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# THE PROBLEM OF THE ANGULAR DISTRIBUTION OF ASCENDING LONG-WAVE RADIATION

E. P. Barashkova, et al

Foreign Technology Division
Wright-Patterson Air Force Base, Ohio

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# FOREIGN TECHNOLOGY DIVISION

# THE PROBLEM OF THE ANGULAR DISTRIBUTION OF ASCENDING LONG-WAVE RADIATION

by

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G. P. Sidorenko



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**13. ABSTRACT**

Measurements were carried out, from on board an aircraft, of the brightness of ascending long-wave radiation in the 3-30 micrometer waveband at different angles α over different types of underlying surface. A clear dependence was found in the nature of the angular distribution of this radiation on atmospheric stratification. Changes in the water-vapor content were not observed to play an appreciable role in the mechanism involved, with the angular distribution of the ascending long-wave radiation at different levels of the troposphere basically determined by the vertical temperature profile.

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| Long Wave Radiation | | | | | | |



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By: Ye. P. Barashkova, L. I. Prokov'yeva,
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## U. S. BOARD ON GEOGRAPHIC NAMES TRANSLITERATION SYSTEM

| Block | Italic | Transliteration | Block | Italic | Transliteration |
|---|---|---|---|---|---|
| А а | *А а* | A, a | Р р | *Р р* | R, r |
| Б б | *Б б* | B, b | С с | *С с* | S, s |
| В в | *В в* | V, v | Т т | *Т т* | T, t |
| Г г | *Г г* | G, g | У у | *У у* | U, u |
| Д д | *Д д* | D, d | Ф ф | *Ф ф* | F, f |
| Е е | *Е е* | Ye, ye; E, e* | Х х | *Х х* | Kh, kh |
| Ж ж | *Ж ж* | Zh, zh | Ц ц | *Ц ц* | Ts, ts |
| З з | *З з* | Z, z | Ч ч | *Ч ч* | Ch, ch |
| И и | *И и* | I, i | Ш ш | *Ш ш* | Sh, sh |
| Й й | *Й й* | Y, y | Щ щ | *Щ щ* | Shch, shch |
| К к | *К к* | K, k | Ъ ъ | *Ъ ъ* | " |
| Л л | *Л л* | L, l | Ы ы | *Ы ы* | Y, y |
| М м | *М м* | M, m | Ь ь | *Ь ь* | ' |
| Н н | *Н н* | N, n | Э э | *Э э* | E, e |
| О о | *О о* | O, o | Ю ю | *Ю ю* | Yu, yu |
| П п | *П п* | P, p | Я я | *Я я* | Ya, ya |

* ye initially, after vowels, and after ъ, ь; e elsewhere. When written as ё in Russian, transliterate as yё or ё. The use of diacritical marks is preferred, but such marks may be omitted when expediency dictates.

FOLLOWING ARE THE CORRESPONDING RUSSIAN AND ENGLISH DESIGNATIONS OF THE TRIGONOMETRIC FUNCTIONS

| Russian | English |
|---|---|
| sin | sin |
| cos | cos |
| tg | tan |
| ctg | cot |
| sec | sec |
| cosec | csc |
| sh | sinh |
| ch | cosh |
| th | tanh |
| cth | coth |
| sch | sech |
| csch | csch |
| arc sin | $\sin^{-1}$ |
| arc cos | $\cos^{-1}$ |
| arc tg | $\tan^{-1}$ |
| arc ctg | $\cot^{-1}$ |
| arc sec | $\sec^{-1}$ |
| arc cosec | $\csc^{-1}$ |
| arc sh | $\sinh^{-1}$ |
| arc ch | $\cosh^{-1}$ |
| arc th | $\tanh^{-1}$ |
| arc cth | $\coth^{-1}$ |
| arc sch | $\text{sech}^{-1}$ |
| arc csch | $\text{csch}^{-1}$ |

---

| Russian | English |
|---|---|
| rot | curl |
| lg | log |

# THE PROBLEM OF THE ANGULAR DISTRIBUTION OF ASCENDING LONG-WAVE RADIATION

Ye. P. Barashkova, L. I. Prokov'yeva,
G. P. Sidorenko

In 1967, using equipment installed on an IL-18 aircraft, the Main Geophysical Observatory (MGO) carried out measurements of the brightness $I_\alpha$ of ascending longwave radiation in the 3-30 micrometer spectral region at different angles α over different types of underlying surface. For these brightness measurements the scanning equipment used was of the same type as that installed on meteorological satellites [1]. The equipment was calibrated in accordance with the method discussed in an article [2] by L. B. Krasil'shchikov. The view angle of the scanner was 4 × 5°, with the scanning conducted in the plane perpendicular to the direction of flight. The duration of a single complete scan from +90 to -90° (reading the angles from the nadir) was 7.5 s. The working scanning angles α were limited by the sides of the aircraft nascelle. Measurements were made at different altitudes from 500 to 9000 meters. Air speed was about 500 km/h.

Fig. 1 shows an example of a brightness recording in relative units from different altitudes above two types of underlying surface. The results of these measurements were used to estimate the relative angular distribution of long-wave radiation $\frac{I_\alpha}{I_0}$, where $I_0$ is the brightness at the nadir. Because of systematic errors in instrument readings for large negative angles, the results for negative scanning angles were excluded from further consideration. It should also be noted that the analysis was based not on the findings of a single scan, but on the average of ten successive scans under invariable measurement conditions.

A total of 106 averaged series were considered, corresponding to various conditions of observation, for scanning angles $\alpha$ of 0, 10, 20, 30, 40, 50, and 60°. Analysis of these data supports the following conclusions.

1. The value of the ratio $\frac{I_\alpha}{I_0}$, for constant scan angles, changes within fairly wide limits. As is evident from the maximum and minimum values for the ratio $\frac{I_\alpha}{I_0}$ cited in Table 1, even with a constant observation altitude ($\alpha = 10°$) above the same surface $0.893 \leq \frac{I_\alpha}{I_0} \leq 1.070$.

2. A different behavior is observed on the part of the ratio $\frac{I_\alpha}{I_0}$ in different situations as a function of $\alpha$. In the majority of cases the ratio $\frac{I_\alpha}{I_0}$ decreases as $\alpha$ increases. In a number of instances, noted mainly during low-altitude measurements, $\frac{I_\alpha}{I_0}$ is seen to increase along with $\alpha$. In measurements over a variegated underlying surface, chaotic fluctuations are found in $\frac{I_\alpha}{I_0}$ as $\alpha$ changes. We should expect that the most correct behavior of $\frac{I_\alpha}{I_0}$ as a function of $\alpha$ would be observed in measurements over a uniform underlying surface, although, as indicated by Table 2, these cases also display a deviation from the monotonic development of the $\frac{I_\alpha}{I_0}$ ratio depending on $\alpha$.

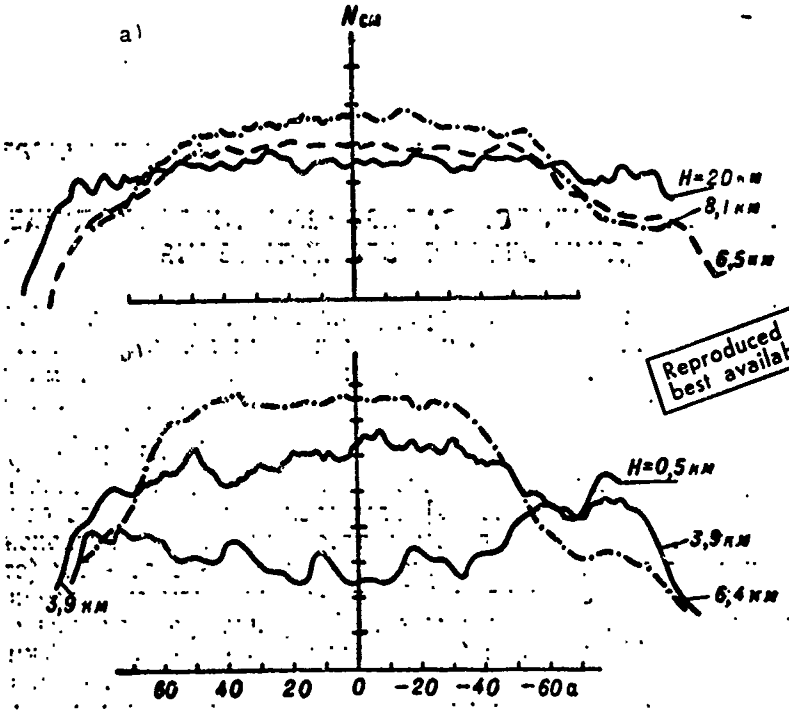




Fig. 1. Example of the recording of the angular distribution of ascending long-wave radiation: 1 - Sea of Okhotsk, 16 September 1967; b - desert, 6 December 1967.

Table 2 lists average values of $\frac{I_\alpha}{I_0} = f(\alpha)$ for different altitudes, abtained in 1967 during vertical soundings of the atmosphere over the Caspian Sea on 24 June, over the Sea of Okhotsk on 16 September, and over the Central Kara Kum Desert on 6 December. Along with the $\frac{I_\alpha}{I_0}$ values, Table 2 gives the aircraft altitude H, the air temperature at the flight altitude $t_в$, the temperature of the underlying surface $t_п$, and the difference $t_п - t_в = \Delta$. The underlying surface temperature was determined on dry land from the data of meteorclogical stations situated along the flight route, and at sea from the results of observations reported by ships and coastal stations. The air temperature was obtained from aircraft meteorograph recordings

Table 1. Extreme values of $\frac{I_a}{I_0}$.

| $a$ | $\max \frac{I_a}{I_0}$ | $H$ | Underlying surface | $\min \frac{I_a}{I_0}$ | $H$ | Underlying surface |
|---|---|---|---|---|---|---|
| 10 | 1,07 | 330 | Sea | 0,89 | 330 | Sea |
| 20 | 1,06 | 330 | Steppe | 0,88 | 330 | • |
| 30 | 1,12 | 450 | Desert | 0,88 | 7783 | • |
| 40 | 1,06 | 330 | Sea | 0,87 | 7783 | • |
| 50 | 1,14 | 330 | • | 0,83 | 9000 | Clouds |
| 60 | 1,16 | 450 | Desert | 0,72 | 9000 | • |

Table 2. Relative angular distribution of ascending long-wave radiation.

| $H$ | ratio $\frac{I_a}{I_0}$ for $a°$ | | | | | | | $t_в$ | $t_н$ | $\Delta t$ | $\gamma_{cp}$ °/км |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | 0 | 10 | 20 | 30 | 40 | 50 | 60 | | | | |
| | | | | Caspian Sea | | | | | | | |
| 500 | 1,00 | 1,00 | 0,99 | 0,99 | 0,99 | 0,99 | 0,99 | 20,4 | 19,0 | −1,4 | −2,8 |
| 3300 | 1,00 | 1,00 | 0,99 | 0,99 | 0,98 | 0,97 | 0,96 | 6,8 | 19,0 | 12,2 | 3,7 |
| 5500 | 1,00 | 0,99 | 0,97 | 0,96 | 0,94 | 0,92 | 0,89 | 0,3 | 19,0 | 18,7 | 3,4 |
| 7800 | 1,00 | 1,00 | 0,90 | 0,88 | 0,87 | 0,83 | 0,79 | −13,3 | 19,0 | 32,3 | 4,2 |
| | | | | Sea of Okhotsk | | | | | | | |
| 200 | 1,00 | 1,00 | 1,00 | 1,00 | 0,98 | 0,97 | 0,97 | 10,6 | 10,0 | −0,6 | −3,0 |
| 2000 | 1,00 | 1,00 | 1,00 | 1,01 | 1,01 | 0,99 | 0,98 | 1,8 | 10,0 | 8,2 | 4,1 |
| 5100 | 1,00 | 0,99 | 1,00 | 1,00 | 0,99 | 0,97 | 0,90 | −9,2 | 10,3 | 19,5 | 3,8 |
| 6600 | 1,00 | 1,00 | 1,00 | 1,00 | 0,97 | 0,95 | 0,89 | −20,0 | 12,0 | 32,0 | 4,9 |
| 8000 | 1,00 | 0,99 | 1,00 | 0,99 | 0,97 | 0,93 | 0,87 | −30,8 | 13,5 | 44,3 | 5,5 |
| | | | | Central Kara Kum | | | | | | | |
| 450 | 1,00 | 1,00 | 1,01 | 1,02 | 1,02 | 1,04 | 1,11 | 16,0 | 2,5 | −13,5 | −30,0 |
| 2700 | 1,00 | 0,99 | 0,98 | 0,97 | 0,99 | 1,00 | 0,98 | −1,6 | 2,0 | 3,6 | 1,3 |
| 3900 | 1,00 | 1,01 | 1,00 | 1,00 | 1,00 | 1,00 | 0,95 | −11,8 | 6,0 | 17,8 | 4,6 |
| 6400 | 1,00 | 1,00 | 0,99 | 0,99 | 0,96 | 0,92 | 0,83 | −32,0 | 8,0 | 40,0 | 6,3 |

Desination: Subscript letters "cp" = "average".

The observations over the Caspian in the lower, near-water, layer (0-0.5 km) reveal a slight inversion; there is virtually no dependence of $\frac{I_a}{I_0}$ on $a$ at the 0.5 km altitude, but at higher levels $\frac{I_a}{I_0}$ falls off as the angle increases. Over the Sea of Okhotsk there is practically no change in $\frac{I_a}{I_0}$ at small angles; beginning with $a = 40°$, $\frac{I_a}{I_0}$ decreases as the angle continues to increase. In readings over Kara Kum, where a powerful inversion was noted in the lower half-kilometer layer, an increase in $\frac{I_a}{I_0}$ was observed at an altitude of 0.5 as a function of increasing $a$. At altitudes of 2700 and 3900 m the $\frac{I_a}{I_0}$ ratio differs little from unity, while at 6400 m there was a decrease in $\frac{I_a}{I_0}$ with increasing $a$.

Table 3. Mean dependence of $\frac{I_a}{I_0}$ on $a$ and $\Delta t$.

| $a°$ \ $\Delta t°$ | —10 | 0 | 10 | 20 | 30 | 40 | 50 | 60 | 70 |
|---|---|---|---|---|---|---|---|---|---|
| 0 | 1,000 | 1,000 | 1,000 | 1,000 | 1,000 | 1,000 | 1,000 | 1,000 | 1,000 |
| 10 | 1,002 | 1,000 | 0,999 | 0,998 | 0,997 | 0,996 | 0,995 | 0,994 | 0,993 |
| 20 | 1,005 | 1,000 | 0.999 | 0,997 | 0,995 | 0,992 | 0,990 | 0,987 | 0,985 |
| 30 | 1,006 | 1,000 | 0,995 | 0,990 | 0,985 | 0,980 | 0,975 | 0,970 | 0,965 |
| 40 | 1,010 | 1,000 | 0,990 | 0,980 | 0,970 | 0,960 | 0,950 | 0,940 | 0,930 |
| 50 | 1,015 | 1,000 | 0,985 | 0,970 | 0,955 | 0,940 | 0,925 | 0,910 | 0,895 |
| 60 | 1,025 | 1,000 | 0,975 | 0,955 | 0,935 | 0,915 | 0,885 | 0,865 | 0,845 |

All this points to a clear dependence in the nature of the angular distribution of ascendant longwave radiation on the stratification of the atmosphere. Moreover, when $a$ = const there is detected a dependence of $\frac{I_a}{I_0}$ on $\Delta t$, the difference of the soil and air temperature: as the latter increases, $\frac{I_a}{I_0}$ decreases.

On the basis of the aggregate of the 1967 observation results, the average dependence of $\frac{I_a}{I_0}$ on Δt was obtained for different angles $a$ (Table 3). From this table it follows that, when $a$ = const, on the average there is observed a linear dependence of $\frac{I_a}{I_0}$ on Δt:

$$\frac{I_a}{I_0} = 1 - b\Delta t.$$

The coefficient b is an increasing function of the angle $a$:

| $a°$ | $b$ |
|---|---|
| 0 | 0,00000 |
| 10 | 0,00011 |
| 20 | 0,00025 |
| 30 | 0,00050 |
| 40 | 0,00100 |
| 50 | 0,00150 |
| 60 | 0,00225 |

and may be written in the form:

$$b = 6{,}25 \cdot 10^{-7} a^2,$$

where $a$ is given in degrees.

Now

$$\frac{I_a}{I_0} \approx 1 - 6{,}25 \cdot 10^{-7} a^2 \Delta t$$

represents the average dependence of the ratio $\frac{I_a}{I_0}$ on the angle and soil-air temperature difference.

Individual values differ substantially from the averages cited in Table 3. The reason for this deviation may be, first of all, the fact that the difference $\Delta t$ does not accurately describe the stratification of the atmosphere; in the second place, the inadequate accuracy in the determination of the temperature of the underlying surface; and, in the third place, the varying water-vapor content in the atmospheric layer.

The results of angular distribution estimates made at the GMTs[1] according to temperature sounding data in the Dolgoprudnaya region, and graciously made available to us by V. G. Boldyrev, also attest to a linear relationship between $\frac{I_\alpha}{I_0}$ and $\Delta t$.

The intensity $I(\alpha, H)$ was computed by the formula

$$I_{\Delta\nu}(\alpha, H) \approx \frac{1}{\pi}\left[B_{\Delta\nu}(T_0)\,\tau(w_H, \alpha) + \sum_i B_{\Delta\tau}(T_i)\,\Delta_i\,\tau(w_H, \alpha)\right] \quad (1)$$

on the supposition of an absolutely black underlying surface and a spherically symmetrical atmosphere.

In formula (1) $B(T_0)$ and $B(T_i)$ are the Planck functions for an underlying surface temperature of $T_0$ and an average temperature of the i-th layer of $T_i$; $\tau$ is the transmission factor.

For all absorbent substances and spectral intervals (with the exception of the transparency window) the transmission factor was adopted following Elsasser [3], and for the transparency window - according to Moller [4]. In the case of overlapping bands:

$$\tau = \tau_{H_2O} \cdot \tau_{O_3}; \quad \tau = \tau_{H_2O}\,\tau_{CO_2}.$$

The effective content of absorbent matter was determined by the formula

[1]Translator's Note - This abbreviation, which could not be found, may refer to some kind of meteorological center.

$$w_n = \int_0^H \rho(\xi) f(\xi) d\xi,$$

where $\rho(\xi)$ is the density of the absorbent matter on the $\xi$ level; $f(\xi) = \left(\frac{P_\xi}{P_0}\right)^n$ - is the correction for pressure; $p(\xi)$ is the pressure on the $\xi$ level; $p_0$ - 1000 mb. For water n = 0.6; for carbon dioxide, n = 0.8; for ozone n = 0.4.

The specific humidity was extrapolated according to the method of M. S. Malkevich, Yu. B. Samson, and L. I. Kaprova [5], with the $CO_2$ content taken to be equal to 0.03% by volume; Ramanathan and Culcarni [6] were consulted for the $O_3$ distribution.

The spectral interval $\Delta\nu$ was taken as equal to 20 $cm^{-1}$. Integration over the spectrum was conducted from 0 to 2000 $cm^{-1}$.

In the computation of $\sum_i$ the atmosphere was broken down into layers 0.5 km in thickness. The results of these calculations are given in Fig. 2a, where the $\frac{I_a}{I_0}$ ratio is presented as a function of the difference $\Delta t$. It should be noted that at all the altitudes considered (1.5; 3.0; 5.5; 9.0; 12.0; 20.0 km), with $a$ and $\Delta t$ constant, close values are noted for $\frac{I_a}{I_0}$.

With changing $a$, just as in the analysis of the measurement results, we have a set of straight lines intersecting at a point ($\Delta t = 0$, $\frac{I_a}{I_0} = 1$) whose angular coordinates increase along with $a$.

In their work [7] Wark, Yamamoto, and Leinesh cite data on the angular distribution of outgoing radiation obtained on the basis of temperature soundings in a cloudless sky at 59 locations situated in different zones throughout the world. In this case also, if as an indirect characteristic of the stratification of the atmosphere one uses the temperature difference of the underlying surface and the upper boundary of the troposphere, on the average one obtains, if $\alpha$ is constant, a linear dependence of the $\frac{I_\alpha}{I_0}$ ratio on $\Delta t$ (Fig. 2b). In either situation (Fig. 2a and b) the angular coordinates of the lines are close, and the relative angular distribution can be approximately written in the following form:

$$\frac{I_\alpha}{I_0} = 1 - 0{,}52 \cdot 10^{-7} \alpha^{2{,}28} \Delta t.$$

From a comparison of the $\frac{I_\alpha}{I_0}$ values taken from [7] with the value of the water vapor $w_\infty$ contained in the atmosphere (Fig. 3) it follows that a rise in $w_\infty$ is accompanied by a decrease in the ratio $\frac{I_\alpha}{I_0}$. This is particularly evident with large values of $\alpha$ and small values of $w_\infty$. However, it must be borne in mind that the dependence shown in Fig. 3 does not describe the "pure" effect of $w_\infty$, since a definite relation is observed between the values of $w_\infty$ and $\Delta t$: as $\Delta t$ increases there is an increase in $w_\infty$, and the increase of both values acts on $\frac{I_\alpha}{I_0}$ in one direction. The true effect of a change in the water vapor content on the value of $\frac{I_\alpha}{I_0}$ can be evaluated only for fixed values of $\Delta t$.

Table 4 lists the results of calculations of the ratio $\frac{I_\alpha}{I_0}$ for identical values of $\Delta t$ and different content $w_\infty$. The calculations were performed using formula (1) with the following additional assumptions:

1) the absorbing substance is water vapor, its density distribution with altitude being described by the formula

$$\rho_z = \rho_0 e^{-\beta z}, \text{ a } \rho_0 = \frac{217 \cdot 10^{-6}}{T_0} q_0,$$

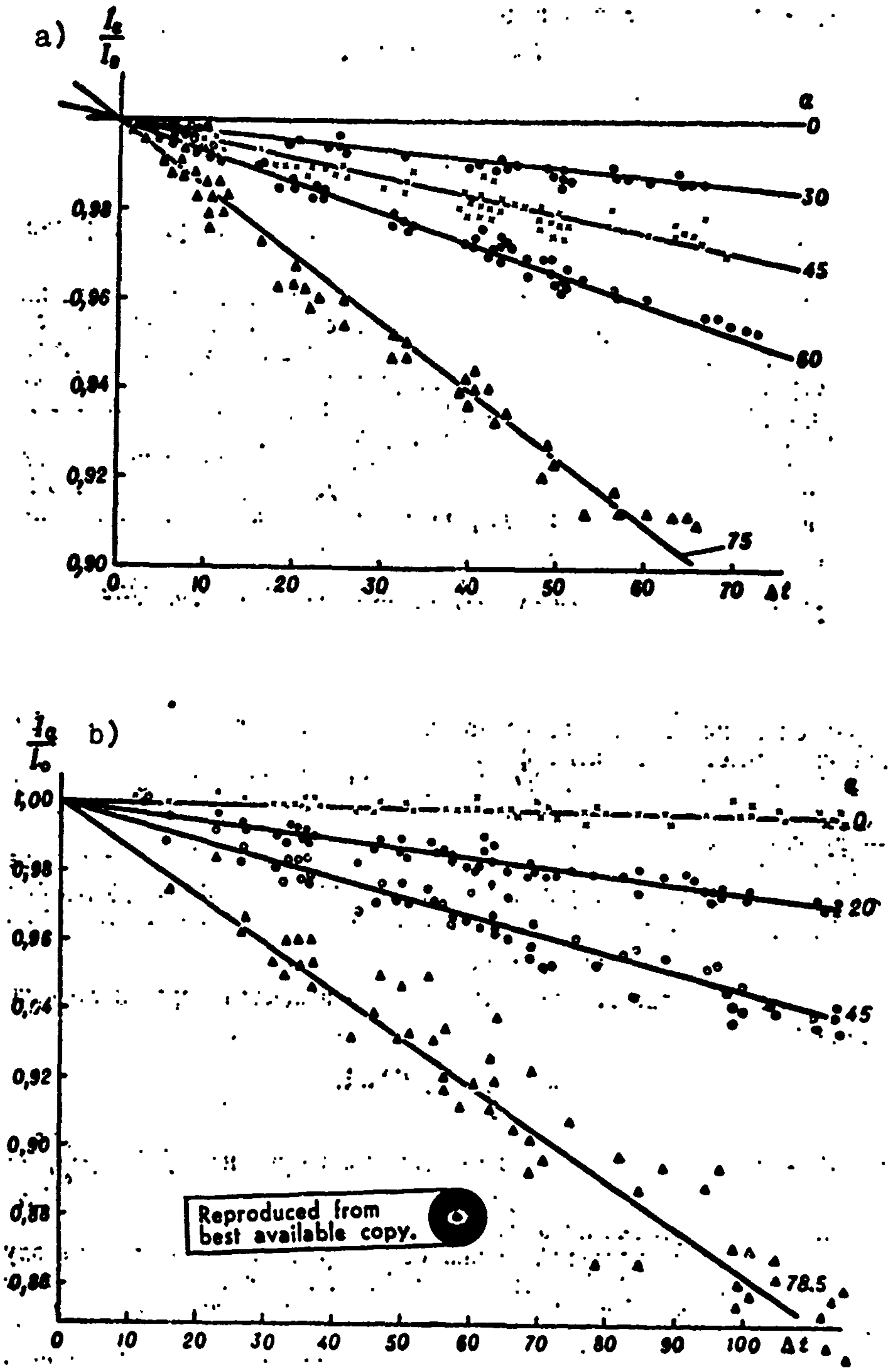




Fig. 2. Dependence of the ratio $\frac{I_a}{I_0}$ on the soil-air temperature difference: a) according to data of V. G. Boldyrev; b) according to data of Wark, Yamamoto, and Leinesh (for outgoing radiation).

where $q_0$ is the elasticity of the vapor in millibars at the underlying surface;

2) the temperature declines linearly with altitude:

$$T_z = T_0 - \gamma z, \quad \gamma = 6 \cdot 10^{-5} \text{ °/см}, \quad \beta = 4{,}5 \cdot 10^{-6} \text{ 1/см}.$$

The transmission factor was written in the form of an exponential function. K. Ya. Kondrat'yev's approach [8] was used to take into account the spectral dependence of the absorption coefficients.

Table 4. Relative angular distribution of ascending long-wave radiation for a fixed value of Δt and varying water vapor content.

| Case | $w_H$ | ratio $\frac{I_\alpha}{I_0}$ for α° | | | |
| --- | --- | --- | --- | --- | --- |
| | | 0 | 30 | 60 | 80 |
| | | $H = 1$ км, $\Delta t = 6°$ | | | |
| 1 | 0,010 | 1,000 | 0,999 | 0,995 | 0,982 |
| 2 | 0,145 | 1,000 | 0,998 | 0,995 | 0,990 |
| 3 | 0,685 | 1,000 | 0,999 | 0,995 | 0,990 |
| 4 | 1,081 | 1,000 | 0,999 | 0,995 | 0,990 |
| 5 | 1,677 | 1,000 | 0,999 | 0,995 | 0,990 |
| | | $H = 4$ км, $\Delta t = 24°$ | | | |
| 1 | 0,022 | 1,000 | 0,996 | 0,975 | 0,942 |
| 2 | 0,314 | 1,000 | 0,996 | 0,980 | 0,947 |
| 3 | 1,483 | 1,000 | 0,996 | 0,980 | 0,950 |
| 4 | 2,338 | 1,000 | 0,996 | 0,980 | 0,951 |
| 5 | 3,621 | 1,000 | 0,996 | 0,980 | 0,952 |
| | | $H = 8$ км, $\Delta t = 48°$ | | | |
| 1 | 0,026 | 1,000 | 0,990 | 0,955 | 0,884 |
| 2 | 0,353 | 1,000 | 0,991 | 0,955 | 0,886 |
| 3 | 1,671 | 1,000 | 0,990 | 0,955 | 0,889 |
| 4 | 2,634 | 1,000 | 0,991 | 0,956 | 0,890 |
| 5 | 4,108 | 1,000 | 0,995 | 0,958 | 0,891 |

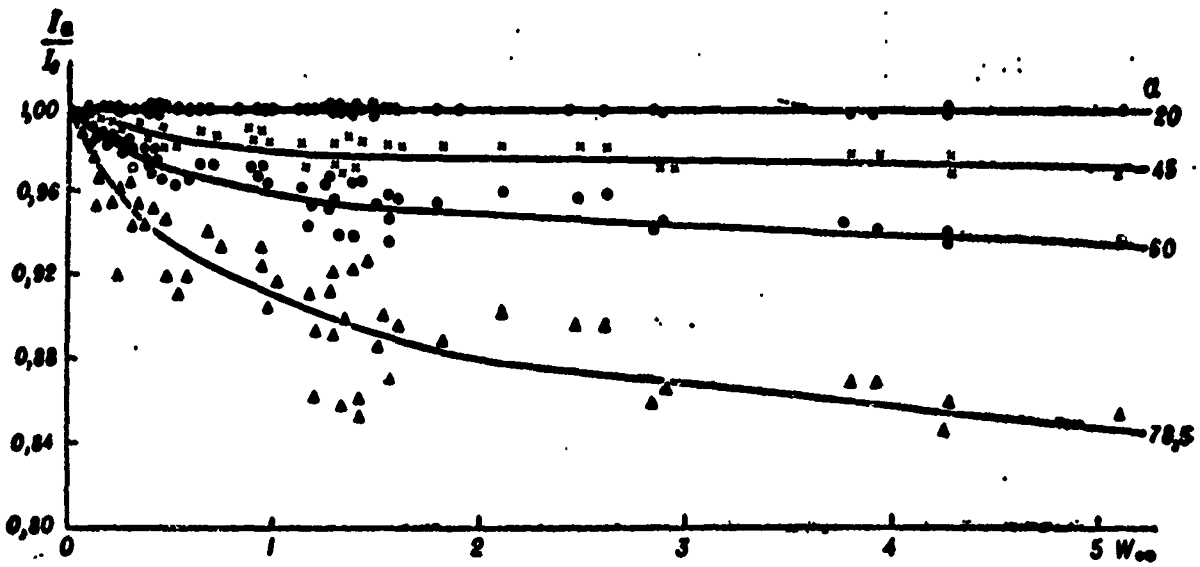


Fig. 3. Dependence of the ratio $\frac{I_a}{I_0}$ for outgoing radiation on the amount of water vapor in the atmosphere.

The initial values of $T_0$ and $q_0$ used in the computations are given below:

| № | $T_0$ °K | $q_0$ мб | $w_\infty$ g/cm$^2$ |
|---|---|---|---|
| 1 | 232,1 | 0,14 | 0,025 |
| 2 | 388,9 | 2,3 | 0,356 |
| 3 | 296,3 | 12,0 | 1,688 |
| 4 | 300,0 | 19,0 | 2,662 |
| 5 | 303,0 | 30,0 | 4,120 |

It follows from Table 4 that a change in the water vapor content has no appreciable effect on the value of the $\frac{I_a}{I_0}$ ratio. The maximum difference between the values of $\frac{I_a}{I_0}$ when $\Delta t$ = const and $a$ = const does not exceed 0.02 $\frac{I_a}{I_0}$. The changes of $\frac{I_a}{I_0}$ as a function of Δt are of the same order of magnitude as those cited in Fig. 2.

Thus, the angular distribution of ascending longwave radiation at different levels of the troposphere will be basically determined by the vertical temperature profile.

Despite differences in computational methods, the findings derived from the data given by different authors are in close agreement; however, as shown in Table 5, they all differ widely from measurement results. The cause of this discrepancy may be the inadequate accuracy of either the computations or the measurements, or both.

Table 5. Comparison of the dependences of $\frac{I_a}{I_0}$ on Δt as obtained by different authors ($a$ = $60^0$).

| Author | Δt° | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | —10 | 0 | 10 | 20 | 30 | 40 | 50 | 60 | 70 |
| 1 | | 1,000 | 0,995 | 0,990 | 0,984 | 0,978 | 0,972 | 0,966 | 0,960 |
| 2 | | 1,000 | 0,992 | 0,985 | 0,978 | 0,972 | 0,965 | 0,958 | 0,952 |
| 3 | max | 1,000 | 0,991 | 0,981 | 0,974 | 0,965 | 0,956 | 0,948 | 0,940 |
| | min | 1,000 | 0,990 | 0,979 | 0,968 | 0,958 | 0,949 | 0,938 | 0,920 |
| 4 | 1,025 | 1,000 | 0,975 | 0,955 | 0,935 | 0,915 | 0,885 | 0,865 | 0,845 |

NOTE: 1 - computations of Wark, Leinesh, and Yamamoto; 2 - computations of Boldyrev; 3 - computations of the authors obtained by integration of the data of Table 4; 4 - measurements.

The assumptions used in the computations may not obtain in the real situation. Thus, for example, by assuming the underlying surface to be absolutely black, we consider its radiation to be isotropic. However, a smooth watery surface displays an acute angular dependence of the reflection factor $r_\lambda$ and, thus, of the emissivity as well:

$$\varepsilon_\lambda = 1 - r_\lambda.$$

Taking into account the angular structure of the emissivity leads to an intensification of the angular dependence of ascending long-wave radiation, as manifest to a greater degree when the water vapor content is low.

Table 6 cites $\frac{I_a}{I_0}$ values obtained for an absolutely black surface (a) and with allowance for the angular distribution of the emissivity (b) according to Novosel'tsev's data [9], but with no allowance for its spectral behavior. These estimations were based on a constant value of $r_\lambda$ corresponding to a minimum when $\lambda = 11$ min.

Table 6.

| | $w_H$ | $\Delta t°$ | $a°$ 30 | 60 | 80 |
|---|---|---|---|---|---|
| а | 0,026 | 48 | 0,990 | 0,955 | 0,884 |
| б | 0,026 | 48 | 0,986 | 0,933 | 0,868 |
| а | 4,106 | 48 | 0,995 | 0,958 | 0,891 |
| б | 4,106 | 48 | 0,992 | 0,950 | 0,880 |

Allowance for the spectral variation of $r_\lambda$ leads to an even greater deviation of $\frac{I_a}{I_0}$ values from those derived for an absolutely black surface. Thus, by bringing the computation conditions into closer accord with the real conditions of observation, the discrepancy between the measured and computed values of $\frac{I_a}{I_0}$ is reduced.

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