

David ANTOINE	SENTINEL-3 OPTICAL PRODUCTS AND ALGORITHM DEFINITION OLCI Level 2 Algorithm Theoretical Basis Document Atmospheric corrections over Case 1 waters	Ref: S3-L2-SD-03- C07-LOV-ATBD Issue: 2.2 Date: July 13, 2010 Page:1
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OLCI Level 2

Algorithm Theoretical Basis Document

Atmospheric corrections over Case 1 waters

(“Clear Waters Atmospheric Corrections” or “CWAC”)

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1. INTRODUCTION

1.1 Purpose and scope

This Algorithm Theoretical Basis document (ATBD) is written for the Ocean and Land Colour Imager (OLCI) of the Earth Observation Mission SENTINEL 3 of the European Space Agency (ESA).

The purpose of this document is to lay out the OLCI atmospheric correction algorithm above Case 1 waters (see Fig. 1). As much as possible, basic principles of the algorithm and the description of its various segments will refer to publications in the peer-reviewed scientific literature or to previous ATBDs. When such literature exists, minimum information will be provided here for the sake of clarity, and the reader will be referred to the relevant literature for further information.

The description of a detailed implementation of the algorithm is out of scope here. Most of the MERIS ATBD 2.7, issue 5.0, is valid in the context of OLCI, in particular the sensitivity studies to non-nominal conditions of operation (residual glint, unidentified turbid waters, cirrus etc...; sections 3.1.1.6 and 3.1.3.2 of the MERIS ATBD 2.7, issue 5.0).

Note: the radiometric quantities are expressed here in terms of reflectance, which is related to the radiance through (see symbols and definitions)

$$\rho(\lambda, \theta_s, \theta_v, \Delta\phi) = \pi L(\lambda, \theta_s, \theta_v, \Delta\phi) / F_0(\lambda) \mu_s \quad (1)$$

1.2 Acronyms

ENVISAT	Environmental Satellite
ESA	European Space Agency
LOV	Laboratoire d'Océanographie de Villefranche
MERIS	Medium Resolution Imaging Spectrometer
NASA	National Aeronautics & Space Administration
nLw	Normalized Water-leaving radiance
OC	Ocean Color
OLCI	Ocean and Land Color Imager
SeaWiFS	Sea-viewing Wide Field-of-view Sensor
Sentinel-3	Third series of "sentinel" (ESA satellites)
SWIR	Short Wave Infra Red
TOA	Top of Atmosphere

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UV Ultra Violet

1.3 Symbols

Symbol	definition	Dimension / units
Geometry, wavelengths		
λ	Wavelength	nm
θ_s	Sun zenith angle ($\mu_s = \cos(\theta_s)$)	degrees
θ_v	Satellite viewing angle ($\mu_v = \cos(\theta_v)$)	degrees
$\Delta\phi$	Azimuth difference between the sun-pixel and pixel-sensor half vertical planes	degrees
Atmosphere and aerosol properties		
$F_0(\lambda)$	Mean extraterrestrial spectral irradiance	$W m^{-2} nm^{-1}$
ε_c	Correction factor applied to $F_0(\lambda)$, and accounting for the changes in the Earth-sun distance. It is computed from the eccentricity of the Earth orbit, $e = 0.0167$, and from the day number D , as	dimensionless
$F_a(\lambda)$	Aerosol forward scattering probability	dimensionless
$F_r(\lambda)$	Rayleigh forward scattering probability (= 1/2)	dimensionless
$\tau_a(\lambda)$	Optical thickness due to aerosol scattering	dimensionless
$\tau_r(\lambda)$	Optical thickness due to Rayleigh scattering	dimensionless
η_r	Contribution of molecules to the total optical thickness (= $\tau_r / (\tau_r + \tau_a)$)	dimensionless
$\tau_{ag}(\lambda)$	Optical thickness due to gaseous absorption	dimensionless
$\omega_a(\lambda)$	Aerosol single scattering albedo	dimensionless
$\omega_r(\lambda)$	Rayleigh single scattering albedo	dimensionless
$P_r(\lambda, \gamma)$	Rayleigh phase function $p_r(\gamma_{\pm}) = P_r(\lambda, \gamma_{\pm}) + [\rho_F(\theta_s) + \rho_F(\theta_v)] P_r(\lambda, \gamma_{\pm})$ where γ_{\pm} is the scattering angle $c \quad \gamma_{\pm} = \pm c s \theta_0 \cos(\theta_v) - s (\theta_0) s (\theta_v) c \Delta\phi$	sr^{-1}
$P_a(\lambda, \gamma)$	Aerosol phase function $p_a(\gamma_{\pm}) = P_a(\lambda, \gamma_{\pm}) + [\rho_F(\theta_s) + \rho_F(\theta_v)] P_a(\lambda, \gamma_{\pm})$	sr^{-1}
ν	Exponent of the Junge law for the distribution of aerosol particles (sensitivity studies)	dimensionless
$c(\lambda)$	Attenuation coefficient for wavelength λ	m^{-1}
$t_{\theta_s}(\lambda, \theta_s)$	Irradiance transmittance for a sun zenith angle θ_s	dimensionless

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	$t_{\theta_s}(\lambda, \theta_s) = E_d(0^+) / (\mu_s \varepsilon_c F_0)$, where $E_d(0^+)$ is the ,	
	where $E_d(0^+)$ is the downward irradiance just above the sea surface	
$td(\lambda, \theta)$	Diffuse transmittance for angle θ	dimensionless
	$td(\lambda, \theta) = L_{TOA}(\lambda, \theta_s, \theta_v, \Delta\phi) / L_{0+}(\lambda, \theta_s, \theta_v, \Delta\phi)$	
$L(\lambda, \theta_s, \theta_v, \Delta\phi)$	Radiance	$W m^{-2} nm^{-1} sr^{-1}$
P	Atmospheric pressure at sea level (subscript 0 for the standard value of 1013.15 hPa).	hPa
RH	Relative humidity	percents
$\rho(\lambda, \theta_s, \theta_v, \Delta\phi)$	Reflectance ($\pi L / F_0 \mu_s$)	dimensionless
	where the product $\pi.L$ is the TOA upwelling irradiance if upwelling radiances are equal to $L(\lambda, \theta_s, \theta_v, \Delta\phi)$, for any values of θ within $0-\pi/2$ and any $\Delta\phi$ within $0-2\pi$.	
	Subscripts t : total reflectance	
	w : water-leaving reflectance	
	path : path reflectance	
	r : Rayleigh reflectance	
	rs: Rayleigh reflectance (single scattering only)	
	a: aerosol reflectance	
	as: aerosol reflectance (single scattering only)	
	ra: heterogeneous aerosol-molecule scattering	
	G : sun glint reflectance	
$\rho^*(\lambda, \theta_s, \theta_v, \Delta\phi)$	Reflectance within a compound atmosphere, containing molecules and aerosols (subscripts as for ρ)	dimensionless
$\varepsilon(\lambda_1, \lambda_2)$	Ratio $\rho_{as}(\lambda_1) / \rho_{as}(\lambda_2)$	dimensionless
$\varepsilon'(\lambda_1, \lambda_2)$	Ratio $[\rho_{path}(\lambda_1) - \rho_r(\lambda_1)] / [\rho_{path}(\lambda_2) - \rho_r(\lambda_2)]$	dimensionless
$f(\tau_a)$	Relationship between the ratio $[\rho_{path} / \rho_r]$ and τ_a	dimensionless
Water properties		
Chl	Chlorophyll concentration	$mg m^{-3}$
$a(\lambda)$	Total absorption coefficient	m^{-1}
$bb(\lambda)$	Total backscattering coefficient	m^{-1}
λ_1, λ_2	Ratio of (bb/a) at λ_1 nm to (bb/a) at λ_2 nm	dimensionless
$R(\lambda, 0^-)$	Diffuse reflectance at null depth, or irradiance ratio (E_u / E_d) (upward and downward irradiances, respectively)	dimensionless
f	Ratio of $R(0^-)$ to (bb/a) ; subscript 0 when $\theta_s = 0$	dimensionless
$Q(\lambda, \theta_s, \theta_v, \Delta\phi)$	Factor describing the bidirectional character of $R(\lambda, 0^-)$ subscript 0 when $\theta_s = \theta_v = 0$	sr
$[\rho_w]N(\lambda)$	Normalised water-leaving reflectance (<i>i.e.</i> , the reflectance if there were no atmosphere, and for $\theta_s = \theta_v = 0$)	dimensionless

Air-water interface

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$\mathfrak{R}(\theta)$	Geometrical factor, accounting for all refraction and reflection effects at the air-sea interface (Morel and Gentili, 1996)	dimensionless
	$\mathfrak{R}(\theta') = \left[\frac{(1 - \bar{\rho})}{(1 - \bar{r}R)} \frac{(1 - \rho_F(\theta'))}{n^2} \right]$ (subscript 0 when $\theta = 0$)	
	where	
	n is the refractive index of water	dimensionless
	$\rho_F(\theta)$ is the Fresnel reflection coefficient for incident angle θ	dimensionless
	$\bar{\rho}$ is the mean reflection coefficient for the downwelling irradiance at the sea surface	dimensionless
	\bar{r} is the average reflection for upwelling irradiance at the water-air interface	dimensionless
	θ is the refracted viewing angle ($\theta = \sin^{-1}(n \cdot \sin(\theta_v))$)	degrees
σ	Root-mean square of wave facet slopes	dimensionless
β	Angle between the local normal and the normal to a wave facet	
p	Probability density of surface slopes for the direction ($\theta_s, \theta_v, \Delta\phi$)	dimensionless
Miscellaneous		
W	Wind speed	dimensionless
X	Aerosol mixing ratio defined on the basis of the $[\rho_{\text{path}} / \rho_r]$ ratio at 775 nm.	dimensionless

1.4 Algorithm identification

This algorithm is identified under reference "SD-03-C7" in the Sentinel-3 OLCI documentation.

In this document, it will be referred to as "CWAC", standing for "Clear-Water Atmospheric Corrections".

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2. ALGORITHM OVERVIEW

2.1 Objectives (*qualitatively speaking*)

The objective of atmospheric correction of the multispectral total reflectances measured by an ocean colour sensor at the Top-Of-the-Atmosphere (TOA), $\rho_t(\lambda)$, is to determine as accurately as possible the part of the total signal that is due to atmospheric scattering and to light reflected by the ocean surface at the exception of direct sun glint (supposedly removed before atmospheric correction is performed). The sum of these contributions is referred to as the “path reflectance”, $\rho_{\text{path}}(\lambda)$. The first step is, therefore, to determine $\rho_{\text{path}}(\lambda)$ and to subtract it from $\rho_t(\lambda)$ to get the signal originating from the ocean after transmission through the atmosphere. The last step is to determine the diffuse transmittance of the atmosphere in order to get the reflectance at sea level, $\rho_w(\lambda)$.

Atmospheric correction is not specifically conceived (designed) to provide aerosol parameters (this is the domain of “aerosol remote sensing”). It is designed to determine $\rho_{\text{path}}(\lambda)$. It is nevertheless feasible to derive the aerosol optical thickness (AOT or τ_a) in the near infrared from the aerosol model selected by the atmospheric correction (see later on).

The presently proposed algorithm is designed to fulfil this task for OLCI TOA observations taken above oceanic Case 1 waters (*cf.* Box/**Fig 1**). It does not include specific steps needed to perform the same task above turbid Case 2 waters.

2.2 Objectives (*quantitatively speaking*)

The marine reflectance is supposed to be derived with a 5% uncertainty in the blue for clear Case-1 waters (low pigment concentration).

This requirement was first expressed by Gordon (1987; 1988; 1997), and was translated in terms of maximum reflectance errors in different bands specifically in the frame of the development of the MERIS atmospheric corrections (Antoine and Morel, 1999).

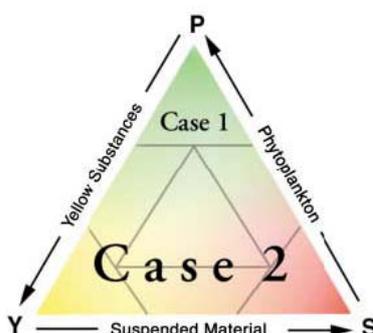
These requirements were set up in order to make possible the discrimination of 10 classes of Chlorophyll (Chl) values within each of the 3 decades of Chl between 0.03 and 30 mg (Chl) m⁻³, and when using a band ratio algorithm to derive Chl. These classes are distributed regularly according to a constant logarithmic increment of 0.1. Shifting from one class to the next (previous) one corresponds to a change in (Chl) by a factor $10^{\pm 0.1}$ (*i.e.*, +25% or -20%).

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The concept of Case 1 and Case 2 waters was introduced by Morel and Prieur (1977) as a binary distinction between optically simple and more complex optical conditions. It provided a basis from which modern marine bio-optics has developed, and onto which the development of ocean colour remote sensing has been built.

Case 1 waters are usually offshore waters (about 95% of the World Ocean), where the optical properties are determined by water itself and by phytoplankton and the ensemble of particles (detritus, bacteria etc.) and dissolved substances associated with them. Optical properties of these waters are commonly indexed on the chlorophyll concentration ([Chl]). This does not mean that the latter is entirely responsible for their changes; [Chl] is used as an index just because it is a ubiquitous pigment, present in all species, and because global relationships have been established between [Chl] and the optical properties. This indexing on a unique component is possible because the other optically-significant components co-vary with it when considering the relationships over the full [Chl] range. This co-variation is much less obvious or even vanishing when looking locally over a small range of [Chl], and much work is presently being done to understand the variability of optical properties around the average, global, relationships established for “typical” Case 1 waters (see, e.g., Siegel et al., 2005; Brown et al., 2008).

As opposed to Case 1 waters, Case 2 waters are those where the co-variation with chlorophyll is no longer valid (or is extremely weak), even when considering a large [Chl] range. These waters are usually coastal waters influenced by river runoffs or sediment resuspensions; in other words, water bodies influenced by land. The multiple combinations of optically-significant components (those present in Case 1 waters plus those coming in addition from land) make their optical properties very complex, and make the inversion of apparent optical properties a difficult task.



The figure shows a triangular representation of Case 1 and Case 2 waters by Prieur and Sathyendranath (1981). The three corners indicate dominance by phytoplankton (P), CDOM (Y), and sediments (S). Case 1 waters are located at the top vertex of this triangle, whereas the rest of its area corresponds to various combinations all belonging to Case 2 waters.

Figure 1: Case 1 and Case 2 waters

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The corresponding requirements were summarized as:

- Atmospheric correction errors must be maintained within $\pm 1-2 \cdot 10^{-3}$ at 443 nm, within $\pm 5 \cdot 10^{-4}$ at 490 nm, and within $\pm 2 \cdot 10^{-4}$ at 560 nm, in order to allow discrimination of 30 reflectance values.
- 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 $\pm 1 \cdot 10^{-3}$ at 443 nm (then $\pm 5 \cdot 10^{-4}$ at 560 nm), or within $\pm 5 \cdot 10^{-4}$ at 490 nm (then $\pm 2 \cdot 10^{-4}$ at 560 nm).

It must be stressed that there are no universal requirements in terms of the acceptable errors for atmospheric corrections. These requirements depend on the subsequent use of the normalized water-leaving reflectances. The above-mentioned requirements remain valid here because the OLCI pigment algorithm for Case 1 waters is also based on a ratio technique. Would the ocean colour products for Case 1 waters be derived by another technique, e.g., a neural network using all bands, these requirements might have to be modified.

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3. ALGORITHM DESCRIPTION

3.1 Baseline: Theoretical Description

3.1.1 The heritage from MERIS

The algorithm proposed here for OLCI is essentially based on the algorithm developed for the MERIS instrument (Antoine and Morel, 1998; 1999). The rationale for this is to ensure the best possible consistency between the two instruments' records (see also comments at the end of section 3). This is fundamental in the present effort towards the construction of "climate-quality data records" (McClain et al., 2006) that should embrace several decades of global and consistent observations. The basic principles and structure of the algorithm are summarised below.

3.1.2 Basic structure of the algorithm

Basic principles for the MERIS atmospheric correction scheme are described in *Antoine and Morel (1998)*, and the practical implementation and tests are developed in *Antoine and Morel (1999)*, as briefly recalled below. In this multiple scattering algorithm, the path reflectance is derived globally to perform the atmospheric correction, instead of assessing separately ρ_a and ρ_r . It is therefore based on the following decomposition of the total reflectance at the top of atmosphere level, ρ_t

$$\rho_t(\lambda) = \rho_{\text{path}}(\lambda) + t_d(\lambda) \cdot \rho_w(\lambda) \quad (2)$$

where ρ_{path} is the atmospheric path reflectance and t_d is the diffuse transmittance along the pixel-to-sensor path (approximated as per *Gordon, et al., 1983*). The reflectance ρ_{path} is formed by all photons reaching the TOA after one or several scattering events in the atmosphere, to the exception of those who entered the ocean.

It was shown that the $[\rho_{\text{path}} / \rho_r]$ ratios are monotonic and unique functions of τ_a for a given aerosol and geometry (*Antoine and Morel, 1998*). Such functions allow an aerosol type to be identified among several predefined models, by using LUTs generated from radiative transfer simulations. These LUTs contain the coefficients of the quadratic relationship between the ratio $[\rho_{\text{path}} / \rho_r]$ and τ_a , for several aerosol models, geometries, and wavelengths. A quadratic expression is used to account for the saturating behaviour of the $[\rho_{\text{path}} / \rho_r]$ versus τ_a relationship for large values of τ_a . For moderate τ_a values ($< \sim 0.5$) a quasi linearity between $[\rho_{\text{path}} / \rho_r]$ and τ_a is actually observed (for non-absorbing aerosols).

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The scheme (see Fig. 2) starts with computing $[\rho_{\text{path}} / \rho_r]$ at two wavelengths in the near IR (λ_{IR1} and λ_{IR2} , where there is no marine signal), where ρ_{path} is measured by the sensor and ρ_r is pre-computed. Because multiple scattering effects depend on the aerosol type, several values of $\tau_a(\lambda_{\text{IR1}})$ correspond to the value of $[\rho_{\text{path}} / \rho_r]$ at λ_{IR1} , each one being associated to a given aerosol model. This set of $\tau_a(\lambda_{\text{IR1}})$ values is converted into the equivalent set at λ_{IR2} , using the spectral attenuation coefficients of each aerosol. Several values of the ratio $[\rho_{\text{path}} / \rho_r]$ at λ_{IR2} correspond to these $\tau_a(\lambda_{\text{IR2}})$ values; they differ according to the aerosol type. Comparing this set of values to the actual $[\rho_{\text{path}} / \rho_r]$ ratio at λ_{IR2} allows the two aerosol models that most closely bracket the actual $[\rho_{\text{path}} / \rho_r]$ ratio to be selected.

A “mixing ratio” is determined as:

$$X = ([\rho_{\text{path}} / \rho_r]^A - [\rho_{\text{path}} / \rho_r]^{A1}) / ([\rho_{\text{path}} / \rho_r]^{A2} - [\rho_{\text{path}} / \rho_r]^{A1}) \quad (3)$$

where the ratios identified by subscripts A, A1 and A2 are respectively the actual ratio (measured at 778 nm) and the two tabulated ratios that most closely bracket $[\rho_{\text{path}} / \rho_r]^A$. X depends on the geometry, aerosol optical thickness and wavelength.

The remaining steps of the algorithm rest on the assumption that the mixing ratio is constant with wavelength (Gordon and Wang, 1994). It is then possible to estimate $[\rho_{\text{path}} / \rho_r]$ for the visible wavelengths from its values at λ_{IR2} and λ_{IR1} , provided that the relationships with τ_a have been established previously for the appropriate wavelengths. Atmospheric correction of the visible observations is obtained by multiplying the ratio $[\rho_{\text{path}} / \rho_r]$ by ρ_r , leading to ρ_{path} , and, by difference with ρ_v , to the marine reflectance. The aerosol optical thickness is obtained at each wavelength as the weighted average of the two values corresponding to the two bracketing aerosol models, again using the mixing ratio obtained from the NIR bands. The accuracy of such a multiple scattering algorithm has been shown to be of about ± 0.002 in reflectance in the blue.

The algorithm implementation for MERIS is described in Antoine and Morel (1999). The most recent implementation into the ESA operational processing environment uses aerosol models from Shettle and Fenn (1979), with a similar set to the 12 aerosol models used in the SeaWiFS processing. The Rayleigh and aerosol LUTs are generated from a successive order of scattering radiative transfer code including polarization.

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This “Case 1 water algorithm” (assumption of no marine signal in the NIR) is applied over all water types, after application of a specific procedure (Moore, et al., 1999), which removes the marine signal, if any, in the NIR bands.

A specific feature has been implemented in addition to the reference scheme for non-absorbing aerosols, which allows absorbing aerosols to be detected using the observations at 510 nm and assumptions about the marine signal at this wavelength (Nobileau and Antoine, 2005). When the presence of blue-absorbing aerosols is revealed by this test, atmospheric correction is based on the dust models proposed by (Moulin, et al., 2001a). This unique feature of the MERIS atmospheric corrections is not further discussed here (but see Antoine and Nobileau, 2006; Nobileau and Antoine, 2005).

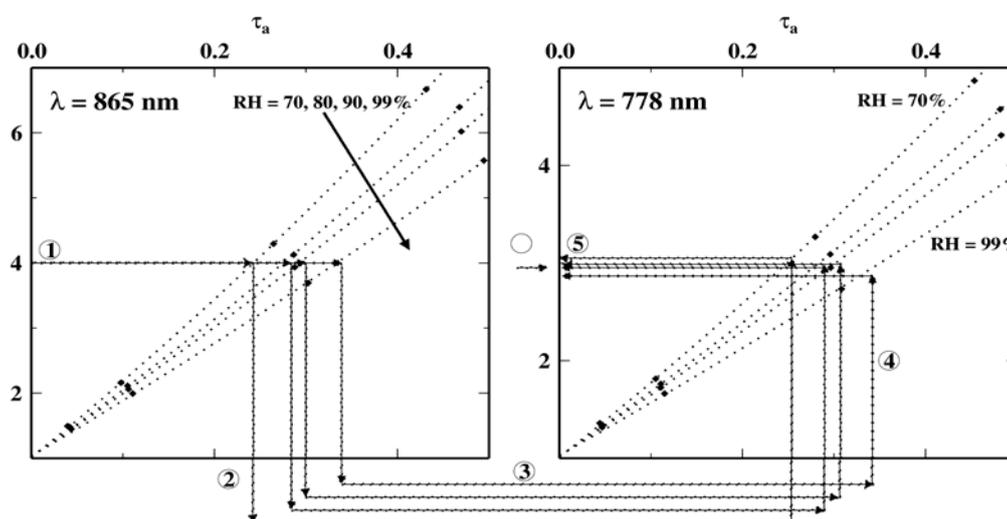


Figure 2: Variation of the path reflectance at 865 and 778 nm as a function of τ_a , and expressed as the ratio $[\rho_{\text{path}} / \rho_{\text{T}}]$, when $\theta_s = 40^\circ$, $\theta_v = 30^\circ$, and $\Delta\phi = \pi/2$. The maritime aerosol model is used, for 4 values of the relative humidity, as indicated. Arrows symbolise a possible way for identifying a couple of aerosol models enclosing the actual aerosol. The circled numbers identify the successive steps of this scheme (see text).

3.2 Evolution: Possible improvements brought by new (published) advances in atmospheric correction of ocean colour observations

3.2.1 Aerosol models

Since the beginning of the present era of ocean colour satellite observations, i.e., since the launch of the NASA SeaWiFS, atmospheric correction of the TOA total signal is based on the use of aerosol models. The rationale for this can be found, for

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instance, in Gordon (1997): briefly speaking, the formerly used, CZCS-type, single scattering algorithms were shown to be unable to meet the accuracy requirements, and consequently some new techniques were needed to account for the multiple scattering effects. Using aerosol models was introduced to better account for multiple scattering effects (e.g., Wang and Gordon, 1994; Gordon, 1997), with the basic idea that using these models would end up with a better extrapolation of these effects from the near IR to the visible. The tests performed on simulated data sets demonstrated that it was indeed the case. The uncertainty in this process mostly lies in the extrapolation from the near infrared to the visible (see below).

The most often used models are the one published by Shettle and Fenn (1979) and those proposed by the WCRP (1986) (thereafter noted S&F79 and WCRP86, respectively). In the MERIS processing, the dust models proposed by Moulin et al. (2001) are also used. The S&F79 and WCRP86 models were not designed for the purpose of atmospheric correction of ocean colour observations, so one may wonder whether they are appropriate for this task.

The question actually recurrently surfaces in the ocean colour community. The sole alternative set of models proposed up to now, however, is the one by Santer et al. (Zagolsky et al., 2007) (at least to our knowledge). This set of models has been built from AERONET (Holben et al., 1998) observations at coastal sites (by definition, there is no open ocean sites in AERONET), so its applicability to the open ocean is not demonstrated.

Therefore, in terms of available sets of aerosol models, the situation is not fundamentally different from what it was 10-15 years ago.

Is this really a key issue, however? Indeed, it must be kept in mind that the role of atmospheric correction is to correct the TOA total signal to get the marine signal with the best possible accuracy, and not to provide aerosol properties (which is the role of aerosol remote sensing, and is using quite different approaches). This sentence reads like a truism, but it must be reminded, however, because the confusion is often made (see, e.g., Yan et al., 2002; and answer by Gordon, 2003), whereas the requirements for the two approaches are different. The quality of aerosol retrievals (in particular the optical thickness) in the near IR will depend on the correctness of the aerosol scattering phase function (in particular in the backward direction), whereas the quality of the extrapolation of multiple scattering effects from the near IR to the visible is more tied to the correctness of the spectral behaviour of the attenuation coefficient.

Another underlying idea was that the processing of ocean colour observations, and in particular the atmospheric correction, cannot be conceived without the final adjustment introduced by the vicarious calibration process. The rationale here is to consider that unavoidable radiometric calibration errors at the sensor level (even the best calibrated instrument cannot do better than ~2% absolute), when coupled to the uncertainty of the atmospheric correction process, cannot alone provide the marine signal with the required accuracy. Therefore, the vicarious calibration process is

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introduced to absorb the remaining uncertainties. Obviously, this cannot work if these “remaining uncertainties” are too large. The corollary is that everything must be done to decrease these remaining uncertainties before applying the vicarious calibration, particularly by improving the atmospheric correction schemes.

Recommendation: use aerosol models from Shettle and Fenn (1979), WCRP (1986), and Moulin et al. (2001), similarly to what was done for MERIS (but see below). Evaluate the IOPA models and the new NASA-OBPG aerosol database (B. Franz, personal communication) in parallel.

3.2.2 Extrapolating the atmospheric path reflectance from the near IR to the visible

One of the main assumptions allowing the present-day algorithms to be applied is that the mixing ratio that can be determined from the near IR bands remains unchanged in the visible so that the extrapolation from the former to the latter is feasible. This “mixing ratio” (X ; see above) is obtained by bracketing the observed aerosol signal in the near IR by choosing the aerosol models that will reproduce the more appropriately the observed aerosol signals (using precomputed LUTs).

The assumption that X is spectrally invariant actually does not hold, and this is likely one important source of uncertainty in the nearIR-to-visible extrapolation process.

It is, therefore, likely that the central issue for atmospheric corrections of ocean colour observations is in the improvement of the extrapolation procedure.

To our knowledge, however, nothing appeared in the literature as to how we can improve this. New aerosol models have been proposed, but this is not really relevant to the extrapolation problem (see above).

Two complementary ways to possibly reduce the extrapolation error would be (1) to increase the number of aerosol models (better “discretisation”), and (2) to use of third band to check the extrapolation based on the two longest near IR bands.

Recommendation for point (1) above: Assuming that the extrapolation from the near infrared to the visible still relies on the assumption of a non-spectral mixing ratio, reducing the uncertainty in this extrapolation would be feasible by having (1) aerosol models distributed evenly in terms of their optical properties (i.e., the spectral extinction coefficients) rather than in terms of relative humidity, and (2) a denser population of aerosol models (within reasonable limits in terms of the size of the LUTs). The implementation of the algorithm for MERIS includes 12 models (using the maritime, coastal and rural types, each for RH=50, 70, 90 and 99%), similarly to what was done in the SeaWiFS algorithms. Doubling this number would double the computation time to generate the LUTs, which is likely acceptable, and could decrease significantly the final error on the marine signal.

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Recommendation on point (2) above: use a band from 744 to 757 nm (~15 nm width) to check the atmospheric correction performed from the bands at 865 and 778 nm. Such a check was already examined for MERIS, using the 709 nm band, but this band has revealed difficult to use because of perturbations by water vapour absorption. Such a band was already proposed as a suitable band for atmospheric correction (IOCCG, 1998). A technique remains to be defined to use this third band either as a confirmation of what is proposed from the use of the two farthest near IR bands, or as an additional constraint. In parallel to this, enlarging the 865 nm band to at least 30 nm instead of 20 nm only would improve the SNR in this band and then would improve certainly the starting point of the atmospheric correction. The transmittance is above 0.998 from 855 to 890 (see also IOCCG, 1998).

Note: using a band further in the visible (e.g. 665 nm) is not recommended because the impact of the marine signal there would require an iterative procedure. This is actually what was done for the processing of the CZCS observations (Bricaud et al., 1987; André et al., 1991), which was mandatory in absence of near IR bands.

3.2.3 Diffuse transmittances

The problem.

The computation of the atmospheric diffuse transmittance, t_d , usually does not receive much attention, whereas it is a fundamental step in the atmospheric correction procedure. Indeed, after the atmospheric path reflectance is estimated, the final step is to divide the TOA marine signal by t_d in order to get finally the water-leaving reflectance (i.e., the signal at the sea level). The error on t_d is, therefore, transferred as an equivalent error on ρ_w . It's likely that accurate atmospheric corrections are often wasted by an inaccurate t_d computation.

In principle, what t_d is supposed to account for is (1) the loss of radiance along the pixel-to-sensor path due to aerosol and Rayleigh scattering out of the field of view and due to absorption, if any, and (2) the gain in radiance due to scattering of marine radiance from neighbouring pixels into the field of view. This second contribution makes t_d depending on the radiance distribution exiting the ocean. Because the radiant field emerging from the ocean is anisotropic (e.g., Morel and Gentili, 1991), the diffuse transmittance in principle depends on the full remote-sensing geometry (i.e., θ_s , θ_v and $\Delta\phi$) and on the water IOPs, which modulate the anisotropy for a given geometry. This was for instance expressed clearly in a recent paper by Gordon and Franz (2008): *"The diffuse transmittance t relates the water*

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component of the TOA radiance to that exiting the water in the same direction. As such t is a property of the ocean–atmosphere system and not just the atmosphere.”

What has been done up to now

Up to now, most of the ocean colour data processing systems use the following approximation (Gordon et al., 1983; see list of symbols):

$$t_d(\lambda, \theta) = \exp\left(-\frac{\tau_{oz} + (1 - \omega_a F_a) \tau_a + 0.5 \tau_r}{\mu}\right) \quad (4)$$

This expression actually ignores point (2) above (i.e., contribution of neighbouring pixels) and approximates only the first contribution to t_d , i.e., the loss of radiance along the pixel-to-sensor path due to scattering out of the field of view and possibly due to absorption.

This is illustrated here on Fig. 3, where the relative difference (in %) between t_d obtained from ocean-atmosphere coupled Monte Carlo simulations and t_d from Eq. (4) is plotted in different conditions (wavelength is 443 nm). The simulations are considered here as the truth.

We see relative errors up to 15% for large aerosol optical thickness, and between 2 and 6% for moderate ones. The errors are larger for clear waters. The errors are significantly larger for $\theta_s = 60^\circ$. Therefore, it is clear that Eq. (4) is not adapted to our requirements.

The comparison of adjacent panels, which are for Chl=0.03 or 1 mg m⁻³, shows clearly a difference for the two chlorophyll concentrations. The difference is small (~2%). It is azimuthally-dependent (because of the anisotropy of the radiant field emerging from the ocean). This dependence is weak, however, in the geometry corresponding to OC remote sensing.

The conclusion is that Eq. (4) must be abandoned and, in principle, the Chl concentration should be known to determine t_d .

Recommendation: it is recommended to abandon Eq (4) and to use lookup tables built from coupled ocean-atmosphere radiative transfer computations, using an average chlorophyll concentration to be determined considering (1) the distribution of the chlorophyll concentration in the global ocean and (2) the way t_a changes with Chl. The azimuth dependence can be neglected (which means that outputs of RT computations will be azimuthally averaged). The entries of these LUTs would be: θ_s , θ_v , λ and τ_a . The Rayleigh optical thickness is implicit in λ , the aerosol type can be fixed provided that only non-absorbing maritime-like aerosols are considered, and the atmospheric pressure can be kept constant at 1013.25 hPa.

A correction for aerosol absorption can be simply introduced as a multiplicative term, in case an absorbing aerosol model from the Moulin et al. (2001) data base is selected:

$$\exp\left[-\left(\frac{(1-\omega_a)\tau_a}{\mu}\right)\right] \quad (5)$$

This correction can be significant for a moderate aerosol optical thickness and a single scattering albedo significantly different from 1 (e.g., 0.8). A correction for atmospheric pressure can also be introduced as a multiplicative term:

$$\exp\left[-\left(\frac{(\Delta\tau_r/2)}{\mu}\right)\right] \quad (6)$$

where $\Delta\tau$ is the difference between the actual atmospheric pressure and the standard value of 1013.25 hPa used to generate the LUTs. This correction is of $\sim\pm 0.4\%$ only for $\mu_s=0.7$ and for changes in atmospheric pressure of $\pm 2\%$ (i.e., from 993 to 1013 and to 1033 hPa).

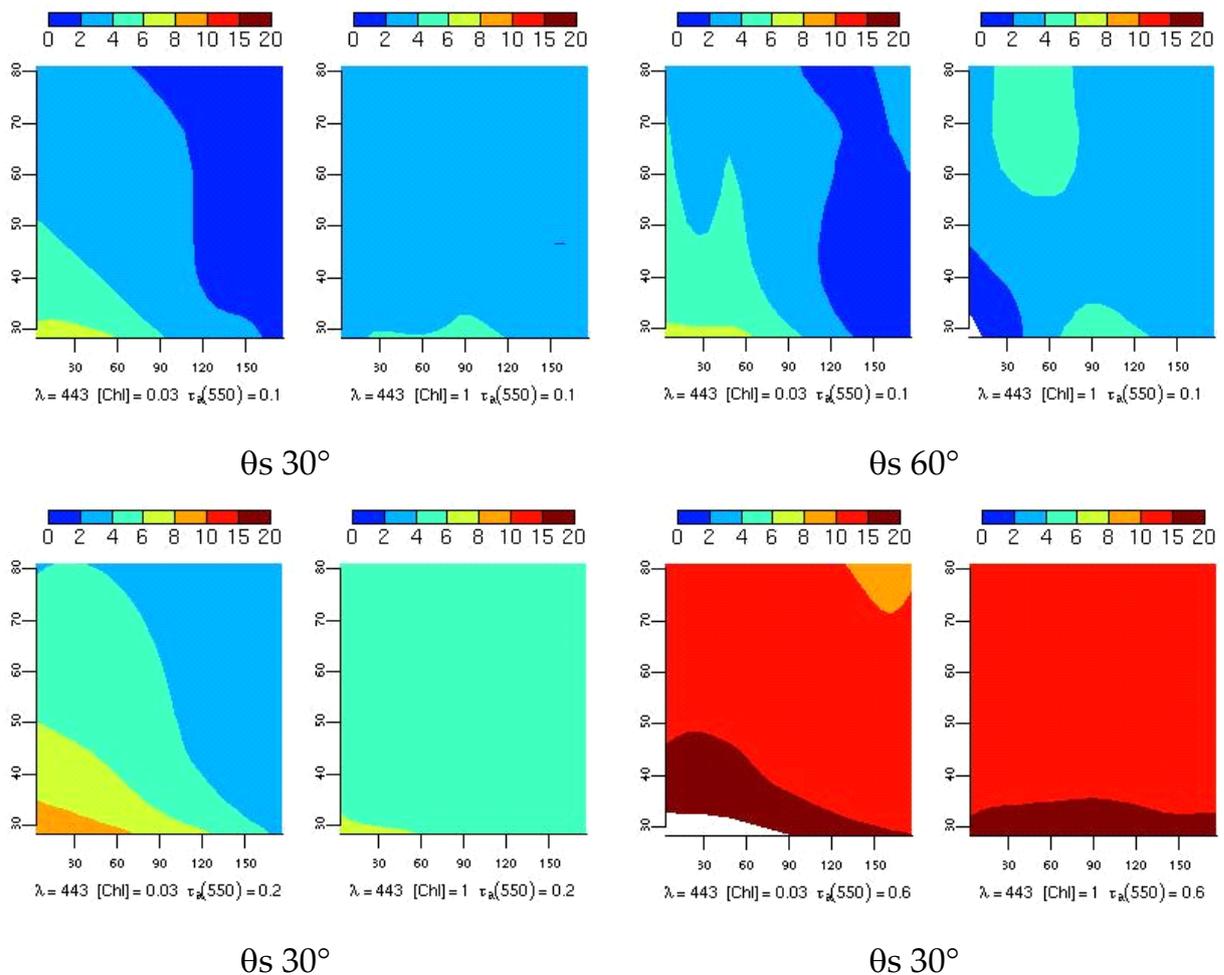


Figure 3: relative difference between t_d obtained from ocean-atmosphere coupled Monte Carlo simulations and t_d from Eq. (4), for the conditions indicated

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3.2.4 Effects of changes in atmospheric pressure

The logic of the algorithm is to rely on lookup tables generated from RT computations where atmospheric pressure is kept constant at the average value of 1013.25 hPa. Therefore, the measured signals have to be corrected for differences in atmospheric pressure before they enter into the processing. The $[\rho_{\text{path}} / \rho_r]$ ratio is sensitive to a change in pressure, so it is not a solution to correct ρ_r from the standard value of pressure to the actual one (which could be done simply by having the atmospheric pressure as an additional entry to the Rayleigh LUT). So, the only solution is to bring ρ_{path} from the actual pressure to the standard one. This was done for MERIS via the following equation:

$$\rho'_p = \bar{\rho}_p \left(1 + \frac{\Delta P}{P} \right) \eta_r \quad (7)$$

where ρ_{path} and ρ'_{path} are the path reflectances for the standard pressure P and the pressure $P' = P (1 + \Delta P/P)$, and $\eta_r = \tau_r / (\tau_a + \tau_r)$. The measured ρ_{path} was corrected to be as close as possible to the value for a standard atmospheric pressure. It was used to form the ratio with ρ_r and to enter into the processing. The converse operation was performed at the end of the procedure, to transform the derived ρ_{path} into its value for the actual atmospheric pressure.

It is here proposed to abandon this equation, and to use a modified version of the Gordon et al. (1988) correction. This correction applies in principle to the multiple-scattering Rayleigh reflectance, and reads:

$$\frac{1 - e^{-\left[\frac{\tau_r}{P} \left(\frac{P}{P_0} \right)^{2.25} \right]}}{1 - e^{-\left[\tau_r \left(\frac{P}{P_0} \right)^{2.25} \right]}} \quad (8)$$

In order to make it suitable for the correction of ρ_{path} , we propose to multiply this correction by η_r in order to account for the fact that the change in pressure is less and less "efficient" in changing ρ_{path} when the aerosol contribution increases.

Note: this modification has to be fully tested before being adopted.

3.2.5 Interpolating within lookup tables

The implementation of the OLCI atmospheric correction will rely on several LUTs. Getting a value from these LUTs is performed by a series of linear interpolations (on geometry in particular), which accuracy has actually never been fully assessed in the MERIS algorithm implementation.

One simple improvement may consist in increasing the discretization of the LUTs, in order to improve the accuracy of these interpolations.

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Another solution is to replace interpolation on some entries by a series expansion, as it is done for instance for the azimuth angle in the SeaWiFS/MODIS processing (Gordon, 1997).

Recommendation: increase LUT discretization as compared to MERIS LUTs. The exact number of entries is TBD. Perform interpolation on cosines of angles instead of angles themselves. This has proven to be efficient for MERIS.

3.2.6 Generation of LUTs (more τ_a 's)

Most of the open ocean is covered by atmospheres with an aerosol optical thickness below 0.1 at 550 nm. Therefore, the discretization of lookup tables should be increased in the range of low optical thickness. The values used for MERIS were 0.03, 0.1, 0.3, 0.5 and 2 at 550 nm. The largest value (2) is unrealistic but is used to better constrain the fit of the $[\rho_{\text{path}} / \rho_r]$ versus τ_a relationship. Recent analysis of the MERIS LUTs (F. Zagolsky, personal comm.) revealed some adverse effects on the 2nd order fit when determining the $[\rho_{\text{path}} / \rho_r]$ versus τ_a relationship. The largest value should be decreased to 0.8.

Recommendation: we propose here to use: 0.01 0.03, 0.05, 0.1, 0.3, 0.5, 0.8 In this case, another quite unrealistic value is introduced (0.01). The rationale here is that ρ_{path} changes more rapidly from its value for a pure Rayleigh atmosphere for initial increases of the aerosol content than it will further change for larger increases of this content (e.g., Antoine and Morel, 1998). Therefore, the relationship has to be better constrained in this domain of low aerosol optical thickness.

3.2.7 Observation geometry: OLCI versus MERIS

The geometry of the Sentinel-3 OLCI observations is different from that of MERIS, in particular with larger view zenith angles. These angles remain, however, within the acceptable limits for radiative transfer calculations to be performed under the “plane-parallel” assumption.

This was shown by Ding and Gordon (1994). The results they presented showed that the effects of the curvature of the Earth on the atmospheric correction of ocean-colour imagery appear to be negligible for solar zenith angles $< 70^\circ$, even for viewing angles as large as 45° .

OLCI goes up to viewing angles of 55° , yet this is lower or comparable to what other sensors such as SeaWiFS and MODIS also reach. There is no need to revise the way atmospheric correction look up tables were generated, as compared to what was done for MERIS. The baseline is still to perform RT computations under the plane parallel assumption.

This is obviously not to say that nothing can be done in this respect. It is not a “sentinel3-specific” issue, however. In particular, the better determination of transmittances is likely to

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already improve the range of angles within which atmospheric correction can be performed with the required accuracy.

3.2.8 Polarization

It is now admitted that polarization should be included into the RT computations onto which the construction of atmospheric correction LUTs are based (for both the Rayleigh and the aerosol signals; see, e.g., Wang, 2006; Meister, 2005; 2006).

Recommendation: the OLCI LUTs should be based on RT computations using a vector RT code. This is actually coherent with what is presently done for MERIS and most of other OC missions.

3.2.9 Including gases within the RT computations

The usual way of dealing with gaseous absorption is to correct the TOA reflectance for the effect of absorption before entering into the atmospheric correction itself, assuming that absorption can be decoupled from scattering effects because it is taking place mostly in the upper atmosphere (this is true for ozone but not necessarily for other gases). Polynomials are often used to perform this correction.

The idea here is that the coupling between absorption and scattering is better accounted for within RT computations (provided that the vertical distribution of scattering and absorption properties is represented correctly) than it will ever be through any empirical or semi-analytical method applied on the TOA signals.

Recommendation: The gaseous absorptions should, therefore, be incorporated into the RT computations needed for the LUTs generation, for average values of each gas concentration (e.g., 350 DU for ozone). Then the correction to be applied when processing the observations only concerns a “residual” change around the average value that was used in the simulation (e.g., +/-50 DU for ozone). The impact of a possible inaccuracy of the correction is, therefore, reduced as compared to the same correction applied to the full range of the parameter in question.

3.2.10 Possible improvements brought by specificities of OLCI: the 400 and 1020 nm bands?

3.2.10.1 The use of the 400-nm band

1 *Improvement of atmospheric correction*

It has been recurrently proposed to use short-wave bands to provide an additional reference point for the atmospheric correction, with the idea that this might be useful to check or constrain the extrapolation of the atmospheric signal from the near IR to the visible.

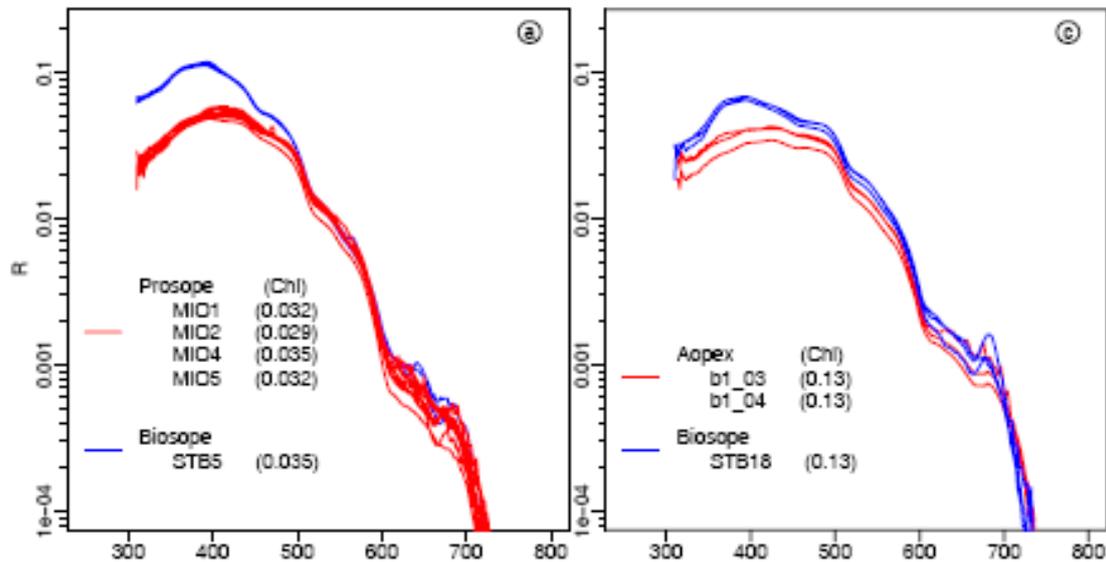


Figure 4: Reproduced from Fig. 6 in Morel et al. (2007). Reflectance spectra for similar [Chl] values as determined in the Pacific Ocean and in the Mediterranean Sea. They illustrate the possible, and very large, range of variation of the ocean reflectance for wavelengths below 440 nm.

In fact, the reflectance of Case 1 waters in the 300-400nm area is (1) large (sometimes this is the spectral range of the maximum in the clearest waters), (2) more varying than it is in the visible for a given chlorophyll concentration (Morel et al., 2007), and (3) so much varying that the option of using this spectral domain for atmospheric correction over Case 1 waters cannot actually be further considered. This is illustrated in Fig. 4.

When Chl reaches high values, the reflectance in the 300-400 nm domain might become smaller and more stable. This could open the way for using this domain as a check point for the atmospheric correction. This possibility remains to be quantified using in situ data from high-Chl Case 1 waters. It is hardly conceivable, however, that this option might be generalized in the frame of an operational processing chain.

2 *Moving the test for clouds from 412 to 400 nm?*

The MERIS algorithms include a test for detection of clouds that is based on a reflectance threshold at 412 nm (Nobileau and Antoine, 2005). This test was actually included for two reasons: (1) the land/ocean/cloud classification scheme was not eliminating enough cloudy pixels from the ocean processing so an additional step of cloud screening was needed, and (2) using the blue allowed thick aerosol plumes to be kept in the processing (they are not bright in the blue) whereas thin clouds are eliminated. Therefore, this was useful to decontaminate images from thin clouds

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(usually a fringe around the brighter clouds eliminated by the tests performed at Level1) and to allow mapping of dust aerosols (another capability of the MERIS algorithms).

There might be some interest to move this test from the 412 nm to 400 nm band in order to improve the discriminative skill of the test. This remains to be tested, however.

3.2.10.2 The use of the 1020 nm band

Using SWIR bands have been proposed recently to overcome the problems of non-zero water-leaving signal in the near IR over turbid Case 2 waters (e.g., Wang and Shi, 2007). This seems a promising solution in this case, although only a moderate improvement is obtained, because of the difficulty of the extrapolation from the SWIR to the visible.

This technique has no real interest over Case 1 waters: there is no gain to expect in the NIR since the marine signal there is 0, and there is some loss to expect because of this additional difficulty in extrapolation.

Therefore, it is not recommended here to use of the 1020 nm band for the OLCI CWAC algorithm.

3.3 Why not proposing some kind of brand new, “advanced”, algorithm?

There are a number of reasons, some of them being very practical and others being more fundamental, for the OLCI CWAC algorithm to be essentially a revision (adaptation) of the MERIS algorithm.

On the practical side, there is the time frame for the development of the OLCI Level-2 algorithms, which is much shorter than was the preparatory phase for MERIS. In the latter case, the “Expert Support Laboratories” worked during ~3-5 years before algorithm prototyping started. It's unrealistic to produce brand news algorithms in the much shorter contractual frame of OLCI algorithms development. The European ocean colour community had no real experience of building a ground segment system at the early times of MERIS; there is now a body of work that can be (and should be) reused.

Another reason is that the instruments are nearly identical.

Still on the practical side, OLCI is not really a “research mission”, and is rather moving toward the operational world. In that sense it has to rely on qualified and proven algorithms, which is the case with the MERIS algorithms.

The continuity requirement is another argument which overlaps practical and more scientific aspects. Building continuous and consistent records over decades (“climate-quality data records”) is indeed more and more put forward as one of the main missions of satellite Earth remote-sensing. Building such consistent records is

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facilitated when the same, or similar, algorithms are used to process the data flow of successive and, hopefully, overlapping missions.

Finally, on a more fundamental aspect, no really brand new, revolutionary, technique has been published in the past 10 or 15 years, at least to my knowledge, which would allow a step change in our ability to perform atmospheric correction of satellite ocean colour observations. Two main paths have been investigated: (1) coupled ocean-atmosphere spectral optimizations or spectral matching algorithms, and (2) neural networks (NN) or similar “statistical” techniques (e.g., genetic algorithms). The motivations for such developments were often in the possibility of performing atmospheric corrections above Case 2 waters or over thick aerosol plumes (such as desert dust aerosols).

The former revealed heavy to implement in operational ground segments, and they rely necessarily on some models, in particular for the ocean, which may not be well adapted in many circumstances.

The latter (NN) necessitates a training, which is inevitably made from simulated data because in situ data bases are still far from being comprehensive enough to be used as training data sets. All these approaches are actually “implementation variants” of the same basic principles, and, therefore, cannot produce significantly different results than more simplistic implementations, such as the ones currently in use for most of the in-orbit ocean colour sensors (e.g. SeaWiFS, MODIS and MERIS).

Simpler techniques, similar to the ones used in the 1980’s and 1990’s for processing of the Coastal Zone Color Scanner (Bricaud and Morel, 1987; André and Morel, 1991), have been recently tidied up and augmented with a capability to process data over sun glint areas with an improved accuracy (“POLYMER” algorithm, Steinmetz, 2006). This type of algorithm relies on a model for the marine reflectance and a simplified description of the atmosphere contribution. The improvement in coverage is dramatic, yet the quality of the marine reflectance has not been to date verified extensively through comparison with in situ data. There is no possibility to derive an aerosol optical thickness from this algorithm. It is therefore mostly designed to provide spatially-extended chlorophyll fields, which is interesting but not meeting the requirements for OLCI, which is supposed to provide a series of advanced ocean colour parameters in addition to the chlorophyll concentration.

In particular, this algorithm and any other algorithm that rely on a model of the marine reflectance (by construction and whatever their degree of sophistication) essentially annihilate the second order variability in the marine reflectance, which is the base of many new and sophisticated algorithms (e.g., PFT derivation from the PHYSAT technique; Alvain et al., 2005). They are definitely adapted to deriving “average-Case1” chlorophyll fields, but not to provide high-quality atmospherically-corrected water-leaving reflectances in all visible bands.

It is on these grounds that updating the MERIS algorithms was considered the best compromise for OLCI.

3.4 Error estimates

3.4.1 Overall uncertainty of the proposed algorithm

An extensive set of sensitivity studies were performed for the MERIS algorithm (see MERIS ATBD 2.7; Antoine and Morel, 2007), which remain valid for the algorithm proposed here.

Since then, a comparison with other atmospheric correction algorithms has been carried out under the auspices of the IOCCG (working group chairman: M. Wang). The report from this exercise will be published in 2010.

It is shown in particular that the MERIS algorithm is performing well and similarly to the SeaWiFS/MODIS, OCTS/GLI and POLDER algorithms. An example of results is presented in Figure 5 below.

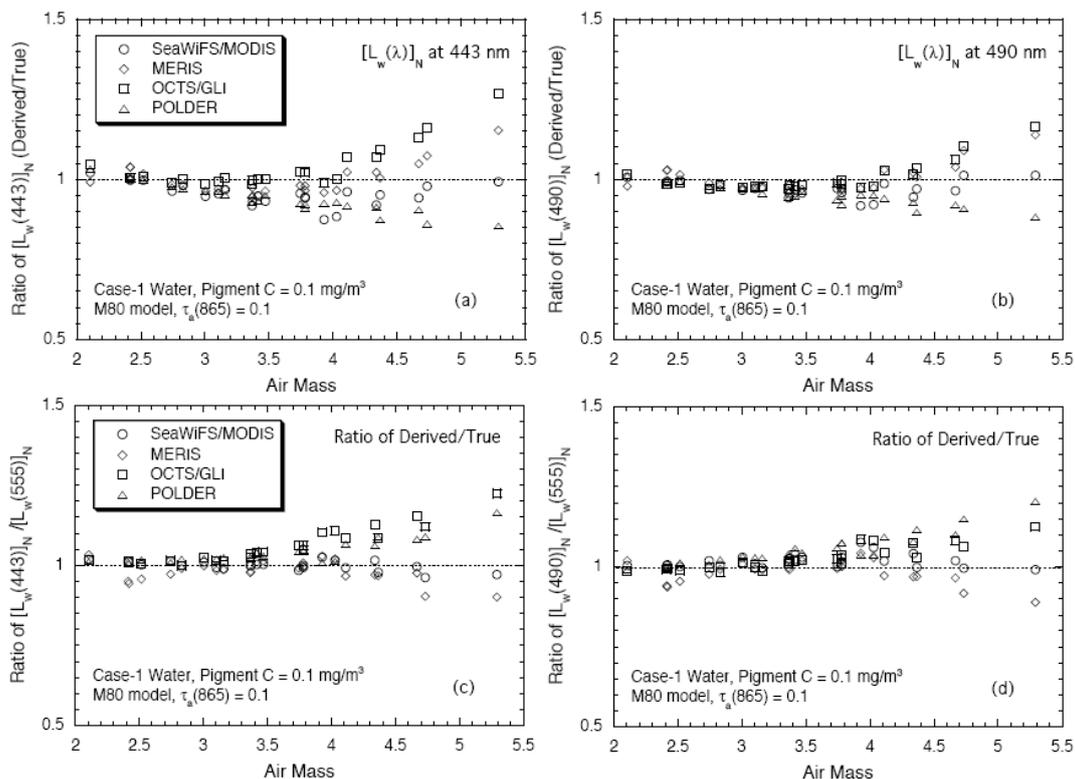


Figure 5: (from Wang et al., under press, IOCCG report #10): Ratio values (derived/true) of various ocean colour parameters as a function of the air mass from atmospheric correction algorithms of SeaWiFS/MODIS, MERIS, OCTS/GLI, and POLDER for (a) $[L_w(\lambda)]_N$ at 443 nm, (b) $[L_w(\lambda)]_N$ at 490 nm, (c) ratio $[L_w(443)]_N/[L_w(555)]_N$, and (d) ratio $[L_w(490)]_N/[L_w(555)]_N$. These results are for the M80 aerosols with aerosol optical thickness of 0.1 at 865 nm and for open ocean (Case-1) water with pigment concentration of 0.1 mg/m³.

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Another way to determine the overall uncertainty of the atmospheric correction algorithm is to analyse matchups with in situ data. This was carried out for MERIS using the BOUSSOLE data set (Antoine et al., 2008). This study demonstrated that the MERIS products were not meeting the requirement accuracy essentially because no vicarious calibration was applied. Such a vicarious calibration will be incorporated in the reprocessing of the entire MERIS archive, due by summer of 2010.

Therefore, the overall uncertainty of the OLCI CWAC algorithm is known from these previous studies.

A significant step forward is however requested by ESA, which is to provide pixel-by-pixel uncertainty estimates.

These estimates have to carry specific information on how the algorithm behaved in the particular geometry and geophysical conditions encountered on every single pixel.

Solutions are proposed in the following section.

3.4.2 Producing Pixel-by-pixel uncertainty estimates

3.4.2.1 *Scaling the TOA radiometric accuracy and uncertainty to the sea level*

The first element of the pixel-by-pixel uncertainty budget originates from the radiometric accuracy (bias in calibration) and uncertainty (instrument noise) at the level of the TOA signal.

In the OLCI configuration, it is planned that pixel-by-pixel radiometric uncertainties will be provided as input to the Level-2 processing chain. These uncertainties are supposed to carry information about radiometric errors at the detector level.

The transfer of these uncertainties on the signal at the sea level depends on the ratio between the marine reflectance and the TOA total reflectance.

This is simply expressed as:

$$\text{Uncertainty at sea level } (0^+) = \text{TOA uncertainty} * [\rho_t/\rho_w] \quad (9)$$

where ρ_t is the TOA total reflectance, and ρ_w the water-leaving reflectance obtained after atmospheric correction.

Note 1: The error on ρ_w originating from atmospheric correction is directly transferred to the $[\rho_t/\rho_w]$ ratio.

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Note 2: Considering that (1) a 1-2% radiometric accuracy is probably the very best that can be achieved for an instrument like OLCI before vicarious calibration and, (2) the $[\rho_t/\rho_w]$ ratio is usually ~ 10 , the uncertainty in the marine reflectance originating from radiometric accuracy at the TOA level would be around 10-20% in average nominal conditions. This is at least twice as large as the required 5% in the blue for clear waters and a clear atmosphere. This is simply an additional illustration that the required accuracy cannot be met without a vicarious calibration. This is also questioning the relevance of simply scaling the radiometric accuracy to the 0^+ level, except if it is accounting for the vicarious calibration.

Note 3: the above note assumes that the radiometric accuracy provided by the L1 processing is of the order of 1-2%. This means a bias of 1-2% (which is precisely supposed to be removed by the vicarious calibration). Another possibility would be to transfer at the 0^+ level the TOA radiometric uncertainties that are quantified by the instrument's noise equivalent radiances ($Ne\Delta L$) or reflectances, again using Eq. 9. This would probably be much more useful and meaningful. Typically this would indicate 1-2% uncertainties at the sea level for marine radiances of about $1.5-2 \text{ mW cm}^{-2} \text{ sr}^{-1} \mu\text{m}^{-1}$ for clear oceanic waters (MERIS $Ne\Delta L$ taken from IOCCG, 1998 are of about $0.0025 \text{ mW cm}^{-2} \text{ sr}^{-1} \mu\text{m}^{-1}$ at 442 nm).

3.4.2.2 Characterizing the atmospheric correction uncertainty

The second element of the pixel-by-pixel uncertainty budget originates from the errors in the atmospheric correction process. These errors are function of the geometry of observation, of the aerosol type and optical thickness (AOT), and of all other parameters entering into this process (uncertainty in wind speed and surface roughness for instance).

When developing and testing the MERIS atmospheric correction algorithm, such uncertainties were quantified in typical situations (see Antoine and Morel, 1997, 1999). This is illustrated below.

The issue here is to generalize such results so that an atmospheric correction uncertainty can be attached to each pixel.

A solution would be to build a lookup table containing coefficients of a relationship between the atmospheric correction error (such as the one shown on Fig. 6) and the AOT, for every geometry and every given aerosol model present in the atmospheric correction lookup tables.

For a given pixel, the coefficients of the relationship between the atmospheric correction uncertainty and the AOT would be interpolated for the geometry of this pixel and between the two aerosol models eventually selected by the atmospheric correction.

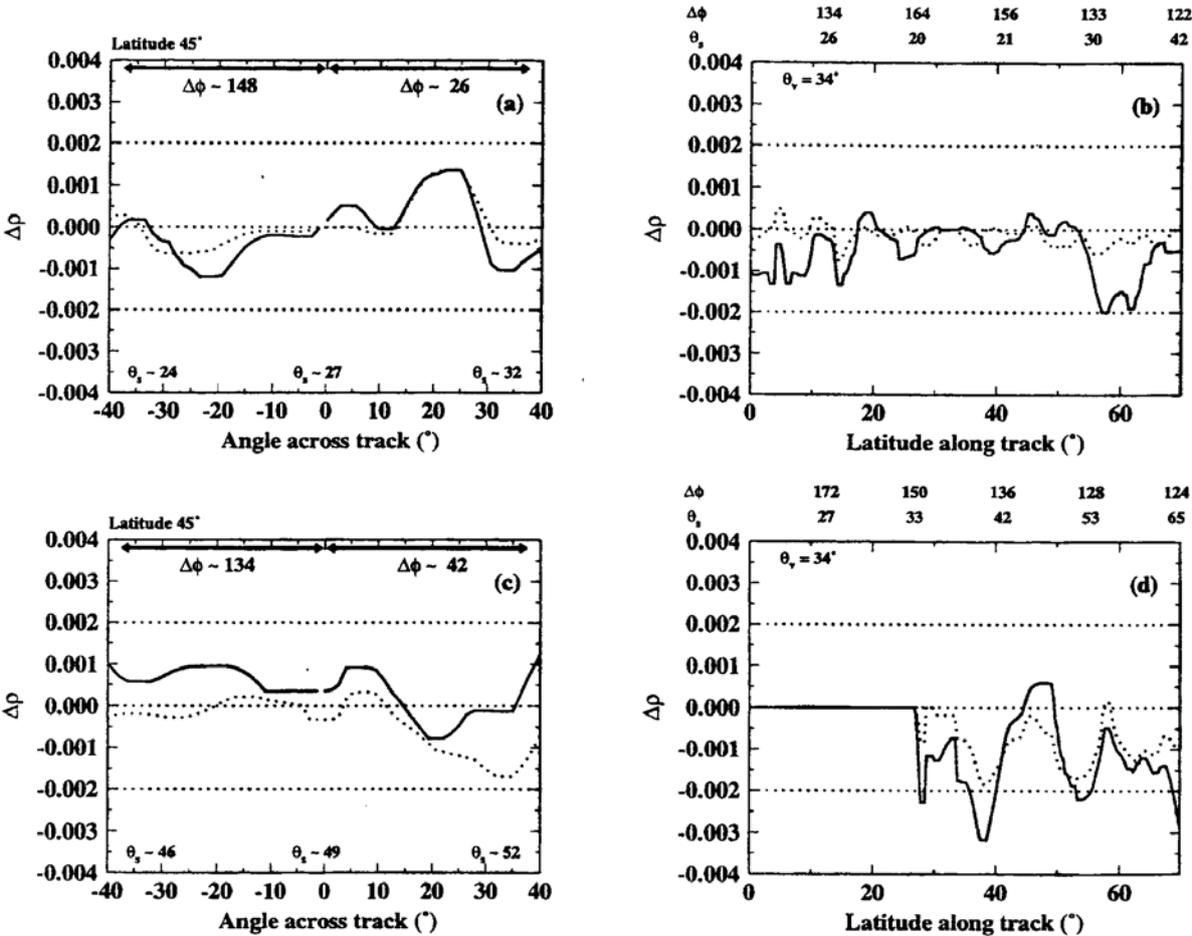


Figure 6. Error in the retrieved marine reflectance at 443 nm (solid curves) and 560 nm (dotted curves) for the maritime aerosol with $RH = 85\%$ and $\tau_a(550) = 0.1$. Panels (a) and (b): summer solstice (21 June). Panels (c) and (d): vernal equinox (21 March). The geometry corresponds either to a MERIS scan at 45° north (panels (a) and (c)), or to a MERIS track from the Equator to the latitude 70° north, the pixel observed being west of the sub-satellite point (panels (b) and (d)). In panel (d), results for latitudes $< 25^\circ$ were discarded as the geometry there corresponded to Sun glint area.

Figure 6. Characterization of the MERIS atmospheric correction errors in typical Earth observation geometries. Figure reproduced from Antoine and Morel 1999

Generating such a table means that atmospheric correction is systematically applied to the simulated data (TOA total reflectances) that were used to generate the atmospheric correction lookup tables, for all aerosol model, geometry and AOT.

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The error at the end of atmospheric correction of these simulated path reflectances should be zero in a perfect correction (because no marine signal was added to the simulation). The error will not, obviously, be null and will conceal errors inherent to the method itself (e.g., interpolation errors, uncertainties due to the assumption of a neutral mixing ratio etc..).

This is, therefore, a characterization of the uncertainty (“noise”) intrinsic to the implementation of the atmospheric correction algorithm.

It cannot quantify the error that would result from selecting an inappropriate aerosol model in a given situation, for instance.

Recommendations:

- Pixel-by-pixel uncertainty estimates should be the quadratic error including the instrument contribution determined from the transfer to the 0+ level of the instrument’s $Ne\Delta\rho$ (Eq. 9), plus the contribution of the atmospheric correction uncertainty determined as described above.
- The average uncertainty of the atmospheric correction should be recurrently assessed through comparison with field data (e.g., Antoine et al., 2008). Such results have to be updated regularly after incorporation of new data. This is a sort of “routine” and mandatory activity that parallels the operational production of the Level-2. Results from these validation activities shall be used to regularly assess whether or not a reprocessing of the OLCI level-2 is requested (i.e., assessing when errors become unacceptable).

3.5 Summary of recommendations

3.5.1 What should be included in the baseline algorithms

- Re-use the core of the MERIS CWAC algorithm (MERIS ATBD 2.7, version 5)
- Re-use the same aerosol models **as presently used for processing of MERIS observations¹**
- Better discretize the aerosol LUTs
- Include polarisation in all radiative transfer computations

¹ The aerosol lookup tables for the present atmospheric correction of MERIS might evolve before the end of the mission (following work and recommendations of the MERIS Data Quality Working Group). Should this happen, the OLCI atmospheric correction lookup tables should be updated accordingly in order to ensure coherency between the two data streams. **This comment actually holds for all implementation variants that will be incorporated in the MERIS processing from now on to the freeze of OLCI algorithms.**

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- Use radiative-transfer-based lookup tables for determining the atmospheric diffuse transmittances
- Modify the correction for changes in atmospheric pressure
- Include the effect of changes in the sun zenith angle on \mathfrak{R}
- Improve LUT interpolation accuracy (interpolate on cosine of angles)
- Include more values of τ_a when generating lookup tables
- Determine pixel-by-pixel uncertainties that account for the instrument noise and the atmospheric correction errors (see section 3.4.2).

3.5.2 What can be examined for further improvements, to be incorporated later on

- Include a band at about 750 nm that would be used as a third band to help the atmospheric correction process.
- Include average gases concentrations within the RT computations and perform only residual corrections in entry of the atmospheric correction process.
- Investigate (specific studies are needed) the use of the 400 nm band instead of the 412 nm band to perform cloud screening (i.e., the Nobileau and Antoine 2005 algorithm).

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4. ASSUMPTIONS AND LIMITATIONS

4.1 Assumptions

- The aerosol models used in the simulations (to generate the lookup tables) are assumed to be good approximations of the actual aerosols over the oceans.
- The “mixing ratio” is wavelength-independent.
- The oceanic diffuse reflectance at 510 nm is varying weakly with the chlorophyll concentration, and it can be derived from a monthly climatology.
- The TOA total reflectance in the visible is represented adequately by summing the atmospheric path reflectance, obtained through radiative transfer simulations over a black, Fresnel-reflecting, ocean, and the product $t_{\rho w}$, calculated independently.
- The plane-parallel atmosphere is a good approximation of the real atmosphere for radiative transfer simulations, at least when the remote sensing configuration is concerned (see Ding and Gordon, 1994).
- It has been assumed here that the algorithm is applied to TOA reflectances that have been previously corrected for gaseous absorption, if any (including water vapour, oxygen and ozone).
- Calibration (Wang Gordon 2002)
- A white caps correction may have been applied to the observations before the entry into the atmospheric correction scheme presented here.
- Pixel-by-pixel error estimates of the TOA radiance (each band) are available.

4.2 Constraints, limitations

- The algorithm is designed to be operated over Case 1 waters. Its application over coastal turbid Case 2 waters (in a situation where pixel classification would have failed) cannot produce correct results.
- Problems could also be encountered over open ocean, when highly reflecting detached coccoliths are present (*e.g.*, Balch et al., 1989, 1991; Gordon and Balch,

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1997), and more generally for any departure of the optical properties from those typical of Case 1 waters (if not identified).

- The algorithm works obviously over cloud free pixels, off cloud shadows, and off the sun glint area.
- The algorithm development relied heavily upon aerosol models, radiative transfer models, as well as bio-optical models. The algorithm reliability is therefore connected to the quality of these models (*cf.* The “Reference model for MERIS level-2 processing”, ESA document PO-TN-MEL-GS-0026, issue 4r1, 13 July 2001).

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