THERMAL RADIATION OF THE EARTH SCATTERED BY AEROSOL LAYERS

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Figure 2: Fig. 2. Relative spectral sensitivity of the radiometer

Abstract

Full Text

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THERMAL RADIATION OF THE EARTH SCATTERED BY AEROSOL LAYERS

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In studies of atmospheric radiation, one usually considers its own radiation (thermal, emission, auroras) and the scattered solar component, which chiefly determine the optical properties of the atmosphere. Indeed, during the day in the short-wave region of the spectrum $\lambda < 4\ \mu$ the most substantial contribution to atmospheric radiation is made by the scattered solar component, while in the wavelength region $\lambda > 4\ \mu$ it is made by the thermal radiation of the atmosphere. At night, in the short-wave region of the spectrum $\lambda < 2\ \mu$, emission radiation and auroras are most intense, while for wavelengths $\lambda > 2\ \mu$ it is the thermal radiation of the atmosphere.

Fig. 1. Radiometer: 1 –protective window; 2 –objective; 3 –optical wedge; 4 –diaphragm; 5 –light filter; 6 –PbTe photoresistor (–190°); 7 –amplifier; 8 –loop oscillograph

Fig. 2. Relative spectral sensitivity of the radiometer

However, in some cases a definite role is also played by the Earth' s thermal radiation scattered by the atmosphere, concentrated in the infrared region of the spectrum. Atmospheric scattering in the region of the Earth's thermal radiation is practically completely determined by aerosol; therefore the Earth' s thermal radiation scattered by the atmosphere will make the greatest contribution primarily in aerosol layers. Naturally, this radiation will be more noticeable in those parts of the spectrum and at those altitudes where it is more intense than the atmosphere' s own radiation and the scattered solar component. The most favorable conditions for observing it occur at night in the atmospheric transmis-

Fig. 3. Results of measurements of the difference of effective radiances: a—at an altitude of 8 km (25 May 1964); b—at an altitude of 9 km (13 May 1964).

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sion bands at altitudes where the atmospheric temperature is sufficiently low. At the same time, the temperature of the Earth' s surface must not be very low. In middle latitudes all these conditions, in particular, are fulfilled in summer during the night hours in the spectral region 3.5-5.0 μ at altitudes of 9-20 km, where a number of investigators ([1–6]) have found clearly expressed aerosol layers in the region of the tropopause and the lower stratosphere.

Experimental studies of atmospheric radiation in the spectral region 3.5-5.2 μ were carried out at night with an infrared radiometer with

from an aircraft at altitudes of 8-9 km in May 1964. The optical layout of the radiometer and the block diagram of the measuring apparatus are shown in Fig. 1; the spectral sensitivity, in Fig. 2. The field of view of the radiometer was $1^{\circ}30'$; changing the direction of the radiant flux was accomplished by rotating an optical wedge, which made it possible to compare the effective radiance B_{eff} of two adjacent portions of the sky separated from one another by an angle $\Delta\theta = 1^{\circ}$.

The results of measurements of ΔB_{eff} for zenith angles $\theta = 50$–90° at aircraft flight altitudes of 8-9 km are presented in Fig. 3. The measurements were made at night on 13 and 25 May 1964 in the Moscow region in the absence of cloud cover. The air temperature at an altitude of 8-9 km was about -45°, and the temperature of the Earth' s surface was 9–11°. According to existing views, in the spectral region 3.5–5.2 μ the most substantial contribution to the radiation of the night sky at an altitude of 8-9 km is made by the thermal radiation of the atmosphere. However, the difference of effective radiances of the night sky recorded in our measurements is considerably larger (approximately by an order of magnitude) than would be expected from the thermal radiation of the atmosphere at a temperature of -45°.

Fig. 3. Results of measurements of the difference of effective radiances: *a*—at an altitude of 8 km (25 May 1964); *b*—at an altitude of 9 km (13 May 1964).

Apparently, under these conditions the most substantial contribution to the atmospheric radiation is made by the Earth' s thermal radiation scattered by aerosol layers of the tropopause and the lower stratosphere. The effective radiance of the Earth's thermal radiation scattered by aerosol layers, when measured at a given altitude and under a given meteorological situation, depends chiefly on the zenith angle θ

$$B_{\text{eff}}(\theta) = B_{\text{eff}}(0) f(\theta), \tag{1}$$

Fig. 4. Effective radiance of the Earth' s thermal radiation scattered by aerosol layers

Figure 4: Fig. 4. Effective radiance of the Earth' s thermal radiation scattered by aerosol layers

where $B_{\text{eff}}(0)$ is the effective radiance at the zenith.

The function $f(\theta)$ for angles θ not exceeding 80° may be represented in the form

$$f(\theta) = (1 - \exp(-\tau \sec \theta))/(1 - \exp(-\tau)), \tag{2}$$

where τ is the optical thickness of the aerosol layers for the working spectral region of the radiometer when observing at the zenith. Then the difference of effective radiances is

$$\Delta B_{\text{eff}}(\theta) = B_{\text{eff}}(0) f'(\theta) \Delta\theta. \tag{3}$$

From equalities (1)–(3), after the simplest transformations, we obtain an expression for the effective radiance

$$B_{\text{eff}}(\theta) = \Delta B_{\text{eff}}(\theta) \cos(\theta) \left[1 - \exp(-\tau \sec \theta)\right]/\tau \operatorname{tg} \theta \exp(-\tau \sec \theta). \tag{4}$$

To calculate the effective radiance from formula (4), it is necessary to estimate the optical thickness of the aerosol layers, τ. This estimate was made from the scatter of the values $B_{\text{eff}}(0)$, calculated for various τ from the results of measurements of $\Delta B_{\text{eff}}(\theta)$:

$$B_{\text{eff}}(0) = \Delta B_{\text{eff}}(\theta) \cos \theta [1 - \exp(-\tau)]/\tau \operatorname{tg} \theta \exp(-\tau \sec \theta). \tag{5}$$

With a correctly chosen τ, the quantities $B_{\text{eff}}(0)$ calculated from the experimental values of $\Delta B_{\text{eff}}(\theta)$ for various angles θ should be identical. In processing the results of our measurements, this condition

is satisfied by the value $\tau \approx 0.15$. Then $B_{\text{eff}}(0) = 12\ \mu\text{W} \cdot \text{cm}^{-2} \cdot \text{sr}^{-1}$, and the dependence $B_{\text{eff}}(\theta)$ calculated from formula (4) is shown in Fig. 4. For the zenith angle $\theta = 80°$, the effective radiance of the Earth' s thermal radiation scattered by aerosol layers is about 50 $\mu\text{W} \cdot \text{cm}^{-2} \cdot \text{sr}^{-1}$.

Fig. 4. Effective radiance of the Earth' s thermal radiation scattered by aerosol layers

The measurement of the effective radiance also includes a small component due to the thermal radiation of the atmosphere. It can be shown that, at an air temperature at an altitude of 8-9 km of (−45°), it does not exceed

1 μW $\cdot$ cm^{-2} $\cdot$ sr^{-1} for the zenith, and 5 μW $\cdot$ cm^{-2} $\cdot$ sr^{-1} for the zenith angle $\theta = 80^\circ$.

In (7) are given the experimental and calculated values of the spectral intensity of the radiance of the atmosphere' s thermal radiation from the Earth, at a temperature of the near-surface layer of about 280° K. It is noted here that the agreement between the calculated and experimental data is satisfactory, but not so good as to regard the results obtained as exhaustive. A particularly noticeable discrepancy between the calculated and experimental values is observed in the atmospheric transmission band 8–12 μ for the zenith direction $\theta = 0$ (elevation angle 90° (7)), where the experimental values exceed the calculated ones by almost a factor of two (by $50 \div 100$ μW$\cdot$cm$^{-2}\cdot$sr^{-1} μ^{-1}). This excess can apparently be explained by the Earth' s thermal radiation scattered by aerosol. From the difference between the experimental and calculated values of the spectral radiance intensity one can roughly estimate the optical thickness of the aerosol at the zenith as $\tau = 0.10$–0.15.

In the direction of the zenith angle $\theta = 82^\circ$ (elevation angle 8°) (7), the experimental and calculated values of the spectral intensity of the atmosphere' s thermal radiation agree satisfactorily. This is explained by the fact that, in observations from the Earth in this direction, absorption and the atmosphere' s own radiation make it difficult to separate out the Earth' s thermal radiation scattered by aerosol.

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Note: Figure translations are in progress. See original paper for figures.