Reports of the Academy of Sciences of the USSR

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1962

SovietRxiv

Abstract

Full Text

Reports of the Academy of Sciences of the USSR

1962. Volume 144, No. 2

GEOPHYSICS

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ON THE THEORY OF TERRESTRIAL ATMOSPHERICS

(Presented by Academician I. M. Vinogradov on 9 XII 1961)

The main discharge of lightning to the ground, the so-called return stroke ([1,2]), radiates a radio-wave pulse localized in time, with a frequency spectrum $f < 50$ kHz. It can reach a receiving point on the Earth' s surface by several paths: 1) through the waveguide formed by the Earth' s surface and the lower ionosphere (a terrestrial atmospheric); 2) along a line of force of the Earth' s magnetic field (a whistling atmospheric ([3])); and 3) through the Earth. Terrestrial atmospherics have not yet contributed to the enrichment of our knowledge of the lower ionosphere. On the contrary, attempts to explain their frequency spectra by means of the waveguide theory of long radio waves introduced great scatter into the numerical values of the conductivities σ and heights h of these layers ([4–6]). The reason for this state of affairs was clarified in works ([7–11]), from which it is also easy to understand the known spectral characteristics of atmospherics ([12]). However, the immediate results of observations are usually time sweeps of the vertical component of the electric field of an atmospheric on an oscilloscope tube.

According to many years of observations ([13]), the most frequently occurring type of the main lightning discharge striking the ground may be represented in the form of an elementary Hertz dipole with a vertical current moment $P(t) = I(t)H$ (m · a), where

$$I(t) = D\left(e^{-At} - e^{-Bt}\right), \quad t \geq 0,\ A = 10^3—10^4,$$
$$B = 10^4—10^5,\ I_{\max} = 10—100 \text{ ka}, \tag{1}$$

H is the effective height, hereafter assumed constant. Taking $I(t) = 0$ for $t \leq 0$ and placing the dipole on the polar axis of a spherical coordinate system at the point $r = a$, we obtain that the time form of any atmospheric $P(t)$ for the field

Figure 1

Figure 1: Figure 1

component E_r at the receiving point $\theta, r = a$, according to ($^{7-12}$), will be

$$E_r(t', \theta) = \sum_{j(k)} E_{j(k)}(t', \theta),$$

where $E_{j(k)}(t', \theta)$ are the components of the pulses of the waves TH_j and TE_k, equal to

$$\frac{1}{10c\sqrt{\pi \sin\theta}} \int_{-\infty}^{\infty} \omega P(\omega) n_{j(k)}^2 e^{-\beta_{j(k)}\theta} e^{-i[\omega t' - \Delta\alpha_{j(k)}\theta]}\, d\omega \left(\frac{\mu\mathrm{V}}{\mathrm{m}}\right), \tag{2}$$

and the spectral density of the current moment (1) is

$$P(\omega) = \frac{iDH}{\sqrt{2\pi}} \left(\frac{1}{\omega + iA} - \frac{1}{\omega + iB}\right), \tag{3}$$

where the time t' is counted from the moment of arrival of the signal with velocity $c = 3\cdot 10^8$ m/sec. Distant terrestrial atmospherics are determined by the first numbers of the waves TH_j and TE_k of the branches $\nu_{j(k)}(a, c)$ of equation (10) from (^{9}), whistling atmospherics by the branches $\nu_{j(k)}(c, \infty)$, and those passing through the Earth by the branches $\nu_{j(k)}(0, a)$ and by a continuous spectrum associated with the singularity at the point $r = 0$. For small θ and t, when the atmospheric has not yet reached the ionosphere, the sum (2) is close to the expression $2 \cdot 10^{-13}(a\theta)^{-1}(dP/dt)$, and it has the form of a pulse of duration 50–150 μsec, composed of two half-waves. To understand the mechanism of its transformation into an extended oscillatory train of waves of a terrestrial atmospheric at large θ, let us note that all harmonic components of (2), except $j = 0$, experience in the spherical Earth–ionosphere waveguide two effects: band-pass filtering in frequency and dispersion of phase velocities.

(see Fig. 3 from (10)). As a result of filtering, in pulse (2), at large θ, a rather narrow frequency band $\Delta\omega_{j(k)}$ is singled out, with center $\omega_{j(k)}^0$ near the minimum $\beta_{j(k)}$. The spectrum $P(\omega)$ for the indicated A and B is almost constant in $\Delta\omega_{j(k)}$,

Fig. 1. Forms (Σ) of the head part of the daytime atmospheric and their components in normal waves TH for distances $\theta = 0.245$ and $\theta = 1$, calculated for $A = 5 \cdot 10^3$, $B = 10^5$, and $DH = 5\sqrt{2\pi C}/1000$, $C = 3 \cdot 10^8$. Along the ordinate axis is the field, along the abscissa axis is time.

therefore the pulses (2) prove to be little sensitive to the form of the lightning discharge and depend mainly on the dispersion of the phase velocities. According to the stationary-phase principle, for given t' and θ in (2), the filling frequency ω dominates, for which on the curves $\Delta\alpha_{j(k)}(\omega)$ of Fig. 3 from (10) the condition $(d\Delta\alpha_{j(k)}/d\omega) = (t'/\theta)$ is satisfied. As a result, (2) with increasing θ

spreads into a sinusoidal train of waves, in which the filling frequency ω decreases with increasing t', tending to $\omega^{\mathrm{cr}}_{j(k)}$. In each (2), except for $j = 0$, the central part of the pulse is filled with oscillations with frequencies close to $\omega^0_{j(k)}$. Its delay and attenuation increase with the number $j(k)$, and, for large θ, in the total pulse $E_r(t', \theta)$ the pulse of the TH_1 waves predominates.

For horizontal lightning discharges at large distances the pulse of the TE_1 wave will predominate. We used these guiding ideas in numerical calculations of (2) on the BESM computer, carried out by V. T. Geras' kin. From Figs. 1 and 2, where part of the calculations for a summer day according to the data of Fig. 3 from ([10]) is presented, it is seen that the calculations confirm these considerations and also agree with experimental data ([14–17]). The sum of the pulses TH_j, $j = 1, 2, 3, ...$ (the TE_k pulses are small) forms the leading part of the atmospheric ($f > 5$ kHz), while the "slow tail" ($f < 0.5$ kHz) is formed by the TH_0 wave pulse.

Fig. 2. Forms of the "slow tails" of the daytime atmospheric for the same data as in Fig. 1. Along the ordinate axis is the field; along the abscissa axis is time.

The deep dip in the spectrum ($0.5 < f < 5$ kHz) is explained by Fig. 3 from ([10]). Because of the absence of a critical frequency for TH_0, its pulse differs from all the others.

Fig. 3. Correction $\Delta v = \left(\frac{v}{c} - 1\right)$ of the mean phase velocity $\bar{v}$ to the speed of light c for the nighttime atmospheric, and corrections $\Delta v(TH_j)$ of the phase velocities v_j of the normal waves TH_j, $j = 1, 2, [3,$ of a spherical waveguide, calculated from the data of Fig. 3 in ([10]) (right-hand graph) by the formula $\Delta v_j = \left(\frac{v_j}{c} - 1\right) \cong 7.5\,\Delta\alpha_j/f$. The dashed curves are the ratios of the amplitudes of the normal waves TH_j at a distance of 1560 km.

At any θ it always consists of two half-waves if $I(t)$ has the form (1), i.e., is similar to the initial pulse dP/dt. With increasing θ its half-waves spread according to the law θ^2, obtained in ([19]). Owing to the poor filtration of TH_0 in the waveguide (a, c), the shape of the "tail"

strongly depends on the character of the discharge. The multiwaviness of the "tails"is apparently caused by the superposition of several normal "tails"produced by repeated principal discharges [(18)]. Allowance for the finite conductivity of the soil (see Fig. 15 in [(7)]) gives an additional attenuation $\Delta\beta = 0.3$–0.5 neper for $\Delta\theta = 1$, without noticeably changing the form of $E_{j(k)}$. Calculations of $E_r(t', \theta)$ for the plane model of the medium [(20)] from the data of Fig. 3 in [(10)] (dashed line) differ strongly from the curves of Fig. 1. Moreover, because of the absence of filtration in the plane waveguide at high frequencies (see the dashed line in Fig. 3 of [(10)] for $\beta_{j(k)}$), the forms of $E_{j(k)}$ prove to be sensitive to the numerical values of A and B, which contradicts the experimental fact of the stability of the form of atmospherics. Nighttime atmospherics at large θ are similar to daytime ones. At small θ and large t', the sum of the pulses $j, k = 1, 2, 3, ...$ of a nighttime atmospheric forms a series of slowly decaying "sinusoids" with frequencies $\omega^{\mathrm{gr}}_{j(k)}$. Their spectrum is almost equidistant, and its summation gives a series of peaks similar to dP/dt, following one another with interval $\tau_p \to 600\,\mu$sec as $p \to \infty$. This constitutes the effect of fragmentation of the nighttime pulse, explained in the language of ray theory by multiple reflection of the initial pulse between the earth and the ionosphere ($h = 90$ km).

All the results set forth above are consequences of the theory of long radio waves, based on experimental data only from radiotelegraph stations with $f > 12$ kc. The data known to us on the spectra of atmospherics have a large scatter (see, for example, the wavy lines in Fig. 3 of [10]), caused by errors in determining the location of the lightning, by ignorance of the structure of the discharge (inclination of the dipole, multipolarity), and by the imperfection of the methods for measuring atmospherics. Including them in the group of reliable data by the SID method [10,11] would lead to a considerable loss of information on layers C, D, and E, and would make the picture of the propagation of terrestrial atmospherics given here more blurred. In order to use terrestrial atmospherics for determining the parameters of the lower layers of the ionosphere, one should compile equations (2) from [10] for a large number of ground stations and spectrum points ω, and also abandon the use of the theory of the plane waveguide [20]. The deceptive simplicity and inadequacy of the latter are illustrated by Fig. 3, where the dependence of the mean phase velocity $\overline{v}$ on frequency f for nighttime atmospherics [6] is presented, derived under the assumption that for all f one wave TH_1 dominates with phase velocity $v_{j=1} > c$. In reality, for $f > 10$ kc and $\theta < 0.5$, the principal contribution to E_r is made by several adjacent waves that change with increasing f. In this case their phase velocities v_j become less than c.

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Received
7 XII 1961

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