# DETERMINATION OF GEOPHYSICAL PARAMETERS FROM MEASUREMENTS OF THERMAL RADIO EMISSION ON THE “KOSMOS-243” ARTIFICIAL EARTH SATELLITE

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## Abstract

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A. E. BASHARINOV, A. S. GURVICH, S. T. EGOROV

# DETERMINATION OF GEOPHYSICAL PARAMETERS FROM MEASUREMENTS OF THERMAL RADIO EMISSION ON THE "KOSMOS-243" ARTIFICIAL EARTH SATELLITE

*(Presented by Academician L. M. Brekhovskikh, 3 VI 1969)*

Measurements of the radio-thermal radiation of the Earth and its atmosphere were carried out on the "Kosmos-243" artificial satellite at wavelengths $\lambda_1 = 8.5$, $\lambda_2 = 3.4$, $\lambda_3 = 1.35$, $\lambda_4 = 0.8$ cm. The measurements were performed with the aim of developing methods for geophysical studies of the atmosphere and the underlying surface from outgoing radiation in the radio-wave range.

To carry out the measurements, four radiometers were installed on board the satellite, with a sensitivity of 0.7° K at wavelengths $\lambda_1$ and $\lambda_2$, and about 2° K at the short wavelengths. The amplification of the radiometers was periodically monitored. At all wavelengths except $\lambda_1$, the radiometer antennas had the same half-power beamwidth, about 3.5°. At wavelength $\lambda_1$ the beam was 2.5 times wider. All antennas were directed toward the nadir. The satellite was launched on 23 IX 1968 into an orbit whose plane was inclined to the equatorial plane by 71.3°. The apogee altitude was 319 km and the perigee 210 km. The measurement data were accumulated in a storage device and then transmitted over a radio-telemetry link.

At wavelengths $\lambda_1$ and $\lambda_2$, absorption in the atmosphere is small; therefore, for the brightness temperature $T_{,\lambda}$ of the outgoing radiation one may write the following approximate formula:

$$T_{,\lambda} = T\ (1 - R_\lambda^2),$$

where $T$ is the effective temperature of the surface (depending on the penetration depth of the radio waves and the temperature distribution in the near-surface layer), and $R_\lambda^2$ is the energy reflection coefficient. For seawater at centimeter wavelengths, $1-R_\lambda^2$ lies in the range $0.35 \div 0.45$, while for sea ice floating on the ocean surface, $R_\lambda^2$ is close to zero. The change in brightness temperature caused by variations of the reflection coefficient was used to determine the characteristics of the ice around Antarctica. The use of radiation at wavelengths $\lambda_1$ and $\lambda_2$ eliminated the influence of cloudiness on these measurements.

The records obtained clearly show an increase in brightness temperature when ice appears in the antenna field of view. During flight over a mass of floating ice, decreases in brightness temperature are also observed, indicating the presence of leads. This makes it possible to estimate the degree of ice compactness. Despite the absence of scanning, from the data obtained it proved possible to determine the ice boundary around Antarctica with two gradations of compactness, 1–5 and 5–10 points, and to mark, within the mass of floating ice, several regions with leads (Fig. 1). It should also be noted that the radio-brightness temperature of the continent of Antarctica differs strongly within the wavelength spectrum: at the shortest wavelength, $\lambda_4$, it is in some cases 40° K lower than at wavelength $\lambda_1$. This difference is apparently determined by the depth of penetration of radio waves into continental ice and by the peculiarities of the temperature regime of Antarctica.

At the short wavelengths $\lambda_3$ and $\lambda_4$, where absorption in the atmosphere must be taken into account, the brightness temperature of the outgoing radiation, without taking into account the inessential

···scattering effects, as a rule, is expressed in the form

$$T_{,\lambda} = T\ (1 - R_\lambda^2) e^{-\tau_\lambda} + \int_0^\infty T(z)\gamma_\lambda(z) \exp\left[-\int_z^\infty \gamma_\lambda(z')\,dz'\right] dz +$$

$$+ R_\lambda^2 e^{-\tau_\lambda} \int_0^\infty T(z)\gamma_\lambda(z) \exp\left[-\int_0^z \gamma_\lambda(z')\,dz'\right] dz, \qquad (1)$$

where $T(z)$ is the temperature at height $z$; $\gamma_\lambda = \gamma_{\lambda O_2} + \gamma_{\lambda Q} + \gamma_{\lambda W}$ is the absorption coefficient, composed of the relatively slowly varying absorption in oxygen, $\gamma_{\lambda O_2}$, water vapor, $\gamma_{\lambda Q}$, and liquid water in hydrometeors, $\gamma_{\lambda W}$; $\tau_\lambda = \int_0^\infty \gamma_\lambda(z)\,dz$. Since under standard conditions $\gamma_{\lambda_3 Q}/\gamma_{\lambda_4 Q} \approx 5$ and $\gamma_{\lambda_3 W}/\gamma_{\lambda_4 W} \approx 0.4$, $\tau_{\lambda_3 Q} \approx 0.12$, it proves possible to determine the integral content of water vapor and liquid water from the spectrum of radio emission. These measurements are most effective over the ocean, where $T_{,\lambda}$ of the surface is considerably lower than its thermodynamic temperature and, as is evident from (1), the presence of water vapor and liquid water in the atmosphere causes an increase in the outgoing radiation. Analysis of the values of $T_{,\lambda}$ at all four wavelengths over a day of flight showed that over the oceans in the Southern Hemisphere the boundary between moist tropical and dry polar air is clearly traced, observed from the sharp change in $T_{,\lambda_3}$. This boundary (Fig. 1) has a width of about 700–1000 km.

**Fig. 1.** Characteristics of the field of drifting ice around Antarctica and the boundary between dry and moist air in the Southern Hemisphere. **1** –ice with

Figure 1

Figure 1: Figure 1

concentration of 1-5 tenths; **2** –5-10 tenths; **3** –openings; **4** –boundary of moist and dry air

To carry out calibrations of shortwave radiometers over the ocean with the aid of an infrared-range radiometer operating in the band⋯

10–12 $\mu$m, cloudless areas were identified, over which it proved possible to calibrate the radiometer scales in brightness temperatures using mean climatological data. The large number of such areas (more than 80 during the first day of flight) made it possible to tie the radiometer scales with an accuracy of about 1–1.5° of brightness temperature. The resulting scale was used to determine the integral content of water vapor in the vertical column of the atmosphere along flight segments over the oceans where radiosounding was carried out. The use of the wavelengths $\lambda_2$, $\lambda_3$, $\lambda_4$ made it possible to perform these determinations also in the presence of cloudiness. To estimate the accuracy, a comparison was made between data obtained with the "Kosmos-243" artificial Earth satellite and radiosonde data over water areas (at a distance of no more than 150 km from the subsatellite point and differing in time by no more than 3 hours from the moment of the satellite pass). The good agreement between the satellite and radiosonde data (Fig. 2) shows that ties to mean climatological data make it possible to obtain a reliable calibration and to determine the integral content of water vapor in the atmosphere from measurements of outgoing radio emission. Simultaneous measurement of $T_{\mathrm{a},\lambda_3}$ and $T_{\mathrm{a},\lambda_4}$ made it possible to obtain information on the vertical distribution of water vapor in a cloudless atmosphere.

**Fig. 2.** Comparison of satellite and radiosonde measurements of the integral moisture content $Q$. 1 –measurements under weak cloudiness; 2 –under heavy cloudiness; straight line –exact agreement.

A clearly expressed correlation is observed between variations of the measured moisture-field profiles and latitudinal changes and variations of the temperature field of the underlying sea surface. For example, Fig. 3 presents profiles, obtained by processing radiometric data, of temperature, moisture, and cloud-water content over the Atlantic Ocean.

**Fig. 3.** Profiles of the ocean surface temperature $T_p$, moisture field $Q$, and hydrometeor field obtained over the Atlantic Ocean. $W$ –integral water content; $c$ –continuous cloudiness, $z$ –considerable cloudiness, $n$ –slight cloudiness.

In individual cases, however, the displacement of masses of dry and moist air leads to a loss of agreement in the measurements of the temperature field and the moisture field. Such a loss is visible in Fig. 3 in the region of 16° N, where there is a deep minimum in the moisture field with a radius of about 300 km.

Detection of hydrometeors was carried out from characteristic peaks on the

Fig. 4. Profiles of radio-brightness temperature: a —Australian continent, $T_{,\lambda_2}$; b —northern region of South America, $T_{,\lambda_2}$; c —in the zone of central Africa, $T_{,\lambda_2}$

Figure 2: Fig. 4. Profiles of radio-brightness temperature: a —Australian continent, $T_{,\lambda_2}$; b —northern region of South America, $T_{,\lambda_2}$; c —in the zone of central Africa, $T_{,\lambda_2}$

profiles of radio-brightness temperatures $T_{,\lambda_4}$, where the signal intensity is determined by the water content in hydrometeor formations. No fewer than 10 gradations of hydrometeor intensity were observed, with cloud water content from 0.03 to 0.3 g/cm$^2$.

**Fig. 4.** Profiles of radio-brightness temperature:
*a* —Australian continent, $T_{,\lambda_2}$; *b* —northern region of South America, $T_{,\lambda_2}$; *c* —in the zone of central Africa, $T_{,\lambda_2}$.

The location of zones of hydrometeor formations correlates with the location of regions of increased cloud humidity (see, for example, Fig. 3, 45° N and 60° N).

Rain clouds and precipitation zones are characterized by a jagged profile shape (Fig. 3, 60° N). Frontal cloudiness is characterized by a comparatively smooth profile shape, extending over several hundred kilometers. The predominant share of the recorded intense hydrometeor formations belongs to the equatorial zone from 10° S to 15° N.

Radio-brightness profiles over homogeneous regions of continents, such as the Sahara zone, the Australian continent, Eastern Siberia, Antarctica, etc., reflect measurements of the temperature of the underlying surface. Figure 4a presents the radio-brightness profile obtained during a pass over Australia. The change in radio brightness from the southern to the northern coast is 16–18°. According to ground meteorological-station data, the difference in near-surface air temperature for these points is +21°; the difference in mean monthly temperatures is 11–13°.

In mountainous regions (the Himalayan Range, the Andes, etc.) a decrease in radio brightness is observed when the field of view moves onto high-altitude regions, owing to the vertical temperature gradient of the atmosphere.

A decrease in radio brightness of continental regions by 30–50°K is observed in places where the cover is moistened, where the degree of blackness does not exceed values of 0.7–0.8. Thus, for example, on the radio-brightness profile during a pass over the northern part of South America in the zone of the left-bank channels of the Orinoco River (6–10° N; 65–70° W), a deep depression in radio brightness is observed (see Fig. 4b). Depressions in the radio-brightness profile are also observed during a pass over Central Africa in the left-bank zone of the Nile River (7–15° N, 23–28° E), see Fig. 4c. In photographs of this region obtained from meteorological satellites, cumulus cloudiness is detected; in this connection it should be considered probable that the soil was moistened as a result of precipitation.

Institute of Atmospheric Physics
Academy of Sciences of the USSR

Institute of Radio Engineering and Electronics
Academy of Sciences of the USSR
Moscow

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