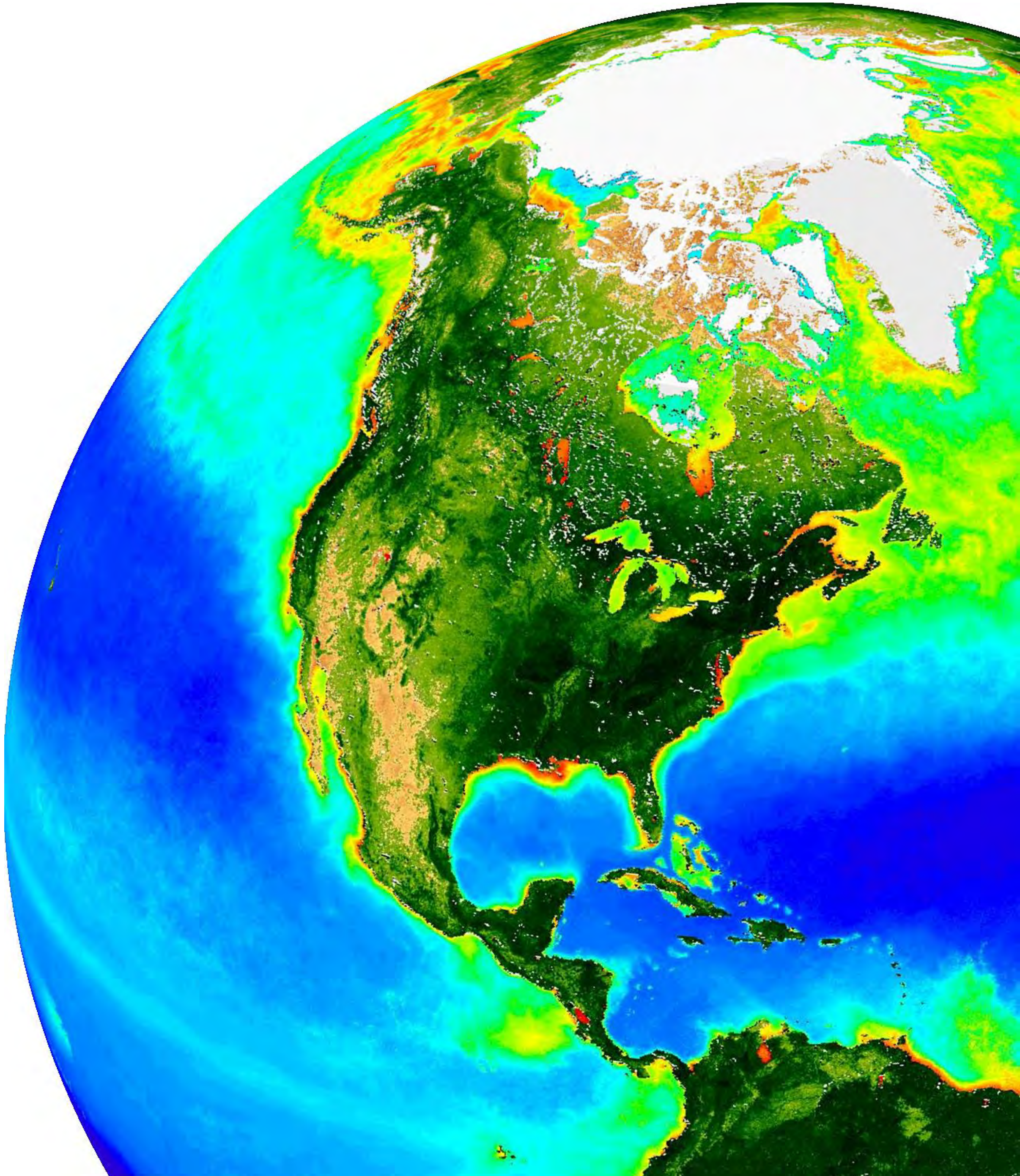
Ocean color satellite atmospheric correction

Jeremy Werdell
NASA Goddard Space Flight Center
(with some tweaks by Curtis Mobley)

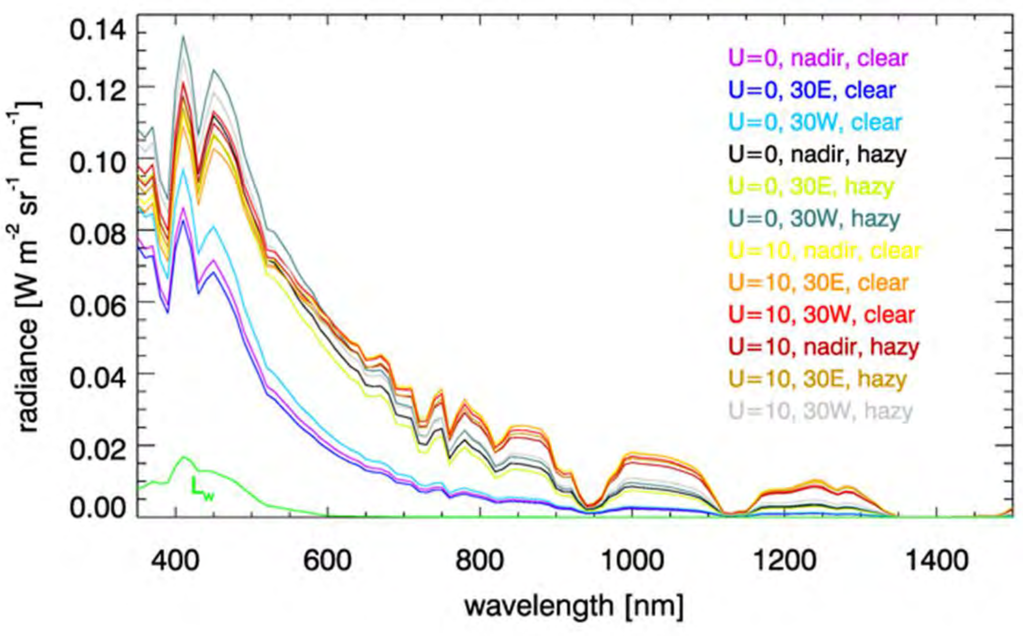
Acknowledgements: Zia Ahmad, Sean Bailey, Bryan Franz, Amir Ibrahim, and Fred Patt

2021 Ocean Optics Summer Course



Today's Lecture

The previous lecture illustrated the atmospheric correction problem



Today's lecture shows one solution

- work through the steps of ocean color atmospheric correction as used by NASA/OBPG
- identify the places where ancillary (external) data are required
- identify the places where in situ data and bio-optical models are used

For the details see

NASA/TM–2016-217551

Atmospheric Correction for Satellite Ocean Color Radiometry

Curtis D. Mobley
Sequoia Scientific, Inc., Bellevue, WA

Jeremy Werdell
NASA's Goddard Space Flight Center, Greenbelt, MD

Bryan Franz
NASA's Goddard Space Flight Center, Greenbelt, MD

Ziauddin Ahmad
NASA's Goddard Space Flight Center, Greenbelt, MD

Sean Bailey
NASA's Goddard Space Flight Center, Greenbelt, MD

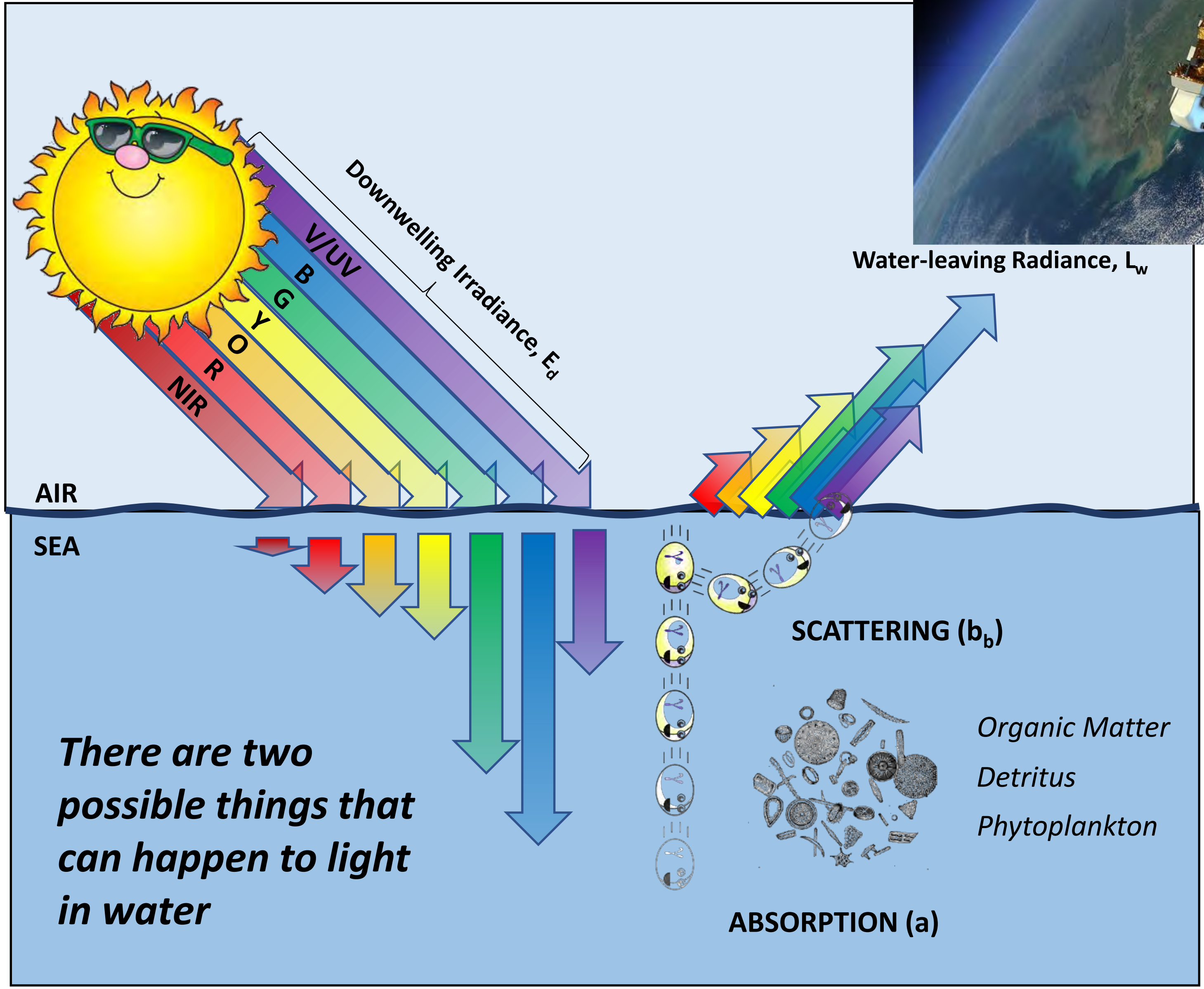
https://oceancolor.gsfc.nasa.gov/docs/technical/NASA-TM-2016-217551.pdf

OCEAN OPTICS
Web Book

Contents

Chapter 10 of *The Ocean Optics Book*

http://www.oceanopticsbook.info



Focus of this lecture

satellite ocean color

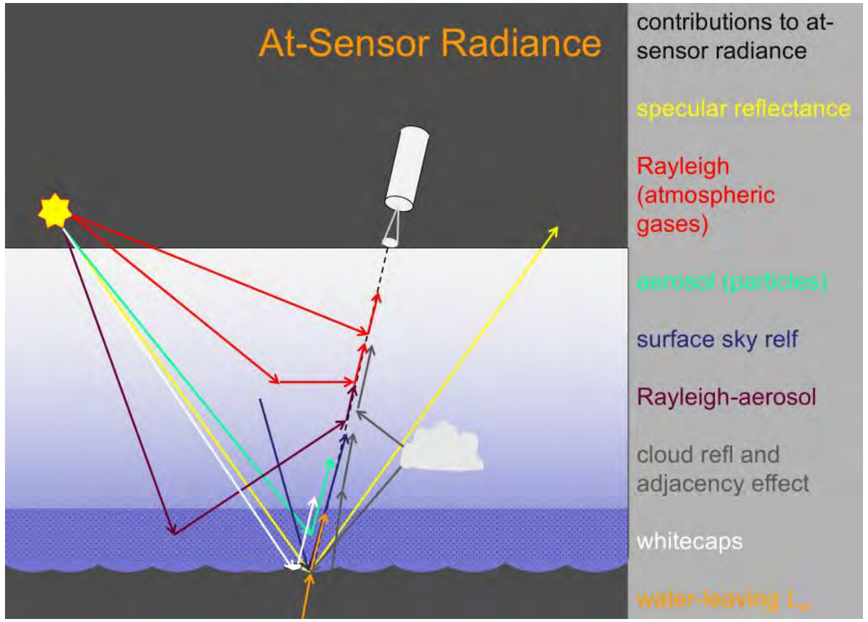
ocean color satellites measure top-of-atmosphere radiances

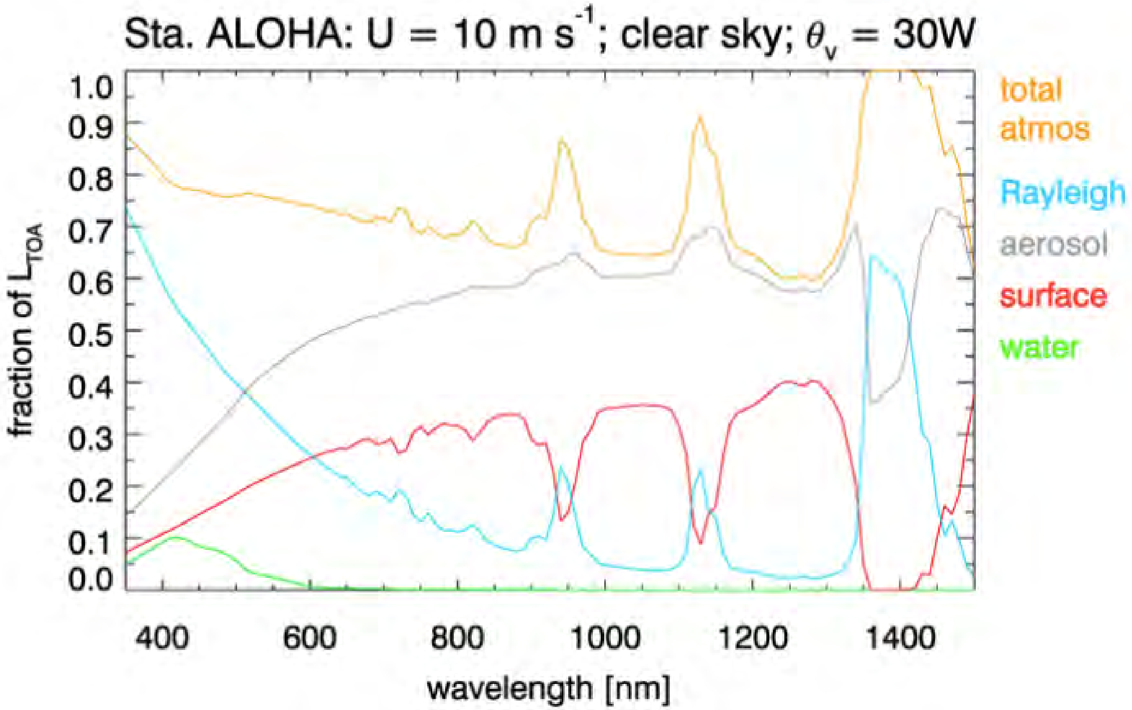
Rayleigh, aerosols, aerosols+Ray, Sun glint, surface-reflected background sky radiance, white caps (foam), water-leaving

$$(1) \quad L_t = L_R + [L_a + L_{aR}] + L_g + L_{sky} + L_f + L_w$$

L_w is what we want

where all radiances are defined at the top-of-atmosphere (TOA)





Processing Constraints

MODIS and VIIRS have about 14 orbits/day

The distance between terminators is ~20,000 km. The MODIS +/-50 degrees of scan angle gives ~1220 pixels, or about 24.4 Mpixels/orbit. For VIIRS the useful scan angle range corresponds to ~2480 pixels, and the along-track resolution is 0.75 km, so about 66.1 Mpixels/orbit. If we limit processing to SZA < 75 degrees, then the numbers are 5/6 of the above, which is 20.3 Mpix for MODIS and 55.1 for VIIRS.

Currently have 2 MODIS and 2 VIIRS in orbit (coming soon, PACE and 3 more VIIRS), so 151 Mpixels/orbit, times 14 orbits/day = 2.1 Gigapixels/day to be processed. **Every pixel gets its own atmospheric correction.**

Processing Constraints

MODIS and VIIRS have about 14 orbits/day

The distance between terminators is ~20,000 km. The MODIS +/-50 degrees of scan angle gives ~1220 pixels, or about 24.4 Mpixels/orbit. For VIIRS the useful scan angle range corresponds to ~2480 pixels, and the along-track resolution is 0.75 km, so about 66.1 Mpixels/orbit. If we limit processing to SZA < 75 degrees, then the numbers are 5/6 of the above, which is 20.3 Mpix for MODIS and 55.1 for VIIRS.

Currently have 2 MODIS and 2 VIIRS in orbit (coming soon, PACE and 3 more VIIRS), so 151 Mpixels/orbit, times 14 orbits/day = 2.1 Gigapixels/day to be processed. **Every pixel gets its own atmospheric correction.**

If atmospheric correction takes 1 second/pixel, this is 67 YEARS to process 1 DAY of collected data!

Atmospheric correction must be done in near real time. (One day to process one day's data.) This requires processing ~34,000 pixels/second (or 0.00003 sec/pixel for one computer; if have 100 computers, then 0.003 sec/pixel)

This is a severe constraint on what atmospheric correction algorithms are used. You WANT the AC algorithms to be as accurate as possible, but they MUST BE computationally FAST

satellite ocean color

Rayleigh · aerosols · aerosols+Ray · Sun glint · surface-reflected background sky radiance · white caps (foam) · water-leaving

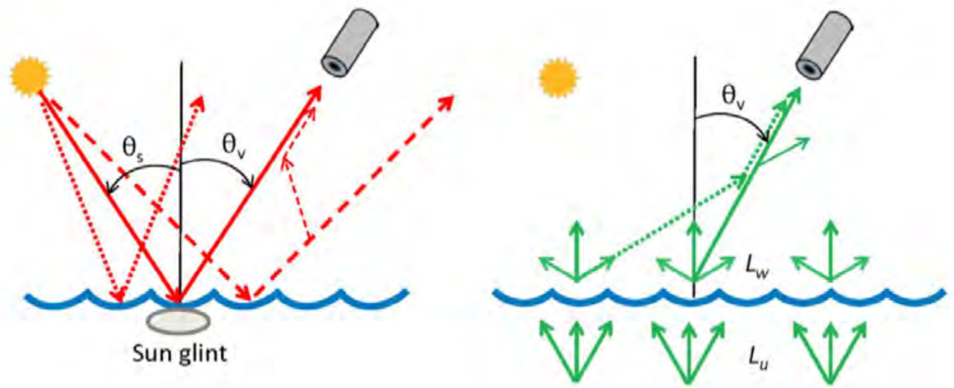
(1) $$L_{\mathrm{t}} = L_{\mathrm{R}} + [L_{\mathrm{a}} + L_{\mathrm{aR}}] + L_g + L_{\mathrm{sky}} + L_f + L_{\mathrm{w}}$$

where all radiances are defined at the top-of-atmosphere (TOA)

(2) $$L_t = L_{\mathrm{R}} + [L_a + L_{\mathrm{Ra}}] + TL_g + tL_f + tL_w$$

where $\mathbf{L_g}$, $\mathbf{L_f}$, and $\mathbf{L_w}$ are now defined at the sea surface, and $\mathbf{L_{sky}}$ is accounted for in Rayleigh correction. **T** and **t** are the direct and diffuse transmittance.

Direct: one particular path connects the source & observer



Diffuse: radiance from all locations & directions can be scattered into the direction of interest

satellite ocean color

Rayleigh | aerosols | aerosols+Ray | Sun glint | surface-reflected background sky radiance | white caps (foam) | water-leaving

(1) $$L_{\mathrm{t}} = L_{\mathrm{R}} + [L_{\mathrm{a}} + L_{\mathrm{aR}}] + L_g + L_{\mathrm{sky}} + L_f + L_{\mathrm{w}}$$

where all radiances are defined at the top-of-atmosphere (TOA)

(2) $$L_t = L_{\mathrm{R}} + [L_a + L_{\mathrm{Ra}}] + TL_g + tL_f + tL_w$$

where $\mathbf{L_g}$, $\mathbf{L_f}$, and $\mathbf{L_w}$ are now defined at the sea surface and $\mathbf{L_{sky}}$ is accounted for in Rayleigh correction. **T** and **t** are the direct and diffuse transmittance.

factor out gaseous diffuse transmissions: (3) $$L_{\mathrm{t}} = \left(L_{\mathrm{r}} + [L_{\mathrm{a}} + L_{\mathrm{ra}}] + t_{\mathrm{dv}} L_{\mathrm{wc}} + t_{\mathrm{dv}} L_{\mathrm{w}}\right) t_{\mathrm{gv}} t_{\mathrm{gs}} f_{\mathrm{p}} + TL_g$$

$\mathbf{t_{gv}}$ is the diffuse transmission by atmos. gases in the viewing direction
$\mathbf{t_{gs}}$ is the diffuse transmission by atmos. gases in the Sun's direction
$\mathbf{t_{dv}}$ is the diffuse transmission along viewing path of the sensor

This "factoring" is done for reasons of computational efficiency. Theoreticians often use (1) or (2); OBPG works with (3).

Justification

Consider the Rayleigh term: $L_{\mathrm{R}} = L_{\mathrm{r}} t_{\mathrm{gv}} t_{\mathrm{gs}} f_{\mathrm{p}}$

The TOA Rayleigh contribution L_R depends on Sun and viewing geometry, absorbing and non-absorbing atmospheric gasses, sea-level pressure, and polarization. This is a serious RT calculation, including polarization.

The L_r term is a “standard” Rayleigh contribution computed using a standard atmosphere and only non-absorbing gases N_2 and O_2, for various Sun & viewing geometries. This can be computed once and placed in a look-up table.

The gaseous transmittances are computed by use of gas absorption coefficients, computed path lengths, and gas concentrations for the various absorbing gases. Compute once and put in a LUT.

The f_p term is a polarization correction, which depends on atmosphere and surface polarization states (modeled Rayleigh and glint Stokes vectors) and the sensor-specific polarization sensitivity with viewing direction. Again, compute once for various inputs and make a LUT.

L_r, the diffuse transmittances, and f_p are pre-computed and stored in look-up tables as functions of Sun and viewing geometry, gas concentrations, instrument polarization sensitivity, etc. **Evaluation of L_R then requires no real-time radiative transfer calculations.**

Ditto for the other terms in the L_t equation

Keep in Mind What We Want: R_{rs}

ocean color satellites measure top-of-atmosphere radiances

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p \; + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(q_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

desired

we desire (normalized)
remote sensing reflectances

Now sequentially step through the meaning & evaluation of each term in these equations

not known

measured

calculated

$$L_t = (L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w) t_{gv} t_{gs} f_p + T L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(q_s) t_{ds} f_s f_b f_l}$$

extraterrestrial solar constant (irradiance)

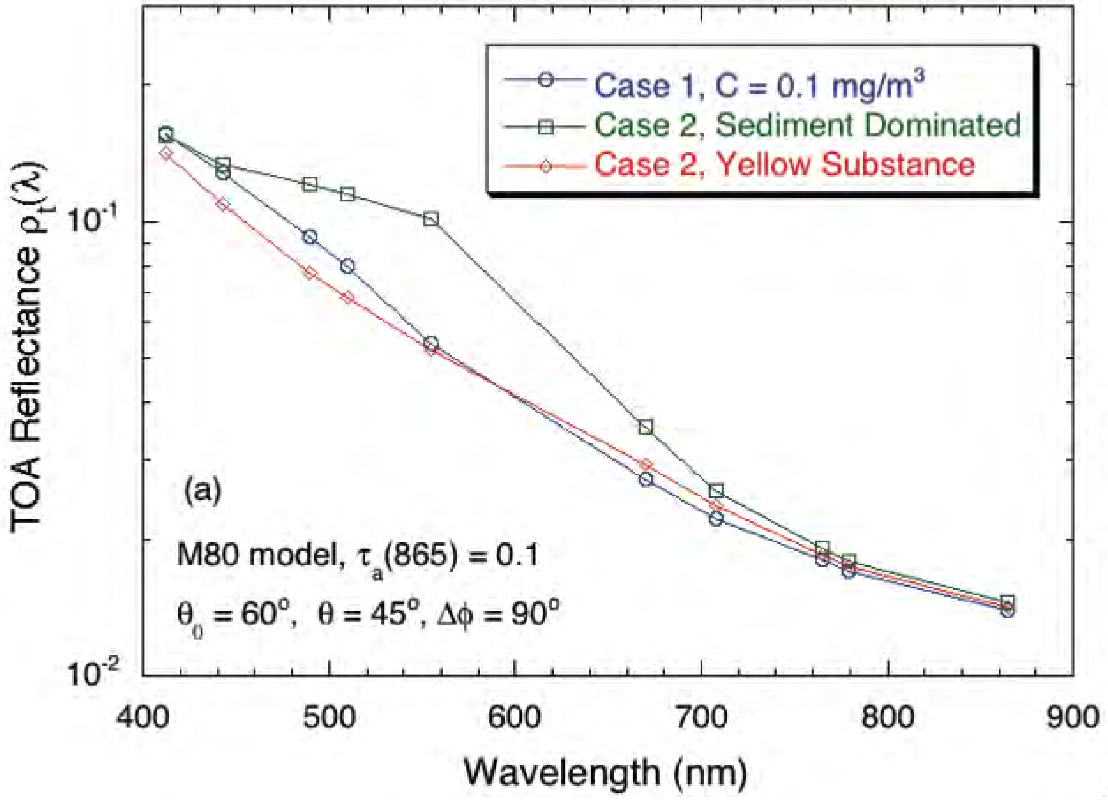
cosine of the solar zenith angle

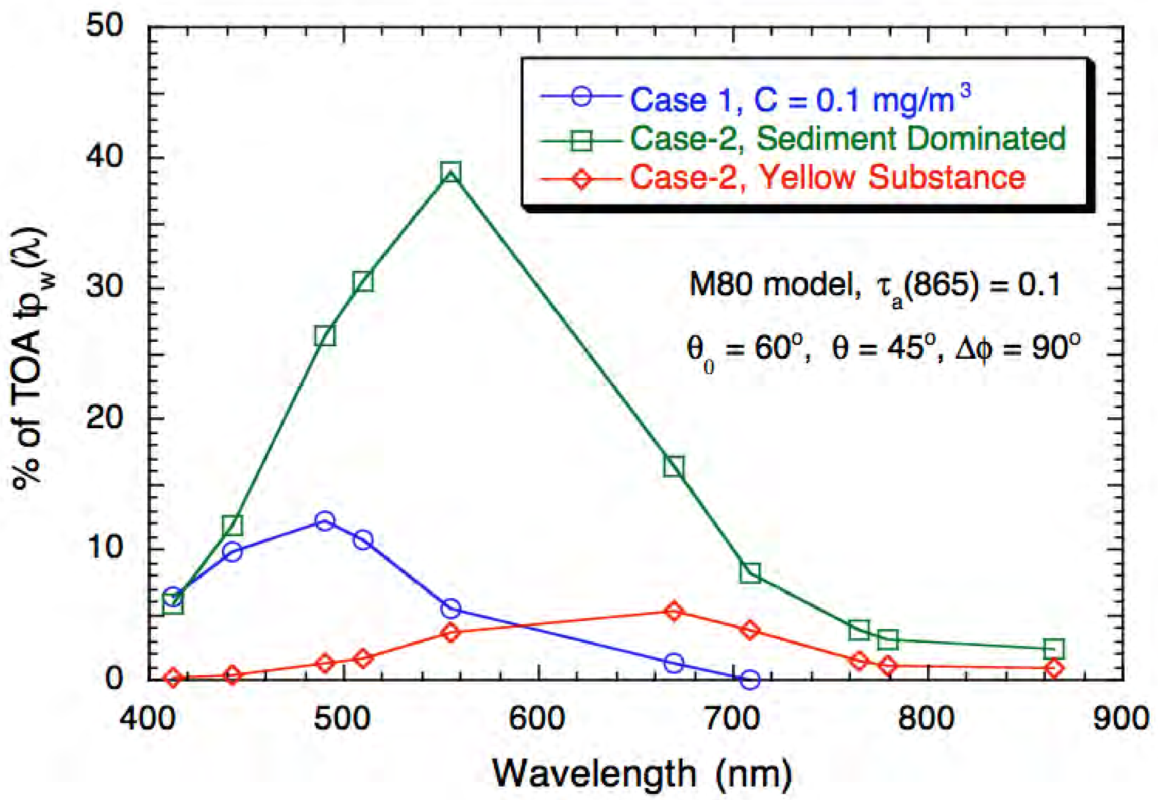
adjustment for the Earth-Sun distance

top-of-atmosphere radiance

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

Menghua Wang, IOCCG Report 10

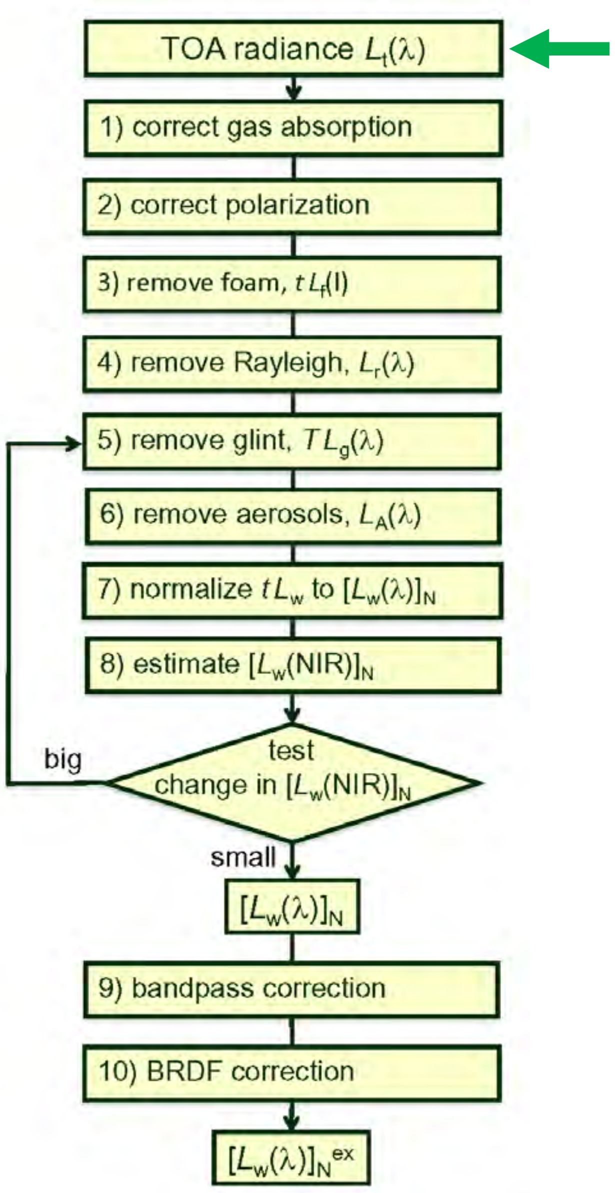


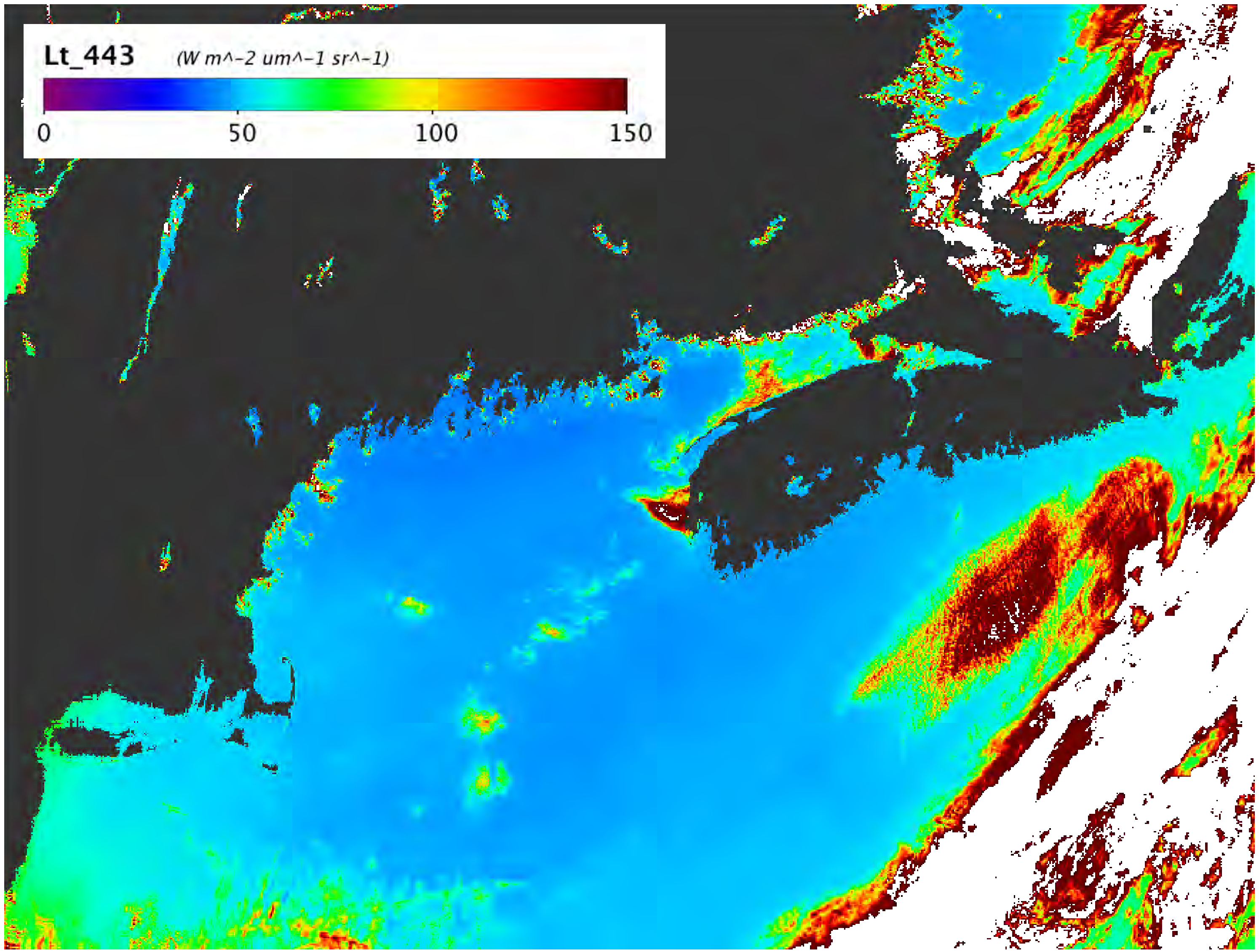


L_w is often <10% of L_t !

0.5% error in atmospheric correction or calibration → 5% error in L_w

processing cadence: L_t





transmittances

diffuse transmittance of gases (g)
in direction of Sun (s) or satellite (v)

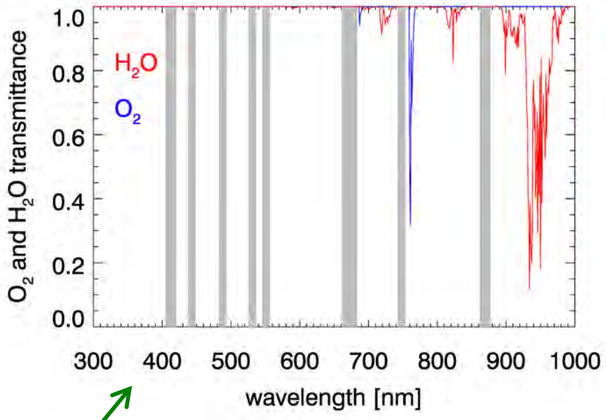
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv} t_{gs} f_p + T L_g$$

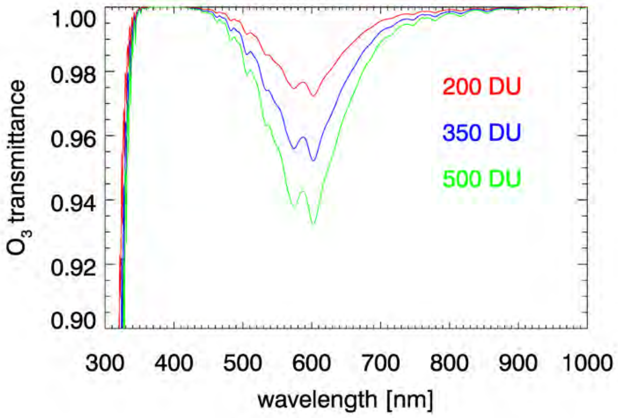
$$R_{rs} = \frac{L_w}{F_0 \cos(q_s) t_{ds} f_s f_b f_l}$$

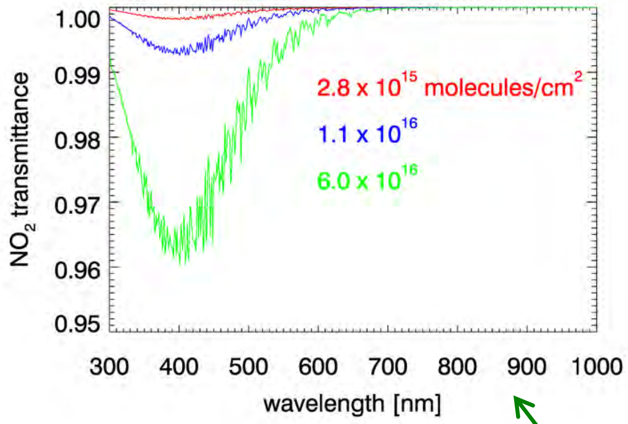
Rayleigh / aerosol diffuse
transmittance (d) in direction
of Sun (s) or satellite (v)

gaseous transmittances

nitrogen dioxide, ozone, oxygen, & water vapor all attenuate sunlight

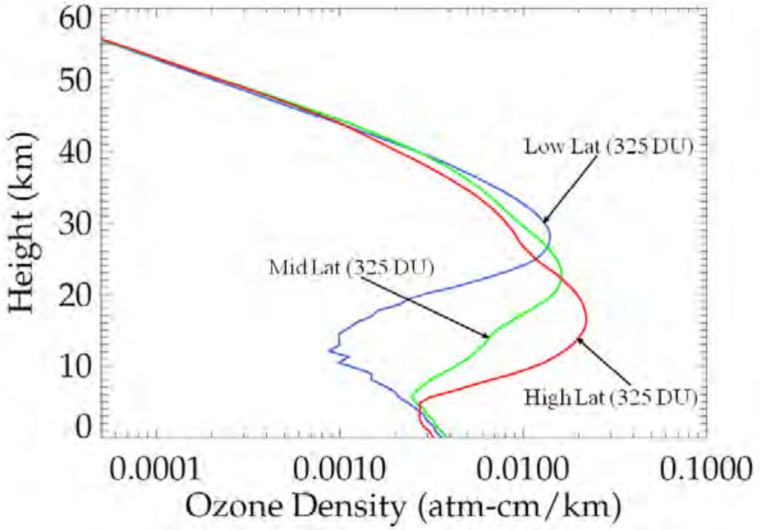


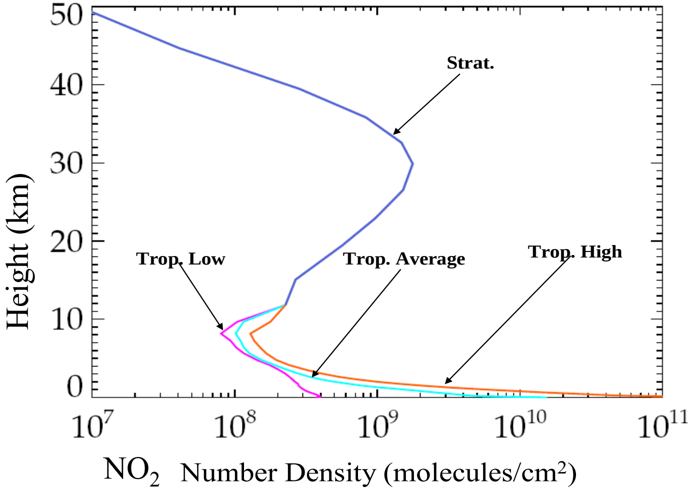




O_3 optically thin & high in atmosphere, but NO_2 dense & near the surface

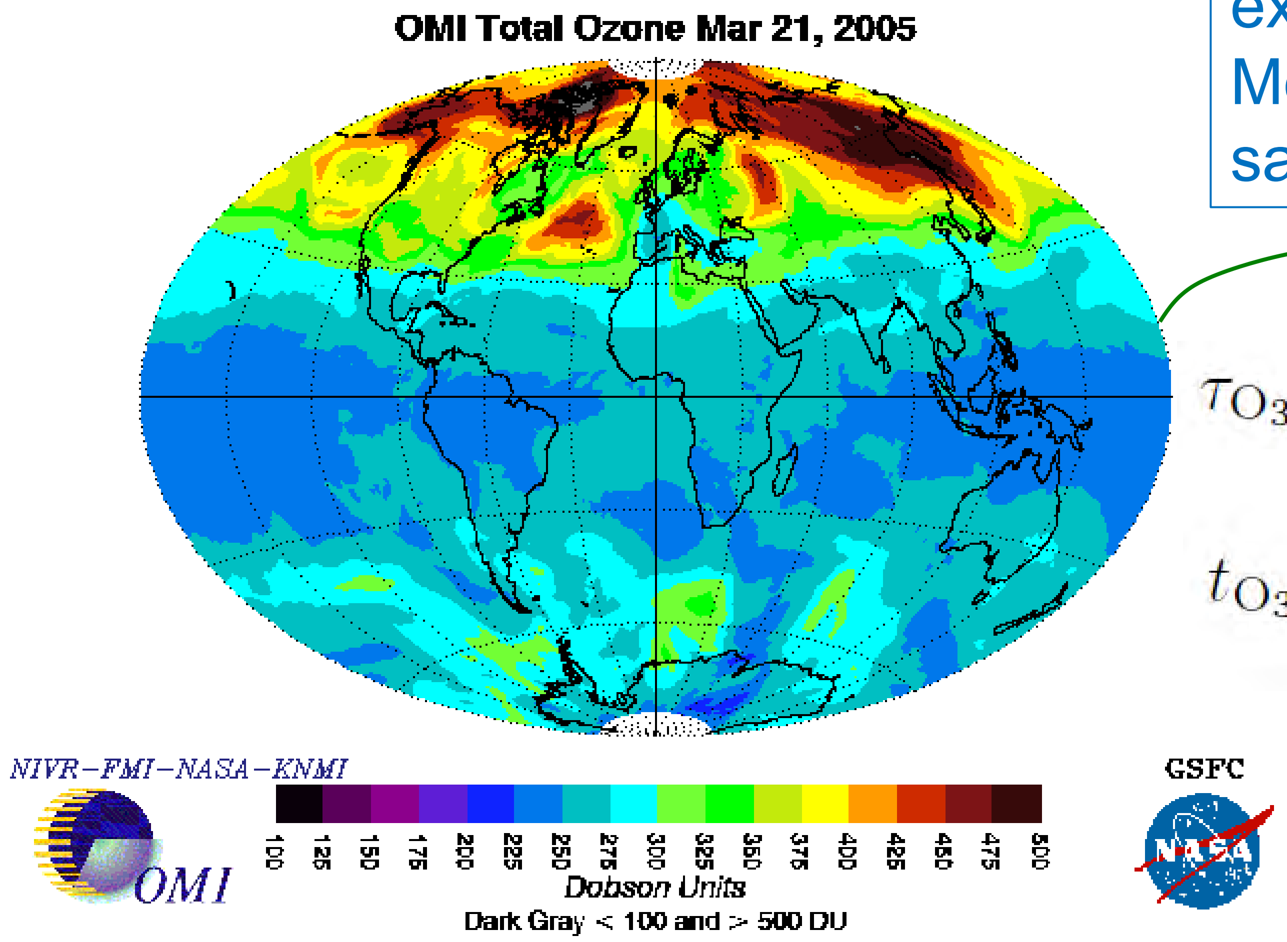
can avoid H_2O and O_2 absorption by choice of band locations





cannot avoid O_3 and NO_2 absorption because they are broadband

calculating gaseous transmittance requires ancillary data



1 DU = 0.01 mm of O3 at STP

example using OMI (Ozone Monitoring Instrument on Aura satellite) ozone measurements

from LUT

$$\tau_{O_3}(\lambda) = [O_3]k_{O_3}(\lambda)$$ for nadir direction

$$t_{O_3} = \exp\left[-\tau_{O_3}\left(\frac{1}{\cos\theta_s} + \frac{1}{\cos\theta_v}\right)\right]$$

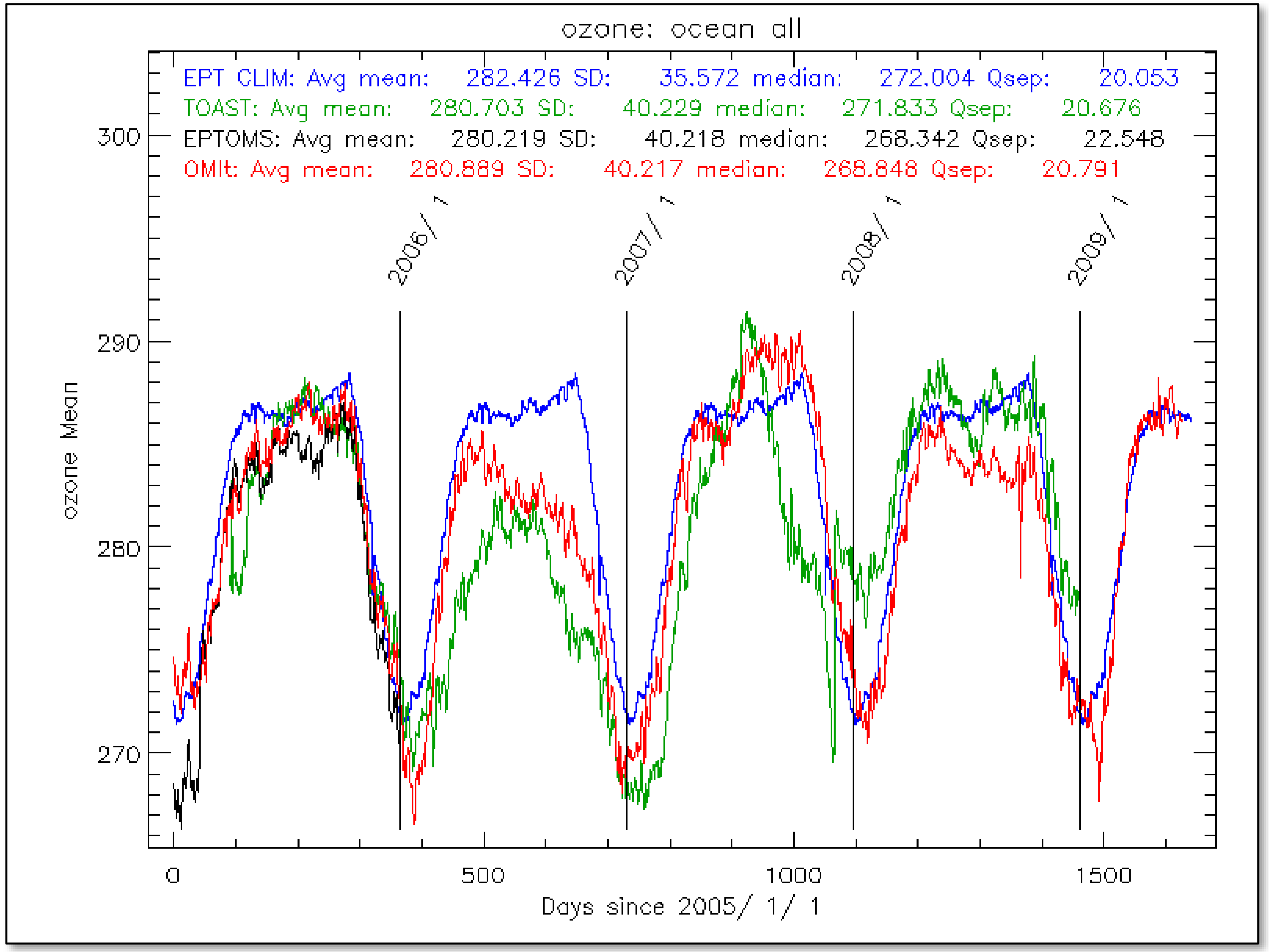
most transmittances are calculated in this manner; see the documentation for the details

all ancillary data are not created equal

comparison of three ancillary sources of O_3:

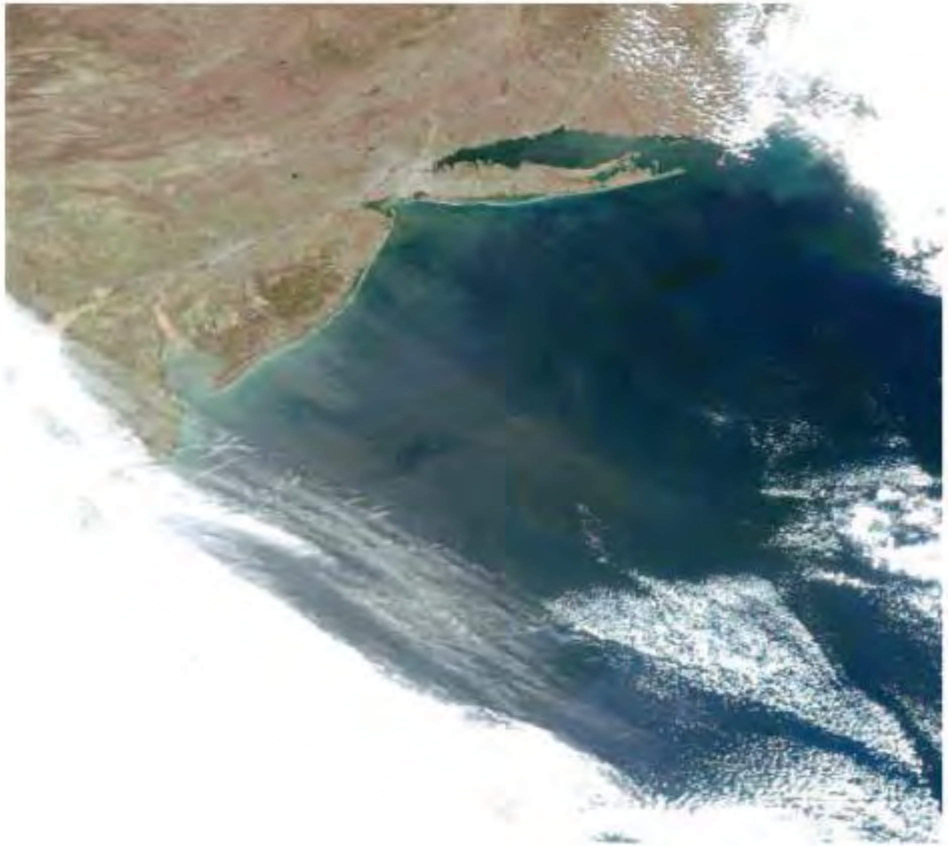
TOAST
OMI
EPTOMS and climatology

small differences in ancillary data can lead to big differences in geophysical products

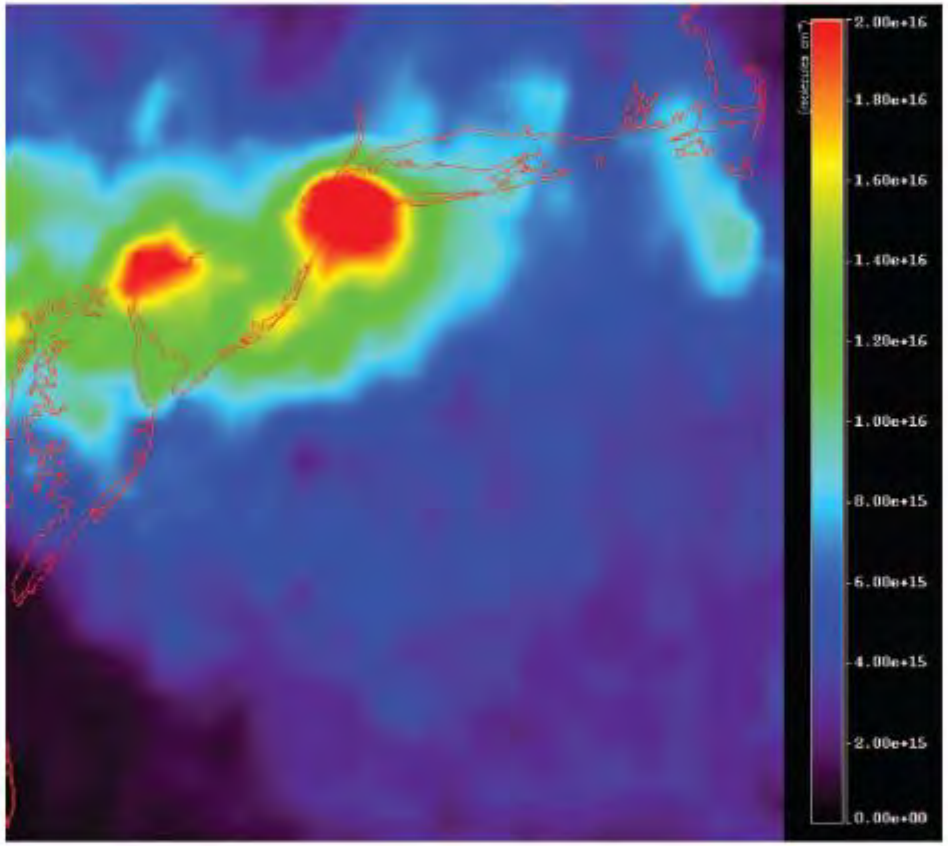


TOAST: Total Ozone from Analysis of Stratospheric and Tropospheric components
OMI: Ozone Mapping Instrument
EPTOMS: Earth-Probe Total Ozone Monitoring Spectrometer
...and there are others

No surprise: Lots of NO_2 near NY City and Philadelphia



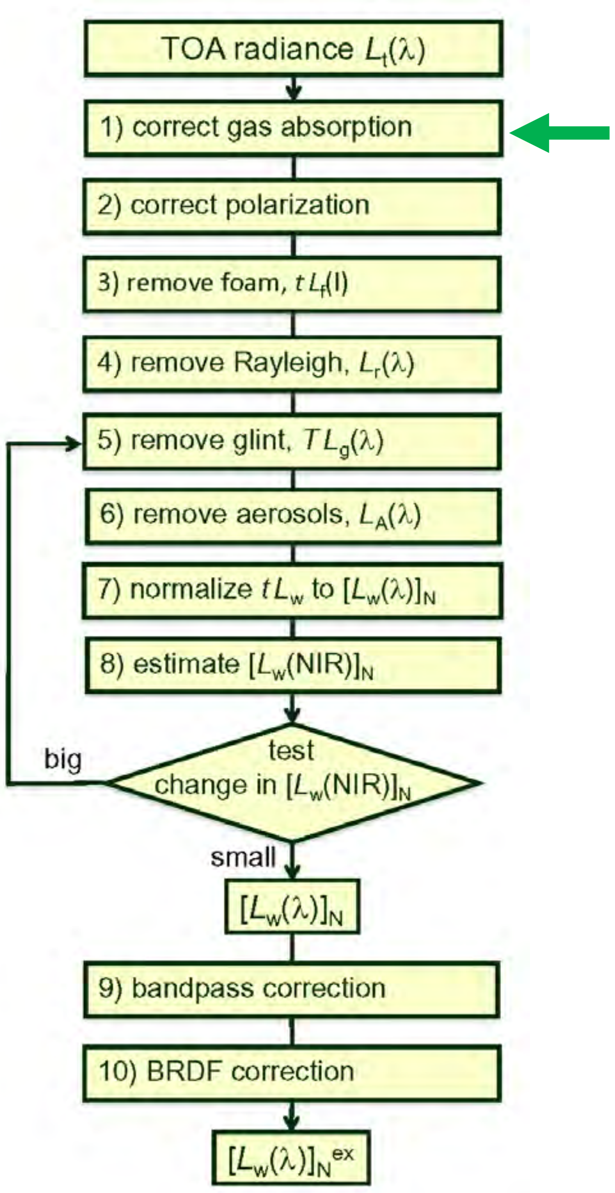
RGB image of a MODIS-Aqua scene from 11 April 2005

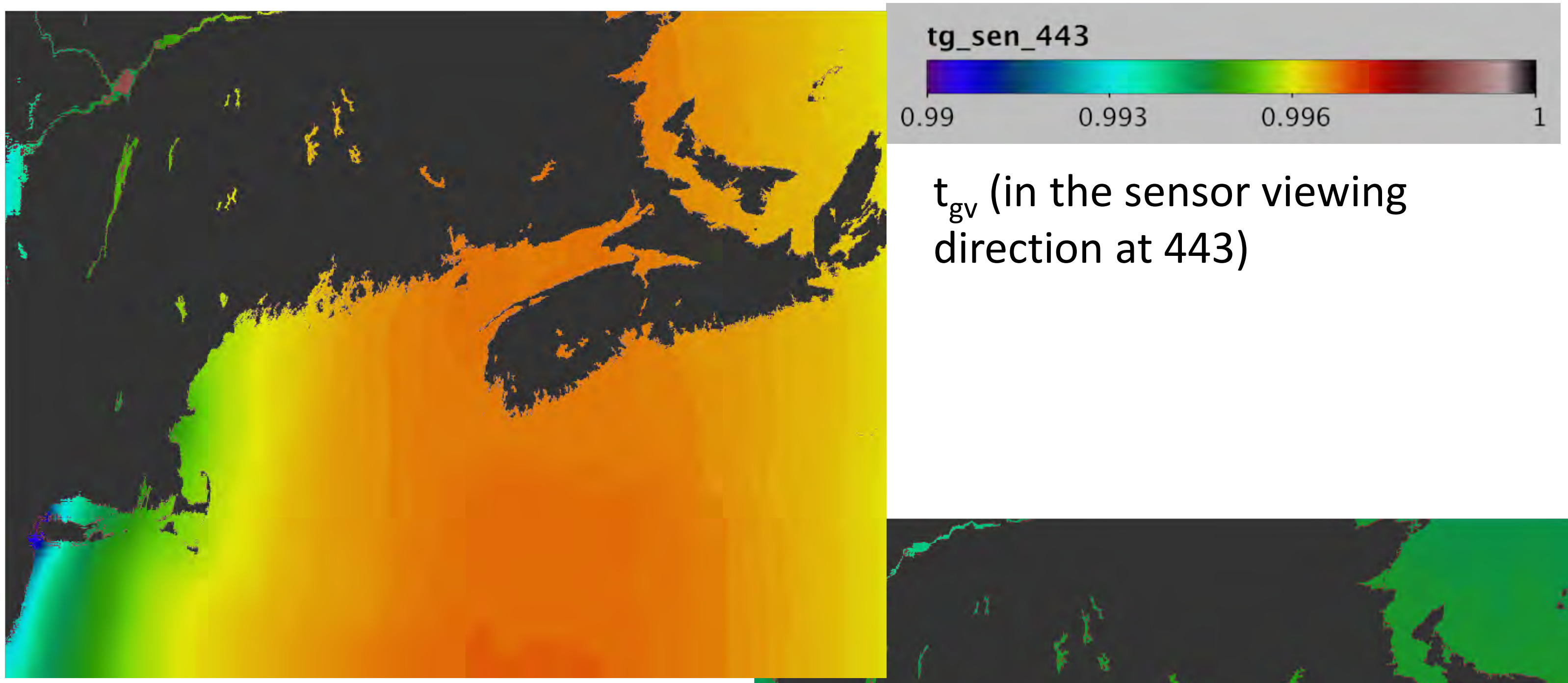


OMI tropospheric NO2 amount on 11 April 2005

From Ahmad et al. (2007)

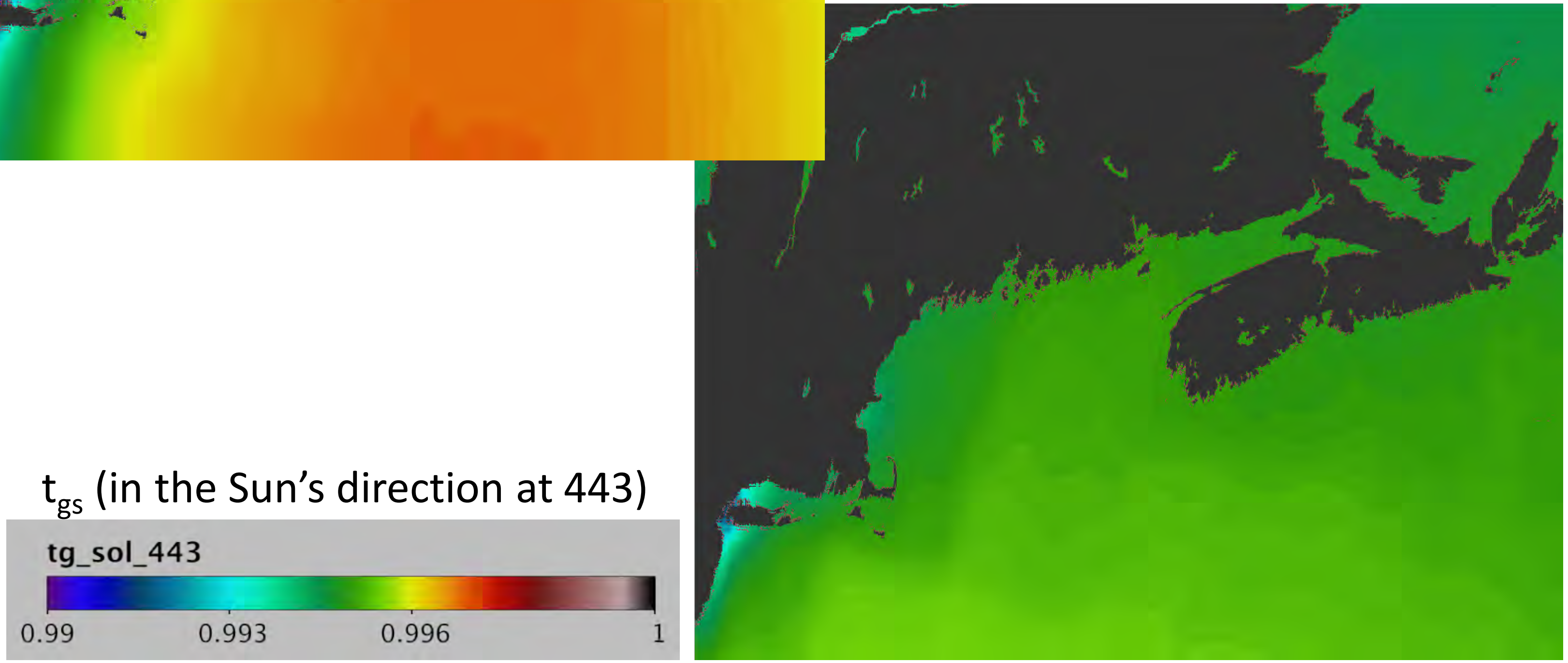
processing cadence: L_t / t_{gv} / t_{gs}





t_{gv} (in the sensor viewing direction at 443)

t_{gs} (in the Sun's direction at 443)



instrument polarization sensitivity

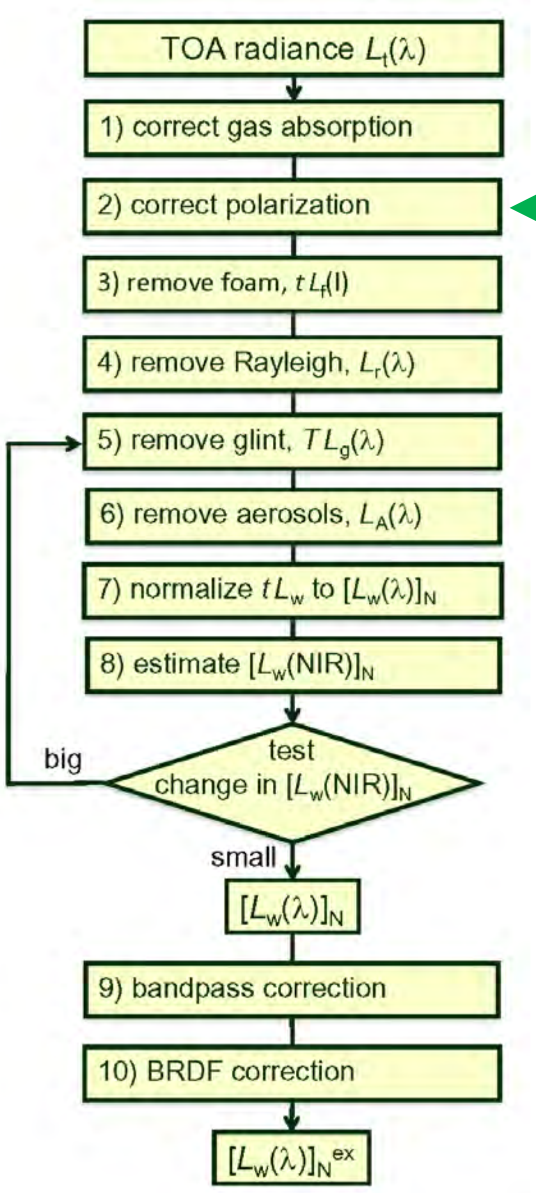
instrument polarization correction factor (pre-launch measurement)

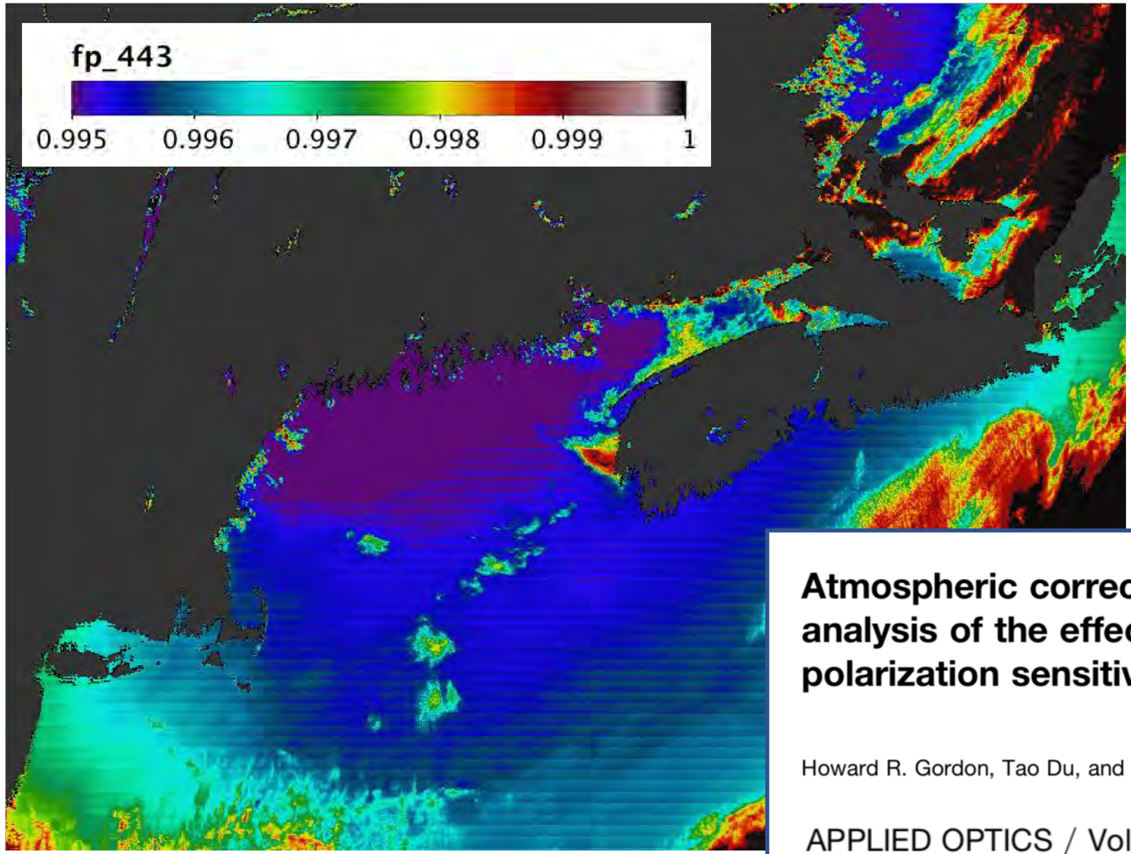
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(\theta_s)\, t_{ds}\, f_s\, f_b\, f_\lambda}$$

MODIS requires an additional correction

processing cadence: L_t / t_{gv} / t_{gs} / f_p





some satellite instruments include depolarizers in their fore optics, which mitigates instrument polarization sensitivity

Atmospheric correction of ocean color sensors: analysis of the effects of residual instrument polarization sensitivity

Howard R. Gordon, Tao Du, and Tianming Zhang

APPLIED OPTICS / Vol. 36, No. 27 / 20 September 1997

Moderate-Resolution Imaging Spectroradiometer ocean color polarization correction

Gerhard Meister, Ewa J. Kwiatkowska, Bryan A. Franz, Frederick S. Patt, Gene C. Feldman, and Charles R. McClain

APPLIED OPTICS / Vol. 44, No. 26 / 10 September 2005

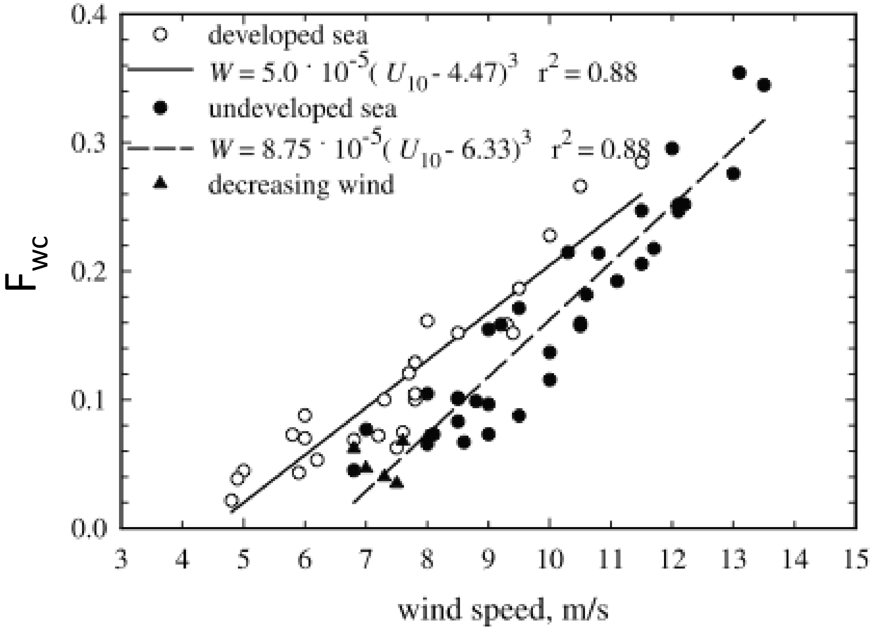
foam & whitecaps

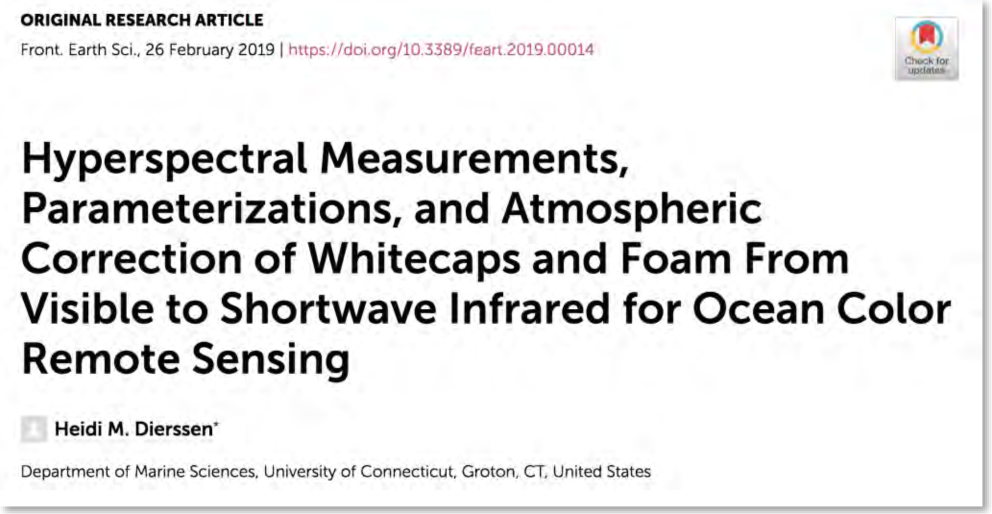
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(q_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

foam & whitecaps

Stramska & Petelski, JGR, 2003





ORIGINAL RESEARCH ARTICLE

Front. Earth Sci., 26 February 2019 | https://doi.org/10.3389/feart.2019.00014

Hyperspectral Measurements, Parameterizations, and Atmospheric Correction of Whitecaps and Foam From Visible to Shortwave Infrared for Ocean Color Remote Sensing

Heidi M. Dierssen*

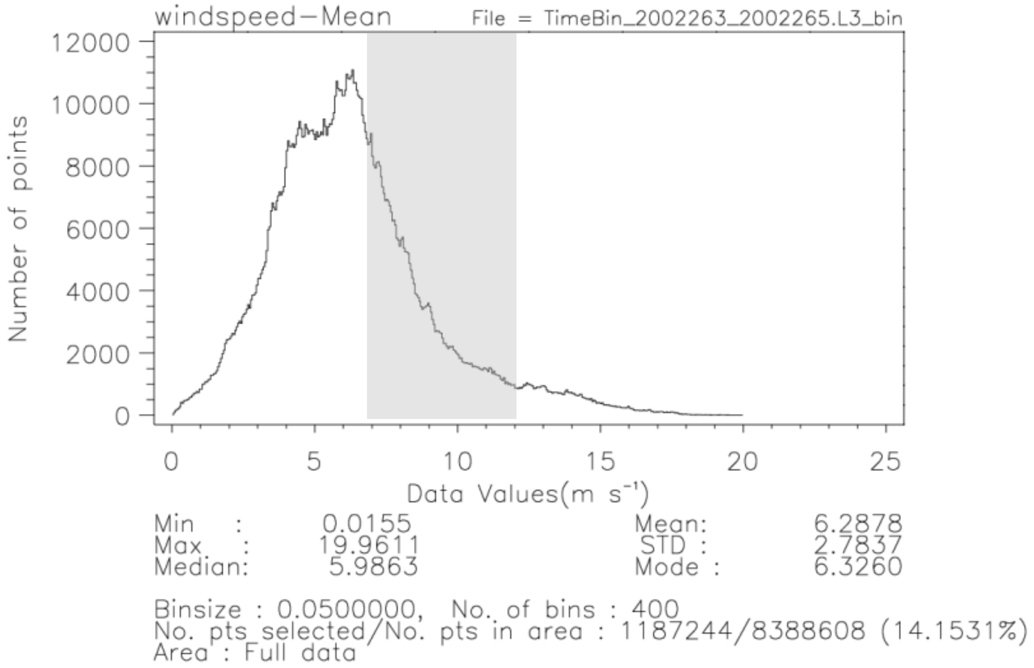
Department of Marine Sciences, University of Connecticut, Groton, CT, United States

$\rho_f = \pi L_f = A F_{wc}$

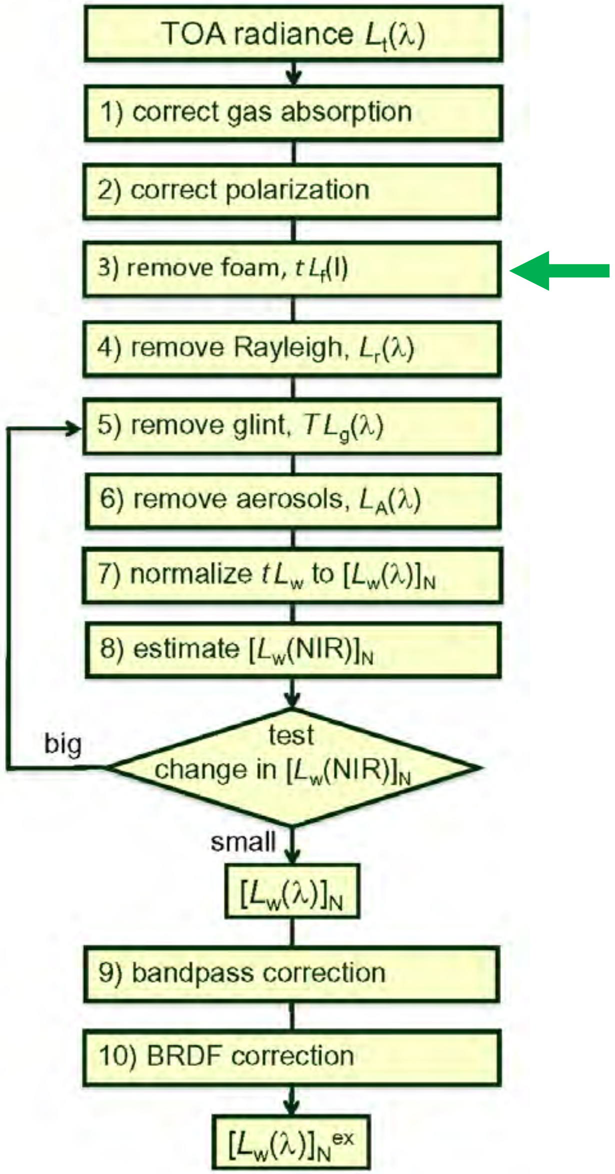
A = 22% (11-33%) from Koepke 1984
+ a correction for decreasing reflectance in red & NIR

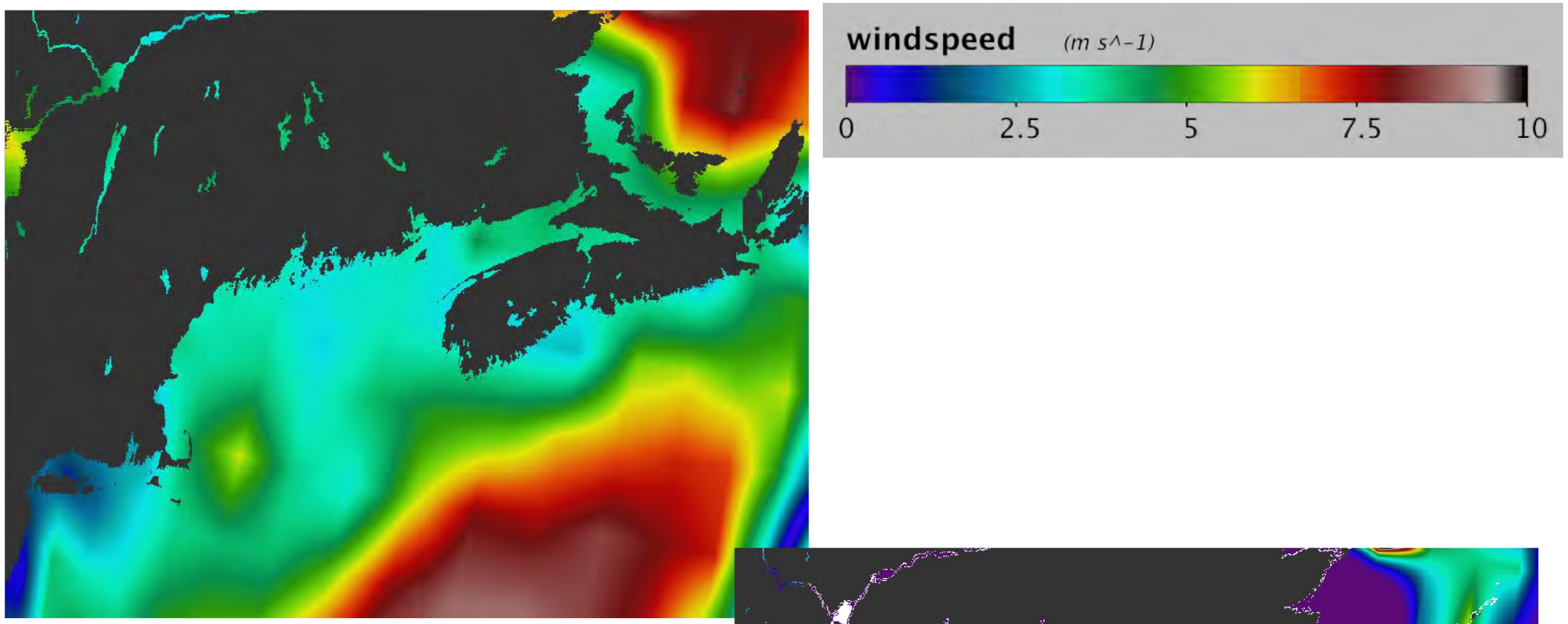
$\rho_f(412) = \pi L_f(412) = 1.925 \times 10^{-5} (U_{10} - 6.33)^3$

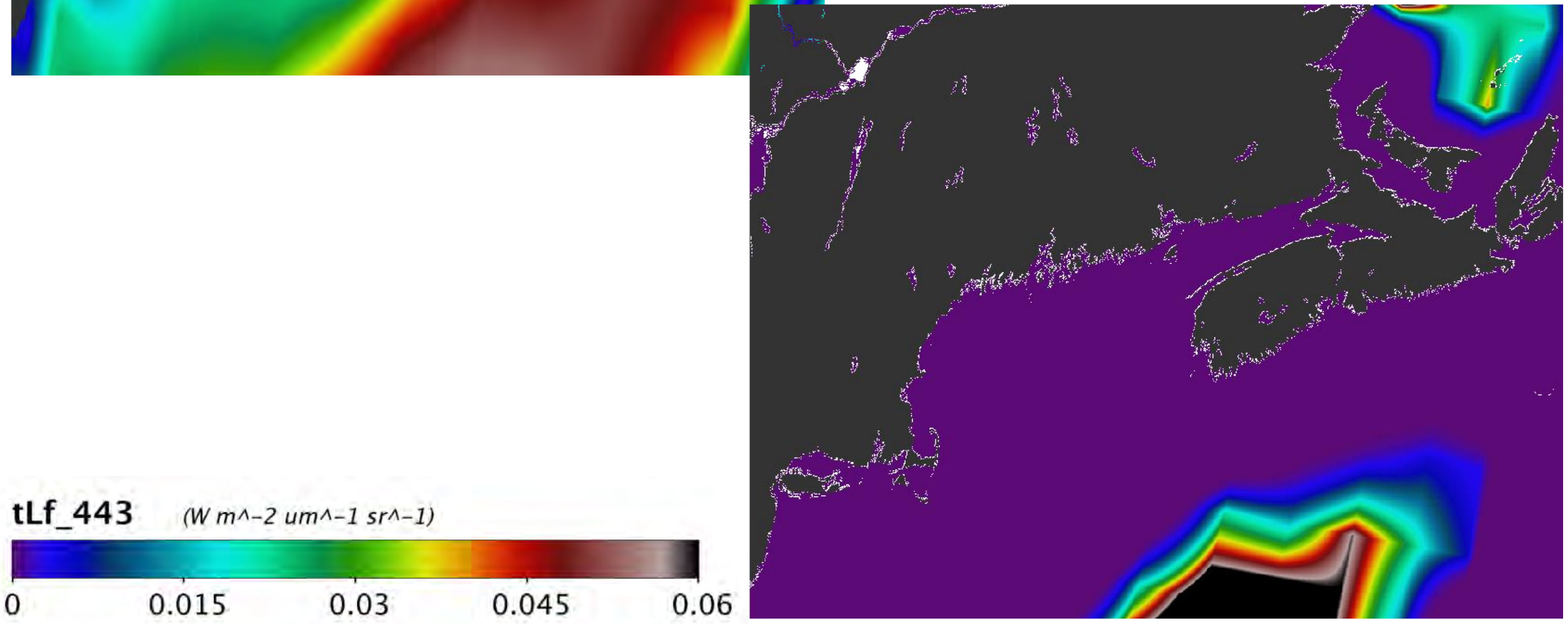
estimation of contribution of whitecaps & foam requires ancillary wind data (NCEP)



processing cadence: L_t / t_{gv} / t_{gs} / f_p - tL_f







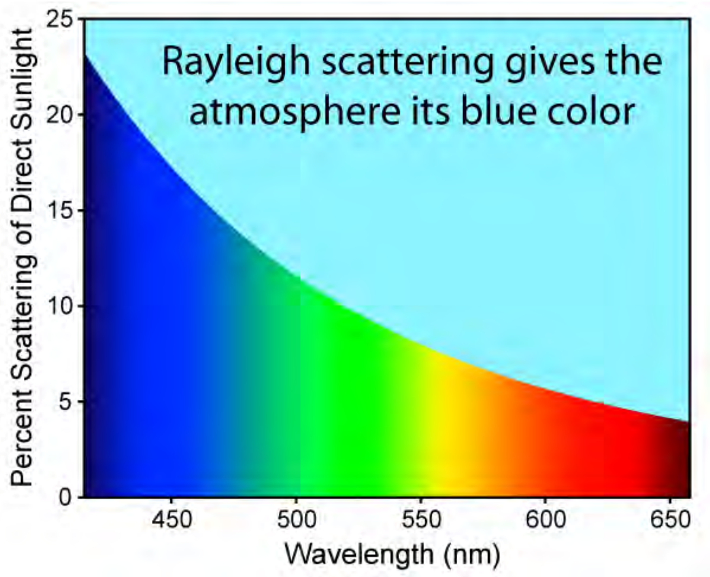
molecular (Rayleigh) scattering

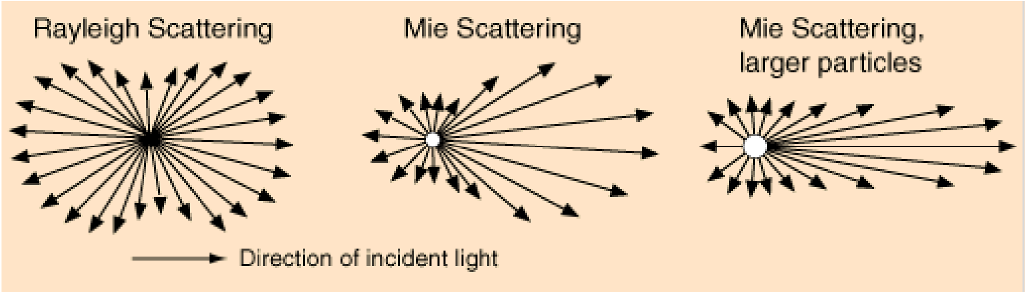
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(q_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

molecular (Rayleigh) scattering

- elastic scattering of electromagnetic radiation by particles much smaller than the wavelength of light (atoms or molecules)
- Rayleigh scattering of sunlight in atmosphere causes diffuse sky radiation – why the sky is blue and the Sun is yellow





http://hyperphysics.phy-astr.gsu.edu/hbase/atmos/blusky.html

- scattering phase function is symmetrical – equal forward & backward

Once more…

$$L_{\mathrm{R}} = L_{\mathrm{r}} t_{\mathrm{gv}} t_{\mathrm{gs}} f_{\mathrm{p}}$$

- Rayleigh TOA contribution (L_R) is factored into the product of a Rayleigh term (L_r), diffuse transmittances, & a polarization correction factor
- develop & maintain one **look-up table** (LUT) for L_r, computed using a standard atmosphere & the non-absorbing gases N_2 and O_2 for various Sun & viewing geometries
- transmittances and f_p calculated separately
- **computational efficiency**: without doing this, the Rayleigh LUT would become a function of other gases affecting transmission (ozone, NO_2, & water vapor), making it too large for efficient operational use

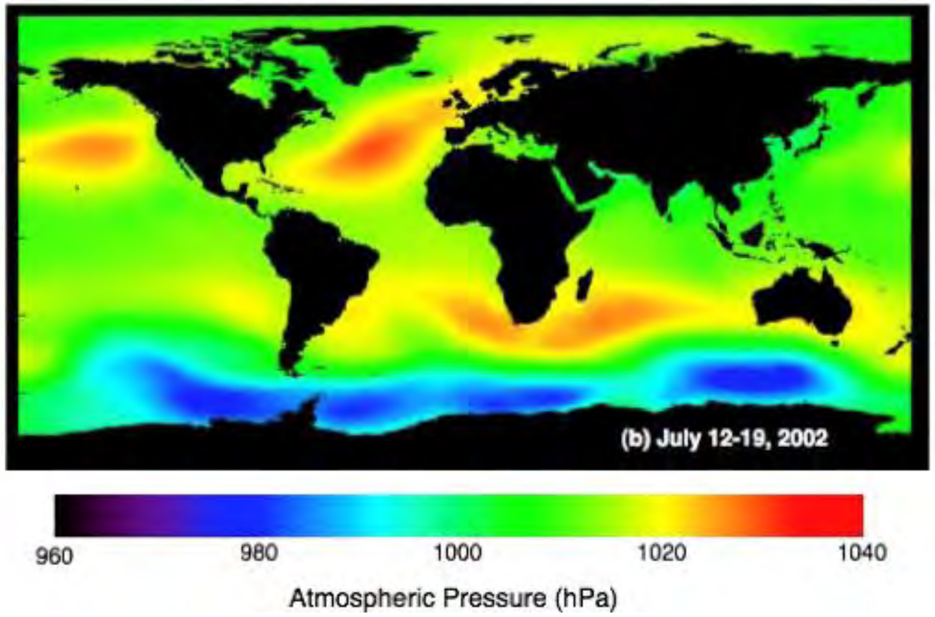
molecular (Rayleigh) scattering

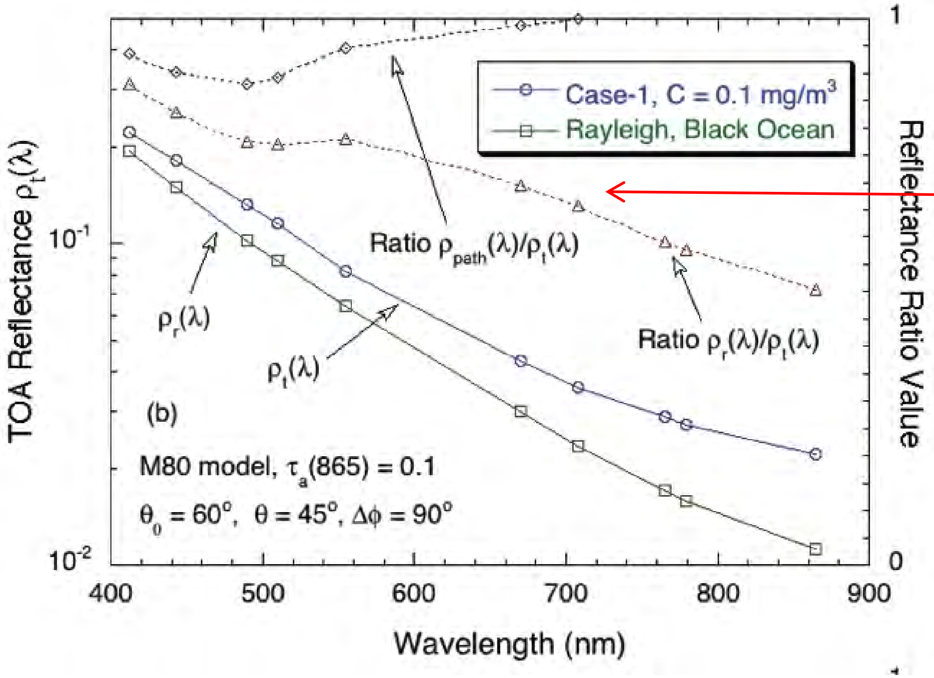
Rayleigh optical properties are calculable (to ~0.2%) – made challenging by a rough, reflective ocean (versus a flat, black ocean)

Rayleigh radiances (with polarization) are retrieved from look up tables given:

- solar & satellite viewing geometries
- wind speed (a proxy for surface roughness (influences L_{sky}))
- atmospheric pressure (∝ # gas molecules, adjusts Rayleigh optical thickness: $\tau_R(P,\lambda) = \frac{P}{P_o}\tau_{Ro}(P_o,\lambda)$

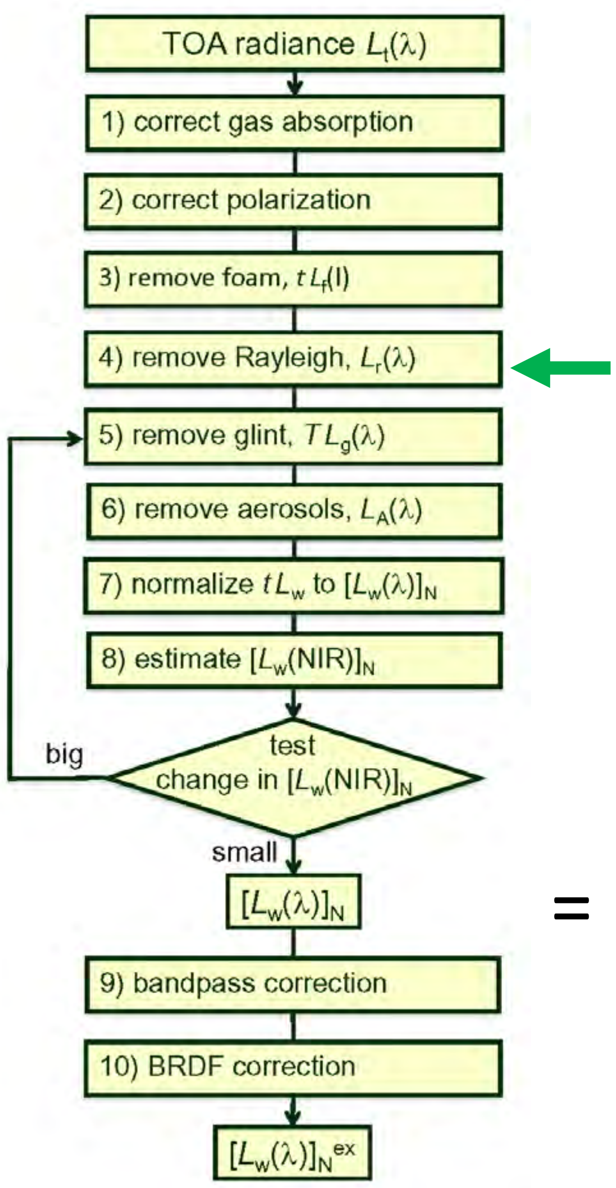
Menghua Wang, IOCCG Report 10

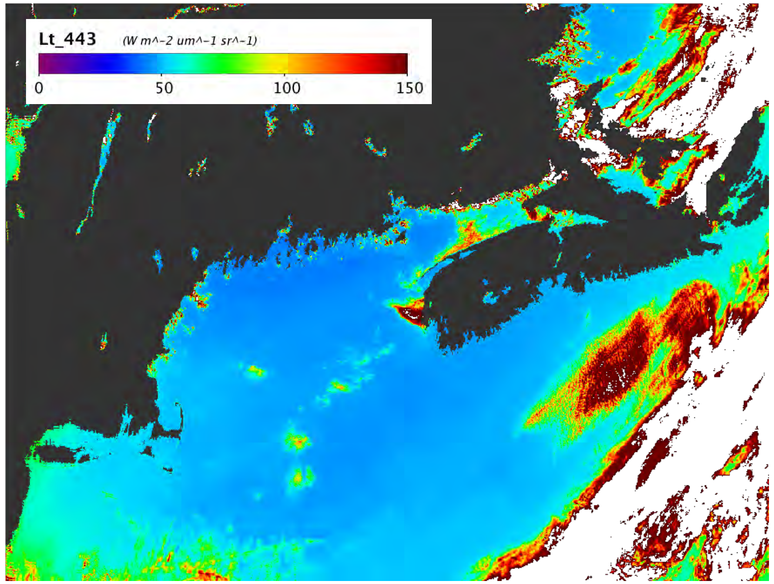




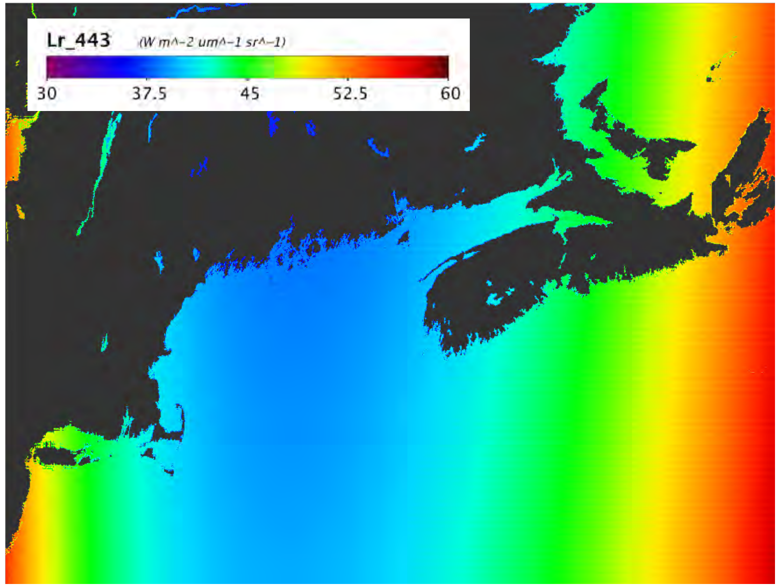
L_r can be 50-90% of L_t

processing cadence: L_t / t_{gv} / t_{gs} / f_p – tL_f - L_r

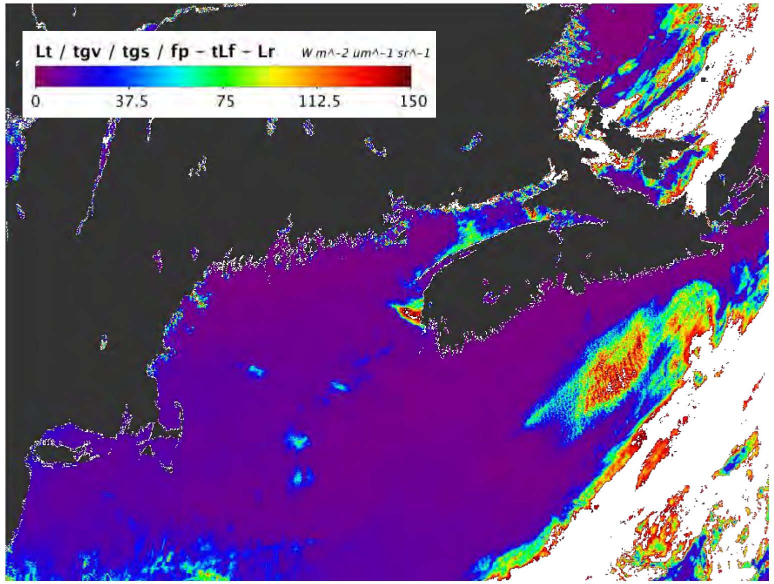




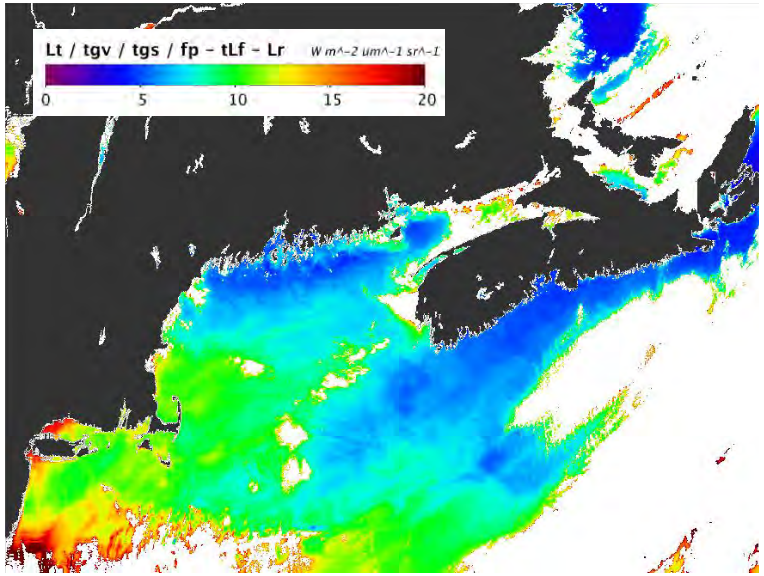
-



=



=



Sun glint

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

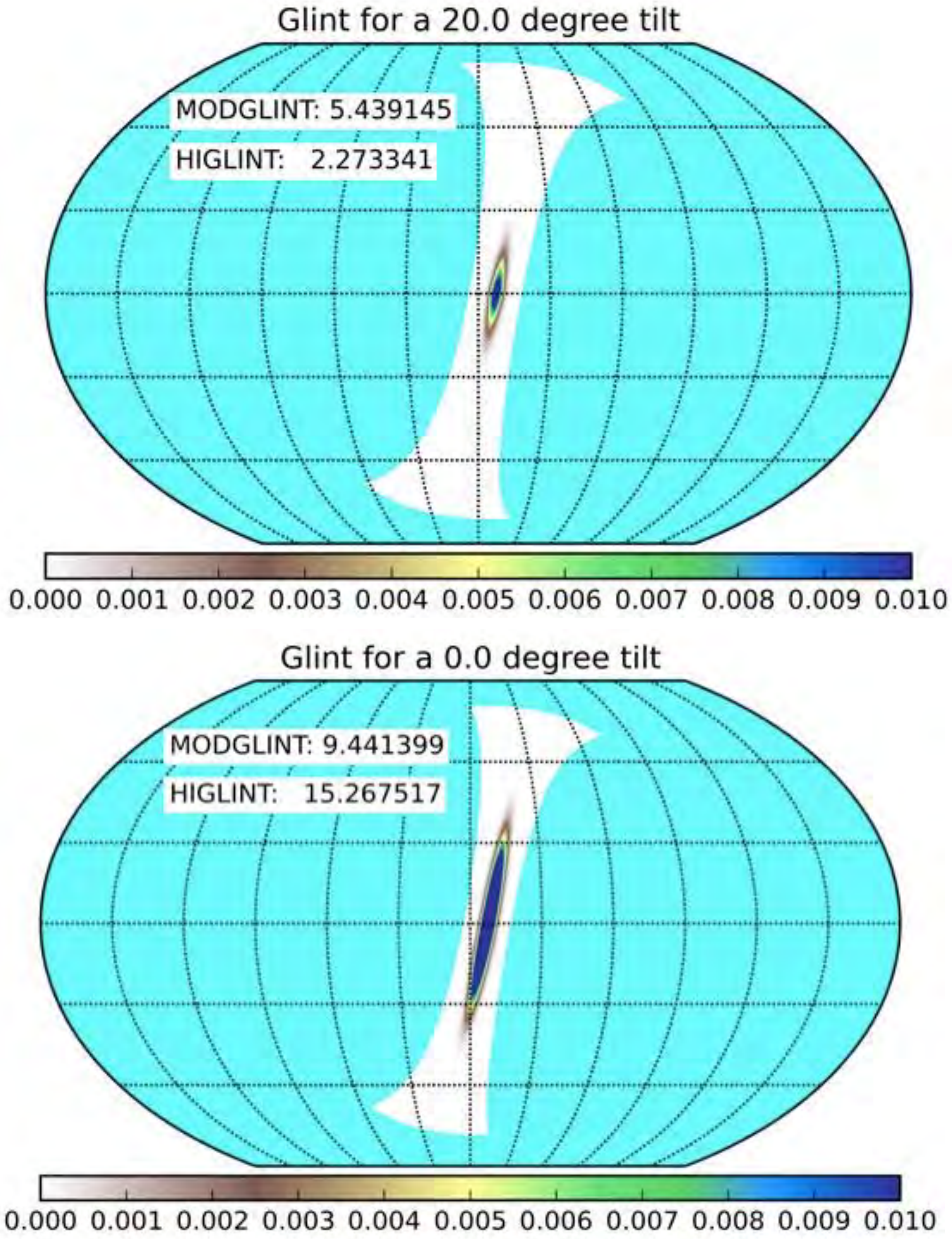
$$R_{rs} = \frac{L_w}{F_0 \cos(q_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

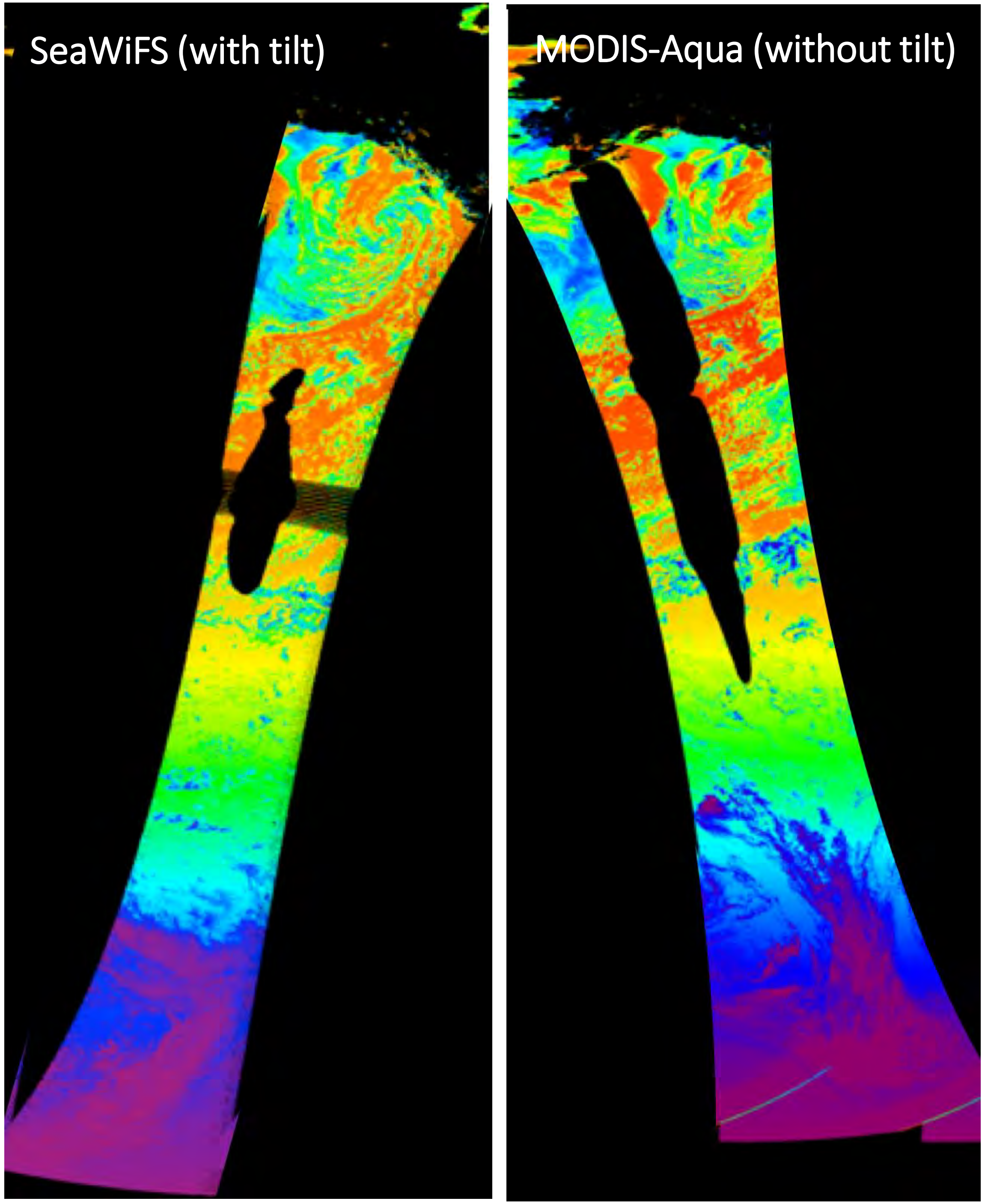
we cannot see “ocean color” through Sun glint

Courtesy NASA Earth Observatory

Ground to Space: A Glittering Path of San Francisco Sunglint

https://earthobservatory.nasa.gov/blogs/earthmatters/2016/11/09/ground-to-space-a-glittering-path-of-san-francisco-sunglint/





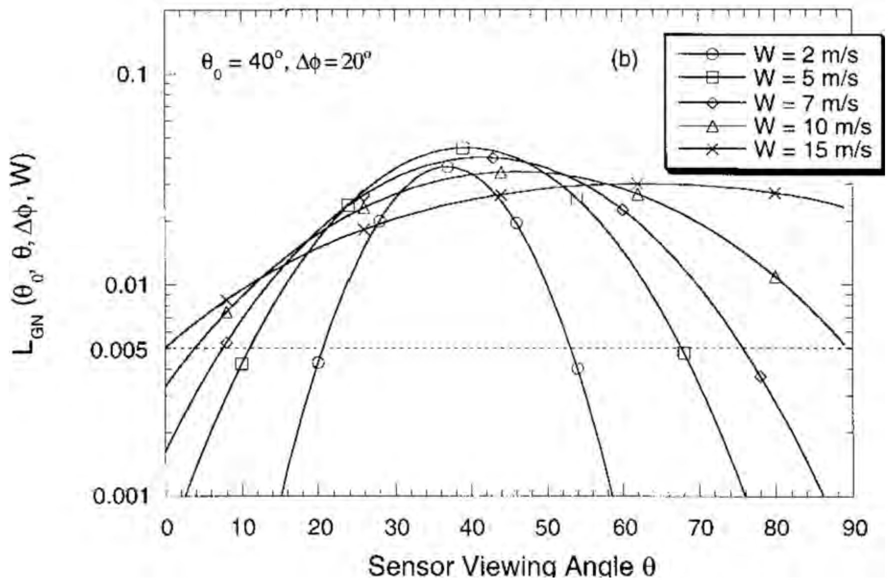
PAR = Photosynthetically Available Radiation (Einstein m^{-2} d^{-1})

Sun glint

Sun glint can be further expressed as:

$$T\,L_g = F_0\,T_0\,T\,L_{GN}$$

L_{GN} is glint radiance normalized to no atmosphere & $F_0 = 1$



L_{GN} from Cox and Munk (1954) requires ancillary wind speed & geometries of Sun & sensor

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

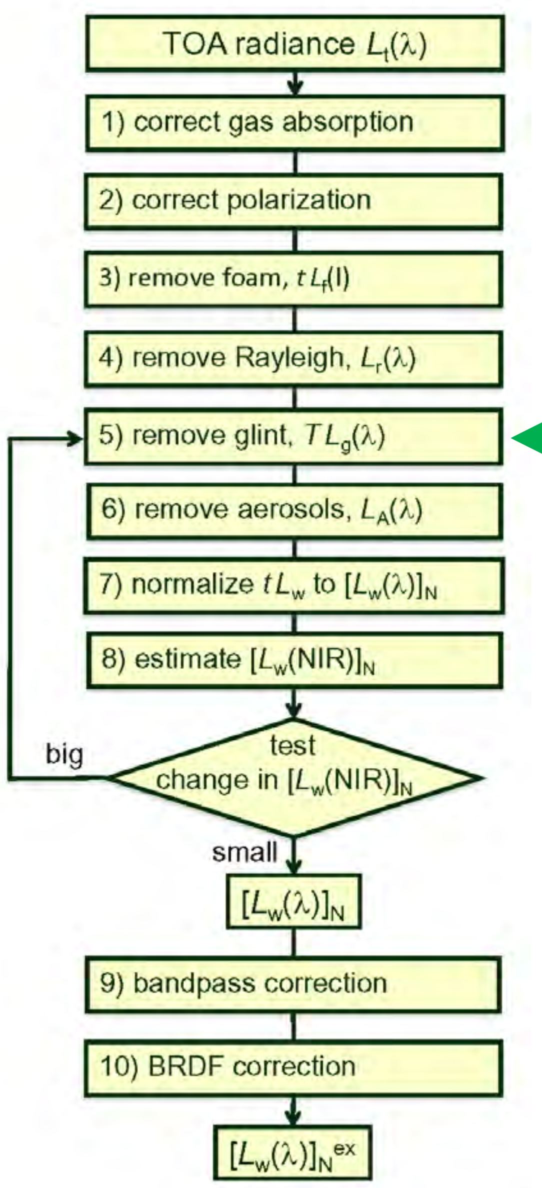
two step iteration since we don't know τ_a:

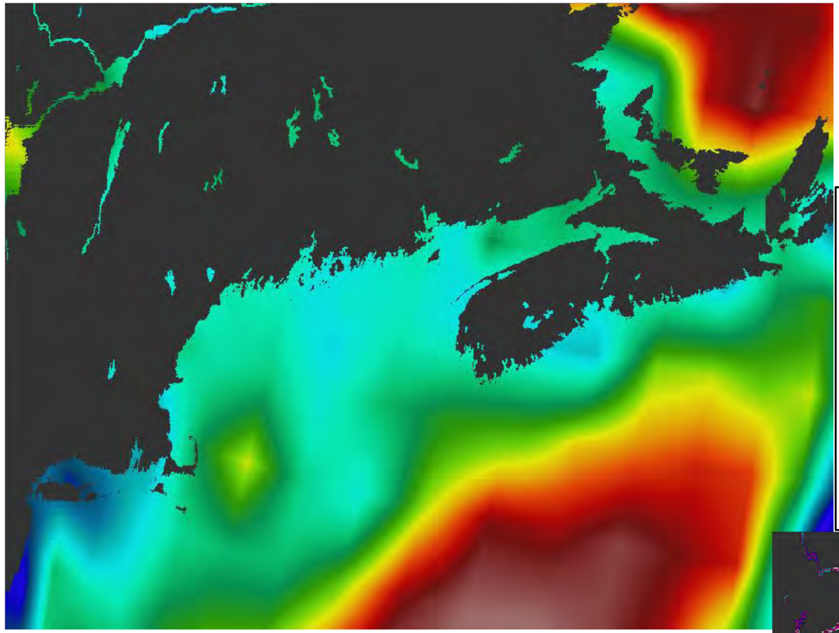
(1) [L_t, τ_a', W] ➔ $L_t^{(1)} = L_t - L_g$ ➔ $\tau_a^{(1)}$

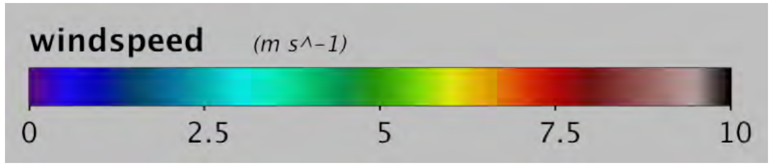
(2) [$L_t^{(1)}$, $\tau_a^{(1)}$, W] ➔ $L_t^{(2)} = L_t^{(1)} - L_g$ ➔ $\tau_a^{(2)}$

with initial guess of $\tau_a' \sim 0.1$ (additional logic included to prevent overcorrection)

processing cadence: L_t / t_{gv} / t_{gs} / f_p – tL_f - L_r - TL_g

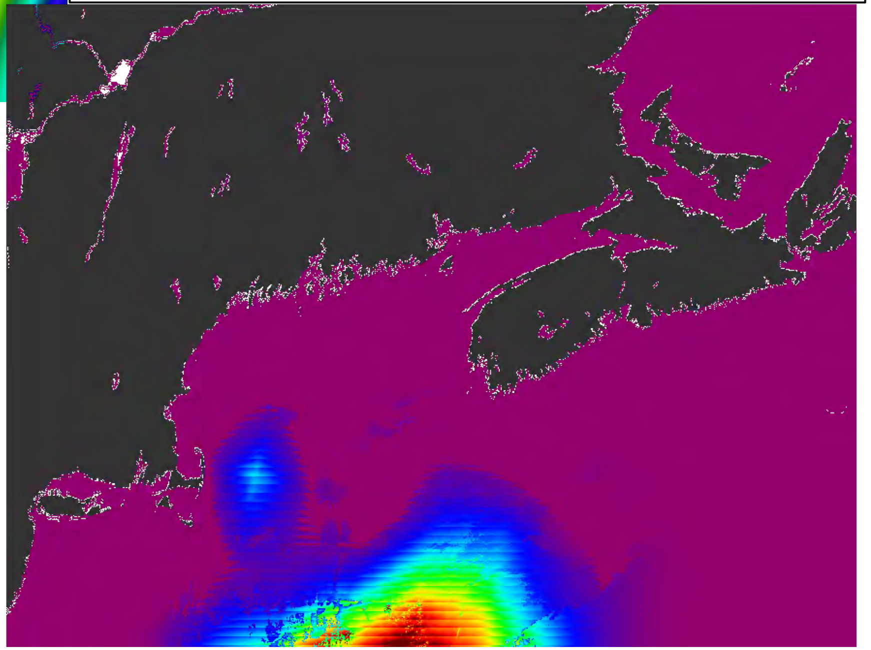


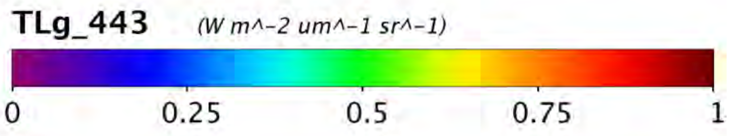




when L_{GN} > 0.005 sr^{-1}
mask the pixel as HIGH GLINT

when L_{GN} ≤ 0.005 sr^{-1}
remove TL_g from L_t





aerosols: the hard part

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(\theta_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

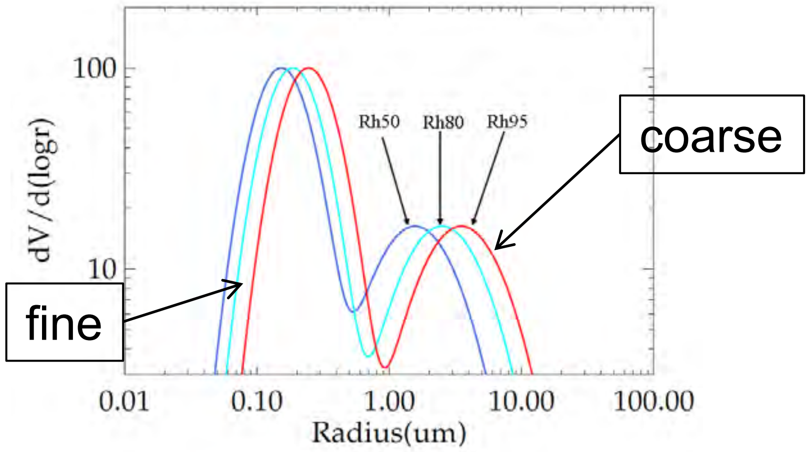
aerosol tables

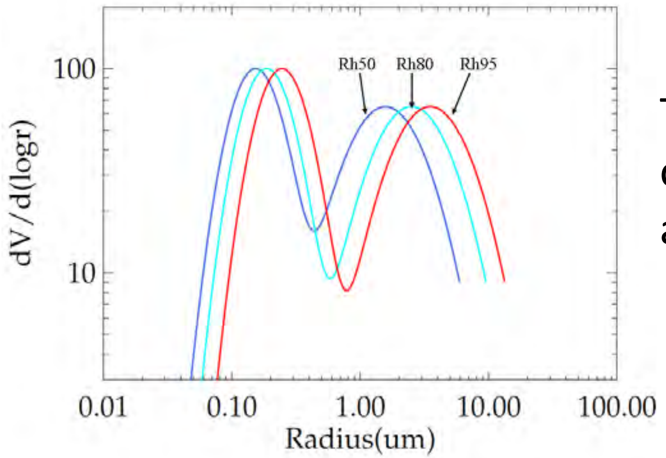
- aerosol properties can be characterized by their particle size distribution (PSD) & their complex index of refraction (m)
- given a PSD & m (& assuming sphericity), aerosol optical properties can be computed using Mie theory:
    - scattering phase function ($\tilde{b}$)
    - single scattering albedo (ω = b / c)
    - extinction coefficient (c = a + b)
- aerosol optical thickness relates to extinction coefficient
    - $\tau_a = \int_0^{TOA} c(z)\, dz$
- aerosol tables are generated for various PSDs (& m's) & are
    - defined by $\tilde{b}$, ω, τ_a (& other variables)
    - navigated using solar & satellite viewing geometries

aerosol tables

- we assume each PSD to be represented by 2 lognormal distributions
  - fine particles (continental & sometimes absorbing)
  - coarse particles (oceanic / sea salt & non-absorbing)

Typical coastal/urban aerosols





Typical open ocean/maritime aerosols

- each PSD modulated by varying relative humidity
  - humidity changes particle size
  - requires ancillary data from NCEP
- 80 aerosol tables total, built from AERONET measurements
  - 10 PSDs
  - 8 relative humidities

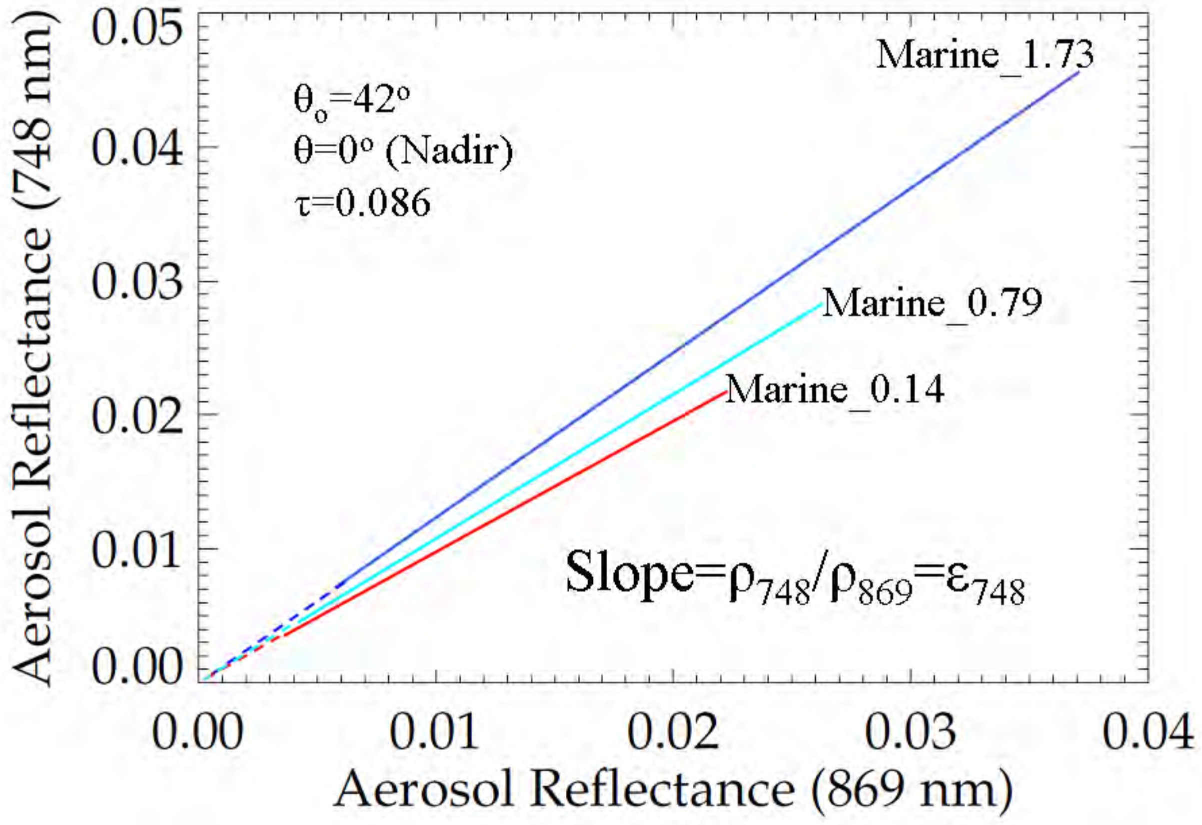
see Ahmad et al., Applied Optics, 2010

aerosol tables

- the Angstrom exponent (α) provides an estimator of particle size
  - high α = small particles
  - low α = large particles
  - defined via $\frac{\tau_a(\lambda)}{\tau_a(\lambda_o)} = \left(\frac{\lambda_o}{\lambda}\right)^{\alpha}$

- aerosol models often defined by epsilon (ε)
  - $\varepsilon(748, 869) = \frac{L_a(748)}{L_a(869)}$

do we know these values yet???



black pixel assumption

final unknowns in top expression

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T L_g$$

in the open ocean, we can assume (???) that L_w in the near-infrared (NIR) is = 0 (i.e., the ocean is *black* in the NIR)

thus, in the NIR (e.g., 748 and 869 nm):
L_a(NIR) + L_{ra}(NIR) = L_t(NIR) – the terms we computed

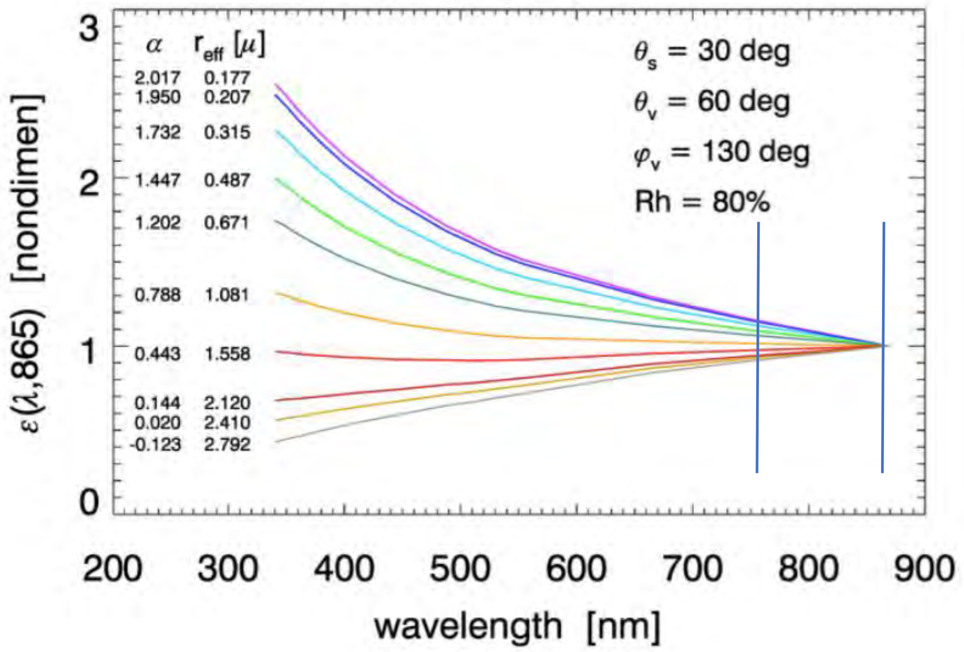
aerosol selection

L_w(NIR) = 0, so L_a(NIR) + L_{ra}(NIR) = L_t(NIR) – (everything previously computed)
how do we estimate L_a(visible) + L_{ra}(visible)?

- let's refer to [L_a + L_{ra}] simply as L_a & ignore single- vs. multi-scattering issues
- select the 10 aerosol tables that match the observed NCEP relative humidity
- compute epsilon values for the 10 tables [$\varepsilon(748,869) = L_a(748) / L_a(869)$]
- perform an iterative determination of the mean $\varepsilon(748,869)$ value & select a final bounding 2 aerosol models
- using 2 bounding models, calculate $\varepsilon(\lambda,869)$ from $\varepsilon(748,869)$
- calculate $L_a(\lambda) = \varepsilon(\lambda,869)\, L_a(869)$

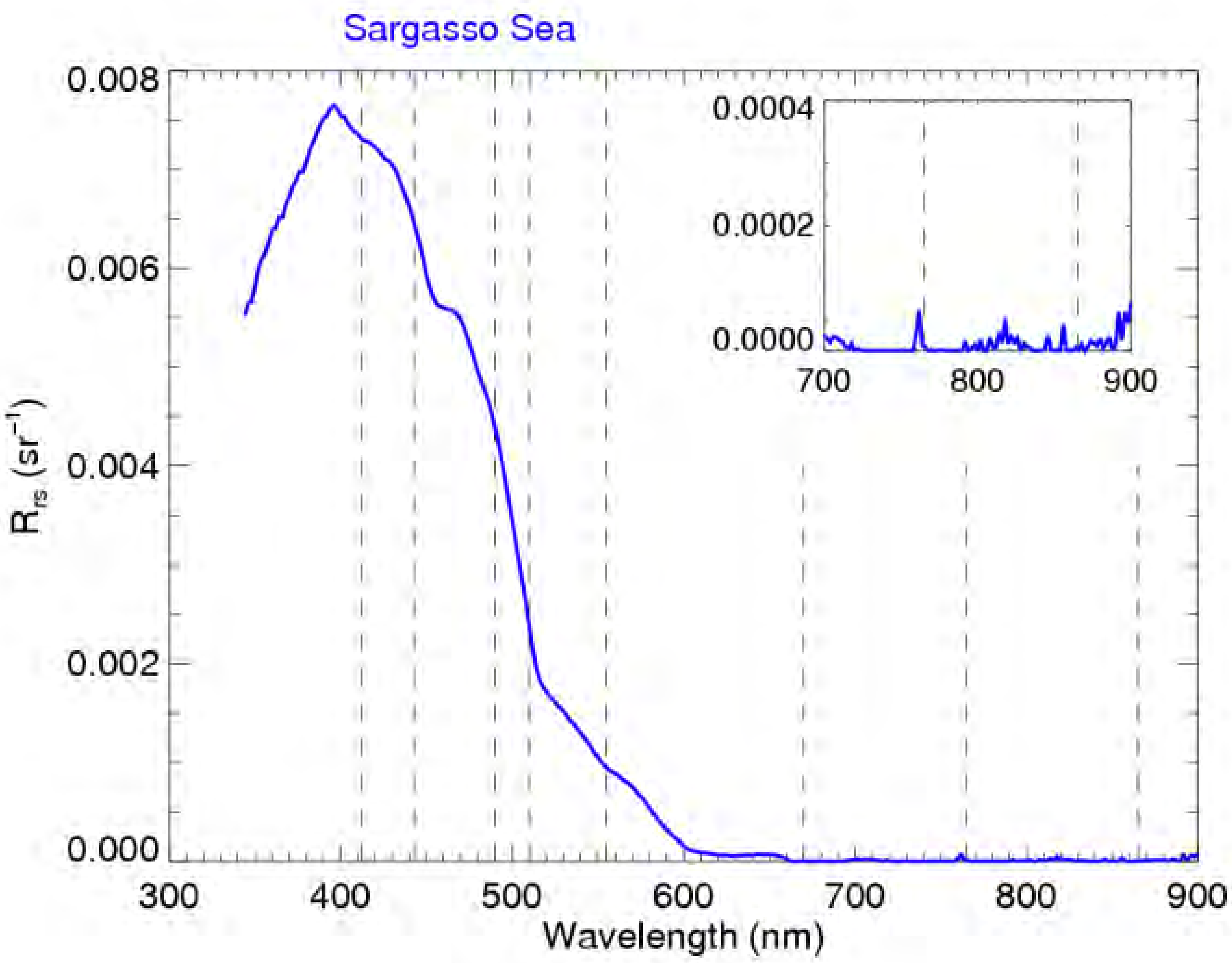
see Gordon & Wang, Applied Optics, 1994

final retrieval of $L_a(\lambda)$ is more accurate than that of τ_a and α; not unlike retrievals of $a(\lambda)$ being more accurate than $a_{dg}(\lambda)$ & $a_{ph}(\lambda)$ in inversion models



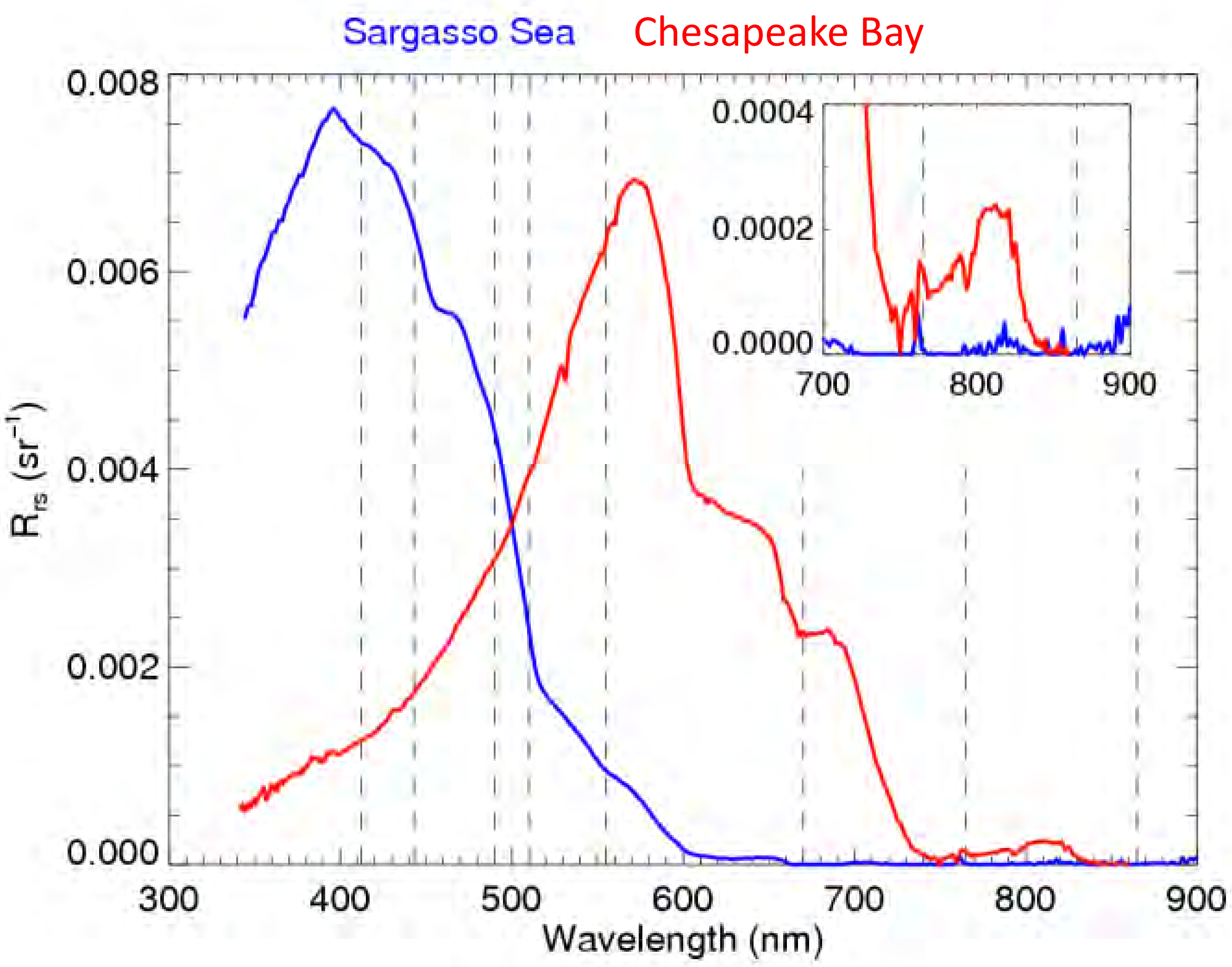
is the black pixel assumption valid?

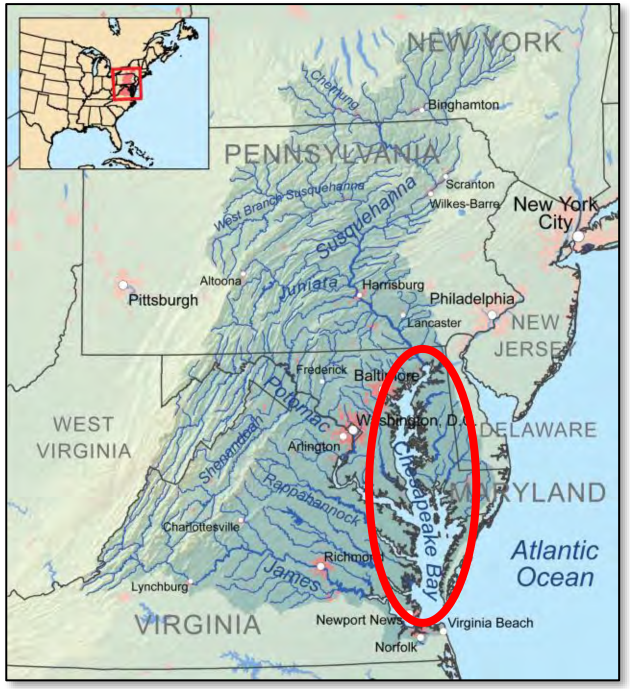
Is R_{rs}(NIR) really black?



is the black pixel assumption valid?

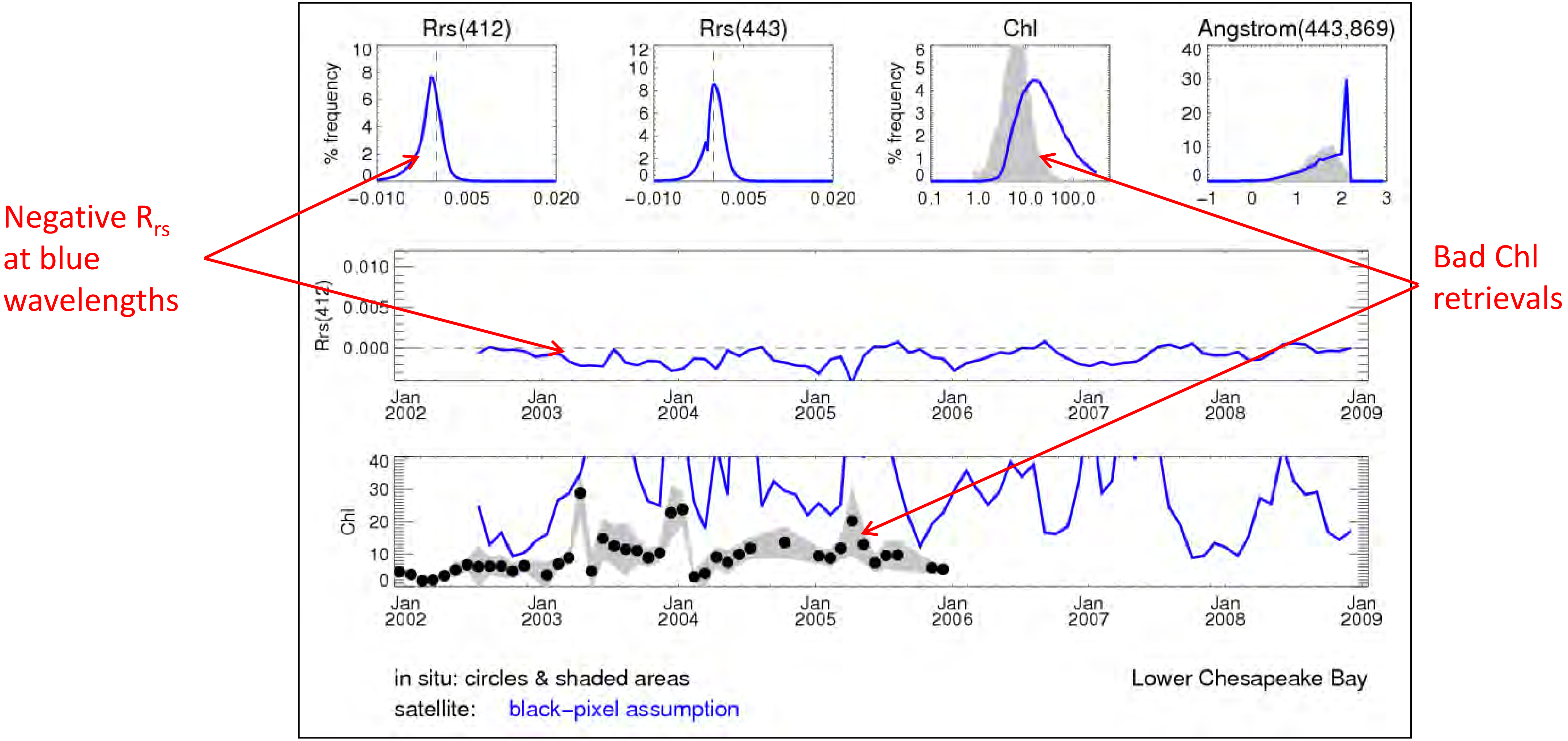
Is R_{rs}(NIR) really black?





is the black pixel assumption valid?

what happens when we don't account for R_{rs}(NIR) > 0?



retrievals using the "black pixel" assumption (e.g., SeaWiFS 1997-2000)

how to proceed with the black pixel assumption?

what to do when $R_{rs}(NIR) > 0$?

many approaches exist, here are a few examples:

assign aerosols (ε) and/or water contributions ($R_{rs}(NIR)$)

e.g., Hu et al. 2000, Ruddick et al. 2000

field data!

use shortwave infrared bands

e.g., Wang & Shi 2007

correct/model the non-negligible $R_{rs}(NIR)$

Siegel et al. 2000	used in SeaWiFS Reprocessing 3 (2000)
Stumpf et al. 2003	used in SeaWiFS Reprocessing 4 (2002)
Lavender et al. 2005	MERIS
Bailey et al. 2010	used in SeaWiFS Reprocessing 6 (2009)

use a coupled ocean-atmosphere optimization

e.g., Chomko & Gordon 2001, Stamnes et al. 2003, Kuchinke et al. 2009

The Bailey et al. (2010) iterative procedure

1. Assume that $R_{rs}(765)$ and $R_{rs}(865)$ are both 0, i.e. make the black-pixel assumption for both NIR reference bands.

2. Complete the atmospheric correction process as described in §9.2. This gives the initial estimate of $\rho_w(\lambda)$, or equivalently $R_{rs}(\lambda)$.

3. Use $R_{rs}(443)$ and $R_{rs}(555)$ from the initial estimate of $R_{rs}(\lambda)$ to get η by the empirical relationship (Lee et al. (2010, Eq. 8); Bailey et al. (2010, Eq. 3))

$$\eta = 2\left[1 - 1.2\exp\left(0.9\frac{R_{rs}(443)}{R_{rs}(555)}\right)\right]. \quad (9.10)$$

bio-optical model

4. Use the initial $R_{rs}(\lambda)$ to get an initial estimate of the chlorophyll concentration Chl. The particular algorithm used to obtain Chl from $R_{rs}(\lambda)$ depends on the sensor.

5. Use this Chl to obtain $a(670)$ via the empirical relationship (Bailey et al., 2010, Eq. 4)

$$a(670) = \exp[0.9389\ln(Chl) - 3.7589] + a_w(670). \quad (9.11)$$

bio-optical model

where the $a_w(670) = 0.439\ \mathrm{m}^{-1}$ is the absorption by pure water.

6. Use $a(670)$ and $R_{rs}(670)$ in Eq. (9.9) to solve for $b_b(670) = b_{bw}(670) + b_{bp}(670)$, where $b_{bw}(670) = 4.26 \times 10^{-4}\ \mathrm{m}^{-1}$ is the backscatter coefficient for pure sea water.

$$R_{rs}(\lambda) = \frac{f(\lambda)}{Q(\lambda)}\frac{b_b(\lambda)}{a(\lambda) + b_b(\lambda)}$$ Eq. (9.9)

f/Q is modeled for Case 1 water as a function of Chl and is assumed known

bio-optical model

The Bailey et al. (2010) iterative procedure

7. Use η from Eq. (9.10) and (Bailey et al., 2010, Eqs. 2b,3) to compute $b_b(765)$:

$$b_b(765) = b_{bw}(765) + b_{pb}(670)\left(\frac{670}{765}\right)^{\eta} \tag{9.12}$$

bio-optical model

where $b_{bw}(765) = 2.38 \times 10^{-4}$ m^{-1}. $b_b(865)$ is computed in the same manner using $b_{bw}(865) = 1.41 \times 10^{-4}$ m^{-1}.

8. Use this $b_b(765)$ and $a(765) = a_w(765) = 2.85$ m^{-1} to get $R_{rs}(765)$ from Eq. (9.9). Similarly, compute $R_{rs}(865)$ using $b_b(865)$ and $a_w(865) = 4.61$ m^{-1}.

9. Use the new, non-zero value of $R_{rs}(765)$ (i.e. $\rho_w(765)$) to remove the non-zero $\rho_w(765)$ contribution to $\rho_t(765)$. Do the same calculation for 865 nm.

10. Return to Step 2 and repeat the atmospheric correction using the black-pixel algorithm. This will give a new (hopefully better) estimate of $R_{rs}(\lambda)$, thus an new estimate of the other parameters, and finally new estimates of $R_{rs}(765)$ and $R_{rs}(865)$ at Step 8. After using the new values of $\rho_w(765)$ and $\rho_w(865)$ to correct for the non-zero water contribution to $\rho_t(765)$ and $\rho_t(865)$, return to Step 2 for a new iteration. Continue iterating until the change in $R_{rs}(765)$ from one iteration to the next is less than 2%, which typically takes 2-4 iterations, or when 10 iterations have been made.

The Bailey et al. (2010) iterative procedure

If this process doesn't converge after 10 iterations, try once more assuming that all NIR reflectance is due to water. If still no convergence, flag the pixel as "atmospheric correction warning." (The pixel might still be useable; your call.)

The above iteration is not done if Chl < 0.3 mg m^{-3}, and is always done if Chl > 0.3. In between use linear combination of the two results.

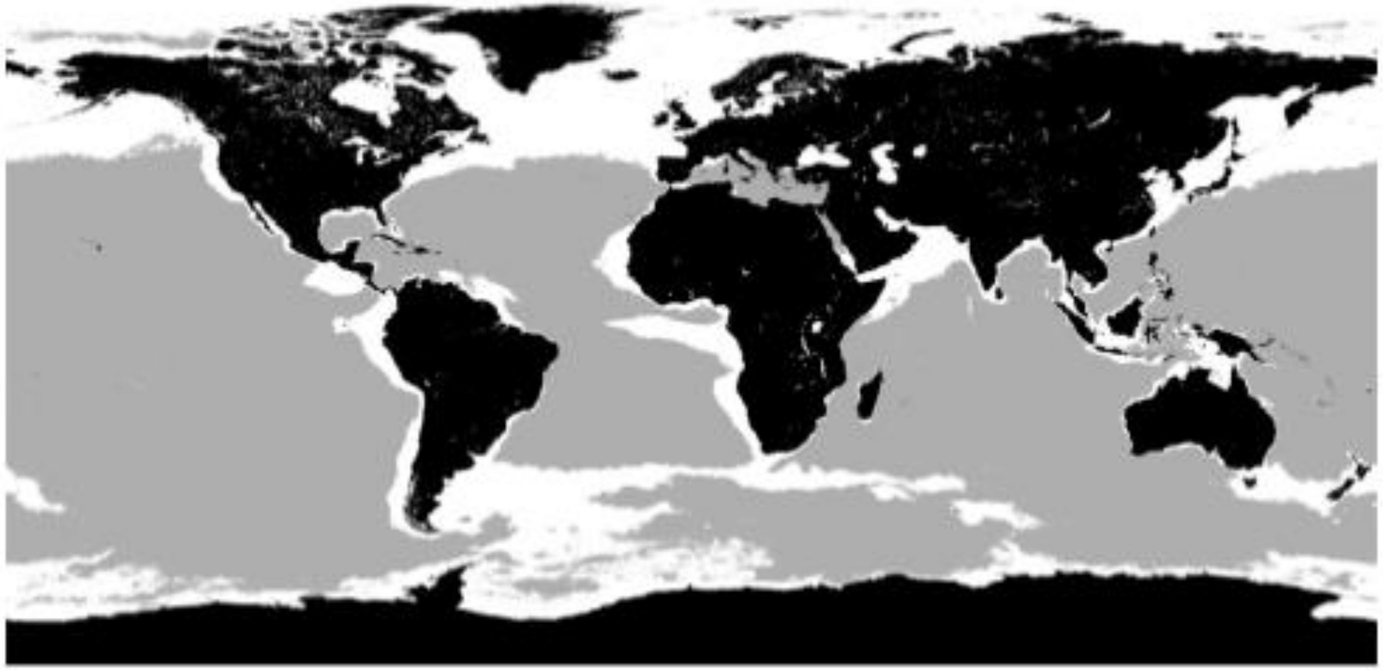
The two NIR wavelength bands depend on the sensor:

Table 9.1: NIR bands used for aerosol correction.

Band Label	Wavelengths [nm]	Nominal Wavelength [nm]
SeaWiFS		
7	745-785	$\lambda_1 = 765$
8	845-855	$\lambda_2 = 865$
MODIS		
15	743-753	$\lambda_1 = 748$
16	862-877	$\lambda_2 = 869$
VIIRS		
M6	739-754	$\lambda_1 = 745$
M7	846-885	$\lambda_2 = 862$

black pixel assumption – a bio-optical model

locations of application of bio-optical model



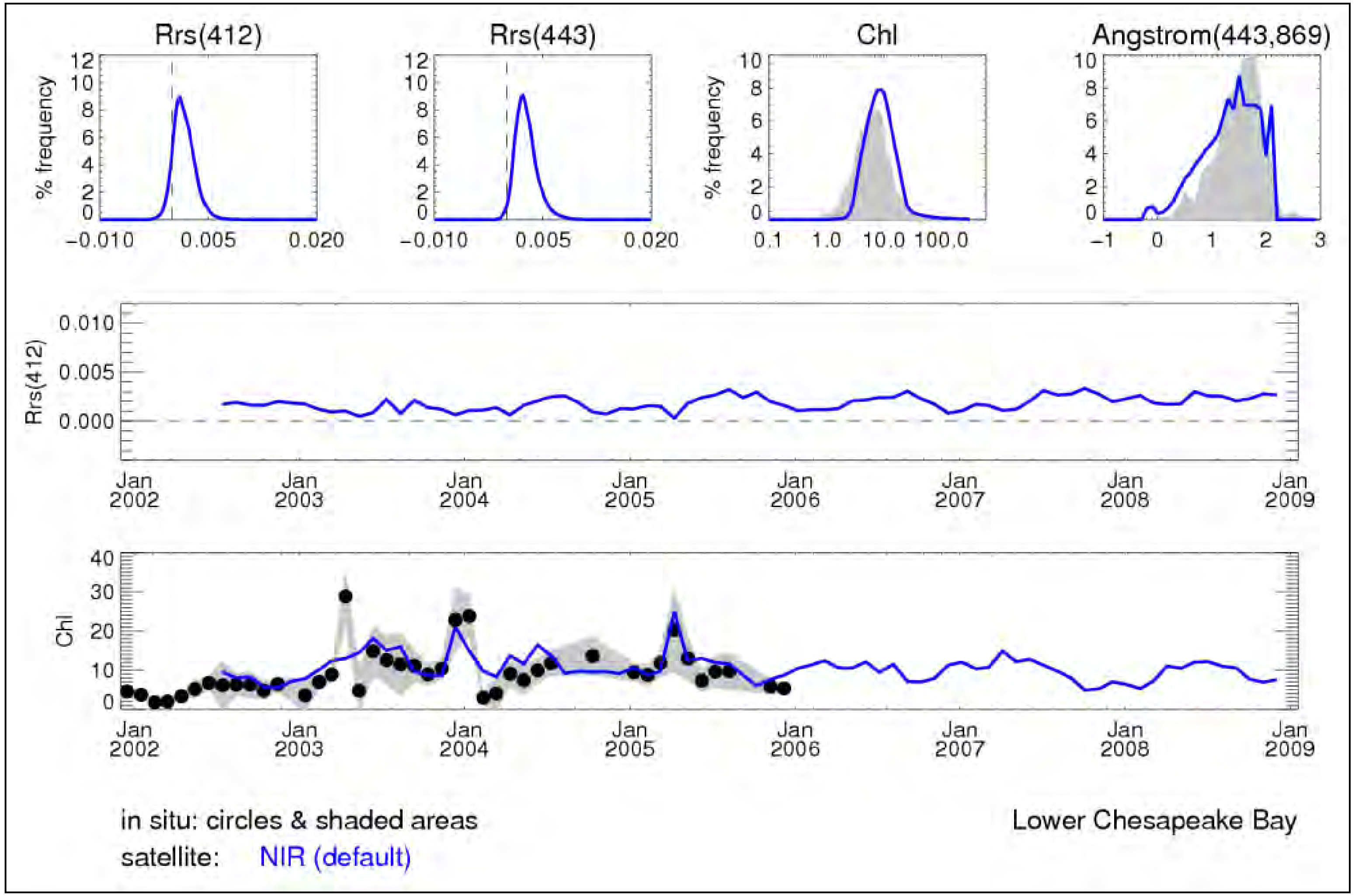
black = land; grey = Chl < 0.3 mg m^{-3}; white Chl > 0.3 mg m^{-3}

not applied when Chl < 0.3 mg m^{-3}
weighted application when 0.3 < Chl < 0.7 mg m^{-3}
fully applied when Chl > 0.7 mg m^{-3}

Bailey et al., Optics Express, 2010

black pixel assumption – a bio-optical model

correction of non-negligible R_{rs}(NIR)



estimate R_{rs}(NIR) using a bio-optical model

operational SeaWiFS & MODIS processing ~ 2000-present

example alternative aerosol selection schemes

ORIGINAL RESEARCH ARTICLE

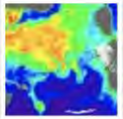
Front. Earth Sci., 31 May 2019 | https://doi.org/10.3389/feart.2019.00116

Check for updates

Multiband Atmospheric Correction Algorithm for Ocean Color Retrievals

Amir Ibrahim[1,2*], Bryan A. Franz[1], Ziauddin Ahmad[1,3] and Sean W. Bailey[1]

- "MBAC"
- uses multiple NIR/SWIR bands instead of 2
- GW94, but no assumption of single-scattering
- Will be used for PACE with its extra wavelengths



Atmospheric correction in presence of sun glint: application to MERIS

François Steinmetz, Pierre-Yves Deschamps, and Didier Ramon

Author Information ▾ Find other works by these authors ▾

Optics Express Vol. 19, Issue 10, pp. 9783-9800 (2011) • https://doi.org/10.1364/OE.19.009783

- "POLYMER"
- spectral matching approach
- https://www.hygeos.com/polymer
- Popular in Europe and better for Sun-glint regions

ORIGINAL RESEARCH ARTICLE Provisionally accepted The full-text will be published soon. Notify me

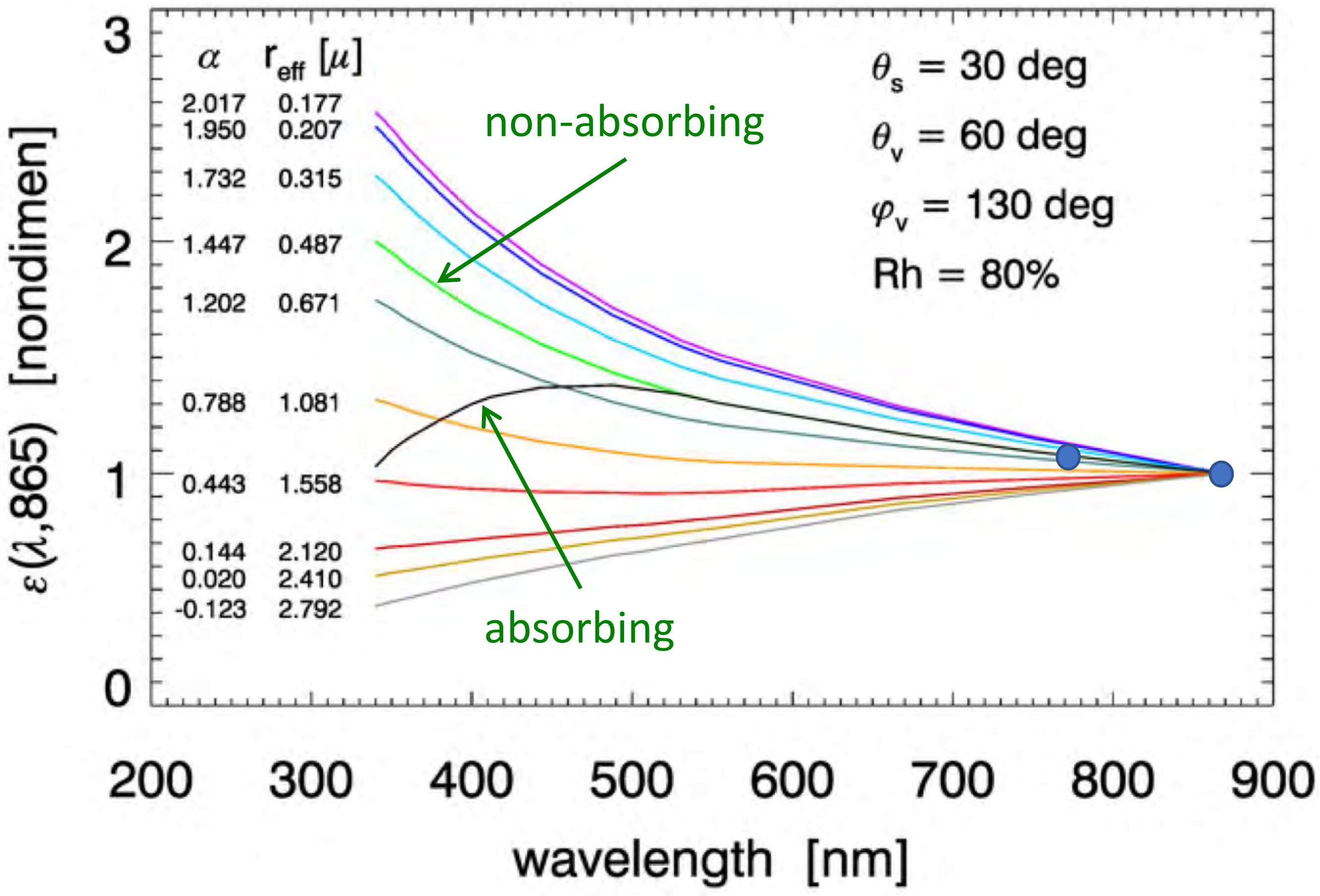
Front. Earth Sci. | doi: 10.3389/feart.2019.00100

Modeling atmosphere-ocean radiative transfer: A PACE mission perspective.

Jacek Chowdhary[1*], Pengwang Zhai[2], Emmanuel Boss[3], Heidi M. Dierssen[4], Robert J. Frouin[5], Amir I. Ibrahim[6], Zhongping Lee[7], Lorraine A. Remer[6], Michael Twardowski[8], Feng Xu[9], Xiaodong Zhang[10], Matteo Ottaviani[11], William R. Espinosa[6] and Didier Ramon[12]

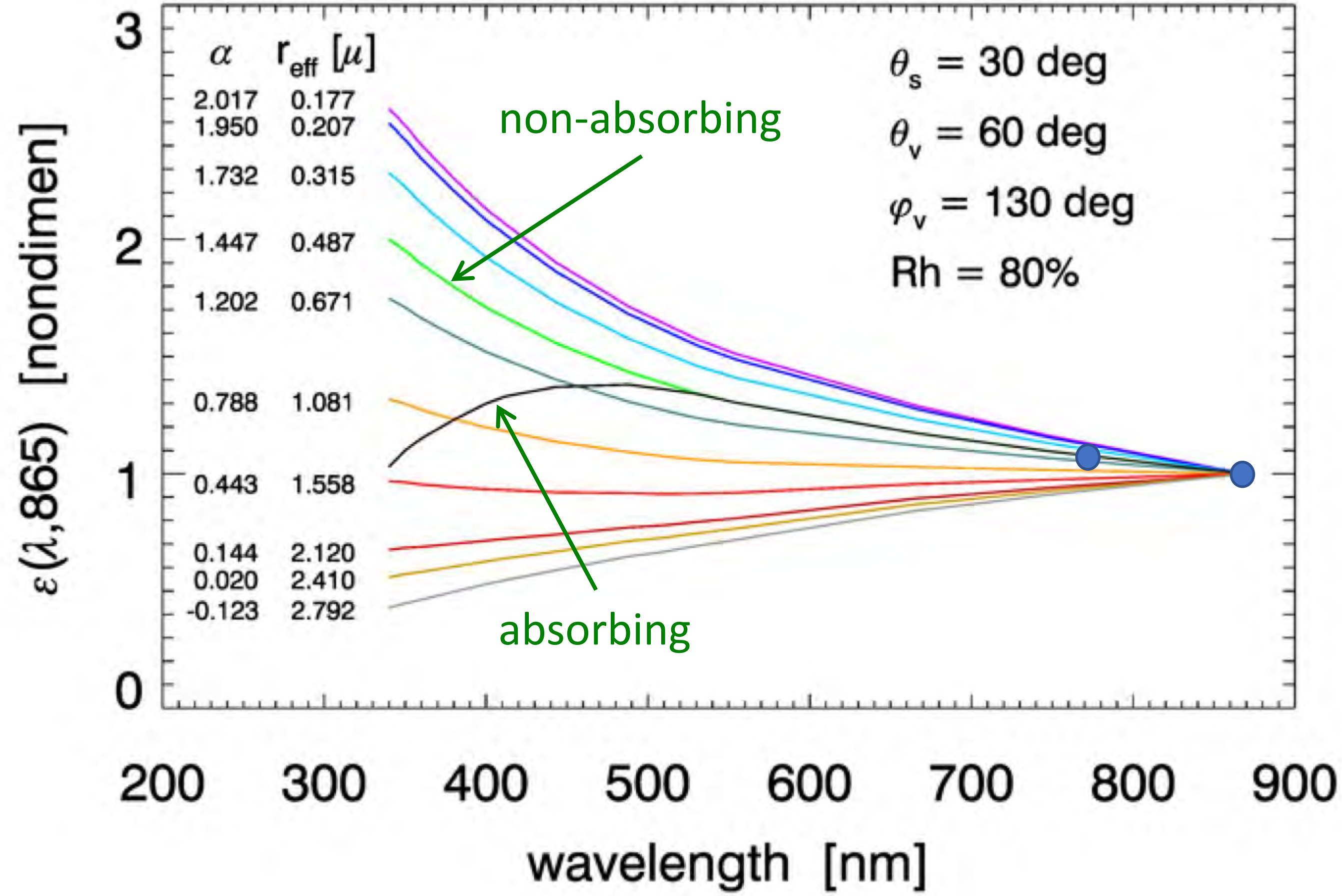
- radiative transfer for coupled ocean-atmosphere systems
- The ultimate goal but computationally expensive

atmospheric correction (currently) assumes no absorbing aerosols



Q: Why not use a wavelength near 350 or 400 nm, which could distinguish between absorbing and nonabsorbing aerosols?

atmospheric correction (currently) assumes no absorbing aerosols

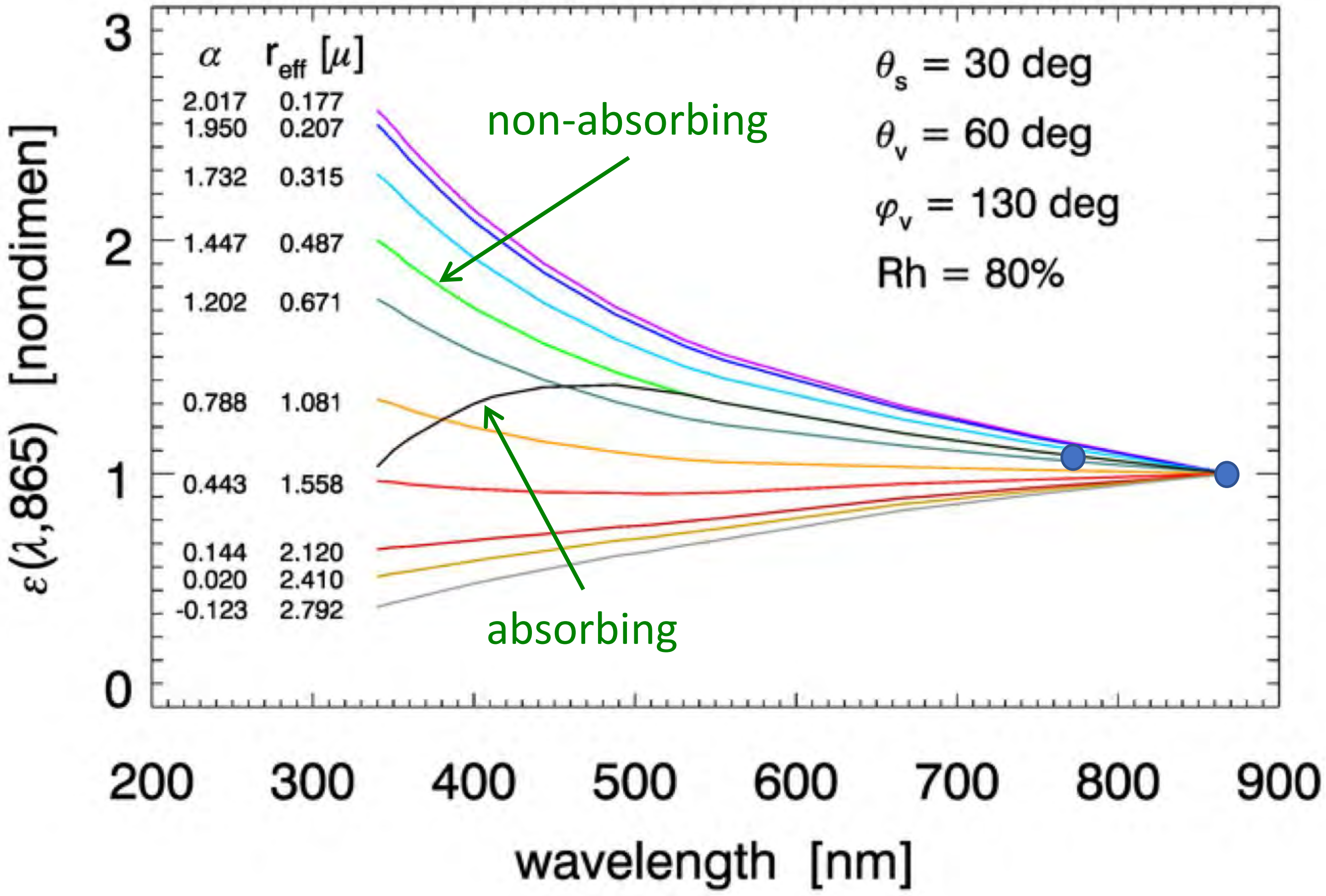


Q: Why not use a wavelength near 350 or 400 nm, which could distinguish between absorbing and nonabsorbing aerosols?

A: the water isn't black

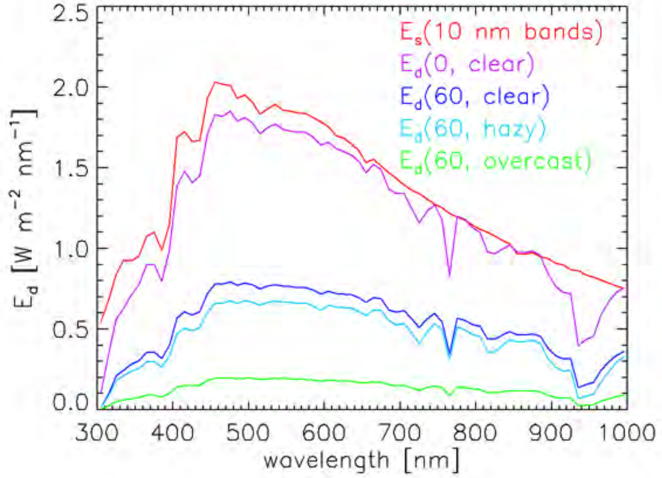
Why not use a wavelength < 300 nm, where the ocean is again black due to high CDOM and water absorption?

atmospheric correction (currently) assumes no absorbing aerosols

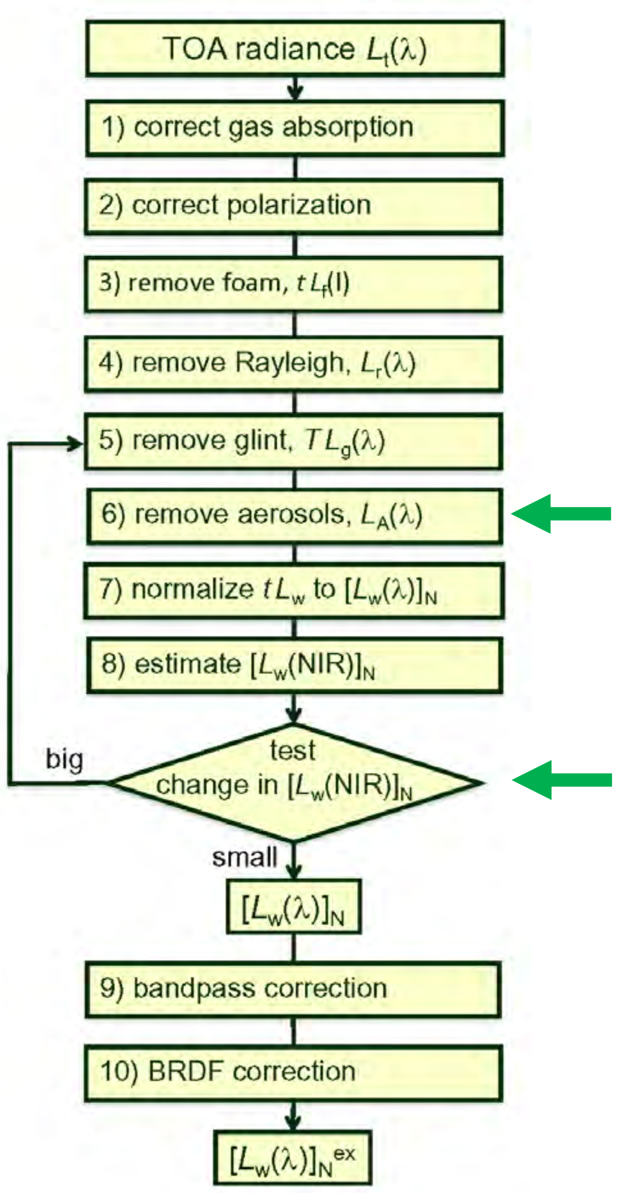


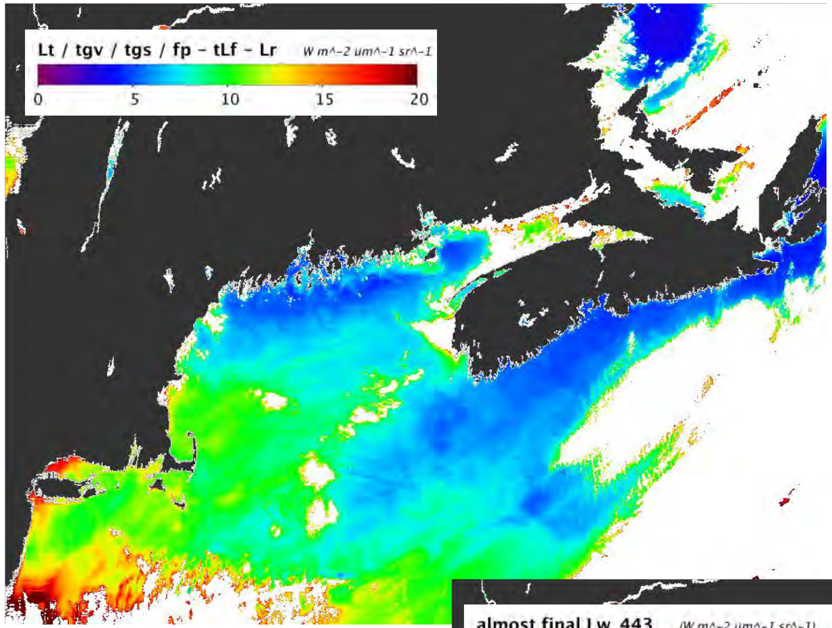
Q: Why not use a wavelength near 350 or 400 nm, which could distinguish between absorbing and nonabsorbing aerosols?
A: The water isn't black

Q: Why not use a wavelength < 300 nm, where the ocean is again black due to high CDOM and water absorption?
A: There ain't no sunlight

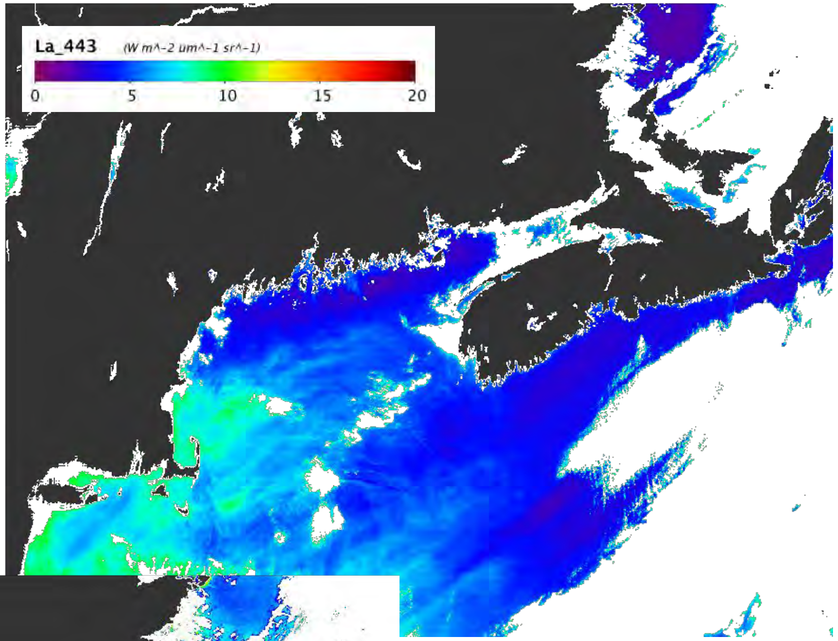


processing cadence: $L_t / t_{gv} / t_{gs} / f_p - tL_f - L_r - TL_g - L_a$

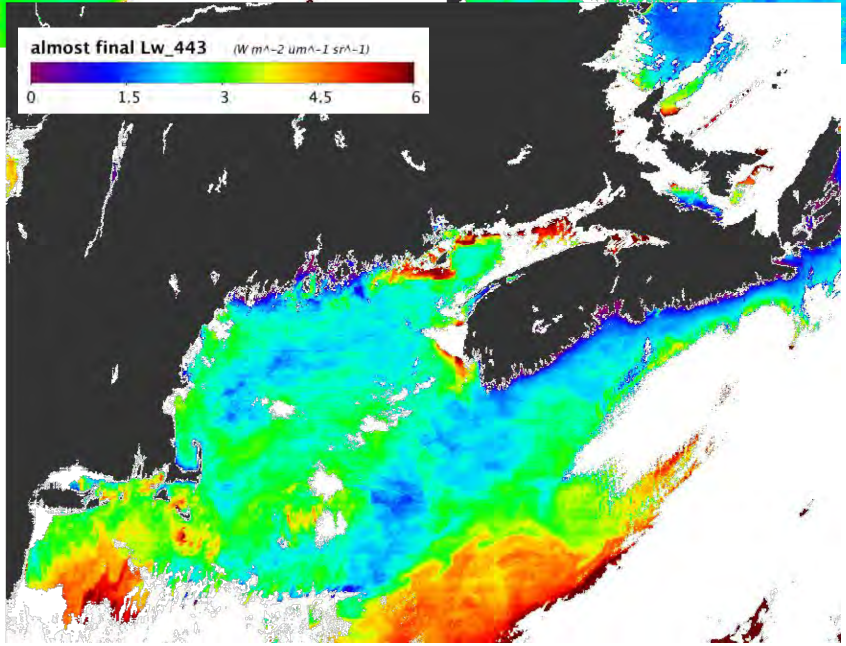




-



=

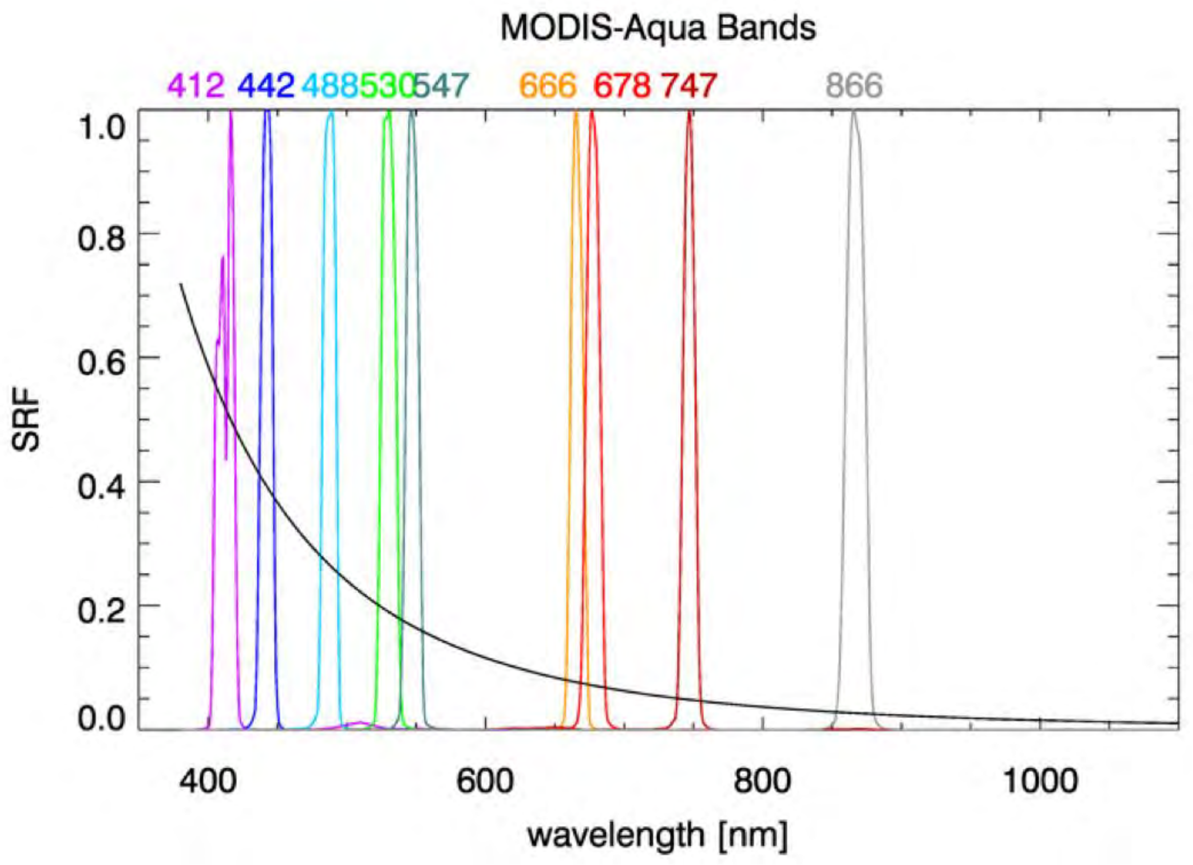


spectral bandpass correction

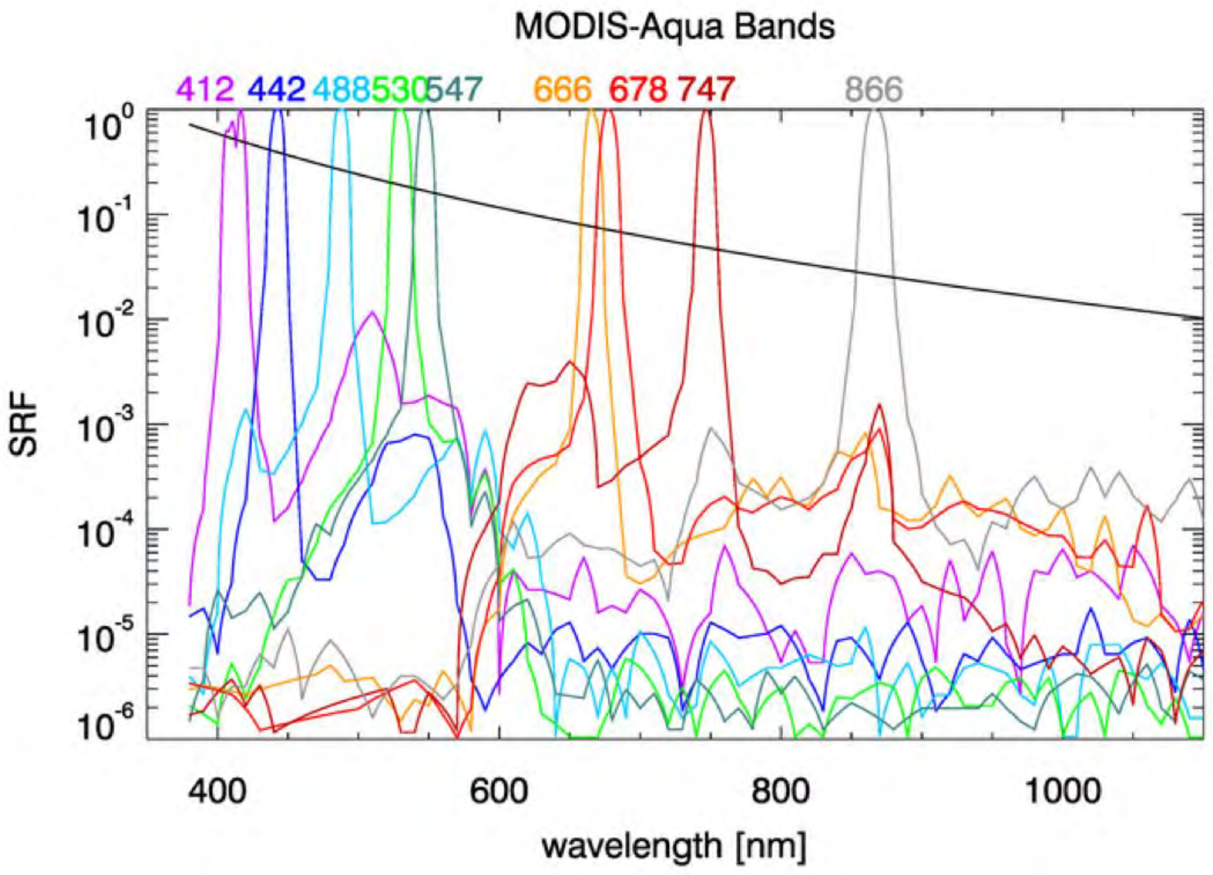
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(\theta_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

instrument spectral bandpasses

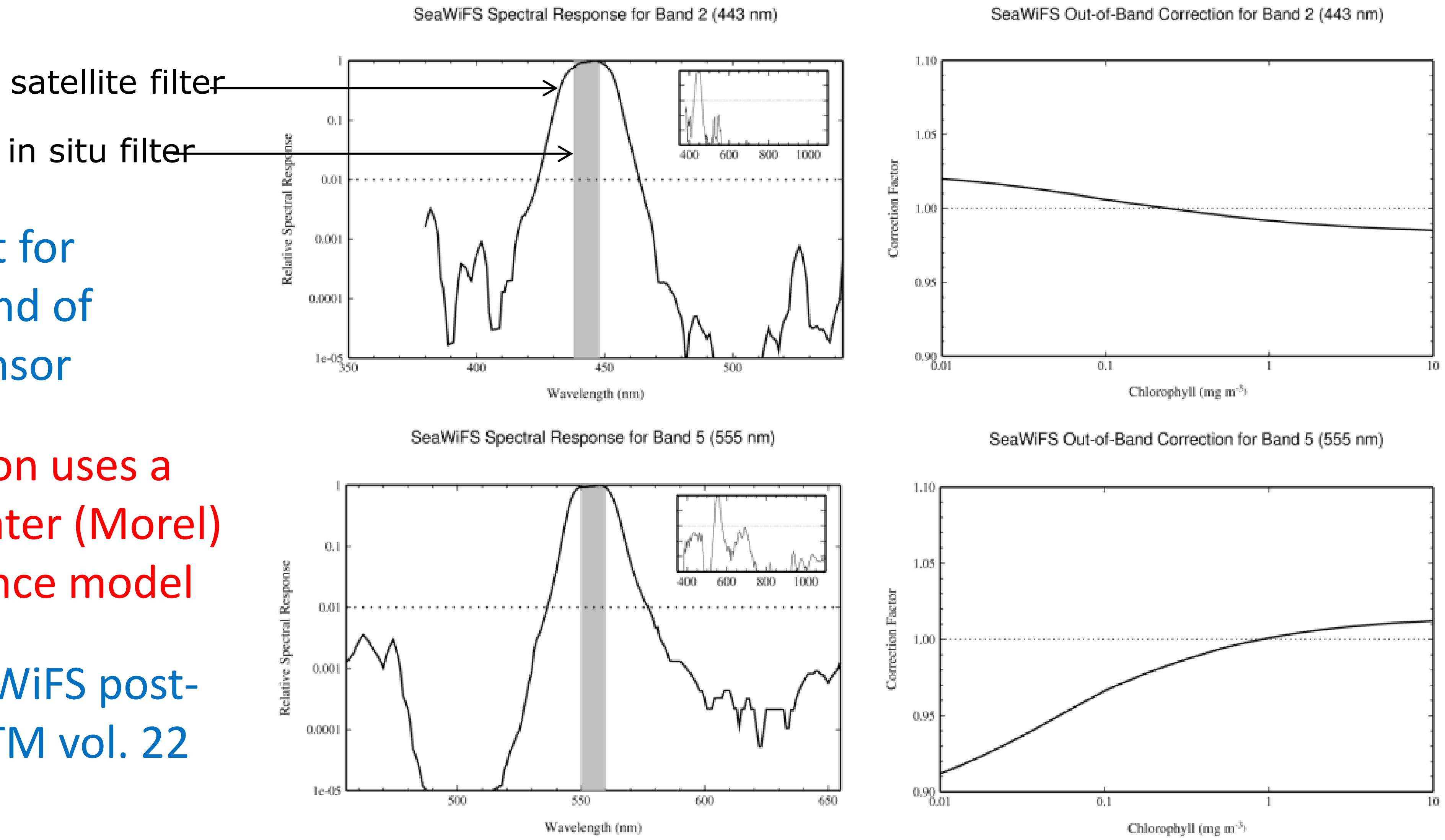


sensor bands appear well separated on a linear axis



however, sensor band response overlaps enough to cause problems

correction for light that comes from outside the nominal wavelength band

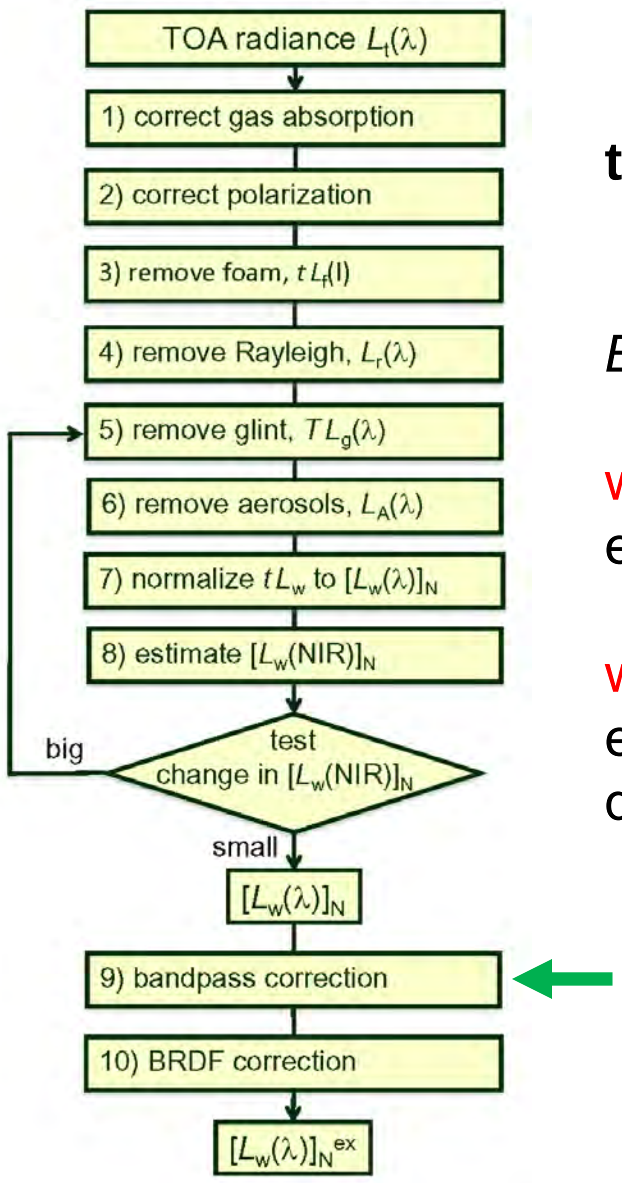


different for each band of each sensor

correction uses a clear-water (Morel) reflectance model

see SeaWiFS post-launch TM vol. 22

processing cadence: spectral bandpass correction



take care when executing satellite-to-in situ match-ups

Example for heritage multi-spectral satellite instruments:

when using multispectral in situ radiometers:
enable the bandpass adjustment

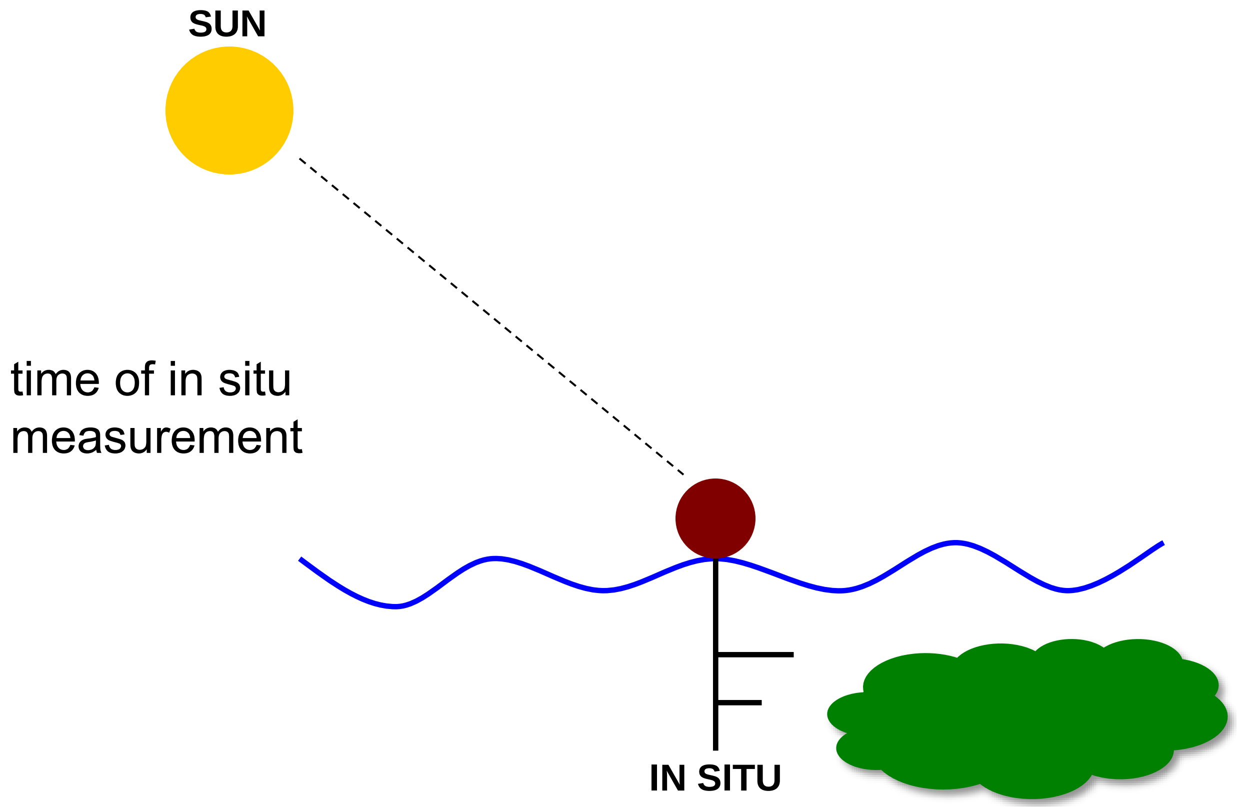
when using hyperspectral in situ radiometers:
enable the adjustment when applying 10-nm filter to in situ R_{rs}
disable the adjustment when applying full-spectral-response to in situ R_{rs}

bidirectional reflectance correction

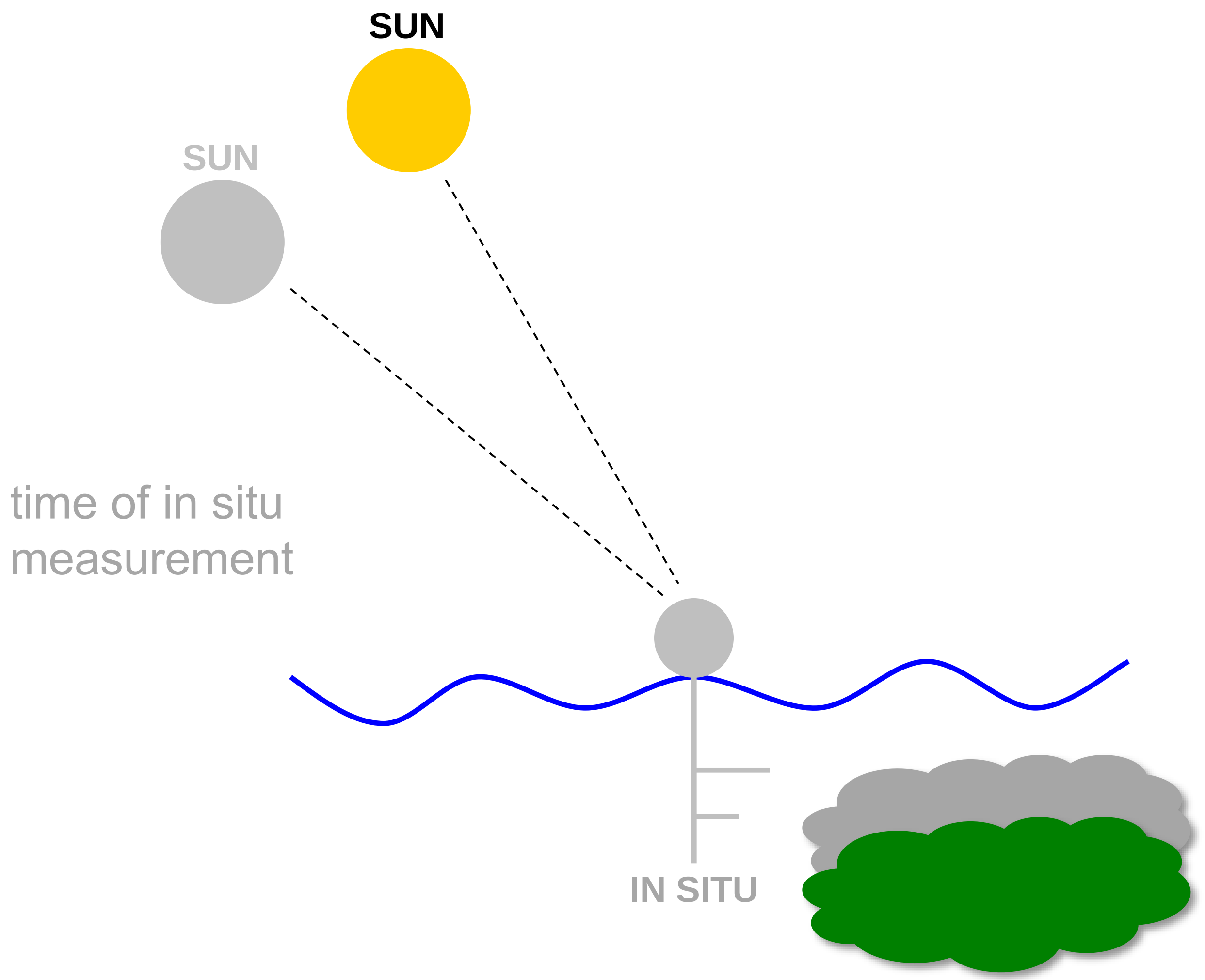
$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(\theta_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

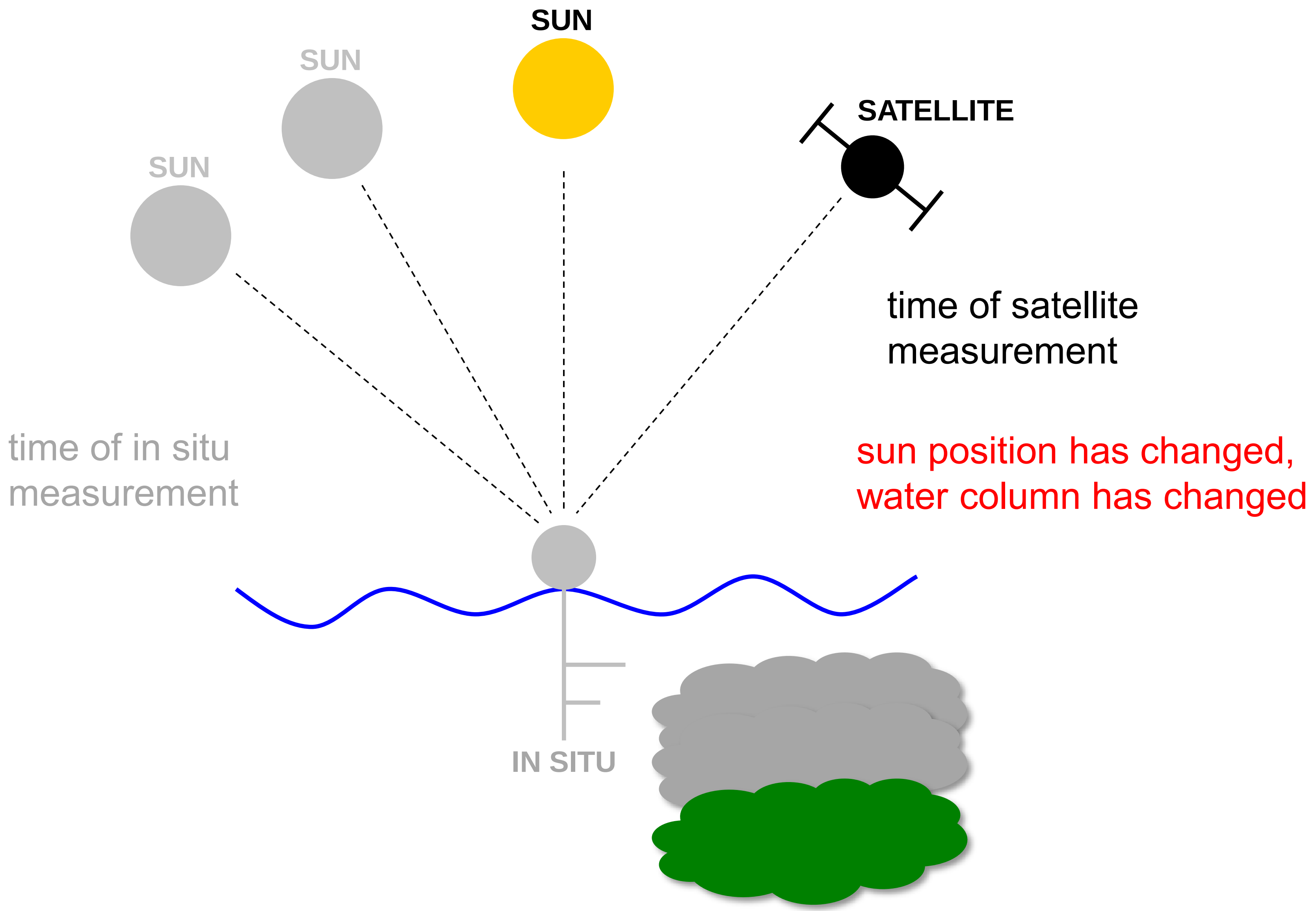
bidirectional reflectance correction



bidirectional reflectance correction



bidirectional reflectance correction



bidirectional reflectance correction

we normalize R_{rs} to account for Sun's changing position in the sky:

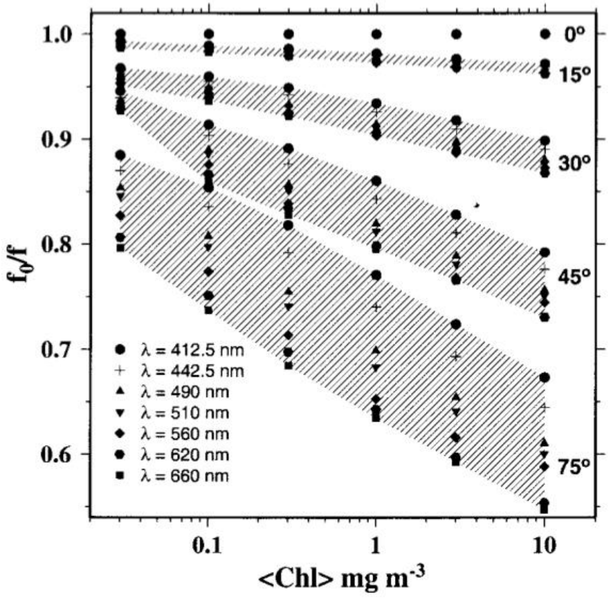
- pathlengths through atmosphere
- transmission of light through air-sea & sea-air interfaces
- angular features of in-water volume scattering functions

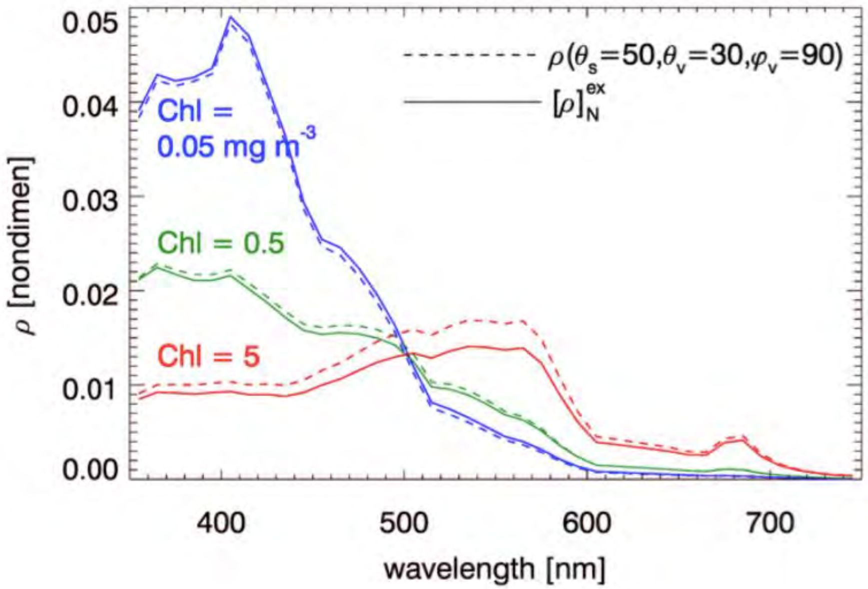
$$[L_w]_N^{ex} \equiv [L_w(\theta_v, \phi)]_N \frac{\mathfrak{R}_o(W)}{\mathfrak{R}(\theta_v', W)} \frac{f_o(\mathrm{ATM}, W, \mathrm{IOP})}{Q_o(\mathrm{ATM}, W, \mathrm{IOP})} \left[\frac{f(\theta_s, \mathrm{ATM}, W, \mathrm{IOP})}{Q(\theta_s, \theta_v', \phi, \mathrm{ATM}, W, \mathrm{IOP})} \right]^{-1}$$

Morel et al., Applied Optics, 2002

$\mathfrak{R}, \mathfrak{R}_0, f, f_0, Q, Q_0$ from look-up-tables based on Chl & geometries of Sun & sensor to normalize all measurements (no subscript) to condition of overhead Sun (subscript $_0$) and no atmospheric losses (not "no atmosphere")

bidirectional reflectance correction





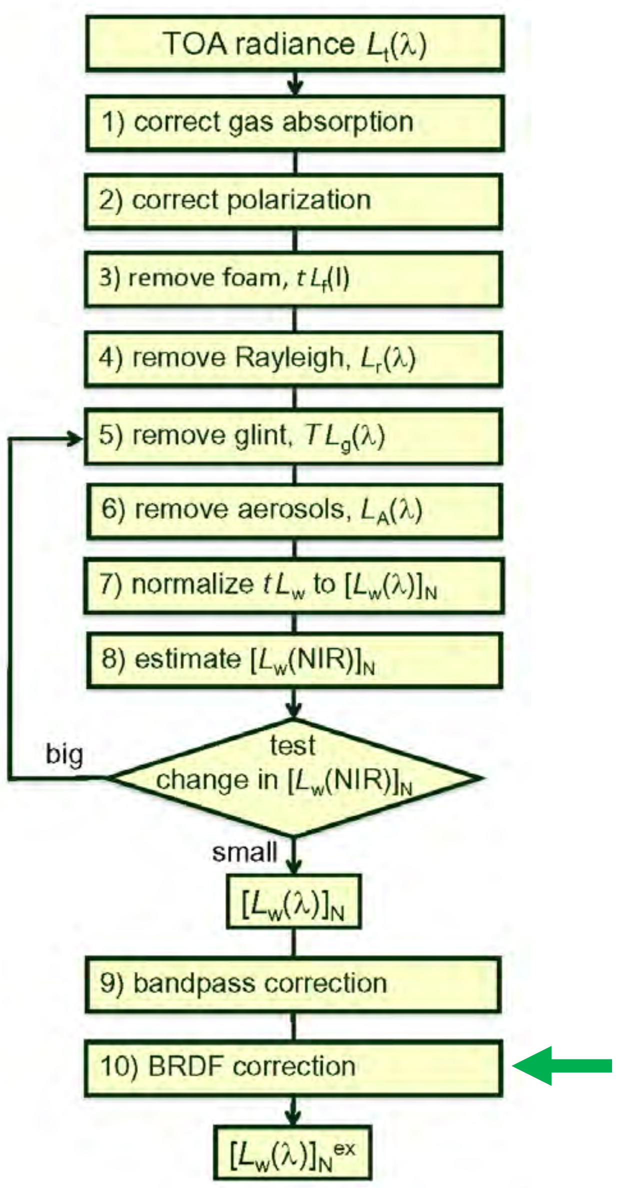
$$[L_w]_N^{ex} \equiv [L_w(\theta_v, \phi)]_N \frac{\Re_o(W)}{\Re(\theta'_v, W)} \frac{f_o(\text{ATM}, W, \text{IOP})}{Q_o(\text{ATM}, W, \text{IOP})} \left[\frac{f(\theta_s, \text{ATM}, W, \text{IOP})}{Q(\theta_s, \theta'_v, \phi, \text{ATM}, W, \text{IOP})}\right]^{-1}$$

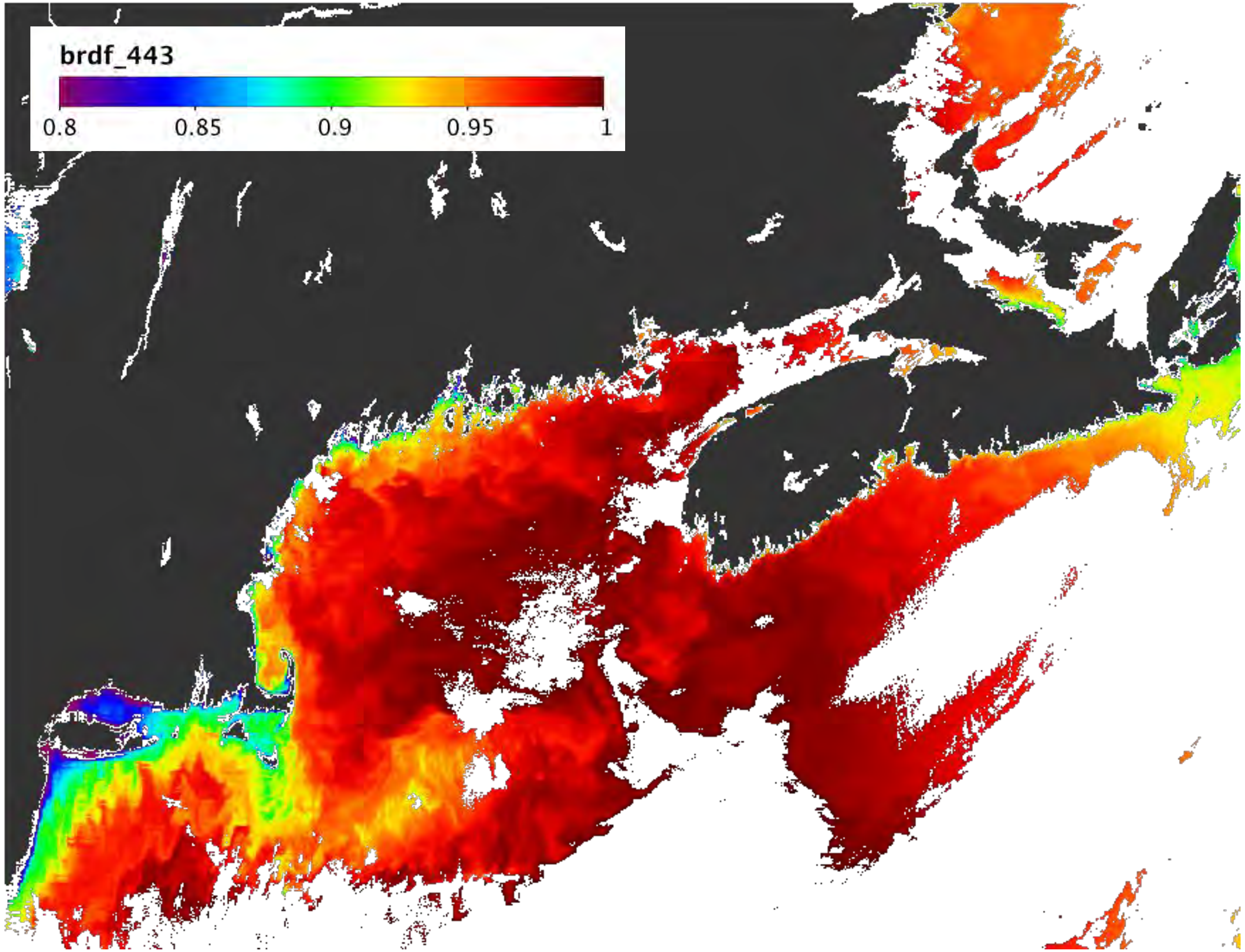
Morel et al., Applied Optics, 2002

$\Re, \Re_0, f, f_0, Q, Q_0$ from look-up-tables based on Chl & geometries of Sun & sensor

to normalize all measurements (no subscript) to condition of overhead Sun (subscript $_0$) and no atmospheric losses (not “no atmosphere”)

processing cadence: BRDF correction





so there there you have it – perfect R_{rs}

$$L_t = \left(L_r + [L_a + L_{ra}] + t_{dv} L_f + t_{dv} L_w\right) t_{gv}\, t_{gs}\, f_p + T\, L_g$$

$$R_{rs} = \frac{L_w}{F_0 \cos(q_s)\, t_{ds}\, f_s\, f_b\, f_l}$$

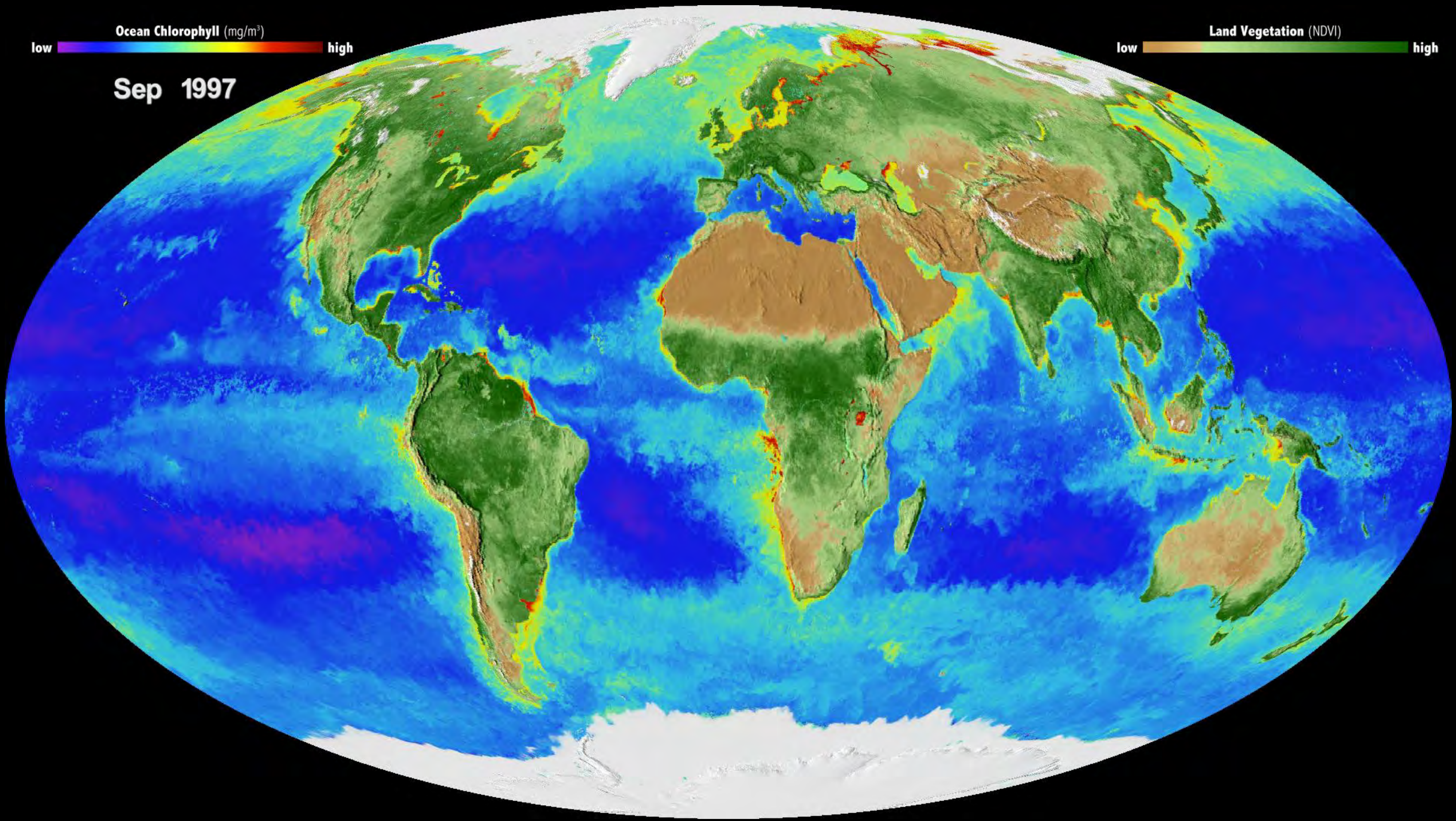
ancillary data requirements

ancillary data	ancillary source	uses
atmospheric pressure	NCEP	Rayleigh
water vapor	NCEP	transmittance
relative humidity	NCEP	aerosol models
wind speed	NCEP	white caps, Sun glint, Rayleigh
ozone	OMI/TOMS	transmittance
NO_2	Sciamachy/OMI/GOME	transmittance
sea surface temperature	Reynolds	bio-optical algorithms
sea ice	NSIDC	masking

look-up tables, coefficients

aerosol models
Rayleigh
Rayleigh optical thickness
ozone absorption
NO_2 absorption
pure seawater absorption, scattering, index of refraction (temp/sal dependent)
f/Q (bidirectional reflectance distributions)
others …

All of this actually does work pretty well most of the time!!

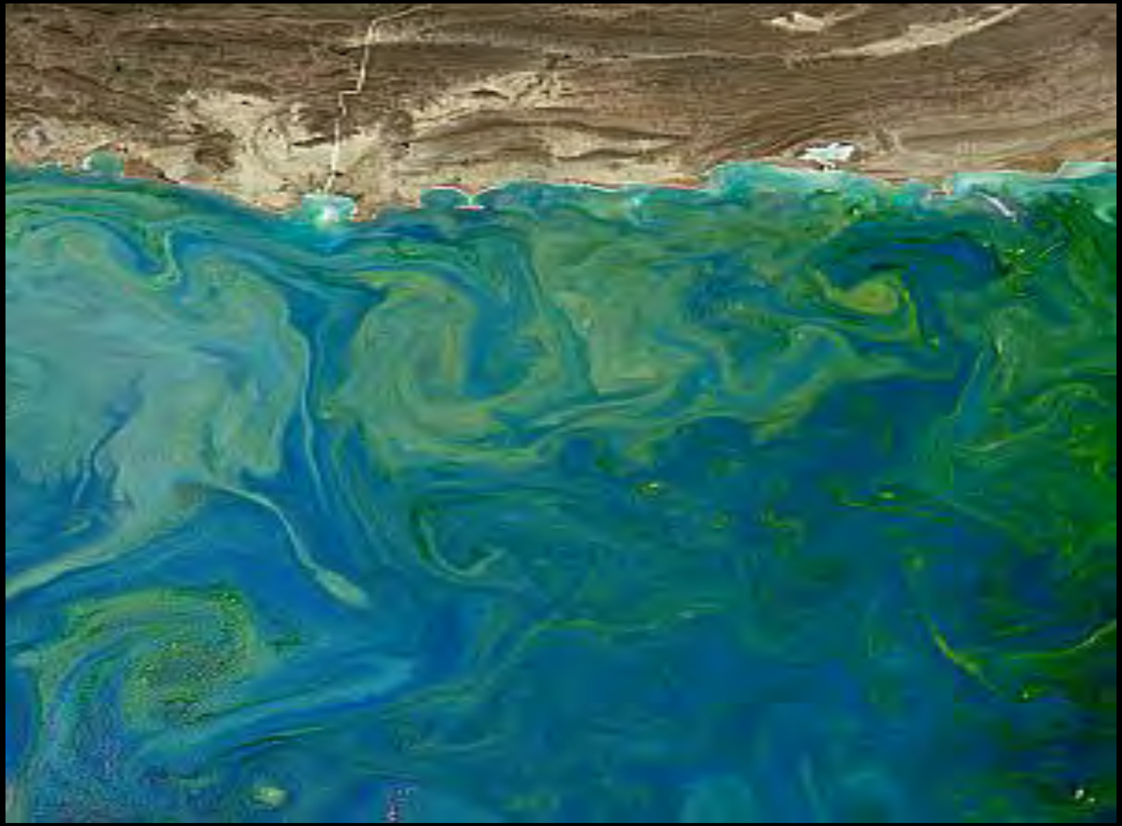


“I know of no competent biologist who considers this an important problem.”

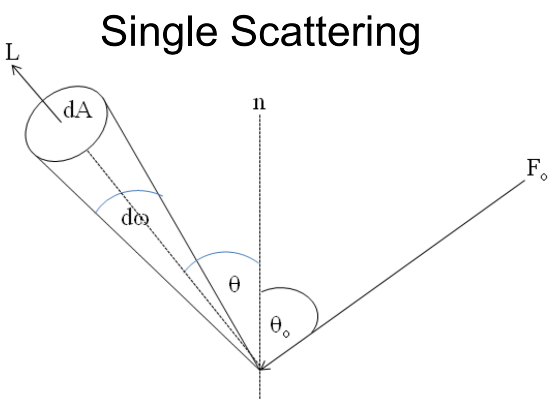
The opinion of an “eminent biologist” who reviewed the proposed Coastal Zone Color Scanner (1978-1986) back in the 1970s.

-- related to me by Howard Gordon

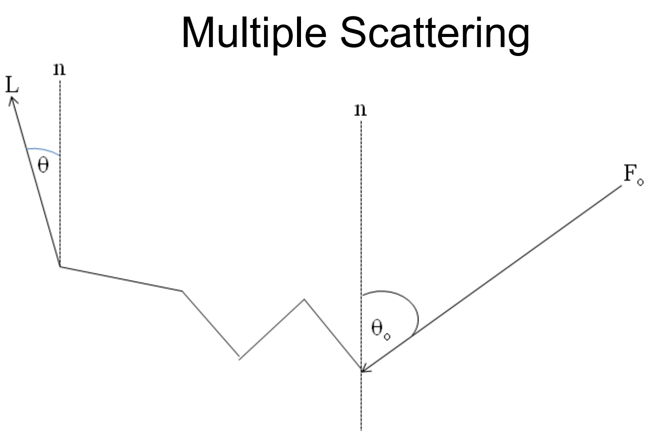
Thank you! Questions?



single vs. multiple scattering



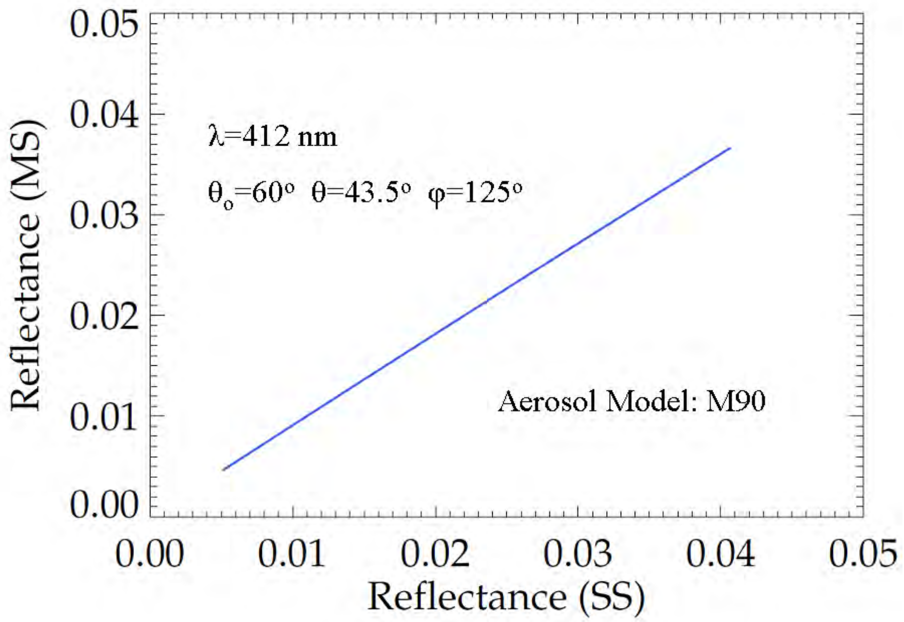
single scattering solution historically used to simplify math (CZCS – present)



real life – $L_a + L_{ra}$

in the atmospheric correction algorithm, the relationship between single (SS) and multiple scattered (MS) reflectances is defined as:

$$ln(\rho_{ms}) = a_0 + a_1\, ln(\rho_{ss}) + a_2[ln(\rho_{ss})]^2$$



see Gordon & Wang, Applied Optics, 1994

single vs. multiple scattering

why bring this up?

- satellites "observe" multi-scattering
- operational atmospheric corrections schemes convert to single-scattering

- what errors are introduced? (plus, error propagation made difficult)
- is this still necessary?