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REMOTE SENSING OF ATMOSPHERE, HYDROSPHERE, AND UNDERLYING SURFACE

Daytime Sky Radiance as a Source of Information on Surface Albedo in IR Spectral Region. Part I

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Abstract—We suggest a methodical justification of determining the surface albedo in the near-infrared (NIR) region using the observations of spectral atmospheric transparency and daytime clear-sky radiance in solar almucantar. The contribution of the component describing the reflection processes to radiance at different angular distances from the Sun is analyzed. The effect of aerosol absorption on radiance components used in albedo determination is estimated. The solar zenith angle and elongation of aerosol scattering phase function are found to affect the final result of albedo calculation.

Keywords: infrared spectral region, optical depths due to scattering and absorption, sky radiance, asymmetry of aerosol scattering phase function

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Global climate changes pointed out by many researchers recently necessitate carefully studying the variations in parameters of the atmosphere and underlying surface responsible for these changes. The surface albedo q is among most important climate-change parameters [1, 2].

In this paper, we explore how observations of sky radiance in the cloudless atmosphere can be used to determine q for small and moderate aerosol scattering depths τ_a in the near-IR spectral region. This approach to radiance analysis was tested earlier in the solar almucantar for scattering angles $\varphi > 90°$ and for $0.01 \le \tau_a \le 0.05$ in shortwave spectral range. It was shown that, in the blue [3] and, especially, in the ultraviolet [4] spectral regions, the following condition holds: after minor aerosol-introduced additions are eliminated, the sky radiance becomes nearly molecular in character. This made it possible to solve certain very significant atmospheric optics problems associated, e.g., with calibration of experimental data [5]. In the IR spectral range, the situation is as follows. For solar zenith angles not too large, i.e., $Z_0 \le 70°–75°$, not only for small scattering angles, but also for large ones, the field of downward scattered radiation is mainly determined by initial scattering, with further contribution coming from surface-reflected radiation. The spectral albedos of typical natural systems are usually larger in this part of the spectrum than in the visible range, especially under summer conditions. The albedo of green vegetation is characterized by a rapid growth at wavelengths longer than 0.7 μm up to ~1.4 μm, where the albedo of vegetation is close to the albedo of snow and clouds [6]. We can mention, e.g., the q observations performed by personnel of the Pedagogical Institute, Almaty, in the steppe in the southeast of Kazakhstan (Fig. 1), vast territories of which during summer are dominated by partially grassed clay soils. This indicates that properly accurate measurements of sky radiance B (with relative error $\Delta B/B$ no greater than 5%) in combination with observations of coefficients of atmospheric spectral transparency with uncertainty no larger than 1–2%, provided that absorption by gases and aerosol particles is negligible, can serve as a basis for devising an indirect method for determining the terrain albedo in the near-IR region of the spectrum.

The urgency of the problem stems, to a certain extent, from the fact that the global AERONET monitoring observations of sky radiance [7] at a few hundred sites worldwide are not always accompanied by direct measurements of albedo q from aircraft.

Different methods for solving inverse problems are usually used to determine any optical parameter of the atmosphere or underlying surface form sky radiance measurements [8–11]. This paper is planned as a base for further development of a simplified (engineering) method for determining q from sky radiance observations in the IR spectral range. Briefly, its purpose will be to derive a number of approximate formulas relating the optical parameters of the atmosphere and underlying surface. Owing to this approach to q determination, there will be no need to solve the radiative transfer equation for each instance of albedo determination from sky radiance.

A short historical excursion is in order. We note that quantitative information about albedo effects on the characteristics of the field of downward diffuse radiation was first obtained by Chandrasekhar [12], and more detailed information by Coulson, Dave, and Sekera [13]. They analyzed the case of a homogeneous molecular atmosphere, containing no aerosol and absorbing no solar radiation. Experimental data on the contribution of snow cover to sky radiance in the visible spectral range were published by Livshits [14]. The basis of his study was to compare sky radiances in solar almucantar under summer and winter conditions for the same values of coefficients of atmospheric transparency. The calculated radiances for different albedos in the case of elongated scattering phase functions were published by Feigelson and coauthors from the Institute of Atmospheric Physics, Academy of Sciences of the USSR [15]. More recently, different authors, such as Smerkalov [16] and others, repeatedly turned to the problem on the relation between sky radiance and surface reflectance.

Of course, the fraction of surface-reflected light in the observed sky radiance will be more significant for outgoing radiation as compared to transmitted light. However, in this case, when q is determined, for example from a satellite, there are independent problems associated with the need to eliminate the effect of atmosphere-scattered light on radiance [17]. An increase in atmospheric turbidity leads to the growth of sky radiance, similar to the case when albedo increases. At last, the problem of q determination from measurements of intensity of outgoing radiation from aircraft or satellites becomes extremely complex when the underlying surface is very patched such as in tundra or mountains, partially or irregularly covered by snow. Complex use of ground-based and satellite data seems to be the most successful solution of this problem [18].

Thus, the scattered radiation reaching Earth's surface will be analyzed by representing the observed sky radiance in the solar almucantar as a sum of three components

$$\begin{aligned} B_n(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q) &= B_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0) \\ &+ B_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0) + B_q(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q), \end{aligned} \tag{1}$$

where $B_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ is the singly scattered radiance, $B_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ is the component caused by multiple scattering, and $B_q(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q)$ is the component due to reflection from the underlying surface with albedo q. Here, τ_m is the molecular scattering optical depth; and η_a is the single scattering albedo of aerosol particles. Reflection will be assumed to be orthotropic, i.e., the last term in (1) will be assumed to be independent of φ. Choosing the spectral interval where almost no gas absorption bands are present, we arrive at the condition $\eta_m \sim 1$ (η_m is the molecular single scattering albedo). It is just on this principle that the AERONET instrumentation is devised [7].

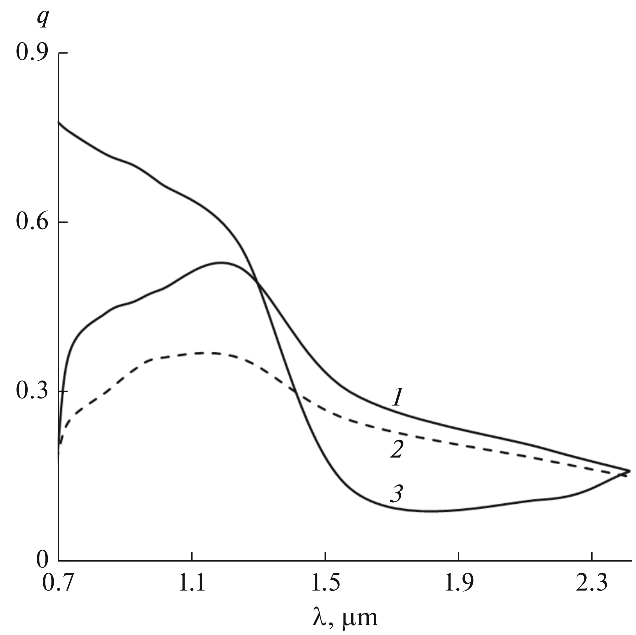

Fig. 1. Average values of spectral surface albedo for Kazakhstan steppe: summer (curve *1*), fall (curve *2*), and winter (curve *3*).

The experimental data will be further processed and analyzed by solving the radiative transfer equation to determine components B_1, B_2, and B_q entering into (1); then, by dividing them into $E_{0,\lambda}e^{-\tau\sec Z_0}\sec Z_0$, where $E_{0,\lambda}$ is the spectral solar constant, we obtain

$$\begin{aligned} f_n(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q) &= f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0) \\ &+ f_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0) + f_q(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q). \end{aligned} \tag{2}$$

The function $f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ is the directional single scattering coefficient, and $f_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ and $f_q(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q)$ are additions to this coefficient due to multiple scattering and reflection of light from the underlying surface. The function f_n is often called the absolute radiance phase function [16]. It should be noted that component $f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ of initial scattering, which is determined from sky radiance, does not depend on absorption depth [14] and has the normalization

$$\tau_1 = 2\pi\int_0^\pi f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)\sin\varphi d\varphi. \tag{3}$$

Here, τ_1 is the (molecular + aerosol) scattering optical depth. In essence, formula (3) quantifies the interrelation between intensities of direct and singly scattered radiation, thus producing many additional conveniences for studying a number of practical tasks and, in

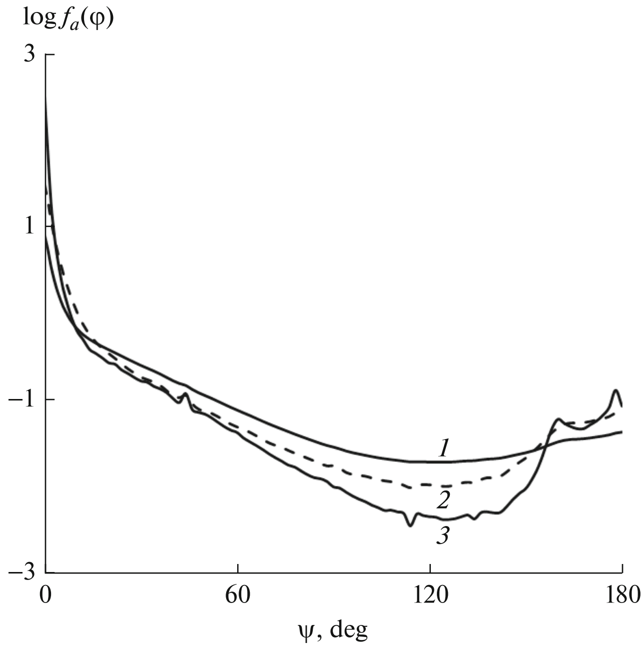

Fig. 2. Initial absolute aerosol scattering phase functions with asymmetry parameters 6 (curve *1*), 9 (curve *2*), and 14 (curve *3*).

particular, those pertaining to the problem of radiative transfer in the atmosphere [3–5, 14, 16, 19].

Integration of formula (2) between the same limits leads to the following equation:

$$\tau_n = \tau_1 + \tau_2 + \tau_q. \qquad (4)$$

The quantity τ_n is often called "the optical depth burdened by multiple scattering and reflection of light from the underlying surface". It should be noted that the use of formula (4) for solving some radiative transfer problems and, in particular, for determining surface albedo is much more efficient than the use of formula (2), and even more efficient than the use of formula (1) [14].

Recent progress in development of numerical methods for solving the radiative transfer equation and, in particular, development of the Monte Carlo method, and the use of contemporary computation means in the radiance calculation make it possible to insure high accuracy in determining components entering into formulas (1) and (2) for any input parameters of the clear-sky atmosphere with minor span time. The process of formation of diffuse light fields can be described in detail if the relative uncertainty of calculating the components entering into (1)–(4) is no worse than 1% in the IR spectral region. This problem is a basis of the current research. The appropriate software for solving the transfer equation by the Monte Carlo method was kindly provided by Zhuravleva [20] and Andreev and Bedareva [21]. Authors of work [21] conventionally assume the solar constant E_0 to be equal to 1. In other words, the calculated data are expressed in units of the solar constant. It is sufficient to multiply the results by the corresponding tabular values of solar constant in order to convert the calculated radiances to standard units of W/(cm^2 nm).

We will consider the behavior of the second and third radiance components for a homogeneous atmosphere. This last simplification is reasonable for points lying in the plane of the solar almucantar [14]. Aerosol is assumed to consist of three fractions: ultramicroscopic, submicron, and coarse fractions [22]. Figure 2 shows three aerosol scattering phase functions for log-normal particle size distribution functions, to be used in subsequent calculations. The refractive index of aerosol material is set to 1.5, and its imaginary part is assumed to be equal to zero. If we characterize the elongation of each of the scattering phase functions by asymmetry factor of scattered radiative fluxes

$$\Gamma_a = \frac{\int_0^{\pi/2} f_a(\varphi)\sin\varphi d\varphi}{\int_{\pi/2}^{\pi} f_a(\varphi)\sin\varphi d\varphi}, \qquad (5)$$

then the Γ_a values are, respectively, 6, 9, and 14.

By adding the directional coefficients of molecular single scattering $f_m(\varphi)$ to absolute aerosol scattering phase functions, and by varying the number of particles (or aerosol scattering optical depth) in the fractions, we can combine summed absolute single scattering phase functions $f_1(\varphi)$ with different asymmetry factors of radiative fluxes Γ_a for a subsequent use in the calculations.

In this paper, the main numerical experiments were performed, unless specifically indicated, with $\tau_m = 0.007$ (wavelength of 1.02 μm), $\tau_a = 0.1$, $\eta_a = 0.95$, and $\Gamma_a = 9$. Figure 3 shows calculations of absolute multiple-scattering functions $f_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ as an example. They were calculated by solving the radiative transfer equation provided that $q = 0$. As shown in [15, 19], multiply scattered light explicitly exhibits diverse angular structures. Moreover, for different scattering phase functions $f_1(\varphi)$, the functions $f_2(\varphi)$ also differ in absolute value.

In this regard, in solving problems on extraction of single scattering phase function $f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ from observed scattering phase function $f_n(\varphi,\tau_m,\tau_a,\eta_a,Z_0,q)$, the sum of components $f_2(\varphi,\tau_m,\tau_a,\eta_a,Z_0) + f_q(\tau_m,\tau_a,\eta_a,Z_0,q)$ should be subtracted for every individual angle φ. If, as in our case, the problem on the use of the radiance scattering phase function $f_1(\varphi,\tau_m,\tau_a,\eta_a,Z_0)$ for determining the surface albedo q is to be solved, the angular structure of both singly and multiply scattered radiation contributions should be taken into consideration in order to extract the term $f_q(\tau_m,\tau_a,\eta_a,Z_0,q)$.

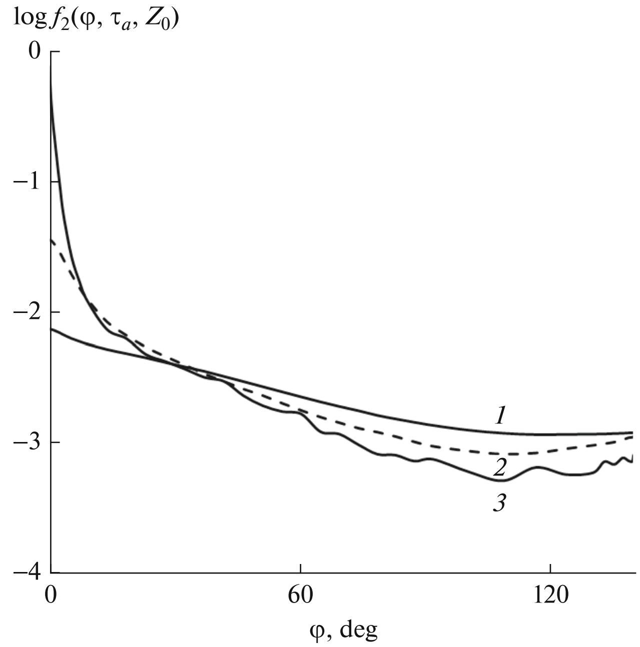

Fig. 3. Angular dependencies of the component of radiance phase function $f_2(\varphi, \tau_m, \tau_a, \eta_a, Z_0)$ of multiply scattered radiation for different asymmetry factors of aerosol scattering phase functions Γ_a: 6 (curve *1*), 9 (curve *2*), and 14 (curve *3*).

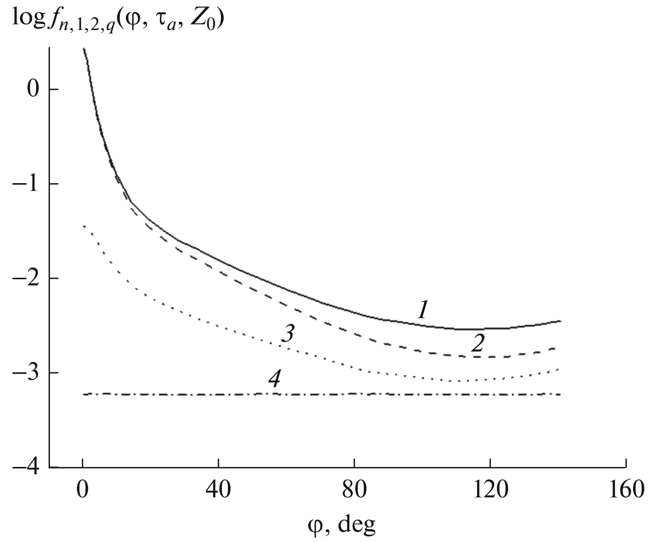

Fig. 4. Angular behavior of components of radiance phase functions $f_{n,\,1,\,2,\,q}$: f_n (curve *1*), f_1 (curve *2*), f_2 (curve *3*), and f_q (curve *4*) for $Z_0 = 70°$ and $q = 0.2$.

We will consider (Table 1) how aerosol absorption influences components entering into formula (4). As was already indicated above, in the process of radiative transfer in a diffuse medium, absorption does not change the absolute values of directional coefficients of initial scattering $f_1(\varphi)$ and, correspondingly, τ_1 values in formula (4). The surface albedo q was set to zero in order to judge absorption of only multiply scattered radiation. All these parameters were kept unchanged in the τ_n calculations, except for η_a, which varied in the range from 1.0 to 0.7.

Table 1 suggests that discrepancies between τ_n, caused by changes in parameter η_a, are very small for fixed solar zenith angles Z_0. Departures from the conservative-scattering ($\eta_a = 1$) case are maximal in essentially those situations when strongly absorbing particles (of the type of urban smog) are present in the atmosphere. These cases are indicated in the last column of Table 1 ($\eta_a = 0.7$) and vary from 0.8 to 3.5%. This suggests that not (2), and not even (1), but, rather, (4) is unconditionally useful for determining albedo q from observed sky radiance in the near-IR spectral region.

We will estimate the contribution of τ_q to τ_n for surface albedo q, varying in the range from 0.1 to 0.7, and for solar zenith angle $Z_0 = 75°$. The τ_q/τ_n values are not large and do not exceed 20% even for high values $q = 0.7$ (Table 2). At the same time, from analysis of angular differences among $f_n(\varphi, \tau_m, \tau_a, \eta_a, Z_0, q)$, $f_1(\varphi, \tau_m, \tau_a, \eta_a, Z_0)$, $f_2(\varphi, \tau_m, \tau_a, \eta_a, Z_0)$, and $f_q(\tau_m, \tau_a, \eta_a, Z_0, q)$ it follows that, if the condition is fulfilled that f_q is independent of φ, as the scattering angle grows, and when $\varphi \geq 60°$, the component f_q approaches f_2 and f_1 (Fig. 4). Thus, we can select such an optimal angular range Δφ, in which surface-reflected light makes a much larger contribution to the downward radiation.

By analogy with formula (4), we introduce the quantity S_n, which differs from τ_n by integration limits:

$$S_n = S_1 + S_2 + S_q, \tag{6}$$

Table 1. The τ_n values for different parameters of the atmosphere

η_a	1.0		0.95		0.8		0.7	
Z_0, deg	60	75	60	75	60	75	60	75
$\tau_a = 0.1$	0.124	0.140	0.124	0.139	0.123	0.13	0.123 (0.8%)	0.136 (2.8%)
$\tau_a = 0.15$	0.193	0.228	0.193	0.227	0.192	0.223	0.191 (1.0%)	0.221 (3.5%)

Table 2. Contribution of τ_q to τ_n (%) for solar zenith angle $Z_0 = 75°$

Γ_a	φ	q			
		0.1	0.3	0.5	0.7
6	$0 \le \varphi \le 180°$	3.0	8.5	13.6	18.3
9	$0 \le \varphi \le 180°$	2.4	7.1	11.3	15.2
14	$0 \le \varphi \le 180°$	2.1	6.1	9.8	13.2

Table 3. Contribution of S_q to S_n (%) for solar zenith angle $Z_0 = 75°$

Γ_a	φ	q			
		0.1	0.3	0.5	0.7
6	$90 \le \varphi \le 150°$	7.6	20.1	29.7	37.3
9	$90 \le \varphi \le 150°$	8.4	21.8	31.9	39.8
14	$90 \le \varphi \le 150°$	10.4	26.1	37.1	45.4

where

$$S_n = 2\pi \int_{\varphi_{\min}}^{\varphi_{\max}} f_n(\varphi)\sin\varphi d\varphi, \quad (7)$$

$$S_1 = 2\pi \int_{\varphi_{\min}}^{\varphi_{\max}} f_1(\varphi)\sin\varphi d\varphi, \quad (8)$$

and

$$S_2 = 2\pi \int_{\varphi_{\min}}^{\varphi_{\max}} f_2(\varphi)\sin\varphi d\varphi. \quad (9)$$

The integration limits can be selected as follows. For the case, analyzed above, the choice of the lower limit $\varphi_{\min}$ is, in essence, already predetermined by all preceding considerations: $\varphi_{\min} = 60°$. The interval $\Delta\varphi$ is set to 60°. If solar zenith angle $Z_0 = 60°$, then $\varphi_{\max} = 2Z_0 = 120°$. This angle corresponds to maximal azimuth $\Psi = 180°$ in the solar almucantar. It is desirable that the upper integration limit $\varphi_{\max}$ in the range of solar zenith angles $60° \le Z_0 \le 75°$ would remain the same in all cases and would be equal to $2Z_0$. Then, the $\varphi_{\min}$ values will vary as $2Z_0 - 60°$.

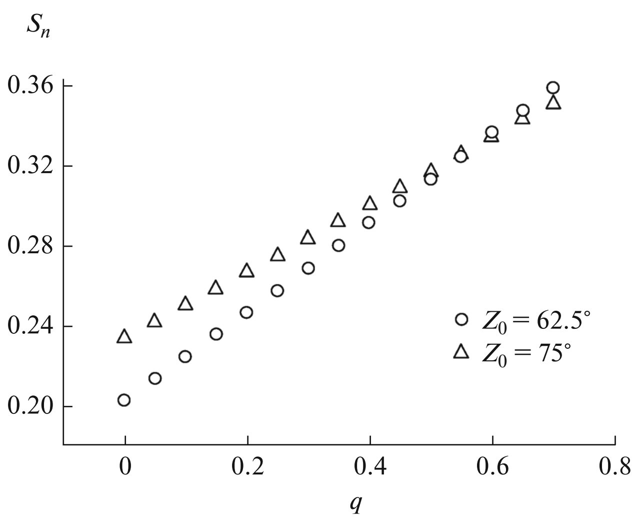

Fig. 5. The function S_n versus surface albedo.

The rationale behind the transition from τ_n to S_n in the q determination can be judged from Table 3, which presents the ratio S_q/S_n for different asymmetry factors of aerosol scattering phase function (in percent). Usually, the contribution of S_q to S_n is markedly larger than the contribution of τ_q to τ_n. When surface albedo is 0.5–0.7, the S_q contribution reaches 30–45%. We note that the values $q \sim 0.5$–0.7 in IR spectral range are quite often encountered in nature [6].

How strongly does the function S_n, to be used subsequently for q determination, depend on surface albedo can be seen in Fig. 5; it shows the S_n calculations for two solar zenith angles 62.5° and 75° when $\Gamma_a = 9$. The calculations cover a wide range of variations in q from 0 to 0.7, including experimental data. The relation between S_n and q is distinctly linear in character, ensuring simplicity of subsequent approximations, to be used in the method devised here.

We will summarize the discussion above. The paper suggests the basics of the method for determining surface albedo from daytime cloud-free sky radiance in the solar almucantar in the IR spectral region. We performed preliminary Monte Carlo calculations in the wavelength region of 1.02 μm and selected appropriate functions to be used subsequently as a basis for albedo determination from measurements. To derive the engineering formulas and grids of radiation parameters, to be used in the method, it is necessary to perform the following complementary studies:

1) to carry out the missing radiance calculations and to analyze them for other wavelengths (such as those from MODIS channels: 0.858, 1.24, 1.64, and 2.13 μm) in the IR range;

2) to explore in detail how the quantities used in the method depend on scattering optical depths;

3) to do the same for a more detailed series of solar zenith angles;

4) after that, to formulate clearly the technology of albedo determination in the IR spectral region. To present conclusion about the accuracy of the method being developed.

These questions will be the subject of research in the near future.

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