NOWPAP / PICES / WESTPAC Joint Training Course on Remote Sensing Data Analysis

Introduction and recent progress in ocean color remote sensing part II:

Correction of the influence of the atmosphere in Ocean Colour RS

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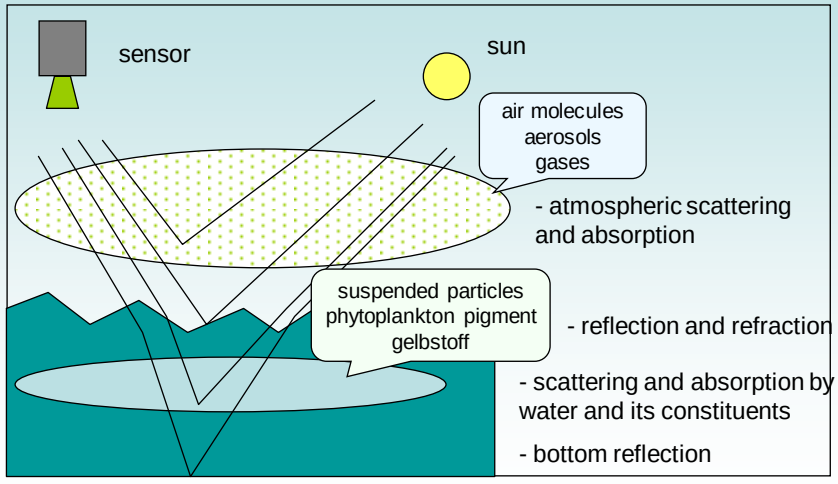
Space Shuttle 335 k
20060720 , NASA

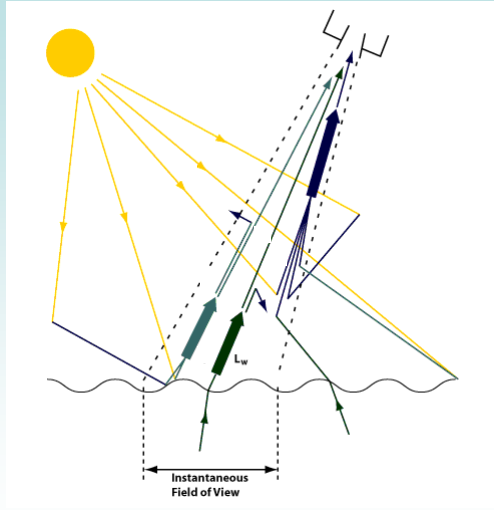
„Atmospheric Correction“

- Goal: to determine the water leaving radiance or water leaving radiance reflectance from top of atmosphere (TOA) radiances or reflectances
- The following quantities have to be determined:
  - Path radiance of the atmosphere
  - Transmittance of solar flux through the atmosphere
  - Transmittance of the radiance at the bottom of atmosphere (BOA) to the satellite
  - Solar light which is reflected at the water surface
    - Direct: sun glint
    - After scattering in the atmosphere: reflected sky light
- Normalisation of the reflectance, i.e. recomputation of the bi-directional reflectance for a sun in zenith and nadir viewing direction
- Further problem is the adjacency effect: sun light, which is reflected by a bright target in the neighbourhood of the water surface and which is then scattred by the atmosphere into the sensor

Basic principles of Water Color RS



Light paths to the sensor



the satellite observes both the **ocean** and the **atmosphere**

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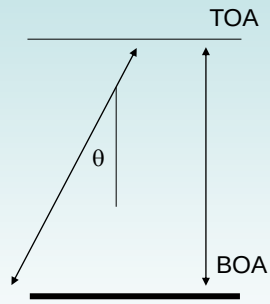
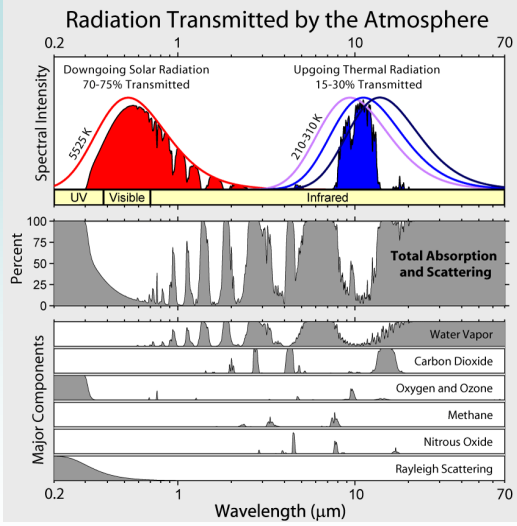
Optical Thickness, Transmittance, Angstrom

- Beam Transmittance $T = I_z / I_0$
- Extinction tau $\tau = -\log(T)$, $T = \exp(-\tau)$
- $T = \exp(-\tau/\cos(\theta))$
- Aerosol Optical Thickness AOT = τ_{ae}

- Angstrom coefficient:
  - Alpha $\alpha = \log(\tau_{\lambda 1} / \tau_{\lambda 2}) / \log(\lambda_1 / \lambda_2)$



Solar Spectrum 1

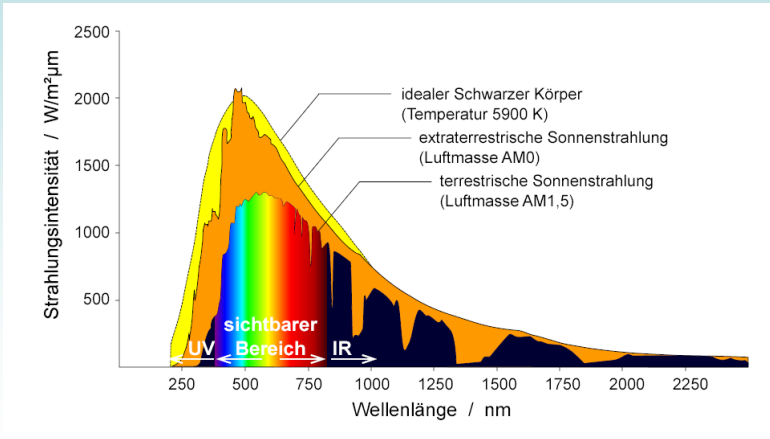


Composition of earth atmosphere:
- 78.09% nitrogen
- 20.95% oxygen,
- 0.247% water vapor (variable)
- 0.93% argon
- 0.038% carbon dioxide
- traces of hydrogen, helium, methane, ozone etc.

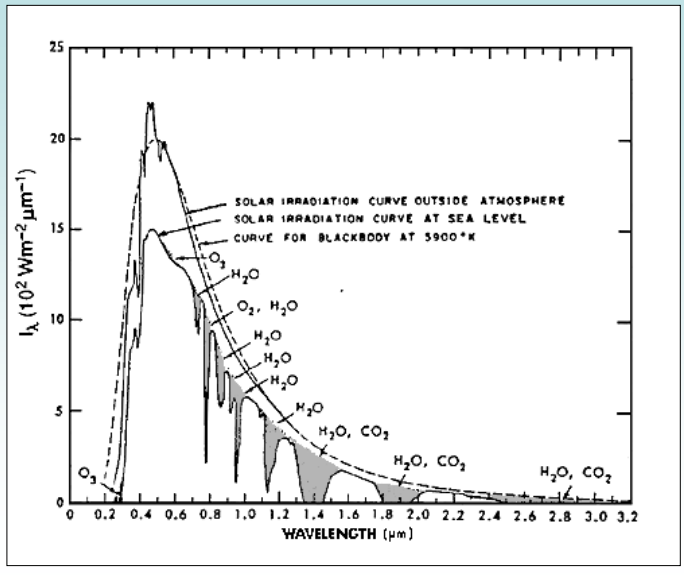
- Gases cause scattering and absorption
- Areas with no absorption are called windows

- Aerosols:
  - Salt cristals
  - Dust from deserts
  - Soot
  - Vulcanic ash

Solar Spectrum 2

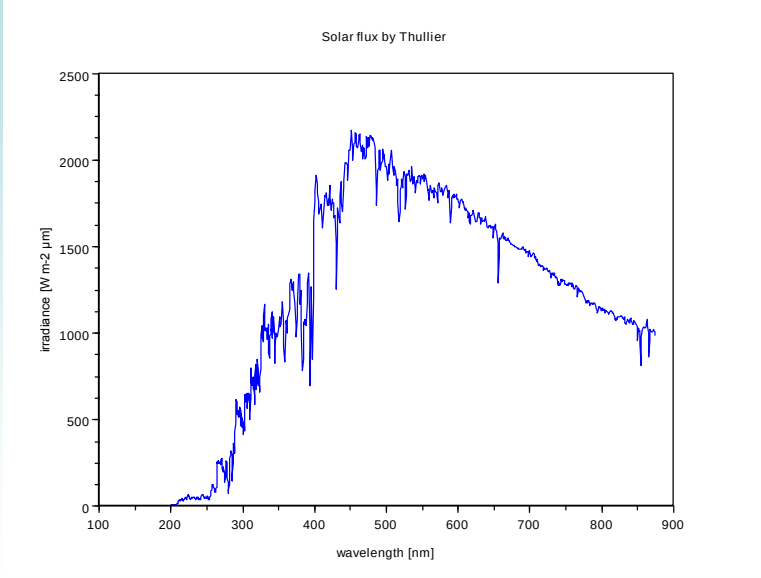


Atmospheric Transmission

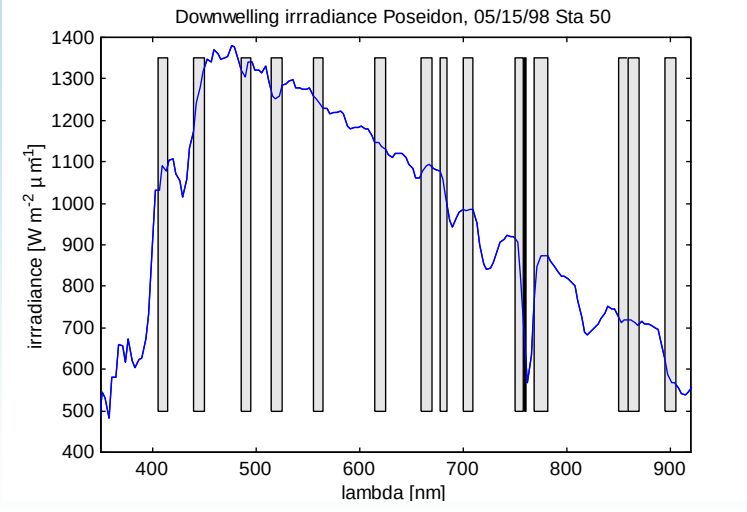


Solar radiation at the top of the atmosphere and the actual radiation at sea level which has been reduced due to absorption by atmospheric gases. The dashed curve is a blackbody at 5900K for comparison with the solar curve outside the earth's atmosphere

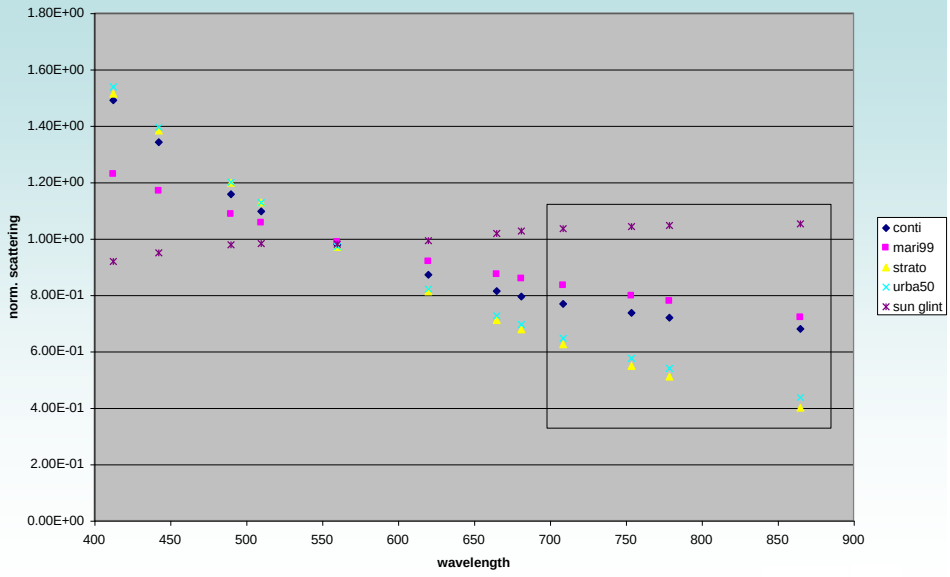
Solar Flux at TOA (Thullier)

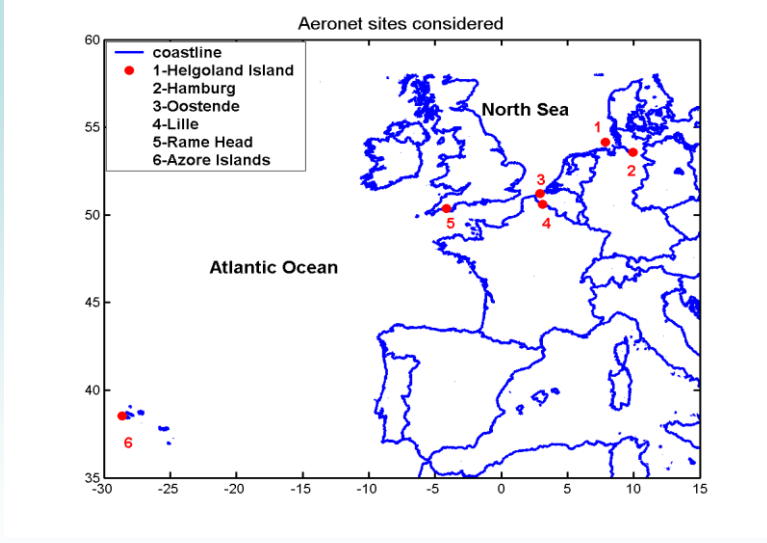


Downwelling irradiance

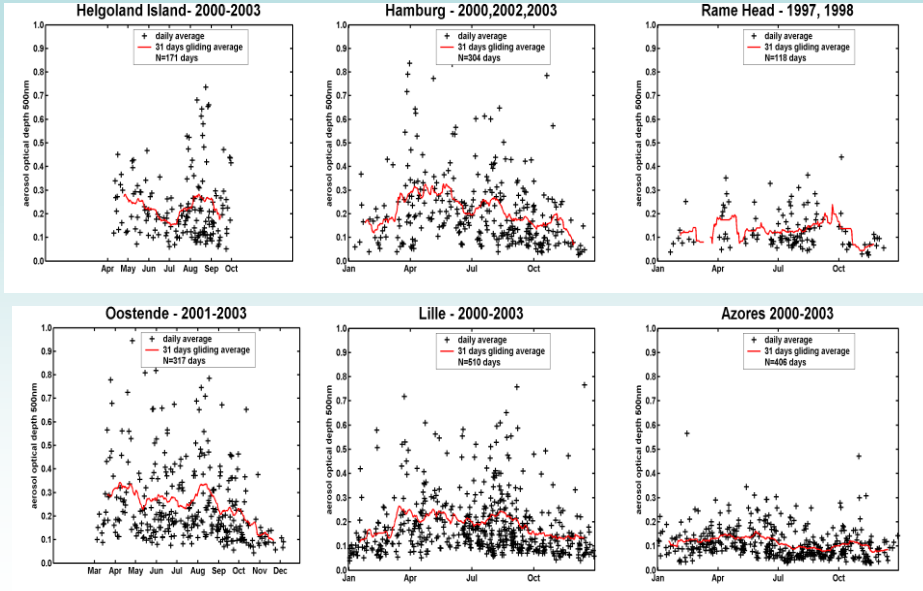


Aerosol Optical Properties

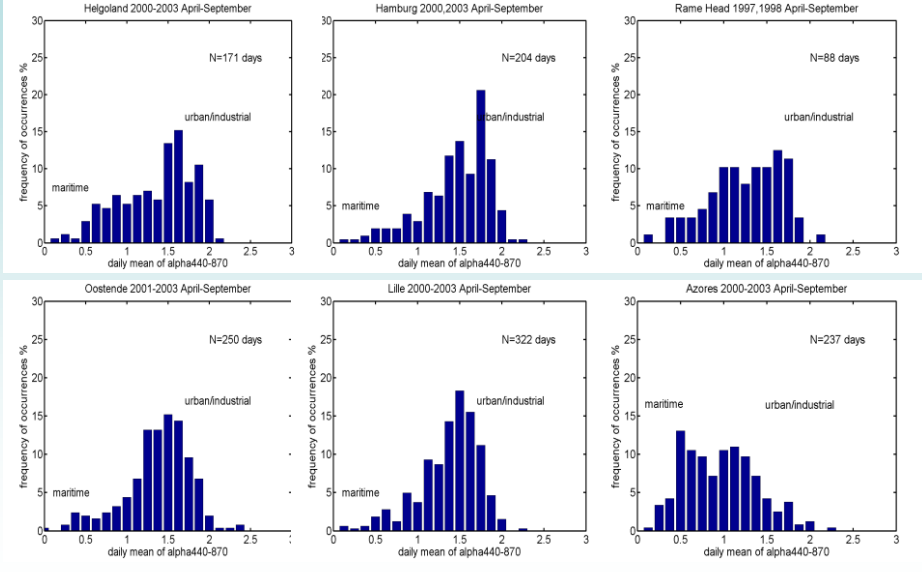




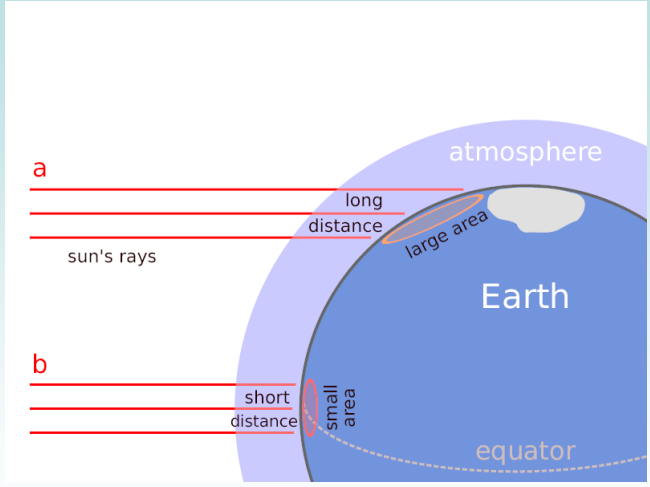
Annual variation of the Aerosol optical depth 500nm



Frequency of occurrences for Angstrom parameter $\alpha 440\_870$



Cosine effect



transmission air-water

Transmission across the air–water interface 45

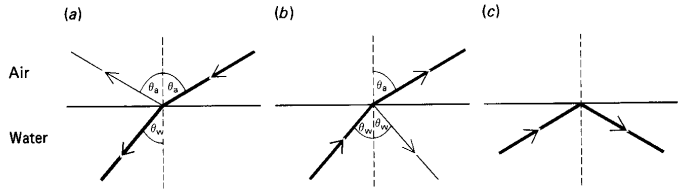


Fig. 2.11. Refraction and reflection of light at air–water boundary. (*a*) A light beam incident from above is refracted downwards within the water: a small part of the beam is reflected upwards at the surface. (*b*) A light beam incident from below at a nadir angle of 40° is refracted away from the vertical as it passes through into the air: a small part of the beam is reflected downwards again at the water–air boundary. (*c*) A light beam incident from below at a nadir angle greater than 49° undergoes complete internal reflection at the water–air boundary.

Snell's Law

$$\frac{\sin\theta_a}{\sin\theta_w} = \frac{n_w}{n_a}$$

$n_w = 1.34$

Specular reflectance

Fresnel's Equation
for unpolarized light

$$r = \frac{1}{2}\left(\frac{\sin^2(\theta_a - \theta_w)}{\sin^2(\theta_a + \theta_w)} + \frac{\tan^2(\theta_a - \theta_w)}{\tan^2(\theta_a + \theta_w)}\right)$$

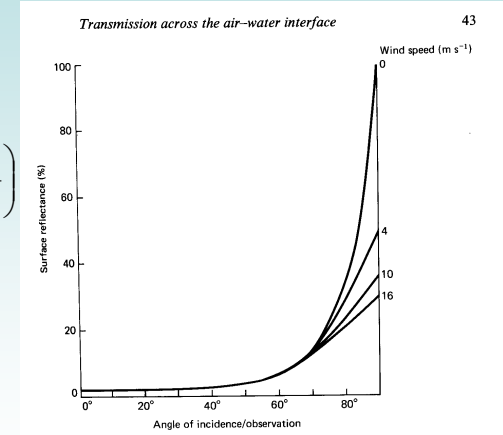


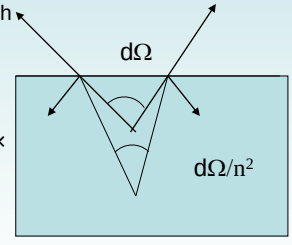
Fig. 2.10. Reflectance of water surface as a function of zenith angle of light (incident from above), at different wind speeds (data of Gordon, 1969; Austin, 1974*a*).

Remote Sensing Reflectance

For comparison with the satellite-sensed signal, it is needed to consider the above-surface remote-sensing reflectance which is the ratio of the upwelling radiance to the downwelling irradiance just above the sea surface

$R_{RS}(\lambda, \theta, \varphi, 0^+) = L_u(\lambda, \theta, \varphi, 0^+) / E_d(\lambda, 0^+)$.

The subsurface upwelling radiance $L_u(0^-)$ passing through the sea surface decreases due to reflection and refraction; the above-surface downwelling irradiance passing through the sea surface decreases due to reflection but it is augmented due to internal reflection of the subsurface upward flux from the sea surface



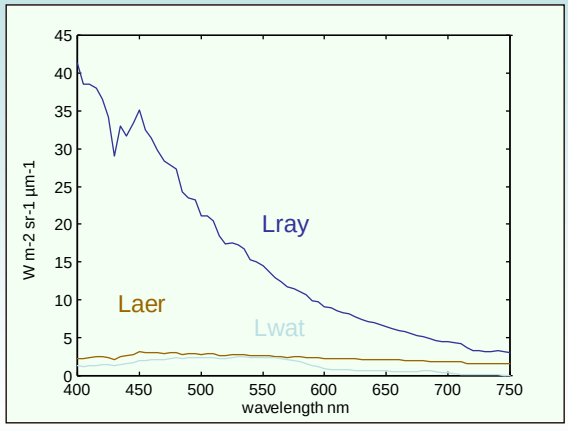
$L_u(0^+) = (t_- / n^2)\, L_u(0^-)$; $E_d(0^-) = t_+ Ed(0^+)/(1-\gamma R) \times$

$R_{RS} = (t_- t_+ / n^2)\, r_{RS} / (1-\gamma R)$; $R_{RS} = \zeta\, r_{RS} / (1-\Gamma\, r_{RS})$;

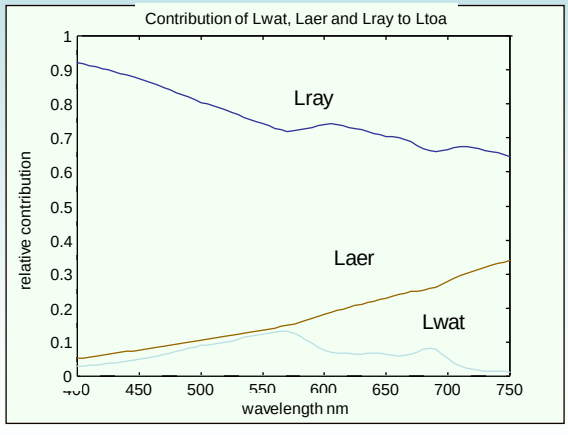
$\zeta = t_- t_+ / n^2 \times$; $\Gamma = \gamma Q$.

For nadir viewing: $\zeta \approx 0.518$, $\Gamma \approx 1.562$, (Lee et al. 1998).

Radiances at Top of Atmosphere (TOA)



The composition of the Radiance Spectrum at Top of Atmosphere



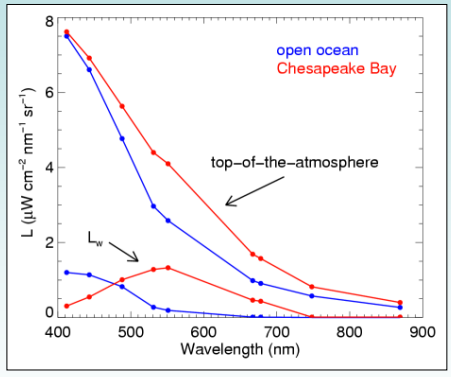
Relative contribution to TOA radiance

water leaving radiance: Lw

aerosol path radiance: Laer

Rayleigh path radiance: Lray

Ocean color



the atmosphere is 80-90% of the total top-of-atmosphere signal in blue-green wavelengths (400-600 nm)

~1% error in instrument calibration or atmospheric model leads to

~10% error in *Lw*(λ)

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Effects of the atmosphere

gaseous absorption (ozone, water vapor, oxygen)

Rayleigh scattering by air molecules

Mie scattering and absorption by aerosols (haze, dust, pollution)

polarization (MODIS response varies with polarization of signal)

Rayleigh (80-85% of total signal)

- small molecules compared to nm wavelength, scattering efficiency decreases with wavelength as λ^{-4}
- reason for blue skies and red sunsets
- can be accurately approximated for a given atmospheric pressure and geometry (using a radiative transfer code)

Aerosols (0-10% of total signal)

- particles comparable in size to the wavelength of light, scattering is a complex function of particle size
- whitens or yellows the sky
- significantly varies and cannot be easily approximated

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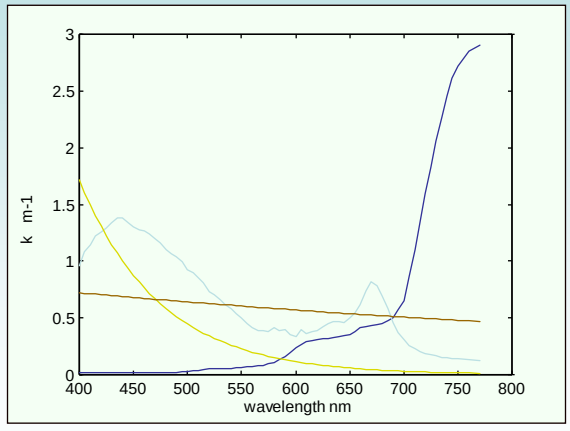
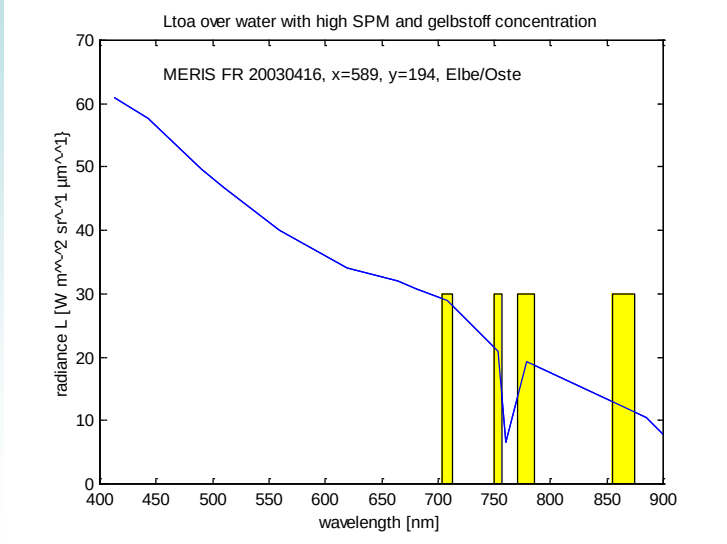
Spectral Attenuation of Water Constituents



Diffuse attenuation k
for pure water, chlorophyll (30µg/l), sus. matter (50 mg/l), gelbstoff

Bands for atmospheric correction



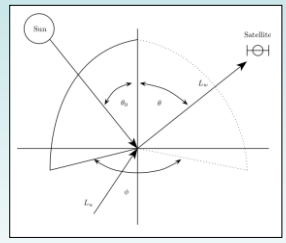
Atmospheric correction

TOA gas pol glint whitecap air aerosol

$t_d(\lambda)\, L_w(\lambda) = L_t(\lambda) / t_g(\lambda) / f_p(\lambda) - TL_g(\lambda) - tL_f(\lambda) - L_r(\lambda) - L_a(\lambda)$

brdf Sun

$nL_w(\lambda) = L_w(\lambda)\, f_b(\lambda) / t_{d0}(\lambda)\, \mu_0\, f_0$



But, <u>we need aerosol</u> to get $L_w(\lambda)$

$L_w(\lambda{=}NIR) \approx 0$ and can be estimated (model extrapolation from VIS) in waters where *Chl* is the primary driver of $L_w(\lambda)$

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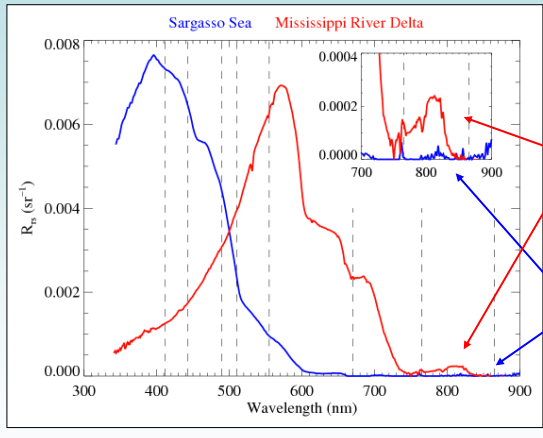
Magnitudes of *Lw*(NIR)



Lw(NIR) ≠ 0 (turbid or highly productive water)

Lw(NIR) = 0 (clear water)

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Aerosol determiniation in visible wavelengths

Given retrieved aerosol reflectance at two λ, and a set of aerosol models fn(θ,θ_0,ϕ).

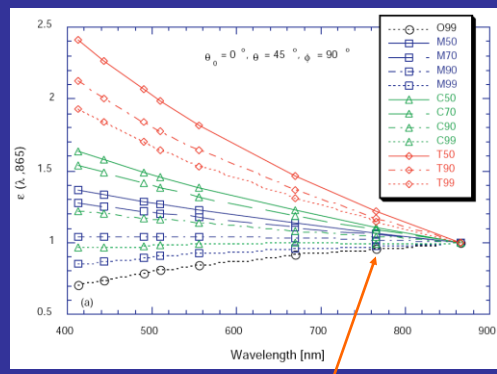
$$\rho = \frac{\pi L}{F_0 \cdot \mu_0}$$

$\rho_a(748)$ & $\rho_a(869)$

$\rho_a(\text{NIR}) \overset{\text{model}}{\Rightarrow} \rho_{as}(\text{NIR})$

$$\varepsilon\,(748,869) = \frac{\rho_{as}(748)}{\rho_{as}(869)}$$

$$\varepsilon\,(\lambda,869) = \frac{\rho_{as}(\lambda)}{\rho_{as}(869)}$$



$\varepsilon\,(748,869)$

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Iterative correction for non-zero *Lw*(NIR)

(1) assume Lw(NIR) = 0
(2) compute La(NIR)
(3) compute La(VIS) from La(NIR)
(4) compute Lw(VIS)
(5) estimate Lw(NIR) from Lw(VIS) + model
(6) repeat until Lw(NIR) stops changing

iterating up to 10 times

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Level-2 ocean color processing

(1) determine atmospheric and surface contributions to total radiance at TOA and subtract, iterating as needed.

(2) normalize to the condition of Sun directly overhead at 1 AU and a non-attenuating atmosphere (nLw or Rrs = nLw/F_0).

(3) apply empirical or semi-analytical algorithms to relate the spectral distribution of nLw or Rrs to geophysical quantities.

(4) assess quality (set flags) at each step

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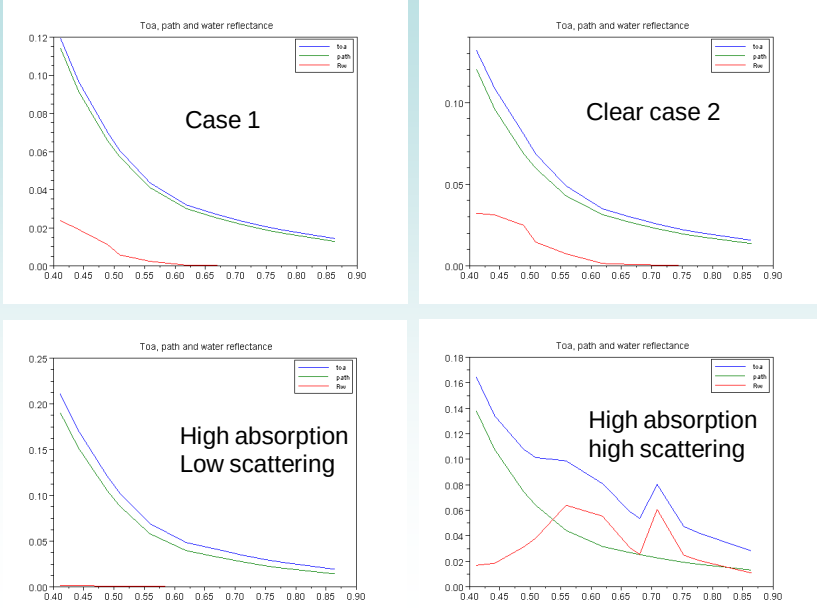
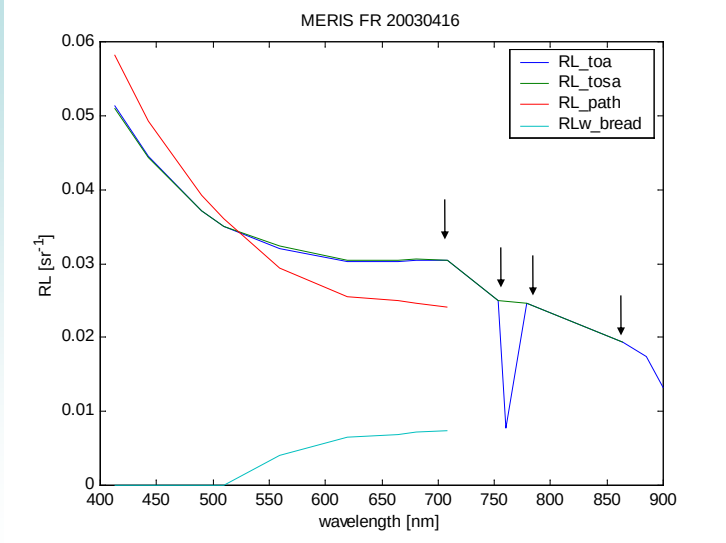
Case 2 water reflectance spectra (simulations)

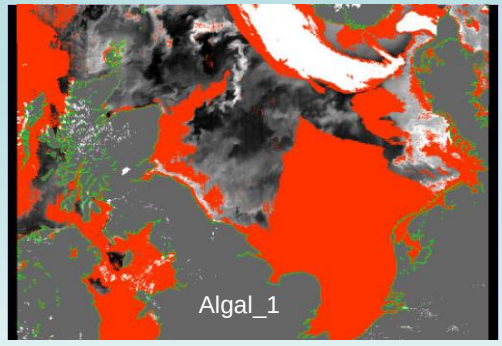


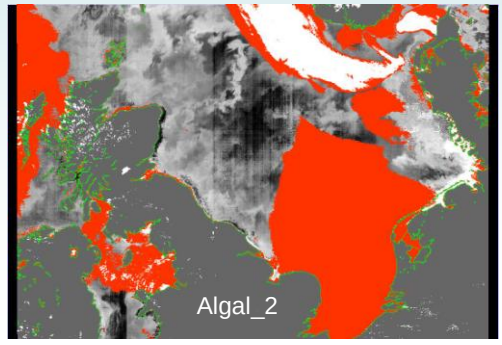
Strange Spectra producing negative reflectances

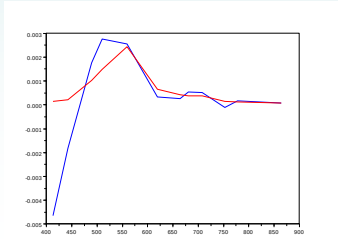


Main Problems

- Atmospheric correction often not sufficient for all 9 bands
- Partly bands 1-2 negativ, or bands 7,8,9 noisy (in case 1 water)

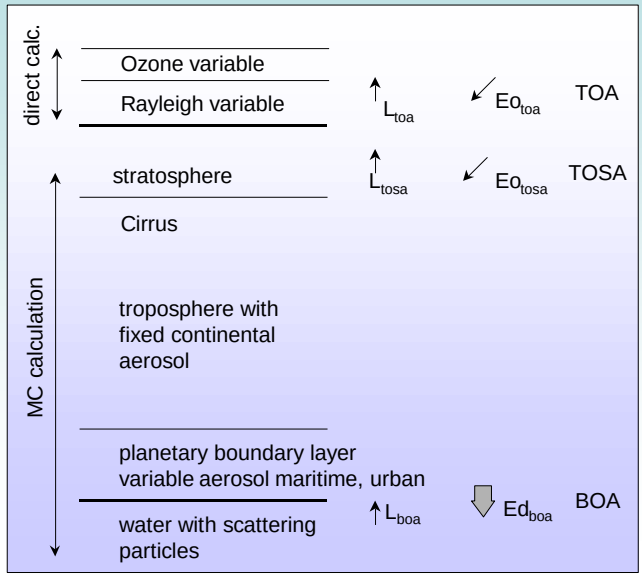




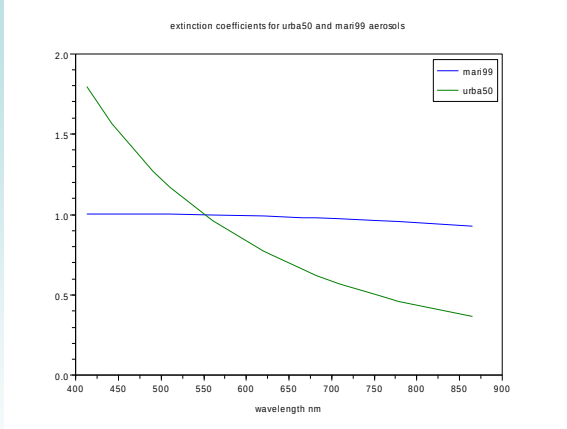


Design of a Model Atmosphere

Model Atmosphere



Attenuation coefficients of boundary layer aerosols



Attenuations coefficients of maritime and urban aerosols, normalized at 550 nm. The maritime aerosol has an angstroem coefficient of – 0.065, the urban aerosol of - 2.15

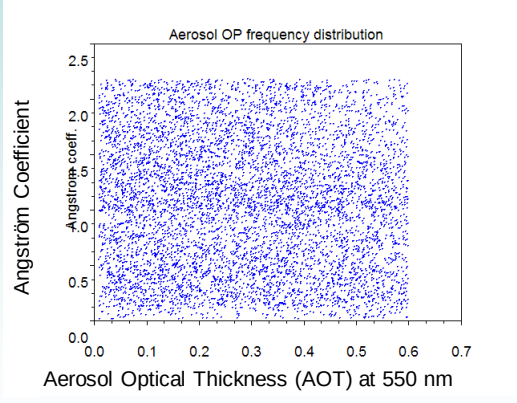
Variables of the atmosphere

Aerosol	***Layer [km]***	***Optical thick-ness at 550 nm***	***Max mean extinction at 550 nm [km^{-1}]***
maritime (99 % rel. hum.)	0 - 3 km	0 - 0.6	0.2
urban (45% rel. hum.)	0 - 3 km	0 - 0.6	0.2
continental	2-12	0 – 0.6	0.165
cirrus	8-11	0 - 0.6	0.2
stratosphere	12-50	0 - 0.003	0.0000625
max. optical thickness		0.6	

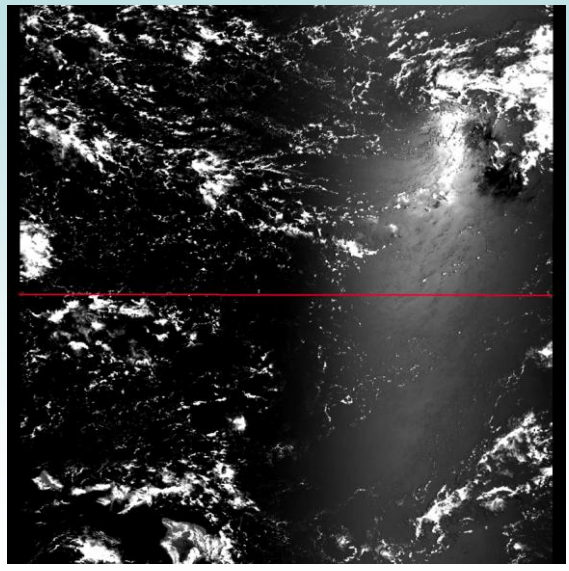
Max optical thickness at 550 nm: 0.6

Wind speed 0 – 8 m/s

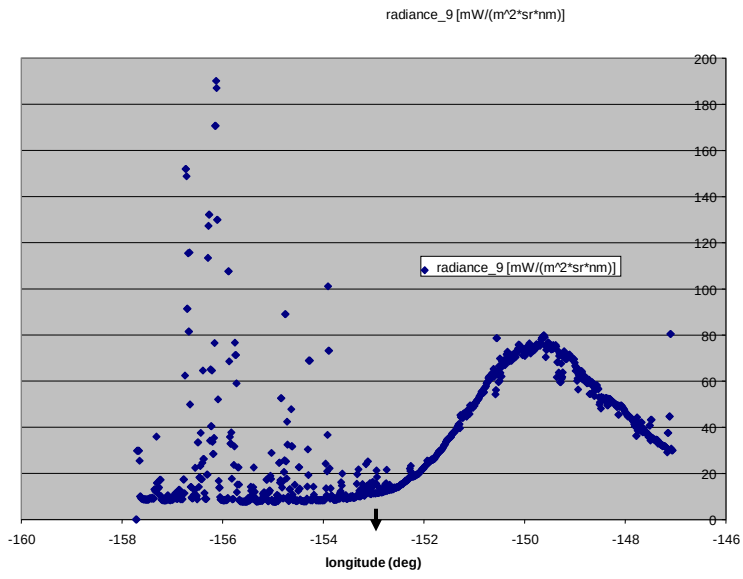
Aerosol Optical Properties used for NN Training data set



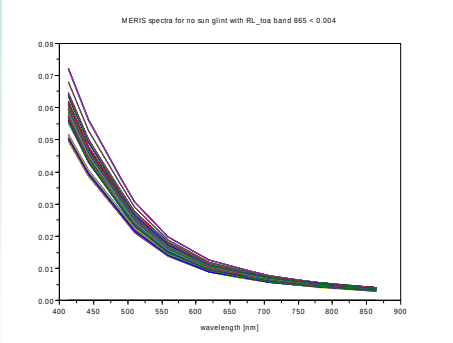
Sun glint problem: Hawai 20030705

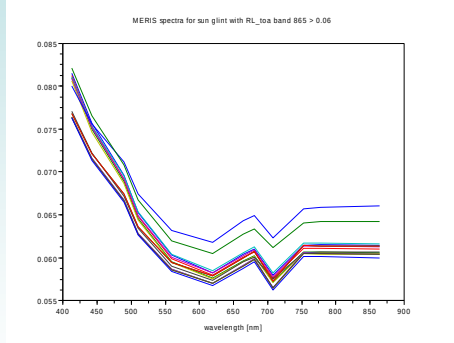


Cross section Hawai scene

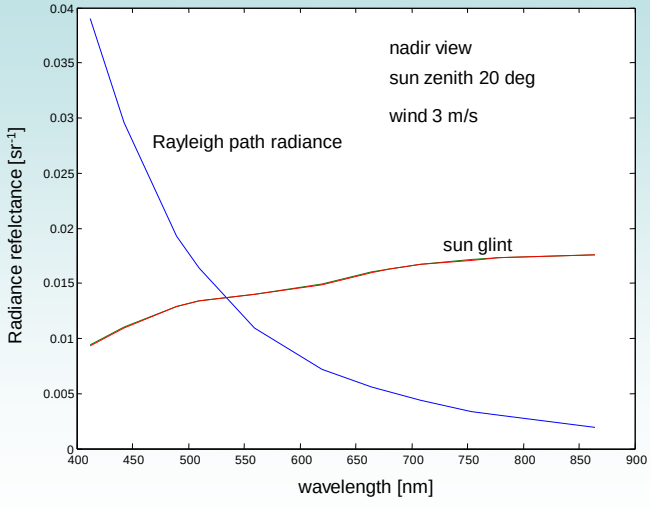


No glint and high glint TOA reflectance spectra

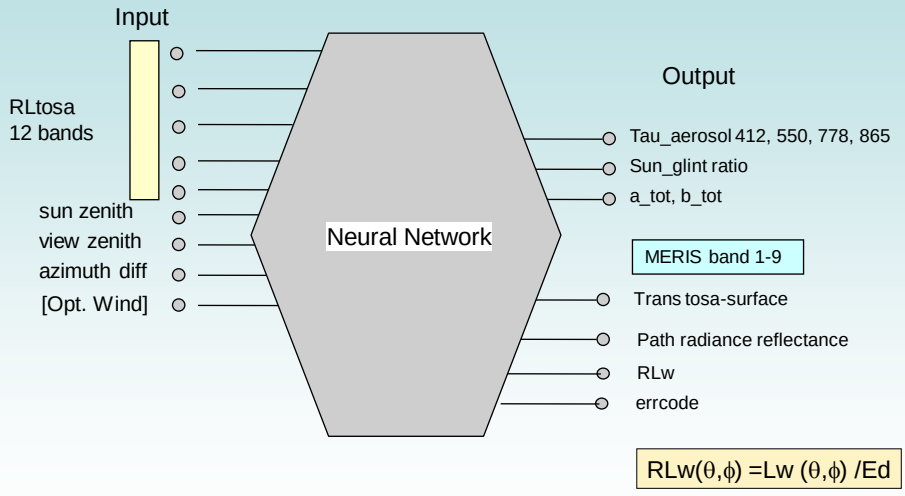




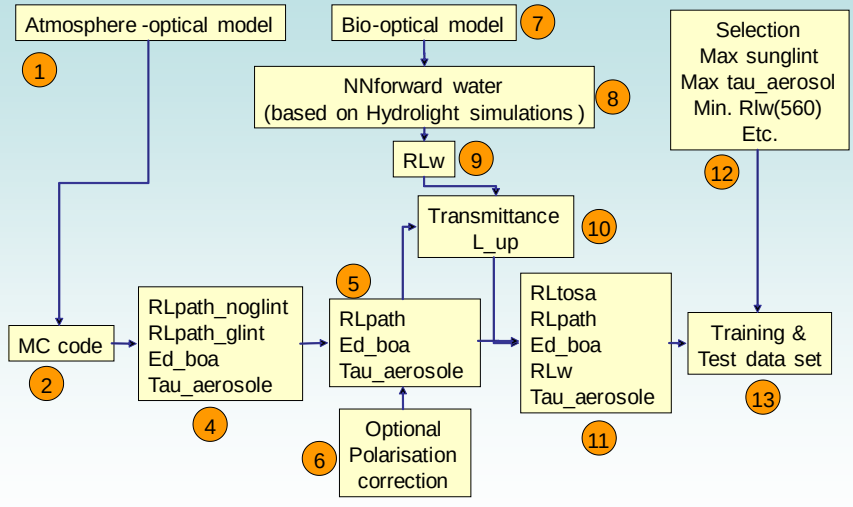
Simulated Rayleigh path radiance reflectance and sun glint radiance reflectance



NN for atmospheric correction - 3rd version in C2R and Glint processor



Training of a neural network for atmospheric correction



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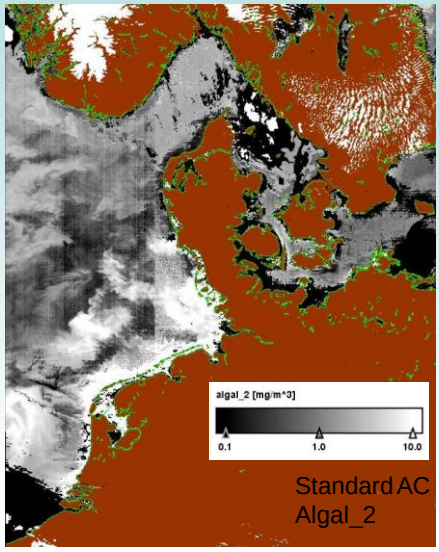
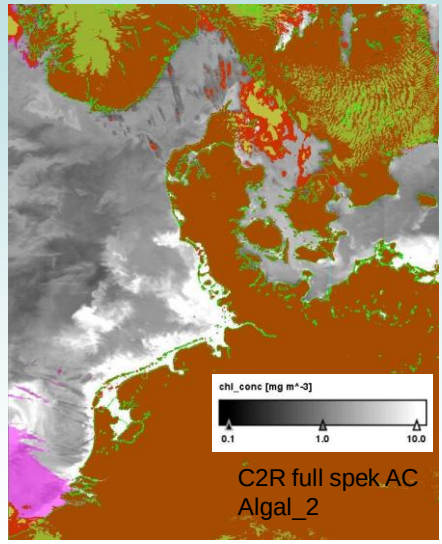
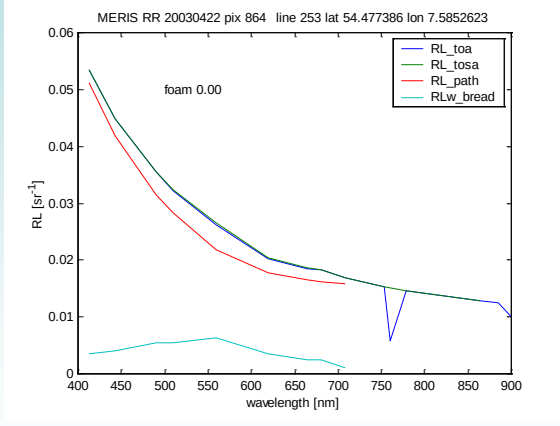


Standard AC
Algal_2

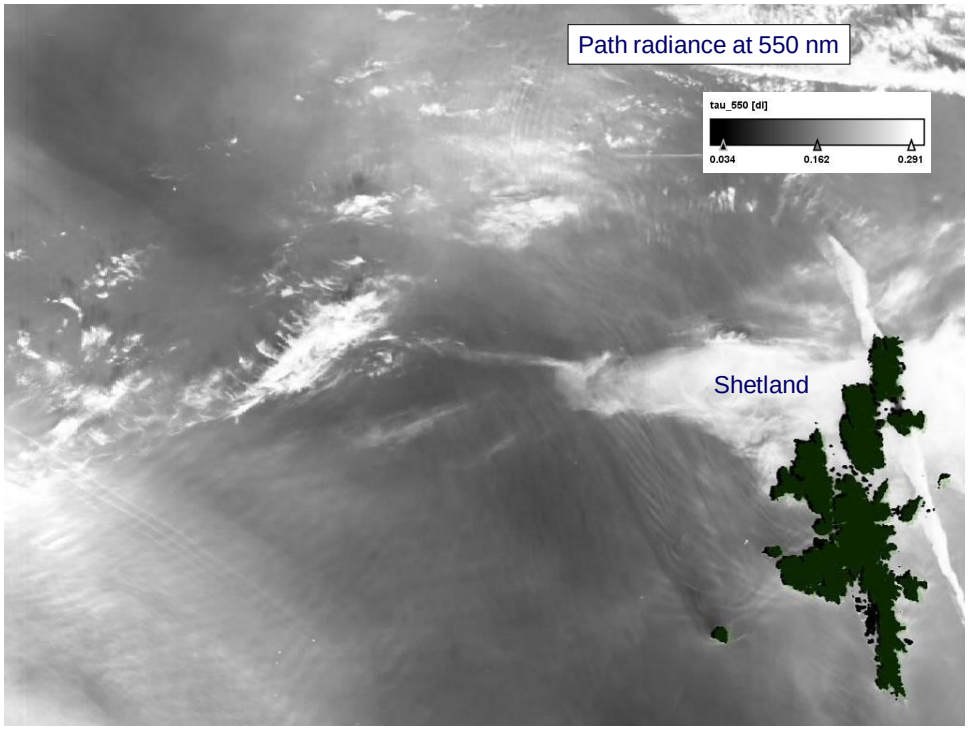


C2R full spek AC
Algal_2

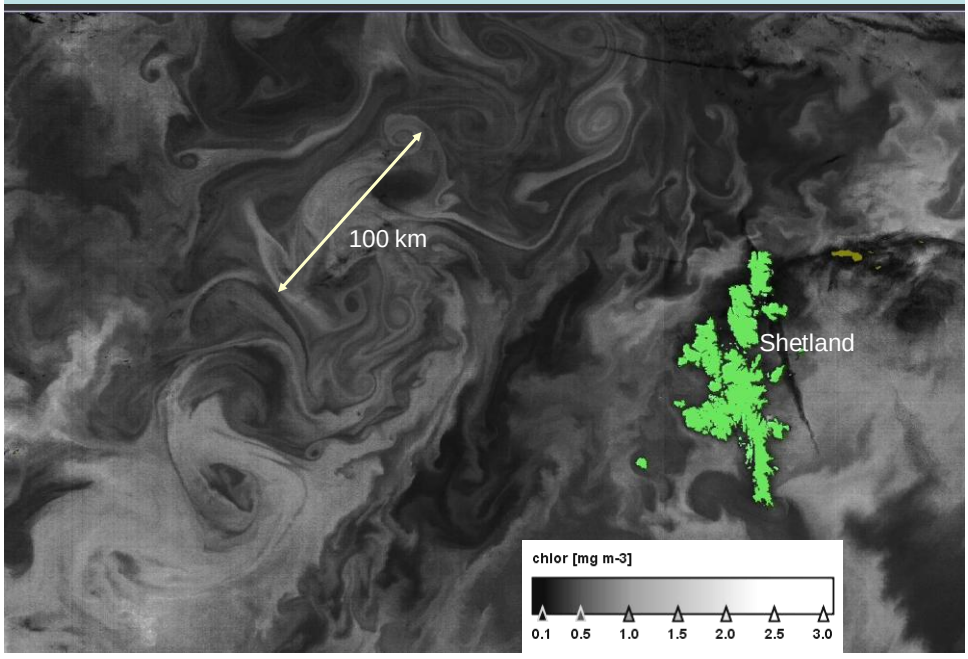
Radiance reflectance: TOA, path, RLw



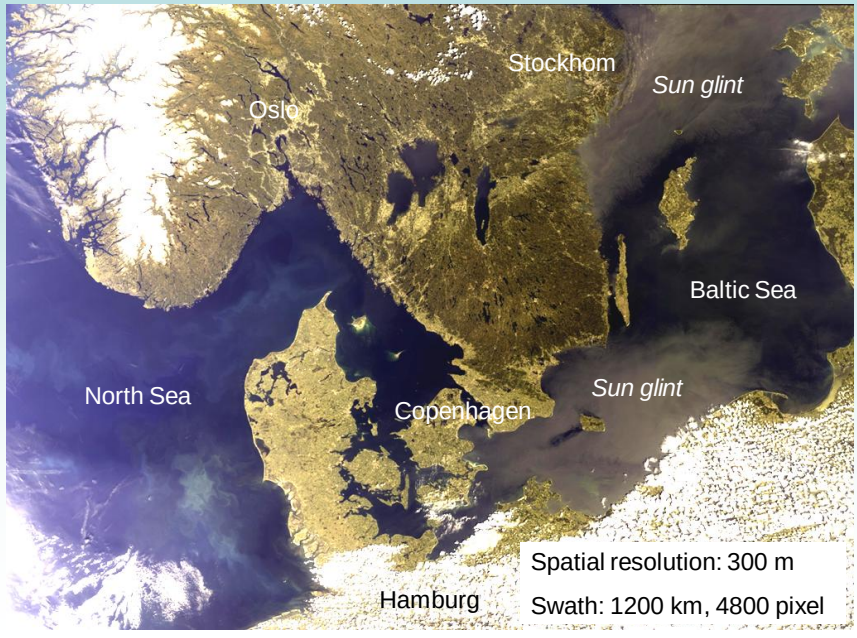

TOA
Radiance reflectance
RLw RGB
Shetland


Path radiance at 550 nm
tau_550 [dl]
0.034
0.162
0.291
Shetland

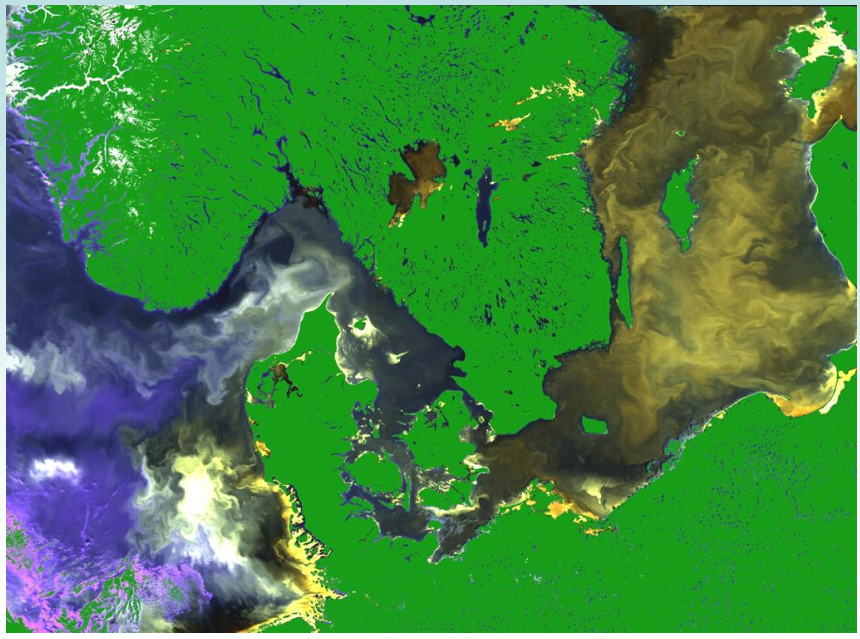
C2R Chlorophyll (MERIS FR)



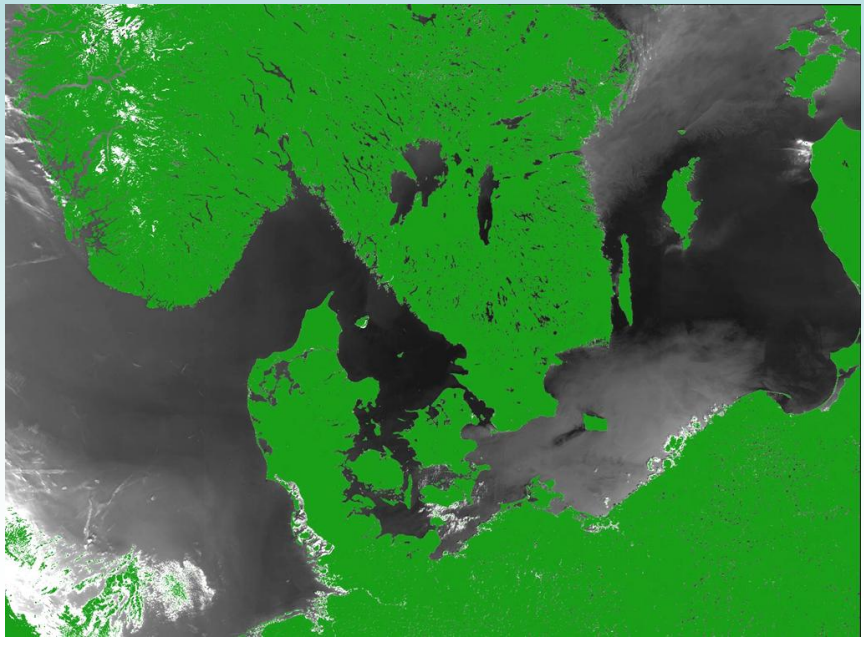
MERIS full resolution: Baltic and North Sea, 20080606



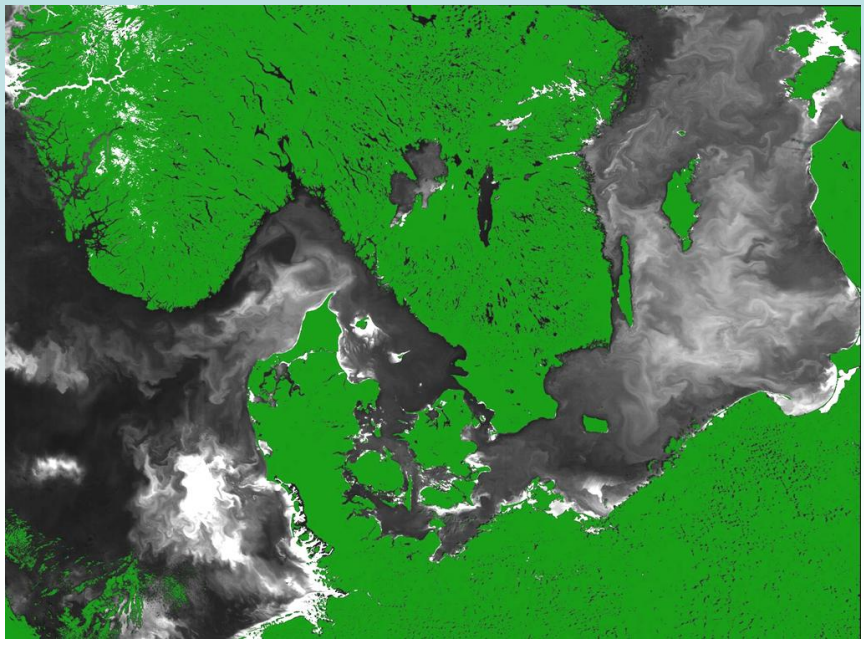
Water leaving radiance reflectance



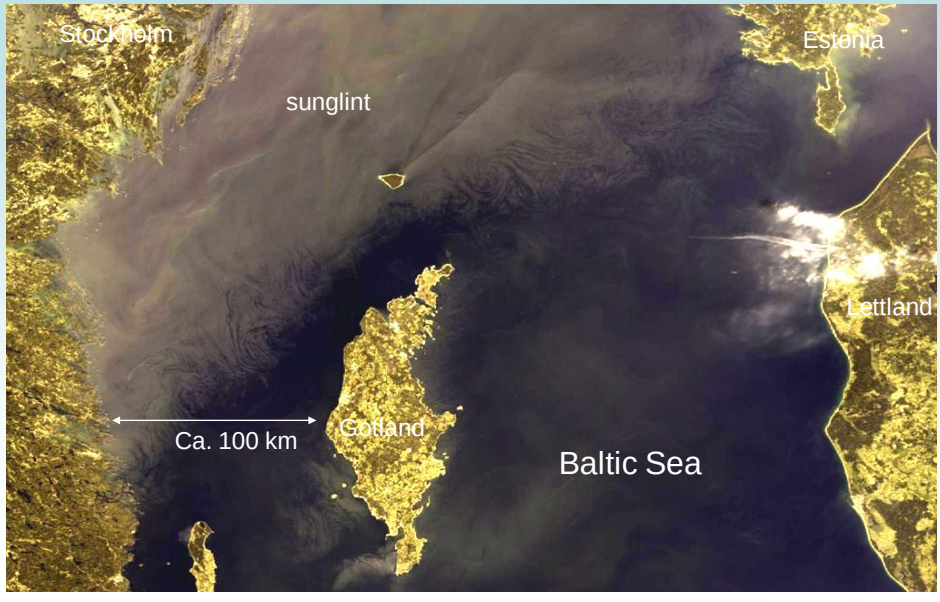
Path radiance reflectance incl. Sun glint band 5 (560 nm)



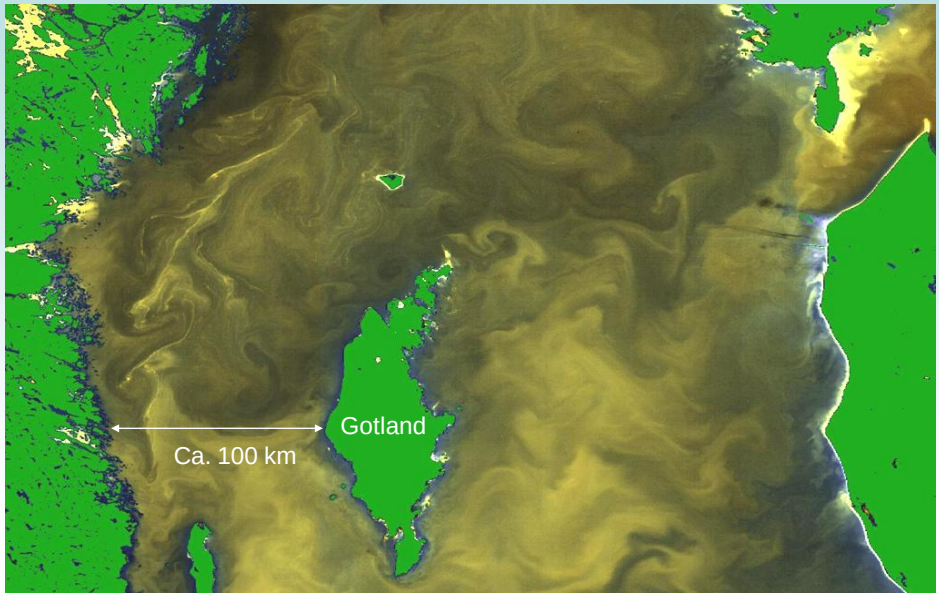
Water leaving radiance reflectance band 5 (560 nm)



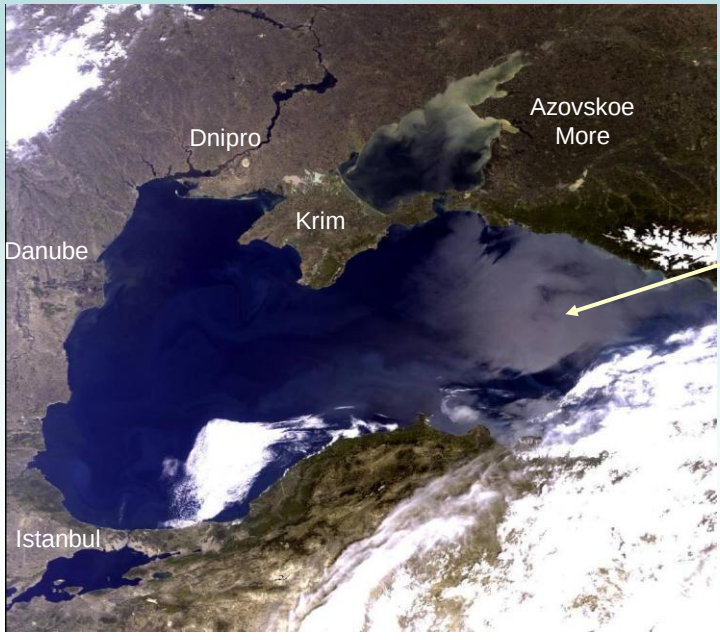
MERIS FR, Area of Gotland, TOA RLw RGB



Water leaving radiance reflectance RGB



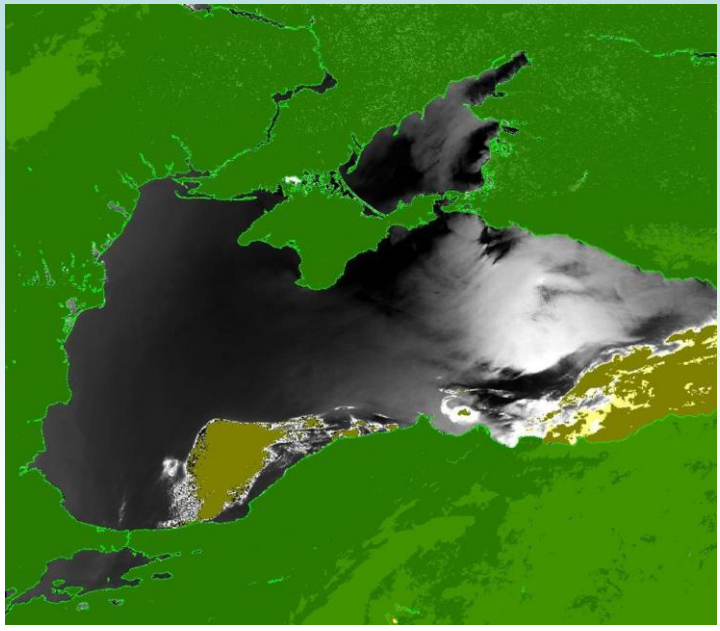
Black Sea MERIS RR May 12 2007



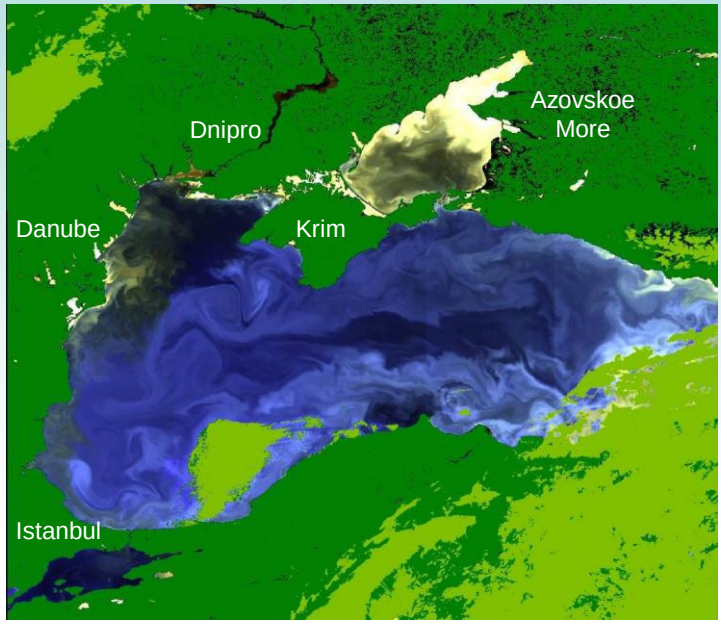
Sun glint

Full swath
RR 1200 m
RL_toa

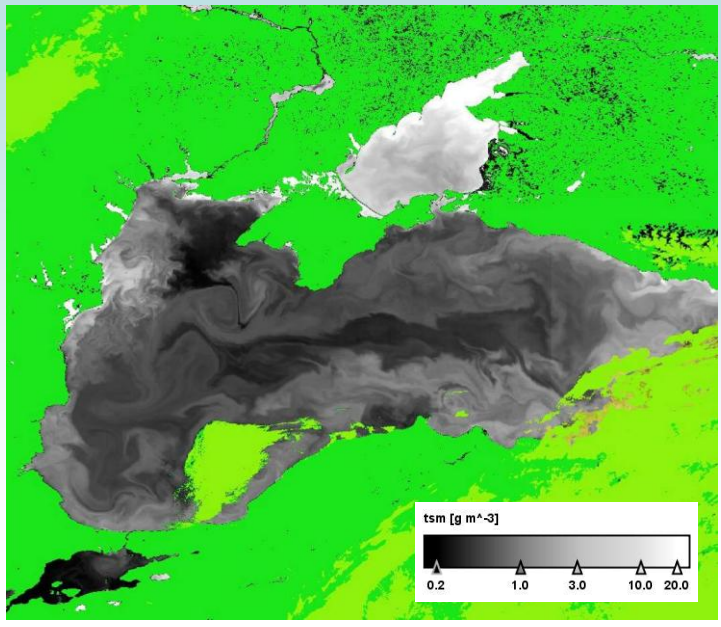
Path radiance reflectance band 5 (560 nm)



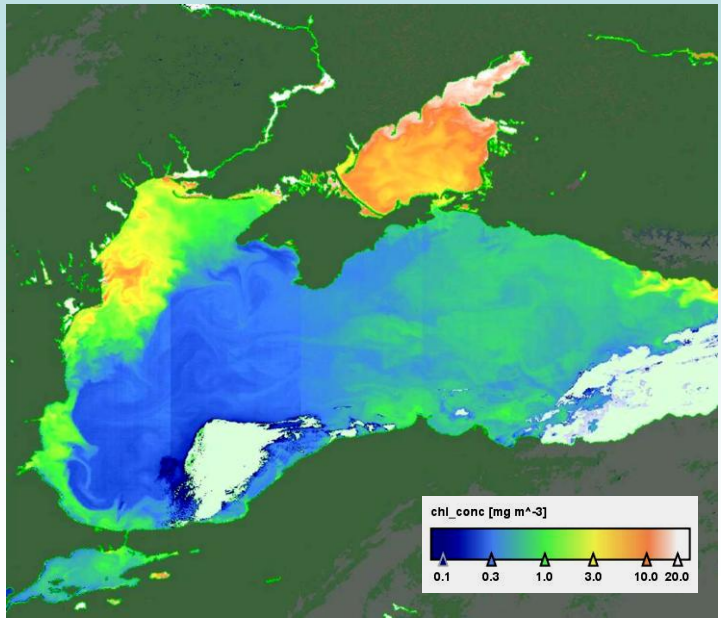
MERIS RR 20070512 water leaving radiance reflectance



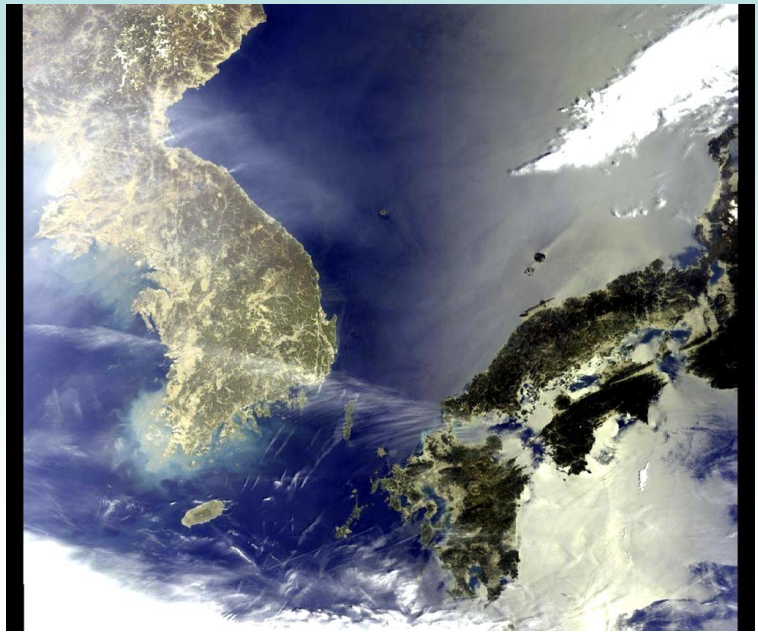
Total Suspended Matter



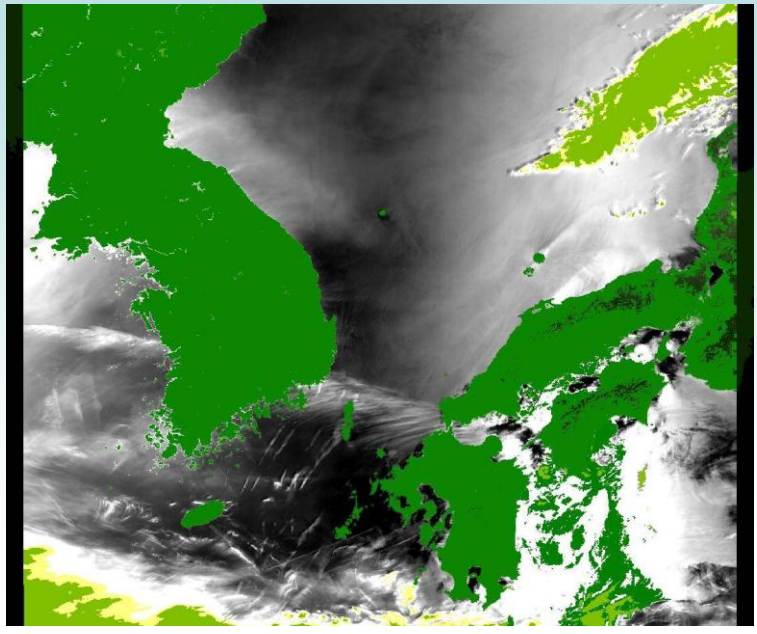
Chlorophyll distribution



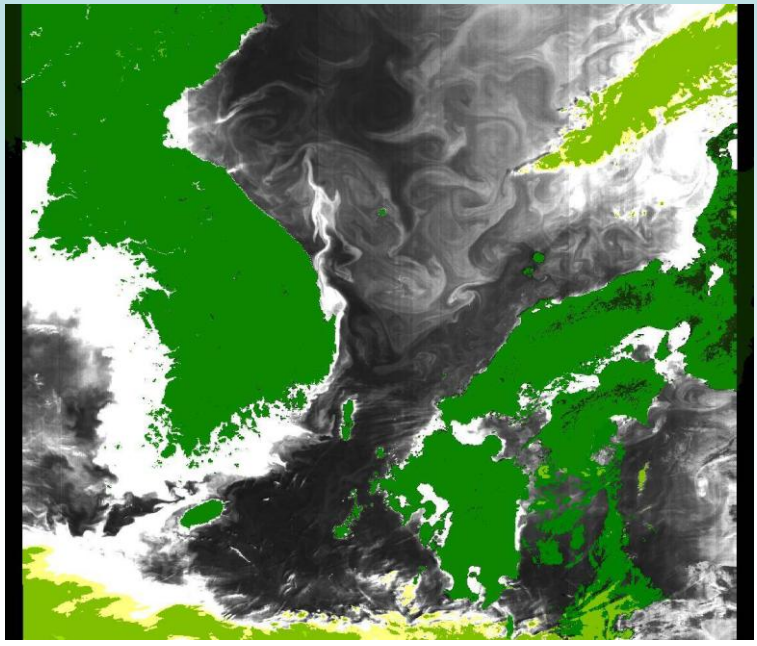
MERIS 20070429 LTOA RGB



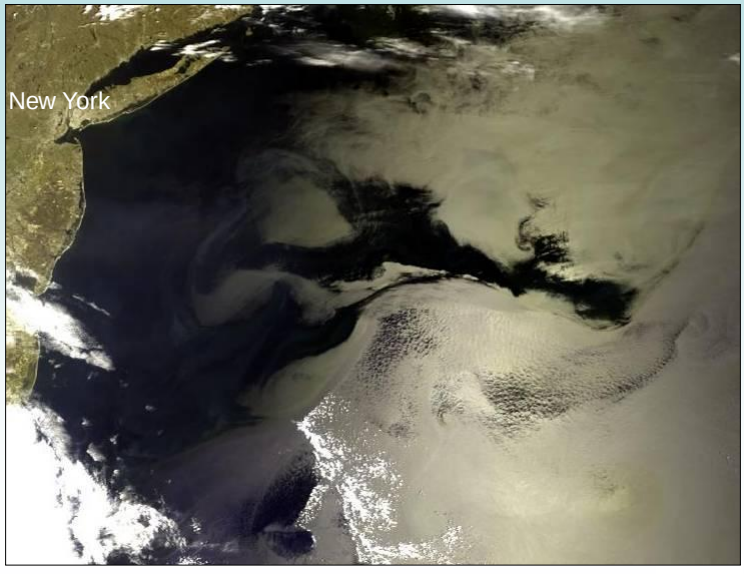
C2R AGC Path radiance reflectance band 5



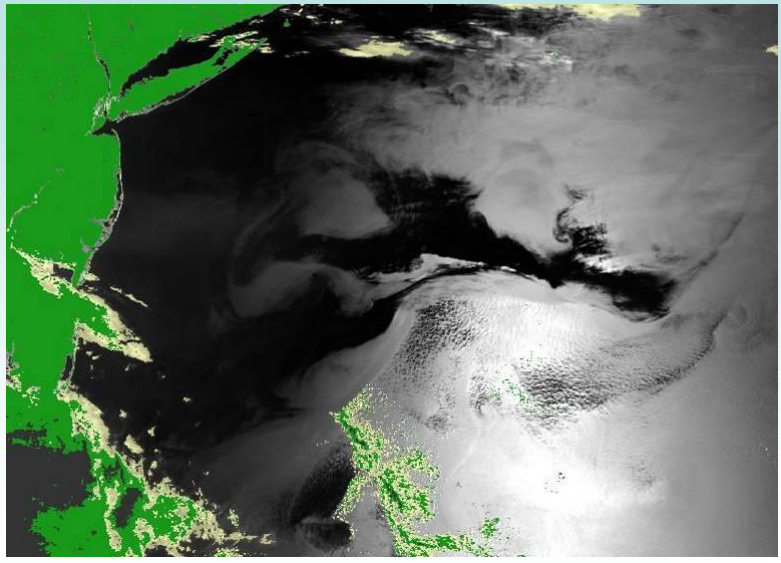
Water leaving radiance reflectance band 5



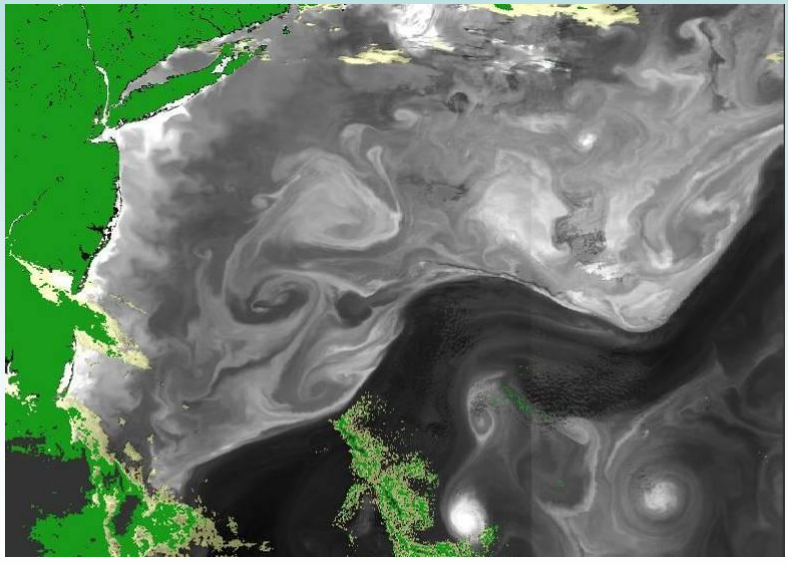
MERIS 20070505: TOA reflectances RGB



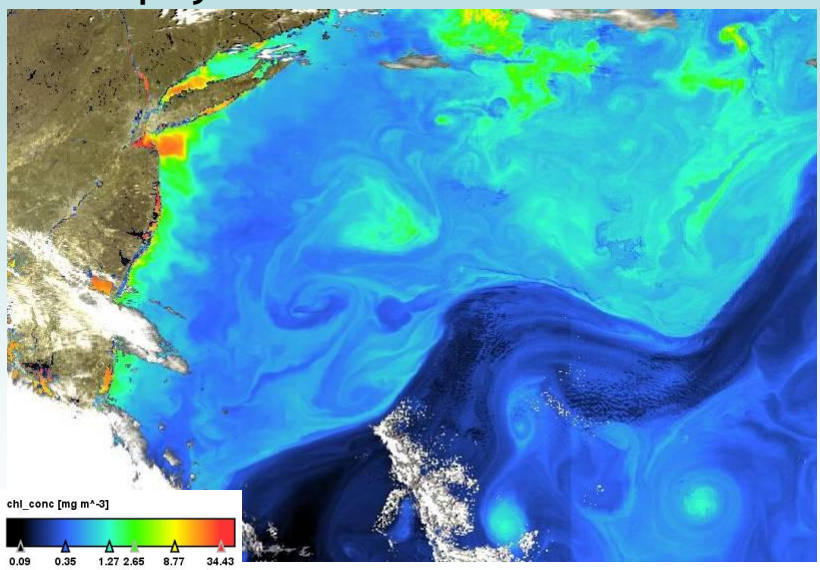
reflectance RLpath MERIS band 5 (560 nm)



reflectance RLW MERIS band 5 (560 nm)



Chlorophyll



MERIS FR USA East Coast 12.6.2008, Signal depth z90