The effect of infrared radiation reflected from the Earth's surface observed in the atmospheric emission spectra in remote sensing of the Earth from space

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The paper analyzes the contribution of the infrared radiation of the Earth's atmosphere, emitted downward and reflected from the Earth's surface, to the observed spectrum of upward thermal radiation directed to space within the 600–2500 cm^{-1} range. Surface emissivity for various ecological systems is taken from HITRAN–96 database. As a result of modeling the spectra of the outgoing thermal radiation with the resolution about 0.1 cm^{-1} with and without the account of the reflected radiation flux we have arrived at a conclusion that the contribution from surface reflectance cannot be neglected if the surface emissivity is below 0.9. In the spectra, the distinctions manifest themselves both quantitatively in wings of relatively strong lines and qualitatively (inversion) for weak lines. Maximum effect is observed at high spectral resolution and if the surface temperature is close to or lower than temperature of the near-ground air. The modeled effects were identified in spectra observed with an IMG sensor installed on ADEOS satellite over the Sahara desert. The necessity is discussed of considering the above effect in the retrieval of atmospheric component profiles from the high-resolution infrared radiation spectra observed from space.

Introduction

In the early papers on thermal sensing of the Earth's atmosphere from space (for example, Ref. 1) an indication was given of the presence of the radiation component due to reflection from the surface flux in the outgoing thermal radiation spectrum. However, in recent papers[2–5] the contribution of the infrared radiation of the atmosphere emitted downward to the surface and reflected upward is commonly neglected. This approximation describes adequately the situation when the surface emittance is close to unity, i.e., for such surfaces as ocean, snow, taiga, and so on, or the low spectral resolution exists.

However, according to literature data on the emittance of different natural systems in the infrared range (see data files in the HITRAN–96 base[6]), the emittance of deserts, forests, agricultural areas in some spectral ranges of the IMG sensor[4] sensitivity, for example, Band 3 (600–2000 cm^{-1}) and Band 2 (2000–2500 cm^{-1}) can be lower than 0.8. By this we mean that about 20% of the atmospheric radiation emitted downward (to the surface) in a given spectral range is reflected in the opposite direction upwards (to the atmosphere). With the advent of the spaceborne high-resolution infrared Fourier spectrometers, such as IMG on ADEOS and others the investigation of the effect of reflected radiation flux on the outgoing thermal radiation spectrum became urgent. It should be noted that this effect is considered when observing the Earth from space in the microwave range[7] where the surface emittance is less than unity. It can be assumed that the contribution from this supplementary fairly large radiation flux can also manifest itself in the spectra of thermal radiation of the Earth when sensing from space the atmosphere using the IMG/ADEOS sensor over ecological systems with low emittance in the infrared range.

The goal of this paper is to model the spectra of outgoing thermal radiation at an experimental resolution of 0.1 cm^{-1} over surfaces with the emittance less than 0.9, both with and without the account of the reflected flux, in order to reveal the characteristic properties of this effect in the observed spectra of the Earth's atmosphere using the IMG/ADEOS sensor.

The infrared radiation transfer in the atmosphere with the account for the radiation reflection from surface

In differential form the equation of the radiation transfer in the atmosphere with the account of aerosol extinction is as follows[8]:

$$dW_\nu = -(K_\nu^{gas} + K_\nu^{abs} + K_\nu^{sca})W_\nu\, dz + (K_\nu^{gas} + K_\nu^{abs})B_\nu\, dz, \quad (1)$$

where K_ν^{gas} is the absorption coefficient of the gas components of the atmosphere; K_ν^{abs} and K_ν^{sca} are the

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coefficients of absorption and scattering by aerosol components of the atmosphere; B_ν is the spectral density of the radiant flux emitted by a unit surface of a black body.

When considering the infrared radiation transfer in the atmosphere in a vertical direction with the allowance for the radiant flux reflected from the surface, proportional to the reflectivity of the surface $(1 - \varepsilon_\nu)$, Eq. (1) can be written as[1,7,9]:

$$W_\nu^{\mathrm{up}} = \varepsilon_\nu B_\nu(T_0) \exp\left[-\int_0^H (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}} + K_\nu^{\mathrm{sca}})\mathrm{d}h\right] + (1 - \varepsilon_\nu)\, W_\nu^{\mathrm{down}} \exp\left[-\int_0^H (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}} + K_\nu^{\mathrm{sca}})\mathrm{d}h\right] + \int_0^H (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}}) B_\nu \exp\left[-\int_h^H (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}} + K_\nu^{\mathrm{sca}})\mathrm{d}h'\right]\mathrm{d}h, \quad (2)$$

where ε_ν is the emittance of the Earth's surface; T_0 is the surface temperature; H is the atmospheric altitude, $W_\nu^{\mathrm{down}}(z = 0)$ is the infrared radiation flux emitted by the atmosphere downwards at the surface level:

$$W_\nu^{\mathrm{down}} = \int_0^H (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}}) B_\nu \exp\left[-\int_0^h (K_\nu^{\mathrm{gas}} + K_\nu^{\mathrm{abs}} + K_\nu^{\mathrm{sca}})\mathrm{d}h'\right]\mathrm{d}h. \quad (3)$$

The first term in Eq. (2) stands for the Earth's surface radiation as a gray emitter, the second term stands for the reflection of the downward going infrared radiation from the surface, and the third component stands for the self-radiation of the atmosphere.

The coefficients K_ν^{gas}, K_ν^{abs}, K_ν^{sca} are the functions of altitude. The absorption coefficient K_ν^{gas} is determined by the altitude profiles of temperature and concentration of atmospheric gases absorbing infrared radiation. The profile and type of atmospheric aerosol determine the coefficients K_ν^{abs} and K_ν^{sca}. The blackbody function B_ν is also the function of altitude as it depends on the temperature profile. The gas absorption coefficient is calculated using line-by-line method and the HITRAN–96 spectral database:

$$K^{\mathrm{gas}}(\nu, h) = K_\nu^* + N_0(h) \sum_{i=1}^{Ng} ni(h) \sum_j S_{ij}[T(h)]\, \Phi_{ij}[\nu_{ij} - \nu, T(h), p(h)], \quad (4)$$

where $p(h)$, $T(h)$, $N_0(h) = p(h)/[k_\mathrm{B}T(h)]$ are the pressure, temperature and molecular concentration in the atmosphere at the altitude h; k_B is the Boltzmann constant; n_i is the relative concentration of the ith gas; S_{ij} is the intensity of the jth line of ith gas; ν_{ij} is the wave number of the line center, cm^{-1}; Φ_{ij} is the spectral line shape. The absorption coefficient $K_\nu^* = K_\nu^{\mathrm{cont}} + K_\nu^{\mathrm{mix}}$ in Eq. (4) allows for such supplementary effects as the continuous absorption by water vapor and the line shift in the Q-branches of the CO_2 absorption bands.

Equation (2) is included in the latest version of the FIRE–ARMS[10,11] software and, if necessary, makes it possible to take this effect of the reflected flux into account in modeling the spectrum of thermal radiation outgoing to space.

Results of modeling and comparison with the observed spectra

Figure 1 shows the results obtained by modeling the outgoing radiation within the range from 1115 to 1145 cm^{-1} (the spectral range from Band 3 of IMG) at the resolution of 0.1 cm^{-1} for the surface with the desert emittance taken from HITRAN–96 with and without the allowance for the second component in Eq. (2).

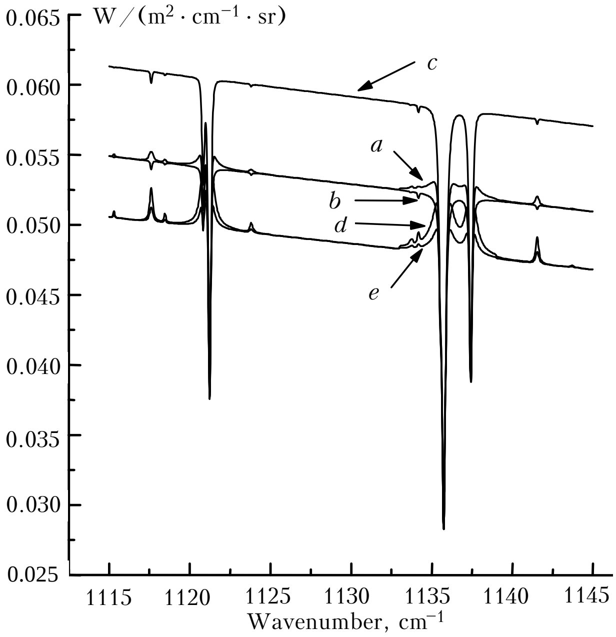

Fig. 1. Model radiation spectra: the fragment of model spectrum over the desert surface in the 1115–1145 cm^{-1} range at the 0.1 cm^{-1} resolution determined by water vapor lines (*a*). The temperature profile is taken from the US standard atmosphere. The surface temperature equals the temperature of the near-ground air. Data on the surface emittance are taken from the HITRAN–96 database for the desert. The contribution of the reflected radiation is taken into account. The same data as for (*a*), but the reflected radiation is not taken into account (*b*); the same data as for (*a*), but the surface temperature $T_\mathrm{s} = 295$ K that is higher than the temperature of the near-ground air T_0 in the model of US standard ($T_0 = 288.2$ K) (*c*); and the case of the temperature inversion (*d*). In the case of standard model the variations are shown in the values of the surface temperature, the temperature of the near-ground air and the air temperature at 1-km altitude, namely, $T_\mathrm{s} = 284$ K, $T_0 = 288$ K, and $T_1 = 287$ K, respectively. The reflected radiation is taken into account. The same as for (*c*), but the reflected radiation is not taken into account (*e*). Inversion of weak lines is not observed close to 1117.6 and 1141.5 cm^{-1}.

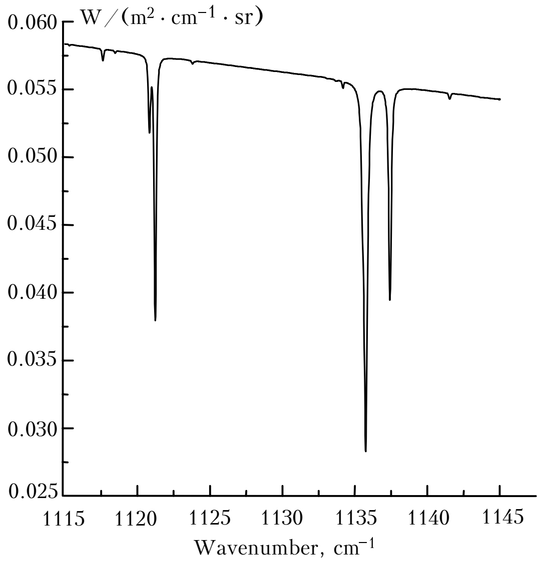

Fig. 2. Fragment of the model spectrum in the range from 1115 to 1145 cm^{-1} at 0.1 cm^{-1} resolution determined by lines of water vapor. The temperature profile is the US standard. The surface temperature equals the temperature of the near-ground air. The value of the surface emittance equals to unity.

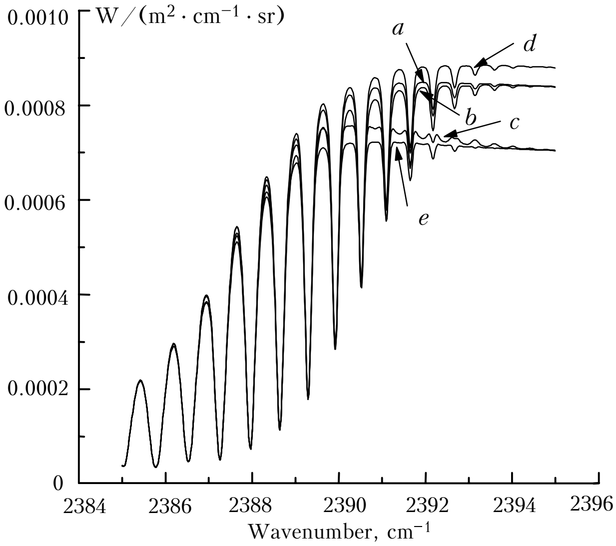

Fig. 3. Model radiation spectra: stands for the fragment of model spectrum over the desert surface in the range from 2385 to 2395 cm^{-1} at 0.1 cm^{-1} resolution determined by the CO_2 lines (*a*). The temperature profile is the US standard. The surface temperature equals the temperature of the near-ground air. Data on the surface emittance were taken from the HITRAN–96 database for the desert. The contribution of the reflected radiation was taken into account; the same data as for (*a*) but the reflected radiation was not considered (*b*); the same data as for (*a*) but the temperature profile is with the temperature inversion (*c*). In the US standard model the values of the surface temperature are varied as well as the values of the temperature of the near-ground air and the air temperature at 1-km altitude; thus T_s = 284 K, T_0 = 288 K, and T_1 = 287 K, respectively. The reflected radiation was accounted for; the same data as for (*c*) but without the account of the reflected radiation (*d*); the same data as for (*c*), but with the value of the surface emittance equal to unity (*e*).

For a comparison the spectrum modeling is given for different surface temperatures at one and the same temperature profile in the atmosphere. It can be seen that the consideration of the reflected radiation increases the model spectrum brightness in the range of wings of strong water vapor lines and results in the inversion of weak lines (the absorption line manifests itself as a radiation line). The contribution of the reflected radiation to the modeled spectrum is considerable if the surface temperature is close to temperature of the near-ground air, while in the case of the temperature inversion this contribution is more important. In the case when the surface temperature is higher than the temperature of the near-ground air, this effect becomes weak. Figure 2 shows for a comparison the results of spectral modeling without the account of the reflected radiation (without the second term in Eq.(2)), but with the surface emittance, $\varepsilon_\nu = 1$.

From the comparison of data obtained it is clear that the effect, associated with the reflection of infrared radiation from the surface, cannot be gained by increasing the surface emittance. Figure 3 shows the calculated spectra of outgoing radiation in the 2385–2395 cm^{-1} range (the spectral range from Band 2, IMG), demonstrating the necessity of the account for the reflected radiation in using this range, for example, for reconstructing the temperature profile.

Figure 4 shows typical spectra of the outgoing radiation observed over Sahara and tropical ocean at nighttime using the IMG sensor from an ADEOS satellite. At nighttime the surface temperature is close to the temperature of the near-ground air, and in the given spectral range the typical effects are clearly seen over Sahara due to the contribution of the reflected radiation. In the spectra over the ocean this effect is negligible because of high ocean emittance.

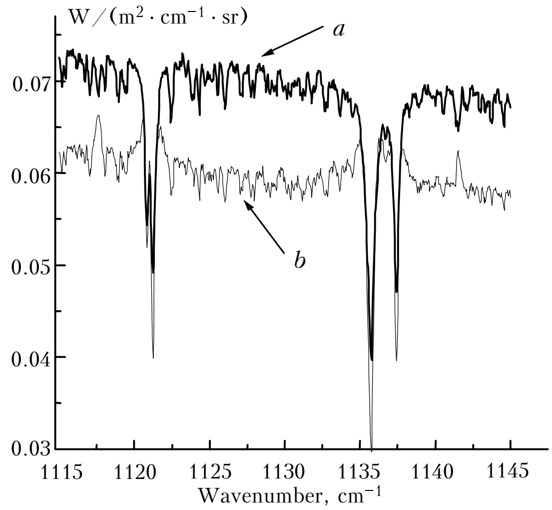

Fig. 4. Fragments of the night-time spectrum (1115–1145 cm^{-1}): night-time spectrum observed by the IMG sensor over the ocean; the surface temperature is about 295 K (*a*); the same fragment of the night spectrum over the Sahara desert (*b*). The surface temperature is about 295 K. The inversion is observed of the same weak lines as shown in the model spectra in Fig. 1.

It should be noted that when reconstructing vertical profiles of atmospheric components, e.g., water vapor, the result depends on the profile of an observed spectral line, and, in particular, on the behavior of line wings. Consequently, the account for reflected radiation for such ecological systems as deserts, broad-leaved forests or agricultural lands enables one to obtain a more correct behavior of the line wings in the spectrum of outgoing radiation and, as a result, to decrease an error when determining the humidity profile in the lower atmosphere. However, it should be kept in mind that at the temperature inversion the model spectrum of outgoing radiation, without the account of the contribution of the reflected flux, has the same peculiarities as in the case of the lack of inversion, but with the account of the reflected radiation. This fact should also be considered when reconstructing the temperature profile.

In conclusion, it should be noted that in spectral ranges with a relatively low absorption of the atmosphere it is necessary to consider the reflected radiation effect if the emittance of an observed surface is less than 0.9. This can be especially urgent when sounding in such spectral ranges as Band 2 and Band 1 of the IMG sensor or in the same spectral ranges of the IASI[12] and TES[13] sensors.

References

1. K.Ya. Kondrat'ev and Yu.M. Timofeev, *Meteorological Sensing of the Atmosphere from Space* (Gidrometeoizdat, Leningrad, 1978), 280 pp.

2. W.L. Smith, H.M. Woolf, and H.E. Revercomb, Appl. Opt. **30**, No. 9, 1117 (1991).

3. A.B. Uspenskii, A.N. Trotsenko, A.N. Rublev, S.V. Romanov, and P.Yu. Romanov, Issled. Zemli iz Kosmosa, No. 2, 3 (1998).

4. *IMG Project Technical Report, IMG Mission Operation & Verification Committee.* CRIEPI, March 30, Japan 1999.

5. X.L. Ma, T.J. Schmit, and W.L. Smith, J. Appl. Meteorol. **38**, 501 (1999).

6. L.I. Rothman, C.P. Rinsland, A. Goldman, S.T. Massie, D.P. Edwards, et al., J. Quant. Spectrosc. Radiat. Transfer **60**, 665–710 (1998).

7. J.B. Kotlyar and B.L. Novak, Issled. Zemli iz Kosmosa, No. 6, 57 (1987).

8. K.G. Gribanov, V.I. Zakharov, Kh. Kobayashi, and A. Shimota, Atmos. Oceanic Opt. **13**, No. 10, 842–845 (2000).

9. Yu.M. Timofeev and A.A. Martynov, Issled. Zemli iz Kosmosa, No. 4, 12 (1996).

10. K.G. Gribanov, V.I. Zakharov, and S.A. Tashkun, Atmos. Oceanic Opt. **12**, No. 4, 358 (1999).

11. K.G. Gribanov, V.I. Zakharov, S.A. Tashkun, and Vl.G. Tyuterev, J. Quant. Spectrosc. Radiat. Transfer **68**, No. 4, 435 (2000).

12. F. Henault, D. Miras, D. Scheidel, and F. Boubolt, in: *Proceedings of the 6th Int. Workshop on Atmospheric Science from Space Using Fourier Transform Spectrometry,* San Juan Capistrano, USA (1995).

13. R. Beer, in: *Proceedings of the 5th Workshop on Atmospheric Science from Space Using Fourier Transform Spectrometry,* Tokyo, Japan (1994).