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Algorithm for Atmospheric Correction of Airborne AVIRIS Ocean Images

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ABSTRACT

The algorithm presented in this paper is based on the previously published analytic theory by the author. That theory is based on the scalar radiative transfer approach and it generalizes the satellite Tanre-Deschampes algorithm to the aircraft situation. The restoration of the atmospheric optical parameters is based on a new empirical relationship between the scattering phase function by aerosols and the total aerosol optical thickness in the near infrared.

Examples of processed with the proposed algorithm airborne AVIRIS images of the Gulf of Mexico and North-West Atlantic in different spectral bands are presented.

Keywords: ocean optics, atmospheric optics, atmospheric correction

1. INTRODUCTION

The major portion of the signal recorded by an airborne optical scanner is due to the atmospheric scattering. In order to correctly process optical information collected by aircraft sensors, it is necessary to have fast and reliable atmospheric correction algorithms. In spite of the fact, that there are several well-established algorithms for processing optical data measured from satellite, [1-7] there is a shortage of similar airborne algorithms. The reason for this lies, possibly, in a greater complexity of airborne algorithms. As an input they need not only upward but also downward radiances, and, consequently, they have twice as many input parameters in comparison with the satellite algorithms.

The algorithm presented in this study is based on a radiative transfer model of Ref. [8] and a one-parameter optical model of aerosol atmosphere. [9] In order to restore an aerosol optical thickness $\tau_0 = \tau_A(745)$ the algorithm uses radiance reflection coefficient computed from the two AVIRIS channels (No. 41 and 42). The near infrared optical thickness τ_0 is used further to calculate atmospheric aerosol optical parameters at wavelengths of visible spectrum. The sea albedo A_s is calculated by subtracting molecular and aerosol components from the measured signal. By properly considering optical processes on the ocean-atmosphere interface it is possible to restore a diffuse reflection coefficient of the ocean R. The values R at 440 and 550 *nm* are finally used to estimate the chlorophyll concentration in the upper ocean layer.

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2. MODEL OF ATMOSPHERIC OPTICAL PROPERTIES

To formulate the optical model of marine atmosphere we combine a one-parameter model of aerosol optical atmosphere proposed by Haltrin [9] and a model of Rayleigh molecular atmosphere published by McCartney. [10]

The total atmospheric optical thickness τ^* is given by

$$\tau^*(\lambda) = \tau_R(\lambda) + \tau_A(\lambda), \tag{1}$$

here λ is a wavelength of light, and $\tau_R(\lambda)$ is the total optical thickness of the Rayleigh component of the atmosphere that can be expressed as follows: [10, 11]

$$\tau_R(\lambda) = 0.36\left(\frac{400}{\lambda}\right)^{4.086}, \tag{2}$$

and $\tau_A(\lambda)$ is a total aerosol optical thickness.

According to Haltrin [9] the aerosol optical thickness τ_A may be expressed through the following version of the Ångstrom [12] law:

$$\tau_A(\lambda) = \tau_0\left(\frac{745}{\lambda}\right)^{\frac{0.08\,\varepsilon}{\tau_0}}, \quad \varepsilon = 1 \pm 0.3, \tag{3}$$

here τ_0 is a total optical thickness of aerosol atmosphere at near infrared, $\tau_0 \equiv \tau_A(745)$, and $\varepsilon \approx 1$ is a tweaking parameter. It will be varied to be slightly different from unity only in the cases when the value $\varepsilon = 1$ causes the sea diffuse reflection coefficient derived from remote data to have a negative value.

The atmospheric scattering phase function $p_a(\lambda, \vartheta)$ averaged over the whole atmospheric depth can be represented as follows:

$$p_a(\lambda, \vartheta) = \left[\tau_R(\lambda)\, p_R(\vartheta) + \tau_A(\lambda)\, p_A(\vartheta)\right] / \tau^*(\lambda), \tag{4}$$

here $p_R(\vartheta)$ is a Rayleigh scattering phase function: [10]

$$p_R(\vartheta) = 0.7629 + 0.7113 \cos^2\vartheta, \tag{5}$$

ϑ is a scattering angle, $p_A(\vartheta)$ is an aerosol phase function of scattering that can be expressed through two empirical functions $A(\vartheta)$ and $D(\vartheta)$ and an aerosol optical thickness $\tau_0 \equiv \tau_A(745)$: [9]

$$p_A(\vartheta) = A(\vartheta) + 5\,\tau_0\, D(\vartheta)\,. \tag{6}$$

Numerical values of functions $A(\vartheta)$ and $D(\vartheta)$ are given in Table 1 of Ref. [9] The functions $A(\vartheta)$ and $D(\vartheta)$ also can be expressed in the following analytic form:

$$\left.\begin{aligned} A(\vartheta) &= 1 + 0.645\,P_2(\cos\vartheta) \equiv 0.6775 + 0.9675\cos^2\vartheta, \\ D(\vartheta) &= 0.4\left[p_{HG}(0.8,\cos\vartheta) - p_{HG}(-0.2,\cos\vartheta)\right], \end{aligned}\right\} \qquad (7)$$

here p_{HG} is a Henyey-Greenstein scattering phase function:

$$p_{HG}(\eta,\cos\vartheta) = \frac{1-\eta^2}{\left(1+\eta^2-2\eta\cos\vartheta\right)^{3/2}} \equiv \sum_{n=0}^{\infty}(2n+1)\,\eta^n P_n(\cos\vartheta), \quad |\eta|<1. \qquad (8)$$

The Rayleigh and aerosol phase functions are normalized according to the formula:

$$0.5\int_0^{\pi} p_R(\cos\vartheta)\sin\vartheta\,d\vartheta = 0.5\int_0^{\pi} p_A(\cos\vartheta)\sin\vartheta\,d\vartheta = 1, \qquad (9)$$

The backscattering probability on air molecules B_R and on aerosol particles B_A are calculated according to the equations:

$$B_R = 0.5\int_{\pi/2}^{\pi} p_R(\cos\vartheta)\sin\vartheta\,d\vartheta = 0.5, \quad B_A = 0.5\int_{\pi/2}^{\pi} p_A(\cos\vartheta)\sin\vartheta\,d\vartheta = 0.5-\tau_0. \qquad (10)$$

By integrating atmospheric scattering phase function (4) over backward direction of the scattering angle, we have the following formula for probability of backscattering in atmosphere:

$$B_a(\lambda) = 0.5\int_{\pi/2}^{\pi} p_a(\cos\vartheta)\sin\vartheta\,d\vartheta = 0.5 - \frac{\tau_0\,\tau_A(\lambda)}{\tau_R(\lambda)+\tau_A(\lambda)}. \qquad (11)$$

Equations (1)-(11) express all optical parameters of marine atmosphere through the one single parameter – an aerosol atmospheric optical thickness at near infrared $\tau_0 \equiv \tau_a(745)$, and one additional tweaking parameter ε usable only in special cases of non-physical negative results for retrieved sea reflectance parameters.

The optical model presented here may be used with any one of the radiative transfer models published in Refs. [1-8]

3. RADIATIVE TRANSFER MODEL

In this study we use an atmospheric radiative transfer model by Haltrin [9] proposed in 1996 for the marine environment. By considering multiple scattering effects and by including an option for processing airborne data this model generalizes an algorithm by French authors (Violllier-Tanre-Herman-Deschampes, [1-2]) proposed for satellites. For the same reasons this algorithm is preferable to the approach by Gordon *et. al.* [3-5]. Alternative algorithm proposed by Frazer *et al.* [6] is based on complex numerical computations and it has a shortcoming to be proprietary. Another algorithm by Gao and Goetz [7] was developed for the continental atmosphere and does not include terms with surface albedo.

For simplicity we present here a simplified version of the algorithm (§3 of Ref. [9]) that coincides with the French algorithm [1-2] in the case of satellite measurements. The coded version of algorithm, used to process AVIRIS data, includes the original algorithm presented in §2 of Ref. [9]

According to Ref. [9] the radiance of an ocean-atmosphere system just below the ozone layer is represented as

$$\rho_{oa} = \rho_R + \rho_A + A_S T_S T_V, \tag{12}$$

here

$$\rho_R = \tau_R \frac{\pi_R(\mu,\mu_s)}{4\mu\mu_s}, \tag{13}$$

is a Rayleigh component of radiance reflection coefficient,

$$\pi_R(\mu,\mu_s) = p_R(-\cos\chi_+) + \left[R_F(\mu_s) + R_F(\mu)\right] p_R(\cos\chi_-), \tag{14}$$

and

$$\rho_A = \tau_A \frac{\pi_A(\mu,\mu_s)}{4\mu\mu_s} \tag{15}$$

is an aerosol component of radiance reflection coefficient,

$$\pi_A(\mu,\mu_s) = p_A(-\cos\chi_+) + \left[R_F(\mu_s) + R_F(\mu)\right] p_A(\cos\chi_-), \tag{16}$$

T_S and T_V are, respectively, transmissions from the Sun to the ocean surface and from the ocean surface to a photometer:

$$T_S = \frac{1}{1 + B_a \tau^* / \mu_s}, \quad T_V = \frac{1}{1 + B_a \tau^* / \mu}, \tag{17}$$

here $\mu_s = \cos z_s$, z_s is a solar zenith angle, B_a is given by Eq. (11), $\mu = \cos\theta$ θ is a zenith viewing angle,

$$\cos\chi_\pm = \mu\mu_s \pm \sqrt{\left(1-\mu^2\right)\left(1-\mu_s^2\right)}\cos\varphi, \tag{18}$$

φ is a viewing azimuth angle measured from the solar plane, A_S is the sea albedo.

In the absence of wind A_S is expressed as follows: [13, 14]

$$A_S \cong \frac{\left[1 - R_F(z_s)\right]^2 R}{n_w^2 - \left\{n_w^2 - \left[1 - R_F(z_s)\right]\right\} R} \tag{19}$$

here $n_w \cong 1.34$ is the seawater refraction coefficient, R_F is a Fresnel reflection coefficient of light from the water surface: [15]

$$R_F(\mu) = \frac{1}{2}\left\{\left[\frac{\mu - n_w\eta}{\mu + n_w\eta}\right]^2 + \left[\frac{n_w\mu - \eta}{n_w\mu + \eta}\right]^2\right\}, \quad \eta = \sqrt{1 - \frac{1-\mu^2}{n_w^2}}, \tag{20}$$

and R is a diffuse reflection coefficient of the sea: [16-19]

$$R = \left(\frac{1-\bar{\mu}}{1-\bar{\mu}}\right)^2, \quad \bar{\mu} = \sqrt{\frac{a}{a + 3b_B + \sqrt{b_B(4a + 9b_B)}}}, \quad 0 \le R \le 1, \tag{21}$$

here a is a seawater absorption coefficient, and b_B is a seawater backscattering coefficient.

4. ATMOSPHERIC CORRECTION ALGORITHM

Our atmospheric correction algorithm consists of the following steps: 1) By using Eq. (12) with $A_S = 0$ for each pixel of AVIRIS image above the sea at 745 *nm* we determine the aerosol optical thickness τ_0. 2) By using Eqs. (12)-(21) we calculate diffuse reflectance coefficient R for each maritime pixel in a number of optical channels. 3) In the next step we check for non-physical values of diffuse reflectance (the diffuse reflectance coefficient R should be between zero and one). 4) Next we repeat steps 1-3 for all maritime pixels tweaking the parameter ε until we find a physical solution for R. 5) In the last step we compute a chlorophyll concentration C_c for each maritime pixel by using a regressional relationship that connects C_c with a color index. Below we discuss these steps in more detail.

The atmospheric aerosol optical thickness τ_0 is calculated by the following equations:

$$\tau_0 \equiv \tau_A(745) = \frac{b}{a} - \sqrt{\left(\frac{b}{a}\right)^2 - \frac{c}{a}}, \tag{22}$$

where

$$a = -5\left\{D(-\cos\chi_+) + \left[R_F(\mu_s) + R_F(\mu)\right]D(\cos\chi_-)\right\} \ge 0, \tag{23}$$

$$b = 0.5\left\{A(-\cos\chi_+) + \left[R_F(\mu_s) + R_F(\mu)\right]A(\cos\chi_-)\right\} \ge 0, \tag{24}$$

$$c = 4\mu\mu_s\left[\rho_{oa}(745) - \rho_R(745)\right]. \tag{25}$$

Because the AVIRIS lacks channel corresponding to 745 *nm*, we interpolate $\rho_{oa}(745)$ using channels No. 41 and 42 ($\lambda_{41} = 742.729980\,nm$, $\lambda_{42} = 752.320007\,nm$):

$$\rho_{oa}(745) = \left.\frac{(\lambda_{42} - \lambda)\rho_{41} + (\lambda - \lambda_{41})\rho_{42}}{\lambda_{42} - \lambda_{41}}\right|_{\lambda = 745\,nm} \equiv 0.76329368\,\rho_{41} + 0.23670632\,\rho_{42}. \tag{26}$$

Numerical subscripts in Eq. (21) and below denote the AVIRIS channel numbers.

Let us introduce the value,

$$\rho^{(-)}(\lambda,\mu,\mu_s)=\rho_{oa}(\lambda,\mu,\mu_s)-\rho_R(\lambda,\mu,\mu_s)-\rho_A(\lambda,\mu,\mu_s), \tag{27}$$

which is, according to Eq. (12), proportional to the sea albedo A_s. By using Eq. (19) we can express the sea diffuse reflectance R through the sea albedo A_S:

$$R=\frac{1}{1-\gamma_w}\cdot\frac{A_s}{\varepsilon_w+A_s}, \tag{28}$$

here

$$\gamma_w=\frac{1-R_F(z_s)}{n_w^2},\quad \varepsilon_w=\frac{n_w^2\gamma_w^2}{1-\gamma_w}\equiv\frac{\left[1-R_F(z_s)\right]^2}{n_w^2-1+R_F(z_s)}, \tag{29}$$

According to Refs. [20-21] the subsurface chlorophyll concentration in the ocean can be estimated by the following regression formula:

$$C_c=\left(1.92\ mg\cdot m^{-3}\right)\left[\frac{R(550)}{R(440)}\right]^{1.8}, \tag{30}$$

or, by expressing diffuse reflectances through the sea albedo A_s,

$$C_c=\left(1.92\ mg\cdot m^{-3}\right)\left\{\frac{A_s(550)\left[\varepsilon_w+A_s(440)\right]}{A_s(440)\left[\varepsilon_w+A_s(550)\right]}\right\}^{1.8}, \tag{31}$$

The sea surface albedo at 440 *nm* is computed from the sea surface albedos estimated for the AVIRIS channels 8 ($\lambda_8=438.76001\,nm$) and 9 ($\lambda_9=448.549988\,nm$):

$$A_s(440)=\left.\frac{(\lambda_9-\lambda)A_{s8}+(\lambda-\lambda_8)A_{s9}}{\lambda_9-\lambda_8}\right|_{\lambda=440\,nm}\equiv 0.87334088\,A_{s8}+0.12665912\,A_{s9}. \tag{32}$$

The sea surface albedo at 440 *nm* is computed from the sea surface albedos estimated for the AVIRIS channels 19 ($\lambda_{19}=546.960022\,nm$) and 20 ($\lambda_{20}=556.830017\,nm$):

$$A_s(550)=\left.\frac{(\lambda_{20}-\lambda)A_{s19}+(\lambda-\lambda_{19})A_{s20}}{\lambda_{20}-\lambda_{19}}\right|_{\lambda=550\,nm}\equiv 0.69199802\,A_{s19}+0.30800198\,A_{s20}. \tag{33}$$

The sea albedo at n^{th} AVIRIS channel is expressed through $\rho_n^{(-)}$ and analytically computed values $\tau_R(\lambda_n)$ and $\tau_A(\lambda_n)$ as follows:

$$\left.\begin{aligned} A_{sn} &= \frac{\rho_n^{(-)}}{T_{sn}\,T_{vn}} \equiv \rho^{(-)}(\lambda_n,\mu,\mu_s)\left[1+\frac{\xi}{\mu_s}\right]\left[1+\frac{\xi}{\mu}\right], \\ \xi &= 0.5\,\tau_R(\lambda_n) + \left(0.5-\tau_0\right)\tau_A(\lambda_n) \equiv 0.18\left(\frac{400}{\lambda_n}\right)^{4.086} + \tau_0\left(0.5-\tau_0\right)\left(\frac{745}{\lambda_n}\right)^{\frac{0.08}{\tau_0}}, \end{aligned}\right\} \tag{34}$$

here λ_n is a wavelength of the n^{th} AVIRIS channel.

Equations (22)-(34) with the supporting Eqs. (1)-(11) constitute a complete algorithm for atmospheric corrections and for estimation of a chlorophyll content using the following six AVIRIS channels: 8, 9, 19, 20, 41, and 42.

5. EXAMPLE OF PROCESSED AVIRIS IMAGES

Figure 1 present images of the Gulf of Mexico bay at three AVIRIS channels (No. 41, 8, 19, 42, 9, and 20).

Figure 2 illustrates computed by Eq. (20) aerosol atmospheric optical thickness τ_A at $\lambda = 745$, 440 and 550 nm.

Figure 3 displays chlorophyll content C_c at the same aquatory.

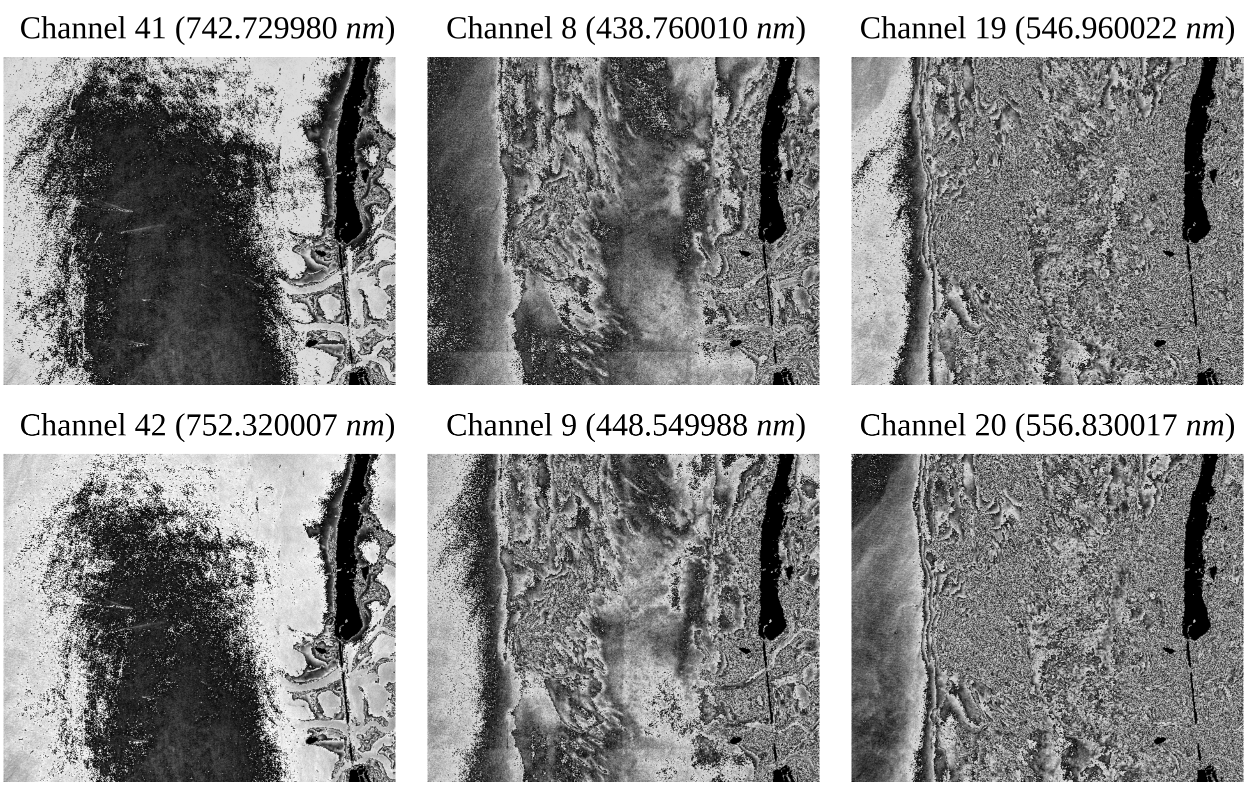

Figure 1. Images from six AVIRIS channels used as input to the described algorithm of atmospheric correction and restoration of subsurface chlorophyll content.

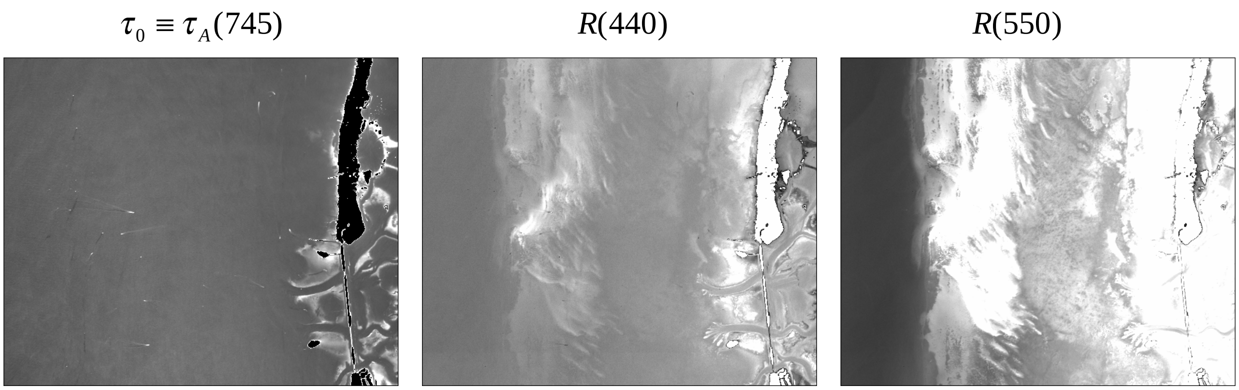

Figure 2. Results of processing AVIRIS images: Atmospheric aerosol optical thickness at 745 *nm*, and diffuse attenuation coefficients R at 440 and 550 *nm*.

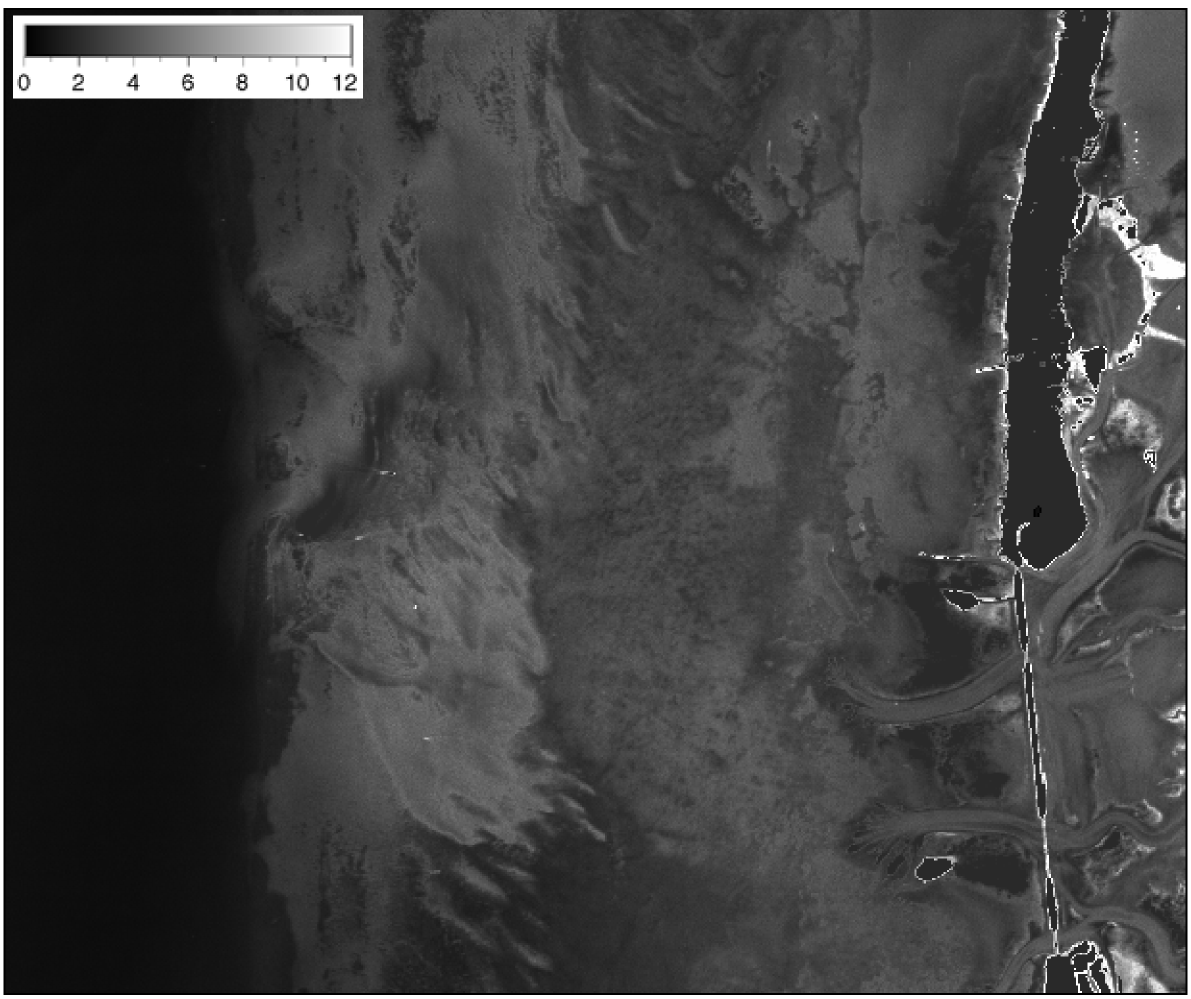

Figure 3. Results of processed AVIRIS images: Subsurface chlorophyll concentration C_c. The density bar in the upper left corner shows concentrations C_c in mg / m^3.

6. CONCLUSION

The algorithm presented in this paper is based on the previously published analytic theory by the author. That theory is based on the scalar radiative transfer approach and it generalizes the satellite Tanre-Deschampes algorithm to the aircraft situation. The restoration of the atmospheric optical parameters is based on a new empirical relationship between the scattering phase function by aerosols and the total aerosol optical thickness in the near infrared.

Examples of airborne AVIRIS images of the Gulf of Mexico and North-West Atlantic in different spectral bands processed with the proposed algorithm are presented. The values of chlorophyll concentrations restored from the AVIRIS images are very close to the values measured *in situ*.

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