On a possible seismomagnetic effect in the topside ionosphere

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Abstract

In this paper we present the results of the computation of the electric and magnetic fields produced in the ionosphere by the near-earth seismogenic disturbance in the vertical atmospheric electrostatic field under different ionospheric conditions. It is shown that in the nighttime ionosphere during solar minimum and inside large-scale plasma bubbles, the magnitude of the transverse electric field can attain ~0.2 and 1.0 mV/m, respectively. The seismomagnetic effect with the magnitude of ~13 pT is predicted in the topside daytime and nighttime ionosphere at any solar activity.

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1. Introduction

There are numerous publications which provide the observational evidence of pre-earthquake ionospheric perturbations (e.g., Pulinets and Boyarchuk, 2004, and references therein: Oyama et al., 2008, 2011; Liu et al., 2009, 2010, 2011; Sharma et al., 2010; Le et al., 2011; Ryu et al., 2014). However, the question if these perturbations are really associated with seismic activities preceding earthquakes remains unresolved (Rishbeth, 2006). It is mainly because of ionospheric variability (over time-scales from hours and days to solar cycles) caused by solar and magnetospheric influences as well as by impact of lower atmosphere (e.g., Prölls, 1995; Rishbeth, 1991; Forbes et al., 2000; Rishbeth and Mendillo, 2001; Mendillo et al., 2002; Zhang and Holt, 2008). Furthermore, physical mechanisms for pre-earthquake seismo-ionospheric coupling are still far from being clearly understood. In the literature, a number of probable drivers responsible for precursory seismo-ionospheric effects have been discussed (e.g., Hayakawa, 1999, 2000; Hayakawa and Molchanov, 2002). One of them is the seismogenic electrostatic field (SEF) that could be seen near the Earth’s surface as a perturbation in the vertical atmospheric electrostatic field $E_z$. Perturbations in $E_z$ have been observed prior to several earthquakes within their preparation zones (Kondo, 1968; Vershinin et al., 1999; Hao et al., 2000; Kamogawa et al., 2004). Before strong earthquakes, the magnitude of $E_z$ perturbation can reach 300–1000 V/m. Hao et al. (2000) have found that the pre-earthquake $E_z$ perturbation’s lateral scale size $R_0$ is related to the imminent earthquake magnitude $M$ as $R_0 \sim \exp(M)$, where $R_0$ is taken in kilometers. Thus, for major earthquakes with $M \sim 8$, a value of $R_0$ can be assumed to be as large as ~3000 km. It is presently unclear what is underlying mechanism for SEF. There has been made an attempt to explain

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generation of SEF by electric currents associated with the stressed rock (Freund, 2000, 2010; Freund et al., 2004, 2009; Freund and Sornette, 2007). Under certain conditions, SEF can penetrate into the ionosphere and modify ionospheric plasma density (e.g., Pulinets and Boyarchuk, 2004; Kuo et al., 2011, 2014; Liu et al., 2011). In contrast, penetration of SEF into the ionosphere is negligibly small according to Denisenko et al. (2008) and Ampferer et al. (2010). Another plausible mechanism for pre-earthquake electric field appearance in the ionosphere was suggested by Oyama et al. (2011) and Sun et al. (2011) who presumed that the electric field could be generated in the ionospheric E layer dynamo region (around the height of 100 km) due to the atmospheric gravity wave which might be induced by the pre-earthquake seismic activity.

In this report, we calculate the perturbations in the electric and magnetic fields, which might be produced by SEF in the ionosphere under different ionospheric conditions.

2. Basic equations

The penetration of SEF into the ionosphere is modeled following the similar formalism to that used by Park and Dejnarakitra (1973) to examine the mapping of thunder-cloud electrostatic fields into the ionosphere. Under steady state conditions, the governing equations are

\[ \nabla \cdot \mathbf{J} = 0 \]  
(1)

\[ \mathbf{J} = \sigma \mathbf{E} \]  
(2)

\[ \mathbf{E} = -\nabla \Phi \]  
(3)

where \( \mathbf{J} \) is the electric current density, \( \sigma \) is the electrical conductivity tensor, \( \mathbf{E} \) and \( \Phi \) are the electrostatic field and potential, respectively. Neglecting the Earth’s curvature, and using cylindrical coordinates \((r, \phi, z)\) centered at a forthcoming earthquake epicenter and with the \( z \) axis pointing vertically upward, we represent the seismogenic perturbation in the vertical atmospheric field near the Earth’s surface by the Gaussian-like spatial distribution

\[ \Delta E_z = E_0 \exp[-\ln(10)(r/R_0)^2] \]  
(4)

where \( E_0 \) and \( R_0 \) are the peak value and the scale size of electric field perturbation, respectively. If one assumes that the geomagnetic field \( \mathbf{B} \) is vertical, and the electrical conductivity tensor depends only on \( z \), the following equation for the electrostatic potential \( \Phi \) can be obtained from (1)–(3)

\[ \partial^2 \Phi / \partial^2 r + (1/r) \partial \Phi / \partial r + (1/\sigma_0) \partial (\sigma_0 \partial \Phi / \partial z) / \partial z = 0, \]  
(5)

where \( \sigma_0 \) is the Pedersen conductivity, and \( \sigma_1 \) is the specific conductivity. At altitudes below 70 km, the conductivity is isotropic \((\sigma_0 = \sigma_p)\) since the geomagnetic field does not affect drifts of charged particles. Above 70 km, the presence of the geomagnetic field results in the anisotropy of the conductivity \((\sigma_0 \neq \sigma_p)\). The Eq. (5) can be solved analytically if the conductivities \( \sigma_0 \) and \( \sigma_p \) depend exponentially on altitude. In the case of isotropic conductivity (setting \( \sigma_0 = \sigma_p = b \exp(z/h) \) where \( b \) and \( h \) are constants), we obtain

\[ \Phi = \int_0^\infty J_0(kr)[A_1(k) \exp(c_1z) + B_1(k) \exp(c_2z)]dk \]  
(6)

where \( J_0 \) is the zero-order Bessel function of the first kind, \( A_1 \) and \( B_1 \) are coefficients, \( c_1 = -1/(2h) - [1/(4h^2) + k^2]^{1/2} \), \( c_2 = -1/(2h) + [1/(4h^2) + k^2]^{1/2} \).

For the anisotropic region, where \( \sigma_0 = b_0 \exp(z/h_0) \) and \( \sigma_p = b_p \exp(z/h_p) \), the solution to Eq. (5) is

\[ \Phi = \int_0^\infty J_0(kr)[A_2(k)J_1(2h_0k) + B_2(k)K_1(2h_0k)]k^2dk \]  
(7)

where \( J_1 \) and \( K_1 \) are the \( q \)-order modified Bessel functions of the first and the second kind, respectively, \( A_2 \) and \( B_2 \) are coefficients, \( v = h_p/(h_p - h_0) \), \( f = 2v_0(h_0 h_p)/h_0^2 \exp[-z/(2\nu_0)] \).

The coefficients \( A_1 \), \( B_1 \), \( A_2 \), and \( B_2 \) are determined from boundary conditions.

The electric field components are given by

\[ E_r = -\partial \Phi / \partial r \]  
(8)

\[ E_z = -\partial \Phi / \partial z \]  
(9)

Since we assume that the geomagnetic field \( \mathbf{B} \) is vertical, \( E_r \) is perpendicular to \( \mathbf{B} \), while \( E_z \) is parallel to \( \mathbf{B} \).

Above 90 km, the geomagnetic field aligned conductivity \( \sigma_0 \) is sufficiently high and much larger the transverse conductivity \( \sigma_1 \), so the geomagnetic field lines of force are nearly equipotential lines for the case of perpendicular electrostatic fields with scale sizes of more than a few tens of kilometers. It makes possible to consider the ionospheric region from \( \sim 90 \) to \( \sim 600 \) km as a thin conducting layer with a geomagnetic field line integrated Pedersen conductivity \( \sum_p \) (Note that the local conductivity \( \sigma_0 \) is negligible above \( 600 \) km.) Thus the continuity equation of electric current can be written at \( z = 90 \) km in the following form:

\[ \sigma_0 E_z = \nabla \cdot \left( 2 \sum_p \mathbf{E}_p \right) \]  
(10)

where \( \nabla \cdot \) denotes the gradient operator in the two dimensions transverse to \( \mathbf{B} \), the factor 2 before \( \sum_p \) accounts for a contribution of the Pedersen conductivity of the magnetically conjugate ionosphere. Note that the relation similar to (10) was previously used as an upper boundary condition while solving the problem of SEF penetration into the ionosphere by Denisenko et al. (2008) and Ampferer et al. (2010). Eq. (10) is explicitly expressed as

\[ \sigma_0 \partial \Phi / \partial z = 2 \sum_p \left[ \partial^2 \Phi / \partial r^2 + (1/r) \partial \Phi / \partial r \right] \]  
(11)

Relations (4) and (11) represent the lower and upper boundary conditions, respectively, to evaluate the electrostatic potential \( \Phi \).

The magnetic effect of seismogenic electric current is described by the Biot–Savart law, which in our case of
azimuthal symmetry, leads to the following estimate of magnetic field perturbation $\Delta B$ in the topside ionosphere

$$\Delta B(r) = -[\mu_0/(8\pi r)] \int_0^{2\pi} \int_0^r 2(\sigma_0 \partial \Phi / \partial r) \exp(z-90) \, \rho \, d\phi \, dr$$

for $z \geq 600$ km (12)

where $r$ is the distance to the $z$ axis, $\mu_0$ is the vacuum permeability. In (12), we take into account that the vertical electric current at $z \geq 600$ km is two times less than the vertical current at $z = 90$ km due to the horizontal outflow of the electric current within the ionospheric layer between 90 and 600 km. The seismogenic magnetic field vector $\mathbf{AB}$ is transverse to the ambient geomagnetic field $\mathbf{B}$ and is aligned with concentric circle lines around the $z$ axis. The direction of $\mathbf{AB}$ is clockwise or anticlockwise (if seen from above) if the vector $\mathbf{E}_0$ is downward or upward, respectively.

3. Conductivity

Below 70 km, we adopt the model conductivity vertical profiles as shown in Fig. 1. The profiles are divided into two altitude sections (from 0 to 40 km, and from 40 to 70 km), in which the conductivity is isotropic and varies exponentially with $z$: $\sigma_0 = \sigma_{01} = b_1 \exp(z/h_1)$, $\sigma_0 = \sigma_{02} = b_2 [\exp(z-z_1)/h_2]$ where subscripts 1 and 2 stand for the lower and second sections, respectively, $z_1 = 40$ km with the values of $b_{1,2}$ and $h_{1,2}$ to approximately fit the atmospheric conductivity models by Cole and Pierce (1965) for the Section 1 and by Swider (1988) for the Section 2. In the anisotropic region between 70 and 90 km, $\sigma_0$ and $\sigma_p$ are exponentially extrapolated from 70 km to their equinoctial midday and midnight values at $z = 90$ km. At altitudes $90 \leq z \leq 600$ km, the conductivities are found from

$$\sigma_0 = e^2 \left[ \frac{N_e}{m_e v_e} + \sum N_i \frac{v_i}{m_i} \right]$$

for $z \geq 600$ km (13)

$$\sigma_p = e^2 \left[ \frac{N_e v_e}{m_e (v_e^2 + v_i^2)} + \sum m_i (v_i^2 + v_i^2) \right]$$

where subscripts $e$ and $i$ denote electrons and the $i$-th ion species, $N_e$ and $N_i$ are the electron and ion densities, $e$ is the electron charge, $m_e$ and $m_i$ are the electron and ion masses, $v_e$ and $v_i$ are the electron and ion momentum transfer collision frequencies, $\omega_e$ and $\omega_i$ are the electron and ion gyrofrequencies. The frequencies $v_e$ and $v_i$ are from Schunk (1988). The required input parameters are taken from the empirical ionospheric model IRI-2012 (http://omniweb.gsfc.nasa.gov/vitmo/iri2012_vitmo.html) and the neutral atmosphere model NRLMSIS-00 (http://ccmc.gsfc.nasa.gov/modelweb/models/nrlmsise00.php). Fig. 2 illustrates altitude profiles of midnight and noon conductivities in the range of 90–600 km computed for specific conditions of March, 2009 at 50°N, 100°E.

At middle latitudes, the integrated along entire magnetic field line Pedersen conductivity attains its minimal values in solar minimum Equinox at night because during Solstice a contribution of the summer hemisphere ionosphere to the conductivity is considerable. We obtain that during solar minimum, in Equinox, the midlatitude height-integrated Pedersen conductivity $\sum_p$ is commonly in the ranges of 5.0–8.0 S and 0.1–0.2 S for day and night, respectively. However, the nighttime $\sum_p$ can be as low as 0.05 S. Within large-scale plasma bubbles extending to middle
latitudes Huang et al. (2007), $\sum_{p}$ might be less than 0.01 S. Under solar maximum conditions, the values of $\sum_{p}$ are several times larger than in solar minimum.

4. Results and discussion

4.1. Electric field

To compute the seismogenic electrostatic potential distribution from (6) and (7), we impose the following boundary conditions:

1. \[-\partial \Phi / \partial z = E_0 \exp\left(-\ln(10) \left(r/R_0\right)^2\right) \text{ at } z = 0\]
2. $\sigma_0 \partial \Phi / \partial z = 2\sum_{p} \left(\partial^2 \Phi / \partial r^2 + (1/r) \partial \Phi / \partial r\right) \text{ at } z = 90 \text{ km}$
3. $\Phi$ is continuous at $z \geq 0$

Fig. 3 illustrates how $E_r$ normalized to $E_0$ depends on $r$ at $z \geq 90$ for several values of $R_0$ if $\sum_{p} = 0.1$ S. $E_r$ first increases to a maximum and then declines gradually. As the scale size $R_0$ increases, efficiency of the SEF penetration into the ionosphere is enhanced.

Fig. 4 shows the maximum magnitude of $E_r$ normalized to $E_0$ at $z \geq 90$ as a function of $\sum_{p}$ for $R_0 = 1000$ km ($E_{r, \ max}$ occurs at $r_{\ max} \approx 740$ km). It is seen that the maximum $E_r$ is $\approx 0.2$ mV/m for $\sum_{p} \approx 0.05$ S and $\approx 1.0$ mV/m for $\sum_{p} \approx 0.01$ S in the case of $E_0 = 1000$ V/m. Thus the seismogenic electric field attains a perceptible value at ionospheric altitudes for a very low $\sum_{p}$ that can be in the nighttime midlatitude ionosphere during solar minimum or inside large-scale plasma bubbles. For daytime $\sum_{p}$, efficiency of the SEF penetration into the ionosphere is insignificant ($E_{r, \ max} \leq 0.002$ mV/m). $E_{r, \ max}$ is reduced by an order of magnitude for a value of near ground atmospheric conductivity $b_1 \sim 10^{-14}$ S/m, that is commonly suggested to be an estimate of mean monthly/yearly near ground conductivity for fair weather conditions.

Note that although our computations are made under an assumption of verticality of magnetic field lines, Park and Dejnakarintra (1977) claim that account of magnetic field tilt is not essential except of being very close to the magnetic equator.

4.2. Magnetic effect

Computations show that the seismogenic magnetic field perturbation $\Delta B$ in the topside ionosphere ($z \geq 600$ km) does not depend on $\sum_{p}$ and is very slightly affected by the difference in the nighttime and daytime conductivities below 90 km. Fig. 5 represents the seismogenic perturbation $\Delta B$ as a function of $r$ for $E_0 = 1000$ V/m and $R_0 = 1000$ km. $\Delta B$ rapidly rises to a maximum at $\approx 740$ km and then reveal a gradual falling off with an increase of $r$. The maximum magnitude of $\Delta B$ is $\approx 13$ pT. The magnetic field perturbation vector $\Delta B$ is azimuthal and perpendicular to the ambient magnetic field force lines.

The seismomagnetic effect is equally pronounced during both daytime and nighttime for any level of solar activity. This effect is too small to be reliably detected by any up-to-date satellite magnetometer, but in the future it will probably be possible when an extremely high-sensitive satellite instrument will be available. The attractive advantage of detection of the seismogenic perturbation $\Delta B$ is that the properly observed spatial distribution of $\Delta B$ would enable in principle to estimate a location of the epicenter of forthcoming earthquake.
Fig. 3. The normalized transverse seismogenic electric field at $z \geq 90$ km plotted against $r$ for different values of the scale size $R_0$ in the case of $\Sigma_p = 0.1$ S/m.

Fig. 4. The maximum magnitude of the transverse seismogenic electric field strength normalized to $E_0$ at $z \geq 90$ km as a function of $\Sigma_p$ for $R_0 = 1000$ km.
5. Conclusion

We have performed the calculation of the perturbations in electric and magnetic fields that could be produced in the midlatitude ionosphere before major earthquakes under different ionospheric conditions. The seismogenic electric field at $z = 600$ km can reach noticeable values only at night in solar minimum and inside large-scale plasma bubbles. The seismomagnetic effect with the magnitude of $\sim 13$ pT is predicted for daytime and nighttime conditions at any solar activity.

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