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Electrodynamic model of the lower atmosphere and the ionosphere coupling

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Abstract

Electrodynamic model of the ionosphere response to seismic related lower atmosphere disturbances is developed. The chain of processes forming the lithosphere–ionosphere coupling starts from the injection of radioactive substances and charged aerosols into the atmosphere. This leads to a change of the vertical electric current in the atmosphere and to modification of the electrical field in the ionosphere. Growth of the electric field and related development of acoustic-gravity wave instability is followed by formation of field-aligned currents and plasma density disturbances stretched along the geomagnetic field. Another effect of the electric field increase is an additional Joule heating of the lower ionosphere which results in an elevation of the F-layer maximum, a decrease of electron density in the maximum of this layer and a growth of light ion density in the upper ionosphere. Thus, the presented model connects disturbances of the key ionosphere parameters with increase of atmospheric radioactivity and injection of charged aerosols into the atmosphere. © 2001 Published by Elsevier Science Ltd.

Keywords: Aerosols; Electric current; Lithosphere–ionosphere coupling; Electric field; Earthquakes

1. Introduction

In a number of satellite experiments various phenomena in the ionosphere and the magnetosphere preceding earthquakes have been revealed (see for example: Gokhberg et al., 1995; Liperovsky et al., 1992; Molchanov, 1993; Buchachenko et al., 1996). Among those are VLF/ELF/ULF emissions (Gokhberg et al., 1982; Chmyrev et al. 1989; Bilichenko et al., 1990; Serebryakova et al., 1992; Molchanov et al., 1993; Parrot, 1994), DC electric field disturbances (Chmyrev et al., 1989), plasma density irregularities (Chmyrev et al., 1997; Afonin et al., 1999), large-scale variations of altitude profile of electron density in the ionosphere (Pulinets et al., 1994), modification of ion composition in the upper ionosphere (Boshkova et al., 1990) and some others. Many of the reported effects require the further experimental confirmation and statistical studies. There is practically no complicated experimental data showing interconnection of the phenomena that have been observed in numerous satellite and ground-based measurements. In this situation it is principally important to create the comprehensive model adequately describing the causal-sequence connections between the separate phenomena participating in seismo–ionospheric interaction.

This paper presents an attempt of developing such model taking into account the observed electromagnetic and plasma effects of earthquakes on the ionosphere. The general idea of the paper is following. Enhanced injection of radioactive substances and charged aerosols into the atmosphere before earthquake together with dynamic transport processes modifies the height distribution of electric conductivity and induces the additional electromotive force (EMF) in some layer of the lower atmosphere. Due to these two effects the main parameters of the global closed electric circuit in the Earth–ionosphere system are modified. First of all an enhancement of the electric current in the Earth–ionosphere

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layer as in element of this circuit occurs. As a result we obtain a growth of the electric field in the lower ionosphere directly proportional to increase of the current intensity. The further scenario is connected with acoustic-gravity wave instability developed in the lower ionosphere when the growing electric field exceeds some threshold value. The instability leads to spatial modulation of plasma density and electric conductivity in the ionospheric E-layer and generation of the related field-aligned currents and plasma density irregularities in the upper ionosphere. The satellite observations of anomalous DC electric field, ULF magnetic pulsations, small-scale plasma inhomogeneities and correlated ELF emissions, as well as the formation of whistler ducts over the seismic zone (Chmyrev et al., 1989, 1997, 1999; Bilichenko et al., 1990; Sorokin et al., 1998; Sorokin et al., 2000; Borisov et al., 2001) could be considered as experimental evidence for these processes. An additional Joule heating of the lower ionosphere in anomalous DC electric field influences on the thermal balance of the ionosphere and results in modification of the height distribution of electron density and the ion composition in F-region and in the upper ionosphere.

Thus the key point of the model is an enhancement of radioactivity and charged aerosols in the atmosphere before earthquake. An importance of these processes for generation of some seismic effects on the ground and in the ionosphere was noticed by Pierce (1976), Pulinets et al. (1994), Molchanov and Hayakawa (1996), Boyarchuk et al. (1998), Sorokin and Chmyrev (1999a, b), Sorokin and Yaschenko (1998, 1999, 2000). An experimental basis for these works is given by King (1986), Virk and Singh (1994), Heincke et al. (1995), who have reported the observations of increasing radon concentration in soil gases and in water of natural sources few days to one week before earthquake.

2. Vertical distribution of the electric field in a conducting atmosphere

Let us consider a disturbance of the electric field beyond the zone of thunderstorm activity resulted from an enhancement of atmospheric radioactivity as well as due to formation of a slow-varying altitude-dependent EMF. The quasi-stationary electrodynamic processes in the Earth–ionosphere layer are described by the equations

$$\nabla \cdot \mathbf{E} = 4\pi(\rho + \rho_s), \quad \mathbf{E} = -\nabla\varphi,$$

$$\frac{\partial(\rho + \rho_s)}{\partial t} + \nabla \cdot (\mathbf{j} + \mathbf{j}_s) = 0, \quad \mathbf{j} = \sigma\mathbf{E}, \tag{1}$$

where $\mathbf{E}$ and φ are the electric field and its potential; $\mathbf{j}$ and ρ are the current and charge densities; σ is the atmospheric conductivity; ρ_s and $\mathbf{j}_s$ are the external charge and current densities arising under the action of EMF. From Eq. (1) we obtain the equation for the electric field in the Earth–ionosphere layer:

$$(\partial/\partial t + 4\pi\sigma)\nabla \cdot \mathbf{E} + 4\pi(\nabla\sigma) \cdot \mathbf{E} = -4\pi\nabla \cdot \mathbf{j}_s. \tag{2}$$

Let us introduce the Cartesian co-ordinates with the axis z directed vertically upwards and with the origin located on the Earth's surface. The lower boundary of the ionosphere coincides with the plane $z = h$. We assume that the atmospheric radioactivity and aerosol concentration vary simultaneously on the horizontal scale exceeding the height of the lower ionosphere. This allows one to use a one-dimensional approximation with all quantities depending on z, the altitude over the Earth's surface. In this approximation only the vertical components of $\mathbf{E}$ and $\mathbf{j}$ are nonzero. The duration of atmospheric radioactivity enhancement is about few days (Virk and Singh, 1994; Heincke et al., 1995). Therefore we can assume that the vertical distributions of the conductivity and the electric field are stationary. In the quasi-stationary approximation, when the EMF varies with the characteristic time exceeding $1/4\pi\sigma_0$, we obtain from Eq. (2):

$$E(z,t) = -E_0(z,t) - \frac{1}{\sigma(z)}\left\{ j_s(z,t) - \frac{\int_0^h j_s(z,t)\frac{dz}{\sigma(z)}}{\int_0^h \frac{dz}{\sigma(z)}} \right\},$$

$$E_0(z,t) = \frac{U}{\sigma(z)\int_0^h \frac{dz}{\sigma(z)}}. \tag{3}$$

This solution was obtained at the condition that EMF appearance in the atmosphere does not change the potential difference between the ionosphere and the Earth surface. Really, the total resistance of the Earth–ionosphere layer amounts to about 200 Ω, and the resistance of the atmospheric column with the cross dimension of the order of hundreds of km, covering the domain of EMF formation, is well above (about 10 MΩ). From equalities (3) we obtain the expression determining the vertical distribution of the electric field in the Earth–ionosphere layer, while varying the atmospheric conductivity in absence of EMF sources:

$$E(z) = \frac{E_0(0)\sigma_0(0)}{\sigma(z)} \frac{\int_0^h \frac{dz}{\sigma_0(z)}}{\int_0^h \frac{dz}{\sigma(z)}}. \tag{4}$$

Index "0" in Eq. (4) denotes undisturbed quantities.

Let us consider a disturbance of the electric field induced by EMF and connected with the formation of external current in the near ground layer of atmosphere. To calculate the disturbed electric field in Eq. (3), it is necessary to employ the altitude dependence of the conductivity and the external current. We represent the spatial-temporal dependence of the external current in the form:

$$j_s(z,t) = j_s(0,t)f(z), \quad f(0) = 1.$$

Let us introduce a dimensionless relative electric field ε and external current density χ:

$$\varepsilon(z,t) = E(z,t)/E_0(z); \quad \chi(t) = j_s(0,t)/j_0.$$

From Eq. (3), we obtain:

$$\varepsilon(z,t) = \frac{f(z) - 1 + [f(z) - k]\varepsilon(0,t)}{1-k};$$

$$k = \int_0^h f(z)\frac{dz}{\sigma(z)} \left[\int_0^h \frac{dz}{\sigma(z)}\right]^{-1}. \tag{5}$$

The dimensionless parameter k characterizes a relationship between the vertical scale of the atmospheric conductivity and the vertical distribution of the external current. With the external current vanishing in the lower ionosphere, one may obtain the simple formulae connecting the dimensionless electric field on the Earth's surface $\Delta\varepsilon(0)$ and in the ionosphere $\Delta\varepsilon(h)$ with the external current $\Delta\chi$ in the Earth–ionosphere layer:

$$\Delta\varepsilon(h,t) = k\Delta\chi(t); \quad \Delta\varepsilon(0,t) = -(1-k)\Delta\chi(t);$$

$$\Delta\varepsilon(h,t) = -[k/(1-k)]\Delta\varepsilon(0,t). \tag{6}$$

These formulae allow one to estimate the value of the field in the lower ionosphere, by its measuring on the Earth's surface, or to restore a magnitude of the casual external current.

3. Ion formation rate under the action of atmospheric radioactivity

The main electric load of the atmospheric current circuit is concentrated in the lower atmosphere. Its conductivity is determined by ionization sources and ion mobility. Various ionization sources, e.g. chemical, electric and radioactive, act in the atmosphere. The main ionizing factor determining the level of conductivity in the near surface layer, is atmospheric radioactivity. The natural radioactivity of the lower atmosphere is mainly connected with such elements as radon, radium, thorium, actinium and their decay products. Radioactive elements enter the atmosphere together with soil gas. They are transferred by air streams upwards up to the altitude of few km. Therewith the ion formation rate amounts to tens of ion pairs in a cubic centimeter per second. An increase in the level of atmospheric radioactivity, e.g. prior to earthquake, leads to an increase in the ion formation rate, and hence the conductivity as well. Assume that the vertical distribution of the radioactive element concentration in the atmosphere is determined by the function $n_r(z)$. The ion formation rate under the action of the ionizing radiation of atmospheric radioactivity $q_a(z)$ is obtained in Appendix A (see Eq. (A.5)). The equilibrium values of electron and ion concentrations, being produced by the source $q_a(z)$, is determined by their recombination processes in air.

4. Conductivity induced by atmospheric radioactivity

To estimate the stationary ion-molecular composition of the atmosphere, we employ a simplified system of ionization-recombination processes (Barth, 1961). In the atmosphere, near the Earth's surface, except light single-charged ions there also exist heavy ions created as a result of adhesion of light ones to aerosols. The concentrations of light positive and negative ions n_+ and n_- forming the lower atmosphere conductivity are determined mainly by their recombination and adhesion to aerosols (Tverskoy, 1949):

$$n_+ = n_- = n = \frac{\beta N}{\alpha}\left[\sqrt{1 + \frac{\alpha q}{\beta^2 N^2}} - 1\right], \tag{7}$$

where N is the aerosol concentration, α is the light ion recombination coefficient, β is the coefficient of light ion adhesion to aerosols. The vertical distribution of the average concentration of soil aerosols is presented by (Gavrilova and Ivliev, 1996). This is an exponential dependence $N = N_0 \exp(-z/H_a)$, where H_a is the characteristic scale of aerosol concentration variation. Dependence of the effective recombination coefficient α on altitude is represented by the equation $\alpha(z) = [5\times10^{-8} + 2.5\times10^{-6}\exp(-z/H)]\ \mathrm{cm^3/s}$ (Medvedev et al., 1980), where H is the scale of exponentially inhomogeneous atmosphere. The atmospheric conductivity σ is expressed in terms of the light ion concentration by the formula

$$\sigma = e(\mu_+ n_+ + \mu_- n_-) \approx 2e\mu n, \tag{8}$$

where $\mu = \mu_0 \exp(z/H)$ is light ion mobility (μ_0 is the mobility on the ground level) (Tverskoy, 1949). To calculate the conductivity, it is necessary to find equilibrium ion concentrations depending on the ion formation rate.

Let us find the ion formation rate q_a due to atmospheric radioactivity. The vertical distribution of concentration of the radioactive elements in the atmosphere is determined by many factors, such as meteorological conditions, turbulent diffusion, gravity, etc. To estimate the effects of increasing atmospheric radioactivity near the Earth's surface on the conductivity and the electric field in the atmosphere, we choose the following vertical dependence of the atmospheric radioactivity: $n_r(z) = n_0 \exp(-z/H_r)$ (here H_r is the vertical distribution scale to be determined below). Substituting this dependence to expression (A.5), we obtain

$$q_a = q_0 F(e^{-z/H_r})/F(1), \tag{9}$$

where

$$F(y) = y\int_0^1 x^{(H/H_r)-1}\mathbf{E}_1[(H/l_{\gamma0})|x-y|]\,dx.$$

The quantity q_0 is the ion formation rate on the Earth's surface. Fig. 1 presents the dependences $q_a(z)/q_0$ on the altitude for different values of the vertical distribution scale H_r of atmospheric radioactivity. As follows from these plots, the vertical distribution of the ion formation rate differs from the

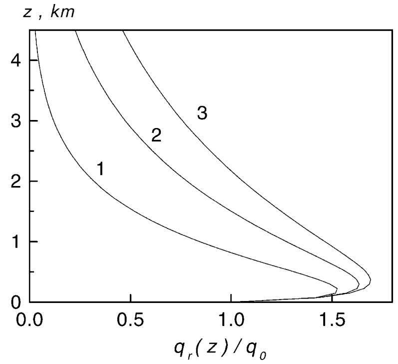

Fig. 1. Altitude dependencies of relative rate of ion production at the different scales of altitude distribution of atmospheric radioactivity: 1—$H_r = 1$ km;2—$H_r = 2$ km; 3—$H_r = 3$ km.

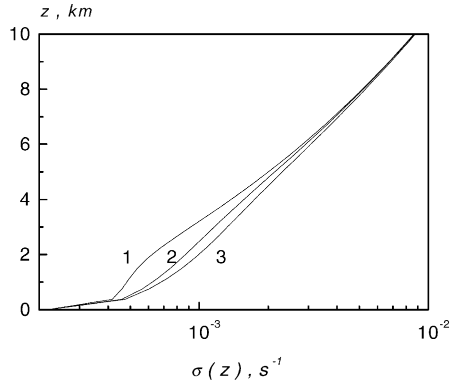

Fig. 2. Altitude dependencies of lower atmosphere conductivity at the different scales of altitude distribution of atmospheric radioactivity: 1—$H_r = 1$ km; 2—$H_r = 2$ km; 3—$H_r = 3$ km.

exponential altitude dependence of atmospheric radioactivity. As H_r increases, a maximum in the vertical distribution of the ion formation appears, its value increases at each altitude. These results seem natural, since the air ionization at a given altitude is determined by the ionizing radiation from the volume elements spaced at a distance of the order of the free path of the gamma radiation, which grows exponentially with the altitude. As $n_r(z)$ decreases slower, a unit volume contains larger amount of radioactive elements, and hence they contribute more significantly to the ionization intensity. Apart from the atmospheric radioactivity, the lower atmosphere is ionized by cosmic rays. The vertical distribution of the ion formation rate q_a resulted from the action of cosmic rays may be approximated by the Chapman function q_s (Ratcliffe, 1960). The total ion formation rate in the lower atmosphere $q(z)$ is a sum $q = q_a + q_s$ of ion formation rates due to the action of atmospheric radioactivity and cosmic rays which are given by formulae (9) and Chapman function with the altitude of ion formation rate maximum z_m and its value at this altitude q_m.

Knowing the vertical distribution function $q_a(z)$, we find a dependence of the induced conductivity by formulae (7) and (8). The calculations of the conductivity were performed for those values of the quantity H_r to which the curves 1–3 in Fig. 1 correspond. In the calculations the following values of the parameters were chosen: the aerosol concentration on the Earth's surface $N_0 = 2000\ \mathrm{cm}^{-3}$, the spatial scale of their vertical distribution $H_a \sim 10^5$ cm (Gavrilova and Ivliev, 1996); the coefficient of light ion adhesion to aerosols $\beta = 4.3\times10^{-6}\ \mathrm{cm}^3/\mathrm{s}$, $q_m \approx 40\ \mathrm{cm}^{-3}/\mathrm{s}$, $z_m \approx 14$ km, $q_0 \approx 10\ \mathrm{cm}^{-3}$ (Tverskoy, 1949); $\mu_0 = 2.3\ \mathrm{cm}^2/\mathrm{Vs}$. Fig. 2 presents results of these calculations. From the plots it follows that in the surface layer a rapid growth of the conductivity is observed up to the altitude $\sim$0.5 km. In the altitude range from 0.5 up to 6–8 km the conductivity increases with the growth of the vertical scale H_r.

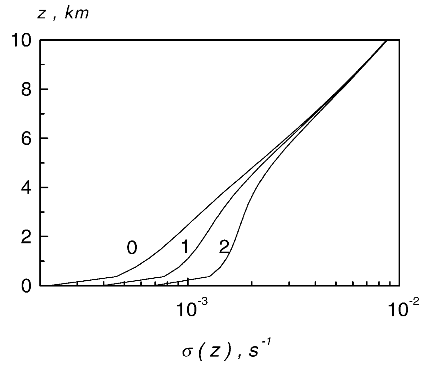

Fig. 3. Altitude dependencies of the atmosphere conductivity at the different rates of ion production near the Earth surface: 0—$q(0)/q_0(0) = 1$; 1—$q(0)/q_0(0) = 2$; 2—$q(0)/q_0(0) = 4$.

Let us estimate a value of the vertical scale H_r of the atmospheric radioactivity distribution. Its value is found from the condition that the potential difference between the Earth and the ionosphere as well as the density of the current flowing between the ionosphere and the Earth are: $U = 3\times10^5\ \mathrm{V} = 10^3$ CGSE and $j = 2\times10^{-12}\ \mathrm{A/m}^2 = 6\times10^{-7}$ CGSE correspondingly. Then the estimates give $H_r = 2$ km, that is of the same order of magnitude as the values presented, e.g., in (Tverskoy, 1949). Fig. 3 presents the calculated vertical distribution of conductivity at $H_r = 2$ km versus an increase in the level of atmospheric radioactivity on the Earth's surface. 2 and 4 times elevation of this level, adopted in the calculations, correspond to the experimental data presented by Virk and Singh (1994) and Heincke et al. (1995). From the plots it follows that the growth of atmospheric radioactivity near the Earth's surface leads to a considerable increase in the atmospheric conductivity in the near Earth layer at the altitudes up to 4 km. Fig. 4 presents the relative

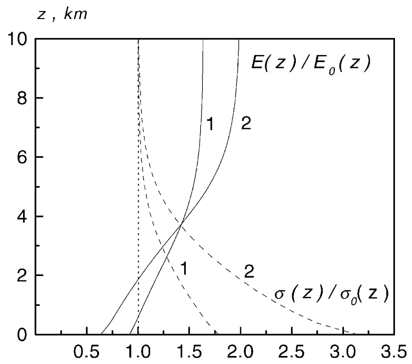

Fig. 4. Altitude dependencies of relative conductivity and the electric field values at the different rates of ion production near the Earth surface: 1—$q(0)/q_0(0)=2$; 2—$q(0)/q_0(0)=4$.

conductivity $\sigma(z)/\sigma_0(z)$ and the relative value of the electric field $E(z)/E_0(z)$, calculated by formula (4) for two values of the relative increase in the atmospheric radioactivity near the Earth's surface and the value $H_r=2$ km. From the plots in Fig. 4 it follows that the electric field near the Earth's surface decreases, while near the lower boundary of the ionosphere it increases few times. This result is in agreement with the satellite data obtained by Chmyrev et al. (1989) wherein a two to three times increase in the electric field was found in the ionosphere over the seismic zone.

5. E.M.F. formation at the aerosol injection

Various mechanisms of forming the external current in the near ground layers of the atmosphere are possible. One of the mechanisms is connected with intensifying the injection of charged soil aerosols into the atmosphere or changing the meteorological conditions under their stable vertical distribution. The quasi-stationary vertical distribution of aerosols may be formed as a result of a turbulent upward transfer and gravitational sedimentation (Fleagle and Businger, 1963). The vertical distribution of soil aerosol concentration is represented in the form $N=N_0\exp(-z/H_a)$ (McCartney, 1976). The aerosol scale height is $H_a=K/w$, where K is the vertical eddy diffusion coefficient and w is the settling speed. Equations, describing concentration N, charge and current densities of aerosol particles injected into the atmosphere (ρ_s, j_s), are obtained in Appendix B (see Eqs. (B.6)). The second equality in (B.6) is 1-dimensional approximation for the equation describing the external charge and current densities in the conducting media. The sense of the equality is as follows. Variations of the external charges in separated volume are determined by two processes. First is their transfer, under the action of EMF, through the surface confining the volume. Second, is diminishing the external charge as result of its relaxation into the environment with conductivity σ. For example, if the total flux of external charges through the surface confining the volume is equal to zero ($\nabla\mathbf{j}_s=0$), then their quantity in this volume decreases as $\rho_s\sim\exp(-4\pi\sigma t)$. The external charges in a conducting medium relax $\sim 1/4\pi\sigma$. For sufficiently rapid processes $t\ll 1/4\pi\sigma$ the continuity equation takes the form $\partial\rho_s/\partial t+\nabla\mathbf{j}_s=0$. Otherwise, in the case of a slow process of external charge formation $t\gg 1/4\pi\sigma$, its density is connected with the external current as $4\pi\sigma\rho_s+\nabla\mathbf{j}_s=0$.

Assuming that the characteristic time scale of the considered processes exceeds the relaxation time $1/4\pi\sigma$ we find from (B.5), (B.6) the equation for altitude distribution of external current in quasi-stationary approximation:

$$\frac{\partial}{\partial z}\left[\frac{1}{4\pi\sigma(z)}\frac{\partial j_s(z,t)}{\partial z}\right]+\frac{w}{4\pi\sigma(z)K}\frac{\partial j_s(z,t)}{\partial z}-\frac{j_s(z,t)}{K}=0. \quad (10)$$

If the vertical scale of the aerosol concentration variation is less than the scale of the conductivity variation, then in Eq. (10) one may assume $\sigma(z)=\sigma_0$. We also assume week dependence of K on altitude. Therewith the solution has the form:

$$j_s(z,t)=j_s(0,t)\exp(-z/H_j), \quad (11)$$

where $H_j=1/\{(w/2K)+[(w/2K)^2+4\pi\sigma_0/K]^{1/2}\}$. Substituting Eq. (11) into second equation in (B.6), we obtain the vertical distribution of the external charge:

$$\rho_s(z,t)=\rho_s(0,t)[\exp(-z/H_j)-H_j\delta(z)], \quad (12)$$

where $\delta(z)$ is Dirac's delta function. The value of the extraneous current and charge on the Earth's surface are given by the equality:

$$j_s(0,t)=4\pi\sigma_0 H_j\rho_s(0,t)=4\pi\sigma_0 H_j q[N_+(0,t)-N_-(0,t)], \quad (13)$$

where N_+, N_- are the number density of charged aerosols. From formulae (11) and (12) it follows that the external currents and charges are distributed mainly at the lower altitudes than aerosols are. This is related to the charged particles being transferred in a conductive medium. If $N_+>N_-$, then the external current is directed upwards. Therewith the field on the Earth's surface is directed downwards and its value increases, while in the ionosphere it decreases. As an example we consider an exponential dependence of the conductivity on altitude $\sigma(z)=\sigma_0\exp(z/H_\sigma)$ (where H_σ is the characteristic scale of conductivity variation). Substituting the vertical distribution of the conductivity and the external current into Eq. (5), we obtain the value of the coefficient $k=H_j/(H_j+H_\sigma)\approx H_j/H_\sigma\ll 1$. Fig. 5 presents the plots of vertical dependence of the relative field for its different values on the Earth's surface calculated by formula (5). It is seen from the plots that the field growth on the Earth's surface corresponds to its decrease or sign change in the ionosphere. The electric field decrease on the Earth's surface or the sign reversal corresponds to a growth of its value in the ionosphere.

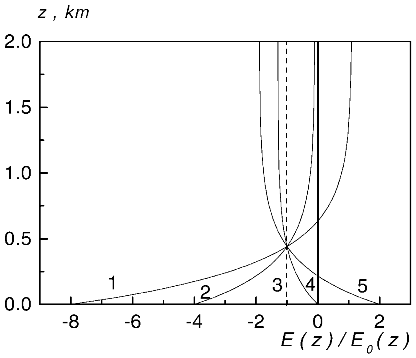

Fig. 5. Altitude dependencies of relative value of the vertical electric field at the different external currents: $k = 0.2$; 1—$j_s(0,t)/j_0 = 9$; 2—$j_s(0,t)/j_0 = 4$; 3—$j_s(0,t)/j_0 = 0$; 4—$j_s(0,t)/j_0 = -1.3$; 5—$j_s(0,t)/j_0 = -4$.

Let us consider a case, when external current changes with height more slowly than conductivity. In this case in last part of formulas (3) we can change the external current altitude dependence to it value in a point of maximum $1/\sigma(z)$ function:

$$\int_0^h j_s(z)\frac{\mathrm{d}z}{\sigma(z)} \approx \bar{j}_s(0)\int_0^h \frac{\mathrm{d}z}{\sigma(z)}.$$

Here $\bar{j}_s(0)$ is average value of the external current near to a surface of the Earth. Expression (3) can be transformed to following one:

$$E(z) = -E_0(z) - \frac{j_s(z) - \bar{j}_s(0)}{\sigma(z)}. \quad (14)$$

On the Earth surface $j_s(0) \approx \bar{j}_s(0)$, therefore $E(0) \approx -E_0(0)$. Expression (14) allows to estimate the electric field value in the ionosphere. At $z = h$, we obtain:

$$E(h) = -E_0(h) - \frac{j_s(h) - \bar{j}_s(0)}{\sigma(h)} \approx -E_0(h) + \frac{\bar{j}_s(0)}{\sigma(h)}$$

$$= -E_0(h) + \frac{\bar{j}_s(0)}{j_0}E_0(h),$$

$$E_0(h) = E_0(0)\frac{\sigma(0)}{\sigma(h)}. \quad (15)$$

Let us estimate the first part of equality (15). Considering the electric field $E_0(0) = 150$ V/m, the atmosphere conductivity $\sigma(0) = 2\times10^{-14}$ S/m near the Earth surface and the low ionosphere conductivity $\sigma(h = 80\ \text{km}) = 10^{-6}$ S/m, we obtain $E_0(h) = 3\times10^{-3}$ mV/m. Second part in Eq. (15) is differed from the first one on the rate of external current value near the Earth surface to the undisturbed atmospheric current $j_0 = 3\times10^{-12}$ A/m^2 = 8.7×10^{-7} CGSE. The external current value is estimated by formula (13): $\bar{j}_s(0) = 4\pi\sigma(0)qH_j(N_+ - N_-)$. To find this value, we assume: $\sigma(0) = 2\times10^{-14}$ S/m = 2×10^{-4} CGSE, $q = e = 4.8\times10^{-10}$ CGSE, $H_j = 2$ km = 2×10^5 cm, $N_+ = 4\times10^3$ cm^{-3}, $N_- = 0$. Then, we obtain: $\bar{j}_s(0) = 8.7\times10^{-4}$ CGSE = 3×10^{-9} A/m^2 = $10^3 j_0$. Hence, we have from (15):

$$E(h) \approx \frac{\bar{j}_s(0)}{j_0}E_0(h) = 3\,\frac{\text{mV}}{\text{m}}.$$

Thus, if the external current is transferred to the altitudes exceeding the vertical scale of atmospheric conductivity near the Earth surface, the electric field in the ionosphere is amplified few orders of magnitude according to the factor $\bar{j}_s(0)/j_0$. The upward transport of external current can be connected with vertical atmosphere convection at heating of the near ground atmosphere. If the vertical gradient of temperature T on absolute value exceeds the critical one (which is about 1/100 deg/m), the atmosphere becomes unstable and the vertical convection arises (Fleagle and Businger, 1963). According to the satellite data (Qiang et al., 1999) the growth of seismic activity is accompanied by heating of the bottom atmosphere. This heating can be connected with “green gas” effect, Joule heating at occurrence of the external currents, etc. Besides the upward transfer of the external current there are few other factor leading to electric field enhancement in the ionosphere. Among those are the increase of the low atmosphere conductivity connected with growth of atmospheric radio-activity and the reduction of low ionosphere conductivity which can be related to upward transfer of hydrogen molecules. The detailed analyses of these processes is a subject of separate paper.

6. Ionospheric effects of DC electric field enhancement

The previous theoretical considerations (Sorokin and Chmyrev, 1999a; Sorokin et al., 1999) required the enhanced DC electric field for explanation of various seismic related phenomena in the ionosphere. Above we have described the mechanism of the field enhancement before earthquake up to the necessary magnitudes. With this result we can try to construct the model of interconnected electromagnetic and plasma disturbances developing in the ionosphere under influence of seismic process.

The diagram illustrating this model is presented in Fig. 6. The chain of processes forming the seismo–ionospheric coupling mechanism starts from the injection of radioactive substances and charged aerosols into the atmosphere (block 1, Fig. 6). This leads to an increase in atmospheric conductivity and forming a vertical electromotive force near the ground level. The growth of lower atmosphere conductivity and appearance of additional electromotive force in the Earth–ionosphere electric circuit is followed by a change in the vertical current and the electric field in the ionosphere (blocks 2 and 3, Fig. 6). Further development is connected with the electric field growth in the ionosphere. It is shown that its enhancement over the definite threshold value in the lower ionosphere results in an instability of the acoustic-gravity waves (AGW) at the Brunt–Vaisala frequency (Sorokin

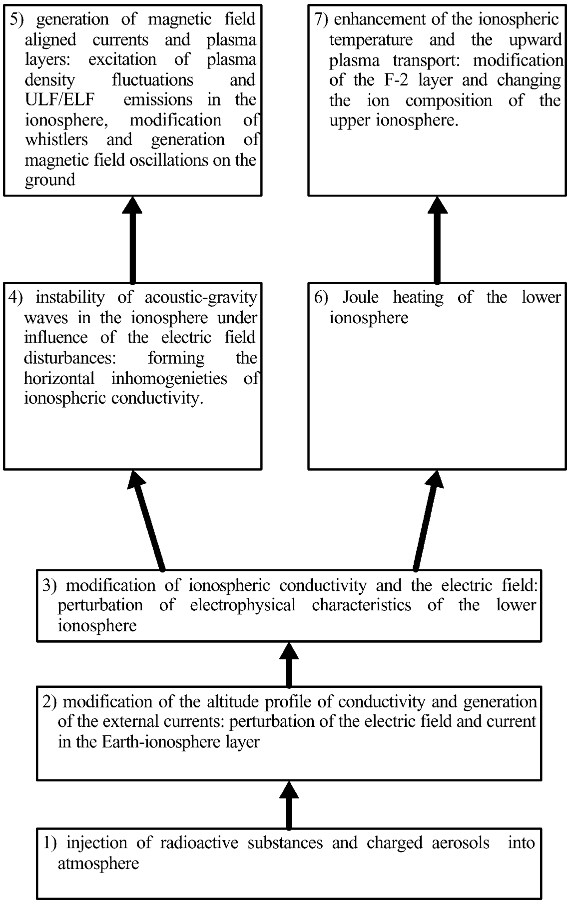

Fig. 6. Diagram of the processes forming the atmosphere–ionosphere coupling.

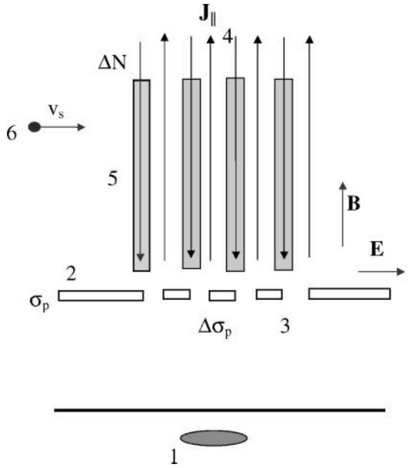

Fig. 7. The scheme of satellite observations of plasma density inhomogeneities and ULF magnetic field oscillations: 1. Earthquake zone; 2. Lower ionosphere; 3. Horizontal inhomogeneities of ionospheric conductivity; 4. Field-aligned electric currents; 5. Field-aligned plasma density inhomogeneities; 6. Satellite.

et al., 1998; Sorokin and Chmyrev, 1999b; Chmyrev et al., 1999). The instability is followed by formation of the horizontal periodic irregularities of ionospheric conductivity with characteristic spatial scale $l = \pi a/\omega_g n(\omega_g)$, where a is the sound velocity, ω_g is the Brunt–Vaisala frequency and $n(\omega_g)$ is the refraction index (block 4, Fig. 6).

Due to high conductivity along the magnetic field lines the polarization electric fields connected with such irregularities are transferred into the upper ionosphere and the magnetosphere. The correspondent field-aligned currents are carried by electrons while the carriers of transverse closure currents are ions. Therefore, the electric field propagation along the magnetic field lines and the generation of closed currents is accompanied by local variations of plasma density. An estimate of the plasma density disturbances is given as $\Delta N/N_0 = v_i c n(\omega_g)E/\omega_i aB$ (where v_i is the collision frequency of ions, ω_i is the gyrofrequency of ions and B is the geomagnetic field magnitude). When a satellite moving with the velocity v_s crosses the periodic plasma irregularities with the spatial scale l and the corresponding structure of field-aligned currents, the fluctuations of plasma density and the transverse disturbances of geomagnetic field with the period $\Delta t = \pi a/v_s\omega_g n(\omega_g)$ are registered (block 5, Fig. 6). The scheme of satellite observations of plasma density inhomogeneities and ULF magnetic field oscillations is presented in Fig. 7. The estimates give $\Delta N/N_0 \approx (1.6$–$16)\%$; $b \approx 5$ nT; $\Delta t \approx (0.3$–$3)$s. These values in the order of magnitude correspond to the satellite data (Chmyrev et al., 1989; Bilichenko et al., 1990; Chmyrev et al., 1997).

Results concerning the excitation of small-scale plasma inhomogeneities can be applied for explanation of modifying the natural whistler characteristics before earthquakes reported by Hayakawa et al. (1993). The physical mechanism of ionosphere–magnetosphere whistler duct formation in terms of above consideration is developed by Sorokin et al. (2000).

Besides that the excitation of horizontal small-scale irregularities of electric conductivity in the lower ionosphere can be used as a basis for generation mechanism of electromagnetic ELF precursors to earthquakes. Such mechanism associated with the emission into the upper ionosphere of whistler mode waves is presented by Borisov et al. (2000). These waves appear due to interaction of thunderstorm related EM radiation with small-scale plasma irregularities excited in the lower ionosphere before earthquakes. EM pulses are generated by lightning discharges and propagate in the subionospheric waveguide with small attenuation in ELF range. Plasma irregularities created above the seismic zone are polarized by EM pulses. The induced nonstationary polarization currents act as sources of whistler mode waves.

Other effects of the electric field enhancement are connected with additional source of the Joule heating in the lower ionosphere, which alters the ionosphere thermal balance determining the ionosphere temperature (block 6, Fig. 6). The theory of F-layer deformation (block 7, Fig. 6) was developed by Sorokin and Chmyrev (1999c). The thermal energy injected in the lower ionosphere is transferred to the top region. Process of thermal interchange in the upper atmosphere determines the altitude distribution of the temperature in this region. If the electric field increases in the ionosphere above the seismic region (Chmyrev et al., 1989) by the quantity $\Delta E = 6$ mV/m $= 2\times10^{-7}$ CGSE, then it results in generation of the thermal flux $q = 2.3$ erg/(cm^2 s) and in the ionosphere temperature growth up to $T \approx 2000$ K. Calculations of the perturbed electron density altitude profiles performed by (Sorokin and Chmyrev, 1999c) show an increase in the maximum altitude and decrease in the maximum value in the profiles. This result is in agreement with experimental data on the top-side ionosphere sounding (Pulinets, 1994).

The calculations (Sorokin and Chmyrev, 1999c) give an ion density enhancement in the upper ionosphere. Such enhancement in the density of atomic oxygen ions [O+] which are dominant in the upper ionosphere should be followed with increase of the light ion densities reported by (Boshkova et al., 1990) as a result of resonance reaction of the charge exchange: $O^+ + H \leftrightarrow H^+ + O$.

7. Conclusion

The vertical distribution of atmospheric electric field connected with excitation of external currents near the Earth's surface and increase of atmospheric radioactivity at the preparatory phases of earthquakes is analyzed. An injection of charged aerosols and their vertical turbulent transfer in atmosphere is considered as a possible mechanism of forming the external currents. Relationship between the vertical component of the electric field in the lower atmosphere and on the Earth's surface is found. The electrodynamic model presented above for the seismo–ionospheric coupling prior to earthquake reduces numerous effects in space plasma to a single cause—the increase of quasi-stationary electric field in the ionosphere. This field is controlled by the lithosphere processes through variations of the lower atmosphere dynamics and the electromotive force effected by the injection of radioactive substances and charged aerosols into the atmosphere.

Thus, the presented model describes the complete causal-sequence chain of seismic induced processes beginning from modification of the lower atmosphere state to excitation of the plasma density variations and ULF/ELF magnetic field oscillations, an elevation of the F-layer maximum and growth of the light ion densities localized in the upper ionosphere. The theoretical results are in a reasonable agreement with the available experimental data. The model can be applied to studies of the ionospheric effects of large-scale natural and technological disasters effecting the electrophysical parameters of the atmosphere in the near-ground layer.

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Appendix A.

Below we obtain the vertical distribution of the ion formation rate as a result of absorption in the atmosphere of the gamma radiation from the decay of radioactive elements being constituents of the atmospheric radioactivity. Since initial angular distribution of gamma radiation is isotropic, then the number of quanta q_γ being generated in a unit volume per unit time and per unit solid angle is equal to $q_\gamma = \kappa n_r/4\pi$, ($\kappa = \ln 2/T$, T is an effective half-life period). The gamma radiation is absorbed due to Compton effect of air molecules on electrons. The distribution function $f(\boldsymbol{r}, t, \theta)$ of quanta propagating with the velocity $\mathbf{v}$ ($|\mathbf{v}| = c$, c is the light velocity) is found from the transfer equation (Leipunskiy et al., 1960):

$$\frac{\partial f}{\partial t} + \mathbf{v}\cdot\nabla f = -\frac{f}{\tau_\gamma} + q_\gamma, \tag{A.1}$$

where $\tau_\gamma = l_\gamma/c$ (l_γ is the mean path length of a quantum), θ is an angle between the velocity vector and the z axis aligned vertically upwards. For the exponentially inhomogeneous atmosphere with the scale H, $l_\gamma = l_{\gamma 0}e^{z/H}$ (where $l_{\gamma 0}$ denotes the mean path length on the Earth's surface). In a one-dimensional and stationary approximation we obtain:

$$f(z,\theta) = \frac{\kappa}{4\pi c\cos\theta} \begin{cases} \displaystyle\int_0^z dz' n_r(z')\exp\left\{\frac{H}{l_{\gamma 0}\cos\theta}\left(e^{-z/H} - e^{-z'/H}\right)\right\}, & 0<\theta<\pi/2, \\ \displaystyle -\int_z^\infty dz' n_r(z')\exp\left\{\frac{H}{l_{\gamma 0}\cos\theta}\left(e^{-z/H} - e^{-z'/H}\right)\right\}, & \pi/2<\theta<\pi. \end{cases} \tag{A.2}$$

The number of quanta in a unit volume $n_\gamma(z)$ is determined by integrating the distribution function (A.2) over the solid angle:

$$n_\gamma(z) = 2\pi\int_0^\pi f(z,\theta)\sin\theta\, d\theta. \tag{A.3}$$

As a result of the Compton effect in the air, the gamma radiation generates a flux of fast electrons whose concentration n_k is found from the formula (Medvedev et al., 1980):

$$n_k = \frac{l_e}{l_\gamma} n_\gamma = \frac{l_{e0}}{l} n_\gamma.$$

Moving in the air, the fast electrons loose energy and, as result of collisions, produce low-energy secondary electrons. The absorbed energy of fast electrons in the air in a unit volume is $n_k \varepsilon$, ε is the energy of fast electron. 33 eV of the absorbed energy is spent on creating an electron–positron pair in the air (Massey et al., 1969). Hence, the number of secondary low-energy electrons in a cubic centimeter produced during the life-time of fast electrons is $n_k \lambda$, where $\lambda = \varepsilon/33$ eV. Thus, we obtain that the mean rate of secondary electron production q_a is determined by the equality:

$$q_a = n_k \frac{\lambda}{\tau_e} = \frac{\lambda n_\gamma}{\tau_\gamma(z)}. \tag{A.4}$$

Substituting (A.2), (A.3) into (A.4) and integrating over the solid angle, we obtain the altitude dependence of the ion formation rate under the action of the ionizing radiation of atmospheric radioactivity with the vertical distribution of its concentration $n_r(z)$:

$$q_a(z) = \frac{\lambda \kappa}{2 l_\gamma(z)} \int_0^\infty \mathrm{d}z' n_r(z') \mathrm{E}_1[k(z,z')], \tag{A.5}$$

where $k(z,z') = (H/l_{\gamma 0})|\mathrm{e}^{-z/H} - \mathrm{e}^{-z'/H}|$, $\mathrm{E}_1(u) = \int_1^\infty (\mathrm{e}^{-ux}/x)\,\mathrm{d}x$ is the integral exponential function (Abramowitz and Stegun, 1980).

Appendix B

Let us obtain an equation describing the external current of charged aerosol particles injected into the atmosphere. The charge of such particle q in the conductive atmosphere decreases in time due to conductivity currents. Integration of the continuity equation $\partial\rho/\partial t = -\operatorname{div}(\sigma \mathbf{E})$ over volume Ω of aerosol particle gives:

$$\frac{\mathrm{d}q}{\mathrm{d}t} = -4\pi\sigma q. \tag{B.1}$$

The vertical motion of aerosol particles includes the gravitational settling with the velocity w and the turbulent transport with the velocity $\xi(t)$:

$$\frac{\mathrm{d}z}{\mathrm{d}t} = -w + \xi(t). \tag{B.2}$$

The particle dynamics can be described by the stochastic differential equations for the probability distribution function (Arnold, 1973). We believe that the velocity of turbulent transport $\xi(t)$ in (B.2) satisfies to the relationships:

$$\langle \xi(t) \rangle = 0 \langle \xi(t)\xi(t') \rangle = 2K\delta(t-t'), \tag{B.3}$$

where angular brackets note the statistical averaging.

Let us introduce the distribution function $f(q,z,t)$ of aerosol particles depending on the electric charge, altitude and time. Under conditions (B.1)–(B.3) it is easily to obtain (Arnold, 1973) the Fokker–Plank equation for the $f(q,z,t)$:

$$\frac{\partial f}{\partial t} - w\frac{\partial f}{\partial z} - 4\pi\sigma(z)\frac{\partial}{\partial q}(qf) = \frac{\partial}{\partial z}\left(K\frac{\partial f}{\partial z}\right). \tag{B.4}$$

Spatial and temporal dependencies of aerosol concentration $N(z,t)$, their electric charge $\rho_s(z,t)$ and current $j_s(z,t)$ densities are expressed as the moments of distribution function $f(q,z,t)$:

$$N(z,t) = \int_{-\infty}^{\infty} f(q,z,t)\,\mathrm{d}q,$$

$$\rho_s(z,t) = \int_{-\infty}^{\infty} q f(q,z,t)\,\mathrm{d}q,$$

$$j_s(z,t) = -w\int_{-\infty}^{\infty} q f(q,z,t)\,\mathrm{d}q - K\int_{-\infty}^{\infty} q\frac{\partial f(q,z,t)}{\partial z}\,\mathrm{d}q = -\rho_s w - K\frac{\partial \rho_s}{\partial z}. \tag{B.5}$$

The equations for moments are obtained from (B.4) and (B.5):

$$\frac{\partial N}{\partial t} - w\frac{\partial N}{\partial z} = \frac{\partial}{\partial z}\left(K\frac{\partial N}{\partial z}\right),$$

$$\frac{\partial \rho_s}{\partial t} + 4\pi\sigma(z)\rho_s = -\frac{\partial j_s}{\partial z}. \tag{B.6}$$

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