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Formation of a Local Atmospheric Electric Field

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Abstract—We have estimated the variations in the atmospheric electrostatic field (AEF, $E_{Z(0)}$) strength in the surface layer caused by variations in conductivity due to radon influences, cosmic ray intensity, changes in the balance of light and heavy ions during sunset and sunrise, and under the effect of the ionospheric electric current potential on the AEF potential. It is shown that the air conductivity varies due to ionization under the effect of radon emanations and is determined by the radon exhalation and turbulent diffusion of the surface air layer, while the cosmic ray intensity affects the surface air conductivity through changes in the ion recombination conditions. A decrease in the air conductivity due to a decrease in the cosmic ray intensity (Forbush decrease) also decreases $E_{Z(0)}$, while a decrease in radon fluxes results in an increase in $E_{Z(0)}$. We have estimated the effect of illumination conditions on the AEF due to variations in the relative concentration of heavy and light ions under the influence of photodetachment and photoattachment processes. The work has been done on the basis of data received from the Paratunka observatory (Kamchatka).

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1. INTRODUCTION

The atmospheric electrostatic field is a sensitive indicator of many geophysical processes. Observations of its variations are already used for monitoring environmental pollution, forecasting earthquakes, and many other practical purposes. The effects of certain factors are also studied, e.g., the effects of radon fluxes from the lithosphere into the atmosphere; earthquakes, etc. (Buzevich et al., 1998; Rulenko et al., 1996; Firstov, 1999); cosmic ray intensity variations (Märcz, 1997; Shumilov et al., 2005; Cherneva and Kuznetsov, 2005; Anisimov and Shikhova, 2005); and the influence of near-Earth space conditions on the non-equipotentionality of the electronosphere (Park, 1976).

In this work, we consider an empirical model of local atmospheric electric field behavior depending on different natural factors built on the basis of long-term data on the atmospheric electrostatic field (AEF, $E_{Z(0)}$) strength in the surface layer at the Paratunka observatory (Kamchatka). Another model based on physical equations is used for the interpretation of the discovered regularities.

If the atmosphere is considered as a horizontal homogeneous medium with an exponentially decreasing intensity of the neutral component, and the ionized components are considered as a small admixture, one may assume that the neutral atmosphere rapidly (on the scale of the homogeneous atmosphere altitude) transforms into a sufficiently well-conductive medium at a certain altitude z_e.

Let z_e be the boundary of the electronosphere with the potential U, and the Earth potential be equal to zero. Then, a current with the density j originates between the electronosphere and the Earth at a constant latitude:

$$j_Z = \sigma(z)E_Z(z) = -\sigma(z)\frac{dU(z)}{dz} = \sigma_0 E_{Z(0)}, \quad (1)$$

where $E_{Z(0)}$ is the strength of the electric field vertical component; $\sigma(z)$ is the air conductivity, $\sigma_0 = \sigma(0)$; and $U(z)$ is the electric field potential.

According to Eq. (1), we obtain

$$E_{Z(0)} = -\frac{U}{\int \frac{\sigma_0}{\sigma} dz}, \quad (2)$$

where the integration is carried out from the ground to the electronosphere z_e level. Thus, the electric field strength depends on integral conductivity and electronosphere potential.

The AEF strength is distributed rather inhomogeneously at different altitudes. If the total Earth–ionosphere potential difference is about 300 kV, E_Z

† Deceased.

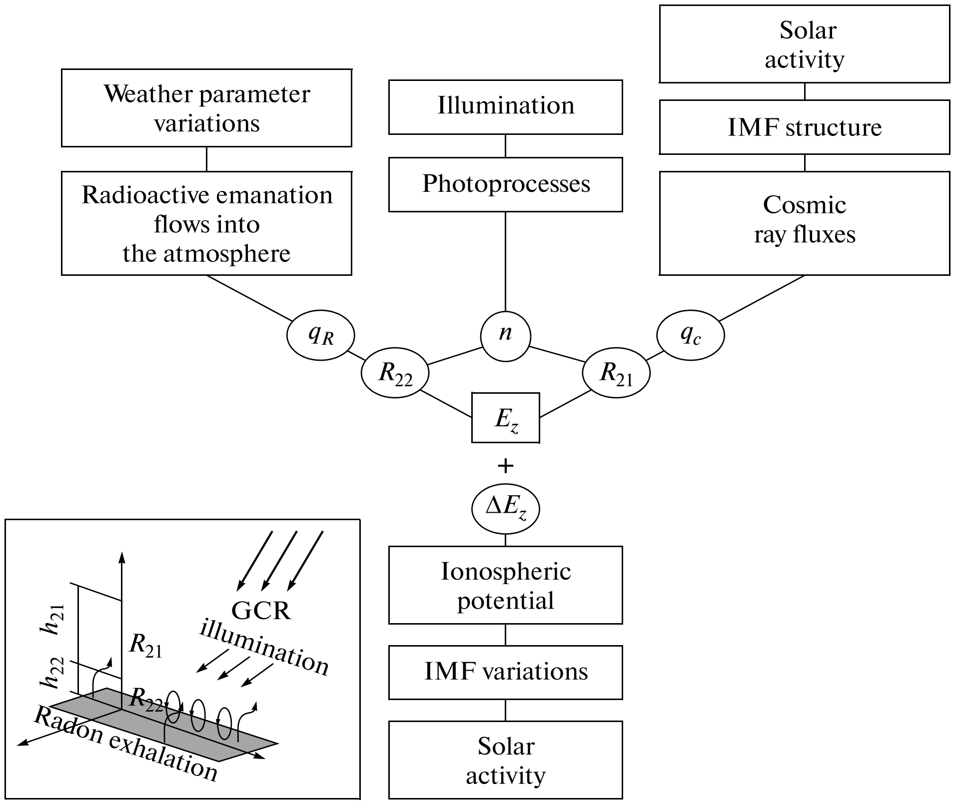

Fig. 1. Scheme of atmospheric electric field formation processes in the presence of factors influencing its value in the surface air layer. The regions of radon and cosmic ray ionization of the atmosphere are marked by R_{22} and R_{21}, respectively.

decreases by about 270 kV in the “lower” 20-km layer and by only 30 kV over the remaining 80 km; therefore, the resistance of the “lower” layer significantly determines the vertical current in the whole Earth–ionosphere column (Kasemir, 1977).

Let us designate the resistance of the upper column, which only depends on external factors, and the resistance of the lower part as R_1 and $R_2 = R_{21} + R_{22}$, respectively, where R_{21} is the resistance of the layer h_{21}, the value of which is determined by the level of cosmic ray ionization, and R_{22} is the resistance of the layer with variable thickness h_{22}, where radon exhalation ionization is added to cosmic rays (Fig. 1). The thickness of the lower “radon” layer depends on the intensity of turbulent air mixing, and the voltage drop here could be written as

$$U_{22} = \frac{UR_{22}}{(R_1 + R_{21} + R_{22})}. \quad (3)$$

Equation (3) is only used for qualitative estimates in the work, while Eq. (2) is used for calculations.

In the work, we have carried out a qualitative complex analysis of the effects of the most significant natural factors on $E_{Z(0)}$ of the AEF, such as radon fluxes into the atmosphere; cosmic ray flux variations; changes in the balance of light and heavy ions during sunset and sunrise; and the effect of the ionospheric electric current potential on the AEF potential.

Interactions between the above-mentioned factors in respect of their effect on the AEF are shown in Fig. 1.

It is necessary to define the notions *regional* and *local* AEFs. It is accepted that the altitude of the “equalizing” layer (isopotential surface, along which the potential is equalized for a rather short time) is about 60 km. Hence, the AEF inhomogeneity on the Earth’s surface caused by the inhomogeneity of potential distribution over the “equalizing” layer should also be of a similar size, which should be considered as the regional scale. There can be several local scales, but their sizes do not exceed the regional scale. One of the scales is the altitude of the upper boundary of the radon mixing layer, which varies from hundreds of meters at calm winter nights to 6–12 and even 12 km on hot and windy summer days. It should be taken into account that winter radon fluxes into the atmosphere can be blocked by the penetration of frost into the surface soil and snow cover. A scale, which is universal for a given region, is supposedly determined by the features of its geological and tectonic structure (Firstov et al., 2006). The conductivity of the surface air layer is firstly determined by the light ion concentration, which not only depends on radon ionization, but a series of factors, due to which local AEFs are formed, masking the unitary (Lobodin, 1980).

The influence of aerosols on the state of the AEF in the surface layer is a different question; it can manifest itself due to three types of aerosols. Firstly, aerosol particles can be collectors of light ions and electrons, to which the former can attach (mainly at night) and then detach (mainly during the day), which can significantly influence the surface layer conductivity (Krasnopevtsev, 1970). Aerosols of the second type are naturally radioactive; this was often observed during active nuclear tests. Power station plumes are sources of such aerosols at present; they can emit radioactive elements into the atmosphere under certain conditions (filter problems and emergency emissions). Finally, aerosols of the third type are particles with different sizes and opposite charges; local electric fields originate during gravitational charge separation in such clouds.

The direct impact of an additional Earth–ionosphere potential difference caused by the nonequpotentiality of the ionosphere and ionospheric currents on electrometer readings stands apart.

2. BRIEF CHARACTERISTICS OF THE OBSERVATION DATA

Routine observations of the electric field in the surface air layer have been carried out at the Paratunka observatory (Kamchatka) since 1996. The volumetric activity of radon (VA Rn) in the subsoil air has been monitored since 1997 (Firstov, 1999; Firstov and Rudakov, 2003; Firstov et al., 2006). The atmospheric electric field is measured with the use of an electrostatic "Pole-2m" fluxmeter along with an air ion composition analyzer. The observatory is also equipped with a weather station and magnetic instruments.

VA Rn in the subsoil air was measured at several levels: in the aeration zone at a depth of 1 m and near the full saturation zone at a depth of 2.5 m. The measurements were carried out using a "REVAR" radiometer with a resolution of 2 cycles per h. In order to distinguish the effect of the ionospheric current potential, weather parameters and magnetic field monitoring data from the Paratunka observatory were used. In order to estimate the effect of Forbush decreases on AEF variations, we used data on cosmic ray intensity received with the use of a neutron monitor in Magadan (Cherneva and Kuznetsov, 2005).

3. FACTORS AFFECTING RADON DISTRIBUTION IN THE SURFACE LAYER AND THE ALTITUDE PROFILE OF THE ATMOSPHERE IONIZATION RATE BY COSMIC RAYS

The main factor determining the local atmospheric electric field is the resistance of the surface air layer, which depends on the ionizing effect of cosmic rays and radon. Diurnal and seasonal variations in radon fluxes into the atmosphere are determined by variations in the atmospheric pressure and different permeability levels of soils depending on seasonal air temperature variations.

The volume power density of an ionization source q_{0R} related to the radon exhalation conditions is defined by the following equation (Baranov, 1955):

$$q_{0R} = 2\xi J\tau/h_D, \tag{4}$$

where ξ is the radon ionization efficiency; $\xi = 3 \times 10^5$ ion pairs per one decay (Physical magnitudes, 1991); q_{0R} is the ionization source power density near the ground; J is radon exhalation or the density of radon fluxes from surface rocks, Bq cm^{-2} s^{-1} (Bq is a unit of radioactivity equal to one decay per s); τ is the radon decay time constant equal to 3.3×10^5 s; and h_D is the reference altitude of the mixing layer. The reference values of J and h_D, at which the mean q_{0R} is equal to 1.5 ion pairs per cm^{-3} s^{-1}, were set equal to 3.7×10^{-7} Bq cm^{-2} s^{-1} and 500 m, respectively.

Weather factors also play a key role in the formation of annual AEF variations. Higher winter values of field strength are connected to the decrease in radon fluxes in the winter months due to the penetration of frost into the soil and thick snow cover in Kamchatka and the significant decrease in the solar illumination intensity (the Paratunka observatory latitude is 53°N), which results in an increase in the heavy ion fraction due to photoattachment. Heavy ions are low mobile, and the surface air layer conductivity also drops in the winter, resulting in an increase in $E_{Z(0)}$ of the AEF.

The dynamic parameters of the atmosphere are valuable in both winter and especially summer, because they determine the degree of radon mixing due to turbulent diffusion, and, hence, its ground-level concentration. In the summer, the surface air layer heats intensively due to solar radiation and gets positive buoyancy. The altitude temperature gradient γ and its relationship with the dry-adiabatic gradient $\gamma_a = g/c_p$, where g is the gravity acceleration and c_p is the heat capacity at constant pressure, are decisive in testing the atmospheric stability. The dry adiabatic gradient is a constant equal to 10° per 1 km (Matveev, 1984). It is evident that if $\gamma > \gamma_a$, the atmosphere is absolutely unstable; if $\gamma_a = \gamma$, the atmosphere is stable; and if $\gamma_a > \gamma$, the atmosphere is absolutely stable.

The first situation can possible be observed only in the summer, the third one is only possible in the winter, while the second situation is possible in any season. The convective instability (Rayleigh–Taylor instability) develops in the case of temperature distribution, which corresponds to the unstable state, resulting in thermal turbulization of the atmosphere.

The turbulent thermal conductivity, which originates in this case, results in a decrease in the temperature gradient, and the atmosphere becomes conventionally stable. In this case, the ionizing factor distribution is described by the equation

$$q_R = q_{0R} e^{-\frac{z}{h}}, \tag{5}$$

where h is the altitude of the mixing layer, q_{0R} is the radon contribution level to ground-level atmospheric ionization.

The altitude dependence of cosmic ray ion formation q_C can be described by the approximate equation of the figure in (German and Goldberg, 1981):

$$q_C = q_{C0} \frac{2\exp(2Z)}{2 + \exp 3.2(Z - 1.4)}, \tag{6}$$

where $Z = z/H$, H is the altitude of the homogeneous atmosphere.

The altitude profile of the cosmic ray ion formation intensity calculated by Eq. (6) is shown in Fig. 2a, which is maximum at an altitude of 13 km. Ionization at the maximum reaches 30 $cm^{-3}\,s^{-1}$, and the altitude range of the maximum ionization region is 16–10, i.e., about 7 km, which indicates the significant effect of cosmic rays on the integral conductivity. The high stability of cosmic ray intensity, except for the Forbush decrease, causes relatively weak variations in the AEF. The current in the Earth–ionosphere column decreases during Forbush effects due to significant increase in the resistance R_1 caused by a decrease in atmospheric ionization (Fig. 1), which results in a drop in the voltage U_{22}, according to Eq. (3), and a decrease in E_Z near the Earth surface. At the same time, the resistance R_{22} is independent of the cosmic ray intensity, i.e., it is virtually invariable.

According to the data on long-term observations at the Hungarian Nagycenk observatory (Märcz, 1997), and in mountainous regions (Krechetov and Filippov, 2000), the correlation between the cosmic ray intensity and $E_{Z(0)}$ is shown. Figure 2b shows $E_{Z(0)}$ as a function of cosmic ray intensity at the Paratunka observatory during 18 events of Forbush decreases according to the data of the Magadan neutron monitor. The dependence is well described by the linear equation ($E_{Z(0)}$, %) = 9.64 (GCR, %) – 0.72, from which it is evident that a decrease in GCR by 3–10% results in a significant decrease in $E_{Z(0)}$ of the AEF by 20–80% (Cherneva and Kuznetsov, 2005).

4. RESPONSE OF THE ATMOSPHERIC ELECTRIC FIELD TO IONIZATION PROCESSES

The air conductivity is determined by the concentration of light ions, their charge, and mobility. The

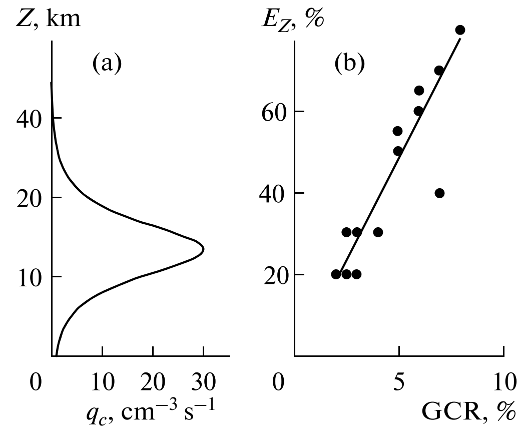

Fig. 2. Average altitude profile of cosmic ray ionization of the atmosphere (a); the dependence of $E_{Z(0)}$ decrease on the decrease in cosmic ray intensity during Forbush decreases (b).

sea-level mobility of an "average" light ion in air is equal to μ $cm^2/(Vs)$ (Smirnov, 1992). Heavy ions, i.e. dust particles or molecular clusters with attached molecular ions or electrons, are low mobile and virtually do not participate in the current formation.

Let us consider a stationary case. Using the balance equation for the number of light and heavy ions in air (Matveev, 1984) with accounting for the coefficients of photoattachment $c = 5.2 \times 10^{-23} cm^3\,s^{-1}$ (Danilov, 167; Smirnov, 1978) and photodetachment $\beta \sim 2 \times 10^{-4}\,s^{-1}$ of light ions from clusters and dust particles, i.e. from heavy ions (Alpert, 1972; Laboratory researches..., 1970), we obtain for the stationary state (for small β)

$$\begin{gathered} N = C/(bn + \beta) \approx C/bn - C\beta/(bn)^2; \\ n = \left[(C^2 + qa)^{1/2} - C\right]/a + C\beta/\left[b\left(C^2 + qa\right)^{1/2}\right], \end{gathered} \tag{7}$$

where n is the density of light and N is that of heavy ions, $a = 6.5 \times 10^{-6}\,cm^3\,s^{-1}$ is the coefficient of recombination of light ions with different signs, $b = 1.6 \times 10^{-6}\,cm^3\,s^{-1}$ is the coefficient of recombination of light and heavy ions, $C = cN_L = 1.4 \times 10^{-3} s^{-1}$ is the coefficient of attachment reaction rate, N_L is the Loschmidt constant, and q is the ionization rate $q = q_R + q_C$.

The conductivity is expressed via mobility by the equation $\sigma(z) = en(z)\mu_0 \exp Z$. Further use of the conductivity equation is convenient in the form $\sigma(Z) = \sigma_0 \frac{n(Z)}{n_0} \exp Z$, where σ_0 is the ground-level conductivity, $Z = z/H$, H is the altitude of the homogeneous atmosphere.

Considering light ions as the only current carriers (Ponomarev and Sedykh, 2006), let us consider two

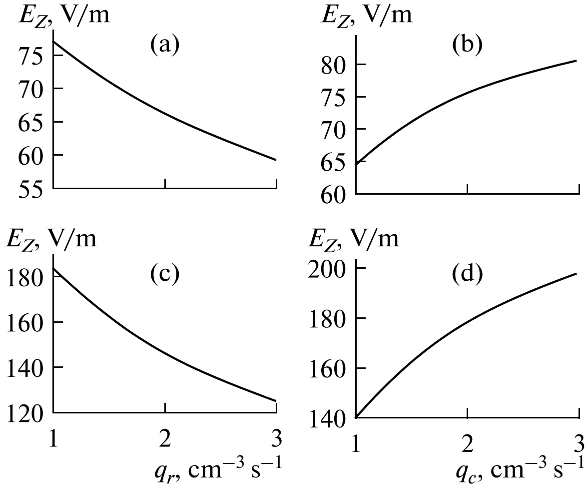

Fig. 3. Ionizer intensity effect on $E_{Z(0)}$ as a function of cosmic ray q_c and radon q_r ionization rates when one of the parameters is fixed at 1.5 pairs of ions per cm^{-3} s^{-1}: (a) the influence of radon ionization intensity variations at a constant cosmic ray ionization rate in the case of the quadratic recombination law; (b) the influence of cosmic ray ionization intensity variations at a constant radon ionization rate in the case of the quadratic recombination law; (c) the influence of radon ionization intensity variations at a constant cosmic ray ionization rate in the case of the linear recombination law; and (d) the influence of cosmic ray ionization intensity variations at a constant radon ionization rate in the case of the linear recombination law. The altitude of the mixing layer is 720 m in all cases.

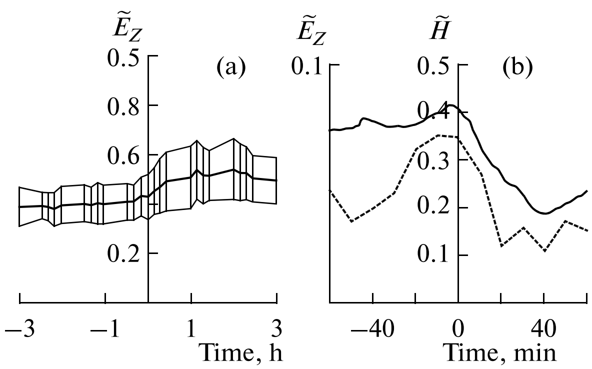

Fig. 4. Separation of the sunrise effect on the AEF by the method of epoch superposition with rms errors (the thin curves). In the calculations, 203 events during days with good weather have been used; the large errors are connected to the use of data for different seasons (a). The separation of AEF variations (b) connected to geomagnetic disturbances (the solid curve corresponds to the H component of the Earth's magnetic field and the dashed one is E_z), the beginning of the bay in geomagnetic variations in the time range near midnight is accepted as the epoch start; the values are normalized in respect of the maximum.

cases, i.e., the implementation of the quadratic ($C \ll qa$) and linear ($C \gg qa$) recombination laws. Below we provide the calculations for the quadratic recombination and the results for both cases.

In order to estimate the part of photoprocesses in n variations, let us separate the term containing C and β as a separate summand:

$$n = (q/a)^{1/2} + \frac{\beta C}{b(qa)^{1/2}}. \tag{8}$$

Only the first term of Eq. (8) is used for all calculations, except for the sunrise effect.

Taking into account the above-mentioned about a and using Eqs. (5) and (6), we find

$$\sigma = \sigma_0 (q_R + q_C)^{1/2} e^{1.5Z}, \tag{9}$$

where $\sigma_0 = e\mu_0/a_0^{1/2}$, e is the electron charge, μ_0 is the light ion mobility, and a_0 is the coefficient of recombination of light ions on the Earth's surface.

Substituting Eq. (9) in Eq. (2) and taking into account Eqs. (5) and (6), we find

$$E_{Z(0)} = -U/H \int \left\{ \varepsilon_R \exp(3 - \alpha Z) + 2\varepsilon_C \exp \frac{5Z}{2 + \exp 3.2(Z - 1.4)} \right\}^{-1/2} dZ, \tag{10}$$

where $\varepsilon_R = q_{0R}/(q_{0R} + q_{0C})$, $\varepsilon_C = q_{0C}/(q_{0R} + q_{0C})$, $\alpha = H/h$, h is the altitude of the mixing layer; and q_{0R} and q_{0C} are the intensities of ground-level ion formation depending on the radon and cosmic ray effects, respectively. According to the calculations, the influence of the mixing layer altitude is not very significant; we have accepted $h = 720$ m. $E_{Z(0)}$ as a function of q_R at the fixed value $q_C = 1.5$ of ion pairs per cm^{-3} s^{-1} is shown in Fig. 3a, and $E_{Z(0)}$ as a function of q_C at the same value of q_R is shown in Fig. 3b.

Above, the dependence between the air cosmic ray ionization and ion recombination processes was shown by the quadratic ion balance equation, but in (Ermakov and Stozhkov, 2004; Stozhkov, 2003; Bazilevskaya et al., 1991) it is shown that the dependence can be represented by the linear law of recombination. It follows that $n \sim (q)^{0.5}$ in the case of the quadratic equation and $n \sim (q)$ in the case of the linear equation. The latter dependence indicates the presence of a stronger real correlation between the atmospheric ion concentration and cosmic ray fluxes as compared to earlier assumptions (Ermakov and Stozhkov, 2004).

The theoretical dependences of $E_{Z(0)}$ for the quadratic and linear recombination laws are shown in Fig. 3. It is of interest that $E_{Z(0)}$ decreases with a growth in q_R and increases with a growth in q_C, which is confirmed by experimental data. Indeed, we observe an almost synchronous decrease in E_Z of the AEF during the Forbush decrease. The experimental data confirming the calculations are given in (Cherneva and Kuznetsov, 2005).

We can assess the sunrise–sunset effects if we rewrite q_{0R} as $q_{0R}^{*} = q_{0R} - C(q_{0R}/a)$ for the hours after sunrise and as $q_{0R}^{**} = q_{0R} - C(q_{0R}/a) + \beta C/b(q_{0R}a)^{1/2}$ for the hours after sunset and then substitute these equations in Eq. (10) instead of q_{0R}. The theoretical assessment of $E_{Z(0)}$ variations is 5% in this case, which is confirmed by experimental data.

Figure 4a shows the averaged variation in the AEF strength during sunrise at the Paratunka observatory obtained by the method of epoch superposition over 37 days of good weather in the spring and summer months in 2004–2005. A six-hour interval is considered, on which the sunrise time is accepted as the epoch start: a smooth 10% increase in $\bar{E}_{Z(0)}$ is observed during two hours after sunrise. The presented data confirm the sunrise effect on AEF $E_{Z(0)}$ variations.

5. EFFECT OF THE IONOSPHERIC POTENTIAL DIFFERENCE ON THE ATMOSPHERIC ELECTRIC FIELD

The Earth's ionosphere is influenced by the potential electric field formed in the magnetosphere due to complex transformation processes of solar wind kinetic energy into electromagnetic energy. The effect of ionospheric potential on the AEF was theoretically considered in (Park, 1976); it was shown that it could be significant near the auroral zone and equal to 10 V/m for $E_{Z(0)}$. Indeed, the response of $E_{Z(0)}$ with a delay of 1–2 days to solar bursts was discovered in AEF measurements at the Zugspitze mountain (Reiter, 1969), which was confirmed by later AEF measurements at the Vostok station and in Antarctica (Frank–Kamenetsky et al., 1999). The ionosphere is a well-conductive medium; therefore, the ionospheric potential gradients form an intense electric current flowing in a quite narrow altitude range (from 100 to 120 km with the maximum near 107 km), which is called the dynamo layer.

Figure 4b exemplifies the epoch superposition ionospheric AEF variation obtained for 39 events at the Paratunka observatory. The beginning of the bay is taken as epoch zero. Events close to the local midnight time have been selected; this is about a 5% value resulting from the statistical errors in the method at the mean value of the electric field of about 120–140 V/m.

6. CONCLUSIONS

The electric field in the Earth–ionosphere capacitor is formed by thunderstorms from all over the world and, hence, conventionally has a unitary variation; it is influenced by local regional factors, which are to be taken into account when identifying an external effect on the basis of observation data, e.g., aerosol pollution or seismic effects. In this work, we have suggested an approximate scheme for accounting for these effects and made rough assessments. However, there evidently exists a prospect for designing a filter allowing for the separation of the above-mentioned factors from AEF $E_{Z(0)}$ variations. Due to the uncertainty associated with the coefficients in the ionization balance equations and other equations, the suggested model cannot provide the reproduction accuracy of surface layer AEF variations required in practice today. At present, empirical models are more convenient for practical purposes; however, the considered model allows for a qualitative analysis of interactions between all the factors included.

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