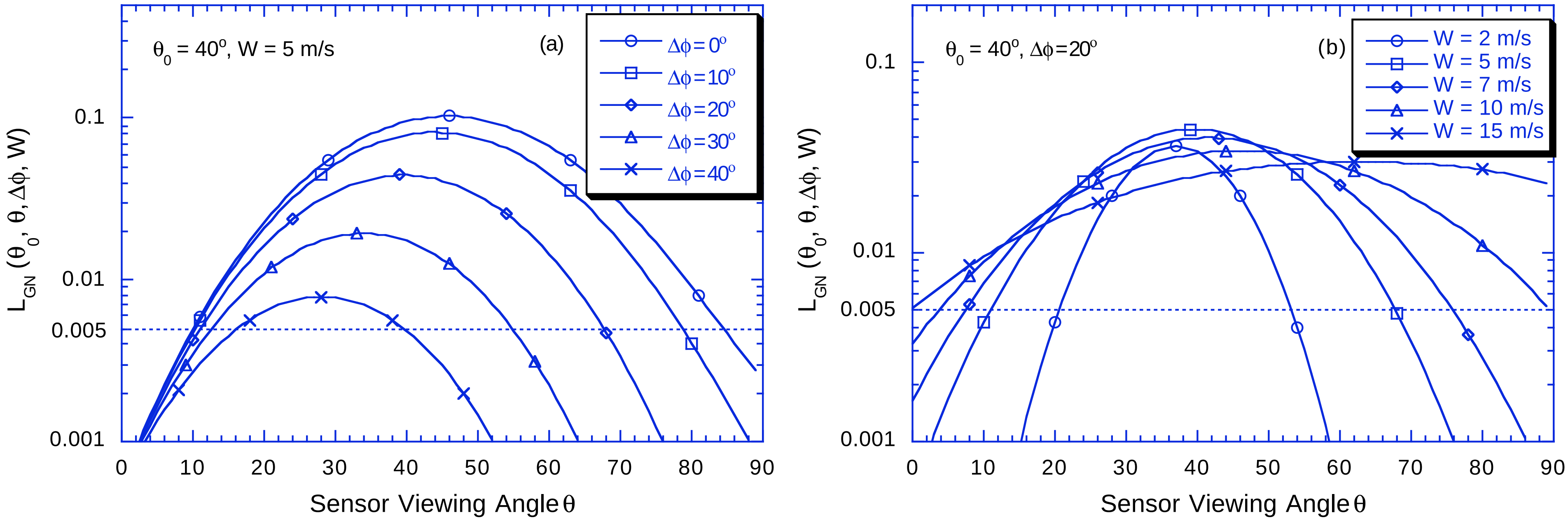
$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}} \underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

Sun Glint Contribution

At the TOA glint radiance is (define normalized glint radiance L_{GN})

$$T(\lambda)L_g(\lambda) = F_0(\lambda)T(\lambda)T_0(\lambda)L_{GN}$$

$$T_0(\lambda)T(\lambda) = \exp\left\{-\left[\tau_r(\lambda)+\tau_a(\lambda)\right]\left[\frac{1}{\cos\theta_0}+\frac{1}{\cos\theta}\right]\right\}$$



Ocean regions with significant contamination by sun glint should be masked and avoided. **Cox and Munk** (1954) model can be used for masking the significant glint contamination regions. Some glint contamination can be corrected.

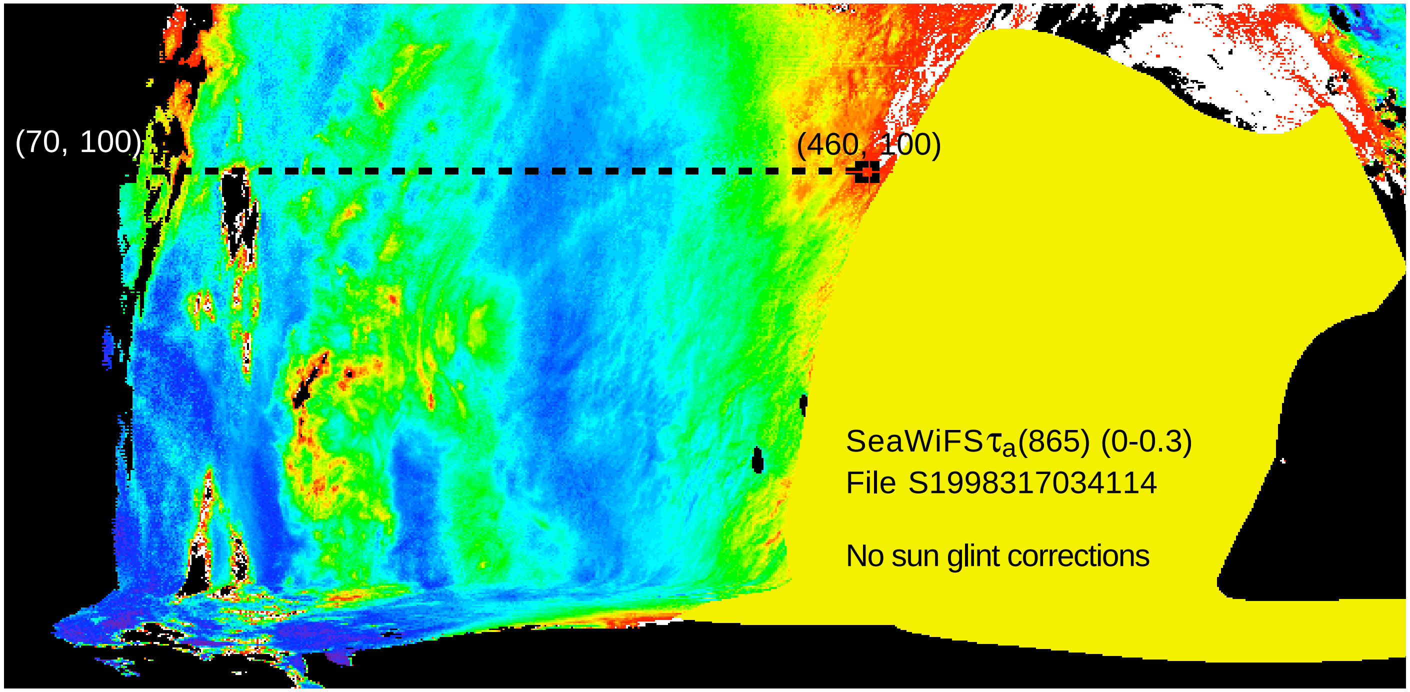
Wang and Bailey (2001) proposed a glint correction algorithm for SeaWiFS & MODIS ocean color data processing using Cox and Munk (1954) model.

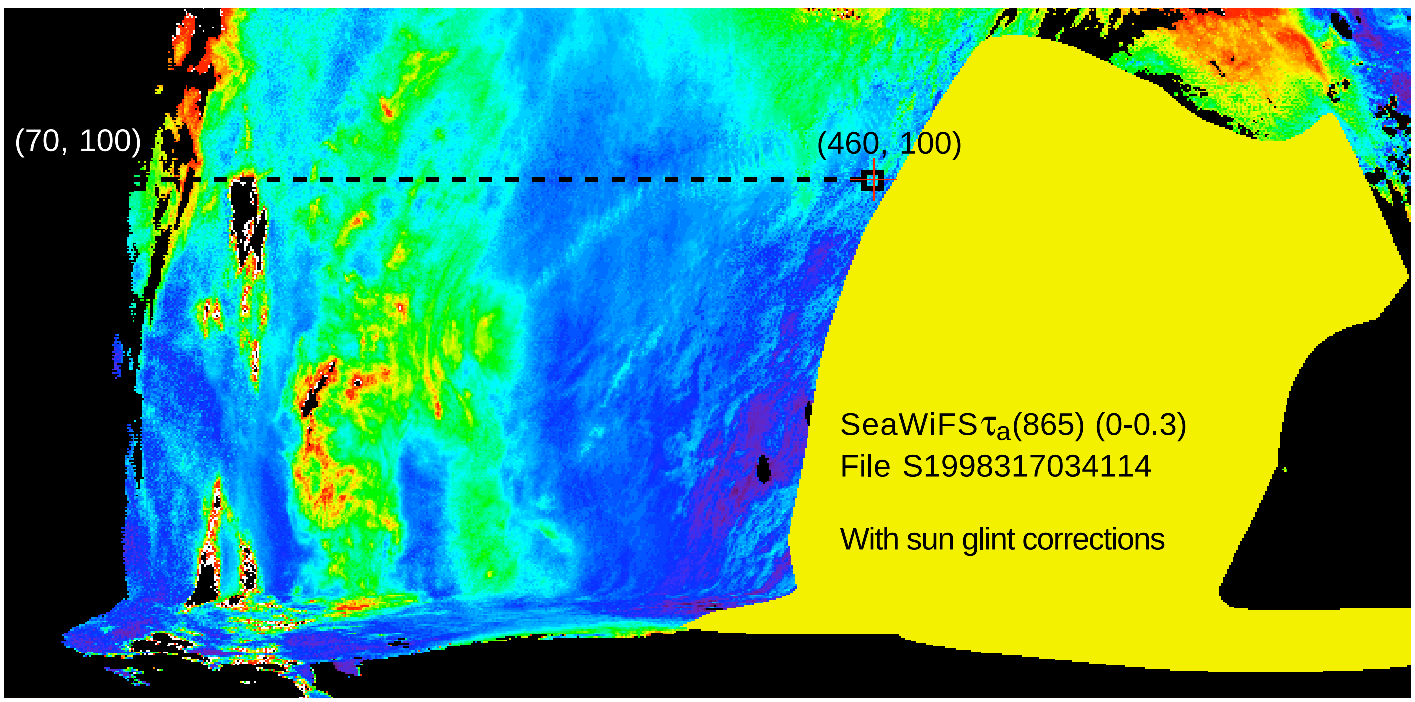
$$\left\{L_t(\lambda), \tau_a'(\lambda), W\right\} \Rightarrow L_t^{(1)\prime}(\lambda) \Rightarrow \tau_a^{(1)}(\lambda) \Rightarrow L_t^{(2)\prime}(\lambda) \Rightarrow \tau_a^{(2)}(\lambda)$$

$$L_t'(\lambda) = L_t(\lambda) - F_0(\lambda) T_0(\lambda) T(\lambda) L_{GN}$$

With the glint correction, ocean color data are reasonable for the normalized sun glint radiance L_{GN} up to ~0.01.

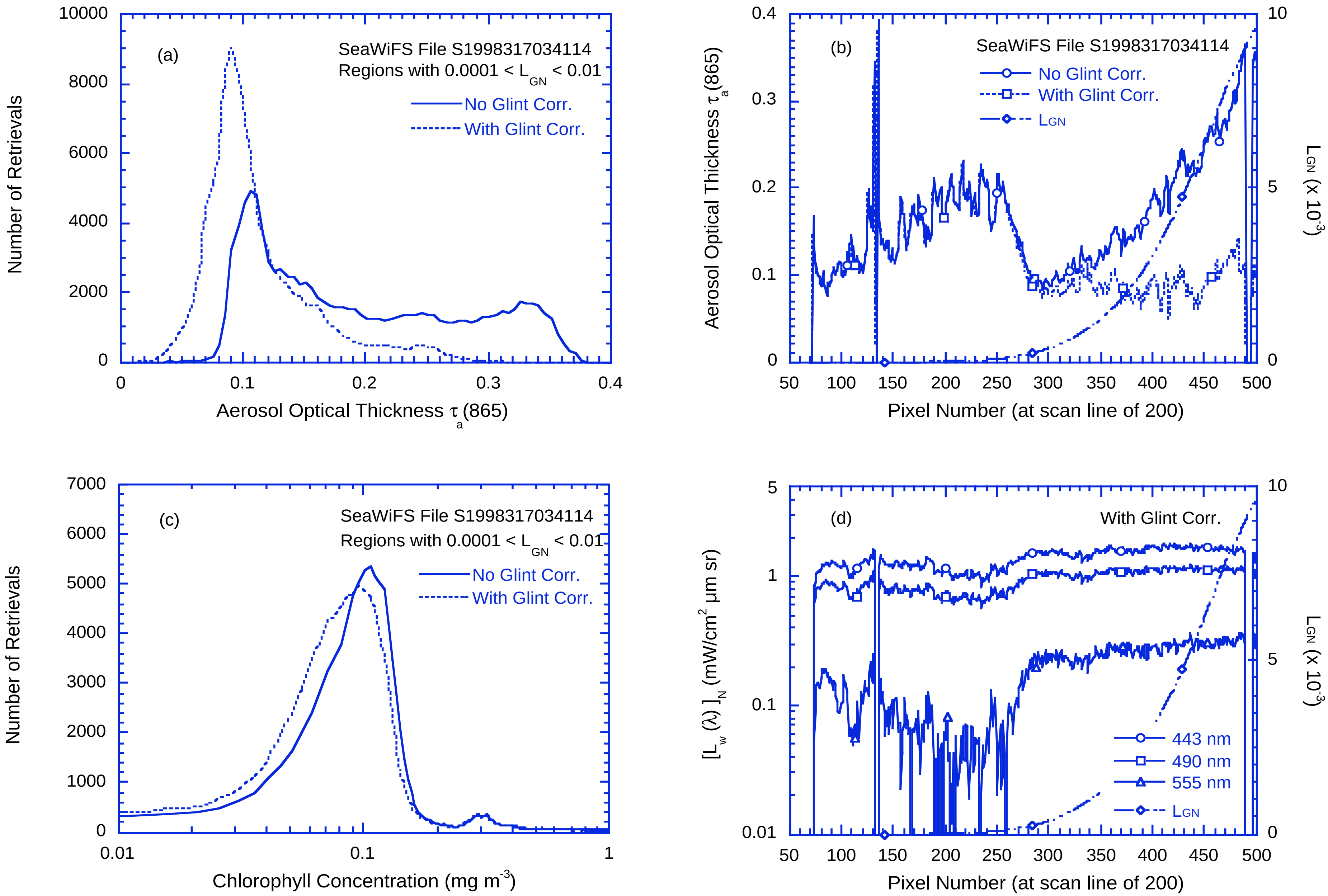
Using MODIS measurements, *Zhang and Wang* (2010) showed that Cox and Munk (1954) model, which was developed 50 years ago, is still among the best.





SeaWiFS aerosol optical thickness (AOT)

Sun Glint Correction



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Atmospheric diffuse transmittance

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}}\underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

The atmospheric diffuse transmittance can be defined as:

$$t(\hat{\xi}) = \frac{L_w^{(TOA)}\left(\hat{\xi}\right)}{L_w\left(\hat{\xi}\right)}$$

Assuming the upward radiance just beneath the surface is **uniform**, *Yang and Gordon* (1997) show that

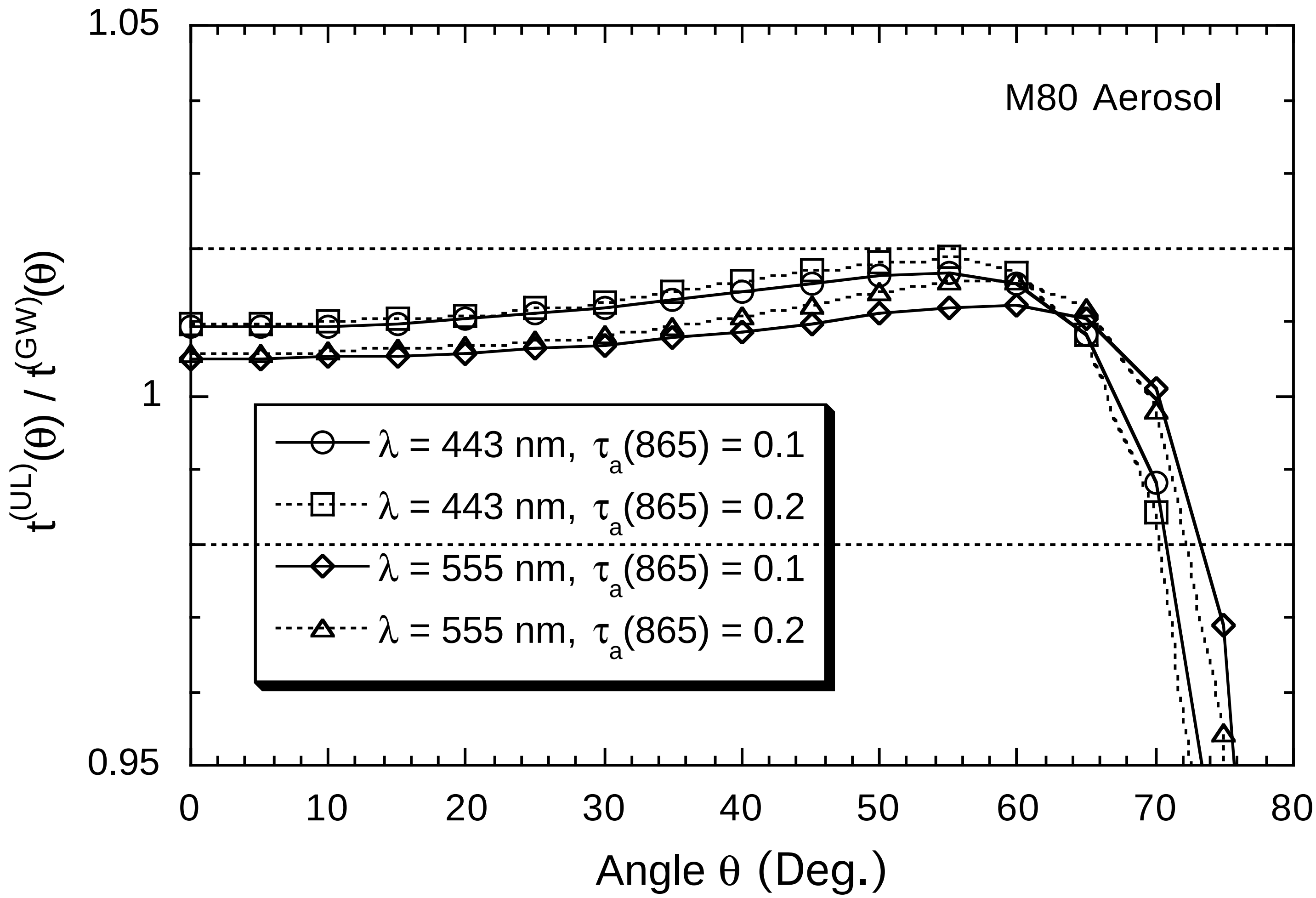
$$t\left(\theta_0\right) = \frac{E_d^-\left(\theta_0\right)}{F_0 \cos\theta_0 \, T_f\left(\theta_0\right)} \quad \text{or more general} \quad t\left(\alpha\right) = \frac{E_d^-\left(\alpha\right)}{F_0 \cos\alpha \, T_f\left(\alpha\right)}$$

Where T_f is the Fresnel transmittance for air-sea interface, E^-_d downward irradiance just beneath the surface, and F_0 is solar irradiance at the TOA.

On the other hand, if we assume the upward radiance just **above** the surface is **uniform**, we can get (IOCCG, 2010)

$$t'(\theta_0) = \frac{E_d(\theta_0)}{F_0 \cos\theta_0} \quad \text{or more general} \quad t'(\alpha) = \frac{E_d(\alpha)}{F_0 \cos\alpha}$$

where E_d is downward irradiance just above the surface, and F_0 is solar irradiance at the TOA.



Here, $t^{(UL)}$ is t' and $t^{(GW)}$ is t. Results show errors are usually within ~2% for solar or sensor angles < 70º for these two systems in computing atmospheric diffuse transmittance.

For a Rayleigh atmosphere with a Fresnel-reflecting surface, the diffuse transmittance can be approximated (Gordon et al., 1983) as:

Eq. (2)

$$t_r(\lambda,\theta)=\exp\left[-\frac{\tau_r(\lambda)}{2\cos\theta}\right]$$

This simple formula can be improved as (Wang, 1999):

Eq. (3)

$$t_r^{(C)}(\lambda,\theta)=\exp\left[-C_r(\lambda,\theta)\tau_r(\lambda)/2\cos\theta\right]$$

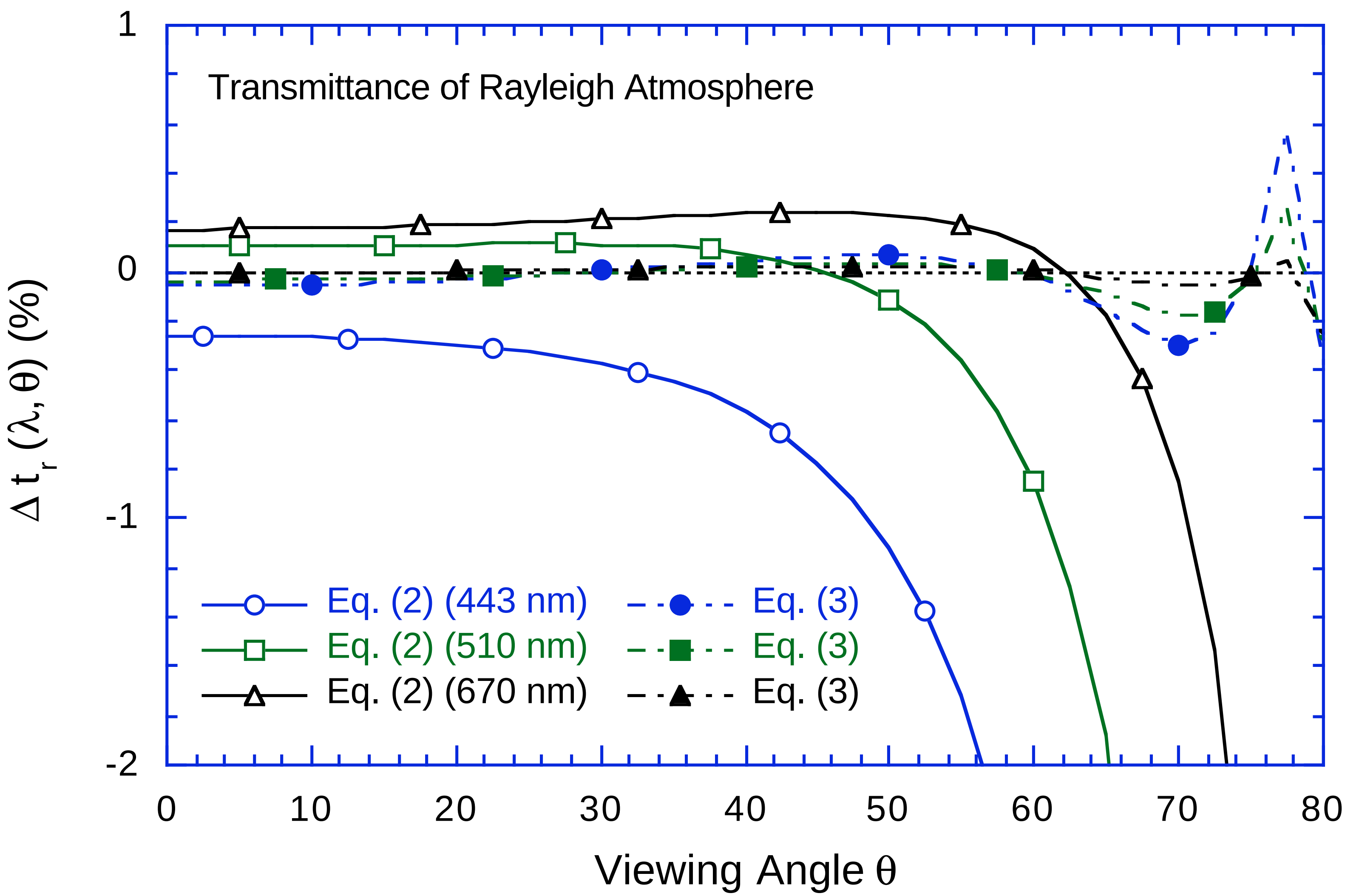
$$C_r(\lambda,\theta)=a_1(\theta)+a_2(\theta)\log_e\left[\tau_r(\lambda)\right]+a_3(\theta)\log_e^2\left[\tau_r(\lambda)\right]$$

$$a_j(\theta)=a_{0j}+a_{1j}/\cos\theta+a_{2j}/\cos^2\theta+a_{3j}/\cos^3\theta$$

TABLE 1. Values of coefficients to fit the diffuse transmittance computations for a pure Rayleigh atmosphere bounded by a flat Fresnel-reflecting ocean.

j	a_{oj}	a_{1j}	a_{2j}	a_{3j}
1	1.0220	-0.1408	-0.0167	0.0037
2	0.0333	-0.1645	0.0442	-0.0027
3	0.0075	-0.0198	0.0057	-0.0004

Results from Improved Approximation



For an atmosphere composed of both air molecules and aerosols with a Fresnel-reflecting surface, the diffuse transmittance can be approximated as (Gordon et al., 1983):

Eq. (6)

$$t(\lambda,\theta) = t_r(\lambda,\theta)\, t_a(\lambda,\theta)$$

$$t_a(\lambda,\theta) = \exp\left[-\left(1-\omega_a(\lambda)F_a(\lambda)\right)\tau_a(\lambda)/\cos\theta\right]$$

$$F_a(\lambda) = \frac{1}{2}\int_0^1 P_a(\Theta)\, d\cos\Theta$$

It can be improved as (Wang, 1999):

Eq. (8)

$$t^{(C)}(\lambda,\theta) = t_r^{(C)}(\lambda,\theta)\, t_a^{(C)}(\lambda,\theta)$$

$$t_a^{(C)}(\lambda,\theta) = \exp\left[-a_o(\lambda)\left(1+\omega_a(\lambda)C_a(\lambda,\theta)\right)/\cos\theta\right]$$

$$C_a(\lambda,\theta) = b_1(\lambda,\theta) + b_2(\lambda,\theta)\log_e\left[a_o(\lambda)\right] + b_3(\lambda,\theta)\log_e{}^2\left[a_o(\lambda)\right]$$

$$b_j(\lambda,\theta) = b_{0j}(\lambda) + b_{1j}(\lambda)/\cos\theta + b_{2j}(\lambda)/\cos^2\theta + b_{3j}(\lambda)/\cos^3\theta + b_{4j}(\lambda)/\cos^4\theta$$

All coefficients are in the next page.

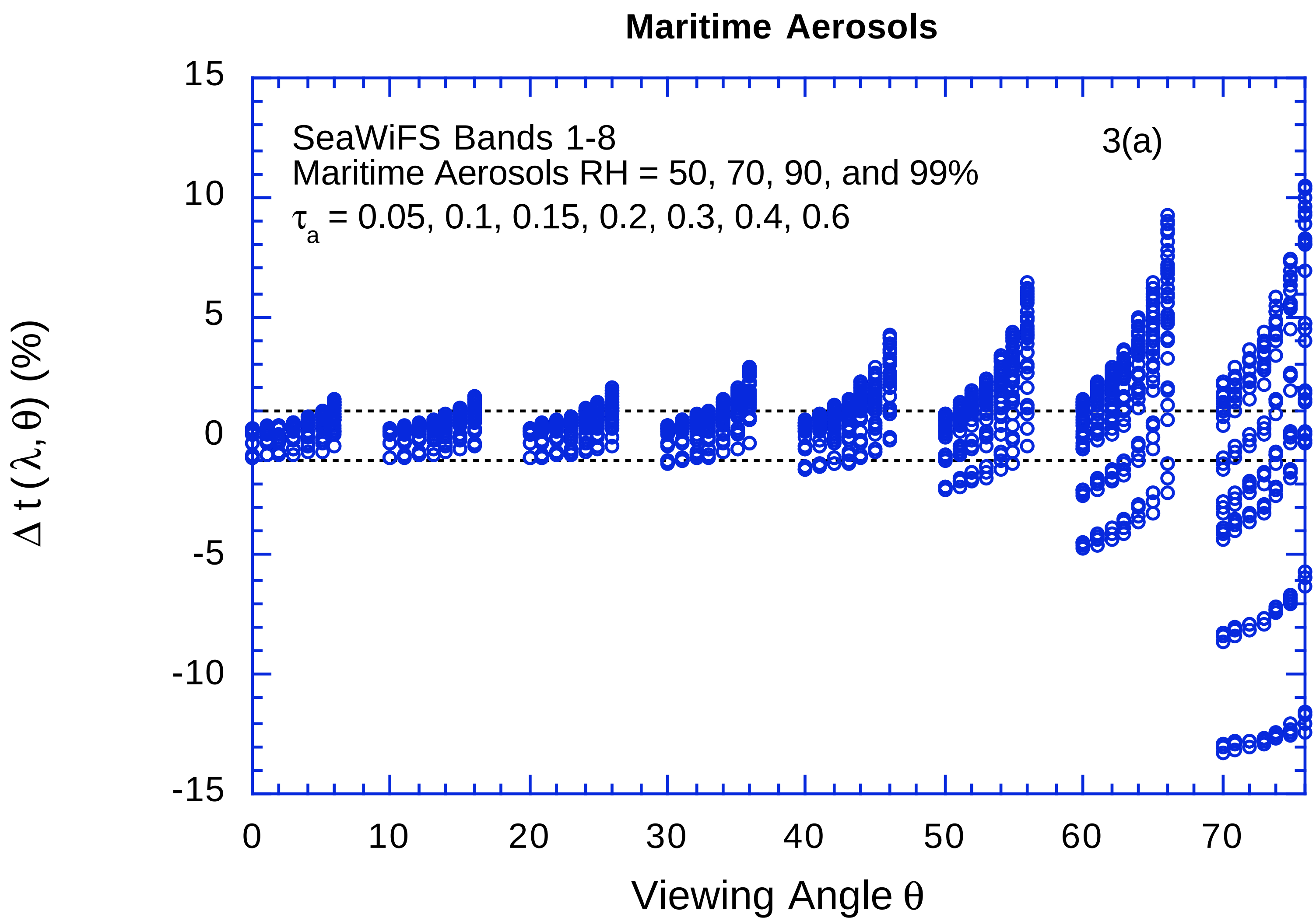
TABLE 3. Values of coefficients to fit the diffuse transmittance computations for a two-layer Rayleigh-aerosol atmosphere bounded by a flat Fresnel-reflecting ocean for the wavelengths of 412, 490, 555, and 865 nm.

λ (nm)	j	b_{oj}	b_{1j}	b_{2j}	b_{3j}	b_{4j}
412	1	-3.869	7.247	-4.545	1.114	-0.095
	2	-2.620	4.335	-2.719	0.685	-0.060
	3	-0.608	0.978	-0.586	0.145	-0.013
490	1	1.009	-2.407	2.197	-0.771	0.088
	2	1.032	-2.654	2.092	-0.657	0.071
	3	0.043	-0.253	0.265	-0.093	0.011
555	1	2.418	-4.397	3.073	-0.888	0.089
	2	2.135	-4.191	2.778	-0.757	0.073
	3	0.249	-0.545	0.405	-0.117	0.012
865	1	1.833	-3.090	2.004	-0.518	0.045
	2	1.848	-3.454	2.148	-0.532	0.046
	3	0.224	-0.472	0.340	-0.091	0.008

TABLE 2. Aerosol optical properties for the 16 aerosol models for the SeaWiFS band 8 (865 nm). The $\tau_a(\lambda_1)/\tau_a(\lambda_8)$ values are for λ_1 = 412 nm and λ_8 = 865 nm.

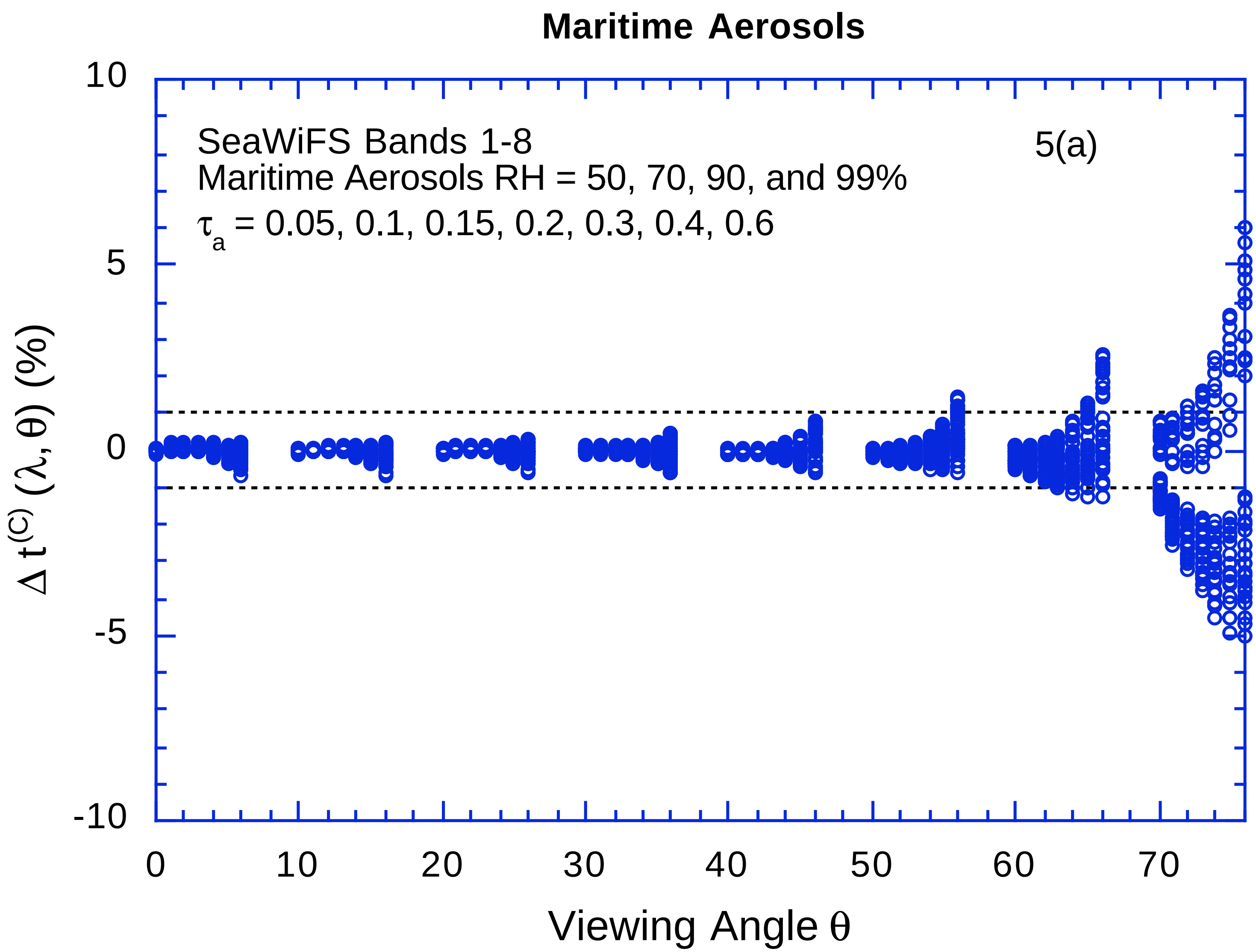
Aerosol Model	ω_a	F_a	$\omega_a F_a$	$\tau_a(\lambda_1)/\tau_a(\lambda_8)$
M50	0.9814	0.9089	0.8920	1.440
M70	0.9859	0.9203	0.9073	1.344
M90	0.9953	0.9381	0.9337	1.175
M99	0.9986	0.9460	0.9447	1.077
C50	0.9705	0.9036	0.8769	1.719
C70	0.9768	0.9146	0.8934	1.595
C90	0.9919	0.9354	0.9278	1.356
C99	0.9974	0.9449	0.9424	1.187
T50	0.9295	0.8827	0.8205	2.771
T70	0.9346	0.8872	0.8292	2.756
T90	0.9698	0.9175	0.8898	2.524
T99	0.9870	0.9346	0.9225	2.197
U50	0.6026	0.8914	0.5372	2.190
U70	0.6605	0.9056	0.5982	2.206
U90	0.8206	0.9314	0.7643	2.141
U99	0.9419	0.9480	0.8929	1.741

Results from Eq. (6) (Old)



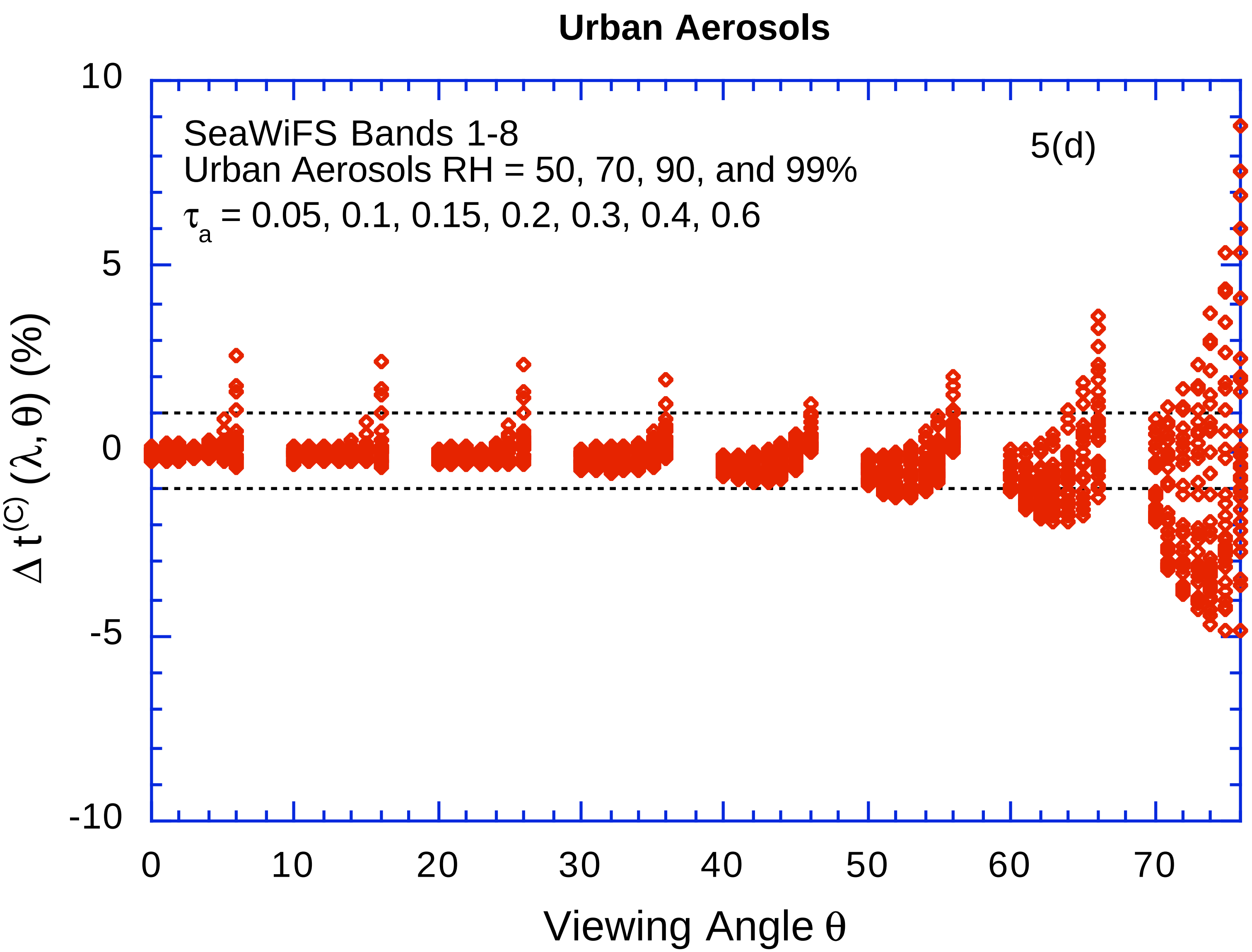
Atmospheric diffuse transmittance

Results from Eq. (8) (New)

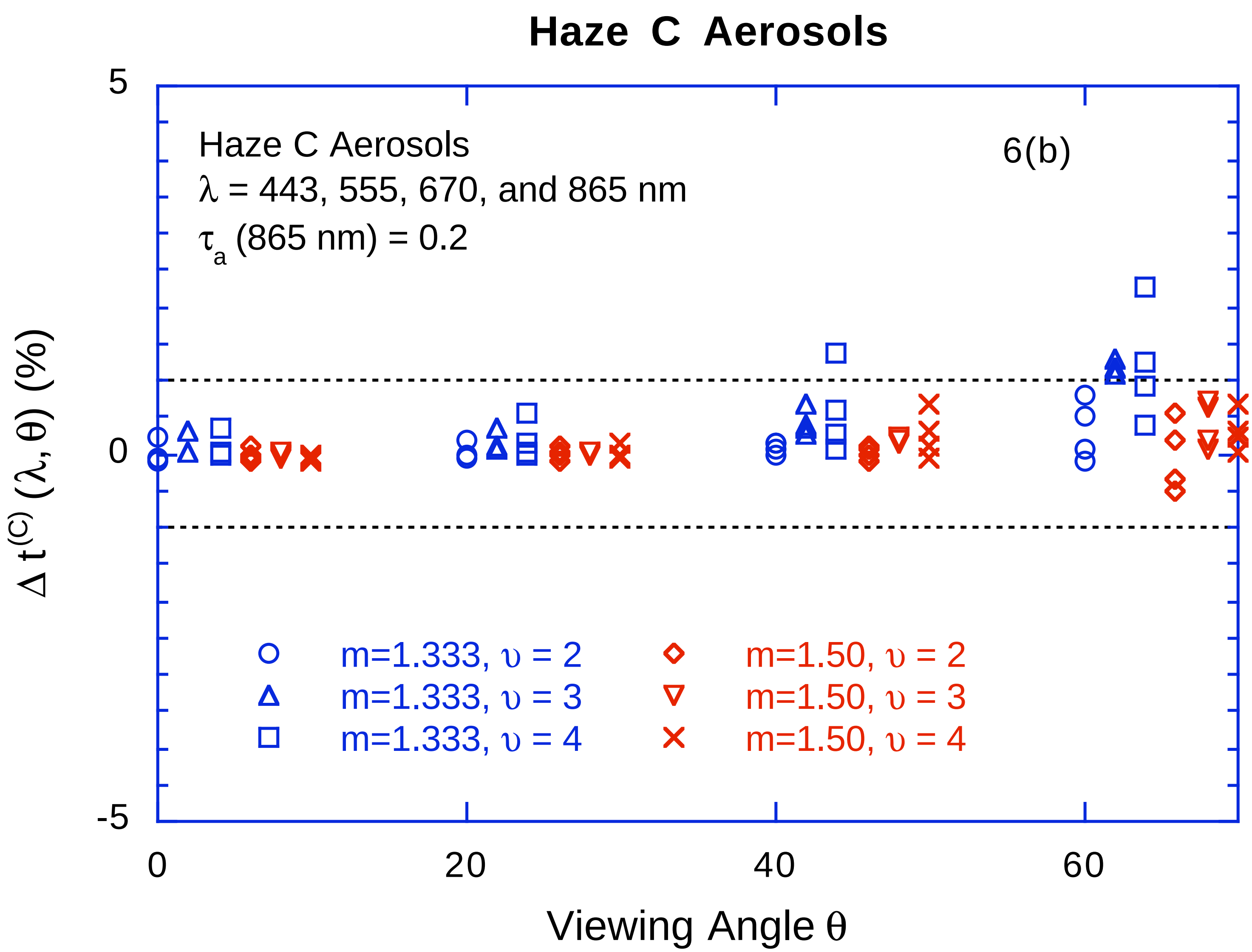


Atmospheric diffuse transmittance

Results from Eq. (8) (New)

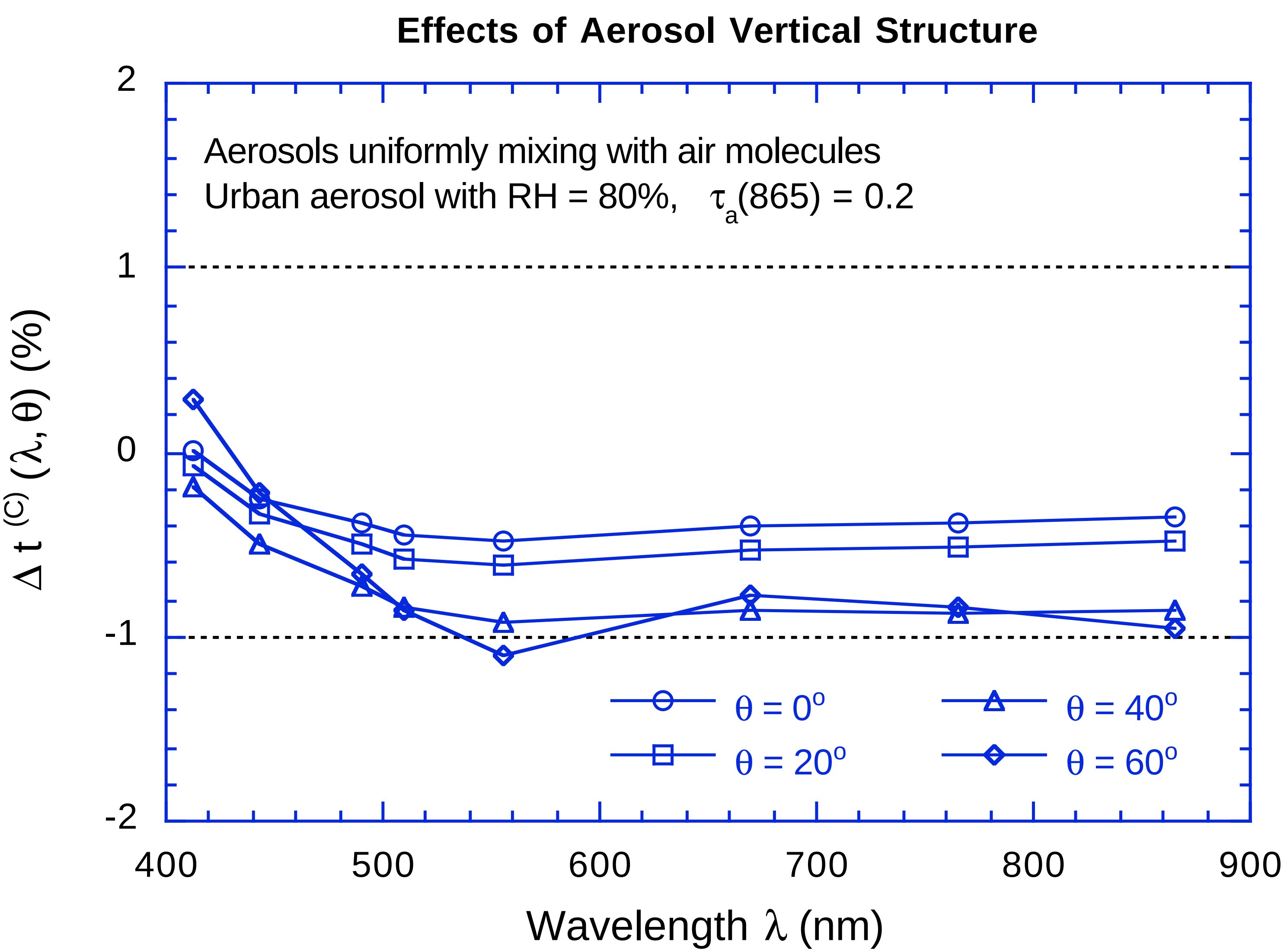


Results from Eq. (8) (New)--Different Aerosol Models



Atmospheric diffuse transmittance

Results from Eq. (8) (New)--Different Vertical Profiles

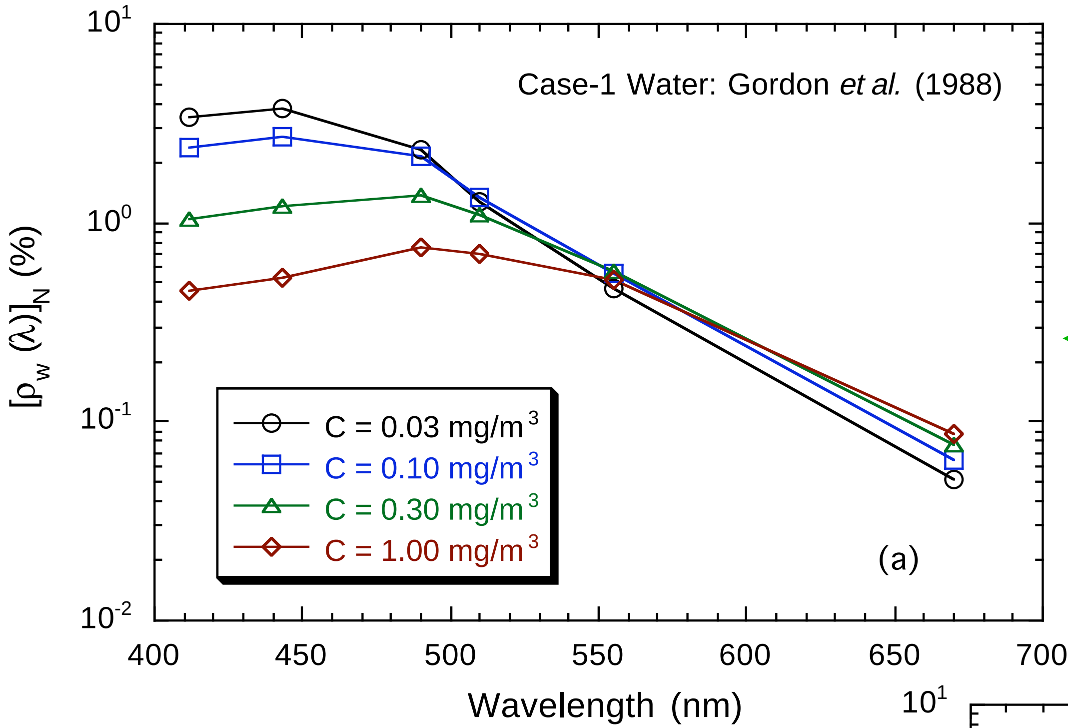


Atmospheric diffuse transmittance

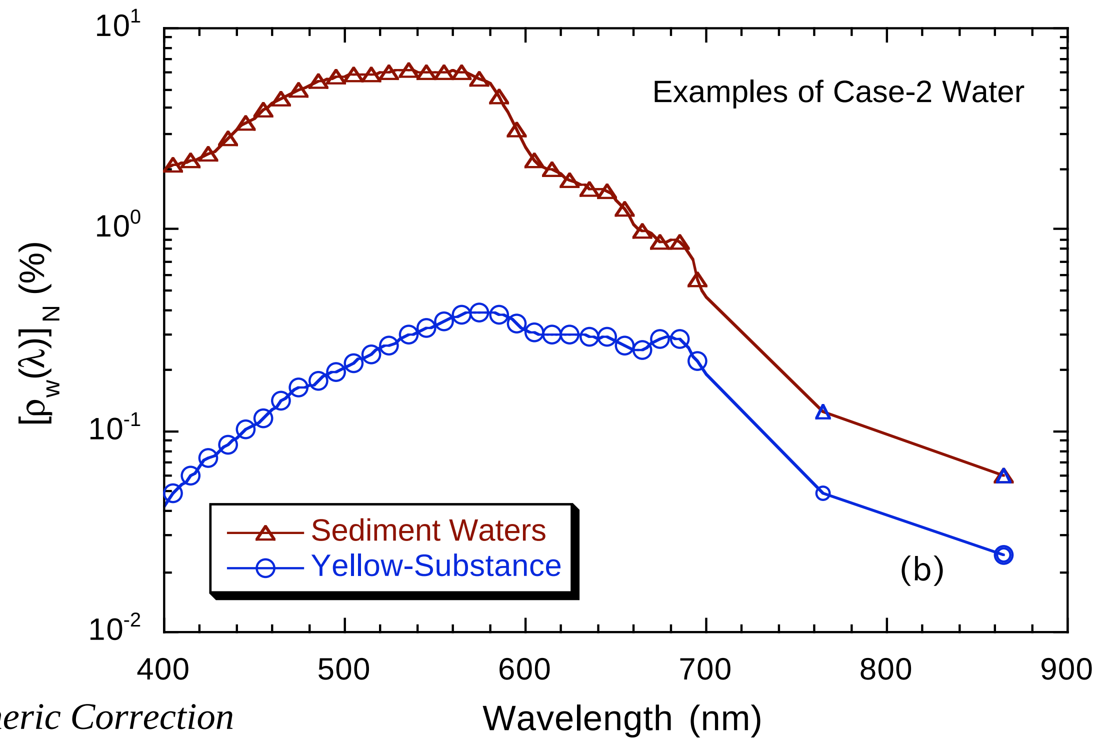
Normalized water-leaving radiance (reflectance)

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

$$+ \underbrace{t(\lambda)t_0(\lambda)}_{\text{Diffuse Transmittance}}\underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$



Ocean Contributions:
Case-1 Waters



Ocean Contributions:
Case-2 Waters

IOCCG Report-10

The normalized water-leaving radiance can be defined (Gordon and Clark, 1981)

$$\left[L_w(\lambda,\theta_0,\theta,\Delta\phi)\right]_N = L_w(\lambda,\theta_0,\theta,\Delta\phi)\frac{\bar{F}_0(\lambda)}{E_d^{(+)}(\lambda,\theta_0)} \cong \left(\frac{d}{d_0}\right)^2 \frac{L_w(\lambda,\theta_0,\theta,\Delta\phi)}{t(\lambda,\theta_0)\cos\theta_0}$$

Morel et al. (2002) and Morel and Gentili (1991; 1993; 1996) extended the definition to account for additional effects due to angular variations in reflection and refraction at the sea surface and for the in-water BRDF, introducing a quantity the exact normalized water-leaving radiance:

$$\left[L_w(\lambda)\right]_N^{Exact} = \left[L_w(\lambda,\theta_0,\theta,\Delta\phi)\right]_N \left\{(f/Q)_{Eff}\right\} \left[\frac{\Re_0(\lambda,\tau_a,W)}{\Re(\lambda,\theta_0,\theta,\tau_a,W)}\right]$$

$$\left\{(f/Q)_{Eff}\right\} = \left\{\left(\frac{f_0(\lambda,IOP)}{Q_0(\lambda,IOP)}\right) \Big/ \left(\frac{f(\lambda,\theta_0,IOP)}{Q(\lambda,\theta_0,\theta,\Delta\phi,IOP)}\right)\right\}$$

where (*f/Q*) is effects of the in-water BRDF, and term $\Re$ ratio accounts for angular variations in all effects of reflection and refraction of radiance at the sea surface.

Some Useful Relationships

The normalize water-leaving radiance (Gordon & Clark, 1981):

$$[L_w(\lambda)]_N = L_w(\lambda) \,/\, \cos\theta_0 \, t(\lambda,\theta_0)$$
$$= F_0(\lambda) \, L_w(\lambda) \,/\, E_d(\lambda,\theta_0)$$

where $t(\lambda,\theta_0)$ is diffuse transmittance at the solar direction, $F_0(\lambda)$ is the solar irradiance at the TOA, and $E_d(\lambda,\theta_0)$ is the downward irradiance measured just above the sea surface.

The normalize water-leaving reflectance is defined as

$$[\rho_w(\lambda)]_N = \pi \, [L_w(\lambda)]_N \,/\, F_0(\lambda)$$
$$= \rho_w(\lambda) \,/\, t(\lambda,\theta_0)$$

$$\Re\left(\lambda,\theta_0,\theta,\tau_a,W\right)=\Re^{(Sun)}\left(\lambda,\theta_0,\tau_a,W\right)\Re^{(View)}\left(\lambda,\theta,W\right)$$

$$\Re^{(Sun)}\left(\lambda,\theta_0,\tau_a,W\right)=\frac{1-\bar{\rho}_f\left(\lambda,\theta_0,\tau_a,W\right)}{1-\bar{r}\,R\left(\lambda,\theta_0\right)}$$

Solar angle (Irradiance) term

$$\Re^{(View)}\left(\lambda,\theta,W\right)=\frac{1-\rho_f\left(\lambda,\theta,W\right)}{m^2}$$

Sensor angle (radiance) term

$$\left[L_w\left(\lambda\right)\right]_N^{Exact}=\left[L_w\left(\lambda,\theta_0,\theta\right)\right]_N\left[\frac{\Re_0^{(Sun)}\left(\lambda,\tau_a,W\right)}{\Re^{(Sun)}\left(\lambda,\theta_0,\tau_a,W\right)}\right]\left[\frac{\Re_0^{(View)}\left(\lambda,W\right)}{\Re^{(View)}\left(\lambda,\theta,W\right)}\right]$$

IOCCG Report #10 & References therein

Solar Zenith Angle Effects on the In Situ nLw Measurements (e.g., MOBY)

Solar angle (Irradiance) term

$$\Re\left(\lambda,\theta_0,\theta,\tau_a,W\right)=\left[\frac{1-\bar{\rho}\left(\lambda,\theta_0,\tau_a,W\right)}{1-\bar{r}\,R\left(\lambda,\theta_0\right)}\right]\left[\frac{1-\rho\left(\lambda,\theta,W\right)}{m^2}\right]$$

Sensor angle (radiance) term

Solar Angle Effects (sea surface) on *nLw*

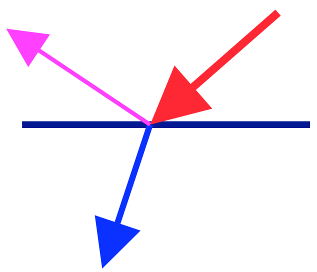
Basic Physics:

Irradiance just below the surface

$$nL_w(\lambda) \propto E_d(0^-, \theta_0), \quad nL_w(\lambda) = 0 \text{ at night } (\theta_0 > 90^o)!$$

Because of sea surface, $E_d(0^-, \theta_0>0) < E_d(0^-, \theta_0=0)$. As θ_0 increase, more photons reflected (instead of entering into water), i.e.,

$$\frac{E_d(0^-, \theta_0 = 0, \lambda, W)}{E_d(0^+, \theta_0 = 0, \lambda, W)} > \frac{E_d(0^-, \theta_0 > 0, \lambda, W)}{E_d(0^+, \theta_0 > 0, \lambda, W)}$$



and

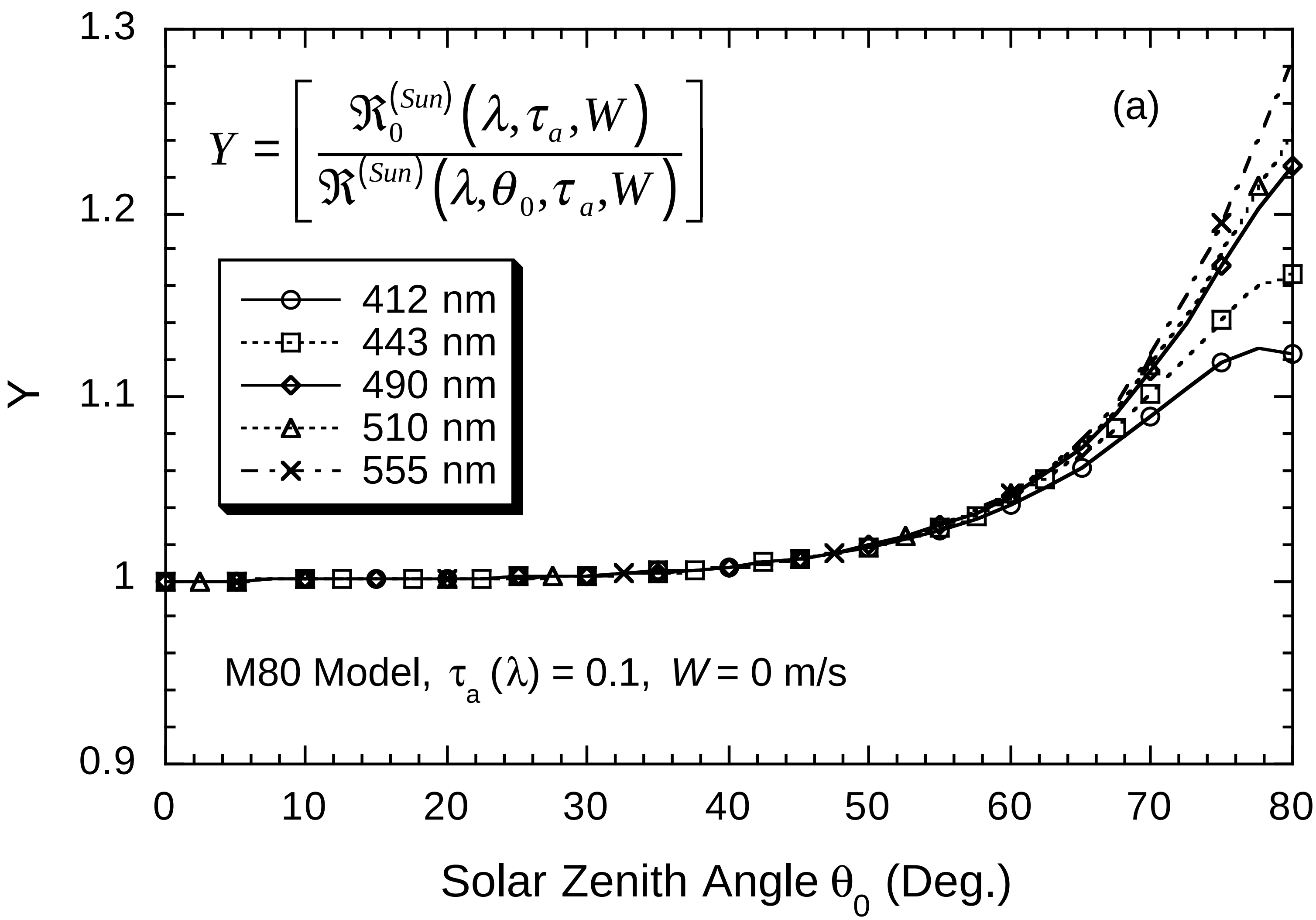
$$nL_w(\theta_0 = 0, \lambda, W) > nL_w(\theta_0 > 0, \lambda, W)$$

This is part of the formulation in Morel & Gentili (1996) and Morel and Mueller (2002) ($\Re(\theta, \theta_0, W)$), but assumed constant (independent of θ_0).

Computing the $\Re$ Terms

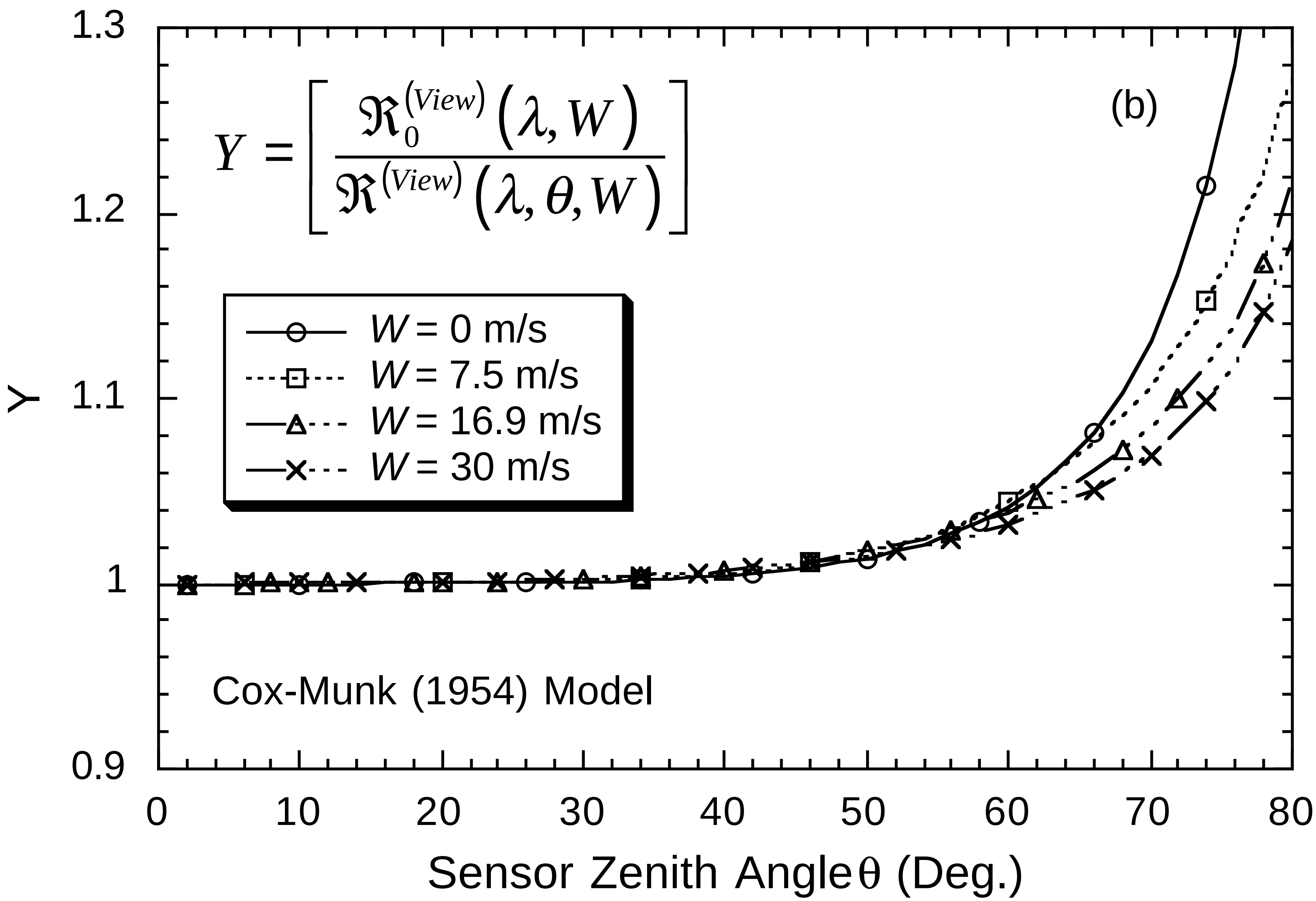
It has been found that, for cases with solar and sensor zenith angles up to ~60^{o}, the sun angle $\Re$ term depends only on the properties of the ocean surface and not the atmosphere (Wang, 2006), while the sensor angle $\Re$ term can be computed accurately with a flat ocean surface (i.e., independent of the wind speed) (Gordon, 2005). Thus, the effects of in angular variation of the surface reflection and refraction on the normalized water-leaving radiance can be calculated easily (Gordon, 2005; Wang, 2006).

The Sun Angle $\Re$ Term



Normalized water-leaving radiance

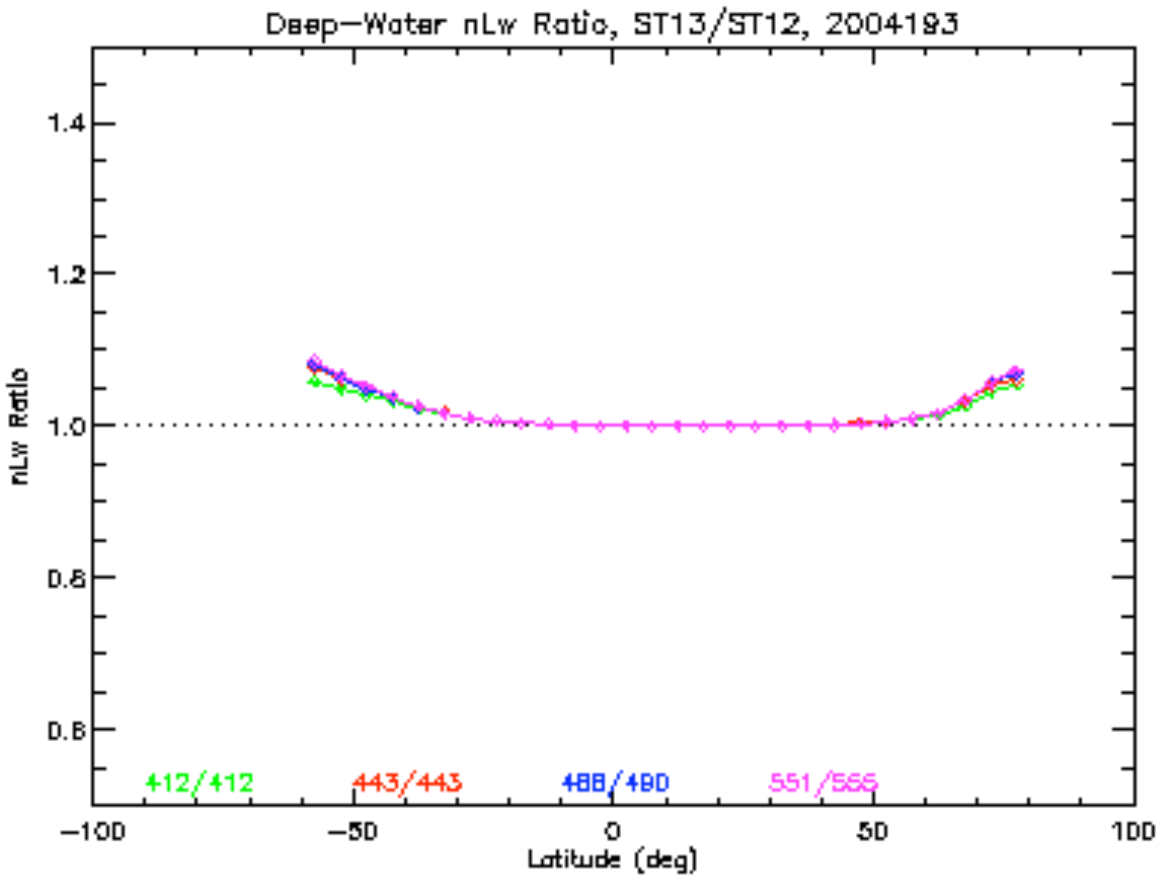
The Sensor Angle $\Re$ Term



Menghua Wang, IOCCG Lecture Series--Atmospheric Correction

Normalized water-leaving radiance

SeaWiFS/MODIS Results



- With correction, MODIS and SeaWiFS results can be more meaningfully compared.
- Satellite and in situ measurements can have better comparisons (if measured at different time between two).

The (*f/Q*) Effects

Within a water body, particularly close to the interface, the upward radiance field is not isotropic, so that a bidirectional reflectance distribution function (BRDF) has to be taken into account.

The (*f*/*Q*) correction factor depends on solar-sensor geometry and the ocean IOPs. The *f* factor is relates ocean upwelling irradiance reflectance to the ocean IOPs and the *Q* factor is defined as ratio of the upwelling irradiance just beneath the ocean surface to the upwelling radiance just beneath the surface.

Please see IOCCG report #10 and many references there in!

Aerosol Contributions

$$L_t(\lambda) = \underbrace{L_r(\lambda)}_{\text{Rayleigh}} + \underbrace{L_A(\lambda)}_{\text{Aerosols}} + t(\lambda)\underbrace{L_{wc}(\lambda)}_{\text{Whitecap}} + \underbrace{T(\lambda)}_{\text{Direct Trans}}\ \underbrace{L_g(\lambda)}_{\text{Sun Glint}}$$

$$+\ \underbrace{t(\lambda)\,t_0(\lambda)}_{\text{Diffuse Transmittance}}\ \underbrace{\left[L_w(\lambda)\right]_N}_{\text{Ocean}}$$

Aerosol Radiance Lookup Tables

Aerosols are highly variable (spatially and temporally), and aerosol radiance contributions depend on aerosol model (size distributions and compositions, or aerosol optical properties: single-scattering albedo, phase function, and spectral extinction coefficients).

- Vector (polarization) RTE used for SeaWiFS/MODIS spectral bands.
- Solar-zenith angles 0^o-80^o (2.5^o), sensor-zenith angles 0^o-76^o ($\sim 2^o$), and relative azimuth angles 0^o-180^o (10^o), .
- Shettle and Fenn (1979) aerosol models (12 models) were used for Oceanic with RH of 99% (O99), Maritime with RH of 50%, 70%, 90%, and 99% (M50, M70, M90, M99), Coastal model with RH of 50%, 70%, 90%, and 99% (C50, C70, C90, C99), and Tropospheric model with RH of 50%, 90%, and 99% (T50, T90, T99).
- A flat Fresnel-reflecting ocean surface is used.

For detail of aerosol lookup implementation see Wang (2007).

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

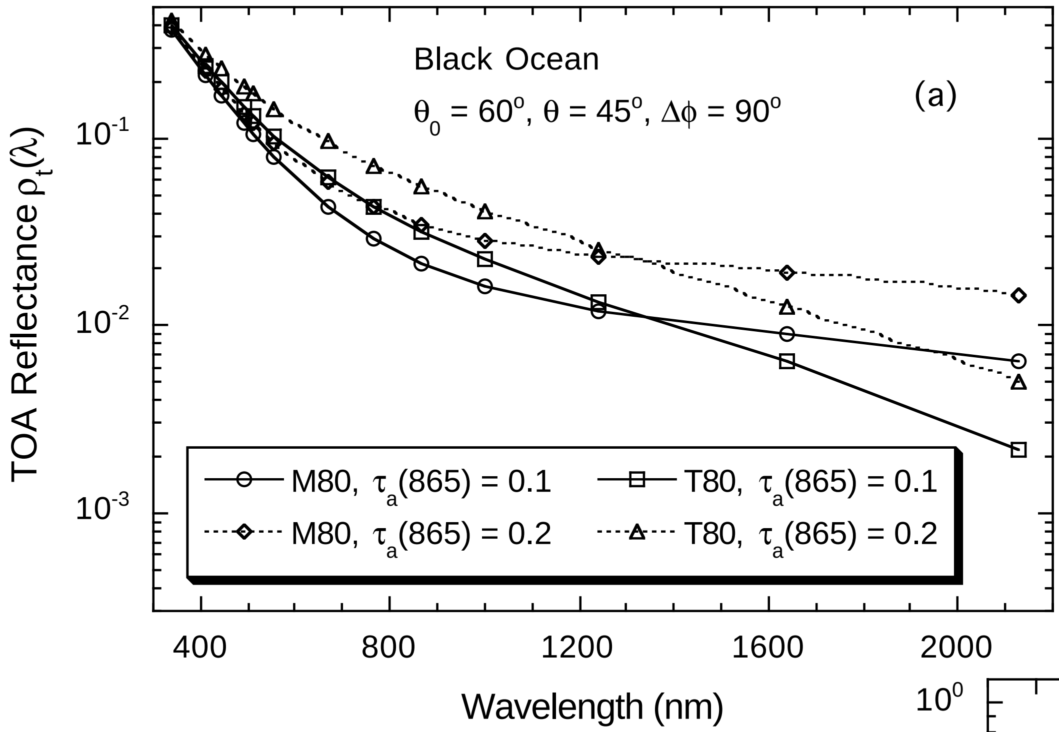
[†] Shettle and Fenn (1979) aerosol models. [††] Gordon and Wang (1994)

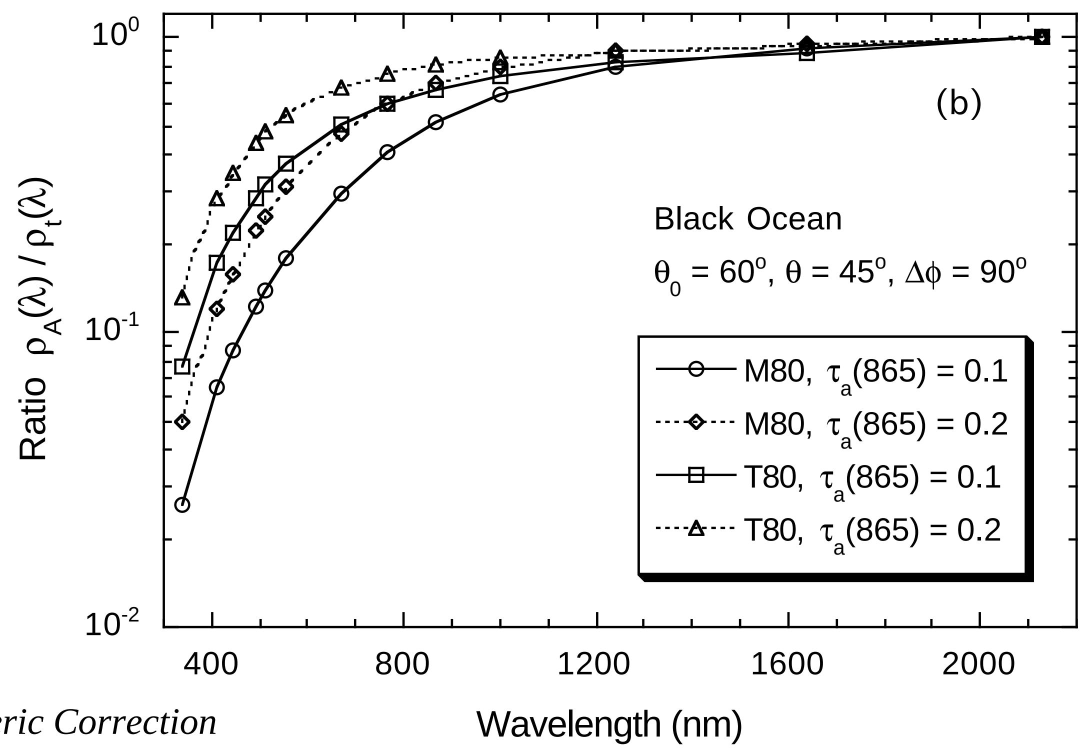
[†††] Shettle (1984) and Moulin et al. (2001).

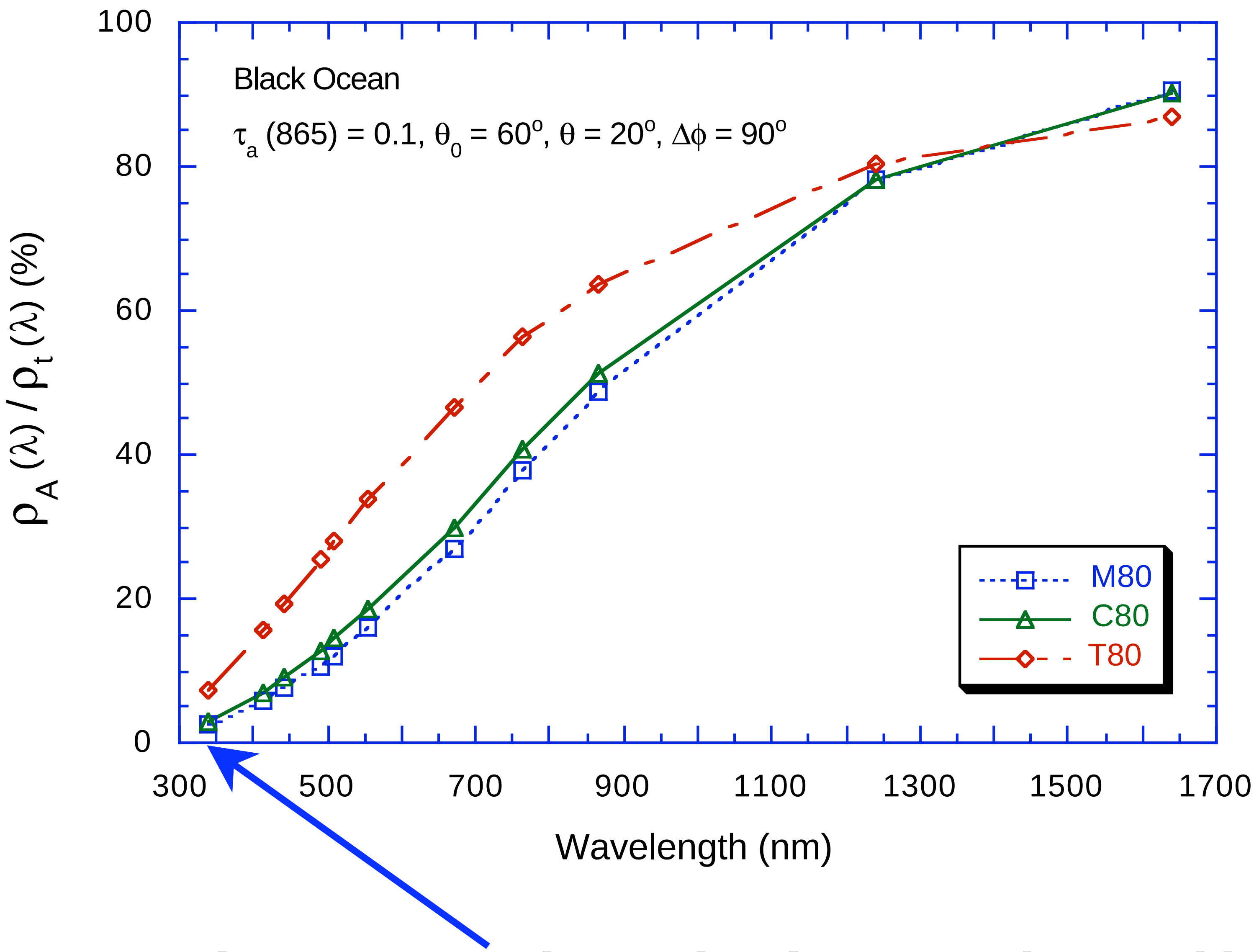
Strongly-absorbing aerosols

Non- and weakly-absorbing aerosols

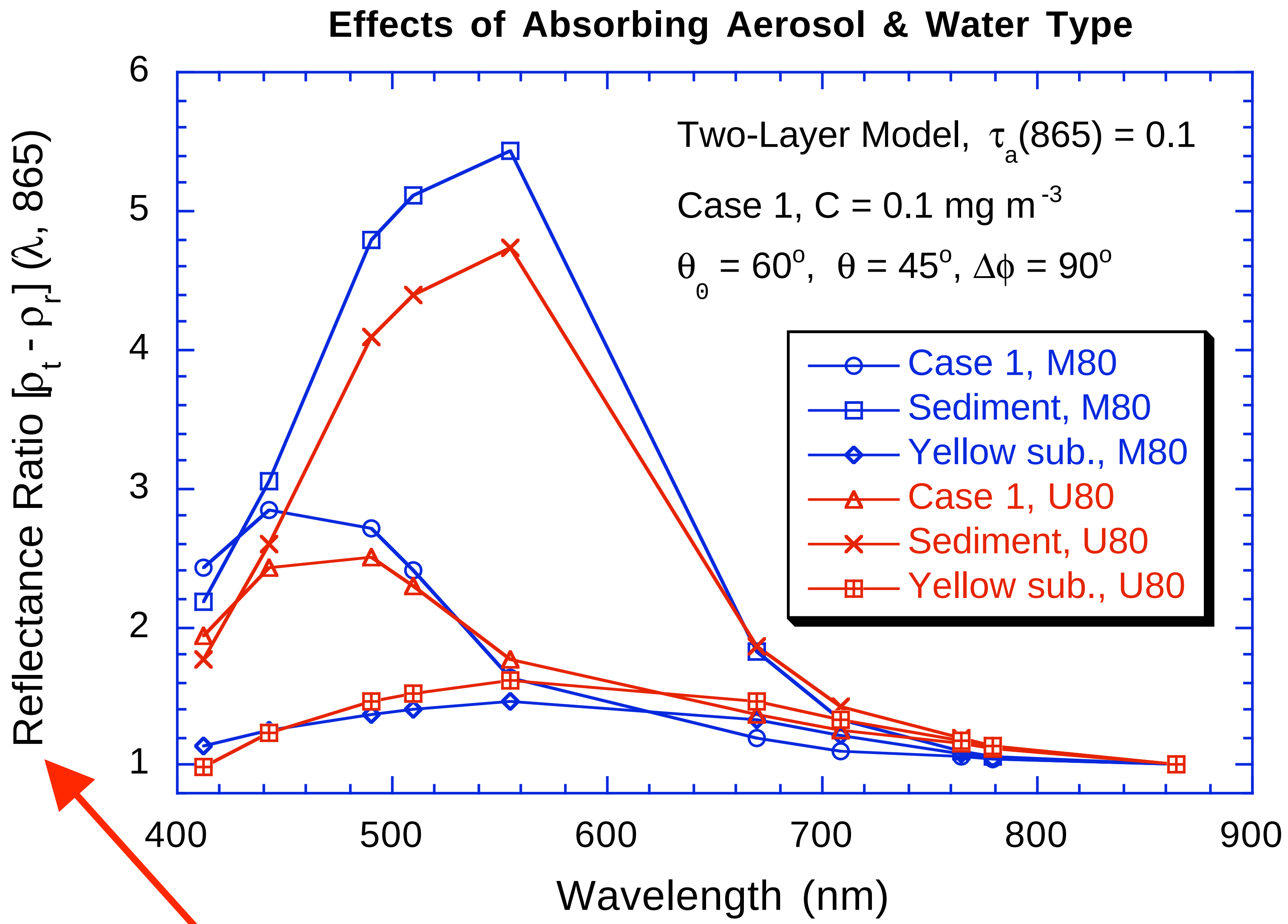
Aerosol Reflectance Contributions



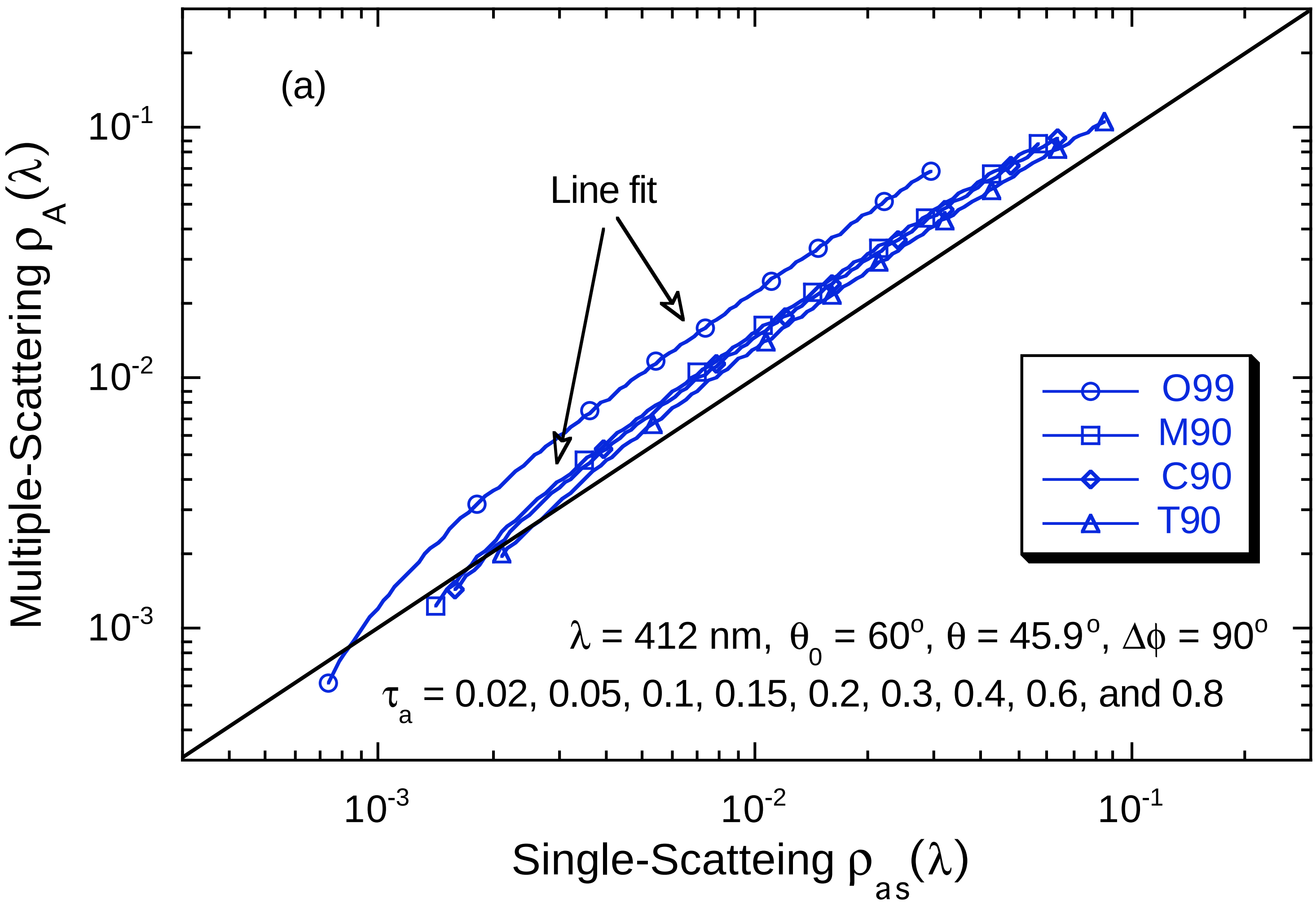




Aerosol correction at the UV bands may not be a problem!

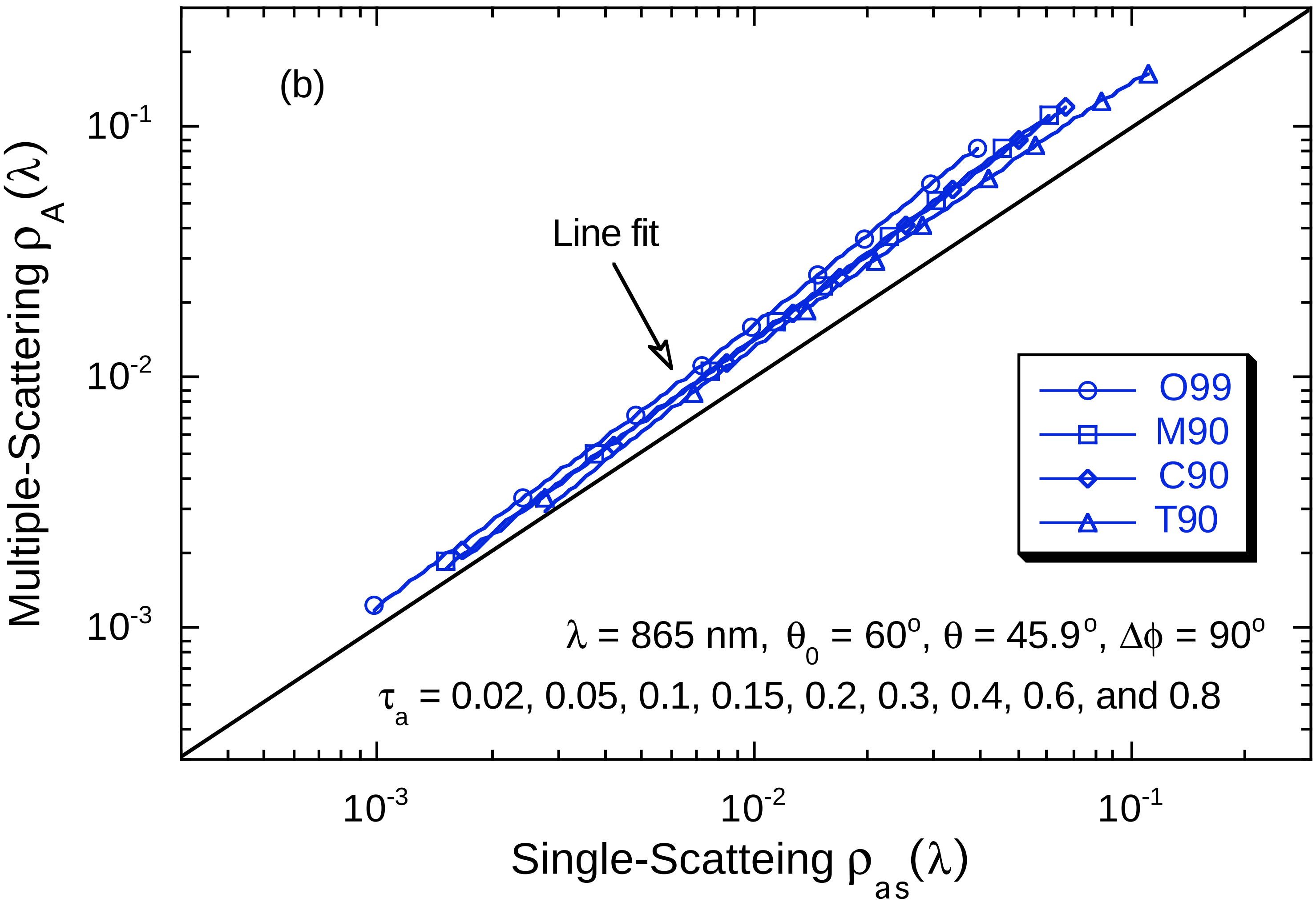


This is $\rho_A + t\,\rho_w$



$$\rho_A(\lambda) \propto \rho_{as}(\lambda) \propto \omega_a(\lambda)\tau_a(\lambda)P_a(\Theta_-,\lambda)$$

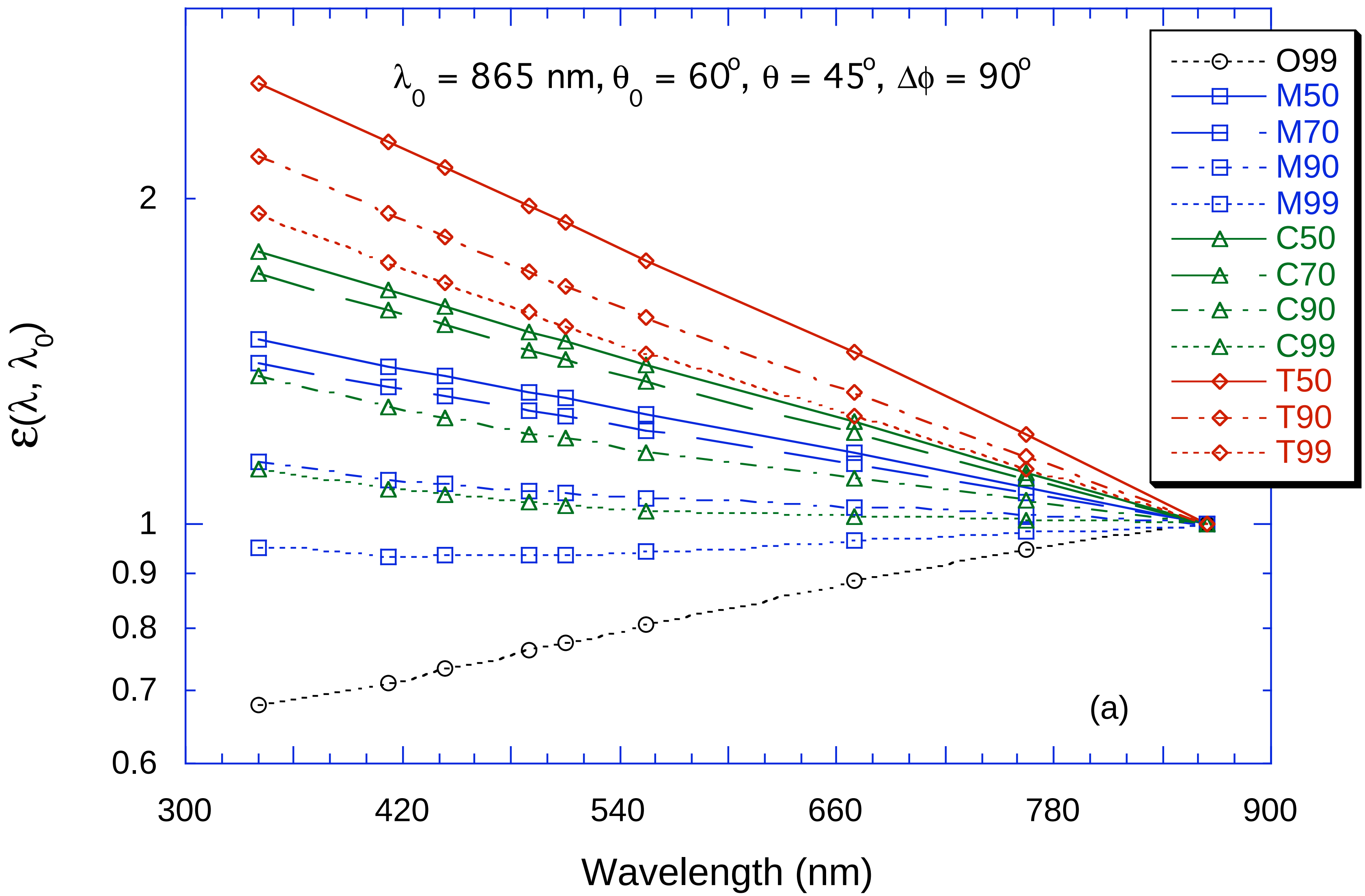
At short blue 412 nm



$$\rho_A(\lambda) \propto \rho_{as}(\lambda) \propto \omega_a(\lambda)\tau_a(\lambda)P_a(\Theta_-, \lambda)$$

At the NIR 865 nm

Aerosol Single-Scattering Epsilon (λ_0 = 865 nm)



Aerosol Single-Scattering Epsilon

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

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

Spectral variation of aerosol reflectance.
Atmospheric correction is for correction of aerosol reflectance (not aerosol optical properties).

Atmospheric Correction Algorithm

MODIS and SeaWiFS algorithm (Gordon and Wang 1994)

$$\rho_t = \rho_r + \rho_A + t\rho_{wc} + T\rho_g + t\rho_w, \quad \rho = \pi L / \mu_0 F_0$$

- **ρ_w is the desired quantity in ocean color remote sensing.**
- $T\rho_g$ is the sun glint contribution—avoided/masked/corrected.
- $T\rho_{wc}$ is the whitecap reflectance—computed from wind speed.
- ρ_r is the scattering from molecules—computed using the Rayleigh lookup tables (vector RTE, wind speed, atmospheric pressure dependents).
- $\rho_A = \rho_a + \rho_{ra}$ is the aerosol and Rayleigh-aerosol contributions —estimated using **aerosol models**.
- For Case-1 waters at the open ocean, ρ_w is usually negligible at 750 & 865 nm. ρ_A can be estimated using these two NIR bands. Ocean is usually not black at NIR for the coastal regions.

Single-scattering approximation for computing ρ_A:

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

Aerosol Correction Procedure (Multiple-scattering)

$$t(\lambda)\rho_w(\lambda) = \underbrace{\left[\rho_t(\lambda) - \rho_r(\lambda) - t(\lambda)\rho_{wc}(\lambda) - T(\lambda)\rho_g(\lambda)\right]}_{\rho'(\lambda)} - \rho_A(\lambda)$$

For open oceans, the NIR $\rho_w(\lambda) = 0$, NIR black ocean assumption.

$$\underbrace{\rho'(\lambda)}_{\text{NIR 765, 865 nm}} \overset{\text{N Models}}{\Rightarrow} \varepsilon(765, 865) \Rightarrow \text{2 Aerosol Models} \rightarrow$$

$$\rightarrow \varepsilon(765, 865) \overset{\text{2 Models}}{\Rightarrow} \varepsilon(\lambda, 865) \rightarrow \rho_{as}(865) \overset{\varepsilon(\lambda, 865)}{\Rightarrow} \rho_{as}(\lambda) \rightarrow$$

$$\rightarrow \rho_{as}(\lambda) \overset{\text{2 Models}}{\Rightarrow} \rho_A(\lambda) \longrightarrow t(\lambda)\rho_w(\lambda) = \rho'(\lambda) - \rho_A(\lambda)$$

The same two aerosol models are used to compute the diffuse transmittances.

Some Other Details

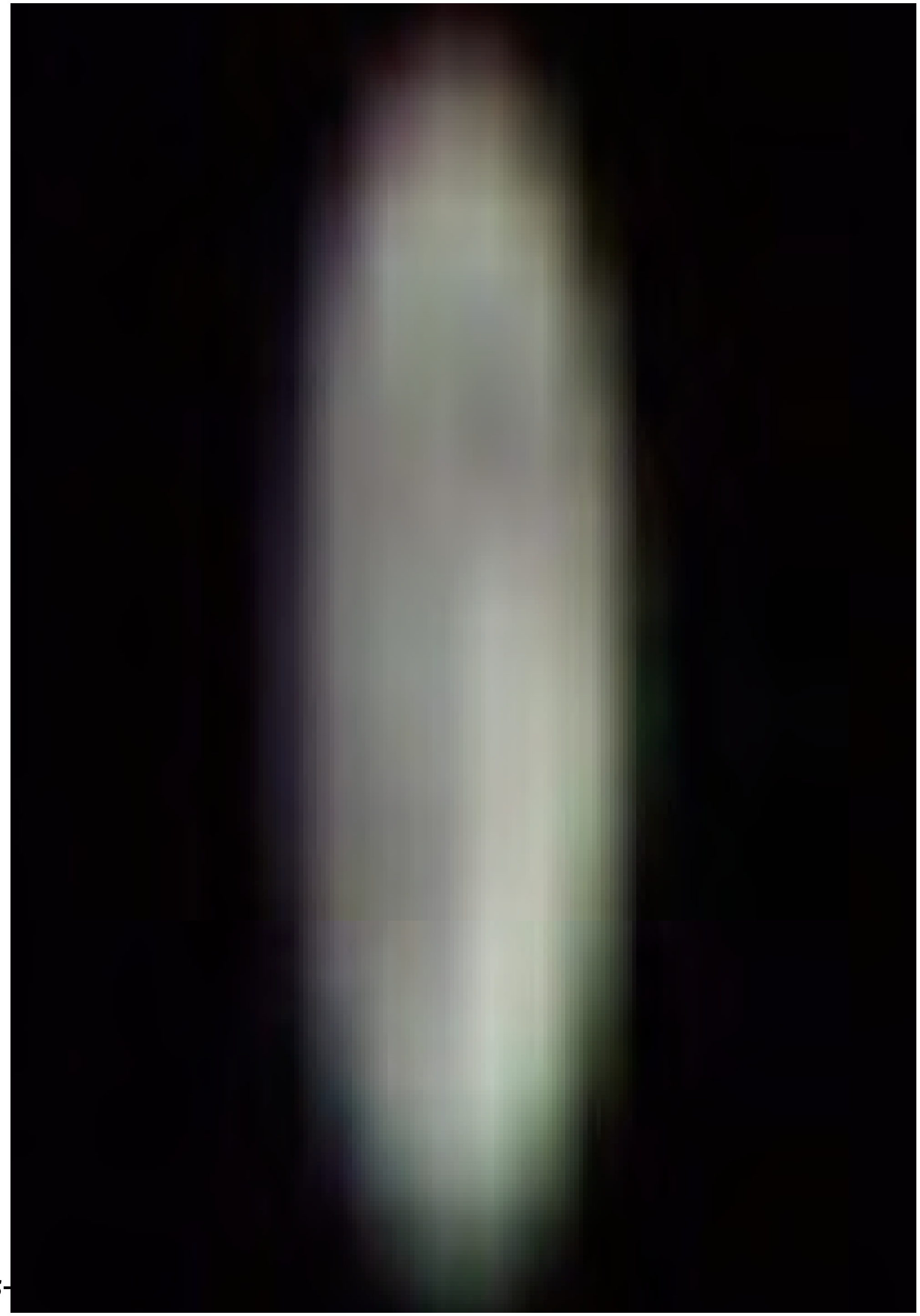
- **Calibrations:** (a)**Lunar calibration** to remove sensor degradation, and (b) **Vicarious calibration** using the in situ MOBY data to set gains. It requires **~0.1%** accuracy.
- **Cloud Masking for SeaWiFS/MODIS:** A simple reflectance threshold technique has been used**,** **$(\rho_t - \rho_r) \leq 2.7\%$** at the NIR (865 nm) band as being identified as clear sky.
- **Surface effects:** Ocean is assumed to be black at the NIR bands**,** modifications are made to account for the NIR ocean contributions at the NIR bands for **productive oceans**.
- A flat **Fresnel-reflecting surface** was used in RTE computations for aerosol lookup tables, while the Cox & Monk (1954) surface roughness model (wind dependent) was used for the Rayleigh lookup tables.
- **Sun glint** is avoided/masked and sun glint contamination is corrected using Wang and Bailey (2001).
- **Aerosol Models:** A set of 12 aerosol models from or derived from the work of **Shettle and Fenn** (1979) was used.
- **Validations:** SeaWiFS ocean color and aerosol products have been validated through the NASA SeaWiFS and SIMBIOS projects, and from in situ data acquired from various field campaigns in global open oceans.

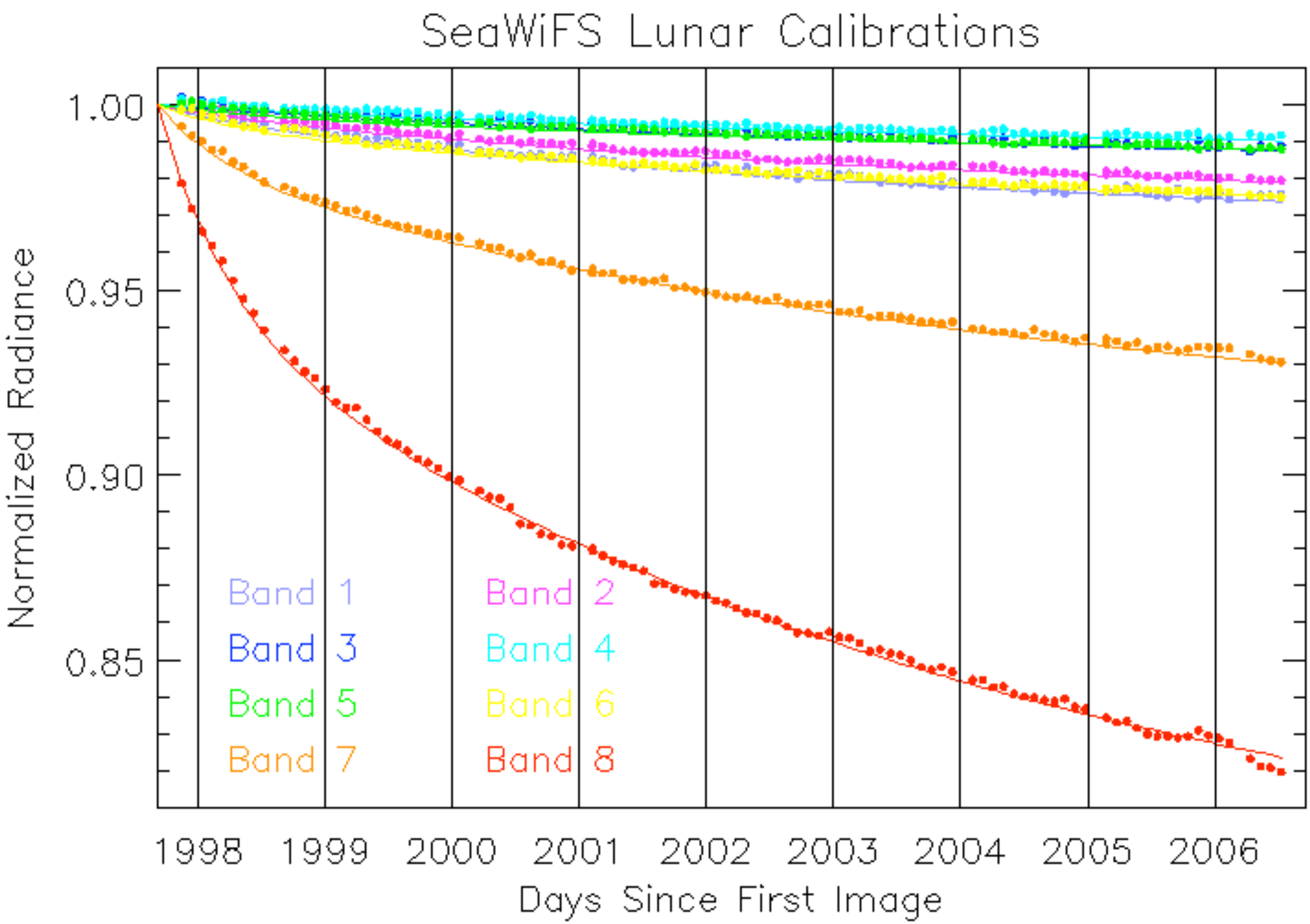
Example of Satellite Ocean Color Sensor

SeaWiFS

Sea-Viewing **Wi**de-**F**ield-of-view **S**ensor

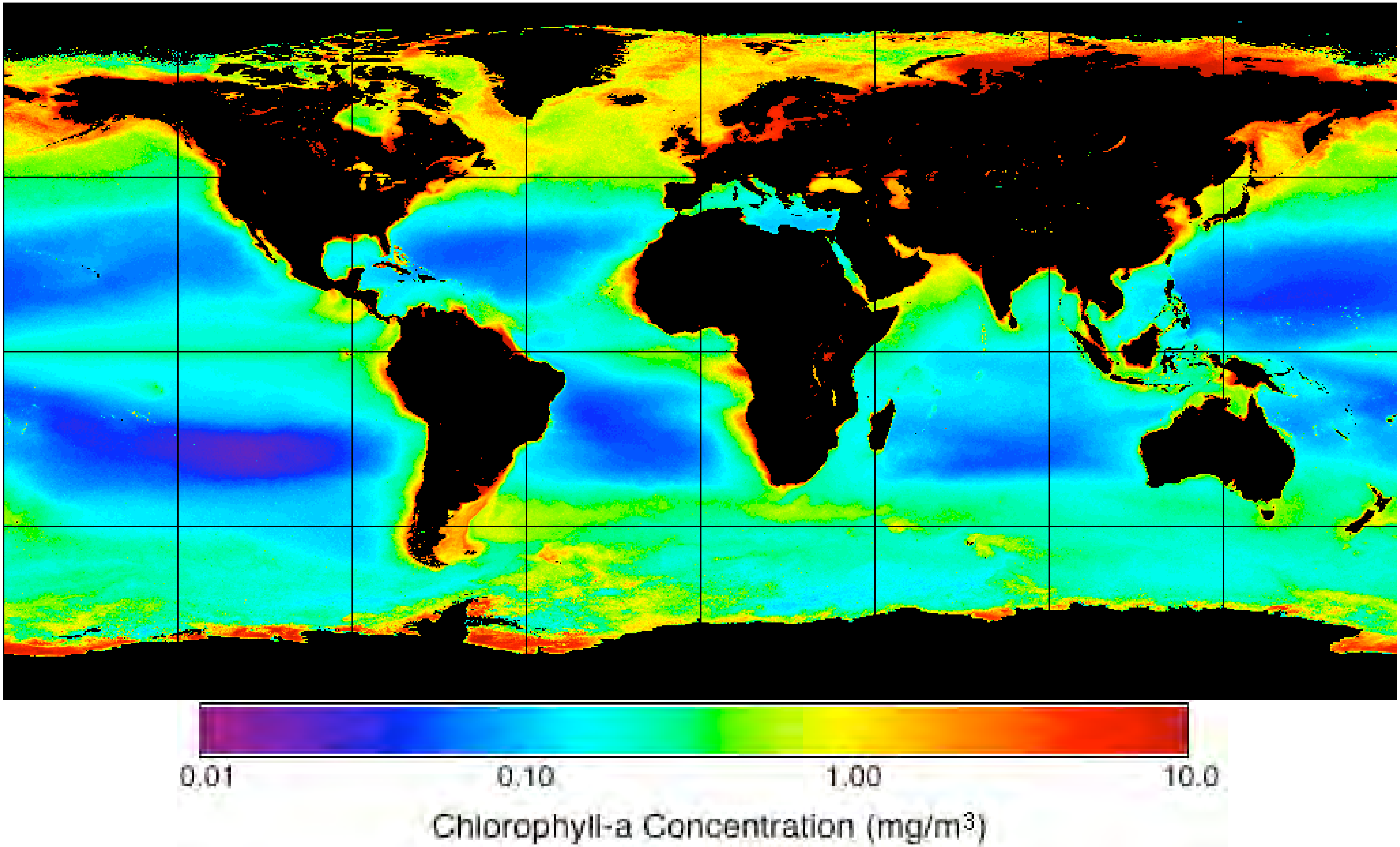
SeaWiFS Looks at the Moon



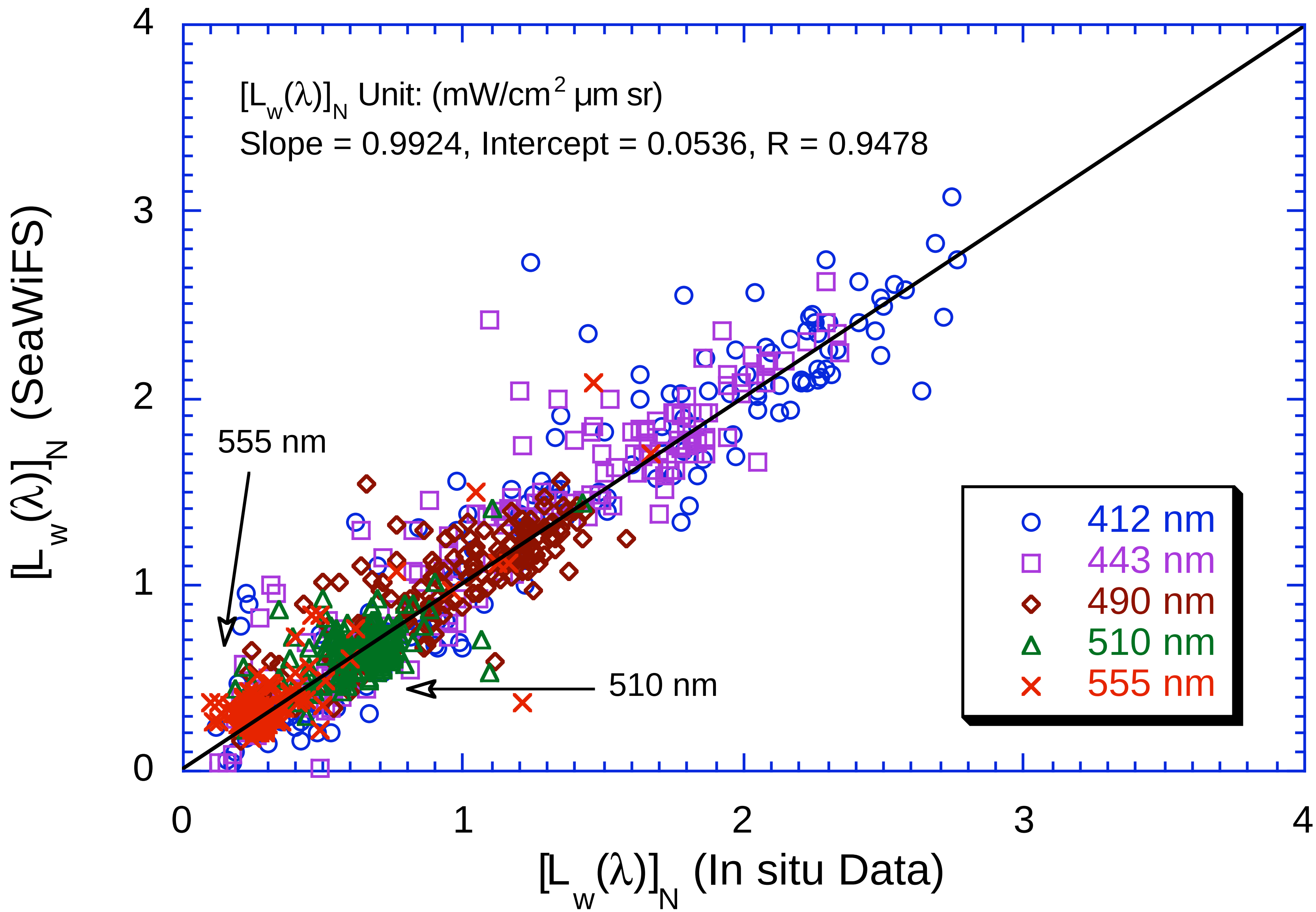


From http://oceancolor.gsfc.nasa.gov

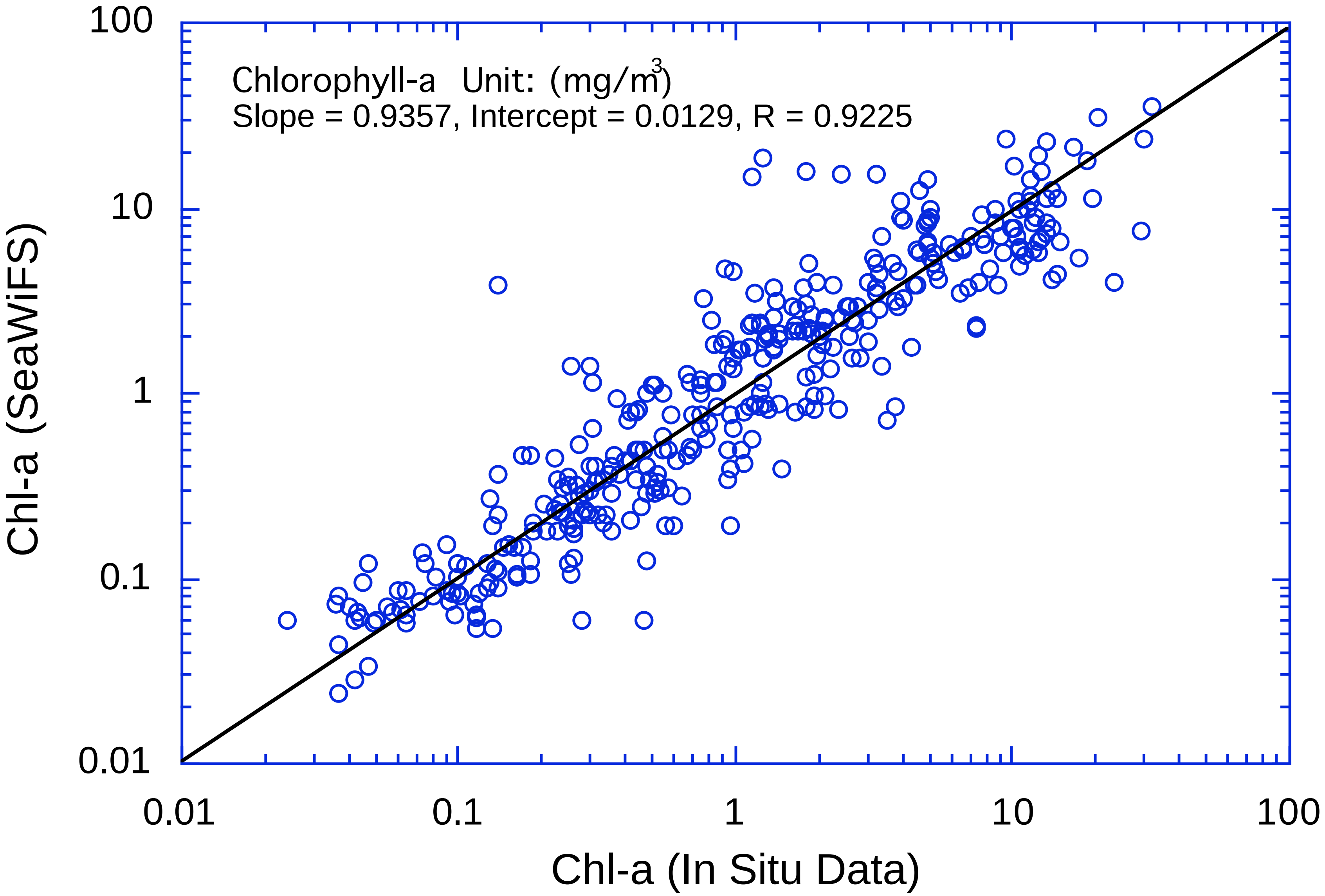
SeaWiFS Chlorophyll-a Concentration
(October 1997-December 2003)



SeaWiFS experiences demonstrate that the atmospheric correction works well in the **open oceans**.



SeaWiFS Chlorophyll-a Comparison



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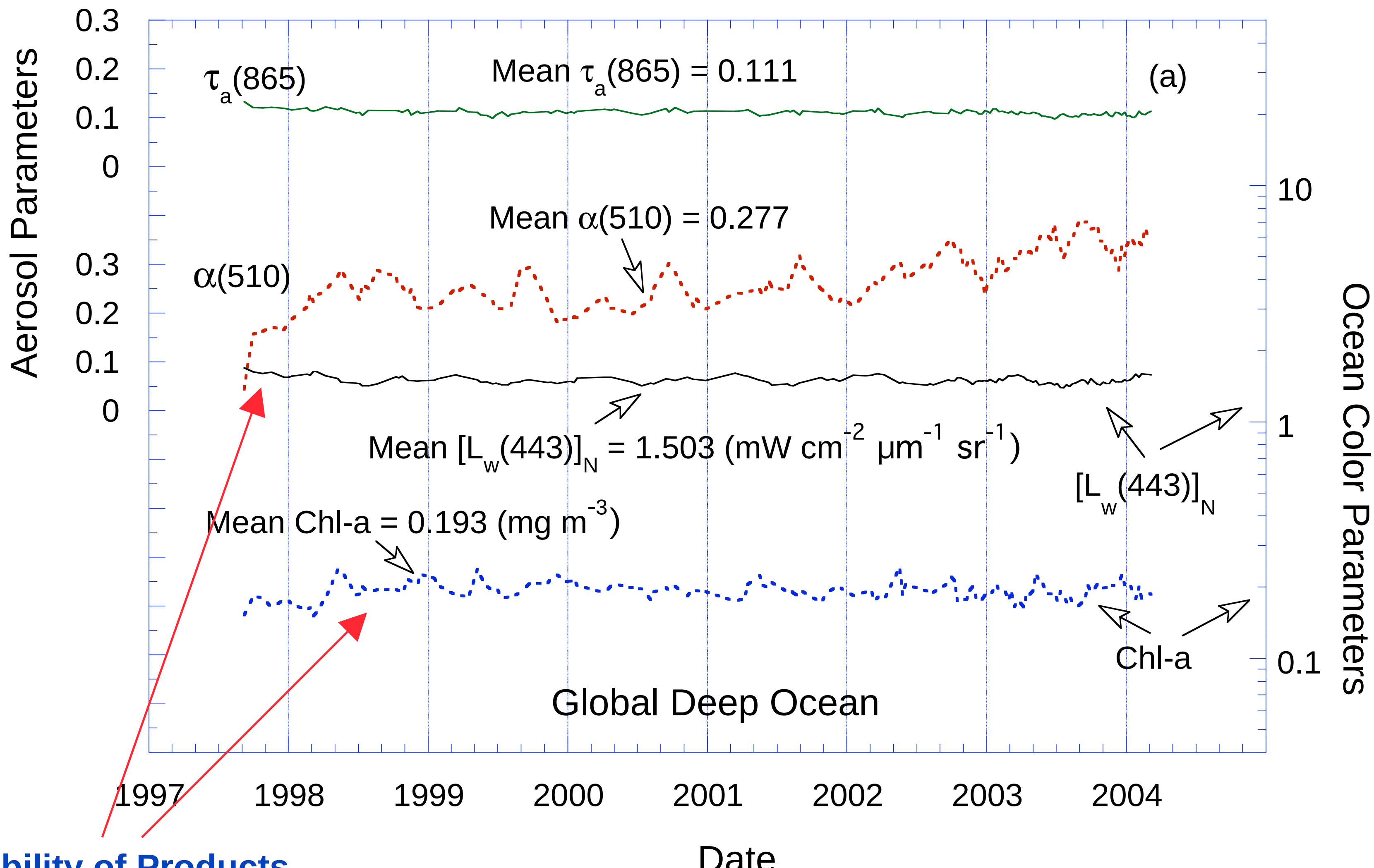
SeaWiFS Global Deep Ocean Results

(Wang et al., 2005)



Stability of Products

SeaWiFS and MODIS Experiences Show:

High quality ocean color products for the global open oceans (Case-1 waters).

Significant efforts are needed for improvements of water color products in the inland & coastal regions:

- **Turbid Waters**
  (violation of the NIR black ocean assumption)
- **Strongly-Absorbing Aerosols**
  (violation of non- or weakly absorbing aerosols)

Questions?

My question: If we use all derived parameters (all radiance contributions) from atmospheric correction (Gordon & Wang, 1994) procedure to re-compute L_t(computed) and comparing with satellite-measured L_t(measured), what you will get?