Atmospheric correction of satellite ocean colour observations

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Lecture content

- What atmospheric correction is supposed to do?
- What are the accuracy requirements for atmospheric correction of satellite OCR? What an algorithm has to achieve to be qualified for OCR atmospheric correction?
- What the total signal measured by the sensor is made of, and how the various contributions vary spectrally
- Some basic principles of OCR atmospheric corrections
- How OCR atmospheric correction can be performed under simplified assumptions in a number of situations
- How most modern OCR atmospheric correction schemes work
- Other "terms" of the atmospheric correction problem (e.g., glint)
- Alternative approaches to the "aerosol-model-based" schemes
- Under which conditions modern OCR atmospheric correction schemes still fail
- Current issues (turbid waters, absorbing aerosols, high spatial resolution sensors)

Setting up of "the scene" : can this really work?

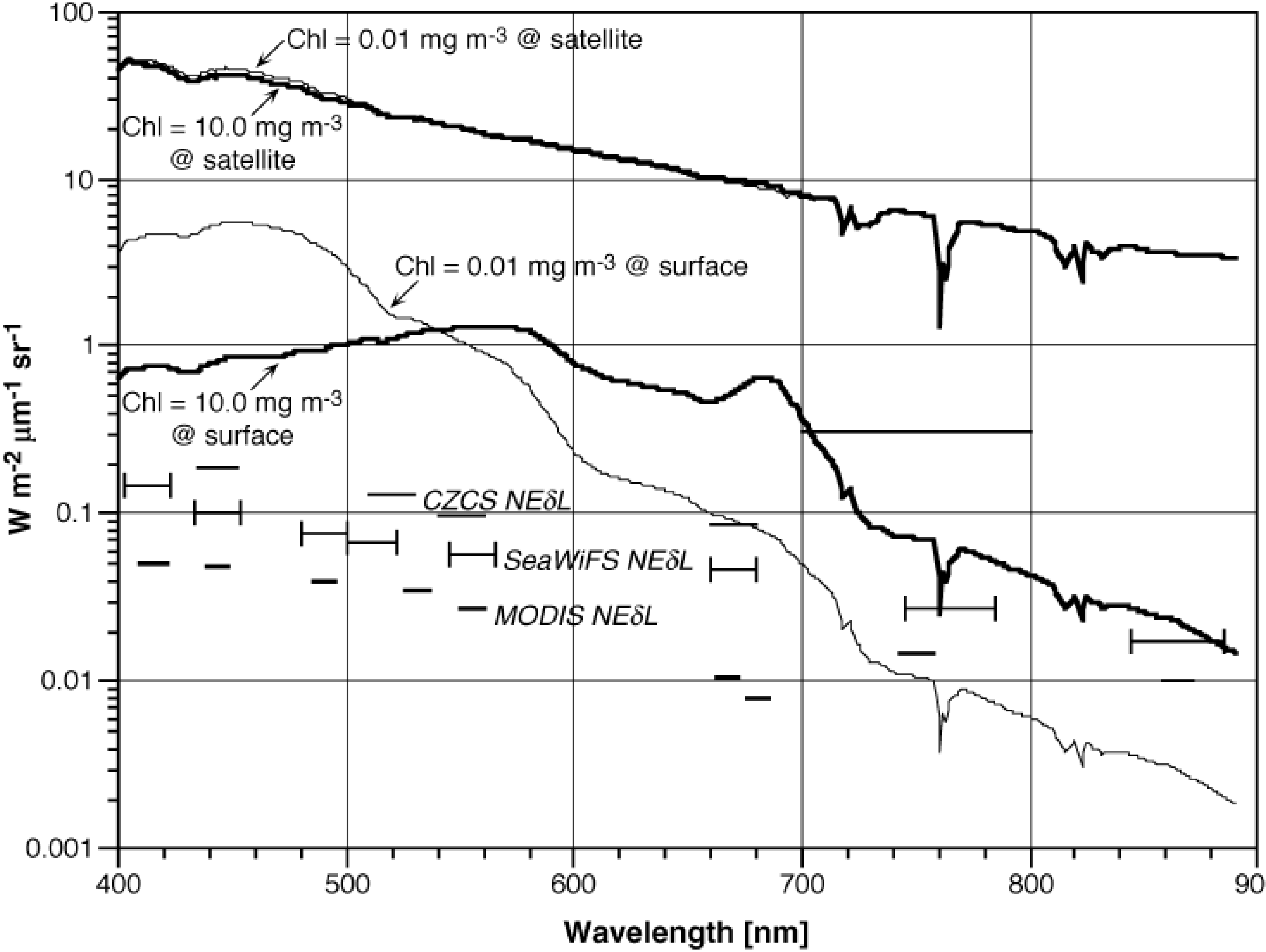
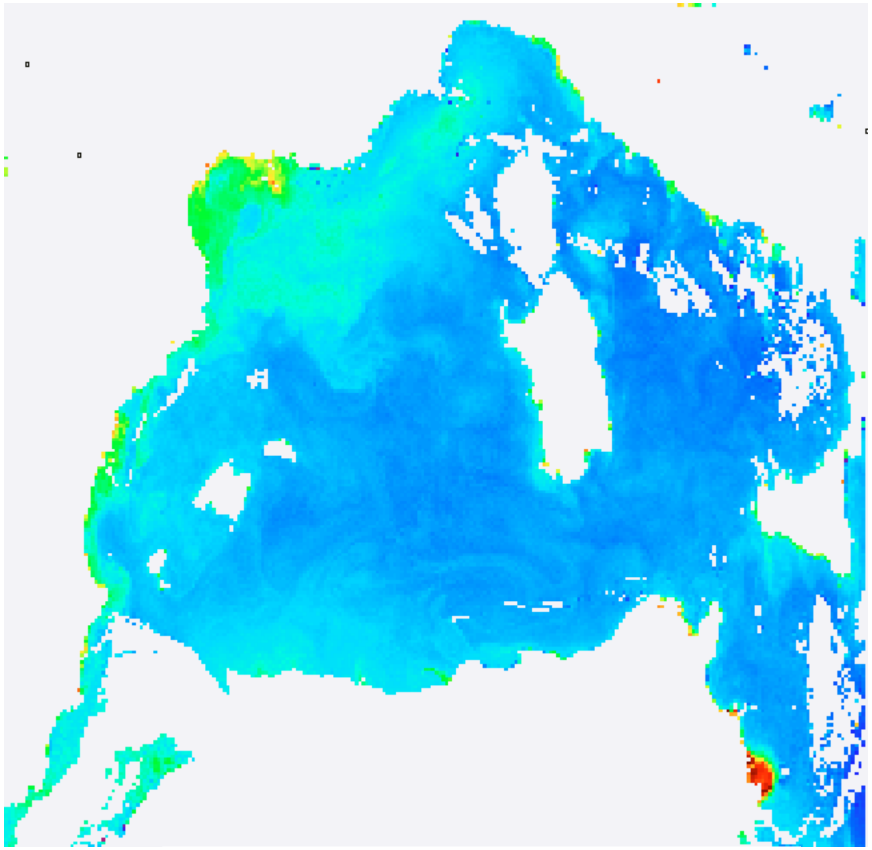


Figure 3 in Hooker SB et al., 1992. An overview of SeaWiFS and ocean color, NASA TM 104566, vol 1, NASA GSFC, Greenbelt, MD 20771

Setting up of "the scene"



We want to end up with this…

Images created from SeaWiFS products from the ocean color web:
https://oceandata.sci.gsfc.nasa.gov/cgi/getfile/S2000181114642.L1A_GAC.Z
https://oceandata.sci.gsfc.nasa.gov/cgi/getfile/S2000181114642.L2_GAC_OC.nc

Atmospheric “correction”?

Estimating the “atmospheric path radiance”, L_{path}, in order to get to the “marine signal”, i.e., the water-leaving radiance, L_w

L_{path} is generated by scattering in the atmosphere (molecules and aerosols). There is also a component of reflection at the air-water interface

The difficulty is: aerosols (quantity and type) are unknown

The difficulty is: L_{path} is about 10 times bigger than L_w…so errors on it have a much larger (relative) effect on L_w

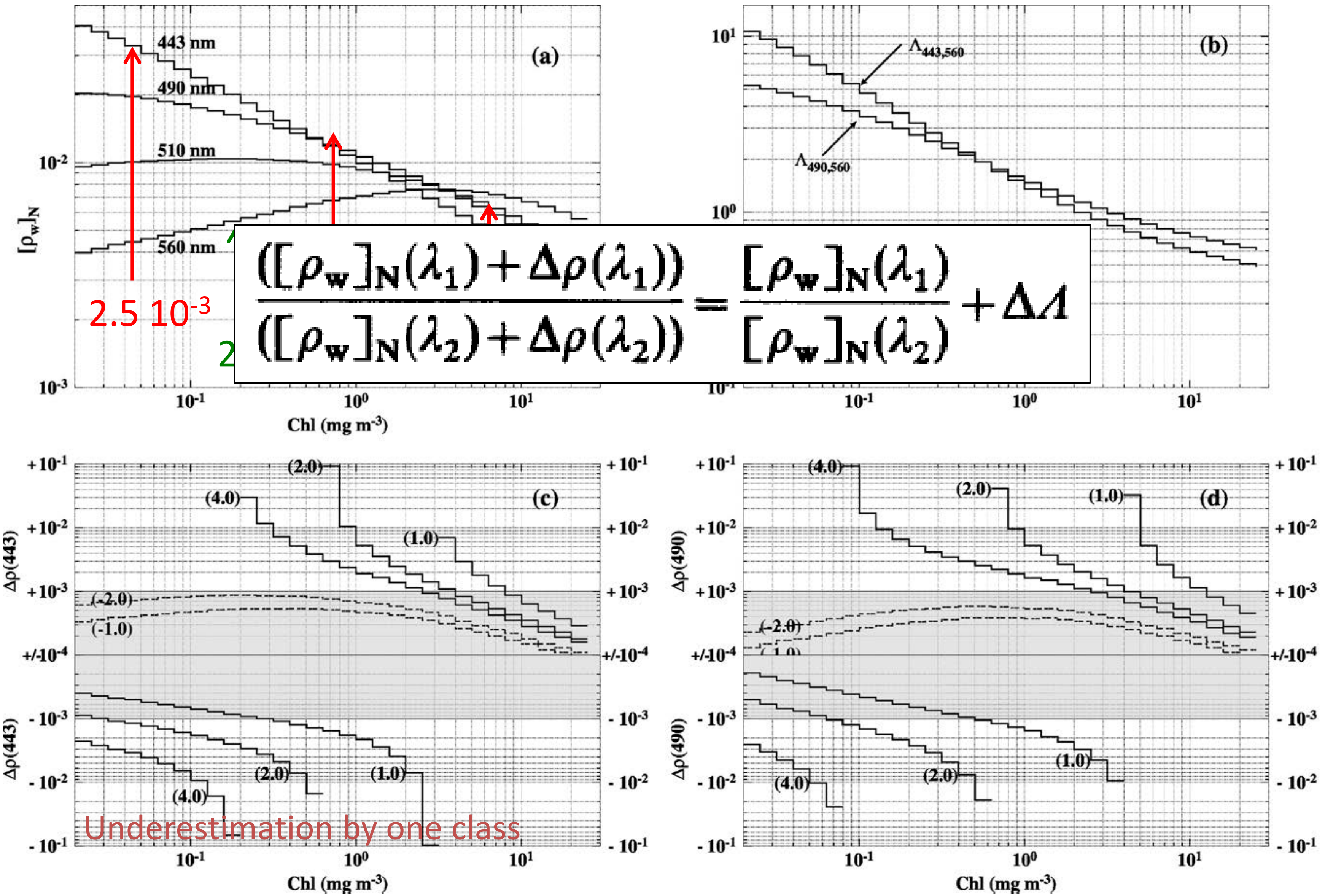
Accuracy requirements

Should normally be the first step for any development of an atmospheric correction algorithm

- "Historically": satellite ocean color was supposed to deliver reflectances or water-leaving radiances with the aim of deriving the chlorophyll concentration
- Therefore, the accuracy requirement was defined *w.r.t*. chlorophyll, when derived from band-ratio techniques
- Basically, the question being: which errors can we tolerate on reflectances in the blue and green that we still can derive Chlorophyll within, say, ±30%?
- The answer was: 5% in blue bands for an oligotrophic ocean (Gordon and Clark, 1981. Appl. Opt. 20:4175-4180)
- Somewhat more sophisticated requirements were derived for more recent missions, e.g., MERIS
- Goal: discriminate 10 values of chlorophyll concentration within each decade from 0.03 to 0.3, 3 and 30 mg m^{-3}, so 30 values.

- Antoine, D. and A. Morel, 1999. A multiple scattering algorithm for atmospheric correction of remotely-sensed ocean colour (MERIS instrument) : principle and implementation for atmospheres carrying various aerosols including absorbing ones, International Journal of Remote Sensing, 20, 1875-1916.
- Wang M, and HR Gordon, 2018. Sensor performance requirements for atmospheric correction of satellite ocean color remote sensing, Optics Express, 26(6), https://doi.org/10.1364/OE.26.007390

Accuracy requirements



$$\frac{([\rho_w]_N(\lambda_1)+\Delta\rho(\lambda_1))}{([\rho_w]_N(\lambda_2)+\Delta\rho(\lambda_2))} = \frac{[\rho_w]_N(\lambda_1)}{[\rho_w]_N(\lambda_2)} + \Delta\Lambda$$

Figure1 from : Antoine, D. and A. Morel, 1999. A multiple scattering algorithm for atmospheric correction of remotely-sensed ocean colour (MERIS instrument) : principle and implementation for atmospheres carrying various aerosols including absorbing ones, International Journal of Remote Sensing, 20, 1875-1916

Accuracy requirements

Conclusions from the MERIS exercise:

- The first requirement (retrieval of 30 reflectance values) requires that atmospheric correction errors be maintained within ±1-2 10^{-3} at 443 nm, within ±5 10^{-4} at 490 nm, and within ±2 10^{-4} at 560 nm.
- This value (±2 10^{-4}) at 560 nm is equal to the Ne$\Delta\rho$ specified for MERIS (see above).
- If it is assumed that atmospheric correction errors in the 440-500 nm domain are about twice the errors at 560 nm, the second requirement (discrimination of 30 (Chl) values) requires errors within ±1 10^{-3} at 443 nm (then ±5 10^{-4} at 560 nm), or within ±5 10^{-4} at 490 nm (then ±2 10^{-4} at 560 nm).
- When expressed as relative errors, all the above requirements represent about 1% of the normalised oceanic reflectances at 443 nm (and often 2-5%), except when (Chl) > 3 mg m^{-3}.
- The situation is about the same for the wavelength couple 490-560 nm, except when Chl < 0.1 mg m^{-3}, and when the Λ490,560 ratio is equal to 1.
- Errors have to be of the same sign at all wavelengths

Accuracy requirements

And what about other applications?

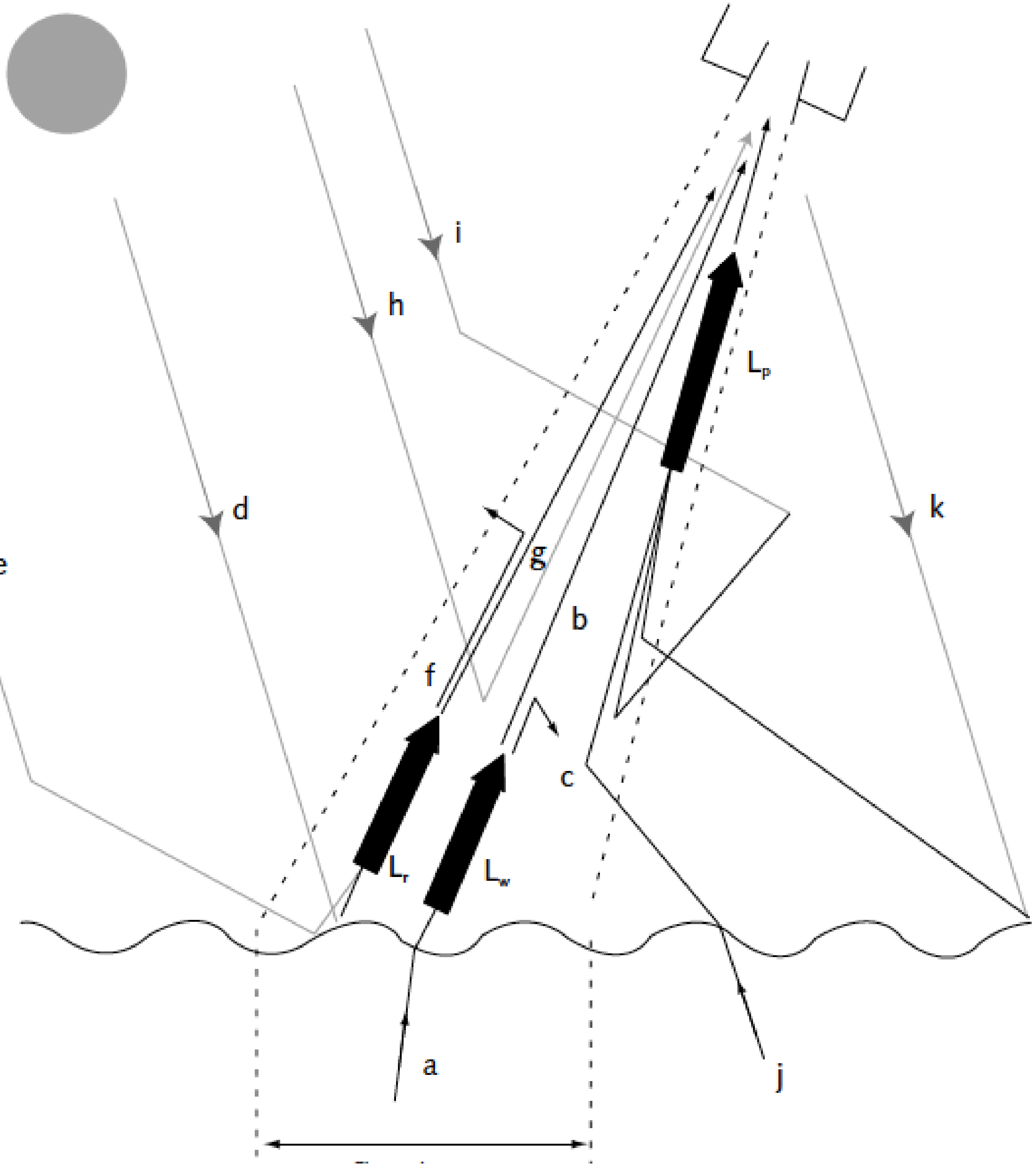
- SPM in turbid waters
- PFTs
- IOPs
- ….

Basically: we do not have any detailed requirements for these other uses of satellite ocean colour

But see:
Wang M, and HR Gordon, 2018. Sensor performance requirements for atmospheric correction of satellite ocean color remote sensing, Optics Express, 26(6), https://doi.org/10.1364/OE.26.007390

Another question is: is my sensor worth it? Do I need to have a sophisticated AC algorithm if my sensor does not have a very high radiometric accuracy?

What makes the TOA total radiance?



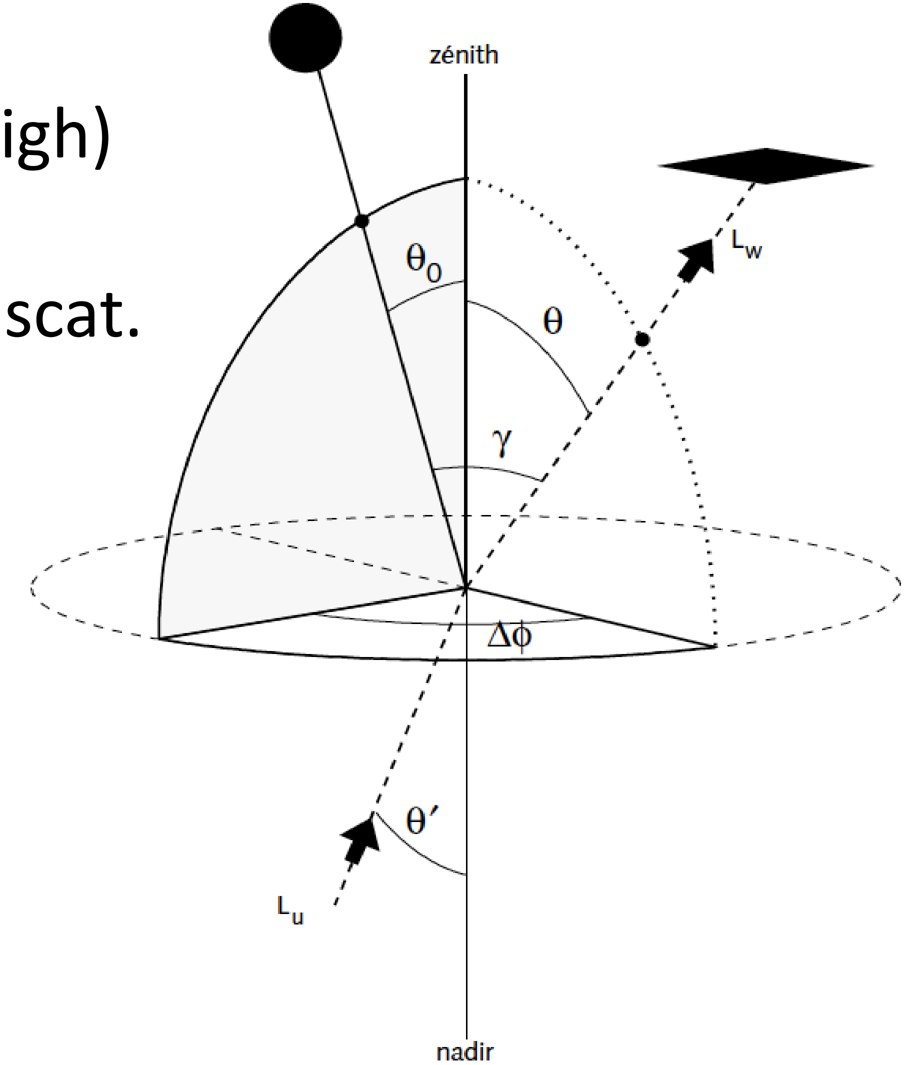
IOCCG Summer Lecture Series 2018. Lecture on Atmospheric corrections 1/2

The TOA total radiance

$$L_t = L_r + L_a + L_{ra} + T.L_g + t.L_{wc} + t.L_w$$

L_r : Molecular scattering (Rayleigh)
L_a : Aerosol scattering
L_{ra} : Molecular-aerosol coupled scat.
L_g : Sun glint
L_{wc} : White caps
L_w : Water-leaving signal
t, T: Transmittances

θ_s or θ_0: sun zenith angle
θ_v: satellite view angle
$\Delta\phi$: azimuth difference



Optical thickness

$$\tau_\lambda = \int_0^{z_{top}} b_{ext}(\lambda) dz$$

A dimensionless quantity related to how much attenuation occurs in the entire atmosphere

Can be defined for:

- Molecular scattering only
- Aerosol attenuation or scattering (or absorption)
- Everything you have in the atmosphere

“single scattering albedo”

b/c,

- Dimensionless
- 1 if no absorption

Molecular (Rayleigh) scattering

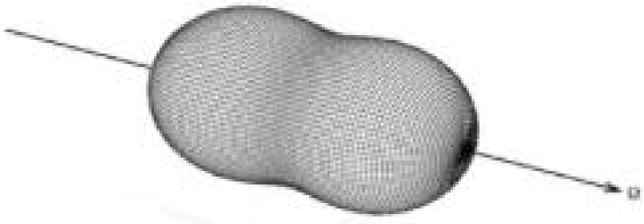
L_r is a large contribution to the total signal

$\tau_r = P/P0.[8.524e\text{-}3.\lambda^{-4} + 9.63e\text{-}5.\lambda^{-6} + 1.1e\text{-}7.\lambda^{-8}]$ with λ in μm

The Rayleigh scattering phase function should be computed as :

$$\tilde{\beta}_r(\psi,\lambda) = \frac{3}{4\pi * 4\left(1+2\gamma\right)} \left[\left(1+3\gamma\right)+\left(1-\gamma\right)\cos^2(\psi)\right]$$

where



$$\gamma = \frac{\delta}{2-\delta}$$

and δ is the depolarisation factor = 0.0279.

Actually δ varies spectrally (0.0279 is an acceptable average over 450-900nm)

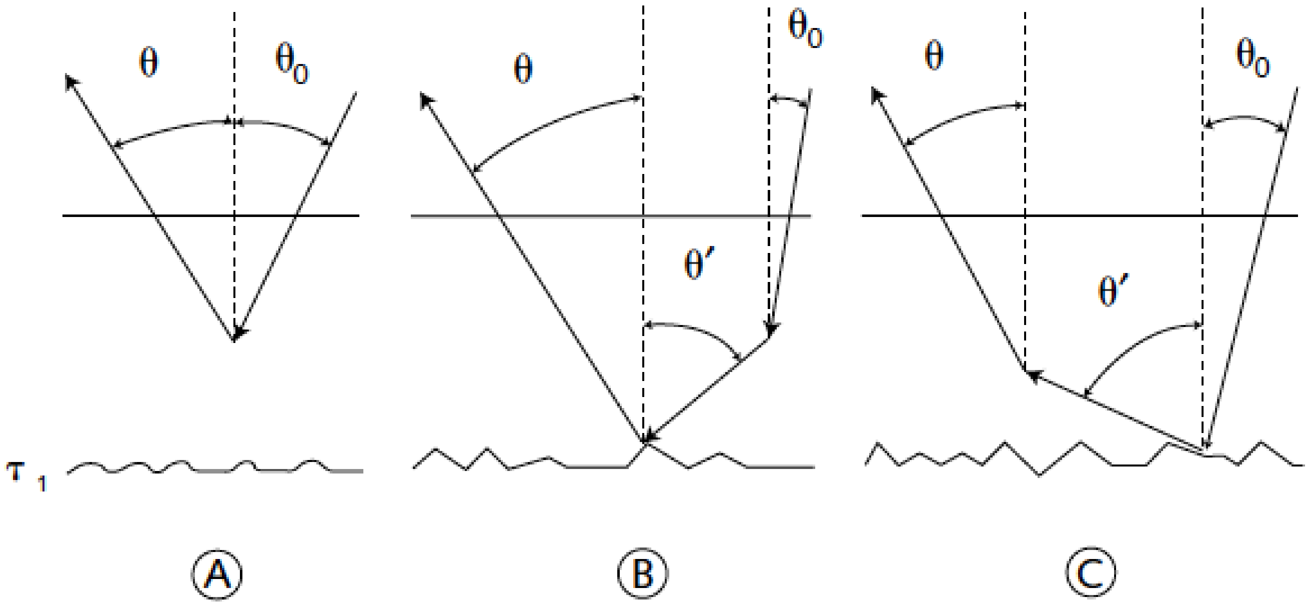
λ	412.5	442.5	490.0	510.0	560.0	620.0	665.0	708.75	778.75	865.0
τ_r	0.314085	0.235229	0.154853	0.131495	0.089808	0.059387	0.044703	0.034542	0.023609	0.015456

Hansen, J.E., and L.D. Travis, 1974: Light scattering in planetary atmospheres. *Space Sci. Rev.*, **16**, 527-610, doi:10.1007/BF00168069.

Molecular (Rayleigh) scattering

L_r : function of the amount of molecules (atmospheric pressure). In a single scattering regime:

$$
\begin{aligned}
L_R(\lambda,\theta_0,\theta_v,\Delta\phi) = & \left[\frac{F_0\,\mu_0\,\omega_R}{\mu_v+\mu_0}\,\rho_R(\gamma^-)\left(1-e^{-\tau_R\left(\frac{1}{\mu_v}+\frac{1}{\mu_0}\right)}\right)\right] && \text{(A)}\\
& +\left[\frac{F_0\,\mu_0\,\omega_R}{\mu_s-\mu'}\,\rho_R(\gamma^+)\,\rho_F(\theta_v)\,e^{-\frac{\tau_R}{\mu_v}}\left(e^{\frac{-\tau_R}{\mu_0}}-e^{\frac{-\tau_R}{\mu'}}\right)\right] && \text{(B)}\\
& +\left[\frac{F_0\,\mu'\,\omega_R}{\mu'-\mu_v}\,\rho_R(\gamma^+)\,\rho_F(\theta_s)\,e^{-\frac{\tau_R}{\mu_0}}\left(e^{\frac{-\tau_R}{\mu'}}-e^{\frac{-\tau_R}{\mu_v}}\right)\right] && \text{(C)}
\end{aligned}
$$



Wang M., 1994. Atmospheric correction of the second generation ocean color sensors, dissertation presented in fulfil-lment of the requirement for
degree of doctor of philosophy (PhD), University of Miami, Coral Gables, Florida, USA.

Molecular (Rayleigh) scattering

L_r can also be computed using radiative transfer, in order to include all orders of scattering

Will need to consider:

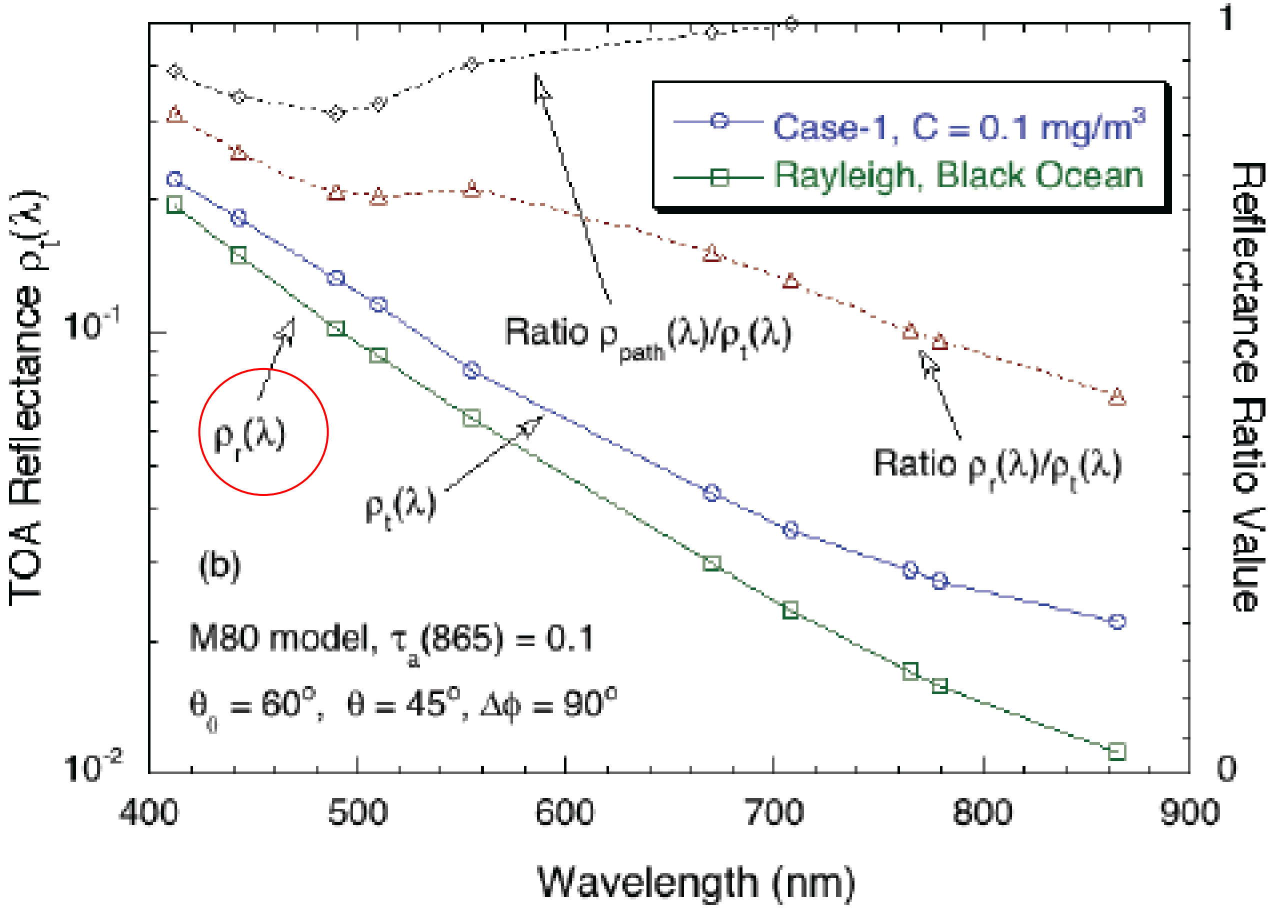
- Wavelength: Rayleigh optical thickness for standard atmospheric pressure
- Geometry
- Actual atmospheric pressure
- Wind speed (surface roughness)
- Polarisation

RT through: Monte Carlo, Successive Orders of scattering, Matrix operator method, …

So you can build a “Look-up Table” (LUT): $L_R(\lambda, \theta_s, \theta_v, \Delta\phi, P, Wind)$

Molecular (Rayleigh) scattering

L_r is a large contribution to the total signal

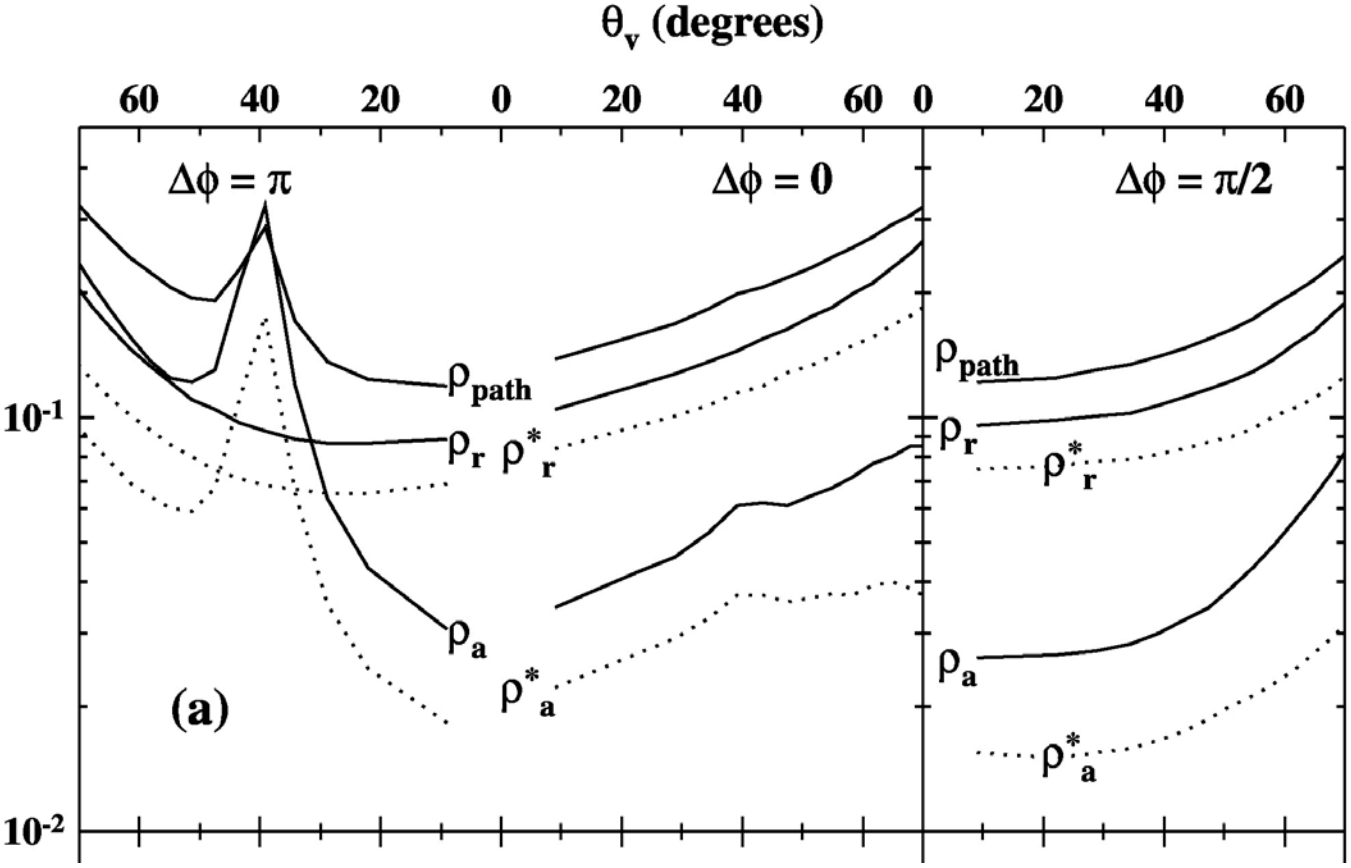


$\lambda^{-4.09}$ spectral dependence

From Figure 3.2 in IOCCG (2010). Atmospheric Correction for Remotely-Sensed Ocean-Colour Products. Wang, M. (ed.), Reports of the International Ocean-Colour Coordinating Group, No. 10, IOCCG, Dartmouth, Canada.

Molecular (Rayleigh) scattering

Variations in the swath



From Figure 4 from: Antoine, D. and Morel, A. (1998). Relative importance of multiple scattering by air molecules and aerosols in forming the atmospheric path radiance in the visible and near-infrared parts of the spectrum, *Applied Optics*, **37**, 2245-2259.

Aerosol scattering

L_a contribution to the total signal is proportional to the amount of aerosol particles in the atmosphere (to a first degree)

Unknown a priori

Aerosol optical properties are determined by their composition (so: refractive index) and size distribution (PSD)

Scattering by particles with size lower or similar to the wavelength: IOP computed through the Mie theory

Because the average particle sizes are close to the visible wavelengths, scattering is highly peaked in the forward direction

Aerosol size distributions

Purely maritime type

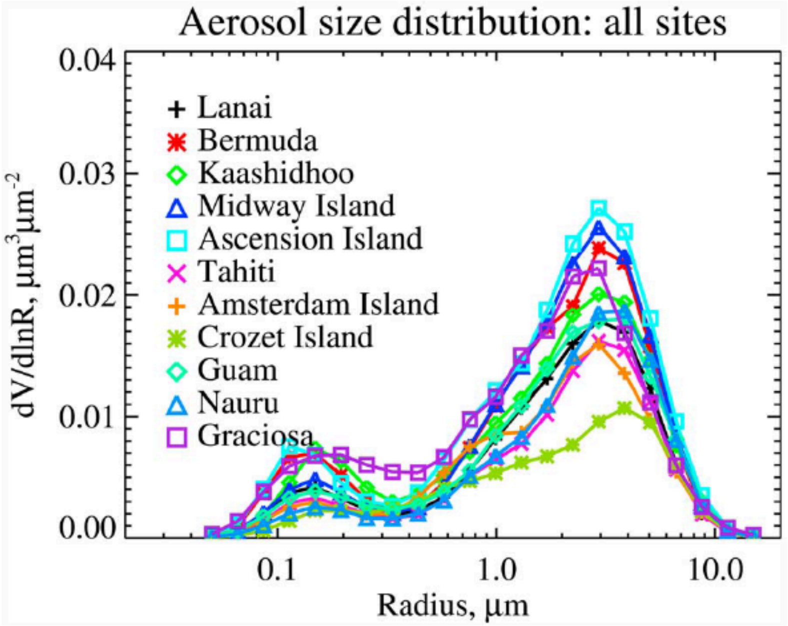


Figure 3 in: Sayer et al., 2012. J. Geophys. Res., VOL. 117, D05213, doi:10.1029/2011JD016689, 2012

Urban / industrial type

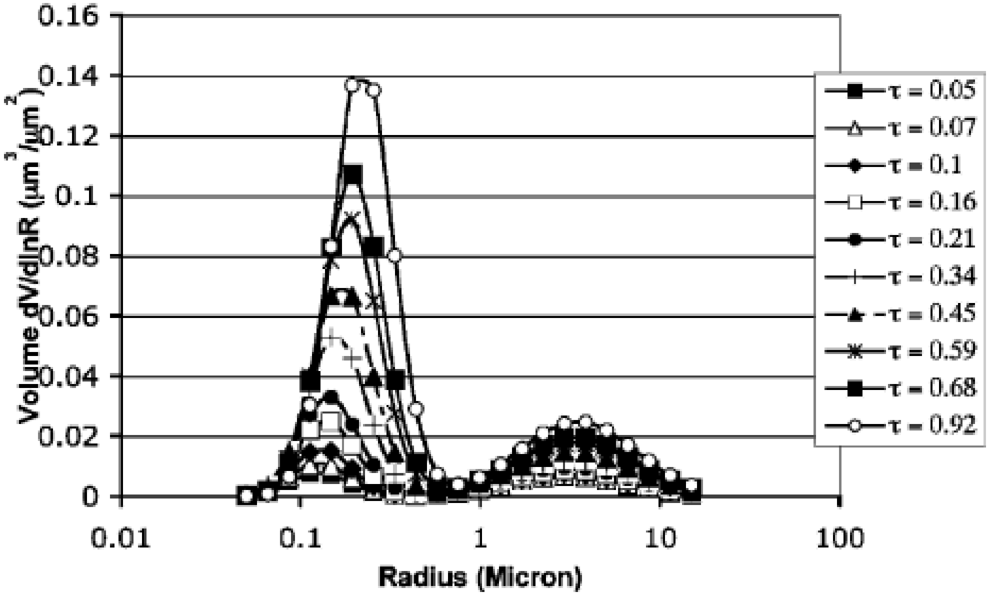


FIG. 10. Volume size distribution for the dynamic urban/industrial aerosol model of Dubovik et al. (2002). Curves are functions of optical depth at 0.44 μm (τ).

Figure 10 in: Levy et al., 2005. Journal of the Atmospheric Sciences—special Section, vol 62, 974-992

The more abundant small particles are, the more “selective” scattering will be

Aerosol refractive indices

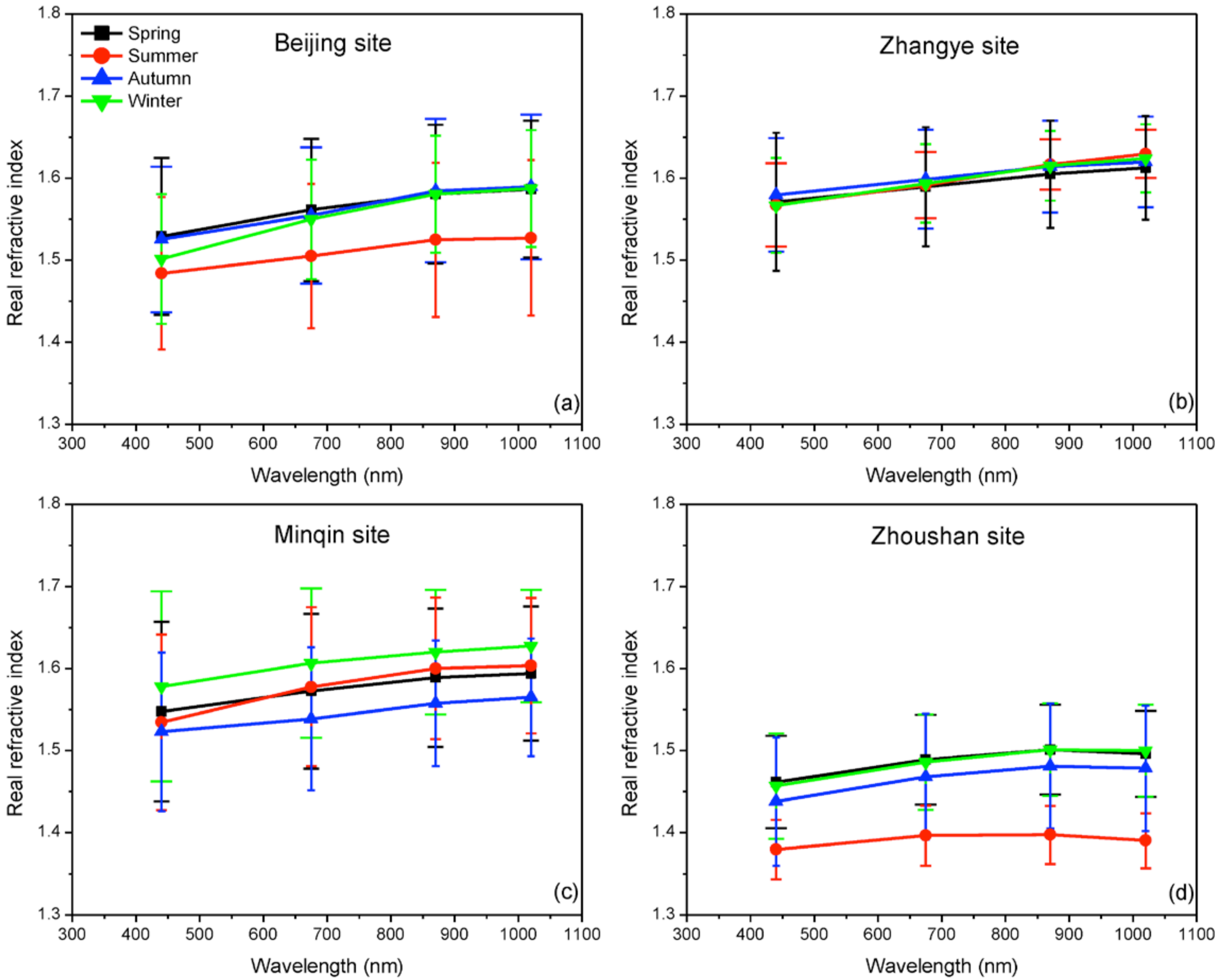


Figure 5 in: Xie et al., 2015. Remote Sens. 2015, 7, 9928-9953; doi:10.3390/rs70809928

Aerosol "single scattering albedo" b/c

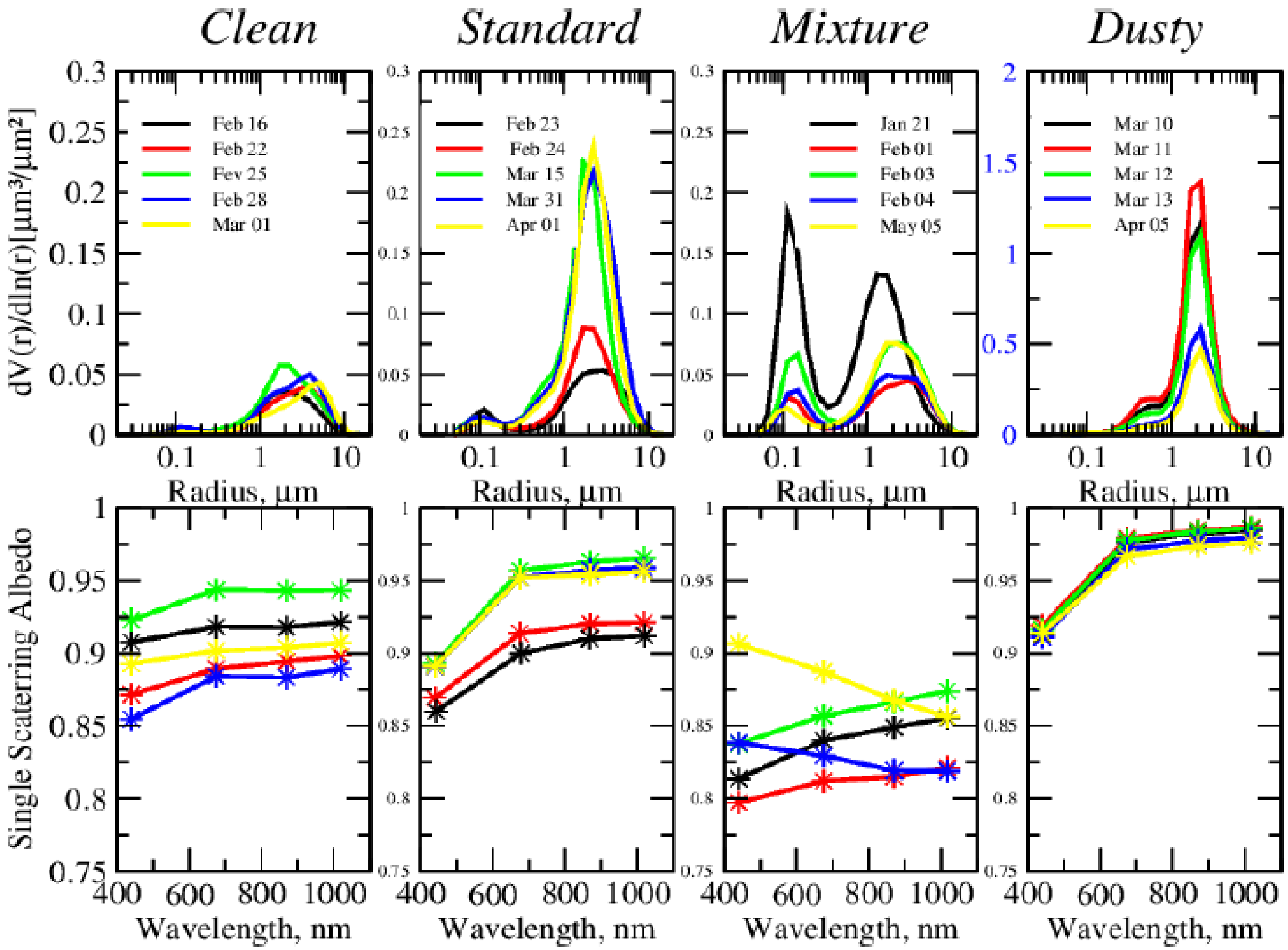
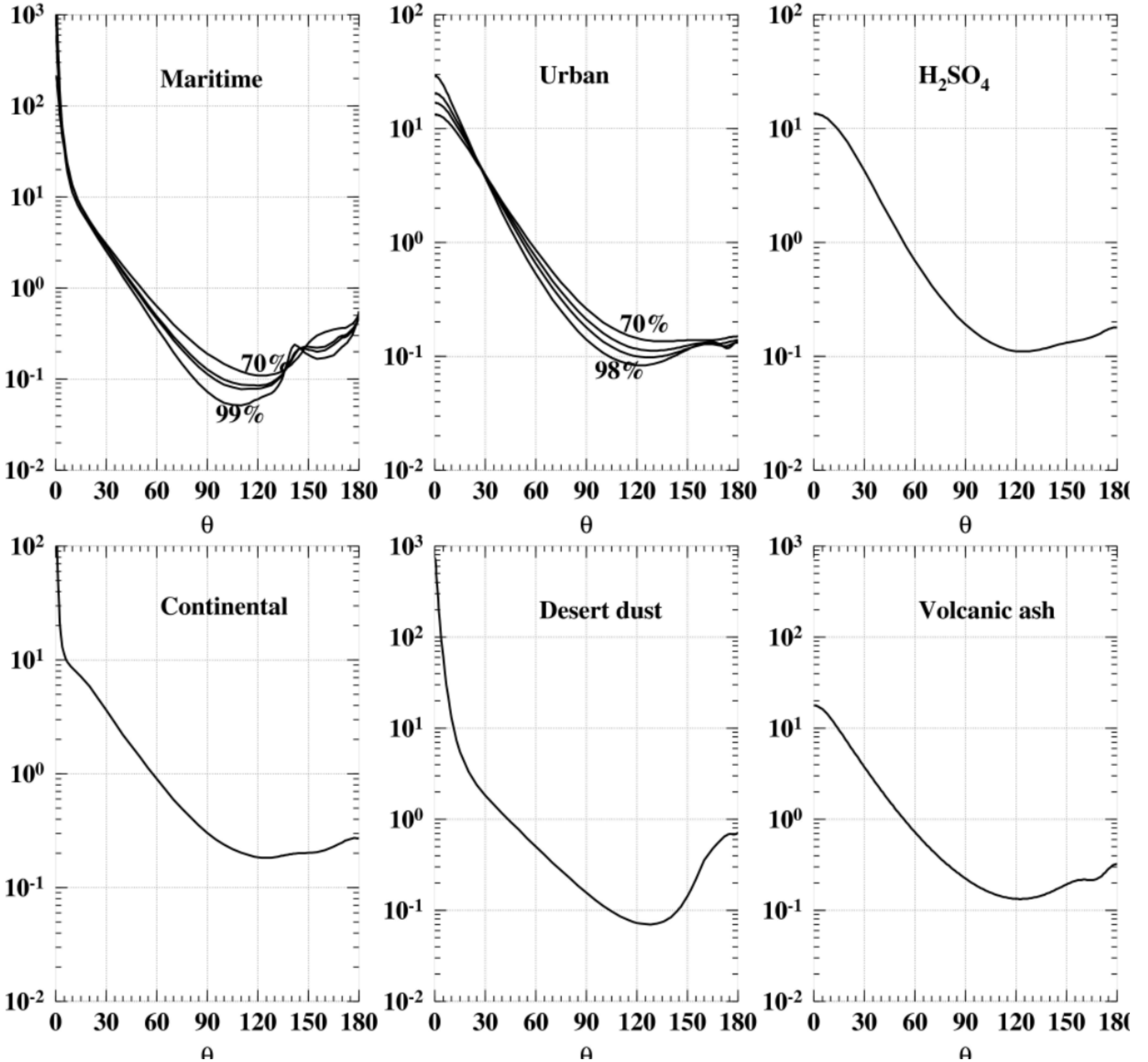


Figure 2 in: Simina Drame et al., 2015. Atmosphere, 6, 1608-1632; doi:10.3390/atmos6111608

Aerosol volume scattering functions

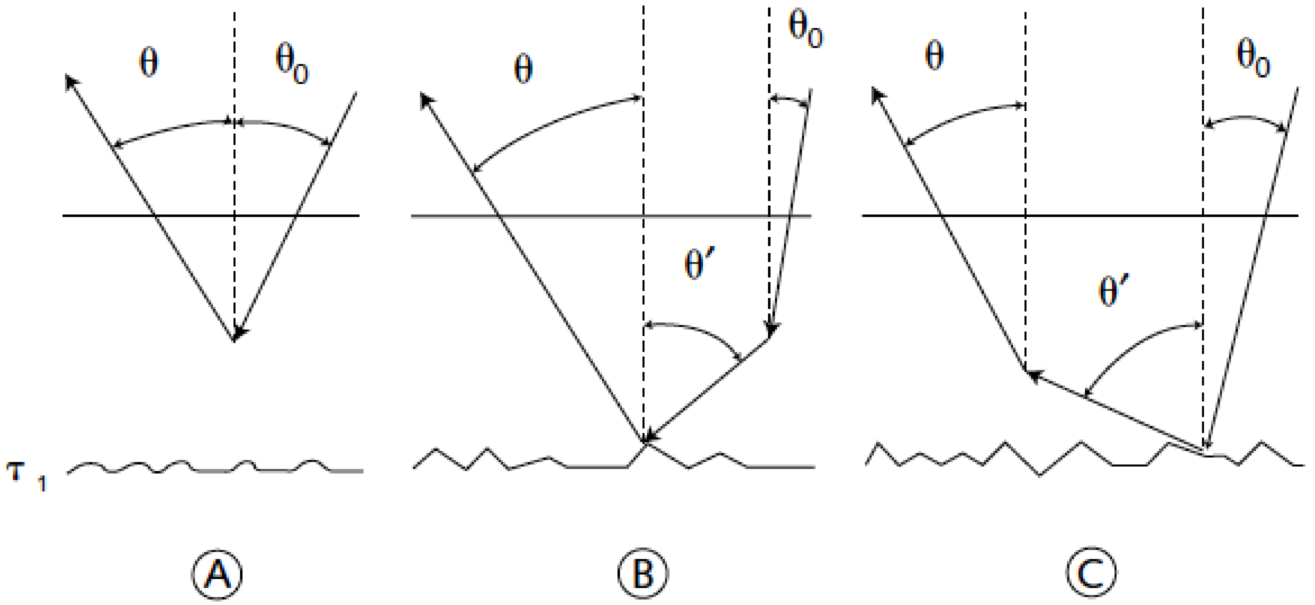


λ=550nm

Aerosol scattering

L_a. Single scattering approximation, again

$$
\begin{aligned}
L_R(\lambda,\theta_0,\theta_v,\Delta\phi) = {} & \left[\frac{F_0\,\mu_0\,\omega_R}{\mu_v+\mu_0}\,p_R(\gamma^-)\left(1-e^{-\tau_R\left(\frac{1}{\mu_v}+\frac{1}{\mu_0}\right)}\right)\right] && \text{Ⓐ}\\
& + \left[\frac{F_0\,\mu_0\,\omega_R}{\mu_s-\mu'}\,p_R(\gamma^+)\,\rho_F(\theta_v)\,e^{-\frac{\tau_R}{\mu_v}}\left(e^{\frac{-\tau_R}{\mu_0}}-e^{\frac{-\tau_R}{\mu'}}\right)\right] && \text{Ⓑ}\\
& + \left[\frac{F_0\,\mu'\,\omega_R}{\mu'-\mu_v}\,p_R(\gamma^+)\,\rho_F(\theta_s)\,e^{-\frac{\tau_R}{\mu_0}}\left(e^{\frac{-\tau_R}{\mu'}}-e^{\frac{-\tau_R}{\mu_v}}\right)\right] && \text{Ⓒ}
\end{aligned}
$$



Wang M., 1994. Atmospheric correction of the second generation ocean color sensors, dissertation presented in fulfil-lment of the requirement for
degree of doctor of philosophy (PhD), University of Miami, Coral Gables, Florida, USA.

Aerosol scattering

L_A, similarly to L_R, can also be computed using radiative transfer, in order to include all orders of scattering

Will need to pick up an aerosol model, and then to consider:

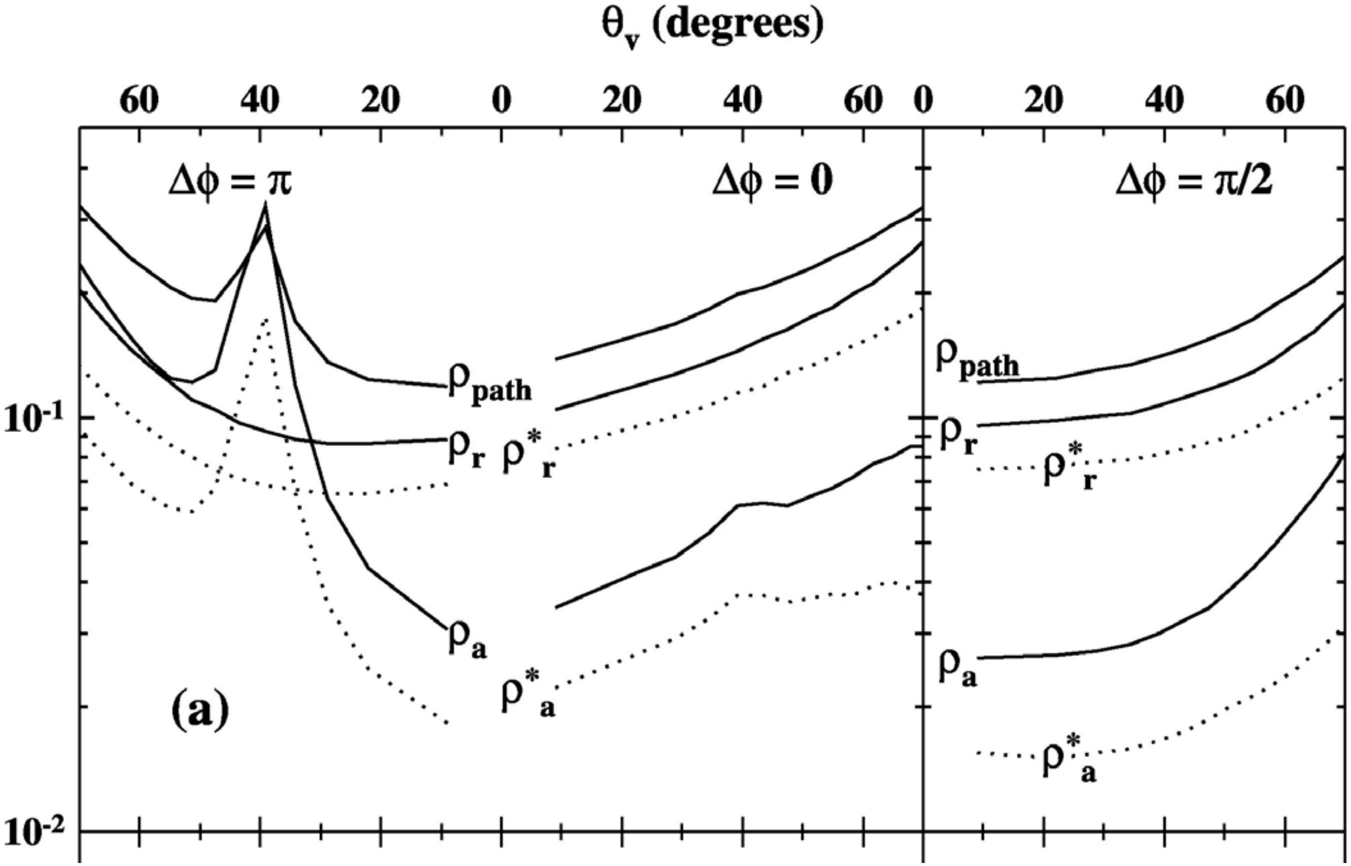
- Wavelength: aerosol optical thickness
- Geometry
- Wind speed (surface roughness)
- (Polarisation)

RT through: Monte Carlo, Successive Orders of scattering, Matrix operator method, …..

So you can build a "Look-up Table" (LUT): $L_A(\lambda, \theta_s, \theta_v, \Delta\phi, \tau_a$, aerosol type, Wind)

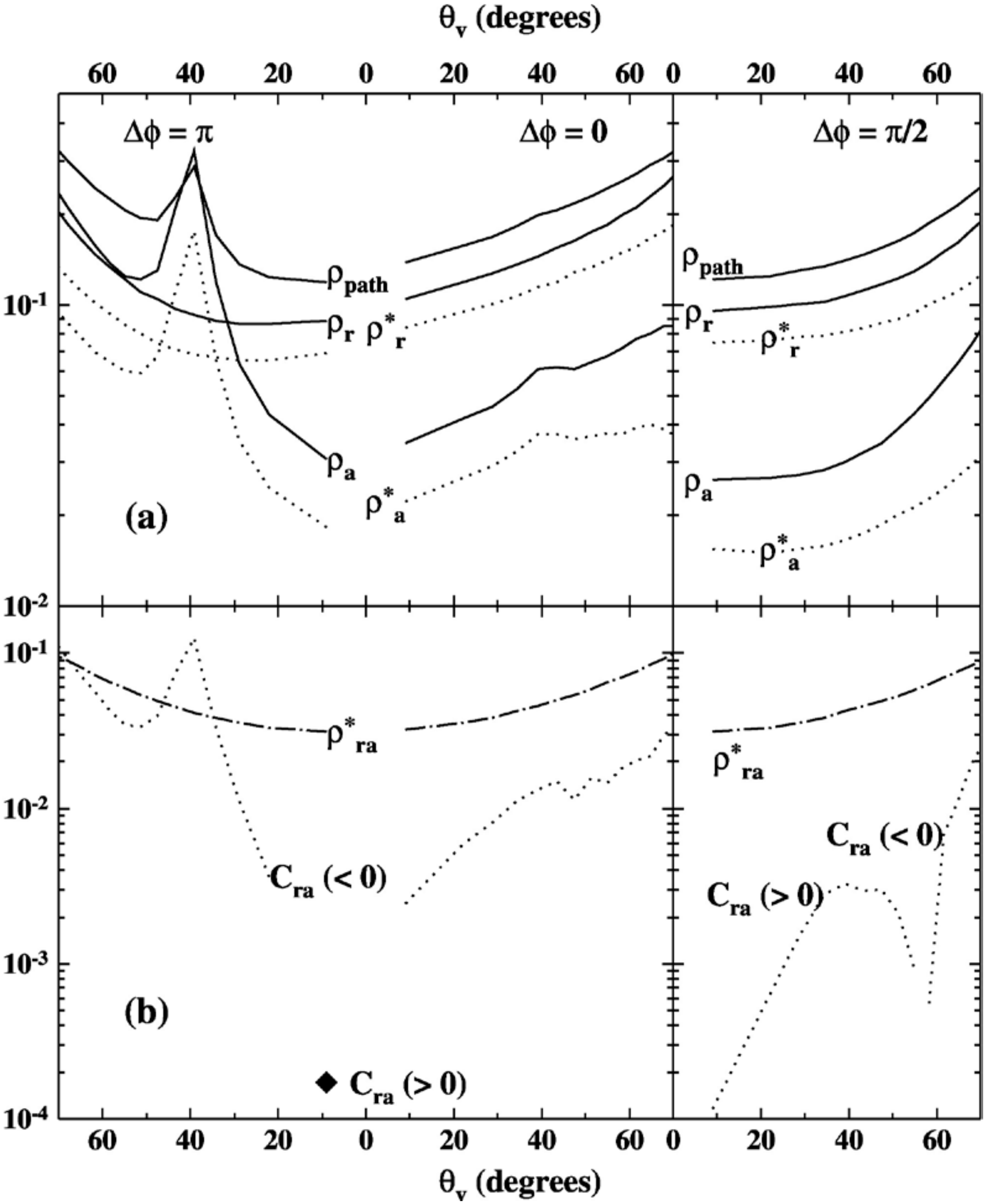
Aerosol scattering

Variations in the swath



From Figure 4 from: Antoine, D. and Morel, A. (1998). Relative importance of multiple scattering by air molecules and aerosols in forming the atmospheric path radiance in the visible and near-infrared parts of the spectrum, *Applied Optics*, **37**, 2245-2259.

Aerosols and molecules together



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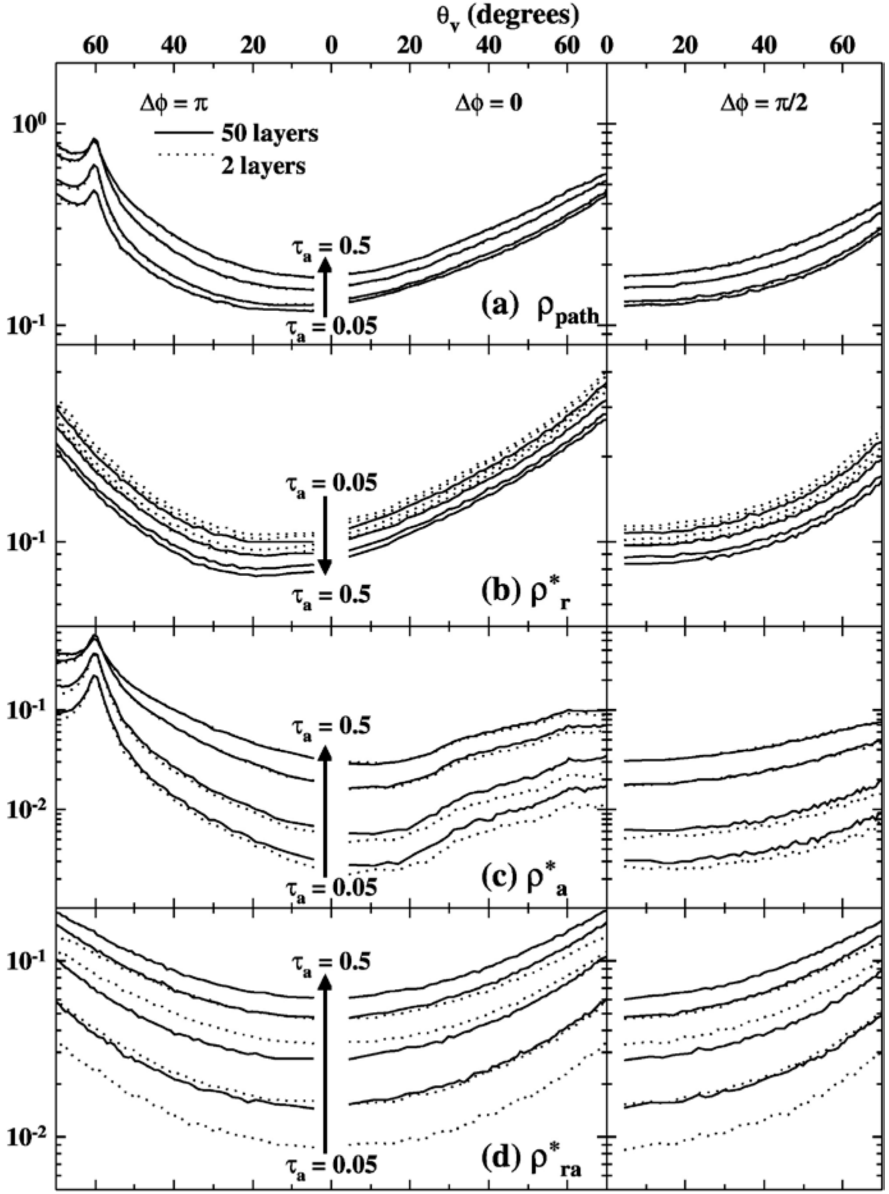

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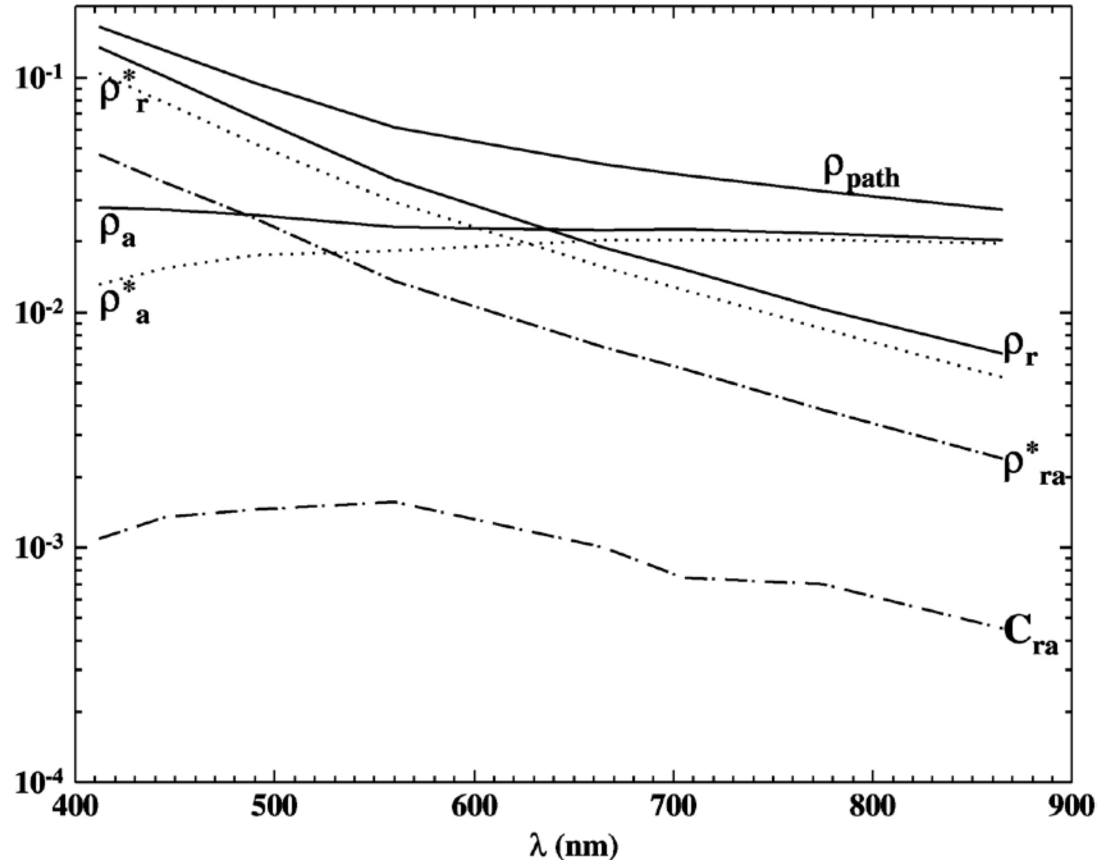
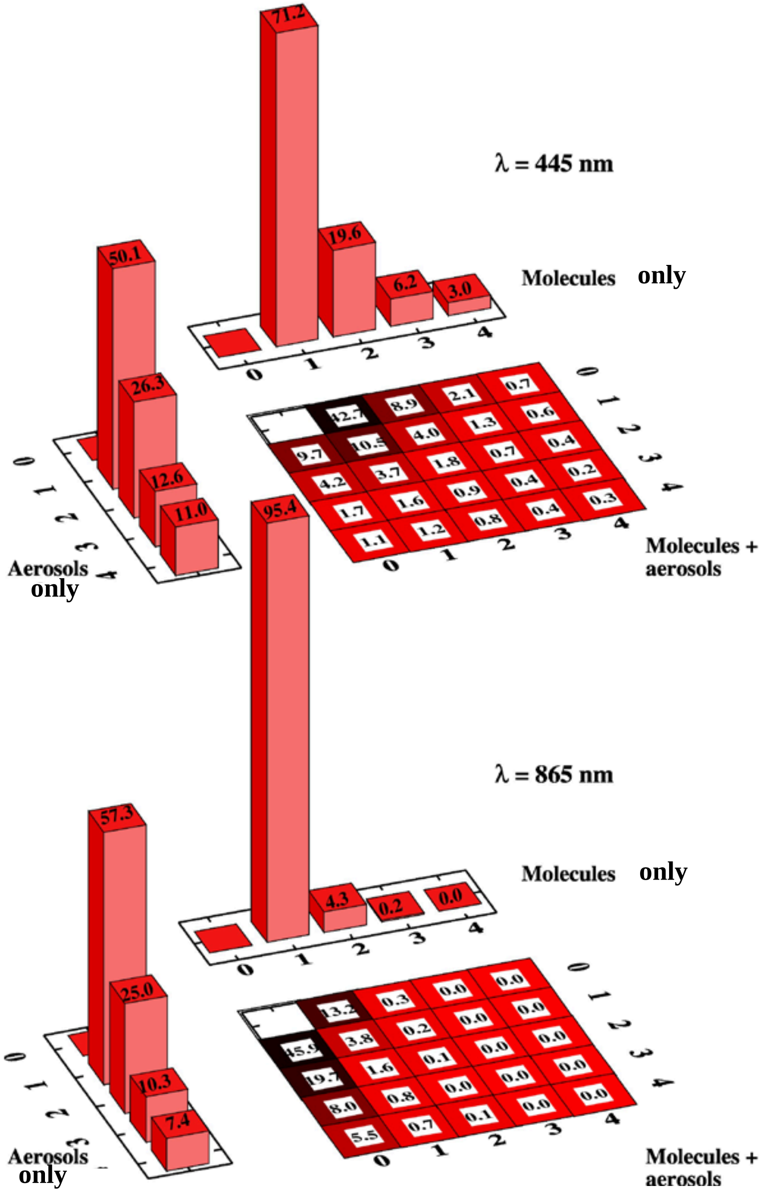
Aerosols and molecules together



Spectral changes and relative importance

Figure 3 from: Antoine, D. and Morel, A. (1998). Relative importance of multiple scattering by air molecules and aerosols in forming the atmospheric path radiance in the visible and near-infrared parts of the spectrum, *Applied Optics*, **37**, 2245-2259.

Aerosol-Rayleigh interaction scattering



$\theta_s = 40°$

$\tau_a(550) = 0.3$

Martine type aerosol

Figure 9 from: Antoine, D. and Morel, A. (1998). Relative importance of multiple scattering by air molecules and aerosols in forming the atmospheric path radiance in the visible and near-infrared parts of the spectrum, *Applied Optics*, **37**, 2245-2259.

Some basic principles of atmospheric correction

Basically: need to “play with” the spectral information and use atmospheric transmission windows to disentangle the various components that form the total signal

Fortunately: at least the Rayleigh signal can be determined from atmospheric pressure only

Unfortunately: aerosols are highly variable in the oceanic environment, and their type and concentration are *a priori* unknown when performing the atmospheric correction.

$$L_t = L_r + L_a + L_{ra} + \cancel{T.L_g} + \cancel{t.L_{wc}} + t.L_w$$

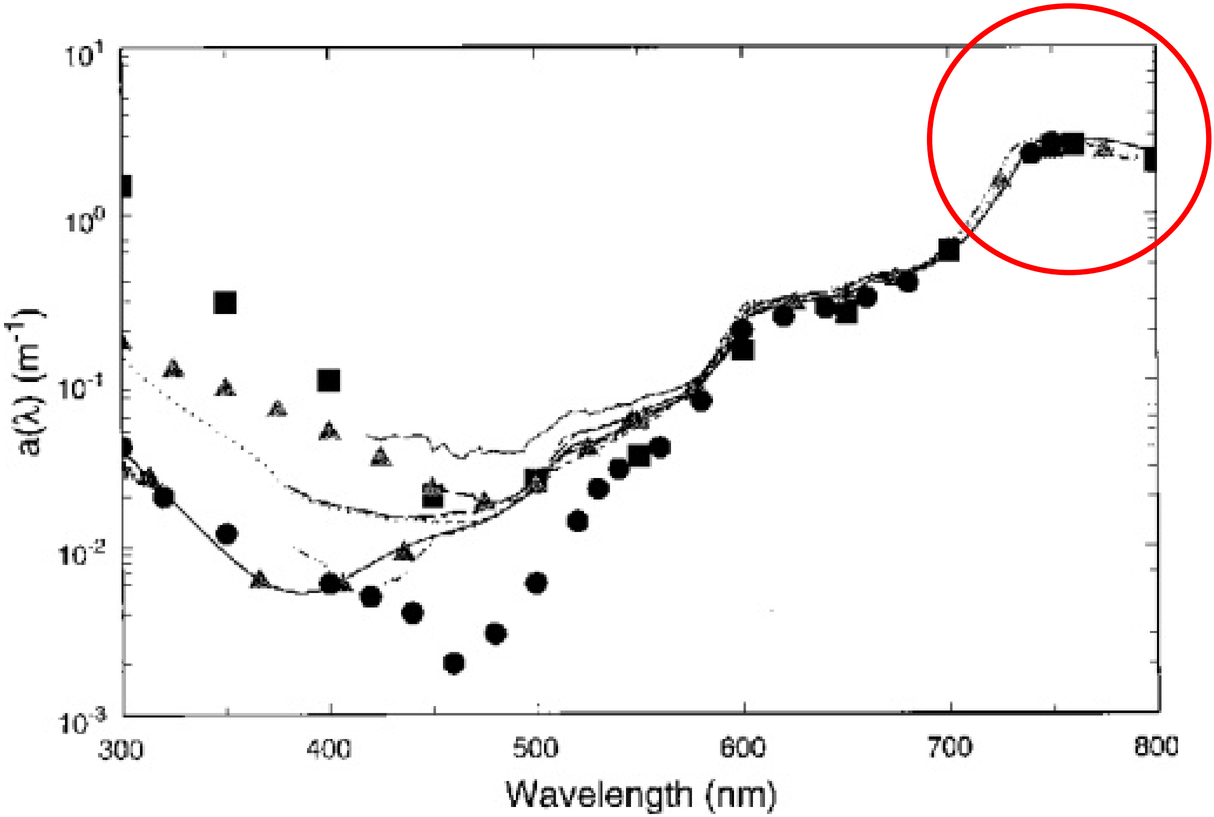
We need to eliminate some terms to get access to the others:

If we select the geometry appropriately, no sun glint: $L_g=0$

If we avoid high wind speeds: no white caps, so $L_{wc}=0$

What can we do for L_w ??

Black pixel assumption



Water absorption is high-enough that few photons can escape the medium, at least when (back) scattering is low.

This is true in Case I waters, as far as the particle content remains low (so this assumption falls out in Eutrophic areas, for instance)

From: Pegau S. et al., 1997. Applied Optics 36(24), 6035-6046 •https://doi.org/10.1364/AO.36.006035

So: L_w=0 for λ>~750nm

Some basic principles of atmospheric correction

Basically: need to “play with” the spectral information and use atmospheric transmission windows to disentangle the various components that form the total signal

Fortunately: at least the Rayleigh signal can be determined from hPa only

Unfortunately: aerosols are highly variable in the oceanic environment, and their type and concentration are *a priori* unknown when performing the atmospheric correction.

$$L_t = \underbrace{L_r}_{\text{Calculated}} + \underbrace{\boxed{L_a + L_{ra}}}_{\text{What's left}} + \cancel{T.L_g} + \cancel{t.L_{wc}} + \cancel{t.L_w}$$

We need to eliminate some terms to get access to the others:

If we select the geometry appropriately, no sun glint: $L_g=0$

If we avoid high wind speeds: no white caps, so $L_{wc}=0$

Black pixel assumption: $L_w=0$

Aerosol signal in the NIR
Ok, so what?

What we need is the aerosol signal in the visible, so that we can get L_w in the visible

We have to infer $[L_a + L_{ra}]$ in the visible from their values in the NIR

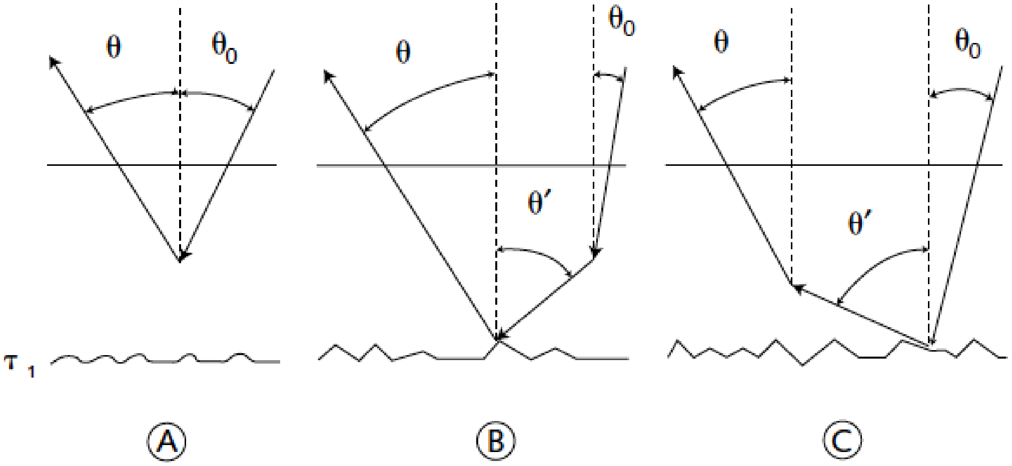
This is, basically, what atmospheric correction of satellite OCR is all about

Various techniques have been developed for that purpose

Initially: assumption of single scattering

Atmospheric correction under the assumption of single scattering

$$
\begin{aligned}
L_R(\lambda,\theta_0,\theta_v,\Delta\phi) = &\left[\frac{F_0\,\mu_0\,\omega_R}{\mu_v+\mu_0}\,p_R(\gamma^-)\left(1-e^{-\tau_R\left(\frac{1}{\mu_v}+\frac{1}{\mu_0}\right)}\right)\right] && Ⓐ\\
&+\left[\frac{F_0\,\mu_0\,\omega_R}{\mu_s-\mu'}\,p_R(\gamma^+)\,\rho_F(\theta_v)\,e^{-\frac{\tau_R}{\mu_v}}\left(e^{\frac{-\tau_R}{\mu_0}}-e^{\frac{-\tau_R}{\mu'}}\right)\right] && Ⓑ\\
&+\left[\frac{F_0\,\mu'\,\omega_R}{\mu'-\mu_v}\,p_R(\gamma^+)\,\rho_F(\theta_s)\,e^{-\frac{\tau_R}{\mu_0}}\left(e^{\frac{-\tau_R}{\mu'}}-e^{\frac{-\tau_R}{\mu_v}}\right)\right] && Ⓒ
\end{aligned}
$$



Under the assumption of small $\tau_{r,}$ the equations simplifies to: ($1 - e^{-x} \rightarrow x$ when x small)

$$L_R(\lambda,\theta_0,\theta_v,\Delta\phi) = \frac{F_0'(\lambda)\,\tau_R(\lambda)\,\bar{\omega}_R(\lambda)\,p_R(\lambda,\gamma\pm)}{4\pi\,\cos(\theta_v)}$$

Where:

$$F_0'(\lambda) = F_0(\lambda)\,\exp\left[-\tau_{O3}(\lambda)\left(\frac{1}{\cos(\theta_v)}+\frac{1}{\cos(\theta_0)}\right)\right]$$

And:

$$p_R(\lambda,\gamma\pm) = p_R(\gamma\pm,\lambda) + [\rho_F(\theta_v + \rho_F(\theta_0)]\,p_R(\gamma\pm,\lambda)$$

Exact same equations can be derived for the aerosol signal, with the aerosol phase function (P_a), aerosol optical thickness (τ_a) etc..

Wang M., 1994. Atmospheric correction of the second generation ocean color sensors, dissertation presented in fulfillment of the requirement for degree of doctor of philosophy (PhD), University of Miami, Coral Gables, Florida, USA.

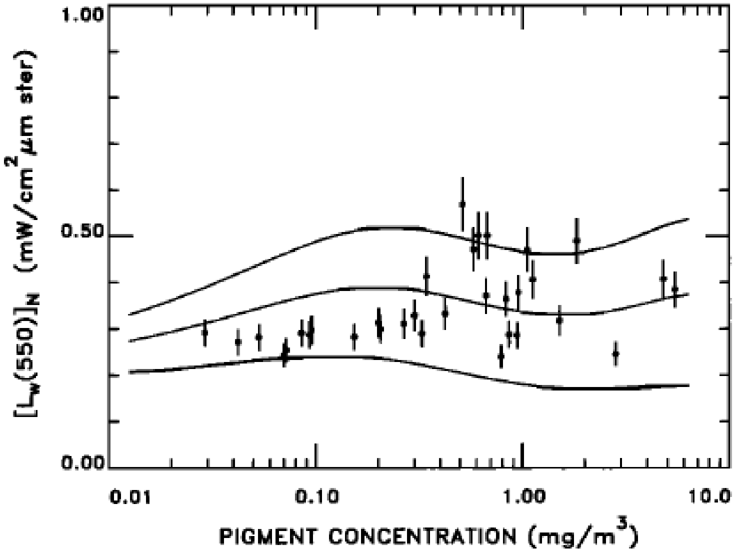
Atmospheric correction under the assumption of single scattering

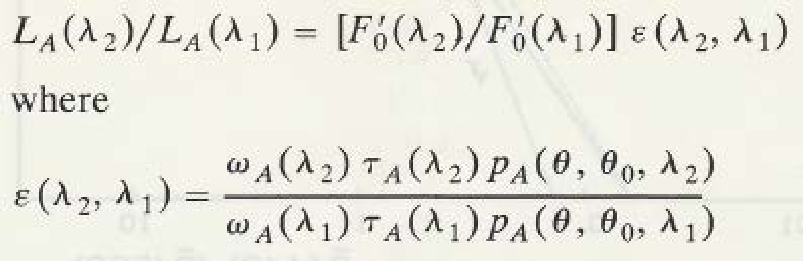
By the way, what if you do not have NIR band? (CZCS)

“Clear Water Radiance Concept” (CWRC)

Gordon HR, 1978. Applied Optics, 17, 1631-1636; Gordon HR and DK Clark, 1981, Applied Optics, 20, 4175-4180

Basically considers $L_w(550)$ as a constant over clear waters, and $L_w(670)$=0





$$L_A(\lambda_2)/L_A(\lambda_1) = [F_0'(\lambda_2)/F_0'(\lambda_1)]\,\varepsilon(\lambda_2, \lambda_1)$$

where

$$\varepsilon(\lambda_2, \lambda_1) = \frac{\omega_A(\lambda_2)\,\tau_A(\lambda_2)\,p_A(\theta, \theta_0, \lambda_2)}{\omega_A(\lambda_1)\,\tau_A(\lambda_1)\,p_A(\theta, \theta_0, \lambda_1)}$$

So:

- Locate clear waters in the image
- Determine ε on the corresponding pixels
- Average them and apply to the entire image

Fig. 3 in Gordon HR et al., 1988, J. Geophys. Res., 93(D9), 10909-10924

Atmospheric correction under the assumption of single scattering

Extension of the “Clear Water Radiance Concept” (CWRC)

Still without NIR bands

Instead of considering L_w as a constant at 550nm and zero at 670nm, and being limited to clear waters, L_w can be modelled.

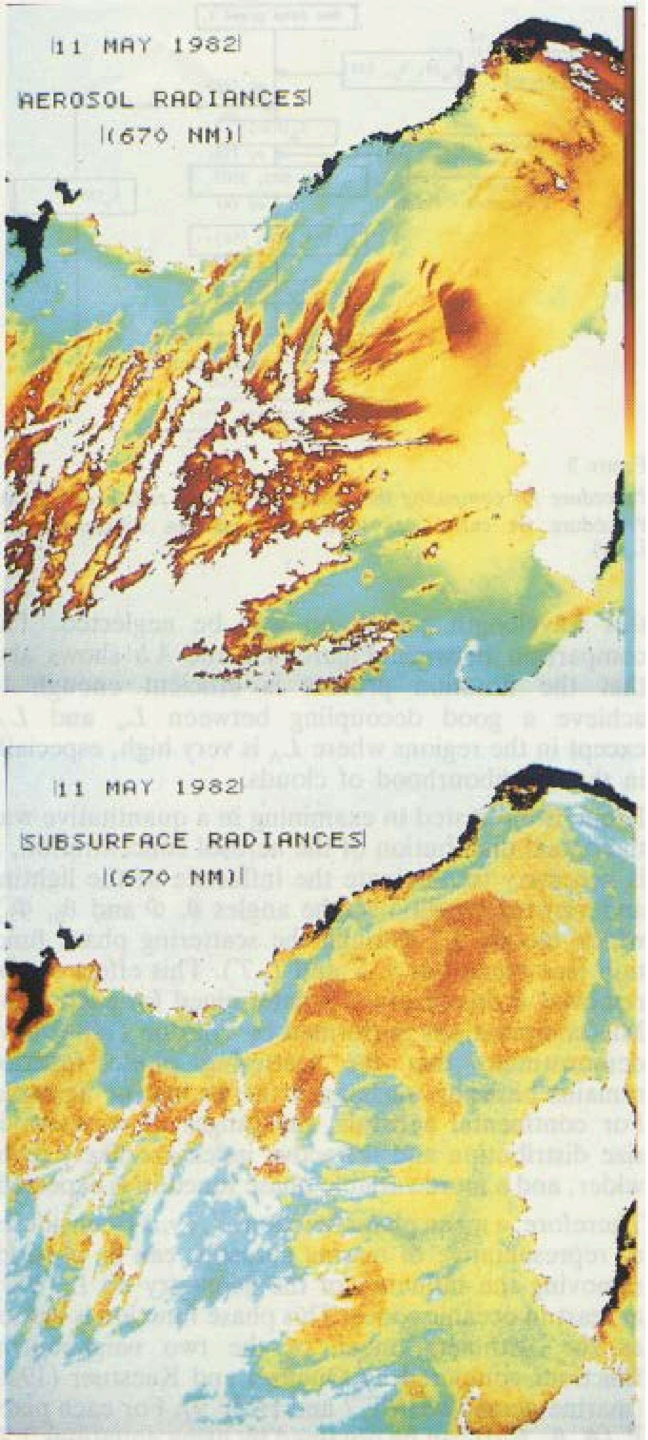
Bricaud, A., and A. Morel (1987). Atmospheric corrections and interpretation of marine radiances in CZCS imagery: use of a reflectance model. Oceanologica Acta, N°SP, 33-50.

André, J.M. and A. Morel (1991). Atmospheric corrections and interpretation of marine radiances in CZCS imagery, revisited, Oceanologica Acta, 14, 3-22.

The underlying model (Morel 1988, Journal of Geophysical Research, 93, 10,749-10,768) is a Case I water reflectance model that needs chlorophyll as input.

Therefore, you need to enter in an iterative system, because you do not know Chl when you start

Atmospheric correction under the assumption of single scattering



It works!

Figure 4 in: Bricaud, A., and A. Morel (1987). Atmospheric corrections and interpretation of marine radiances in CZCS imagery: use of a reflectance model. Oceanologica Acta, N°SP, 33-50.

Atmospheric correction under the assumption of single scattering

What if you now have two NIR bands? (SeaWiFS) λ_{IR1} and λ_{IR2}

No longer need to make an assumption that $L_w(550)$ is constant or that it can be modelled as a function of chlorophyll

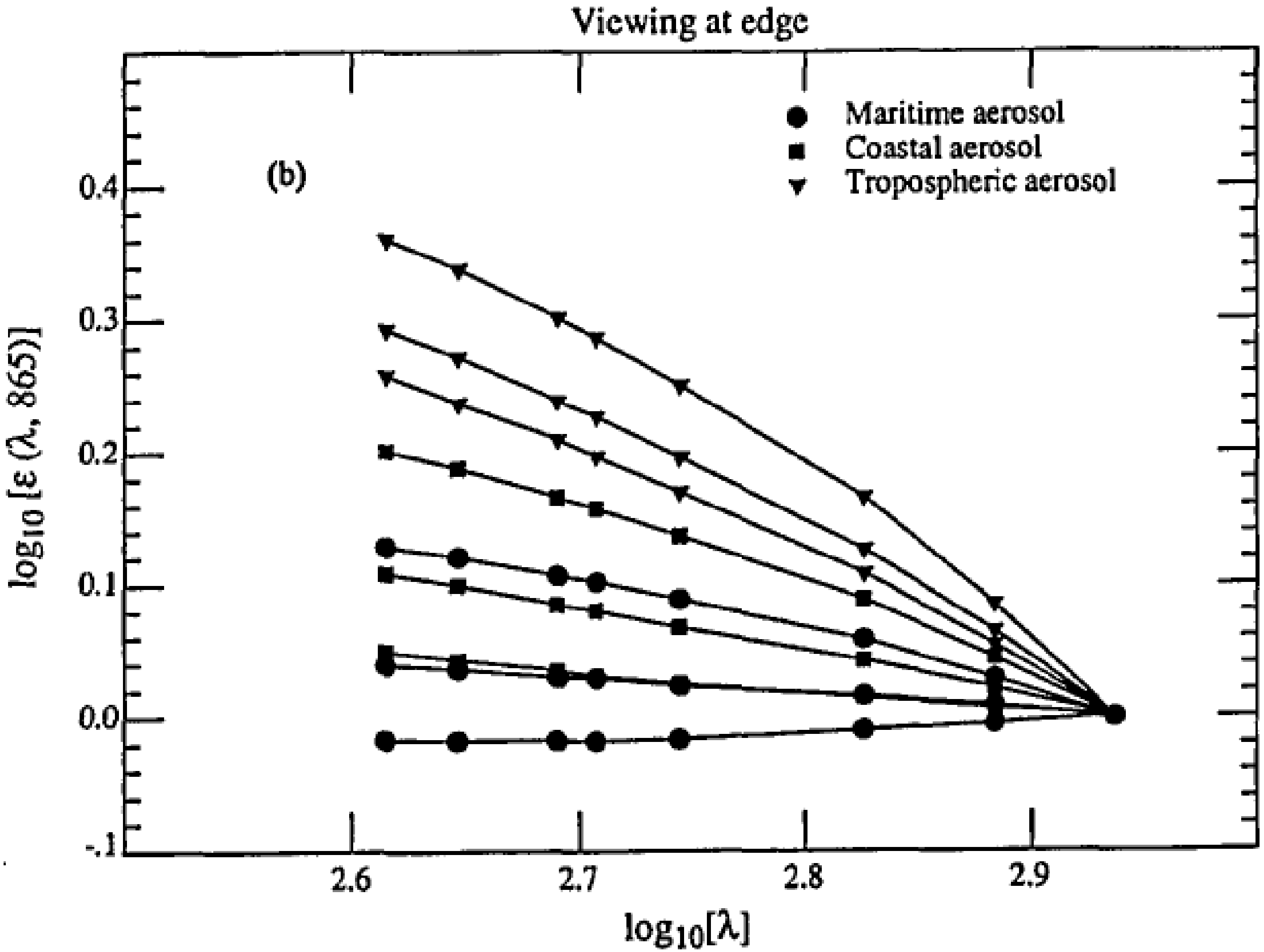
We can directly determine $\varepsilon(\lambda_{IR1}, \lambda_{IR2})$ from the two NIR bands

Then, we need a law to determine $\varepsilon(\lambda_{VIS}, \lambda_{IR2})$:

$$\varepsilon(\lambda_{VIS}, \lambda_{IR2}) = \left(\frac{\lambda_{IR2}}{\lambda_{VIS}}\right)^{n}$$

Assumes linear relationship between $\log(\varepsilon)$ and $\log(\lambda)$ of slope $-n$

The relationship between log(ε) and log(λ) is actually not linear



From: Gordon HR and M Wang, 1994, Applied Optics, 33(3), 443-452

We need another way of extrapolating the aerosol contribution from the NIR to the VIS

- The single scattering approximation does not allow reaching the accuracy we need
- That is where the use of aerosol models for atmospheric correction of satellite OCR observations has been introduced, in the mid 1990s
- What an “aerosol model” is?
- A particle size distribution (PSD) + an index of refraction (n, complex)
- From these, the MIE theory can be used to calculate optical properties, such as the VSF, the scattering and absorption coefficients
- Then, these optical properties can be used as inputs for RT calculations, from which the spectral dependence of the aerosol signal can be studied

Aerosol size distributions

The more smaller particles are present, the more “selective” scattering will be

Purely maritime type

Urban / industrial type

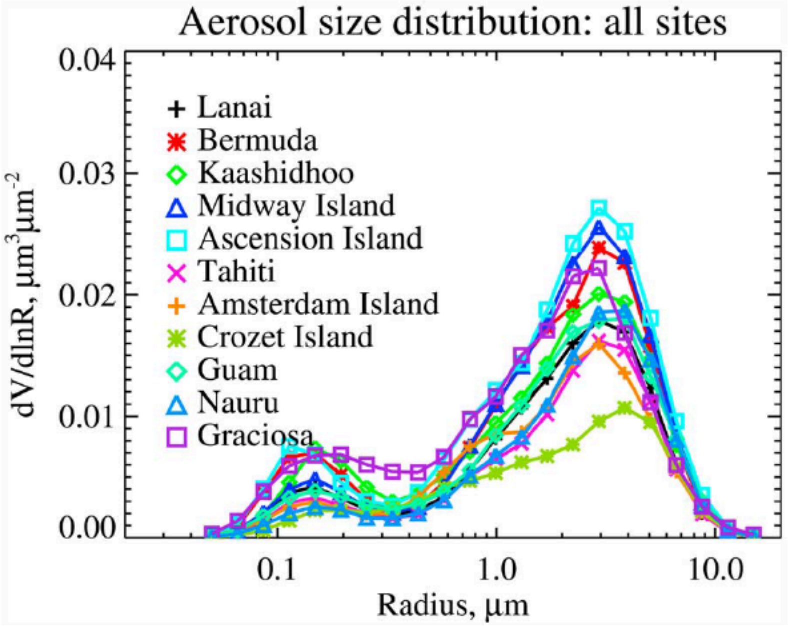


Figure 3 in: Sayer et al., 2012. J. Geophys. Res., VOL. 117, D05213, doi:10.1029/2011JD016689, 2012

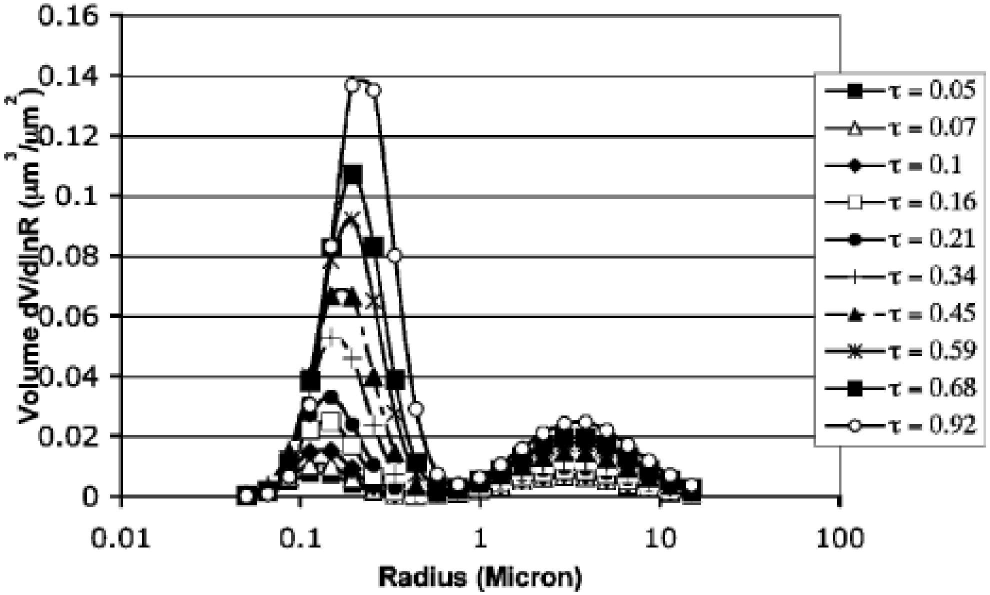


FIG. 10. Volume size distribution for the dynamic urban/industrial aerosol model of Dubovik et al. (2002). Curves are functions of optical depth at 0.44 μm (τ).

Figure 10 in: Levy et al., 2005. Journal of the Atmospheric Sciences—special Section, vol 62, 974-992

Aerosol refractive indices

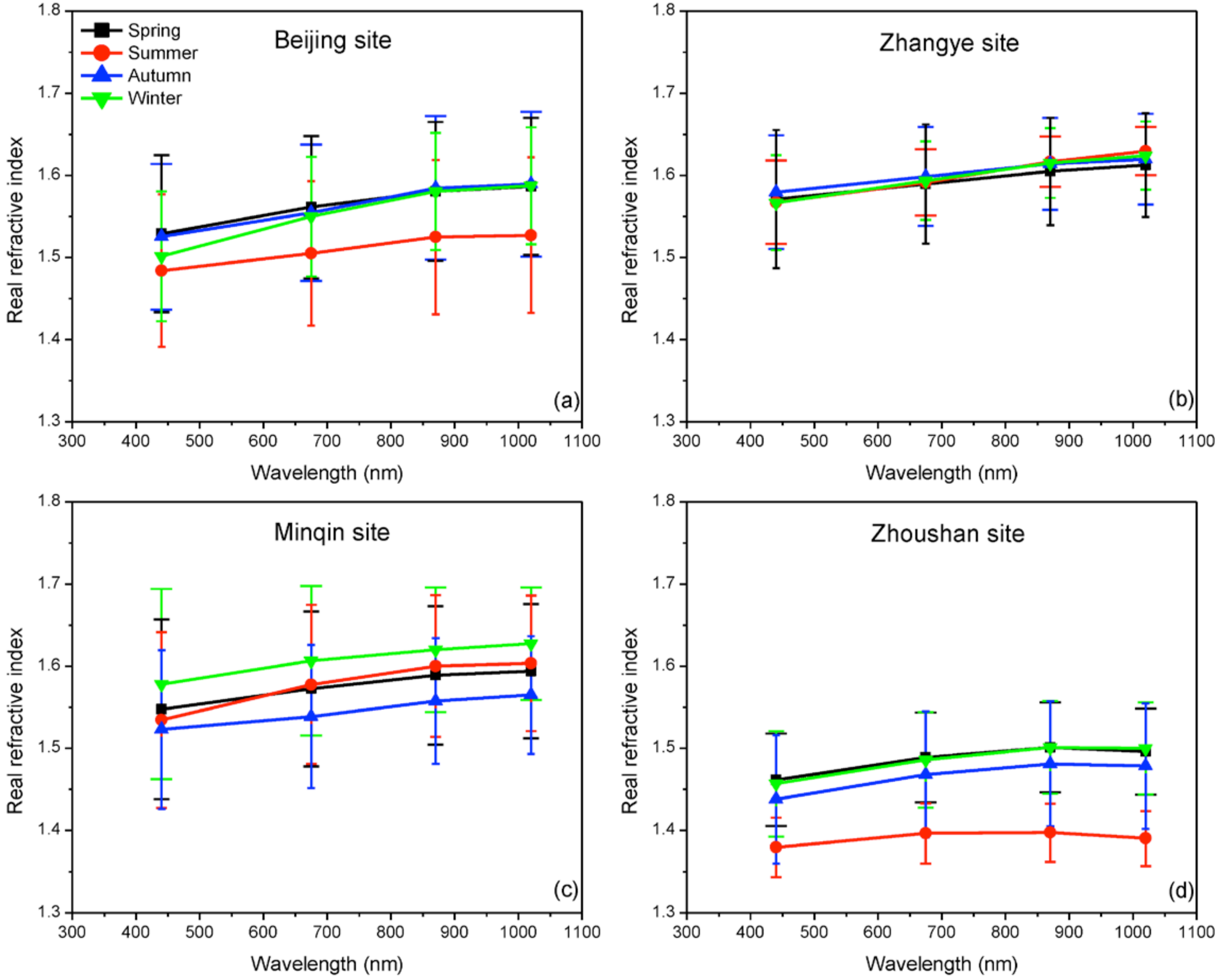


Figure 5 in: Xie et al., 2015. Remote Sens. 2015, 7, 9928-9953; doi:10.3390/rs70809928

Aerosol "single scattering albedo" b/c

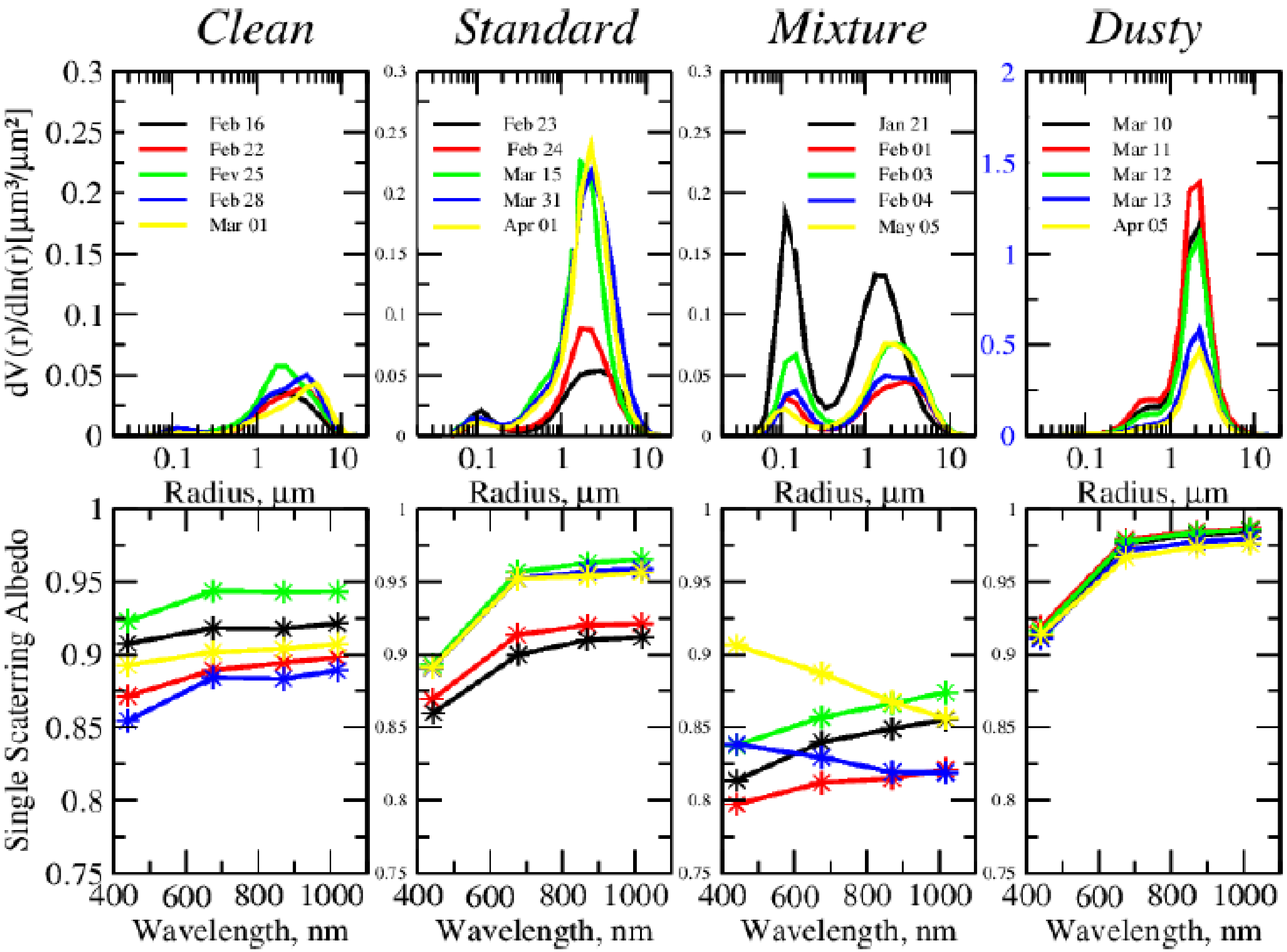
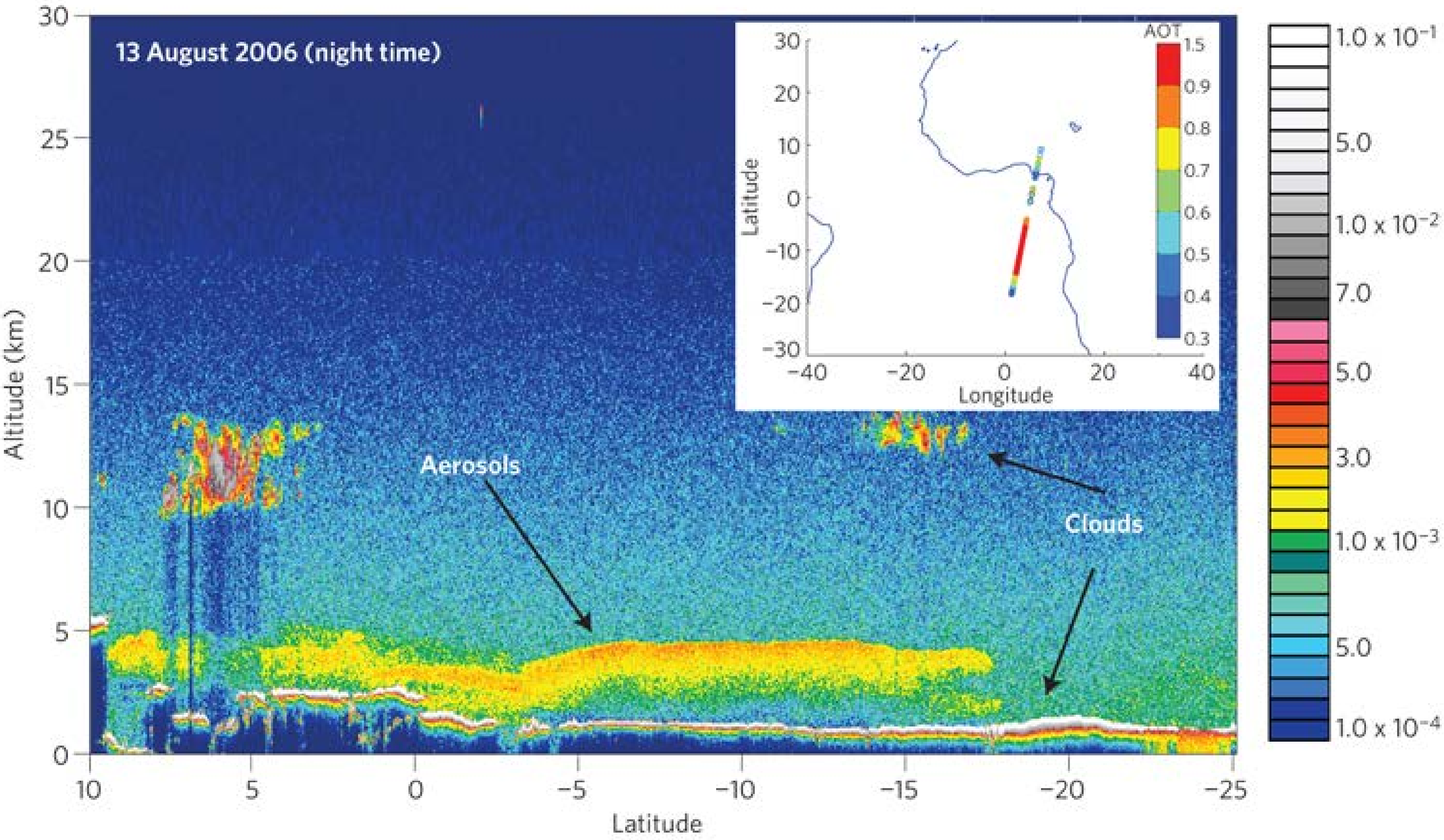


Figure 2 in: Simina Drame et al., 2015. Atmosphere 2015, 6, 1608-1632; doi:10.3390/atmos6111608

Where aerosols are, however?



Vertical structure in principle needed for running RT calculations

From: https://www.nasa.gov/topics/earth/features/calipso-aerosol.html

We have to specify a vertical structure

We do not know it *a priori*. Fortunately, when aerosols do not absorb light, their repartition with altitude (vertical profile) does not matter: the impact on the calculated top-of-atmosphere path radiance is minimal

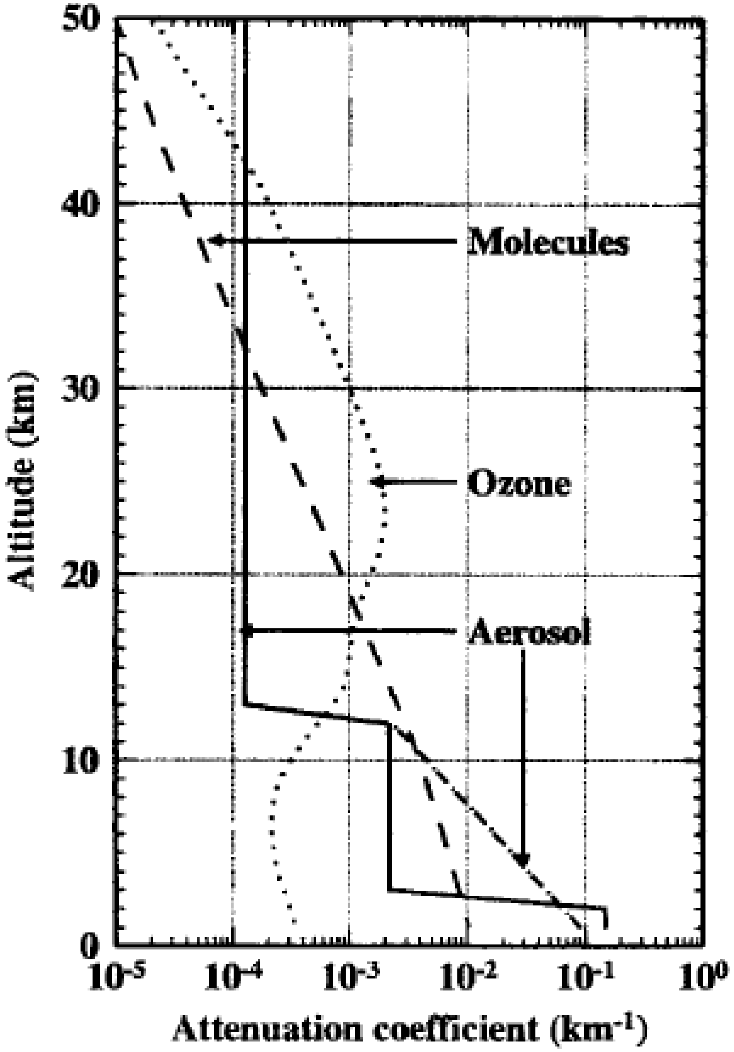
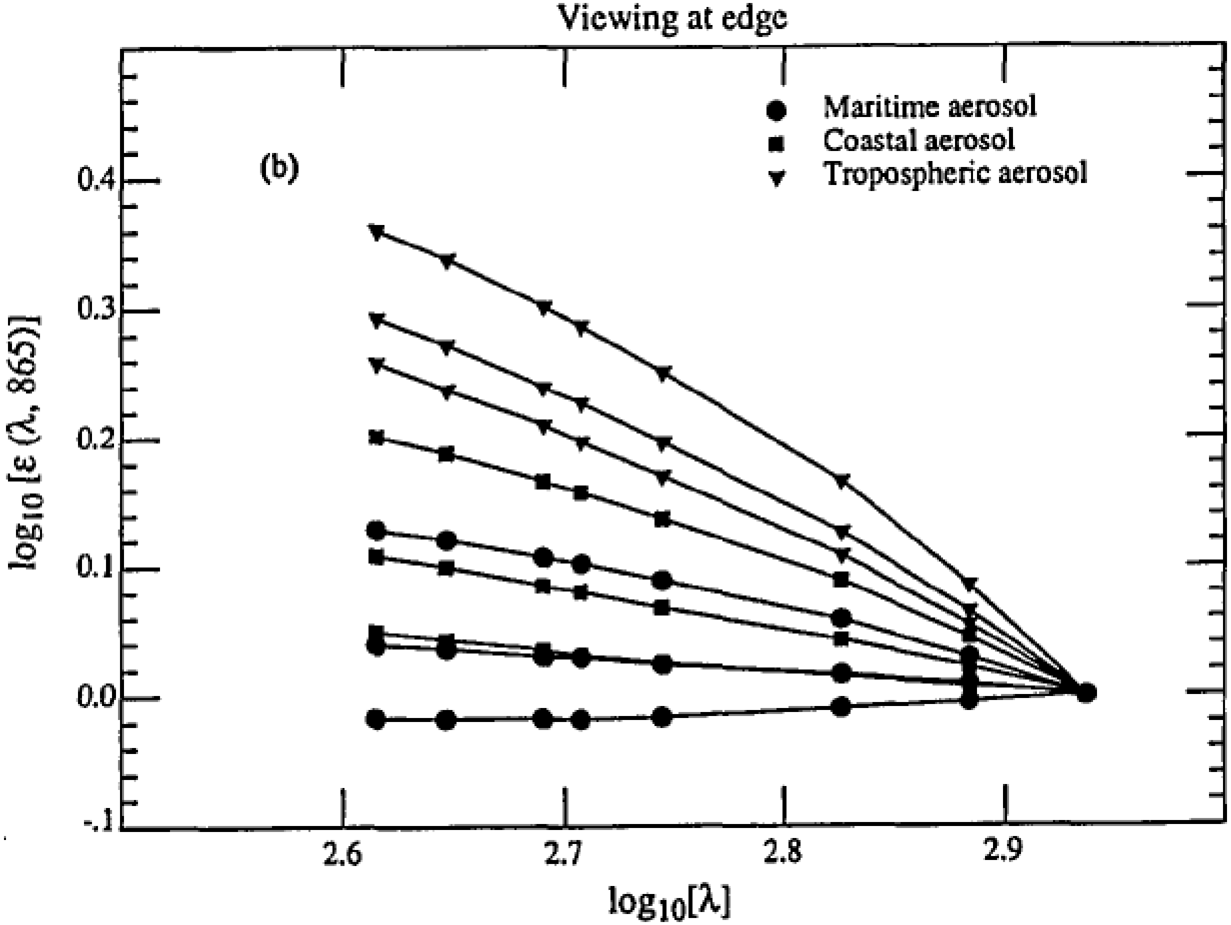


Fig. 4 in Antoine and Morel, 1999. A multiple scattering algorithm for atmospheric correction of remotely-sensed ocean colour (MERIS instrument) : principle and implementation for atmospheres carrying various aerosols including absorbing ones, International Journal of Remote Sensing, 20, 1875-1916.

Ok, at this point:

- We consider situations where $L_G=L_{wc}=0$
- Then we use the black pixel assumption, so that $L_w=0$ in the NIR
- We therefore have determined L_A+L_{RA} at NIR bands
- We also know how to compute L_R in a multiple scattering regime for any spectral band and geometry and wind speed
- We also know how to compute L_A in a multiple scattering regime for any spectral band and geometry and wind speed
- We still do not know how to use this to determine L_A+L_{RA} at visible bands

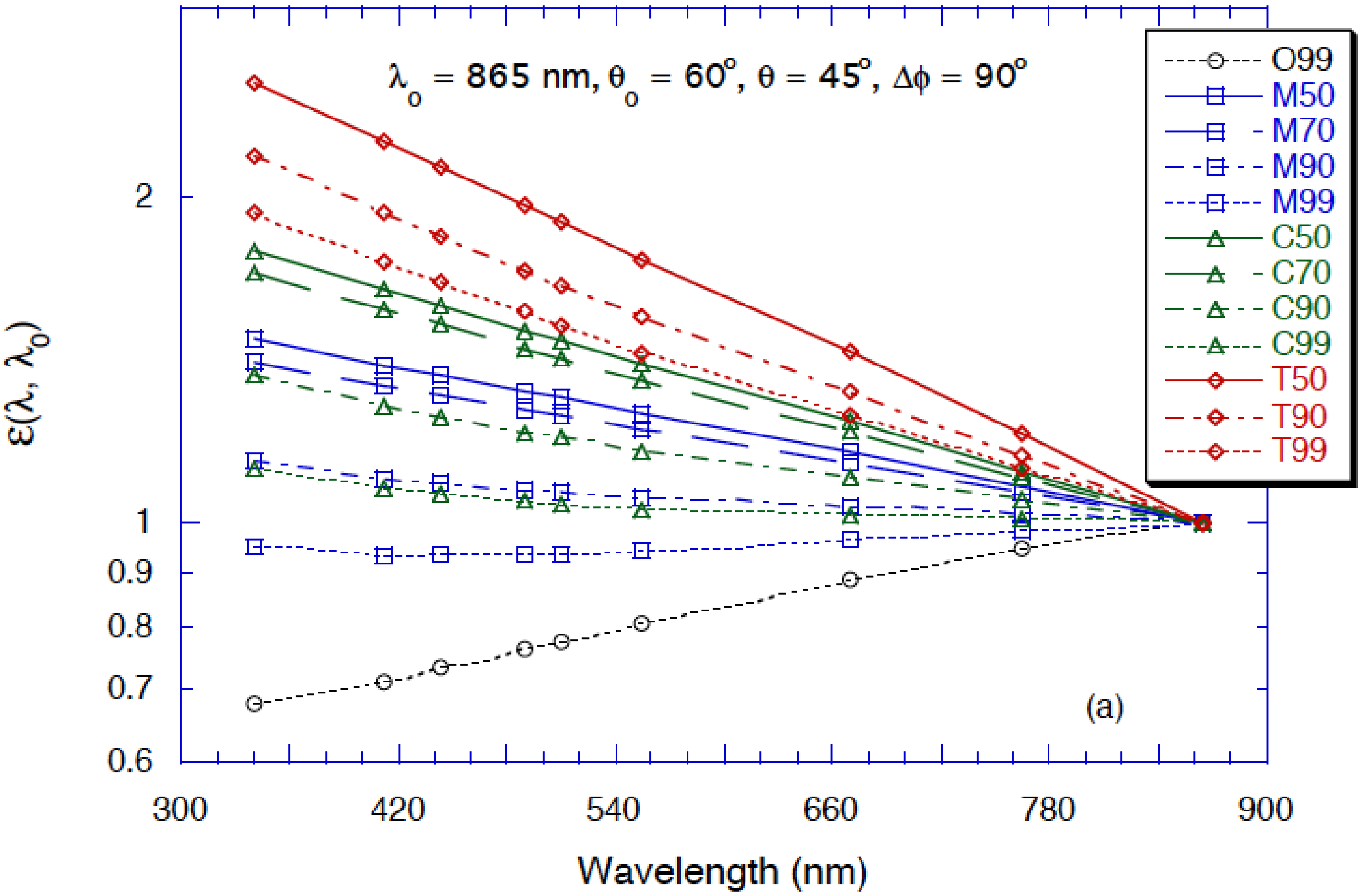
Modeling $\varepsilon(\lambda_1,\lambda_2)$



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

From: Gordon HR and M Wang, 1994, Applied Optics, 33(3), 443-452

Inferring $\varepsilon(\lambda_1,\lambda_2)$ using RT results (LUTs)



From Menghua Wang, IOCCG summer lectures 2016

The SeaWiFS/MODIS algorithm Gordon & Wang

- Starts by calculating ρ_t - ρ_r at all bands
- Then Determine $\varepsilon(\lambda_1,\lambda_2)$ from 2 NIR bands
- Navigate the LUTs and find the 2 closest values to the NIR ε
- They determine two “bracketing aerosol models”
- Calculate a “mixing ratio” from this, and assumes it is constant with wavelength
- Then we can use these LUTs to find $\varepsilon(\lambda_{vis}$, 865). There will be one value for each of the “bracketing models”, and the mixing ratio will allow interpolation between these two values
- Assumes the curves do not cross !!

Gordon, HR, 1997. J. Geophys. Res., 102(D14), 17081-17106

The MERIS algorithm

Starts from ρ_t/ρ_r instead of $\rho_t - \rho_r$

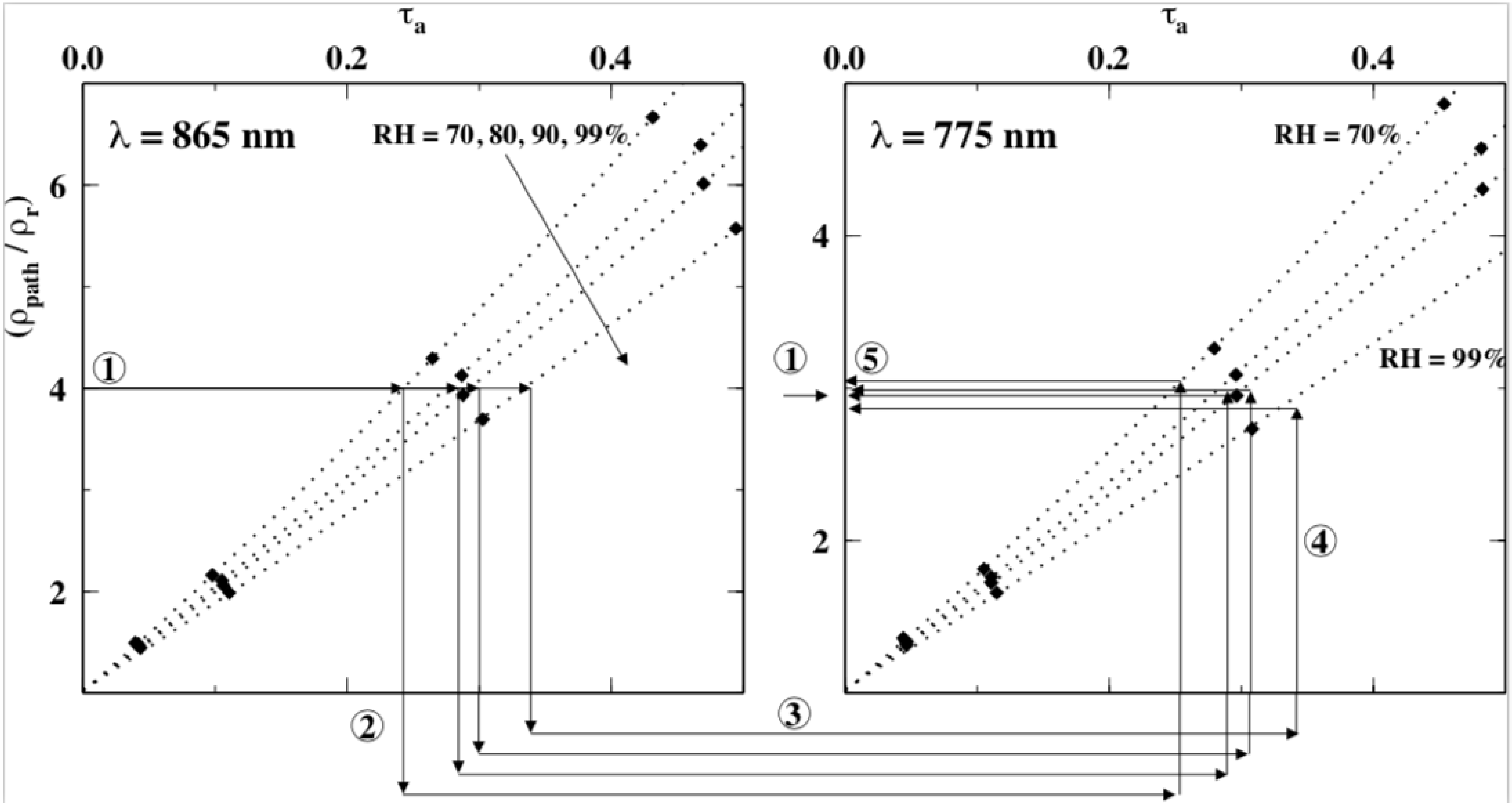
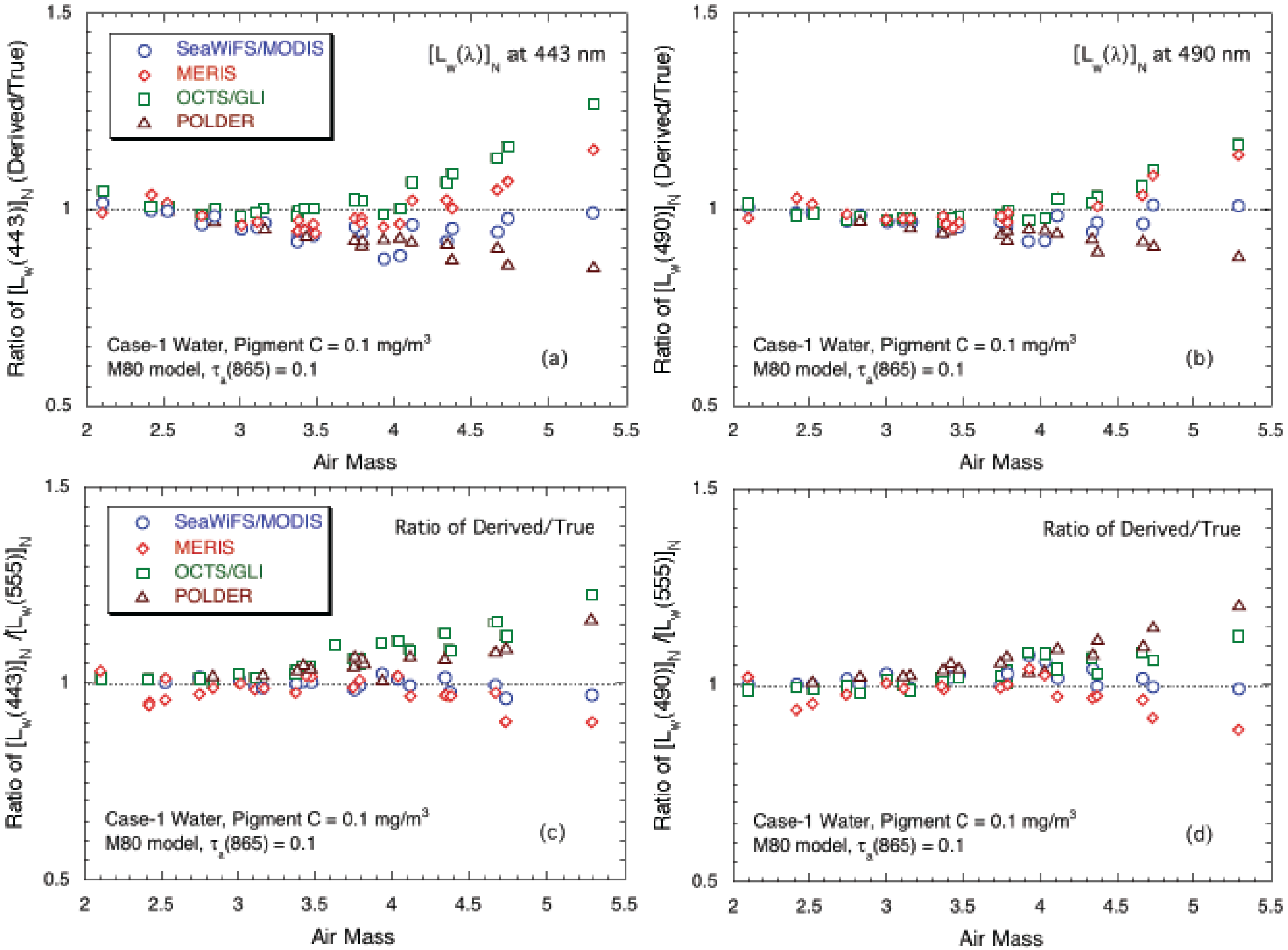


Fig. 3 in Antoine and Morel, 1999, International Journal of Remote Sensing, 20, 1875-1916.

Overall performance



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

Sun glint

Basically occurs when the viewing angle is within a close range the sun zenith angle, and the azimuth difference is around 180°

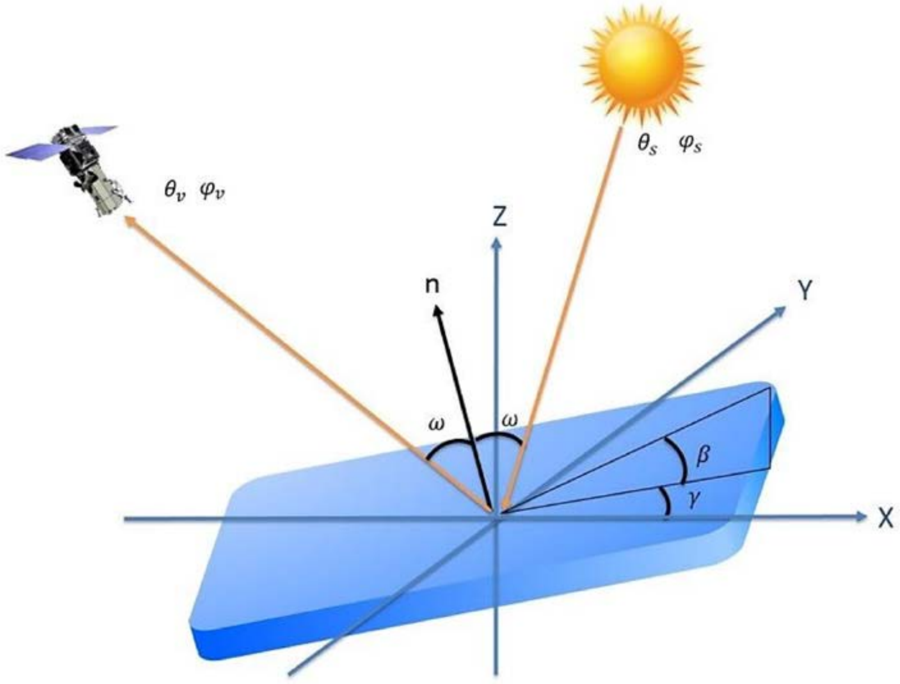


Figure 5 in Martin et al., 2016. Remote Sens. 2016, 8, 37; doi:10.3390/rs8010037

If the surface is perfectly flat: image of the sun; probably blurred by atmospheric transmission

If it is not: have to describe the effect of light being redirected in a range of directions, as a function of the sea state

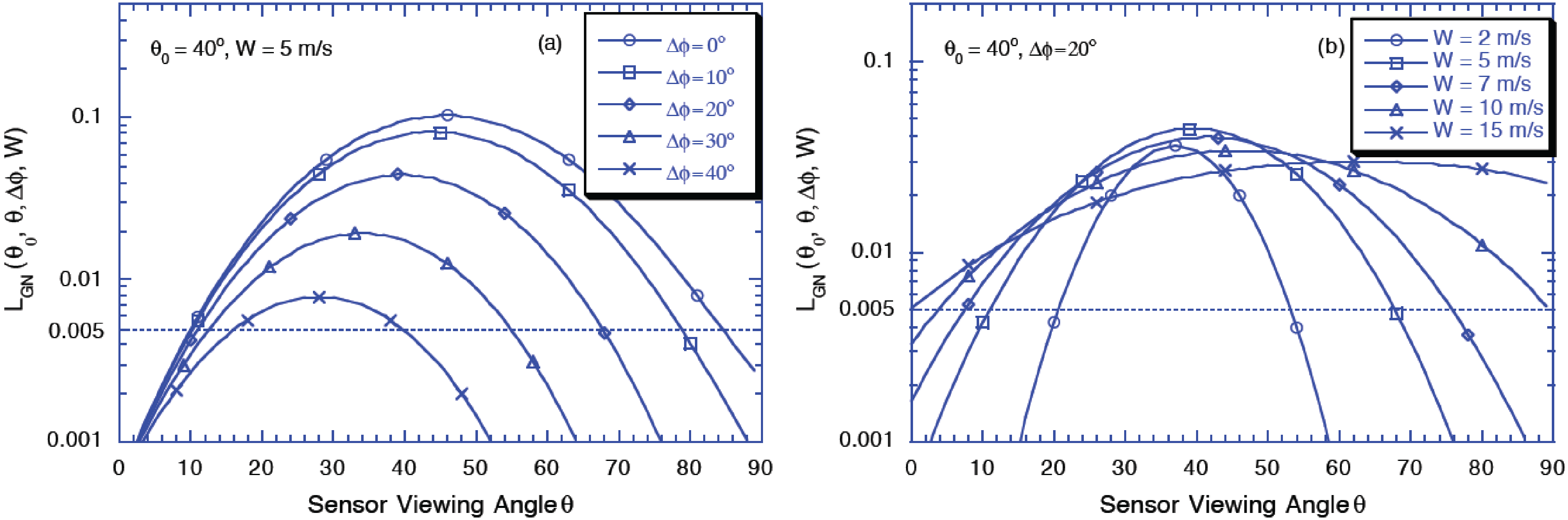
The Cox and Munk, 1954 model (J. Opt. Soc. Am. 44(11), 838-850) is still in use and has proven to work well

$$\sigma^2 = 0.003 + 5.12\ 10^{-3}\ W$$

$$\rho_G = \pi\,\rho_F\ p(\theta_s, \theta_v, \Delta\phi) \,/\, (4 \cos(\theta_v) \cos(\theta_s) \cos^4(\beta)) T(\theta_v)\ T(\theta_s)$$

$$p(\theta_s, \theta_v, \Delta\phi) = \frac{1}{\pi\,\sigma^2} \exp\left(\frac{-\tan^2(\beta)}{\sigma^2} \right)$$

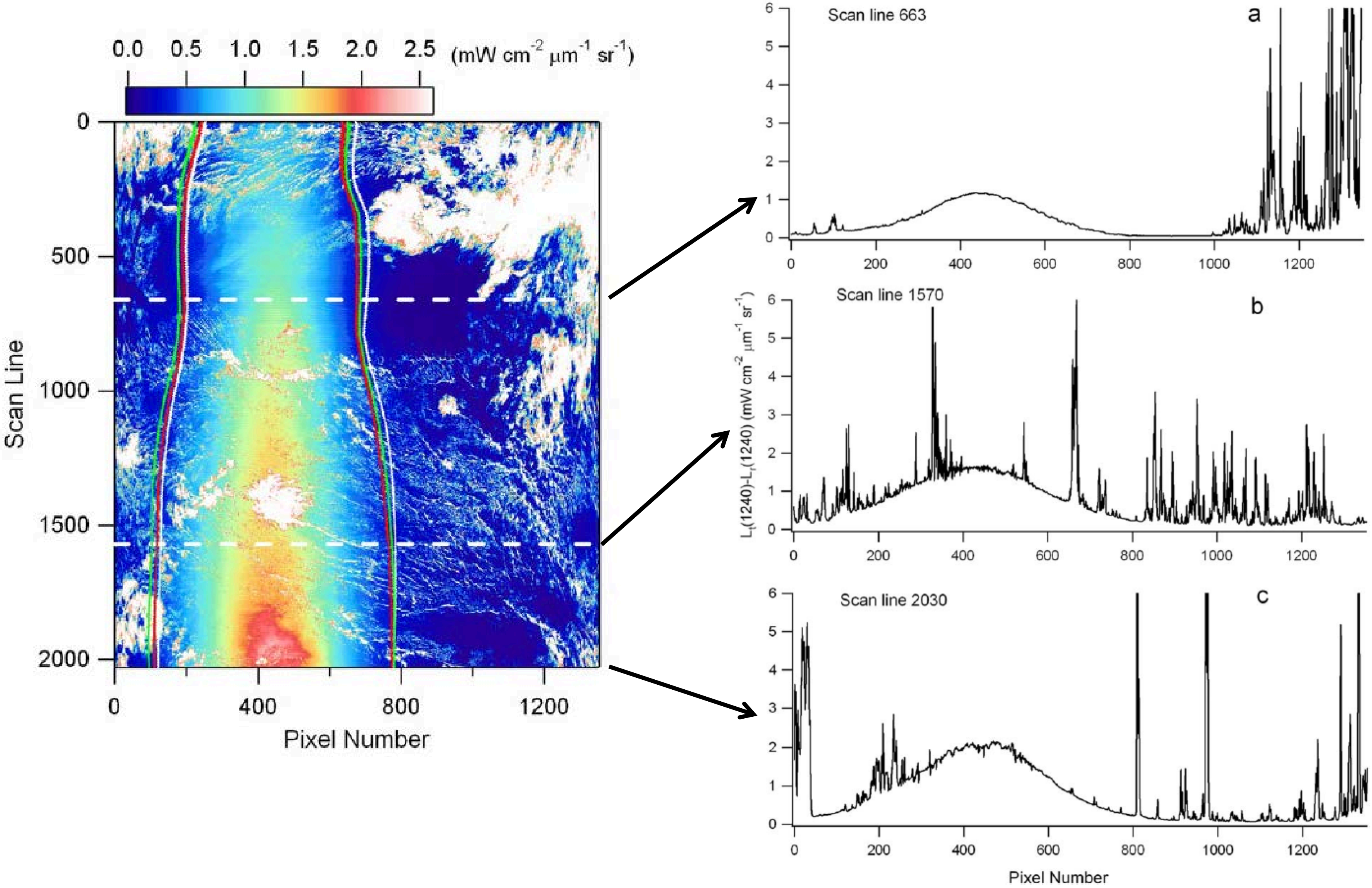
Sun glint



Wang M and SW Bailey, 2001. Correction of sun glint contamination on the SeaWiFS ocean and atmosphere products, Appl. Opt. **40**, 4790-4798

Theoretical calculations, as a function of the wind speed and geometry of observation

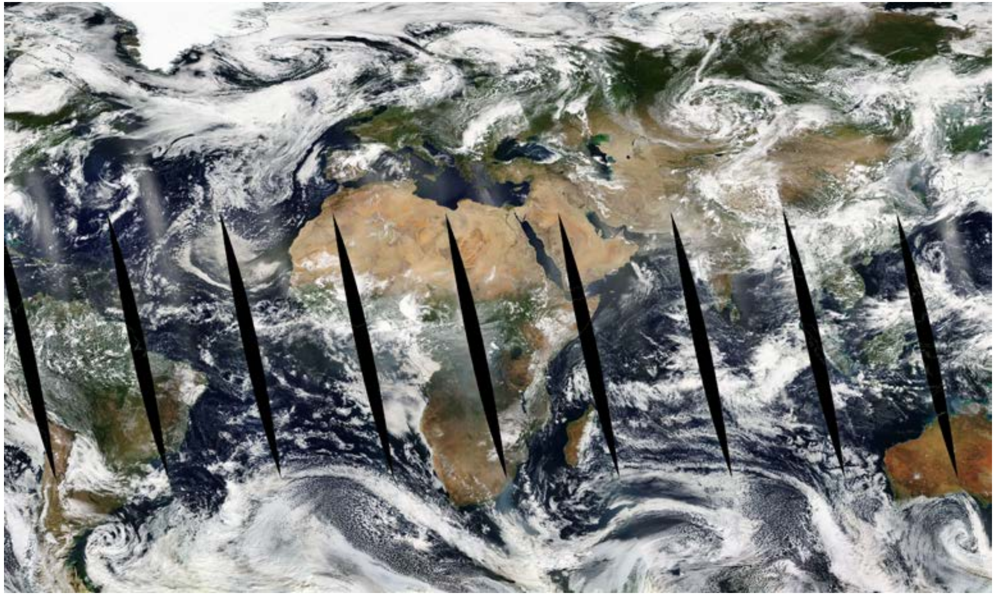
Sun glint

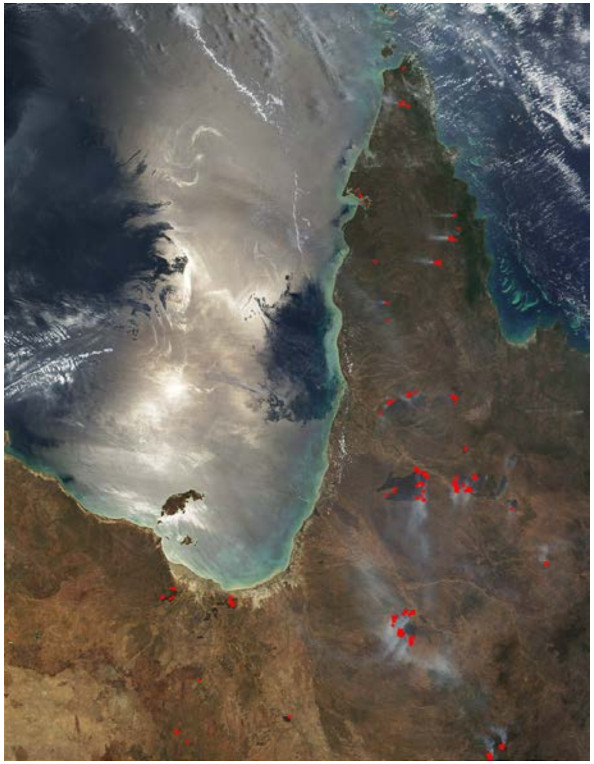


Figs 2 and 3 from Zhang and Wang, 2010, Journal of Quantitative Spectroscopy & Radiative Transfer, 111, 492–506

Sun glint

https://earthobservatory.nasa.gov/IOTD/view.php?id=84333

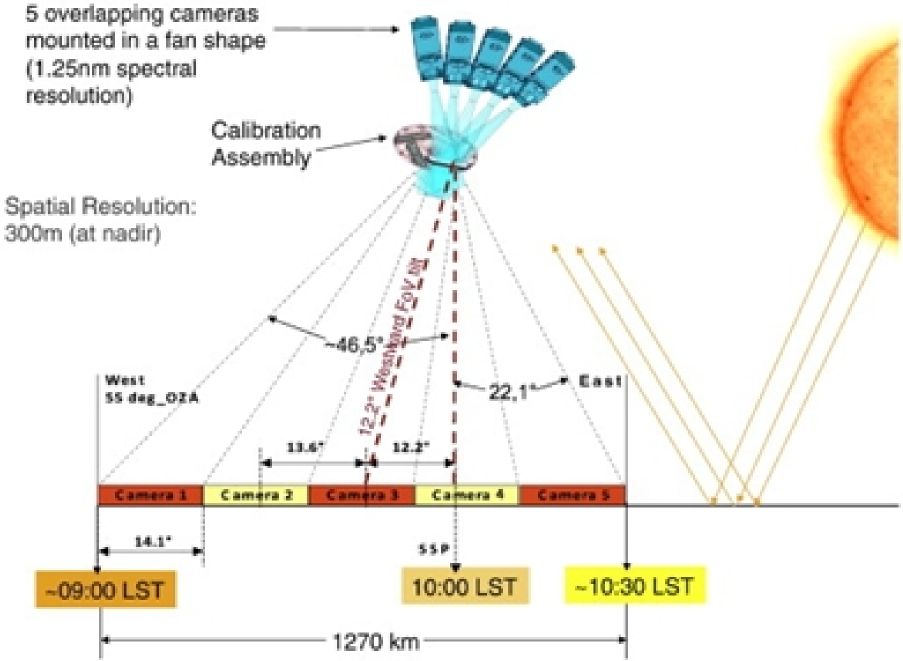




https://earthobservatory.nasa.gov/IOTD/view.php?id=90532

https://visibleearth.nasa.gov/view.php?id=57109

Minimizing sun glint impact: tilt capability



Sentinel3/OLCI: tilting across track

From: https://sentinel.esa.int/web/sentinel/user-guides/sentinel-3-olci/coverage

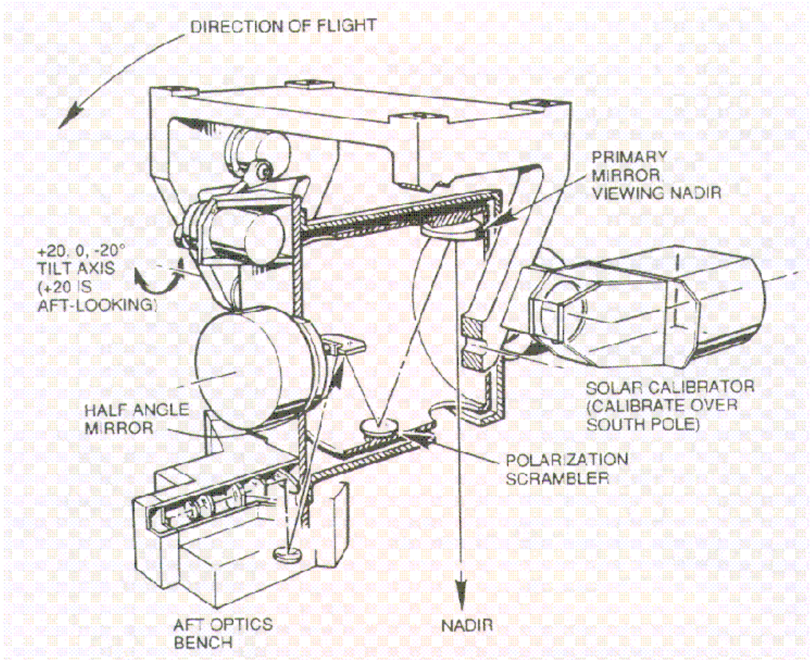


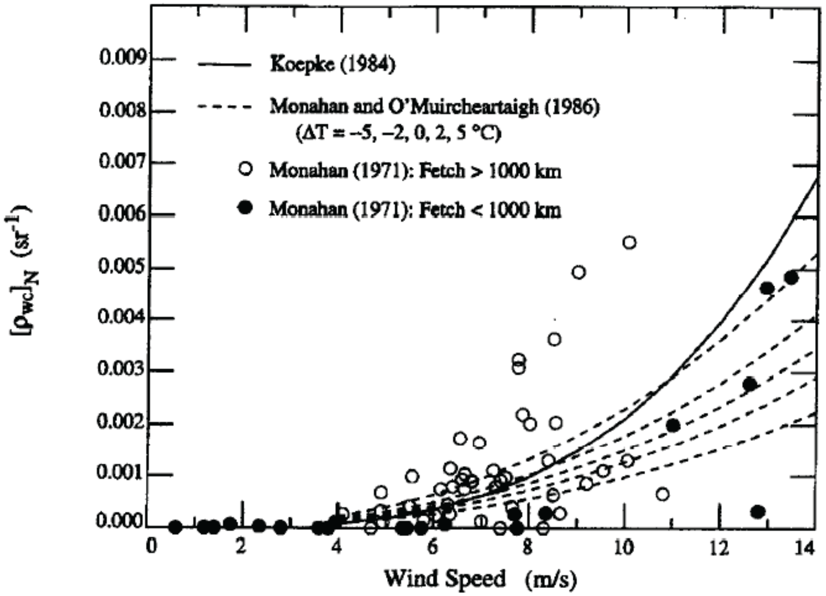
Fig. 7. Schematic of the SeaWiFS scanner assembly which scans from west to east.

SeaSTAR/SeaWiFS : tilting along track (+/-20°)

From: https://oceancolor.gsfc.nasa.gov/SeaWiFS/SEASTAR/SPACECRAFT.html

White caps / foam

Coverage if function of wind speed (also temperature actually)

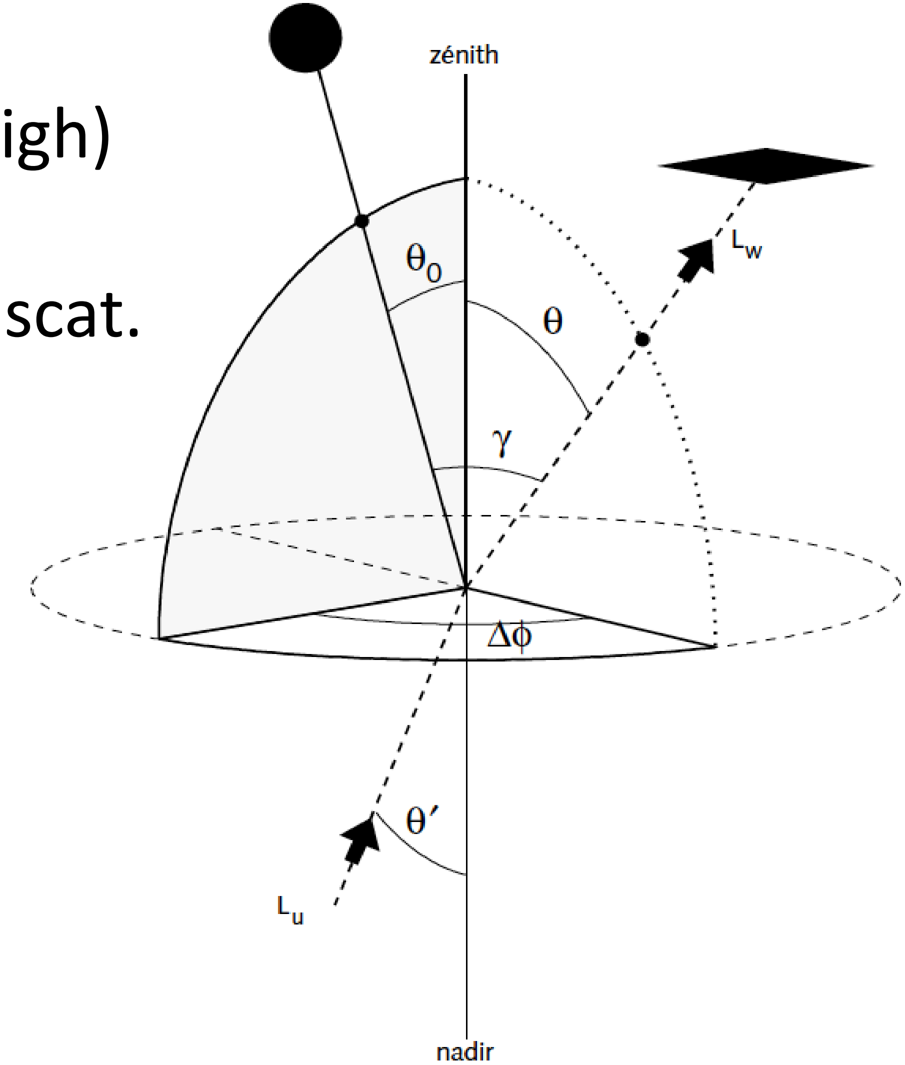


$$\rho_{wc}(\lambda)=\left[2.95\,10^{-6}\ ws^{3.52}\right]\rho_{ef}(\lambda)$$

From Gordon HR and M Wang, 1994. Applied Optics Vol. 33, Issue 33, pp. 7754-7763 https://doi.org/10.1364/AO.33.007754

Transmittances

$$L_t = L_r + L_a + L_{ra} + T.L_g + t.L_{wc} + t.L_w$$



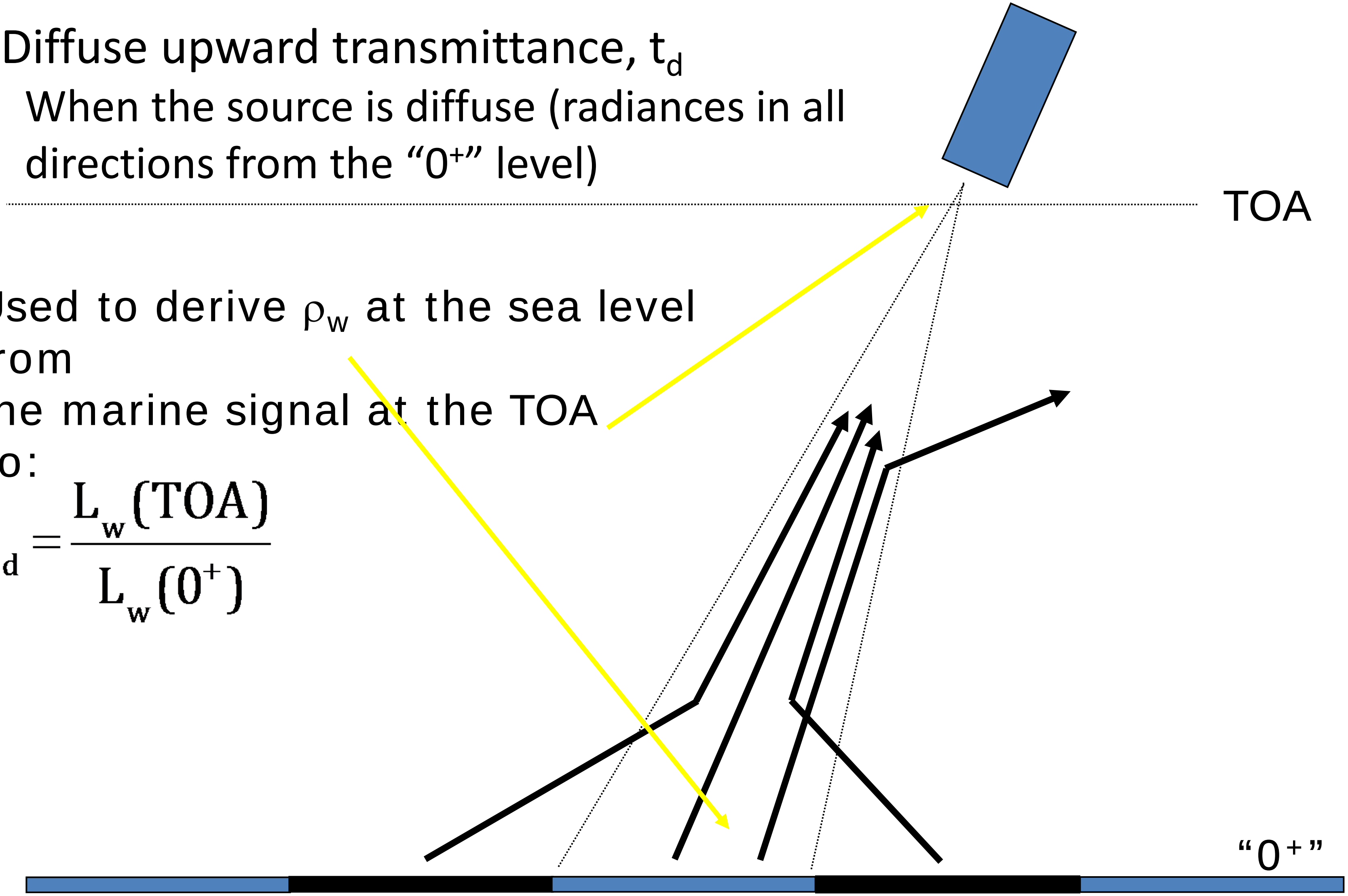
L_r : Molecular scattering (Rayleigh)
L_a : Aerosol scattering
L_{ra} : Molecular-aerosol coupled scat.
L_g : Sun glint
L_{wc} : White caps
L_w : Water-leaving signal
t, T: Transmittances

θ_s or θ_0: sun zenith angle
θ_v: satellite view angle
$\Delta\phi$: azimuth difference

Transmittances

1- Diffuse upward transmittance, t_d

When the source is diffuse (radiances in all directions from the "0^+" level)



Used to derive ρ_w at the sea level from the marine signal at the TOA so:

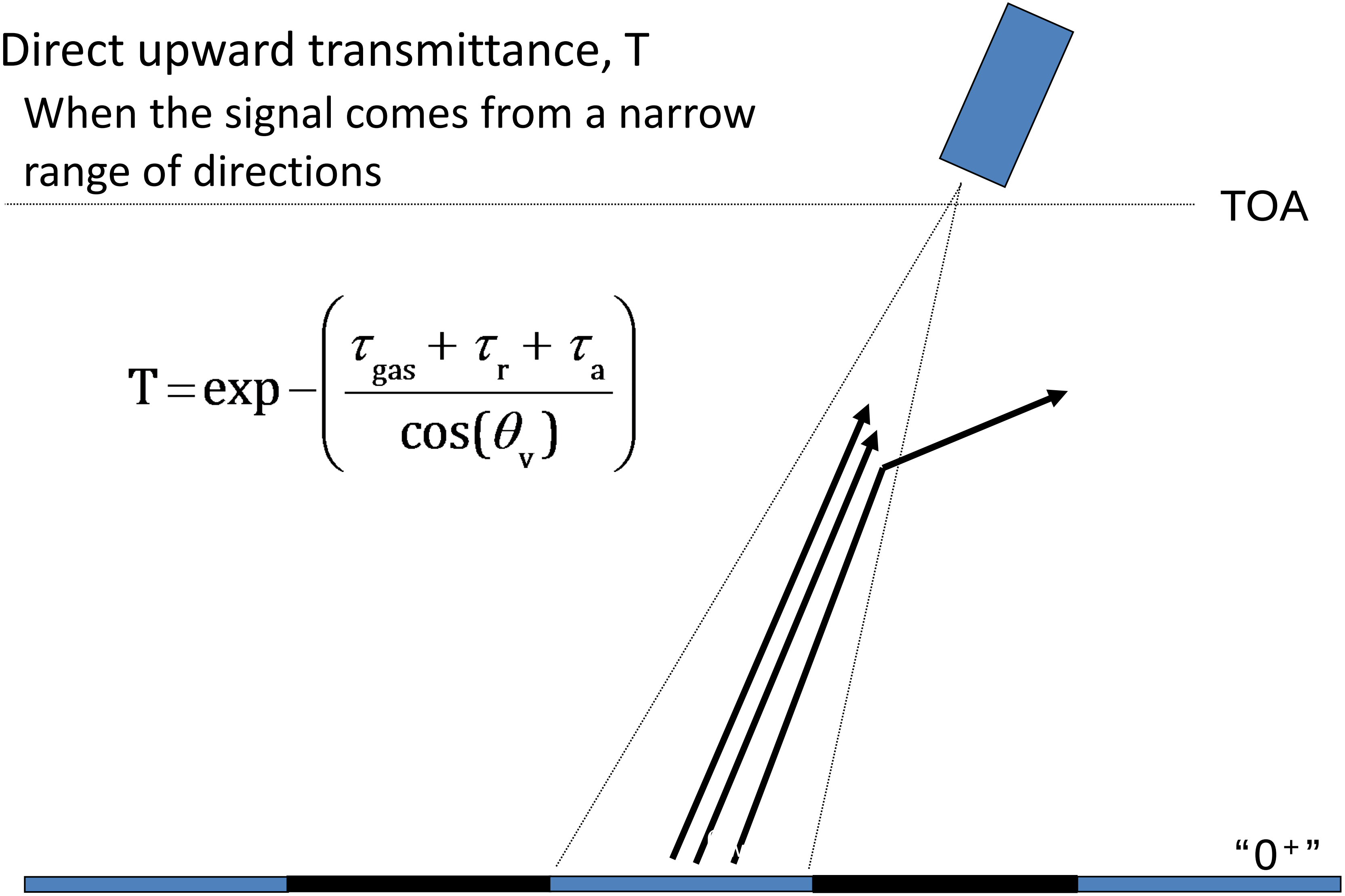
$$t_d = \frac{L_w(\mathrm{TOA})}{L_w(0^+)}$$

Transmittances

2- Direct upward transmittance, T

When the signal comes from a narrow range of directions

$$T = \exp -\left(\frac{\tau_{gas} + \tau_{r} + \tau_{a}}{\cos(\theta_{v})}\right)$$

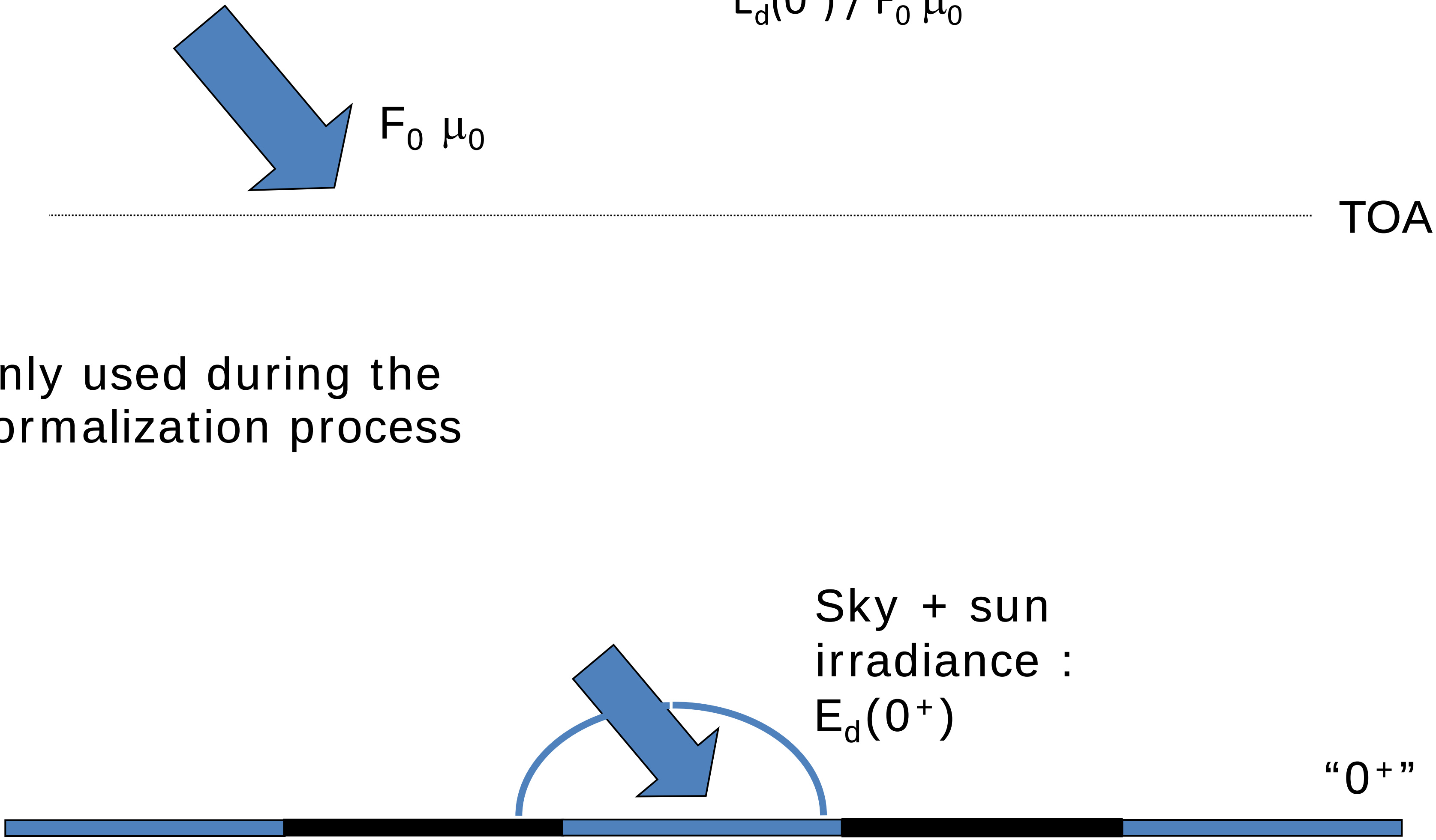


SUN GLINT

Transmittances

3 - Downward atmospheric transmittance for irradiance:

$E_d(0^+) / F_0 \mu_0$



Only used during the normalization process

Transmittances

1- Approximation for t_d

$$t_d = \exp -\left(\frac{\tau_{gas} + (1 - \omega_r f_r)\tau_r + (1 - \omega_a f_a)\tau_a}{\cos(\theta_v)} \right)$$

ω's: single scattering albedos (b/c)
f's: forward scattering probabilities

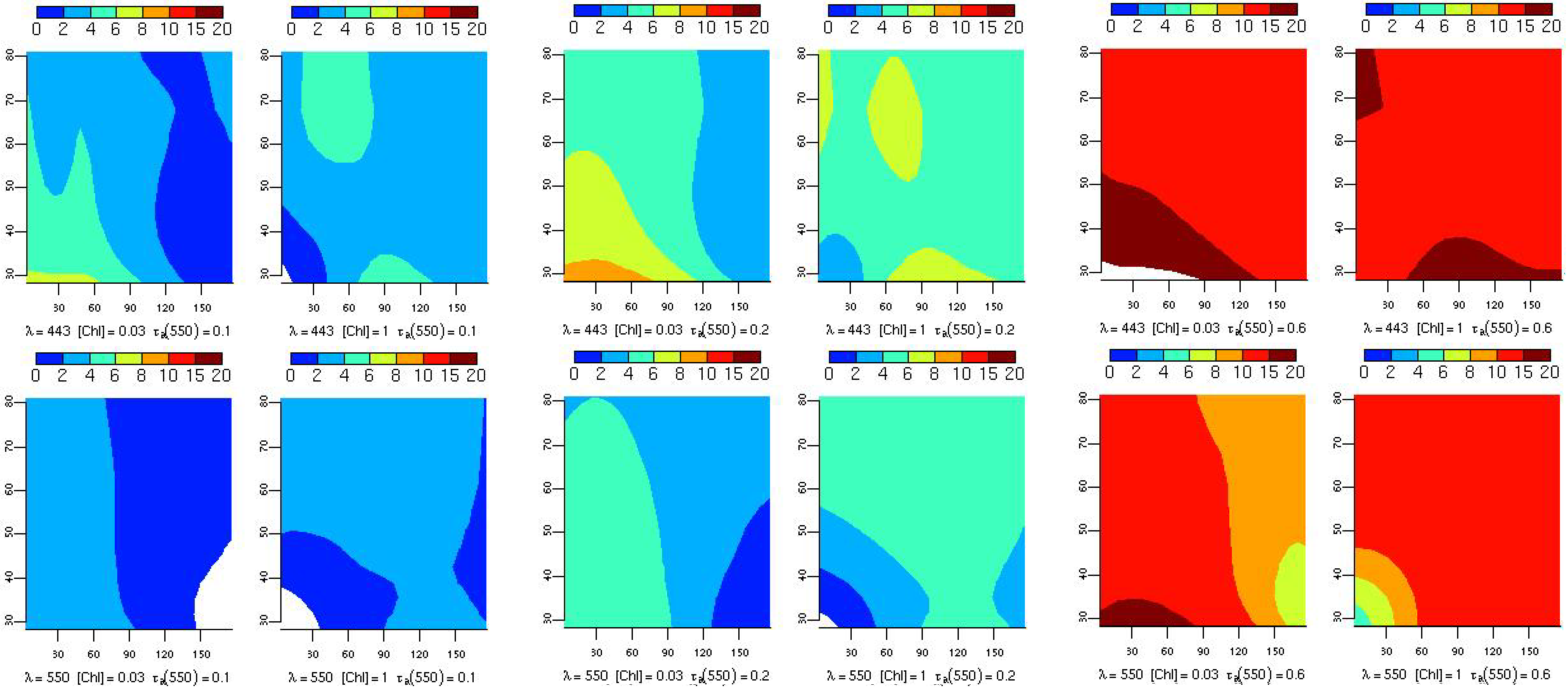
For the Rayleigh part: ω is 1 and f is 0.5
For the aerosol part: ω is close to 1 for non-absorbing aerosols, and f is often around 0.8
The expression therefore simplifies as:

$$t_d = \exp -\left(\frac{\tau_{gas} + 0.5\,\tau_r + 0.2\,\tau_a}{\cos(\theta_v)} \right)$$

And, when gaseous absorption is taken apart and the aerosol optical thickness is low, the expression is further reduced to the impact of Rayleigh only

Transmittances

How good the approximation for t_d is?



$\lambda = 443$ [Chl] = 0.03 $\tau_a(550) = 0.1$

$\lambda = 443$ [Chl] = 1 $\tau_a(550) = 0.1$

$\lambda = 443$ [Chl] = 0.03 $\tau_a(550) = 0.2$

$\lambda = 443$ [Chl] = 1 $\tau_a(550) = 0.2$

$\lambda = 443$ [Chl] = 0.03 $\tau_a(550) = 0.6$

$\lambda = 443$ [Chl] = 1 $\tau_a(550) = 0.6$

$\lambda = 550$ [Chl] = 0.03 $\tau_a(550) = 0.1$

$\lambda = 550$ [Chl] = 1 $\tau_a(550) = 0.1$

$\lambda = 550$ [Chl] = 0.03 $\tau_a(550) = 0.2$

$\lambda = 550$ [Chl] = 1 $\tau_a(550) = 0.2$

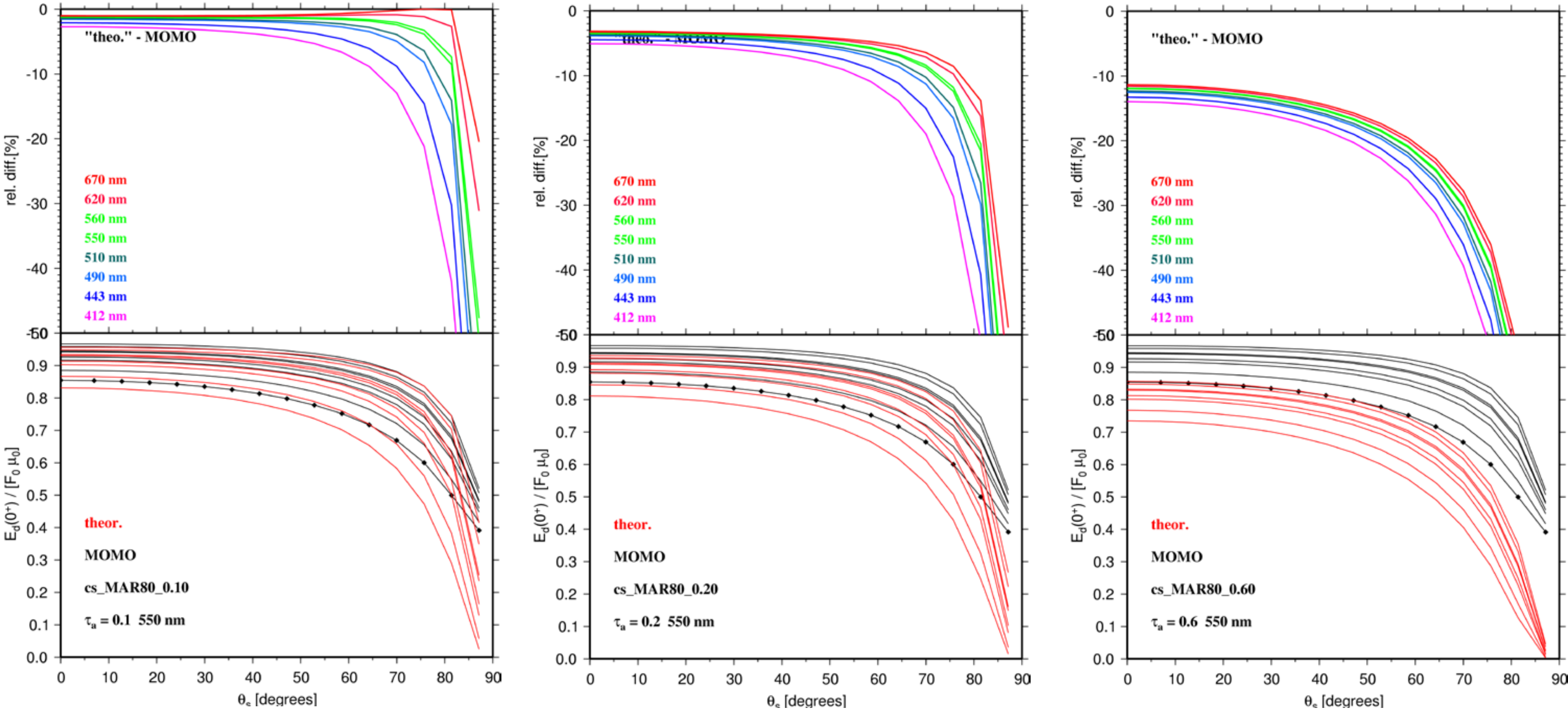
$\lambda = 550$ [Chl] = 0.03 $\tau_a(550) = 0.6$

$\lambda = 550$ [Chl] = 1 $\tau_a(550) = 0.6$

Comparison of t_d obtained via Monte Carlo RT simulations ($\theta_s = 60°$ and via the simplified expression (Δ in %))

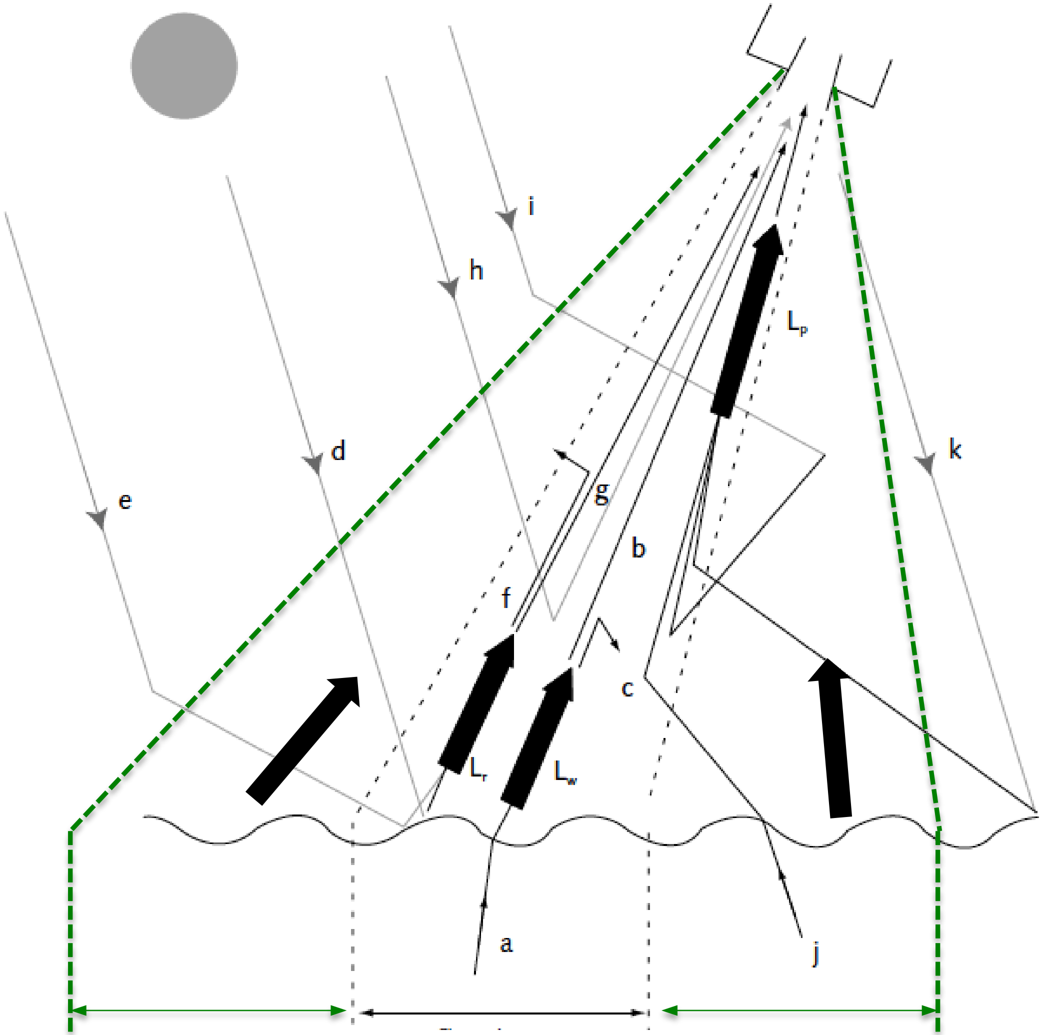
Transmittances

Can we use the t_d approximation for T ?



Now we use RT computations to derive T (and t_d)

Adjacency effects



Radiance from neighboring pixels can actually reach the sensor via forward scattering in the iFov direction of that pixel.

The importance of the effect will increase with:

- Heterogeneity between adjacent pixels
- Amount of scattering

The impact of neighboring pixels will decrease as their distance from the pixel increase

Adjacency effects exist in all cases, yet they are significant only when a strong difference in reflectance exist between adjacent pixels, e.g., cloud border or land mass.

Turbid waters

The difficulty here is that the marine signal in the NIR is no longer negligible.

Two main options:

- 1) using another spectral domain where L_w is zero!
  That is where the short wave infrared (SWIR) can help

- 2) Re-introduce a model for the marine component, and then use an appropriate technique to disentangle the marine and atmospheric signals.
  Include: Spectral matching techniques, spectral optimization techniques, neural networks, genetic algorithms, or more simple iterative schemes

Turbid waters: using SWIR bands

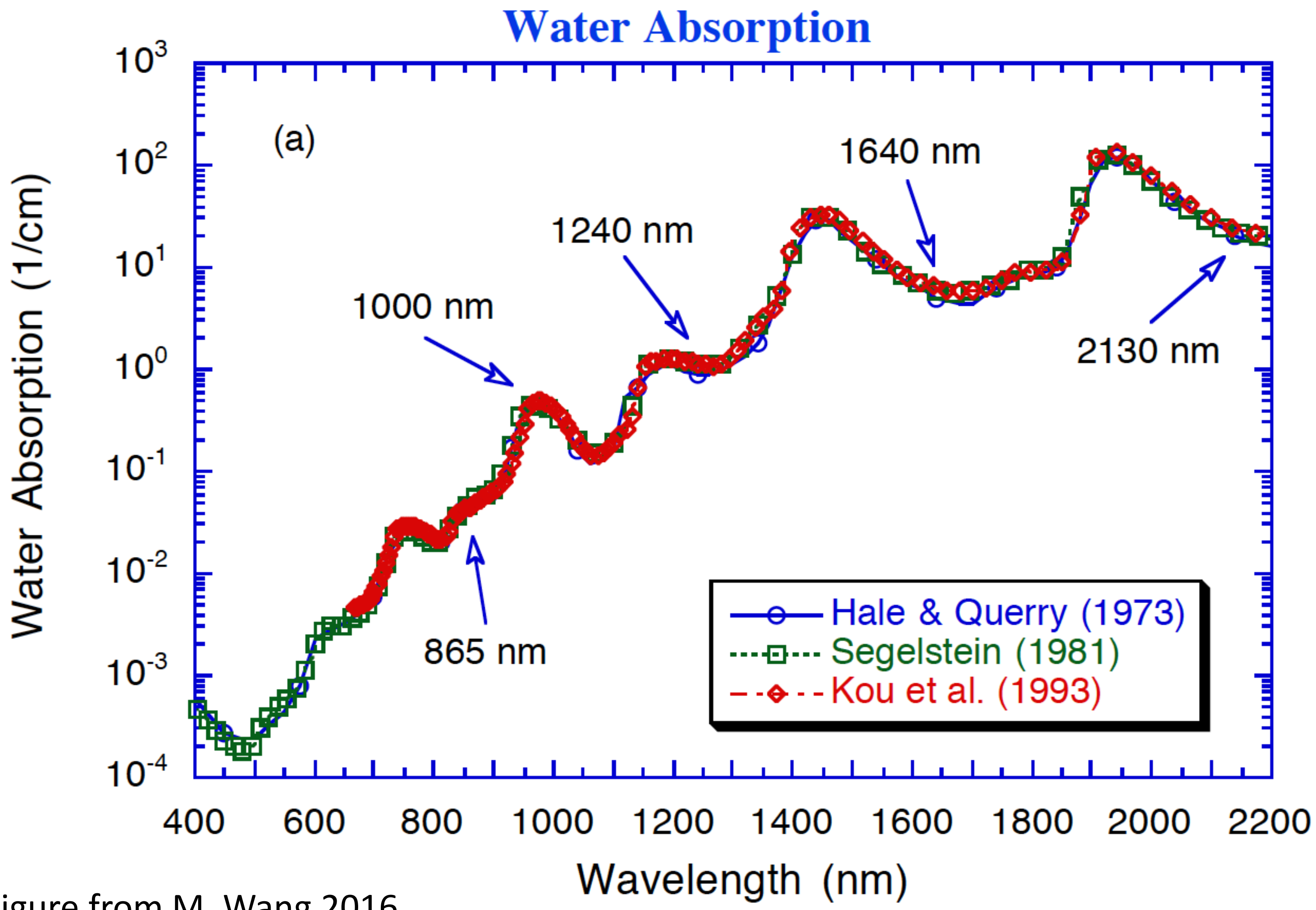


Figure from M. Wang 2016

Turbid waters: using NIR & SWIR bands

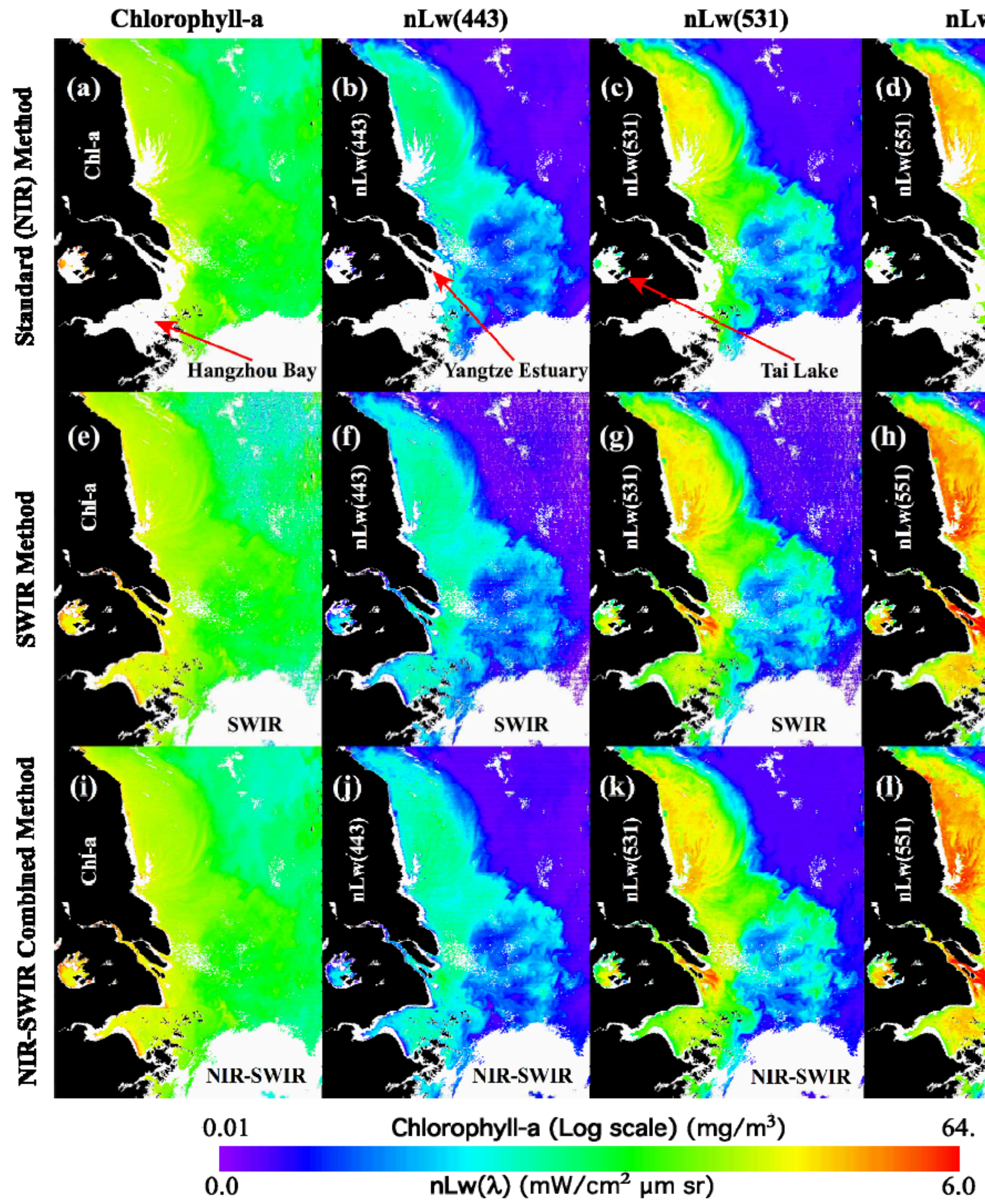
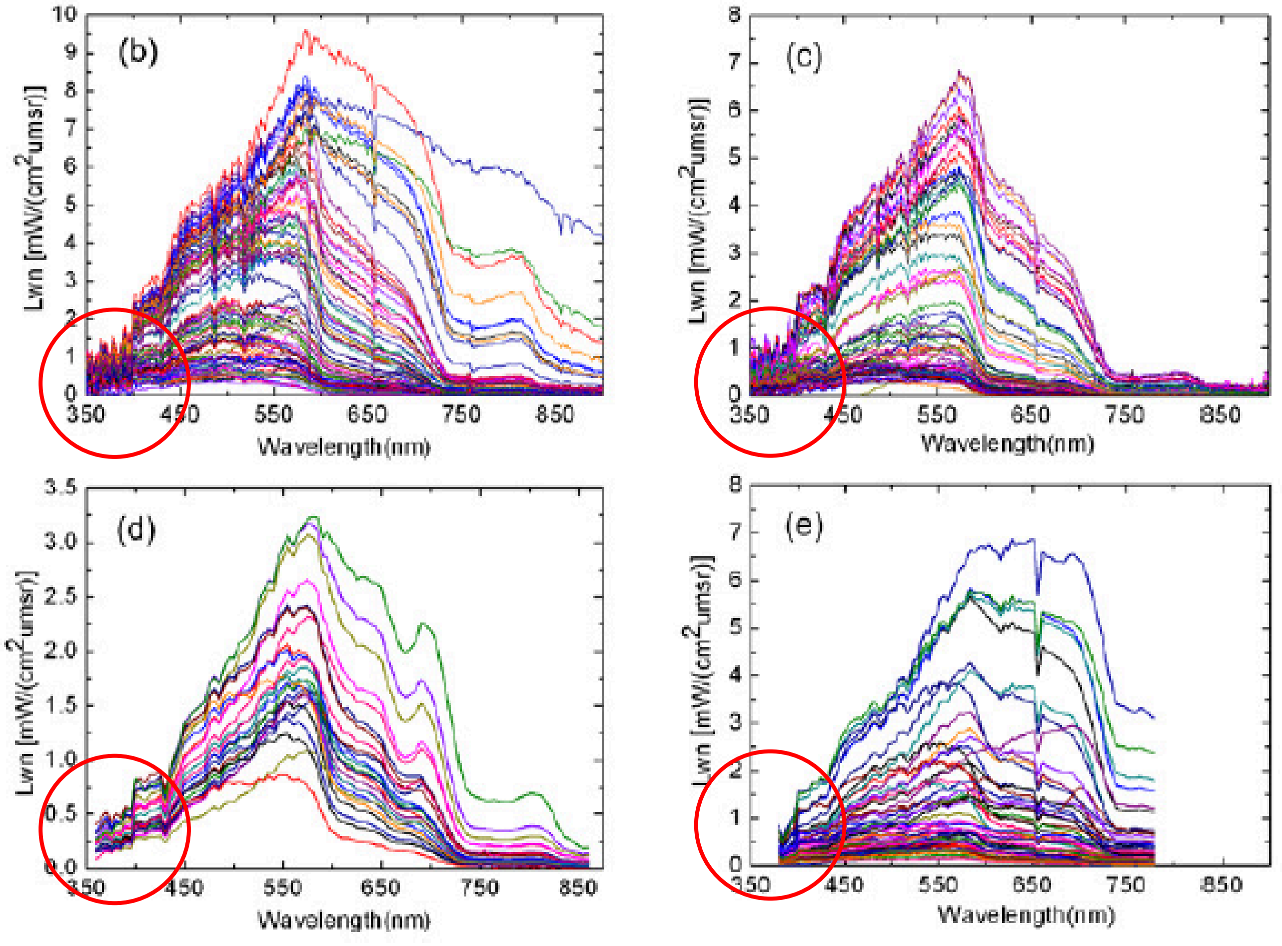


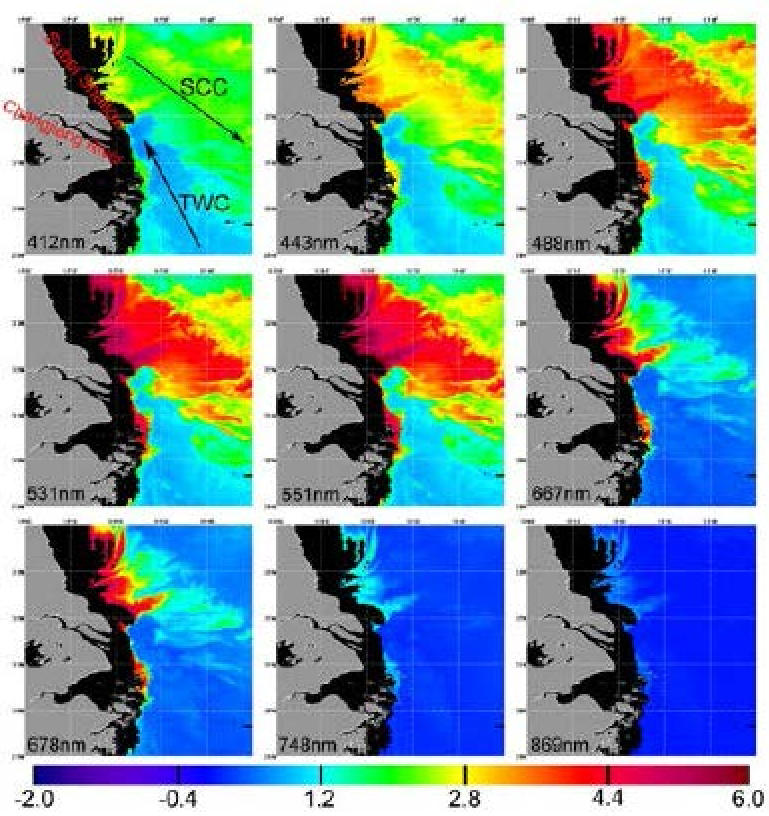
Figure 5 in: Wang and Shi, 2007, Optics Express, 15(24), 15722-15733

Turbid waters: using near UV bands

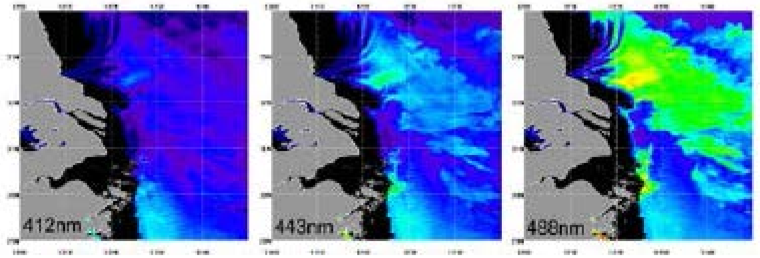


From Fig. 1 in He et al., 2012. OPTICS EXPRESS, 20(18), 20754-20770

Turbid waters: using near UV bands



Using the near UV bands



Using the classical near IR bands

From Figs. 4 and 5 in He et al., 2012. OPTICS EXPRESS, 20(18), 20754-20770

Other methods

An example: “POLYMER”

(“POLYnomial based algorithm applied to MERIS”)

The total reflectance is first corrected by sun glint (rough correction) and for molecular scattering, ending up with:

$$\rho'(\lambda) = \rho_{ag}(\lambda) + t(\lambda)\rho_w^+(\lambda)$$

Where:

$$\rho_{ag}(\lambda) = \Delta\rho_{gli}(\lambda) + \rho_{aer}(\lambda) + \rho_{coupl}(\lambda)$$

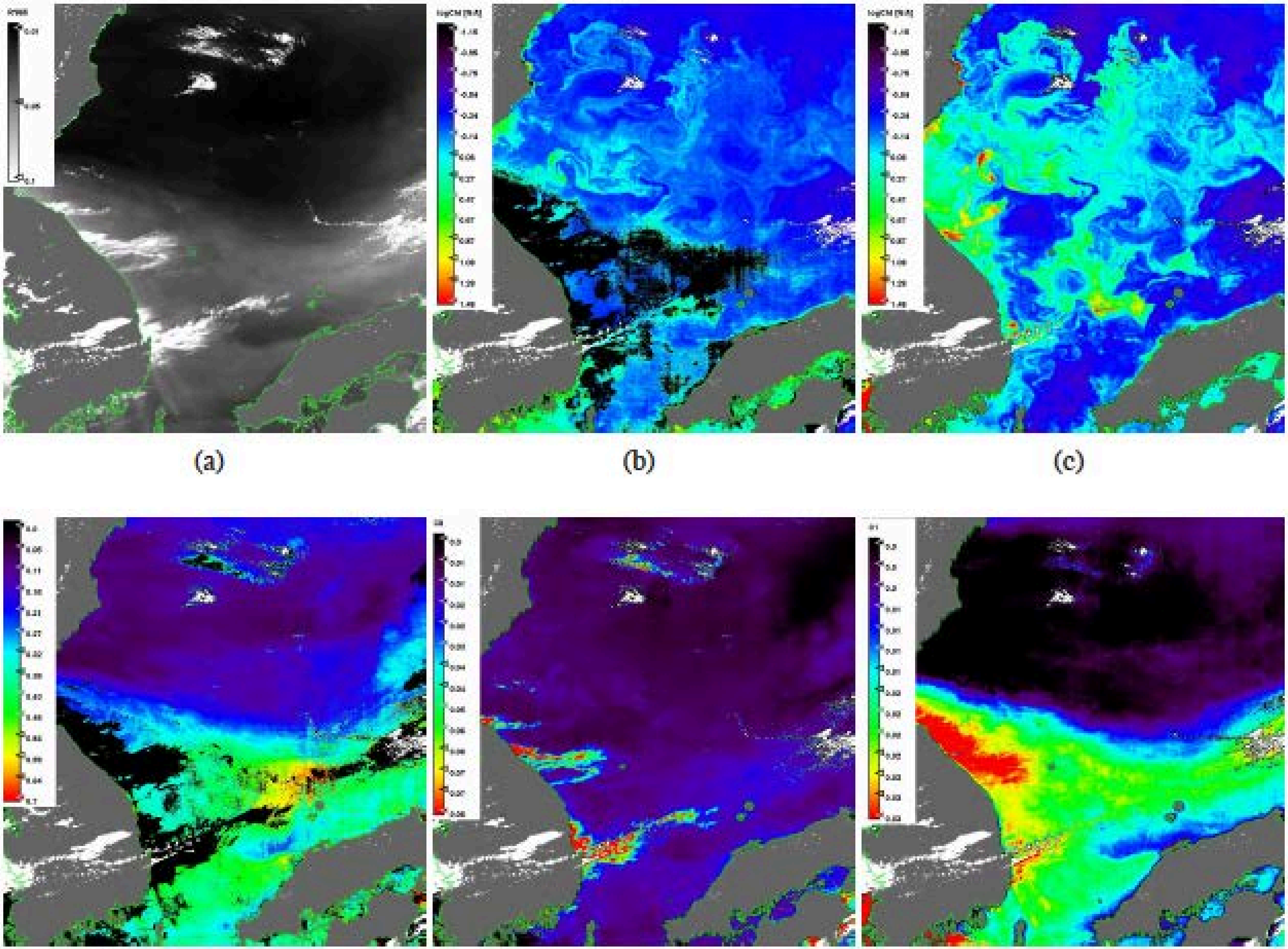
Then, this term is expressed as:

$$\rho_{ag}(\lambda) \approx T_0(\lambda)c_0 + c_1\lambda^{-1} + c_2\lambda^{-4}$$

The coefficients **c**0 , **c**1 and **c**2 are estimated by least square fitting of the observation, where the ocean reflectance is modelled (about) as per Morel Maritorena 2001. That’s a spectral matching algorithm

- Steinmetz, F., P.-Y. Deschamps, and D. Ramon, 2011. Atmospheric correction in presence of sun glint: application to MERIS, Opt. Express 19, 9783-9800
- Morel, A., and S. Maritorena, 2001. Bio-optical properties of oceanic waters: A reappraisal. Journal of Geophysical research, 106, 7763-7780

POLYMER, continued



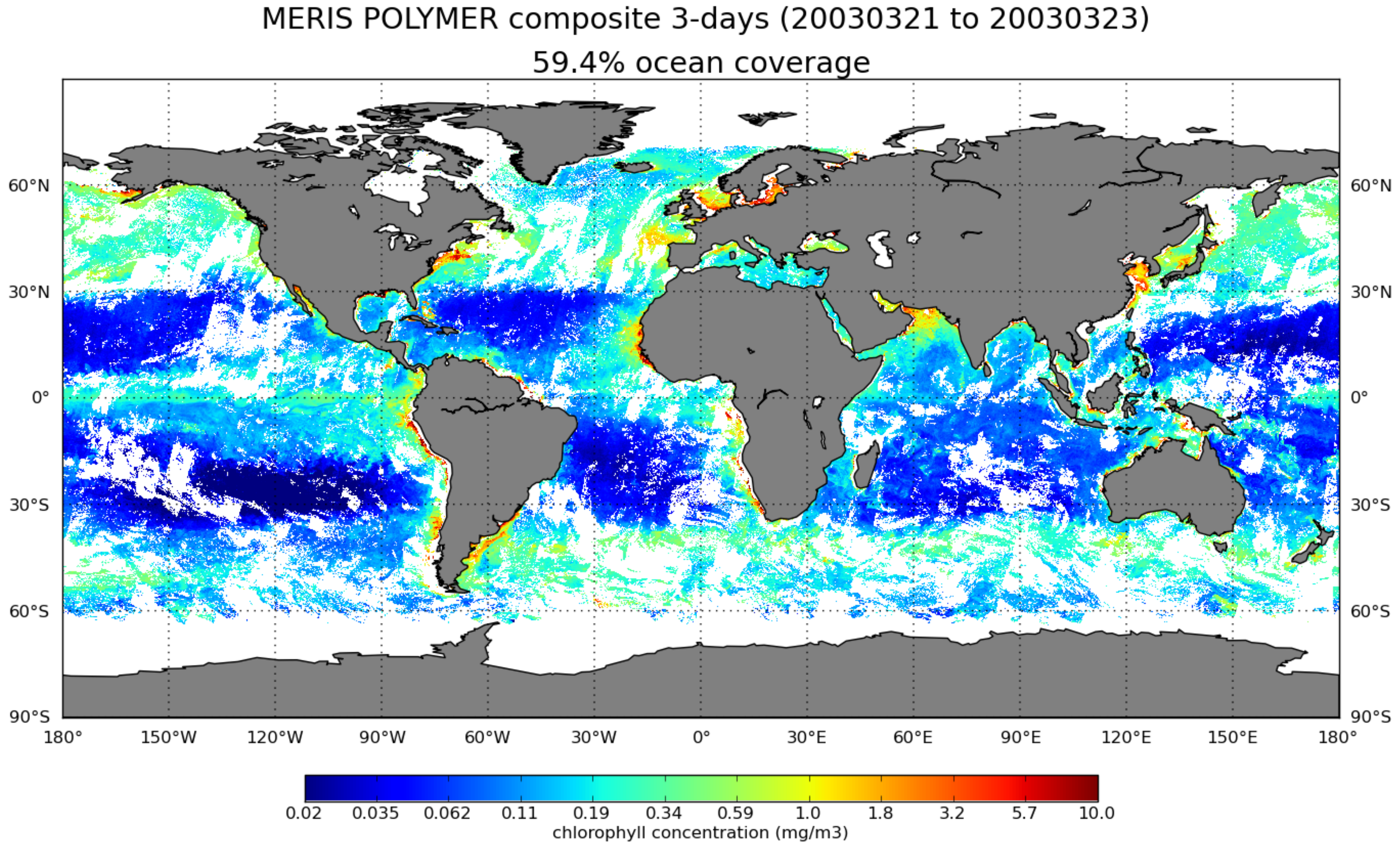
(a) (b) (c)

(d) (e) (f)

Fig. 5 in Steinmetz, et al., 2011

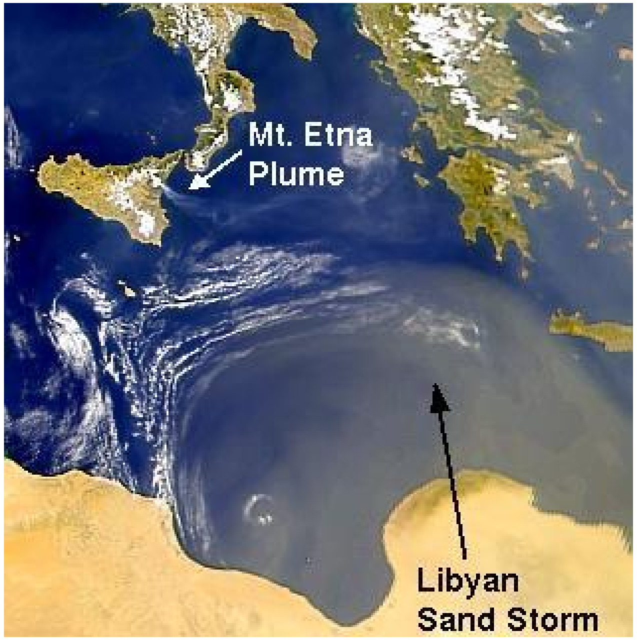
MERIS image from Sea of Japan / East Sea, 2004-03-13.

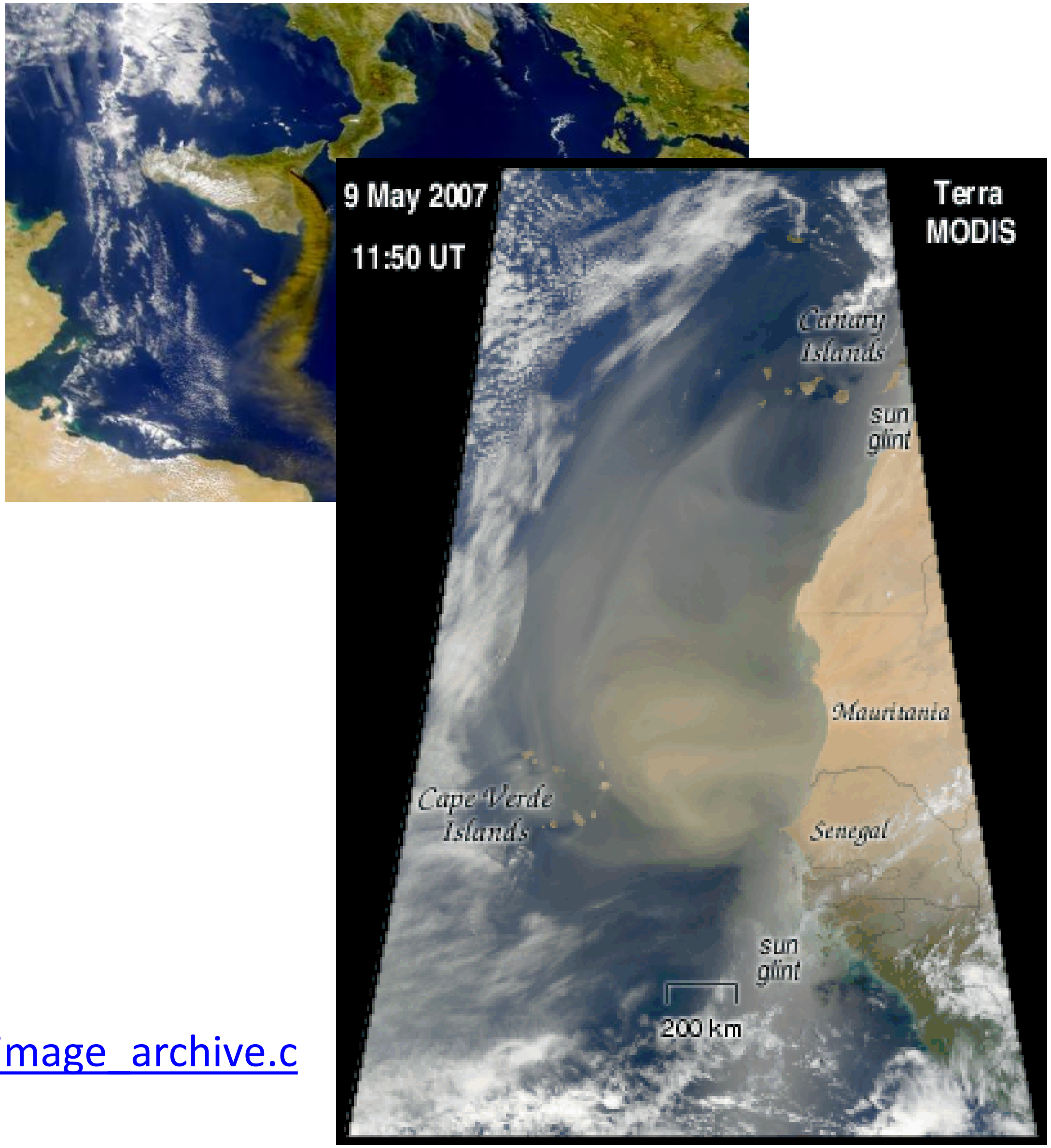
POLYMER, continued



From: https://www.hygeos.com/polymer

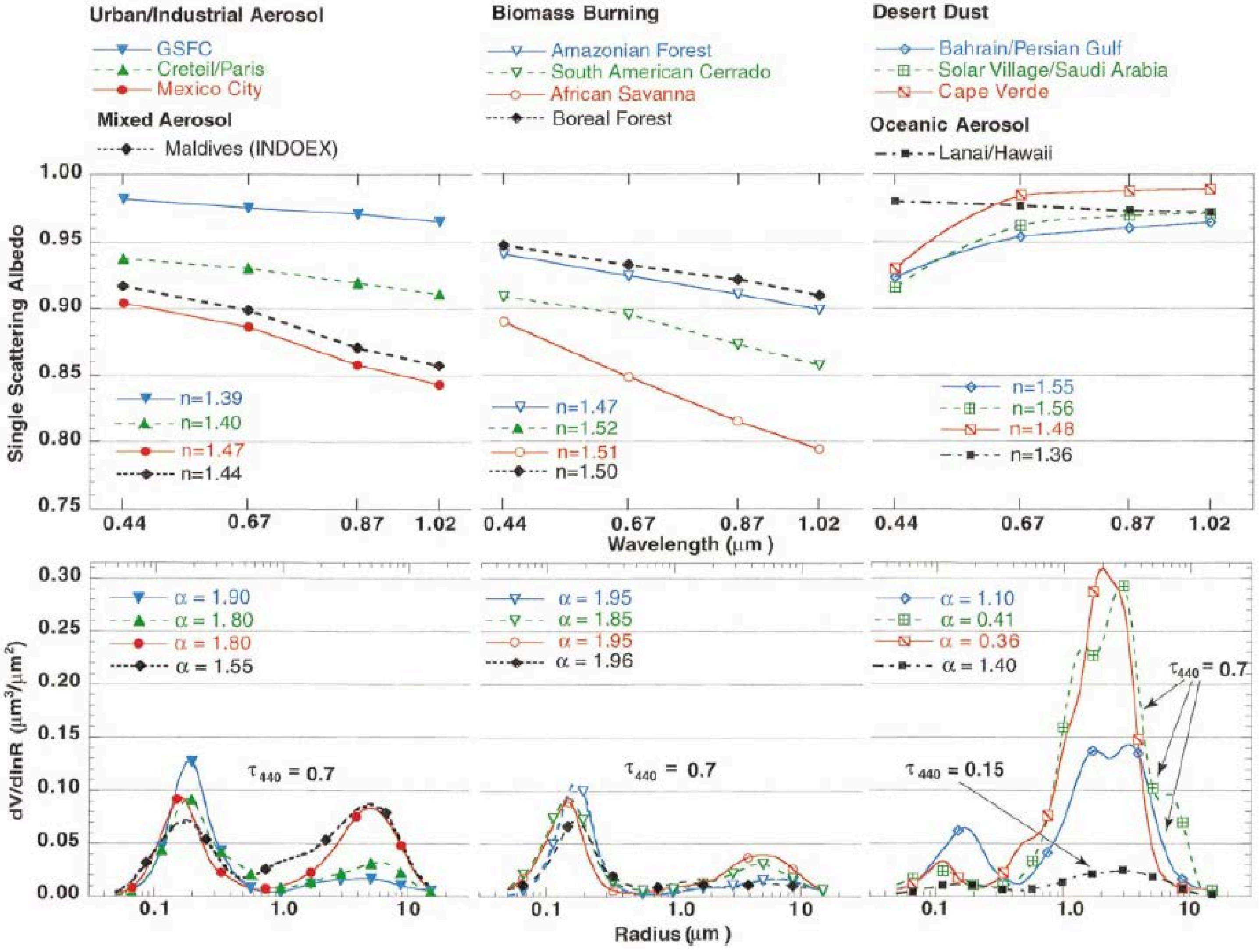
Absorbing aerosols





All images from:
https://oceancolor.gsfc.nasa.gov/cgi/image_archive.cgi?c=DUST

Absorbing aerosols



Dubovik et al., 2002. J. Atm. Sci., 59, 590-608

Absorbing aerosols

Now the issue is that the altitude and thickness of the absorbing aerosols layer make a big difference on the TOA signal

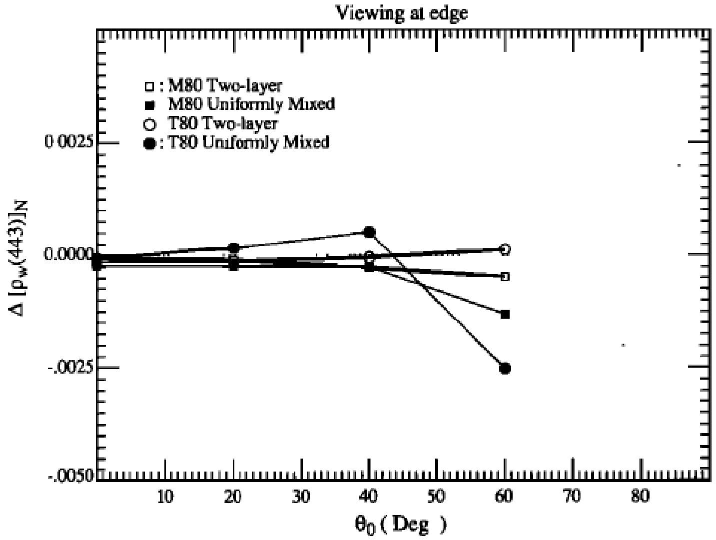


Figure 14a. Effect of the vertical distribution of aerosol on $\Delta[\rho_w(443)]_N$ as a function of θ_0 at the edge of the scan for the T80 and M80 aerosol models with $\tau_a(865) = 0.2$. Note that the correction algorithm assumes that the "two-layer" stratification is correct.

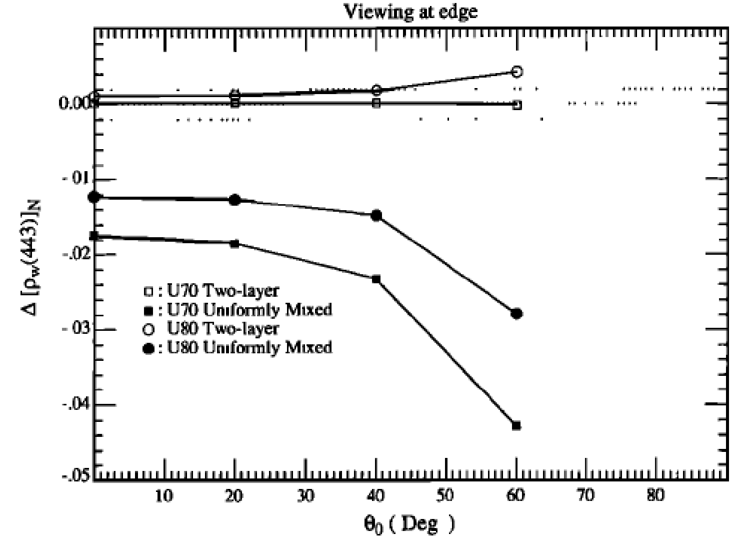


Figure 14b. Effect of the vertical distribution of aerosol on $\Delta[\rho_w(443)]_N$ as a function of θ_0 at the edge of the scan for the U80 and U70 aerosol models with $\tau_a(865) = 0.2$. Note that the correction algorithm assumes that the "two-layer" stratification is correct.

From: Gordon, HR, 1997. J. Geophys. Res., 102(D14), 17,081-17,106

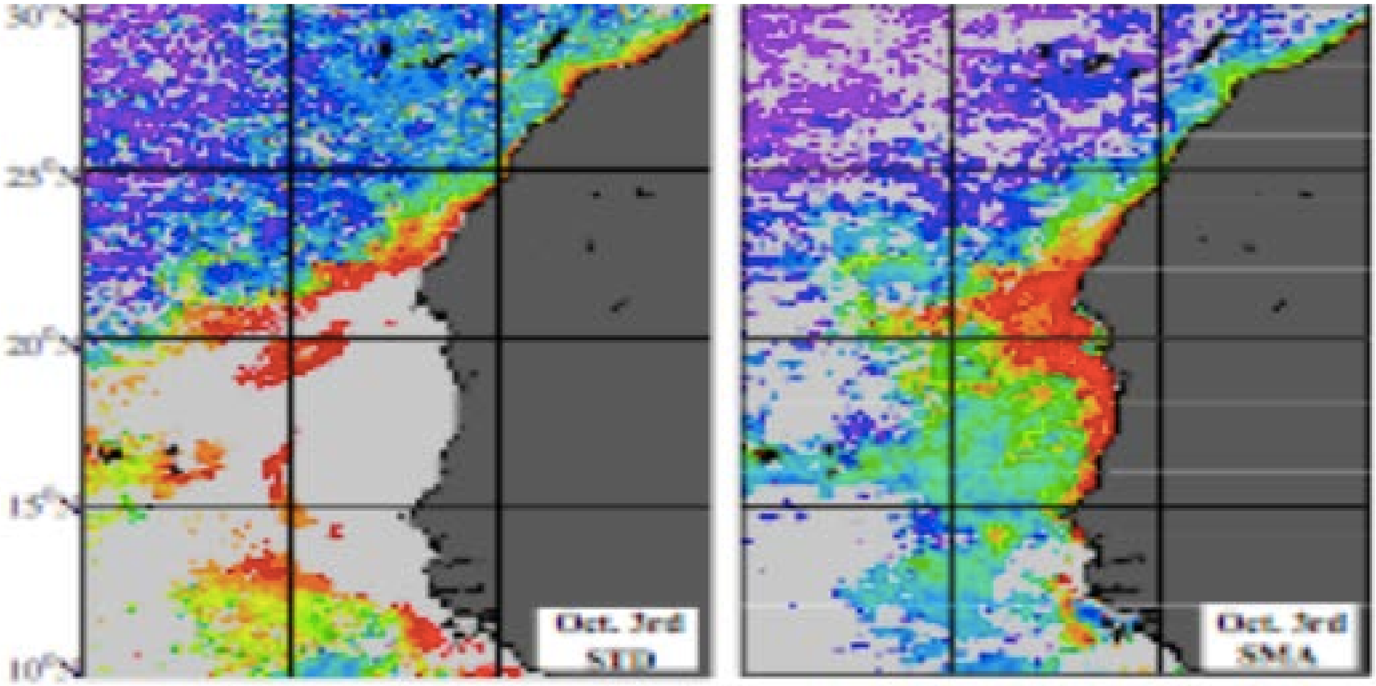
Absorbing aerosols

Solution: spectral matching algorithms (SMA), e.g.: Gordon et al, 1997.

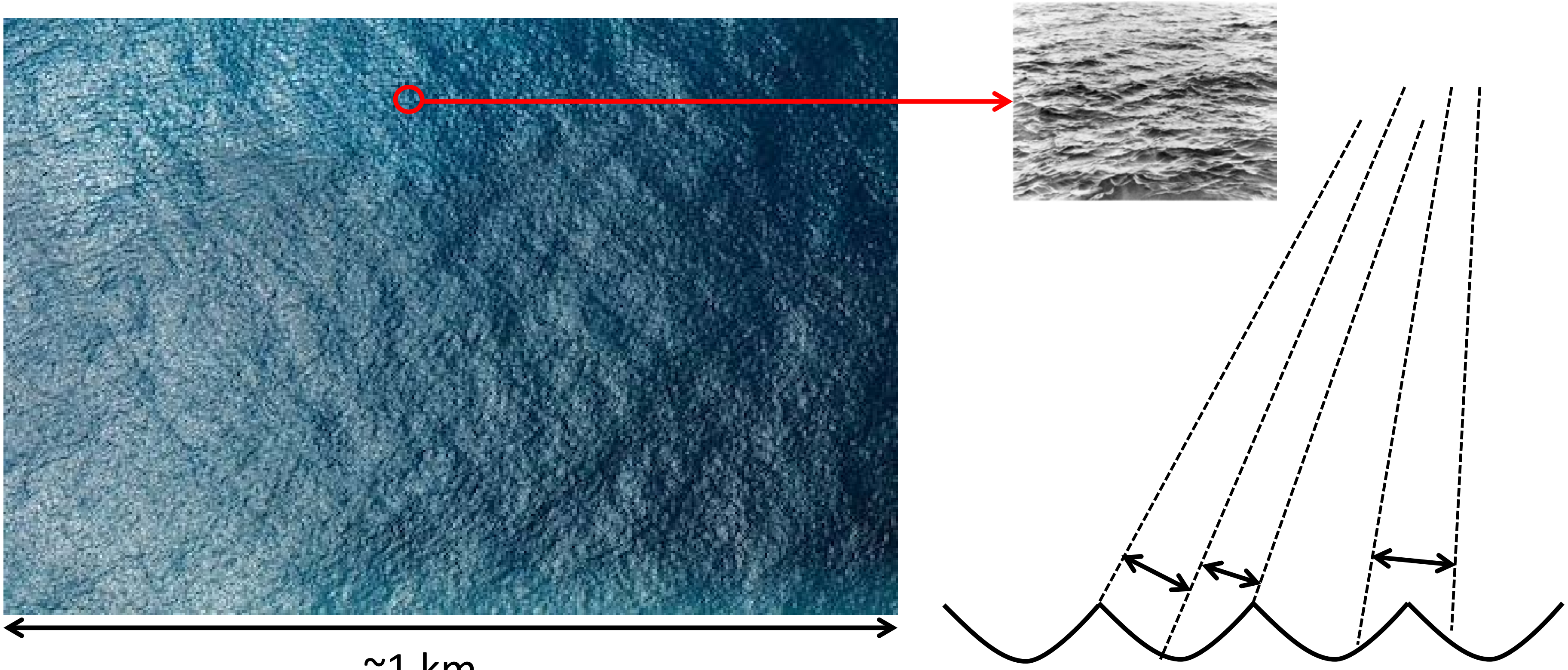
Retrieves the properties of the ocean and atmosphere simultaneously using reflectances from all bands.

Black pixel assumption

Case I water model



High spatial resolution sensors: Surface effects



Valid for medium resolution sensors

$$p(\theta_s,\theta_v,\Delta\phi)=\frac{1}{\pi\,\sigma^2}\exp\left(\frac{-\tan^2(\beta)}{\sigma^2}\right)$$

$$\sigma^2 = 0.003 + 5.12\ 10^{-3}\ W$$

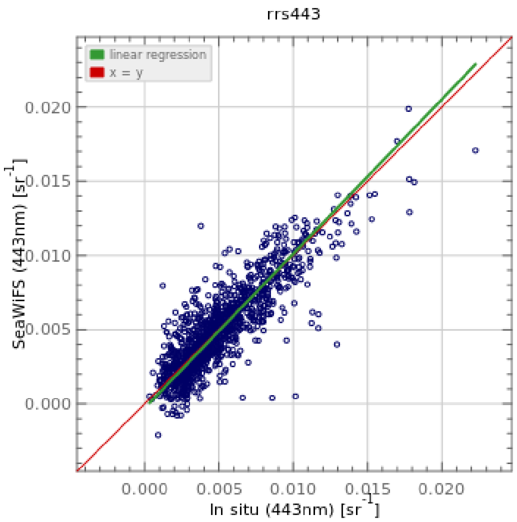
$$\rho_G = \pi\,\rho_F\ p(\theta_s,\theta_v,\Delta\phi)\ /\ (4\cos(\theta_v)\cos(\theta_s)\cos^4(\beta))T(\theta_v)\ T(\theta_s)$$

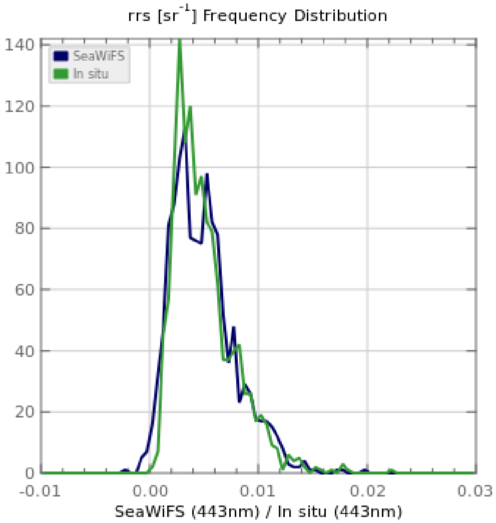
Cox C. and W. Munk, 1954. Statistics of the sea surface derived from sun glitter. J. Mar. Res., 13, 198-227.

Validation of atmospheric corrections

Two main ways of doing it (not exclusive)

- Validating the end-result of AC: water-leaving radiance
- Validating the quantities that the AC produces (basically the path radiance)

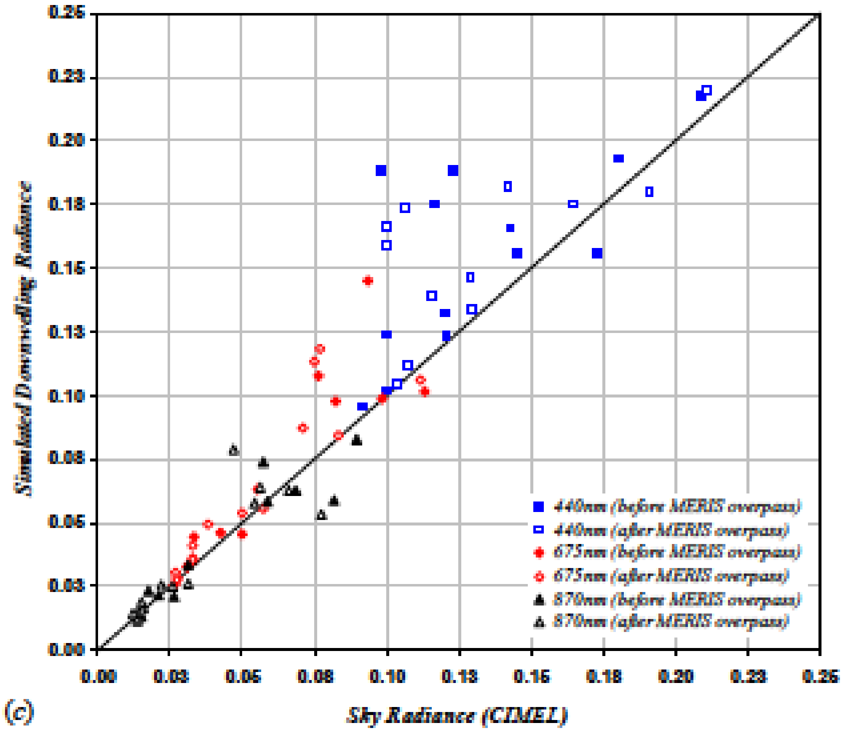


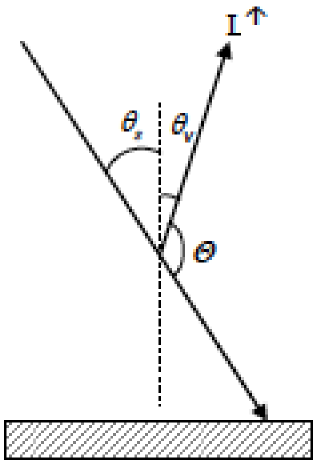


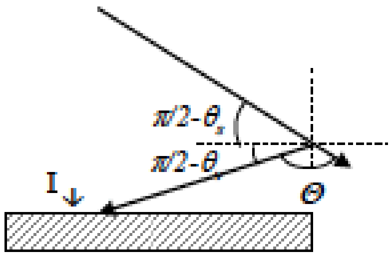
https://seabass.gsfc.nasa.gov/search/?search_type=Perform%20Validation%20Search&val_sata=5&val_products=11

Validation of atmospheric corrections

Validating the path radiance (but issue with the scattering angle)



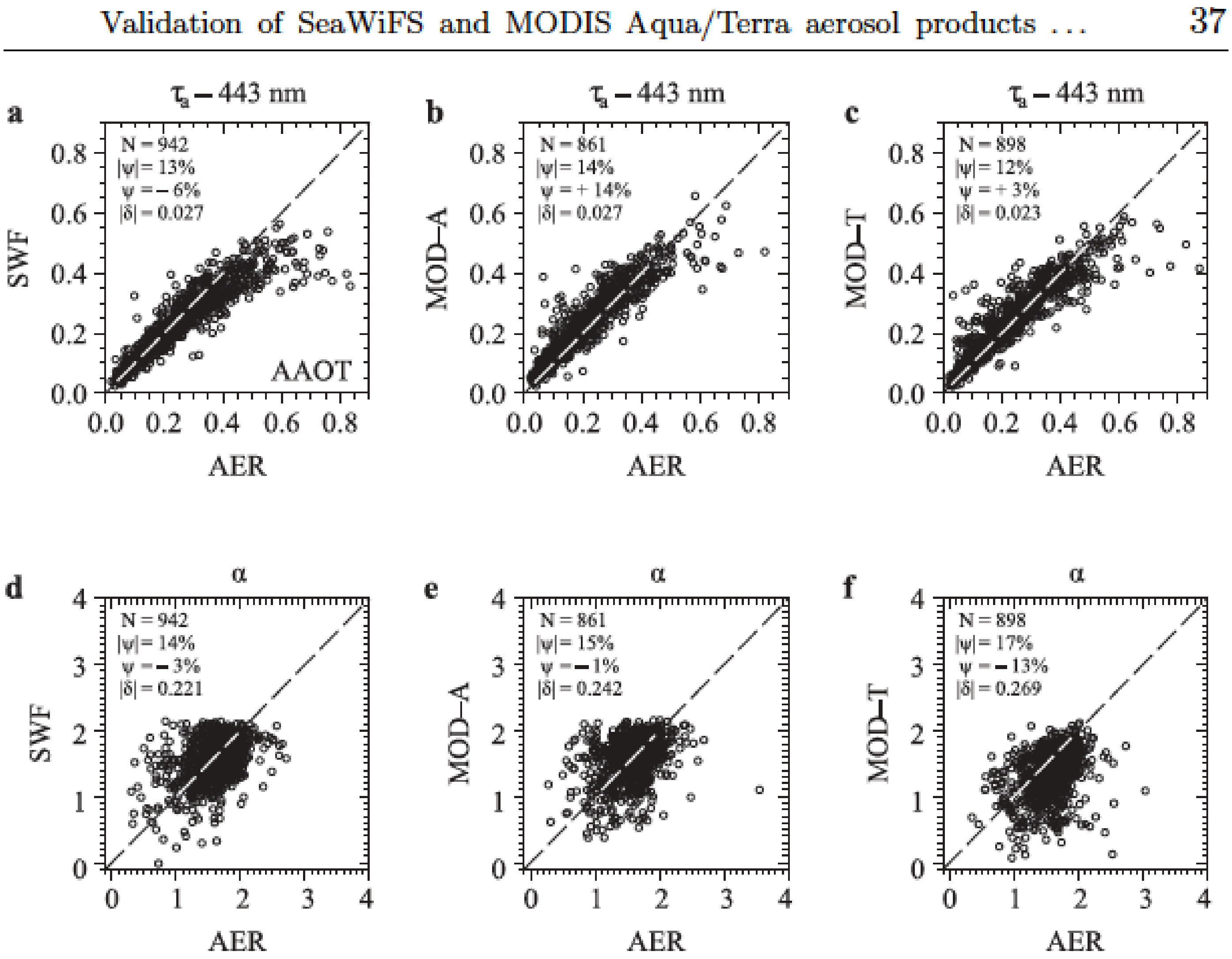




From Figs 4 and 5 in Zagolski et al., ESA proceedings

https://earth.esa.int/pi/esa;jsessionid=C443838F48BED636FD0241B8EB415424.jvm1?colorTheme=03&type=upload&ts=1487899749891&table
progressreport&targetIFramePage=%2Fweb%2Fguest%2Fpi-community%2Fsearch-results-and-projects%2Fapplication&cmd=image&id=7641

Validation of “by-products” of AC



Validation of SeaWiFS and MODIS Aqua/Terra aerosol products . . . 37

From: Mélin et al., 2013. OCEANOLOGIA, 55 (1). 27–51. doi:10.5697/oc.55-1.027

Atmospheric correction of satellite OCR is not aerosol remote sensing

The simple way of saying why is: "**What we (the Ocean colour community) try to eliminate as the unwanted signal is what the aerosol community is interested in**"

More precisely, however:

As reminded previously in this lecture: the goal for atmospheric correction of satellite OCR is the extrapolation of the aerosol signal in the NIR to the VIS
We may do this well, and still have poor-quality aerosol by-products, such as the aerosol optical thickness (AOT)

Why ???

THANK YOU