METHODS FOR ATMOSPHERIC CORRECTION OF THE DATA OF OPTICAL REMOTE MEASUREMENTS

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Atmospheric correction of the data of optical remote measurements is defined as a procedure of reconstructing the current values of the atmospheric optical parameters with subsequent inversion of the optical transmission operator which converts the brightness coefficient of the underlying surfaces into the brightness of the upwelling radiation. A classification of methods for atmospheric correction of ocean and land images is proposed. This classification distinguishes between methods of successive approximation, spectral, angular, variational, etc. The general characteristic of each class is given. The most developed methods are considered individually. The problem of the parameterization of atmospheric optical properties is considered. General conclusions concerning the principles of construction of algorithms for atmospheric correction are made.

INTRODUCTION

Aerospace video information about the natural resources of the Earth obtained by means of planes, satellites, and maimed space stations contains distortions associated with the errors of orientation, instrumental noise, and various atmospheric phenomena. Atmospheric correction is one of the components of systems for digital ground-based processing of aerospace data and now forms an independent section of remote sensing.

Reviews of methods of atmospheric correction[1–4] have touched upon the problem of taking account of the effect of the atmosphere in imaging the ocean surface and cover a small part of the papers that have been published in this field. The purpose of the present review is a broader account of the methods of accounting for the effect of the atmosphere in the case of remote optical sensing of natural surfaces.

The problem of atmospheric correction consists in the elimination of distortions which are introduced by a scattering and absorbing medium in remote determination of the brightness coefficients of underlying surfaces. This problem is solved by inverting the atmospheric optical transmission operator,[5] which converts the brightness coefficient of the underlying surface into the brightness field of the radiation reflected by the system "underlying surface-atmosphere." Since the spatial-temporal instability of atmospheric aerosol[6] necessitates the reconstruction of its optical parameters instantaneously at the moment of shooting, we shall understand the procedure of reconstructing the current values of the optical parameters to mean the atmospheric correction of the data of remote measurements with subsequent inversion of the atmospheric transmission operator.

MODELS OF RADIATION TRANSFER

A generalized model of solar radiation transfer-through the Earth's atmosphere above dry land Is formed with the help of the boundary problem for the transfer equation[5,7]

$$LI = SI; \quad I\Big|_{\substack{z=0\\ s\in\Omega_+}} = \pi S_\lambda \delta(s - s_0); \; I\Big|_{\substack{z=h\\ s\in\Omega_-}} = RI. \tag{1}$$

Here $L = (s, \nabla) + \alpha(z)$ is the differential operator,

$$S : SI = \frac{\sigma(z)}{4\pi}\int\limits_{\Omega} I(z, r, s')f(\cos(\gamma))ds'$$

and

$$R : RI = \int\limits_{\Omega_+} R_u(s, s')I(h, r, s')\mu' ds'$$

are the integral operators of scattering and reflection, $I = I^\lambda = I(z, r, s)$ is the spectral brightness, λ is the wavelength in μm, z is the vertical coordinate, $r = \{x, y\}$ is the vector of horizontal coordinates, $s = \{\mu, s_\perp\}$ is the propagation vector of the radiation, $s_\perp = \sqrt{1-\mu^2}\{\cos\varphi, \sin\varphi\}$, $\mu = \cos\theta$, θ and φ are the zenith and azimuth angles, s_0 is the direction of propagation of the Sun's rays, Ω is the unit sphere, Ω_- and Ω_+ are the upper and lower hemispheres, h is the height of the scattering atmosphere, $z = 0$ and $z = h$ are the heights of the upper atmospheric boundary and the underlying surface, πS_λ is the solar constant, $\alpha(z)$ and $\sigma(z)$ are the attenuation and scattering coefficients, $f \equiv f(\cos(\gamma))$ is the scattering phase function, $\cos\gamma = s \cdot s'$, γ is the scattering angle, and

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$R_u(r, s, s_0)$ is the brightness coefficient of the underlying surface. The optical properties of the surface are characterized by the albedo

$$q \equiv q(r, s_0) = \frac{1}{\pi} \int_{\Omega} R_u(r, s, s_0) \mu ds$$

in the case of isotropic reflection. In this case

$$RI = \frac{q(r, s_0)}{\pi} \int_{\Omega} I(h, r, s') \mu' ds';$$

in addition, the simplifying assumption $q(r, s_0) \equiv q(r)$ is often used.

The solution of the boundary problem (1) In the case of isotropic reflection was investigated in Refs. 5 and 8. In the particular case $q(r) \equiv \bar{q} = \text{const}$ the solution is expressed by the well-known formula[9,10]

$$\bar{T} = D + \bar{q}E\Psi_0(1-\bar{q}c_0)^{-1}, \tag{2}$$

where T is the radiance averaged over the horizontal coordinates, D is the radiance due to the atmospheric haze, E and Ψ_0 are one-dimensional transmission functions, πE is the swan radiance of the lower boundary, ψ_0 is the norm of the optical spatial-temporal characteristic, and c_0 is the spherical albedo of the atmospheric layer for $\bar{q} = 0$.

In Refs. 5 and 11 the albedo $q(r)$ was expressed in terms of the radiance I measured at the upper boundary $z = 0$ in the following way:

$$q(r) = Z(r)\left[E + \int_{-\infty}^{\infty} \bar{O}(r - r')Z(r')dr'\right]^{-1}. \tag{3}$$

In the particular case $q(r) \equiv \bar{q} = \text{const}$, Eq. (3) simplifies to[5,10]

$$\bar{q} = (\bar{I} - D)\left[E\Psi_0 + c_0(\bar{I} - D)\right]^{-1}. \tag{4}$$

It is easy to see that Eqs. (2) and (4) are reciprocal. The following notation was used in Eq. (3):

$$\bar{O}(r) = \frac{1}{\pi} \int_{\Omega_+} O(h, r, s)\mu ds ,$$

$O(z, r, s)$ is the point spread function.

$$Z(r) = \frac{1}{(2\pi)^2} \int_{-\infty}^{\infty} \Psi^{-1}(0, p, s)\hat{U}(p)\exp[-ip(r + \tilde{r})]dp,$$

$$\hat{U}(p) = \int_{-\infty}^{\infty} (I - D) \exp[i(p, r)]dr,$$

and $\Psi(z, p, s)$ is the optical spatial-temporal characteristic of the atmosphere. The radiative characteristics D, E, Ψ_0, c_0, Ψ, and O, which determine the effect of the optical transmission operator, are independent of $q(r)$; therefore, the use of Eqs. (3) and (4) is fundamental for the algorithms of atmospheric correction. The performance of algorithms for reconstructing the value of $q(r)$ deteriorates if these formulas are not used.[12,13] Instead of Eq. (3), the formula of the method of spatial-temporal characteristics[5,8] can be used. Relations analogous to Eqs. (3) and (4) hold true for any height in the atmosphere. A description of methods for the numerical calculation of the functions D, E, Ψ_0, c_0, Ψ, and O can be found in Refs. 8 and 14–16.

Accounting for the anisotropy of reflection from the underlying surface complicates the solution of the inverse problem of reconstructing the brightness coefficient $R_u(r, s, s')$ since the latter is a function of the angular variables.[17] Methods for inversion of the optical transmission operator of the atmosphere are insufficiently developed in this case.[17–19]

The inhomogeneities of the ocean surface albedo are negligible. Equations (2) and (4) apply if the radiation reflected from the ocean surface does not fall within the detector.[20,21] The brightness coefficient of the ocean $r_0 \approx \bar{q}$ is related to the brightness of the radiation emanating from the water thickness I_w: $I_w = r_0E(1 - r_0c_0)^{-1} \approx r_0E$. Proper consideration of the interaction of the radiation with the wavy ocean surface and the water thickness[22,23] assumes more a complicated formulation of the lower boundary condition for Eq. (1), and on the right side of Eq. (2) terms appear which are responsible for the radiation reflected from the surface and refracted by it.

ATMOSPHERIC PARAMETRIC MODELS

Let us now consider the optical atmospheric models used in the correction algorithms. The optical properties of the atmosphere are determined by the composition of the aerosol and its gaseous components[24,25] and are characterized by the macroparameters

$$\tau_0 = \int_0^h \alpha(z)dz, \quad \alpha(z), \quad \sigma(z), \text{ and } f(\cos\gamma),$$

which are input quantities for all radiation transfer models. The mathematical formulation of inverse problems for reconstructing the actual values of the above quantities on the basis of remote measurement data depends on the way the functions $\alpha(z)$, $\sigma(z)$, and $f(\cos\gamma)$ are parametrized. Their domain of definition is the main a priori information regarding the formulation of the inverse problems.

It is convenient to represent the attenuation coefficients in the form of a sum

$$\alpha(z) = \alpha_A(z) + \alpha_M(z) = \alpha_A(z) + \sigma_M(z) + \beta_M(z),$$

whose components correspond to aerosol attenuation and molecular scattering and absorption. In this way we separate out the aerosol attenuation coefficient $\alpha_A(z)$, which is the most difficult to determine. It is the sum of the aerosol scattering and absorption $\alpha_A(z) = \sigma_A(z) + \beta_A(z)$. The optical thickness of the atmosphere is calculated from the formula

$$\tau_0 = \tau_A + \tau_M = \tau_A^s + \tau_M^s + \tau_0^a,$$

where

$$\tau_0^a = \tau_A^a + \tau_M^a, \qquad \tau_A = \int_0^h \alpha_A(z)dz,$$

$$\tau_M = \int_0^h \alpha_M(z)dz, \quad \tau_A^s = \int_0^h \sigma_A(z)dz, \quad \tau_M^s = \int_0^h \sigma_M(z)dz,$$

$$\tau_A^a = \int_0^h \beta_A(z)dz, \qquad \tau_M^a = \int_0^h \beta_M(z)dz.$$

The quantity $\beta_M(z)$ is due primarily to ozone (O_3) absorption, but also to H_2O and CO_2 (Refs. 26 and 27). As a rule, we neglect the absorption due to H_2O and CO_2: $\beta_M(z) \simeq \beta_{O_3}(z)$. We even often neglect the absorption due to O_3 taking $\beta_M(z) = 0$. The mean statistical profiles $\alpha_A(z)$, $\sigma_M(z)$, and $\beta_{O_3}(z)$ are given in Refs. 28–30. The representation of $\alpha_A(z)$ and $\sigma_M(z)$ in terms of exponential $\alpha_A(z) = \alpha_0^A \exp(-z/H_A)$ and $\sigma_M(z) = \sigma_0^M \exp(-z/H_M)$ is typical (Refs. 31–33). The quantities

$$H_A = \frac{1}{\tau_A}\int_0^{\tau_A} Z(\tau'_A)d\tau'_A, \quad H_M = \frac{1}{\tau_M}\int_0^{\tau_M} Z(\tau'_M)d\tau'_M,$$

where τ'_A and τ'_M are the corresponding optical variables, stand for the center .of gravity of the aerosol and molecular layers located above the underlying surface. It was shown in Ref. 34 that the procedure of reconstructing the surface albedo $q(r)$ depends besides τ_A and; τ_M^s on H_A and H_M but depends on $\alpha_A(z)$ and $\sigma_M(z)$. This fact justifies the feasibility of parameterization of the real altitude distributions by the values of H_A and H_M when τ_A and τ_M^s are known. In Ref. 34 the typical values of $H_M = 8$ km and $H_A = 0.8$ km are indicated over convential conditions land masses and in Refs. 33 and 35 $H_A \approx 1$–1.7 km and $H_M \approx 8$–9.2 km are indicated over the ocean. The actual absorption $1 - \omega_0 = \tau_0^a / \tau_0$ is often neglected or assumed to be constant with altitude. Values of $\omega_A = \tau_A^s / \tau_A = \text{const}$ are given in Ref. 36 which are typical for urban ($0.54 \le \omega_A \le 0.64$), suburban ($0.78 \le \omega_A \le 0.87$), and rural ($0.89 \le \omega_A \le 1.0$) aerosols.

For many correction algorithms the spectral behavior of

$$\tau_0^\lambda = \tau_0^{s,\lambda} + \tau_0^{a,\lambda} = \tau_A^{s,\lambda} + \tau_M^{s,\lambda} + \tau_A^{a,\lambda} + \tau_M^{a,\lambda}.$$

has great significance. The value of $\tau_M^{s,\lambda}$ is constant and can be calculated from the formula[1,27] $\tau_M^{s,\lambda} = 0.00879\lambda^{-4.09}$. Values of $\tau_M^{a,\lambda} = \tau_{O_3}^\lambda$ are given in tabular form[1,31] for different values of λ. The order of magnitude of $\tau_A^{a,\lambda}$ is indicated in Ref. 1. In the case of the Junge aerosol particle size distribution $d\bar{n}/d\log\bar{r} \sim \bar{r}^{-\nu}$, where $\bar{n}$ and $\bar{r}$ are the particle density and radius and $\nu > 0$, we have $\tau_A^\lambda = A\lambda^{-B}$, where $A = 0.01$–0.24, $B = 0.8$–1.5. In Ref. 37 an approximation is proposed which sums the spectral behavior of τ_A^λ and τ_M^λ : $\tau_0^\lambda = a + b\lambda^{-1} + c\lambda^{-4}$. For some models τ_A is associated with the meteorological visibility range S_M. The simplest formula for calculating the last value is $S_M = 3.9\,\sigma_o^A$ (Ref. 6).

The altitude-weighted scattering phase function for each λ is written in the form $f = uf_M + (1 - u) f_A$, where $f_M = 0.7629 + 0.7113 \cos^3\gamma$ and f_A are the Rayleigh and aerosol scattering phase functions and $u = \tau_M^{s,\lambda} / (\tau_M^{s,\lambda} + \tau_A^{s,\lambda})$. The function f_A can be approximated by the sum $f_A = 1 + \sum_{i=1}^{N} x_i \cdot P_i(\cos\gamma)$, $1 \le N \le 3$, where $P_i(\cos\gamma)$ is the Legendre polynomials.[9,38] However, a more exact approximation is $f_A^{g_1,g_2} = vf_{g_1} + (1-v)f_{g_2}$, where $f_g = (1 - g^2)/ /(1 + g^2 - 2g\cos\gamma)^{3/2}$ is the Henyey-Greenstein scattering phase function.[39,33] When $v = 0.983$, $g_1 = 0.82$, and $g_2 = -0.55$, $f_a^{g_1,g_2}$ provided a good approximation the scattering phase function for marine aerosol,[26] and with the values of v, g_1, and g_2 changed it approximates the scattering phase function for continental aerosol.[29] In many papers f_A is calculated on the basis of the Mie theory.[34,36] In this case the reconstruction problem becomes one of reconstructing the power ν. If we parameterize the scattering phase function using the quantities u, v, g_1, and g_2 and take the actual absorption $1 - \omega_0^\lambda = \tau_o^{a,\lambda} / \tau_0^\lambda$ into account, it turns out that the quantity u has the following dependence: $u = \tau_M^{s,\lambda} / \omega_0^\lambda\tau_0^\lambda$. The dependence[40] $f_A = C(\gamma)[\tau_A^\lambda]^{K(\gamma)-1}$, where $C(\gamma)$ and $K(\gamma)$ are given in Ref. 41,

$$\bar{\mu}(\gamma)\tau_0^\lambda - f_M\tau_M^{s,\lambda} = C(\gamma)\left[\tau_0^\lambda - \tau_M^{s,\lambda}\right]^{K(\gamma)},$$

and $\bar{\mu}(\gamma)$ is the brightness index ($\tau_A^{a,\lambda} \approx 0$) can be used as *a priori* information.

Thus, the optical atmospheric model is parameterized by the known quantities $\tau_M^{s,\lambda}$ and H_M^λ

and unknown vector of optical parameters Y^{λ}. One of the possible representations of this vector is $Y^{\lambda} = \{\tau_A^{\lambda}, \omega_0^{\lambda}, v, g_1, g_2, H_A^{\lambda}\}$. The procedure of determining $Y^{\lambda} \equiv Y$ based on remote sensing data forms a part of the atmospheric correction algorithm.

CLASSIFICATION OF RECONSTRUCTION METHODS FOR ATMOSPHERIC OPTICAL PARAMETERS

Trial-and-error methods. The simplest trial-and-error method involves the use of standard statistical models for the atmosphere which are typical of the given region.[11,42,43]

An improved trial-and-error method of approximation of the values Y is based on numerical solutions of the direct problem (1) obtained by different numerical methods in transfer theory: the Monte Carlo method,[5,11] the method of successive approximations,[14,16] the method of spherical harmonics,[44] the Green' s function method,[19,27,8] and others. Usually when solving Eq. (1) we assume isotropic reflection or the absence of horizontal inhomogeneities in the surface albedo. Based on the results of numerical calculations of I for different $(\bar{q}, Y)$, a table is constructed, and a comparison is made between the measured and calculated values of I (Refs. 3 and 45). Then those values of $(\bar{q}, Y)$ are determined from the table for which the experimentally recorded and numerically calculated radiance of upwelling radiation agree. For example, in Ref. 46 airborne measurements of I were compared with the solution of the direct problem on the basis of the assumption of Lambertian reflection from the surface. The values of τ_A were selected based on the results of the comparison. The scattering phase function was calculated for the standard particle size distribution with $\omega_A = 0.96$. The measurements were carried out above a dark object (forest, water) in order to decrease the error in assigning $\bar{q}$.

The main disadvantage of the trial-and-error methods for arriving at a value of Y consists in the absence of any mathematical guarantee of uniqueness of the solution. This fact necessitates that we have available *a priori* information about the type of underlying surface and the atmospheric state of the atmosphere.

Variational methods. Let $J_k(q, Y)$ be experimentally recorded functionals of the radiation field which are functions of the state of the system "underlying surface-atmosphere" (q, Y), where $1 \le k \le K$. We understand $J_k(q, Y)$ to stand for radiation fluxes measured experimentally with different angular, spectral, and spatial resolution. The standard (unperturbed) state of the system and its state at the moment of shooting (perturbed) are characterized by the vector-parameters $Z_0 = (q_0, Y)$ and $Z_0' = (q_0', Y')$, respectively. The problem of determining the variations In the target parameters $\delta Z_m = \{\delta Z\}_m$, where $1 \le m \le M$, based on the known variations in the functionals $\delta J_k = J_k(Z_0) - j_k(Z')$ was formulated in Ref. 47. We find the sought-after values of $(q, Y) = (q_0, y_0) + (\delta q, \delta Y)$ by experimentally determining the deviations δJ_k and solving the indicated variational problem for $\delta Z = (\delta q, \delta y)$.

The variational problem was formulated based on the assumption that δZ is small, applying linear perturbation theory and the technique of adjoint functions. The variations δJ_k are related to the variations δJ_m by means of the response functions $W_{k,m}$, which in their turn are expressed in terms of the importance functions of the meteorological information I_k^* (Refs. 47 and 7). The Importance functions I_k^* can be found from the solution of the boundary problems associated with Eq. (1). As was shown in Ref. 7, the equations separate in Y and the albedo variations $\tilde{q}(r)$ in the case of the linear perturbation theory approximation. This fact simplifies the formulation of the problem.

The variational method, which has been successfully used in neutron physics, has not yet found widespread use in the processing of the information obtained by means of satellites. The fact that response functions can be used to select the most informative functionals J_k can be considered as an advantage of this very promising method.

Spectral methods are the most representative.[1,31,20,48,49] The main idea consists in the application of the spectral characteristics of the components of the atmosphere and of the underlying surface. The effect of absorption of solar radiation by the ocean in the red spectral region for $\lambda_0 \ge 0.67$ μm was used in Ref. 20. The oxygen absorption band In the vicinity of $\lambda_0 = 0.76$ μm, which remains stable under different atmospheric conditions, was used In Ref. 49. The use of spectral methods implies the utilization of the spectral behavior of the optical atmospheric parameters, which allows one to relate Y^{λ} with Y^{λ_0}. Spectral dependences make it possible to determine the quantity Y^{λ} for the entire optical spectral range. This is necessary for the determination of the spectral transmission function.[50]

Angular methods. The vector Y is determined from the radiance measurements in different directions. Most suitable for this purpose are scanning measurements, and the use of photographs is not excluded owing to the existence of a difference between the angles of incidence of the light beams at different points of the image. Angular measurements from satellites were first used to determine τ_0 in Ref. 51 together with the use of an approximate solution of the transfer equation. On the basis of the exact solution of Eq. (1) the study of the angular method was continued In Refs. 52 and 53. Analysis Indicates that the most favorable geometry is the observation of the same local region at different angles since in this case the equation for Y is formulated independently of q. Unlike the spectral methods, the calculations are performed independently for any point of the spectral region λ. This fact saves us from the necessity of using approximate spectral relations among the target optical parameters.

Other methods. In accordance with the classification suggested here, a number of methods should be assigned to a mixed type. For example, the method of successive approximations[48] is based on an idea from Ref. 20 which, in its turn, can be combined with an iterative procedure.[1] We often try to avoid the question of determining of Y and rather find the radiative characteristics appearing in the image correcting algorithms directly. For this purpose regression analysis of images obtained in several channels[54] is used, and local test regions which allow us to extend the atmospheric conditions over the entire decoding territory[55] are employed, but the method proposed in Refs. 20, 33, and 56 reduces to the determination of the ratio $\tau_A^{\lambda} / \tau_A^{\lambda_0}$. The problem of reconstructing the scattering law using the known values of I at the lower and upper boundaries of the scattering layer was formulated in Refs. 57 and 58. However, it does not satisfy the requirements of the remote sensing problem in which I is fixed only at the upper boundary. The effect of the atmosphere on the quality of identification of natural objects has been was studied in quite a few papers,[53,43,59,60] in which quantitative estimates of the atmospheric correction for different underlying surfaces were given.

Adaptive correction. Because of the inadequacy of the optical atmospheric models, the exact solution the problem of reconstructing the parameter vector Y does not necessarily lead to the optimum solution of the generalized problem of identification of natural objects. For this reason, feedback should be provided in systems for processing information from satellites to variations of the optical parameters which should introduce additional corrections to the value of Y with the aim of increasing the probability of identification. An atmospheric correction algorithm which depends on the quality of identification of natural objects becomes an adaptive algorithm.[10]

A block diagram for radiative correction of images from satellites which includes feedback and its performance in the open-loop regime are given in Ref. 61. The modulus of the difference between the initial surface albedo q^* and the reconstructed surface albedo q^{δ}: $\delta q = |q^{\delta} - q^*|$ was taken as the criterion of identification. Absolute accuracy of reconstruction is achieved when $\delta q = 0$ if we close these feedback loops with respect to the perturbing influences represented by the components of the vector δY. These feedback loops act by varying δY. In addition, it is necessary to know the albedo of the test region q^*. The iterative algorithm[1,62] in which the number of reconstructed gradations in C_{chl} or the difference between the reconstructed and the measured brightnesses of the radiation emanating upward from the ocean I_W is the qualitative criterion can be considered as an example of adaptive correction. The other criteria are based on the concepts of cluster analysis.

ATMOSPHERIC CORRECTION OF OCEAN IMAGES

The surface of the ocean has its own features. The ocean albedo is small ($\bar{q} \sim 0.03$), and atmospheric haze can produce a signal which is a factor of 5–10 greater than the useful signal.[50,63] From the viewpoint of radiation transfer at the boundary between two media the surface of the ocean is simpler. This fact makes it possible to create an informative model of the brightness coefficient of the ocean r_0, which takes into account the radiation I emanating from the water column, the Fresnel reflection, the effect of foam, and the relation between r_0 and the biological parameters of the ocean water. The relation between r_0 and the chlorophyll concentration C_{chl} make it possible to regard r_0 as an intermediate result in the atmospheric correction of measured values of I on the way to a quantitative determination of the bioproductivity of ocean water. The reflection of the direct solar radiation incident at the ocean surface forms the solar track in the reflected radiance field, which is usually eliminated by means of instrumentation, namely, by selection of the field of view of the instrument and the zenith singles of observation. Aspects of the mathematical formulation of r_0 and its relation with C_{chl} are discussed in Refs. 1, 4, and 63.

Gordon's method. This method was proposed in Ref. 20 and has since gained wide acceptance. Its ideas were further developed in a number of papers, among which Refs. 1, 2–4, 32, 33, 48, 56, 62, and 64–66 should be noted. It was checked out in the course of processing of CZCS data (the coastal zone color scanner placed onboard the satellite NIMBUS-7). The main point and assumptions of this method are the following:

1. The irradiance measured in the nadir direction is represented in the form

$$I^{\lambda} = D_M^{\lambda} + D_A^{\lambda} + T^{\lambda} \cdot I_W^{\lambda}, \tag{5}$$

where D_M^{λ} and D_A^{λ} are the components due to haze caused by scattering by air molecules and aerosol, respectively, and is the transmission function. According to Eq. (2), $T^{\lambda} = \psi_0$ and for the simplest approximation $T^{\lambda} = \exp(-\tau_0 / |\mu|)$. The quantity D_A^{λ} is calculated in the single-scattering approximation.

2. In the smooth open ocean with clear water (C_{chl} < 0.3 mg/l) at a certain wavelength λ_0 (λ_0 = 0.67 or 0.75 μm) the incident radiation is completely absorbed.

3. The scattering phase function f_a is independent of λ.

4. The existence of a proportionality factor between D_A^{λ} and $D_A^{\lambda_0}$ is assumed

$$D_A^{\lambda} = \tilde{\alpha}(\lambda, \lambda_0) D_A^{\lambda_0}. \tag{6}$$

The solar track is neglected in Eq. (5). This is valid for $\theta_0 \geq 30°$ (Ref. 31). The application of the single-scattering approximation to calculate $D_A^{\lambda_0}$ is justified since $\tau_A^{\lambda_0}$ is small (~ 0.1) at this spectral point. Obviously, for $\lambda = \lambda_0$

$$I^{\lambda_0} = D_M^{\lambda_0} + D_A^{\lambda_0}. \tag{7}$$

From Eqs. (5)–(7) it follows that

$$T^{\lambda} \cdot I_w^{\lambda} = I^{\lambda} - D_M^{\lambda} - \tilde{\alpha}(\lambda, \lambda_0) \cdot (I^{\lambda_0} - D_M^{\lambda_0}), \tag{8}$$

from which it follows that the target value I_W^{λ} is expressed in terms of T^{λ}, I^{λ}, I^{λ_0}, D_M^{λ}, $D_M^{\lambda_0}$, and $\tilde{\alpha}(\lambda, \lambda_0)$. The quantities I^{λ} and I^{λ_0} are measured and the quantities D_M^{λ}, $D_M^{\lambda_0}$, and T^{λ} are calculated. On the basis of these assumptions an explicit form of the coefficient $\tilde{\alpha}(\lambda, \lambda_0)$ is obtained[20,2,33]

$$\tilde{\alpha}(\lambda, \lambda_0) = \varepsilon(\lambda, \lambda_0) \cdot {}_{\lambda}/S_{\lambda_0}, \tag{9}$$

where $\varepsilon(\lambda, \lambda_0) = \omega_A^{\lambda} \cdot \tau_A^{\lambda} \cdot f_A^{\lambda} / \omega_A^{\lambda_0} \cdot \tau_A^{\lambda_0} \cdot f_A^{\lambda_0}$. By virtue of assumption (3) and the approximate equality $\omega_A \simeq 1$ we have $\varepsilon(\lambda, \lambda_0) \approx \tau_A^{\lambda} / \tau_A^{\lambda_0}$. Applying the Angström law $\tau_A^{\lambda} \sim \lambda^{-\nu}$, it follows that $\tau_A^{\lambda} = (\lambda / \lambda_0)^{\nu} \cdot \tau_A^{\lambda_0}$ and

$$\varepsilon(\lambda, \lambda_0) = \tau_A^{\lambda}/\tau_A^{\lambda_0} = (\lambda_0/\lambda)^{\nu}, \tag{10}$$

where $0 < \nu < 1$. From Eq. (10) it follows that it is necessary to know the value of ν or the ratio $\tau_A^{\lambda} / \tau_A^{\lambda_0}$ to calculate $\tilde{\alpha}(\lambda, \lambda_0)$. The way to calculate $\tilde{\alpha}(\lambda, \lambda_0)$ from the satellite data was indicated in Ref. 65. For clear water, in the yellow, green, and red channels of the CZCS it is possible to approximately assume

$$I_w^{\lambda} = I_{w,1}^{\lambda} \zeta \cdot \exp\left\{- \frac{1}{\xi} \cdot \left[\tau^{a,\lambda}/2 + \tau_0^{a,\lambda}\right]\right\}, \quad (\tau_0^{a,\lambda} \approx \tau_{0_3}^{a}),$$

where $I_{w,1}^{\lambda} = 0.498$, 0.3, and < 0.015 [mV/cm^{2}sr] for $\lambda_1 = 0.52$, 0.55, and 0.67 μm, respectively. In what follows we determine $\tilde{\alpha}$ (0.52, 0.67), $\tilde{\alpha}$ (0.67, 0.67), and $\tilde{\alpha}$ (0.448, 0.67) from Eq. (8) by means of extrapolation.

For the real situation in turbid water zones where $I_w^{0,67} > 0$ the method was modified as follows:[1,62]

1. The darkest pixel near the center of the scanner is selected which corresponds to the clear water zone, for which we initially set $I_w^{0,67} = 0$ and $\varepsilon(\lambda_1, 0.67) = 0.67/\lambda_1 = 1$. The quantities $I_w^{0.443}$, $I_w^{0.52}$, and $I_w^{0.55}$ are calculated from Eq. (8).

2. An iterative procedure is used. The quantity $I_w^{0.67}$ is refined according to the equation

$$I_w^{0.67} = I_w^{0.443} \cdot 0.0829 \cdot \left[I_w^{0.443}/I_w^{0.55}\right]^{-1.661}.$$

The quantities $\tilde{\alpha}$ (0.52, 0.67), $\tilde{\alpha}$ (0.55, 0.67), and $\tilde{\alpha}$ (0.443, 0.67) are found from Eq. (8) and, in addition, the quantities $I_w^{0.443}$, $I_w^{0.52}$, and $I_w^{0.55}$ are calculated. This procedure is repeated until the quantities $I_w^{0.67}$ and $I_w^{\lambda_1}$ converge to their limiting values.

Taking the combined effect of molecular and aerosol scattering into account results in the need to consider $\varepsilon'(\lambda, \lambda_0)$ instead of $\varepsilon(\lambda_1, \lambda_0)$ for the short wave channel

$$\varepsilon'(\lambda, \lambda_0) \approx \varepsilon(\lambda_1, \lambda_0)\left[1 + C_{M,A}^{\lambda_1}/D_A^{\lambda_1}\right], \tag{11}$$

where $C_{M,A}^{\lambda_1} = I^{\lambda_1} - D_M^{\lambda_1} - D_A^{\lambda_1}$. Thus, this method is reduced the to calculation of the ratio of the aerosol optical thickness $\varepsilon(\lambda, \lambda_0)$ or $\varepsilon'(\lambda, \lambda_0)$. The target quantity I_w^{λ} is determined from Eq. (8). The chlorophyll concentration C_{chl} is determined from the chromaticity index $I_c = \log\left[I_w^{0.443} / I_w^{0.55}\right]$.

A modification of this method was used to process airborne experiments in Ref. 32. A comparison of the reconstructed quantities with the quantities obtained in the subsatellite experiments justifies the efficiency of this method.

The Khalturin method. An advanced modification of the method for determination of the brightness coefficient of the sea r_0 from satellite measurements of I^{λ} in the nadir direction[48] was suggested in Refs. 31 and 67 that makes it possible to calculate τ_A^{λ} at any point of the trajectory. This modification was tested in the course of airborne experiments.[67,68] The main idea of Gordon's method ($I_w^{\lambda} = 0$) is also used.

The brightness coefficient of the system "ocean-atmosphere" measured for the nadir direction can be approximately written as[48,1]

$$\rho = T_0(\rho_M + \rho_A + \rho_0). \tag{12}$$

Here

$$T_0 = T_{0_3}(1) \cdot T_{0_3}(\zeta), \quad T_{0_3}(\zeta) = \exp(-\tau_{0_3}/\zeta),$$

$$\rho_M = \frac{1}{4\zeta}\left[f_M(\cos(\frac{\pi}{2} + \theta_0)) + \tilde{R} \cdot f_M(\cos\theta_0)\right] \cdot \tau_M \tag{13}$$

is the brightness coefficient of the Rayleigh atmosphere including the Rayleigh atmospheric haze and radiation reflected from the surface and scattered by the air molecules,

$$\tilde{R} = T_a(1)R_f(0) + T_a(\zeta)R(\zeta),$$

$$T_a(\zeta) = \exp(-\tau_M + \tau_A)/2),$$

$R_f(\zeta)$ is the Fresnel reflection coefficient, $\zeta = \cos\theta_0$, θ_0 is the zenith angle of the Sun,

$$\rho_A = \frac{1}{4\zeta}\left[f_A(\cos(\frac{\pi}{2} + \theta_0)) + \tilde{R}\cdot f_A(\cos\theta_0)\right]\cdot\tau_A \quad (14)$$

is the brightness coefficient of the aerosol atmosphere, which takes account of the contribution from aerosol haze and radiation reflected from the surface and scattered by aerosol, and $\rho_0 = r_0 \cdot T_a(1)$.

It was determined in Ref. 31 that for the marine aerosol at the point $\lambda = \lambda_0 = 0.745$ μm the relation

$$f_A^{\lambda_0} = A(\gamma) + k_1\cdot\tau_A^{\lambda_0}\cdot D(\gamma) \quad (15)$$

holds, where $A(\gamma)$ and $D(\gamma)$ are known tabulated functions and $k_1 \simeq 5$.

This original relation makes it possible to express $\tau_A^{\lambda_0}$ in terms of the measured and calculated quantities. From Eqs. (12)–(15) neglecting $\rho_0^{\lambda_0}$, we obtain

$$\tau_A^{\lambda_0} = b/2a - \sqrt{(b/2a)^2 + c/a}\,, \quad (16)$$

where

$$a = -\left[D(\frac{\pi}{2} + \theta_0) + \tilde{R}\cdot D(\theta_0)\right],$$

$$b = A\cdot(\frac{\pi}{2} + \theta_0) + \tilde{R}\cdot A(\theta_0),$$

$$c = T_0^{\lambda_0}\cdot\tau_M^{\lambda_0}\left[f_M(\cos(\frac{\pi}{2} + \theta_0)) + f_M(\cos\theta_0)\right] - 4\zeta\rho^{\lambda_0}.$$

In the process of deriving formula (16) errors were made which insignificantly affect the result since the values

$$r_0 = \frac{1}{T_0}\cdot\left[\rho/T_a - \rho_M - \rho_A\right] \quad (17)$$

are in good agreement with the field measurements.[68]

The Badaev-Halkevich method[49,69-72] consists in the determination of the quantities τ_A^λ, $\alpha_A(z)$, f_A^λ, and r_0 with the help of satellite measurements in the oxygen band centered at $\lambda_0 = 0.76$ μm and in the transparency window centered at 0.74 μm. A set of measurements in the bend centered at 0.76 μm is required to reconstruct $\alpha_A(z)$. As was indicated in Ref. 69 the error in the reconstruction of $\alpha_A(z)$ amounts to 25% and increases to 50% for a complicated profile. The typical accuracy is near 30% (Ref. 49). The solution of the transfer equation in the single-scattering approximation was used to formulate the inverse problem. The quantities $\tau_A^{\lambda_0}$ and $f_A^{\lambda_0}$ are optimally extrapolated over the other spectral regions. The extrapolation error for the known average values of $\bar{\tau}_{\lambda_0}$ and $\bar{\tau}_{\lambda_1}$, correlation coefficients $r_{\tau\tau}(\lambda_1, \lambda_0)$ and standard deviations $|(\sigma_{\lambda_1}, \sigma_{\lambda_0})$ is less than 15–20%. Statistical relationships between τ_A^λ and the scattering phase function f_A^λ averaged over the altitudes, or the normalized brightness coefficient $\bar{\mu}^\lambda(\gamma)$ obtained with the help of ground-based measurements are required.

The method was checked by means of a closed scheme for errors in the measurement of I^λ near 1–3% taking account of the errors in the employed information,[48] and was also tested based on data obtained with the help of aircraft carriers,[70] the satellites Interkosmos 20 and 21[71,72] and the orbit station Salyut-7.[73] A comparison of reconstructed values of r_0 and C_{chl} with the shipboard data[71] indicates that the error in the reconstruction of C_{chl} is 15–20%, i.e., this method allows the identification of 5–6 gradations of plankton in the sea water.

The requirement of an *a priori* statistical relationship between τ_A^λ and f_A^λ should be considered as one of the disadvantages of this method. To improve its reliability it is necessary to introduce a control set of spectral intervals to estimate the accuracy of the results and integrate high-accuracy photometric MKS instruments and high-resolution scanners.

ATMOSPHERIC CORRECTION OF DRY-LAND IMAGES

The brightness coefficient of the Earth's surface is very informative. Using its value we can identify the type of underlying surface, its projective plant canopies and a number of other quantities. For example, the relation between $\bar{q}$ and the humus concentration C_h in the soil is given in Ref. 74, which makes it possible to directly relate the direct relation of I and C_h. However, there is no single theory relating the brightness coefficients of the underlying surface and the parameters of natural resources owing to the variety of natural plant canopies; therefore, the purpose of the atmospheric correction remains the calculation of the quantities $\bar{q}$, $q(r)$, and $R_u(r, s, s_0)$.

The Kaufman method. Let ns now examine the method[34] for determination of the quantities τ_0, $\bar{q}$, ω_0, and H_A. We assume that the image includes a boundary between surfaces with essentially different physical properties. The scattering phase function is assumed to be fixed. Let $q^\pm$ be the albedo of the reflecting surface on both sides of the boundary between the two reflective media and $I(-\infty)$, $I(+\infty)$, and $I(\pm 0)$ be the values of I at the points located to the riqht and left side of the boundary and in the vicinity of it. The sizes of the region of mutual influence of the radiative images of natural objects $x_{0.5}$ and its amplitude are given by the equalities

$$x_{0.5} = \frac{I(x_{0.5}) - I(+0)}{I(+\infty) - I(+0)}, \quad (18)$$

$$A_c^{\pm} = \frac{I(\pm\infty) - I(\pm 0)}{I(\pm 0) - \bar{I}}, \quad (19)$$

Based on physical reasons, the expression

$$I(\pm 0) = D + \frac{E}{1 - \bar{q}c_0}\left[\bar{q} \cdot A_0 + q^{\pm}T\right], \quad (20)$$

was obtained in Ref. 34 which can be derived rigorously using a formalism developed in Ref. 75:

$$T = \exp(-\tau_0 / |\mu|), \quad A_0 = \Psi_0 - T_0.$$

The amplitudes of the mutual effect of the right and left sides of the boundary obtained on the basis of Eq. (20) and formula (2), in which it is necessary to replace I and q by $I(\pm\infty)$ and $q^{\pm}$, agree and are equal to

$$A_c^{\pm} = \frac{A_0 \cdot T + q^{\pm}c_0}{1 - q^{\pm}c_0} \quad (21)$$

The values of τ_0 and $q^{\pm}$, are determined by means of an iterative procedure.

1. The quantities f, $\omega_0 = 1$, and $q^{\pm} = I(\pm\infty)/\zeta S_\lambda$, which is the albedo of the system "Earth-atmosphere", are specified.

2. The quantities $I(\pm\infty)$ are calculated from the experimental, data and $A^{\pm}$ — from Eq. (19). τ_0 is then adjusted in such a way that the equality (21) satisfied.

3. The quantities $q^{\pm}$ are determined from Eq. (4) in which I is replaced by $I(\pm\infty)$, etc.

Since ω_0 affects the result of the reconstruction of the quantities $q^{\pm}$, we can calculate ω_0 if one of the quantities $q^{\pm}$ is known. The values of ω_0 and τ_0 are adjusted on the basis of the experimental relation $\omega_0 = (\tau_0 - 0.025)/\tau_0$. Initially we set $\omega_0 = 1$ and calculate τ_0 on the right and left sides of the boundary ($\tau^{\pm}$). We obtain the new value of ω_0 from the above relation in which we make the substitution $\tau_0 = (\tau_0^+ + \tau_0^-)/2$. This procedure is repeated until τ_0 approaches its limit.

Thus, we can calculate $\tau_0^{\pm}$ and $q^{\pm}$ for a fixed value of ω_0 or τ_0 as well as ω_0 and one of the quantities $q^{\pm}$ if we know the other one. The technique was checked using LANDSAT data. Good agreement with the subsatellite data was obtained. The other method of determining ω_0 was suggested in Ref. 36. It was demonstrated that the quantity $\bar{q}_0$, which is determined from the equality $\rho_I - \bar{q}_0 = 0$ ($\rho_I = \bar{I} / \zeta S_\lambda$), depends strongly on ω_0. We can estimate ω_0 using the dependence of $\rho_I - q$ on $\bar{q}$, which is constructed for different values of ω_0.

A method[76] for determining the quantity H_A for $\tau_0 > 0.2$ is based on the fact that $x_{0.5}$ is a linear function of H_A. From Eq. (18) it follows that $I(x_{0.5}) = 0.5 \cdot [\bar{I}(+\infty) - \bar{I}(+0)]$, where $\bar{I}(+\infty)$ and $\bar{I}(+0)$ are one-dimensional quantities independent of H_A. The function appearing on the left side of the equality can be calculated using the technique suggested in Refs. 75 and 8. Adjusting the value of $x_{0.5}$ for different H_A in such a way that the last equality is satisfied, we obtain $x_{0.5}$ as a function of H_A. This dependence is used to obtain H_A. In order to make use of this dependence, it is necessary to find $I(x_{0.5})$ and $x_{0.5}$, from a satellite photograph and then $H_A = \text{function}^{-1}(x_{0.5})$.

The angular method. This method can be applied to any type of underlying surface including the ocean surface. The recent development of the angular method is associated with the utilization of exact concepts of the radiation transfer theory. The scheme of observing of a certain object along different directions was used in Refs. 18, 52, S3, 61, and 77. In Ref. 78 different objects were observed along different directions.[78]

The scheme of the angular method can be described briefly as follows: [52,S3]

1. It is assumed that the average radiance (2) and its variation $\tilde{I} = I - \bar{I}$ can be distinguished.

2. The optical properties of the atmosphere are parametrized: $Y = \{\bar{Y}, H\}$, $\bar{Y} = \{\tau_0, x_1\}$, where $x_1/3$ is the mean cosine of the scattering angle $\sigma(z) \simeq (\sigma_0^A + \sigma_0^M)\exp(-z / H)$. For more complicated versions / is described by a greater number of parameters.

3. The discrepancy functional $\Phi_{\bar{q},1,2,3}[\bar{Y}] = (\bar{q}_1 - \bar{q}_2)^2 + (\bar{q}_1 - \bar{q}_3)^2$, which satisfies the equality $\Phi_{\bar{q},1,2,3}[\bar{Y}] = 0$ for $\bar{Y} = \bar{Y}^*$ and $i = 1, 2, 3$ is minimized in order to determine $\bar{Y}$, where $\bar{q}_1 = (\bar{I}_1 - D) \cdot [E\Psi_{0,1} + c_0(\bar{I}_1^* - D_1)]^{-1}$. Here the true values of the parameters and the quantities which depend on them are denoted by *. It can be easily seen that the surface albedo is eliminated from the equations for $\bar{Y}$. The index i denotes different angular measurements.

4. The value of H is calculated by inverting of the dependence $\zeta(H)$ at the point $H = H^*$, where $\zeta = \zeta(H)$ is determined by an analysis of $\tilde{I}$ taking account of the approximation

$$|\Psi| = [T + A_0(1 + \zeta^2|P|^2)^{-1/2}].$$

The stability of the algorithm with respect to measurement error (less than 3%) independent of the selection of the initial approximation $\bar{Y}^{(0)} = \{\tau_0^{(0)}, x_1^0\}$ in the' domain of variation of the arguments

$$W = \{\tau_0, x_1\}: \tau^{(1)} \le \tau_0 \le \tau^2, \quad x_1^{(1)} \le x_1 \le x_1^{(2)}\}$$

was demonstrated in Ref. 52. In Ref. 53 the vector $Y' = \{\tau_0, x_1, g, d_0, d_1\}$ was reconstructed where the parameters d_0 and d_1 characterize the phase function of reflection from the underlying surface. The method was checked by numerical experiments using a closed scheme.

As calculations[78] have shown the dependence of I on τ_0 is stronger than Its dependence on f; therefore the phase function is assumed to be fixed. This assumption enables us to decrease the number of target parameters. The regularities of radiation transfer which were used in the algorithm are also noted in Refs. 34 and 76. Indeed, it was shown there[34,76] that H_A has practically no effect on the brightness averaged over the horizontal coordinates. On the other hand, is proportional to the size of the region of mutual influence $x_{0.5}$, which is directly related to Ψ.

CONCLUSION

A classification of the methods of atmospheric correct ion is proposed in this review. The most developed methods, the efficiency of which has been checked with the help of numerical or field experiments, were studied separately. A study of the methods of atmospheric correction enables us to formulate general principles for their construction and recommendations on their improvement. Along with the development of instruments, one of the promising trends is the introduction of exact solutions for boundary problems of transfer theory in the algorithms for atmospheric correction. The above theory makes it possible to establish the following general principles for the solution of inverse problems.

1. The quantities q and Y can be reconstructed independently. In the case of the angular method[61] this is achieved by eliminating q when we formulate the equations for Y. In the case of the variational approach, this principle is derived using linear perturbation theory.

2. The problems of reconstructing q and Y are weakly nonlinear. The reason for this is the near-linear dependence of the radiance I on q[55,15,14] and Y.[53]

3. The exact solution of the problem of reconstructing the albedo of an isotropically reflecting surface for known Y has been obtained.[11] The important problem is the parameterization of the optical properties of the atmosphere and underlying surface. This parameterization must provide, first, the required accuracy of the atmospheric correction algorithm. Second, the number of parameters should not be great, in addition, each parameter has to be Independent of the others and important from the viewpoint of model description of the optical properties of the physical substance. The parameterization $Y = \{\tau_A, \omega_0, v, g_1, g_2, H_A)$ is near-optimal.

The feasibility of one or another method is determined by the structure of the remote measurement data and the properties of the underlying surface. The application of alternative methods as well as an increase in the number of measurement channels is intended to improve the reliability of the classification the natural objects. Since different methods were used under different atmospheric conditions and the measurements were performed with different instruments, the results were given in different formats, and at present it is impossible to classify these methods uniquely according to their accuracy. Therefore, the immediate problem now is to numerically simulate the operation of the atmospheric correction algorithms to analyze their accuracy under the same atmospheric conditions.

The exchange of data obtained from satellites which provide remote sensing data on natural resources is of commercial significance abroad.[79] The information about the natural resources of the Earth and the environment is available not only for organizations but also for individuals from different countries of Europe, Asia, and America. This fact significantly explains the international character of the investigations which involve the data obtained with the help of the NIMBUS and LANDSAT satellites and the close relation between the theoretical and applied investigations. In a number of cases atmospheric correction is an element of the data processing system.

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