

Solar Wind Control of Electric Fields and Currents in the Ionosphere

Y. I. FELDSTEIN and A. E. LEVITIN

IZMIRAN, Troitsk, Moscow Region 142092, U.S.S.R.

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KERN (1966) substantiated theoretically the feasibility of using geomagnetic field variations on the earth's surface to calculate the spatial structure of the current system. Algorithms of the electric field and current calculation being used at present are examined. The methods for discriminating the geomagnetic variations controlled by various solar wind parameters are discussed. The calculation results obtained for the assumption of the uniform ionospheric conductivity are presented. This approximation made it possible to account for the existing regularities in the distribution of field-aligned currents under definite solar wind conditions and to predict the character of the three-dimensional current system in the intervals with the positive B_z IMF. The methods for and the results of calculating the electrostatic potential, the electric field, the ionospheric and field-aligned currents in the high-latitude ionosphere for nonuniform conductivity are presented in detail. The calculation results are compared with the direct measurements of the respective parameters.

1. Introduction

The electric and magnetic fields in the Earth's magnetosphere give rise to the large-scale convective motions of plasma. The changes of the motions in space and time which are closely associated with the solar wind create the electromagnetic weather of the near Earth space environment. On the Earth's surface the electromagnetic weather leaves its signature in the geomagnetic field variations which respond keenly to any changes of the electric fields in the Earth's magnetosphere and ionosphere. The geomagnetic field variations at high latitudes (at corrected geomagnetic latitude $\Phi \geq 60^\circ$) are of special interest because in this zone the remote magnetospheric regions interact directly with the solar wind via magnetic field lines.

A number of international research programs involved in the collection of observational data at the Earth's upper atmosphere and magnetosphere have recently been completed resulting in data bases which are unparalleled in the history of science. Present-day science is faced with the most important and urgent task of systematizing and classifying these data bases and of finding the relationships among the diverse processes they describe.

International Geophysical Year (IGY) data analyses yielded two concepts which facilitated the ordering of the space-time description of the occurrence of geophysical events, namely the concept of the auroral oval (FELDSTEIN *et al.*, 1969) and the concept of the auroral (magnetospheric) substorm (AKASOFU, 1968). The oval

concept made it possible to order the events observed at high latitudes because the auroral oval is closely associated with large-scale magnetospheric structure and may be used as a natural, spatial coordinate system to describe the processes occurring in the high-latitude, upper atmosphere. The substorm concept facilitated the temporal ordering of the events since the character of the processes in the upper atmosphere and in the magnetosphere is essentially contingent upon the location and the rate of injection and dissipation of solar wind energy inside the magnetosphere.

In recent years, it has become obvious that the occurrence of the high-latitude geophysical events in the Earth's environment is controlled to a great degree by the interplanetary medium parameters, namely, by the direction and intensity of the interplanetary magnetic field (IMF) vector, and by the solar wind plasma velocity and density (AKASOFU and CHAPMAN, 1972; NISHIDA, 1978; LYATSKY, 1978; LYATSKY and MALTSEV, 1983; SERGEEV and TSYGANENKO, 1980). The quantitative relations were established between the electric and magnetic fields and electric currents at altitudes above 100 km and the interplanetary magnetic fields. As a result, the third concept was formulated, namely, the interplanetary medium parameters control the electromagnetic processes in the high-latitude magnetosphere (the ionosphere included) which give rise to the geophysical events in this region of the Earth's environment. The International Magnetospheric Study Project (IMS) has made it possible to further develop the concept using complex experimental data. Subsequent years saw the construction of the quantitative models of the space-time distributions of the large-scale electric fields and of the ionospheric and field-aligned currents in the auroral and polar cap regions. The most comprehensive models were constructed on the basis of the geomagnetic field variations observed on the Earth's surface. Thus, the long-standing dream of geophysicists became real, namely, the algorithm was developed to infer the three-dimensional current system and the quasi-steady state electromagnetic field of the ionosphere from ground-based geomagnetic data.

The feasibility of using the ground-based magnetic data to obtain the electric fields and currents of the ionosphere has initiated an essentially new and very important phase in the development of the geophysics. Novel numerical methods for solving the equations which relate the magnetic field on the Earth's surface H to the ionospheric electric field potential ϕ made it possible to infer the real-time spatial distribution of fields and currents in the entire high-latitude region from the continuous recording of geomagnetic variations at stations in the existing network. At the same time, the vast volume of geomagnetic data accumulated during quite a number of years has made it possible to carry out the statistical studies, namely, to relate the geomagnetic field variations to the interplanetary medium parameters.

In numerous works, the combination of ground-based and space-craft measurements proved to be a powerful tool to study the diverse and complicated events in both space and time. As stressed by VASYLIUNAS (1970, 1972), TVERSKOY (1972), GUREVICH *et al.* (1975, 1976), and BOSTRÖM (1975), the magnetosphere and the ionosphere must be examined as a whole. This idea was realized by LYATSKY and MALTSEV (1972), and JAGGI and WOLF (1973) and was developed further via extensive, computer simulation models. The numerical simulations of the Rice University group have become a power tool to study the dynamics of the magnetosphere-ionosphere system (HAREL *et al.*, 1981).

In the subsequent sections of this work we shall trace the evolution of the ideas set

forth by CHAPMAN (1935) and ALFVÉN (1950) concerning the character of the system responsible for the geomagnetic field variations observed in high latitudes. Subsequently KERN (1966) theoretically proved the feasibility of using geomagnetic field variations on the Earth's surface to calculate the spatial structure of the current system for a uniform anisotropic conductivity of the ionosphere. Then, KRYLOV (1974) and GUREVICH *et al.* (1975) developed methods for the same calculations using nonuniform ionospheric conductivity. We shall also examine the electric field and current calculation algorithms used at present elsewhere. Since the calculations are based on the high-latitude geomagnetic field variation data, we shall briefly discuss the methods for discriminating the geomagnetic variations controlled by various solar wind parameters. The calculation results are compared with the direct measurements of the respective parameters in the near Earth space.

2. Magnetic Field Variations on the Earth's Surface and Associated Current Systems

The spatial distribution of magnetic field vectors on the Earth's surface do not permit an unambiguous definition of the current system which caused a given observed geomagnetic variation. The solution of such a problem requires that definite assumptions to be introduced concerning the current configuration, the conductivity distribution, the ionospheric thickness etc. The current systems most vividly discussed were proposed by Birkeland, Chapman, Vestine, and Alfvén, namely, the Chapman-Vestine system (VESTINE and CHAPMAN, 1938) and the Birkeland-Alfvén system (BIRKELAND, 1913; ALFVÉN, 1950). Later on, the magnetic field variations at the Earth's surface were explained using both pure ionospheric and three-dimensional current systems (KIRKPATRICK, 1952; BOSTRÖM, 1964; AKASOFU and MENG, 1969).

The unambiguous experimental evidence for the existence of the field-aligned currents above the ionosphere was obtained by analyzing the satellite magnetic measurements at altitudes of ~ 1000 km (ZMUDA and ARMSTRONG, 1974 and the reviews by POTEMRA, 1983; AKASOFU, 1984; FELDSTEIN *et al.*, 1982a; SAFLEKOS *et al.*, 1982). The measurements were compared against simultaneous theoretical calculations which showed that, for uniform ionospheric conductivity and radial magnetic field lines along which the field-aligned current, j_{\parallel} , is streaming, the field-aligned current-produced magnetic signal on the Earth's surface is completely cancelled by the magnetic field of the ionospheric Pedersen currents (FUKUSHIMA, 1976). This theorem is satisfied for both planar and spheric ionospheres and for any distribution of the currents streaming to and from the ionosphere.

Fukushima's works impelled most of the researchers to conclude that the real three-dimensional current system in the high-latitude ionosphere cannot in principle be calculated using ground-based magnetic data only. This opinion prevailed despite KERN (1966) where he proposed a scheme to compute the spatial distribution of the ionospheric and field-aligned currents using values of the magnetic field \mathbf{H} on the Earth's surface. This situation prevailed until MISHIN and POPOV, 1969) used Kern's relation to calculate the values of j_{\parallel} from the magnetic field of the SD -variation in May-August, 1958. Figure 1 shows the field-aligned current isolines obtained by Mishin and Popov for the northern (top) and southern (bottom) hemispheres. Current flowing into the ionosphere is drawn with the solid lines, and currents from

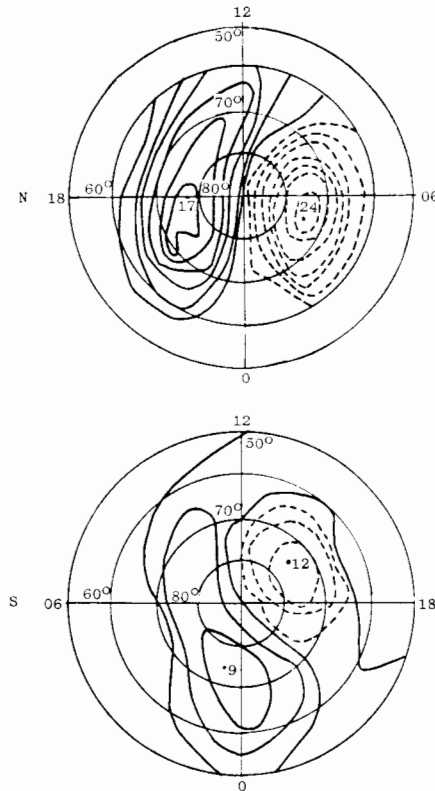


Fig. 1. The distribution of j_{\parallel} in the northern (above) and southern (below) hemispheres inferred from the data from the disturbed solar-diurnal geomagnetic field variation SD in May–August, 1958 (MISHIN and POPOV, 1969). The dashed lines show the field-aligned current flowing out of the ionosphere, the solid lines show the field-aligned current flowing into the ionosphere. The contour interval is 3.2×10^{-7} A/m².

the ionosphere with the dashed lines. Their criticism of the Kern formula was most probably motivated by the strange patterns in the spatial distribution of j_{\parallel} that they obtained. Present-day knowledge of large-scale field-aligned currents during magnetic disturbances (IMF $B_z < 0$), does not agree with the j_{\parallel} distribution shown in Fig. 1. We must note here that some scientists still have doubts as to the validity of the methods for inferring three-dimensional current systems from the ground-based magnetic data only.

For simple models of the nonuniform ionospheric conductivity, the relationship between the equivalent and field-aligned currents was represented by analytical expressions by LYATSKY and MALTSEV (1972). The theory for simulating the electric fields and currents in the high-latitude ionosphere from ground-based geomagnetic data was developed in a series of works (the relevant references may be found in KRYLOV, 1974 and GUREVICH *et al.*, 1975, 1976). The authors have derived the equation for the electric field potential ϕ in the Earth's magnetosphere and ionosphere. For non-uniform ionospheric conductivity, the equation must be solved by numerical methods. FAERMARK (1980) described the difference method calcula-

tion procedure to infer the electric fields and currents in the latitudinal zone $\geq 60^\circ$ from the geomagnetic variations obtained by AFONINA *et al.* (1980) and LEVITIN *et al.* (1982). The calculations are made in terms of a particular model of the conductivity of the high-latitude ionosphere.

MATVEEV and SHPYNEV (1975) following Kern's method developed an equation for computing electrostatic potential which is analogous to that obtained by GUREVICH *et al.* (1975) and allows for nonuniform ionospheric conductivity. The solution was aimed at the analytical representation of the geomagnetic variation field from a confined, nonuniformly-distributed network of stations (MISHIN *et al.*, 1977). The scheme of the solution was as follows: 1) The current function Ψ is determined by selecting the optimal spectrum of spherical harmonics for the initial magnetic variation field. 2) The magnetic and electrostatic potentials are represented by a spherical harmonic expansion. 3) N linear equations for the potential expansion coefficients are formed on the set of N points at which the conductivity, its derivative, and the current function are given. A similar method was used by MISHIN *et al.* (1980; 1981 and 1982 and the references therein) to infer the electric fields and currents in the high-latitude ionosphere from ground-based geomagnetic data.

In the late seventies, KISABETH (1979) developed a different method for finding the parameters of three-dimensional current systems from the data of a chain of geomagnetic stations (see also KISABETH and ROSTOKER, 1977). The method does not solve the equation for the potential, rather it selects ionospheric current intensities which permit the best convergence of the calculated magnetic disturbance produced by a given current system with the real magnetic signal detected by a meridional chain.

KAMIDE *et al.* (1981) presented the height-integrated ionospheric current to be a sum of the toroidal and poloidal currents related to the equivalent current function Ψ (see also KERN, 1966). The logic of the inference of fields and currents from ground-based geomagnetic data is as follows: 1st obtain the initial magnetic field variations using the data from a network of high-latitude observatories; 2nd determine the potential of the initial magnetic field ($H = -\nabla U$) and the current function Ψ ; 3rd select a conductivity model (Σ_H, Σ_P); 4th numerically solve the equation for electrostatic potential ($E = -\text{grad } \phi$) by the difference method on the network with a latitude step 1° and a longitude step $\sim 15^\circ$; 5th compute the horizontal current distribution; and 6th compute j_{\parallel} as the divergence of the horizontal current. Numerical procedures are described in more detail by KAMIDE and KROEHL (1984).

Thus, independently of each other, several groups of researchers have developed and applied the calculational methods which make it possible to infer the distribution of the potential ϕ , the electric field E , the horizontal ionospheric current j_{\perp} , and the field-aligned current j_{\parallel} from the variation of the geomagnetic field H on the Earth's surface. In their calculations, each of the groups used a particular conductivity model, a particular data base for the magnetic field variations, and a particular method of numerical calculations.

3. The High-Latitude Geomagnetic Variation Field Structure Relevant to the Solar Wind Parameters

Both magnetospheric substorm fields and the magnetic field variations at high latitudes closely related to the parameters of the solar wind and of the frozen IMF

(AKASOFU, 1977, 1981; NISHIDA, 1978, 1983; BOSTRÖM, 1975; FRIIS-CHRISTENSEN, 1981; ALFVÉN, 1977, 1981; FELDSTEIN, 1976; TROSHICHEV, 1982; KAMIDE, 1982, 1984; STERN, 1983; SAFLEKOS *et al.*, 1982). The variations involve the entire high-latitude region from the auroral zone to the pole and are observed continuously during quiet and disturbed periods. The DP2 variations relevant to the B_z IMF component and described first by NISHIDA (1968a, b) are the characteristic example of such geomagnetic field variations. SVALGAARD (1968) and MANSUROV (1969) noted the characteristic field variations relevant to the IMF sector structure. Later on, FRIIS-CHRISTENSEN *et al.* (1972), SUMARUK and FELDSTEIN (1973), and KAWASAKI *et al.* (1973) showed that these variations were controlled by the B_y component. Based on the close relationships of all three geomagnetic field components in the near-pole region to the B_y component, FELDSTEIN *et al.* (1975) and FRIIS-CHRISTENSEN and WILHJELM (1975) proposed a regression method for extracting the B_y component-controlled fraction of the high-latitude magnetic field variation. The method was used later by MAEZAWA (1976) for the B_z component. The method was then used by MISHIN (1978) and AFONINA *et al.* (1979) to find the geomagnetic field variation fraction controlled by solar wind velocity, V , and density, n . The near-pole magnetic field variations controlled by the positive B_z values were identified by IWASAKI (1971), MAEZAWA (1976), KUZNETSOV and TROSHICHEV (1977), and others.

The regression analysis method permits the magnetic field components on the Earth's surface to be related quantitatively to the solar wind plasma parameters and to the IMF components. The analysis was carried out by several groups with different sets of parameters. For example, LEVITIN *et al.* (1982) have presented hourly means of the horizontal and vertical components of the field variations on the Earth's surface in the $\Phi \geq 60^\circ$ latitude region to be the series

$$H_E^i = K_i^{B_y} B_y + K_i^{B_z} B_z + K_i^{B_x} B_x + K_i^{nV^2} n \left(\frac{V}{100} \right)^2 + K_i^{V^2} \left(\frac{V}{100} \right)^2 + K_i^{|B_y|} |B_y| + H_0^i. \quad (1)$$

The series is assumed to be sufficient when describing the features of the geomagnetic field variations for the quasistationary processes in the Earth's magnetosphere as related to the interplanetary medium conditions. The addition of new terms fails to improve the agreement between the observed and model-calculated values of H_E^i . The coefficients K_i are found by the least squares method from the condition of the best description of the disturbed field through the IMF components and the interplanetary parameters. Each of the series terms is a geomagnetic field response to a certain interplanetary medium parameter, while the residual term H_0^i is the extrapolation of the mid-latitude S_q current system to the high-latitude zone. The determination of the coefficients, K_i , by the least squares method for each UT hour and for each observatory has been described in detail by MAEZAWA (1976), TROSHICHEV and TSYGANENKO (1979), and LEVITIN *et al.* (1982). The coefficients $K_i^{B_x}$ and $K_i^{|B_y|}$ are much smaller than $K_i^{B_y}$ and $K_i^{B_z}$, while the terms depending on n and V may be combined with H_0^i to constitute a single constant, $H_0 = K_i^{nV^2} n (V/100)^2 + K_i^{V^2} (V/100)^2 + H_0^i$, which characterizes the mean conditions in solar wind ($n=4$ particles/cm³ and $V=500$ km/s from LEVITIN *et al.*, 1982). Thus, if the magnetic field variations are presented as a series in solar wind parameters, the mean values of the solar wind plasma parameters may be limited to the trinomial

$$H_E^i = K_i^B B_y + K_i^B B_z + H_0. \quad (2)$$

The representation (1) has to be used in case of the extremal values of n and V .

The series-expansion (2) of the hourly means of the high-latitude field variation components was used by MAEZAWA (1976), BELOV *et al.* (1977), TROSHICHEV and TSYGANENKO (1979), and TROSHICHEV *et al.* (1979). A similar expansion was made by FRIIS-CHRISTENSEN (1979) using the 20-min means of the field components from the Greenland meridional chain of magnetic observatories. Two sets of the coefficients K_i were derived for the $B_z < 0$ and $B_z > 0$ data sets, respectively. The use of the hourly or 20-min means of the field components and the fairly simple H_E^i representation in terms of a series of the interplanetary medium parameters have permitted the description of only the quasisteady-state geomagnetic field variations. A somewhat different approach has to be used in describing the rapid and complicated magnetic field variations during magnetospheric substorms which are closely related to the intrinsic magnetospheric processes giving rise to the local changes of the current system geometry. For example, the mean error in calculating H_E^i from the relation (2) was estimated (LEVITIN *et al.*, 1982) to be ~ 30 nT, in polar cap and ~ 50 nT in auroral zone. The substorm-time disagreement may rise to ~ 100 nT because the applied methods for separating the high-latitude geomagnetic field variations relevant to the interplanetary medium parameters make it possible to describe only the fairly smooth geomagnetic field fluctuations related directly to the interplanetary medium parameters. Another and more original approach to describing the pulsed field variations during substorms seems to be necessary.

The equivalent currents used below have been obtained (BELOV *et al.*, 1977; LEVITIN *et al.*, 1982) by the regression analysis method and are due to a 1-nT change of the B_y , $B_z < 0$, $B_z > 0$ components and in case of $B_z = B_y = 0$.

MISHIN *et al.* (1978, 1982 and the references therein) made use of another method for finding the on-surface geomagnetic field variations controlled by the IMF components and by the solar wind parameters. The magnetic disturbance potential, U , on the Earth's surface is presented to be the series

$$U(\Theta, \Lambda) = R_E \sum_{k=1}^{\infty} d_k G_k(\Theta, \Lambda), \quad (3)$$

where G_k is the eigenfunction of the Laplace equation. The coefficients d_k are calculated using the measured values of the X , Y , Z -components of the geomagnetic field vector from the network of stations $\{\Theta_i, \Lambda_i\}$ ($i = 1, \dots, N$). The current function, Ψ_e , which describes the equivalent ionospheric current, is determined through the same coefficients d_k . The finite number of observation points restricts the length of the series (3). Therefore, the "optimal spectrum" (definite values of K) is selected by estimating the errors in the coefficients of the spherical series. Such a selection is made using the orthogonalization of the basis function on the chosen set of reference points.

After that, the coefficients, d_k , are subjected to regression analysis whose regression series contains terms comprising the solar wind parameters, the universal time (UT), and the geomagnetic activity indices. The length of the series is varied and may include up to 40 terms, namely, $b = \text{const}$, V , V^2 , V^3 , $V \cos(\text{UT})$, $V \sin(\text{UT})$, n , n^2 , $n^{1/3} V$, nV , nV^2 , B_y , B_y^2 , B_y^3 , $B_y V$, $B_y^2 V$, $B_y V^2$, B_z , B_z^2 , B_z^3 , $B_z V$, $B_z^2 V^2$, $B_z V^2$, $b \cos(\text{UT})$, $b \sin(\text{UT})$, $b \cos(2\text{UT})$, $b \sin(2\text{UT})$, AE , AE^2 , $B_y B_z$, $B_y^2 B_z$, $B_y B_z^2$, $n \cos(\text{UT})$, $n \sin(\text{UT})$,

$B_1 \cos(UT)$, $B_1 \sin(UT)$, $B_2 \cos(UT)$, $B_2 \sin(UT)$, etc.

The reference points $\{\Theta_i, \Lambda_i\}$, whose data are used to determine d_k , amount to about 100, while the time interval of data sampling is some 10 days. The night-time values of the X , Y and Z -components of the geomagnetic field vector (quiet geomagnetic conditions) are taken to be the reference level for X , Y and Z variations.

The drawbacks of the analysis method described above are (i) the necessity for the reference level of the geomagnetic variation field to be introduced before making the correlation analysis and the associate possible distortion of the real current systems controlled by the solar wind parameters; (ii) the application of the spherical harmonic analysis in the high-latitude zone, especially in polar cap, where the magnetic observatories are rare and distributed nonuniformly (there are less than ten of them in polar cap), while geomagnetic variations are local. As a result, the very high harmonics of the expansion are inferred from the observation point data which are not very sensitive to the variations localized in the polar cap; (iii) a great number of arguments in the regression equation relating d_k to the solar wind parameters. As a result, the model gets tedious, the expansion terms depending on each other appear, and the "elementary" geomagnetic variations controlled by B_1 and B_2 components become impossible to separate; (iv) a low reliability of the results because they were obtained for small sample size without grouping into the $B_z > 0$ and $B_z < 0$ IMF time intervals.

BAZARZHAPOV *et al.* (1983) have also employed the regression analysis method in a manner similar to that used by MAEZAWA (1976), BELOV *et al.* (1977), FRIIS-CHRISTENSEN (1979), and TROSHICHEV and TSYGANENKO (1979). They have obtained a geomagnetic variation model based on the regression equations relating the hourly means of the X , Y , Z -components of the geomagnetic field vector at given stations to the solar wind parameters.

4. Can the Three-Dimensional Current System Be Inferred from the Geomagnetic Field Variations on the Earth's Surface?

In Section 2 we discussed the algorithms proposed by various groups for restoring the character of the current system responsible for the geomagnetic field variations observed on the Earth's surface. The natural question arises: what is the extent to which the results obtained are correct? The question may be answered in only one way, namely, by comparing the model calculation results with direct space-born measurements, either instantaneously at a given point in time or averaged to obtain a global picture, or by comparison with radar data (LEVITIN *et al.*, 1977a, b; FELDSTEIN *et al.*, 1982b, c, 1984a; BAKER and KAMIDE, 1985).

KAMIDE *et al.* (1982) and WOLF and KAMIDE (1983) tested the representativity of the method for calculating the magnetospheric fields and currents from ground-based magnetic data. They used a two-step procedure in which (1) the magnetic signal at the Earth's surface was calculated for given electromagnetic conditions in the magnetosphere and (2) the distributions of the magnetospheric fields and currents were calculated from a magnetic disturbance on the Earth's surface using an appropriate algorithm, whereupon the distributions obtained were compared. However, the attempts were aimed rather at verifying the correctness of the mathematical formalism involved than at testing the physical assumptions used to

restore the magnetospheric fields and currents from the geomagnetic variations at the Earth's surface.

By 1976–1977, when FAERMARK *et al.* (1976) and LEVITIN *et al.* (1977a, b) started using the ground-based magnetic data to restore the current system in the Earth's magnetosphere, the space-time distribution of the large-scale field-aligned currents had been safely inferred from the observations on board the OGO, TRIAD, and ISIS-2 satellites (SUGIURA, 1975; ZMUDA and ARMSTRONG, 1974; IJIMA and POTEIRA, 1976a, b; POTEIRA, 1977; BURROWS *et al.*, 1976; MCDIARMID *et al.*, 1977). The corrected geomagnetic latitude-local geomagnetic time coordinates are used in Fig. 2(a)–(d) to present several schemes of the field-aligned current distribution inferred from the Triad observations. The disturbances transverse to the force lines of the main geomagnetic field, and hence the field-aligned currents, exist at any time of a day and at any magnetic activity level. The field-aligned currents flow into and out of the ionosphere at auroral latitudes. During the 1400–2300 (0000–1000) LT interval, the field-aligned current is inward (outward) on the equatorward side of the oval and is outward (inward) on the poleward side of the auroral oval. According to ZMUDA and ARMSTRONG (1974), magnetic disturbances are observed continuously from 1000 to 1400 LT and do not involve any definite characteristic direction of the field-aligned currents (Fig. 2(a)). According to IJIMA and POTEIRA (1976b), however, field-aligned current zone 3 is located poleward of zone 1 during magnetically quiet periods (Fig. 2(b)). In contrast to zone 1, the zone 3 field-aligned current flows into the ionosphere in the afternoon sector and out in the prenoon sector, but the chaotic direction of j_{\parallel} is preserved near noon (the oblique shading in Fig. 2(b)). Zone 3 is associated with the dayside cusp, the j_{\parallel} intensity in that zone increases with enhanced the southward B_z component and is independent of the B_y component. The character of the relationships of j_{\parallel} to the IMF components in zone 3 was established by IJIMA and POTEIRA (1976b) using the limited data of 11 orbits of the satellite. However, the pre- and afternoon traversals of the dayside cusp were not separated and the intensities and signs of the B_y component during the traversals were not indicated. In this model, the field-aligned current zones 1 and 2 exist during both quiet and disturbed periods, while zone 3 disappears during disturbed periods (Fig. 2(c)). The polar cap is characterized by small-scale transverse disturbances of the geomagnetic field (Fig. 2(d)) which are most frequently (>75% of the cases) detected in the dawn sector from 0300 to 0900 MLT for $75^\circ < \Phi < 82^\circ$. In the north polar cap, the disturbances are more frequent when the IMF is antisunward than in case of the sunward IMF. During the periods of persistent and intensive $B_z > 0$, the large-scale field-aligned currents at auroral oval latitudes degenerate into small-scale currents.

Following LEVITIN *et al.* (1977a, 1982), Figures 3(a)–(d) present the results of the model calculations of the space-time distributions of j_{\parallel}^M inferred from the magnetic field variation data. The height-integrated ionospheric conductivity was assumed to be uniform, thereby permitting j_{\parallel}^M to be calculated from the simple relation

$$j_{\parallel}^M (\text{A/km}^2) = \frac{10}{4\pi} \frac{\Sigma_p}{\Sigma_H} \text{div } H (\text{nT/km}) .$$

The allowance for the nonuniform ionospheric conductivity in the real summer conditions does not in practice effect the character of the space-time distribution of j_{\parallel} .

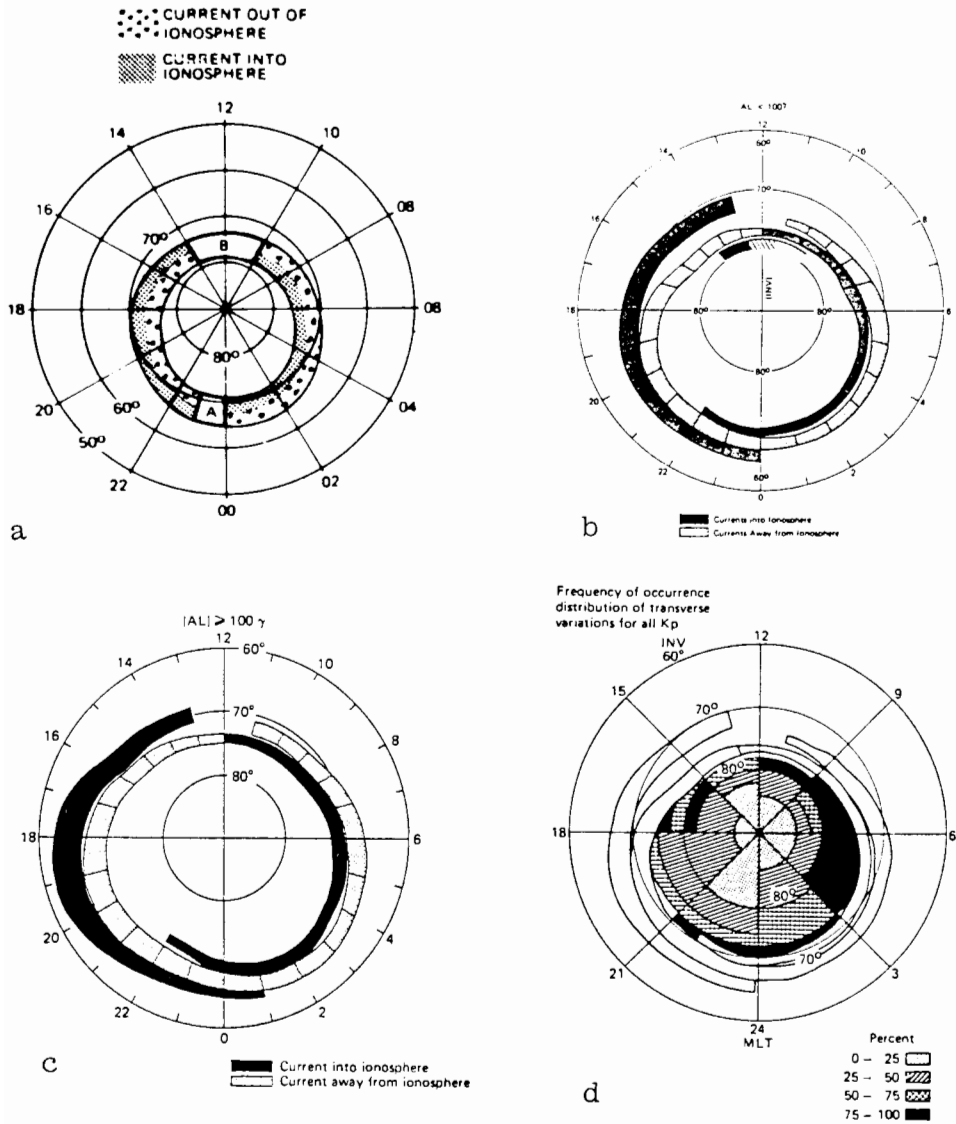


Fig. 2. The space-time distribution of j_i inferred from TRIAD data. (a) According to ZMUDA and ARMSTRONG (1974). The currents flowing both into and from the ionosphere are located in region A. B is the irregular region. (b) Magnetically quiet intervals according to IJIMA and POTEMRA (1976b) (c) Magnetically disturbed period according to Iijima and Potemra. (d) Field-aligned currents in the polar cap according to SAFLEKOS *et al.* (1978).

The values of j_i^M shown in Figs. 3(a)–(d) involve a calibration of different effects including the magnetic field of the current induced in the Earth, the height of the current-carrying ionospheric layer, etc. The isolines contour the regions of equal-intensity field-aligned currents. In Fig. 3(a) the field-aligned currents $j_i^M(H_0)$ independent of the IMF direction cover the entire high-latitude region and are divided

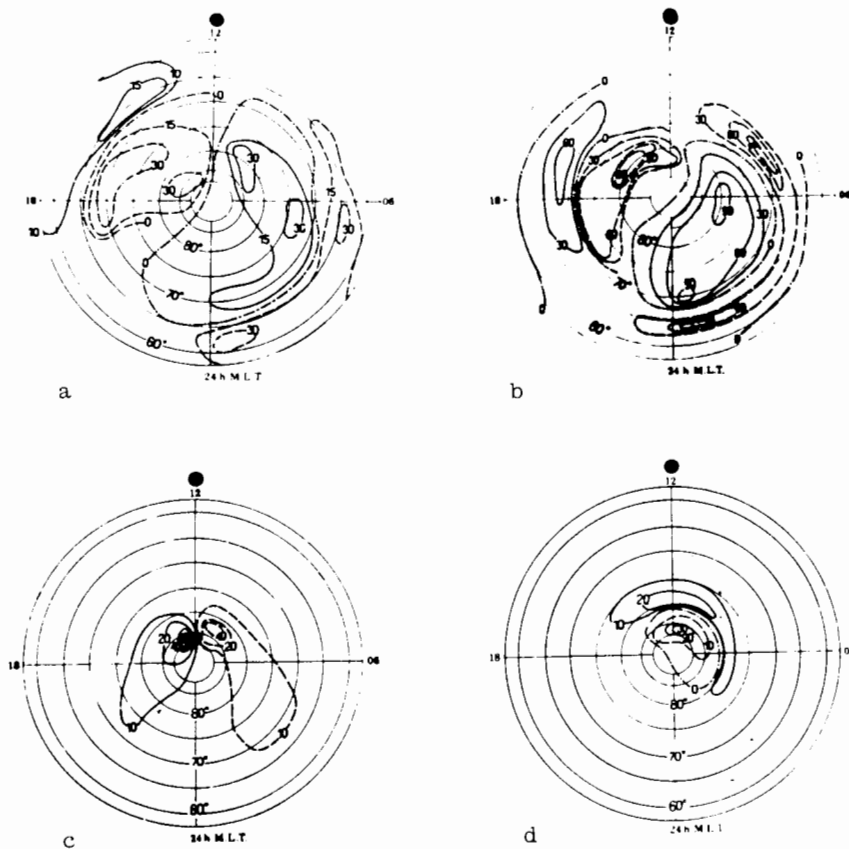


Fig. 3. The results of model-simulating for the space-time distribution of j_{\parallel} in summer (LEVITIN *et al.*, 1977, 1982) obtained in various interplanetary medium situations. The height-integrated ionospheric conductivity is taken to be homogeneous. The value of the isolines of j_{\parallel} are in units of $3 \times 10^{-9} \text{ A/m}^2$. Corrected geomagnetic latitude-local geomagnetic time coordinates are used. The solid lines show the flowing currents into the ionosphere and the dashed lines show the outflowing currents. (a) $B_y = B_z = 0$, $n = 4 \text{ particle/cm}^3$, $V = 500 \text{ km/s}$. (b) $B_z = -1 \text{ nT}$. (c) $B_z = 1 \text{ nT}$. (d) $B_y = 1 \text{ nT}$.

by the 2200–1100 MLT meridian into two zones where the dawn (dusk) sector currents flow into (out of) the ionosphere in the poleward region and flow out of (into) the ionosphere in the equatorward region. The maximum density of the currents is $\sim 0.3 \times 10^{-6} \text{ A/m}^2$. The space-time structure of $j_{\parallel}^M (B_z < 0)$ in Fig. 3(b) is similar to that of $j_{\parallel}^M (H_0)$. Figure 3(c) presents the distribution of $j_{\parallel}^M (B_z > 0)$. The currents are located mainly in the polar cap region, flow into the ionosphere in the dusk sector, and flow out in the dawn sector. The currents $j_{\parallel}^M (B_y)$ controlled by the azimuthal IMF component are mainly concentrated near the dayside cusp and cover the day sector from dawn to dusk. For $B_y > 0$ ($B_y < 0$) the currents flow into (out of) the ionosphere on the equatorward side and flow out of (into) the ionosphere on the poleward side (Fig. 3(d)). The intensity and direction of j_{\parallel}^M at a particular spatial point of the corrected geomagnetic latitude—local geomagnetic time coordinates is a function of the IMF.

There is, therefore, a probability for a satellite to record j_{\parallel} different not only in value but also in direction at the same point, but at different UT time.

Like any model aimed at quantitative description of the regularities in the real situations, the model proposed must

- (i) agree with the main regularities found experimentally in the field-aligned current distribution,
- (ii) permit interpretation of individual observed deviations from such regularities,
- (iii) predict new, yet unknown, regularities.

The j_{\parallel}^M model (Figs. 3(a)–(d)) inferred from the observations of the geomagnetic field variations at the Earth surface satisfies the above conditions.

The general space-time j_{\parallel}^M structure relevant to the IMF-independent geomagnetic variations (Fig. 3(a)) agrees with the structure of the large-scale field-aligned currents observed by TRIAD (Fig. 2(a)) although the j_{\parallel}^M intensity in the maxima is somewhat lower due to the smearing of the field-aligned currents over a large area compared with the data of individual orbits. The permanent solar wind “streaming” around the magnetosphere cause the permanent existence of the region 1 and 2 field-aligned currents. Geomagnetic activity during $B_z < 0$ periods is accompanied by an enhancement of the field-aligned currents in zones 1 and 2 with simultaneous conservation of their spatial distribution structure (Figs. 3(b) and 2(c)). The dependence of the zone 1 j_{\parallel} intensity on the value of the southward ($B_z < 0$) component agrees with the results of the direct measurements (MCDIARMID *et al.*, 1978; IJIMA and POTESRA, 1982). The zone 1 field-aligned current intensity increases as the $B_z < 0$ component enhances, but does not vary with increasing $B_z > 0$. This circumstance agrees with the proposed model because it may be interpreted as the existence of two field-aligned current sources in zone 1 of which one is independent of IMF, while the second is controlled by the southward IMF component. The j_{\parallel}^M distribution in the day sector is determined in a critical manner by the IMF intensity and direction. It is the IMF variability that is responsible for the occurrence of the irregular region in Fig. 2(a). By forming their various combinations, the currents $j_{\parallel}^M(H_0)$, $j_{\parallel}^M(B_z)$, and $j_{\parallel}^M(B_y)$ give rise to the variability of the resultant field-aligned current in the high-latitude dayside sector. The distribution of j_{\parallel}^M presented in Fig. 3(d) is in a good agreement with the model concept concerning the character of the B_y component-controlled field-aligned currents proposed on other grounds by STERN (1973), LEONTYEV and LYATSKY (1974), VOLLAND (1975b), and LYATSKY (1978) and then inferred from the ISIS-2 data by MCDIARMID *et al.* (1979) and from TRIAD data by WILHJELM *et al.* (1978). The chaotic j_{\parallel} directions in zone 3 near local noon in Fig. 2(b) reflect the complicated distribution of the field-aligned currents and the difficulties faced in the analysis of the experimental data. The model calculations of the field-aligned currents in the polar cusp and polar cap (FELDSTEIN *et al.*, 1982c) have shown that the j_{\parallel}^M distribution corresponds to the scheme of IJIMA and POTESRA (1976b) for certain interplanetary conditions and to the scheme MCDIARMID *et al.* (1979) for other conditions. It should be pointed out that the interpretation of the measurement data of particular satellite orbits in terms of one or another model is essentially dependent on the satellite trajectories.

FELDSTEIN *et al.* (1982a) discussed in detail the models for the large-scale magnetospheric field-aligned currents proposed by IJIMA and POTESRA (1976a, b),

HIJIMA *et al.* (1978), and MCDIARMID *et al.* (1979) and concluded that the model shown in Fig. 3 explained the experimental data better.

It is now accepted that the high-latitude distribution of the field-aligned currents is largely controlled by IMF (POTEMRA *et al.*, 1984). This is completely in agreement with the following conclusion (LEVITIN *et al.*, 1977a): "The results of the satellite observations of the field-aligned currents have to be treated not on the basis of the classification according to geomagnetic activity level, but as being contingent on the IMF orientation in the following basic situations: $B_1 > 0$ and $B_2 > 0$, $B_1 > 0$ and $B_2 < 0$, $B_1 < 0$ and $B_2 > 0$, $B_1 < 0$ and $B_2 < 0$ ".

Generally, the magnetic field variations detected by a satellite in the dawn sector are described by the two-layer model for the field-aligned currents flowing into the ionosphere in the near-pole part of the oval and flowing out in the equatorward part of the oval. However, the following cases of anomalous direction were reported by MCDIARMID *et al.* (1977) at 0400–1000 MLT: a field-aligned current flow out of the ionosphere at the poleward edge of the oval and into the ionosphere at its equatorward edge along the ordinary case within the oval. For all cases of the anomalous field-aligned currents the B_1 component was northward. However, the j direction was often normal at high values of $B_2 > 0$. These experimental data can be explained by the model displayed in Figs. 3(c) and (d) and were explained (LEVITIN *et al.*, 1977b) by the dependence of the field-aligned currents not only on B_2 but also on B_1 IMF component. In particular, on several orbit of the satellite the anomalous j direction was due to high values of the $B_1 > 0$ rather than to $B_2 > 0$. In the dawn sector, the normal j direction $B_2 > 0$ must be observed for $B_1 < 0$. The field-aligned current model displayed in Figs. 3(a)–(d) has made it also possible to account for the occurrence of the three-layer j system in the dawn sector on 1–3 of the S3-3 satellite orbits (MOZER *et al.*, 1980).

In the proposed model, a field-aligned current system with maximum intensity at afternoon hours (current inflow) and prenoon hours (current outflow) appears in summer on the dayside of the polar cap during $B_2 > 0$ intervals. According to Fig. 3(c), the density of these currents reaches $\sim 10^{-7}$ A m⁻² at $B_2 = +1$ nT. The direction of the model field-aligned currents coincided with the zone 3 current direction (HIJIMA and POTEMRA, 1976b). However, the modeled currents cover the entire dayside polar cap sector up to the pole, and failed to form narrow bands at $\Phi < 80^\circ$ (see Fig. 2(b)), and were controlled by the $B_2 > 0$ component, rather than by $B_2 < 0$ component. This bold prediction which was to be verified and could be used to test the applicability of all the methods for restoring the three-dimensional current system from the ground-based magnetic observation data. In 1977–1980 our j^M system for $B_2 > 0$ (Fig. 3(c)) provoked vivid discussions and certain distrust, while the method for separating the $B_2 > 0$ component-controlled magnetic field variations on the Earth's surface was cast doubted by BAZARZHAPOV *et al.* (1979), and PUDOVKIN and ZAITSEVA (1983).

The existence of the $B_2 > 0$ component-controlled field-aligned currents in the dayside sector of polar cap which flow into the ionosphere in the afternoon sector and flow out in the prenoon sector was convincingly demonstrated by MCDIARMID *et al.* (1980) and by SAFLEKOS and POTEMRA (1980). The preliminary analysis of the TRIAD data (POTEMRA *et al.*, 1980), demonstrated intense j in the polar cap for $B_2 = 9.8$ nT. COWLEY (1981b) found a fairly high correlation between the intensity of the transverse disturbances of magnetic field inferred from the TRIAD and ISIS-2

data and positive values of the B_z component. The existence of a large scale field-aligned current system in polar cap during summer which is controlled by the $B_z > 0$ component was also confirmed recently by IJIMA (1984) and ARAKI *et al.* (1984). By now, the most comprehensive study of the space-time distribution of field-aligned currents during intensive $B_z > 0$ condition has been carried out by IJIMA *et al.* (1984). Figure 4(a) presents the composite diagram of the spatial distribution of the large-scale field-aligned currents inferred from the data of nine MAGSAT orbits traversing the high latitudes of the southern hemisphere on January 8, 1980. From 1100 to 2300 UT, B_z was stable and equal to ~ 8 nT. Therefore, the consecutive orbits may be treated as a scanning of the steady-state spatial distribution of the field-aligned currents over the polar cap in summer. The large-scale field-aligned currents flowing into the ionosphere in the prenoon sector and out in the afternoon sector can be seen on the polar cap dayside on $\Phi > 80^\circ$. The field-aligned currents at $\Phi < 80^\circ$ belong to zones 1 and 2. Figure 4(b) presents the intensities of magnetic disturbances along the orbits. The near-pole currents are much more intensive than in zones 1 and 2 and reach $\sim 0.8 \times 10^{-6}$ A/m² in the day sector at $\Phi > 83^\circ$. These estimates have been inferred from the satellite magnetograms on the assumption that the currents j_{\parallel} are of the form of infinite "curtains" extended across the orbits. Since in reality the j_{\parallel} configuration may be treated as "curtain" type in only a rough approximation, the real j_{\parallel} may differ from these estimates (KISABETH and ROSTOKER, 1977; DREMUHINA *et al.*, 1985). The comparison of Figs. 4(a) and (b) with Fig. 3(c) shows that the positions, directions, and intensities of the currents are similar. The data of the MAGSAT and TRIAD measurements presented by IJIMA *et al.* (1984) must be treated as excellent evidence for the existence of the $B_z > 0$ component-controlled large-scale field-aligned current system in the near-pole regions. The characteristic features of the system, namely, the location of the most intensive currents on the dayside, flowing into ionosphere in the afternoon sector and out in the prenoon sector, the relationship to the $B_z > 0$ component intensity, were inferred from the observations of the magnetic field variations on the Earth's surface. The results of IJIMA *et al.* (1984) confirm the existence of this $B_z > 0$ current system rather than actually discovering a "new Birkeland current system".

The field-aligned current model displayed in Figs. 3(a)–(d) is identical to the pattern of the field-aligned currents in the dayside cusp region for the northern hemisphere for various IMF directions proposed by TROSHICHEV and GIZLER (1980) and TROSHICHEV (1984). In contrast to the model which gives the field-aligned current direction and intensity for each hour of MLT and for each degree of latitude, the pattern gives but a general idea about the location of the currents by substituting a line for the regions where they are present. A somewhat different pattern of field-aligned current distribution was proposed by PUDOVKIN *et al.* (1984). The main differences are due to the location and direction of the j_{\parallel} controlled by the B_y component. Namely, the field-aligned currents j_{\parallel} are located exclusively in zone 1, their directions are different at prenoon and afternoon hours and do not depend on the B_y sign of the B_y component. Present-day satellite data unambiguously confirm the pattern presented here as reported by TROSHICHEV (1984), WILHJELM *et al.* (1978), DOYLE *et al.* (1981), VANNERSTROM *et al.* (1984).

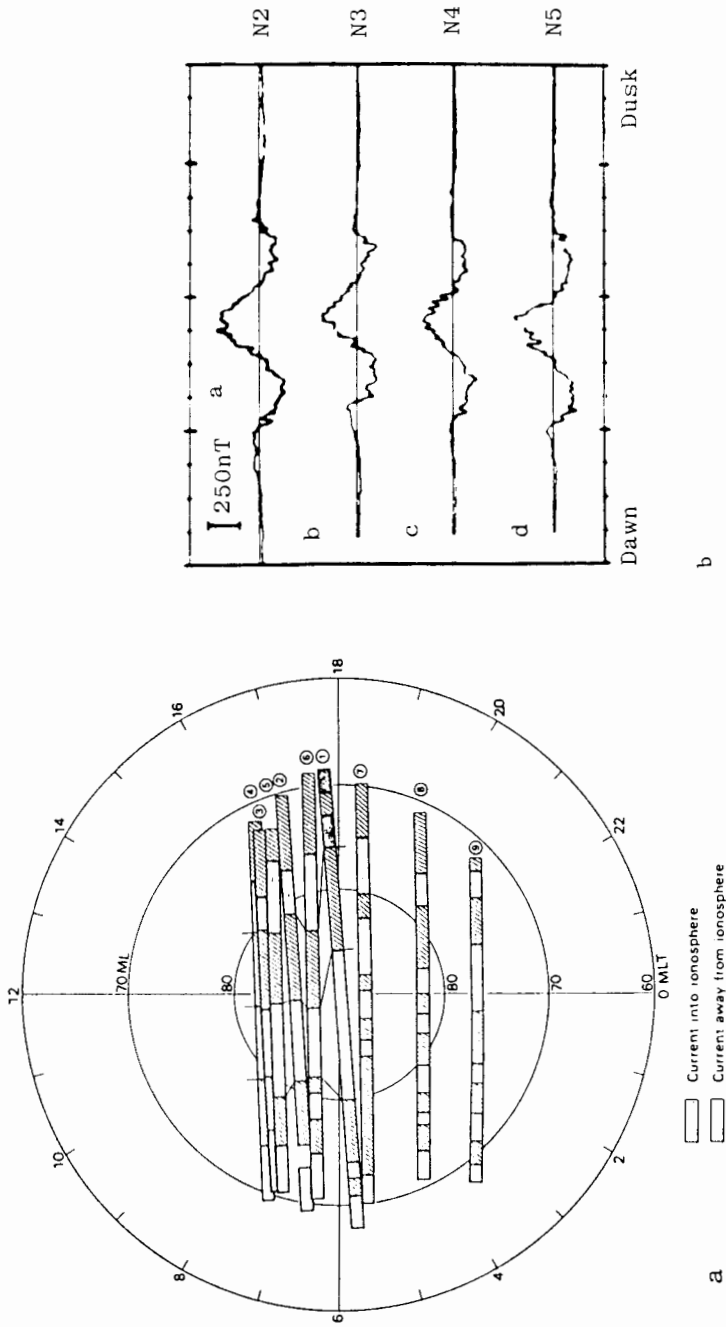


Fig. 4. (a) A composite diagram of spatial distribution of large-scale field-aligned currents during nine MAGSAT orbits across the south polar cap on January 8, 1980 from 1100 to 2300 UT. The B_z component equals ~ 8 nT and is stable throughout the interval. Eccentric geomagnetic dipole latitude-local geomagnetic time coordinates are used. (b) The magnetic disturbance vector transverse to the main geomagnetic field directed to the Sun along orbits 2 through 5 (IITMA *et al.*, 1984).

5. Determination of the Electromagnetic Parameters of the Upper Atmosphere in Terms of a Model for Nonuniform Ionospheric Conductivity

In the case of nonuniform ionospheric conductivity, the electric field potential is determined from the second-order partial differential equation

$$-\frac{1}{\tau} \left[\left(\tau \Sigma_H \phi_\theta \right)_\theta + \frac{1}{\tau} \left(\Sigma_H \phi_\Lambda \right)_\Lambda \right] + \frac{1}{\tau} \left[\left(\Sigma_P \phi_\Lambda \right)_\theta - \left(\Sigma_P \phi_\theta \right)_\Lambda \right] = \frac{1}{\tau} \left(\tau \Psi_\theta \right)_\theta + \frac{1}{\tau^2} \Psi_{\Lambda\Lambda} \quad (4)$$

where θ and Λ are respectively the colatitude and longitude in spherical coordinates; $\tau = \sin \theta$. In (4) Ψ is the current function related unambiguously to the variation of the horizontal magnetic field on the Earth's surface and calculated for various classes of the geomagnetic field variations. In the numerical calculations of ϕ on the basis of (4), the height-integrated ionospheric conductivity tensor is usually specified on a grid with a 1° step in latitude and a 15° step in longitude (1^h of local geomagnetic time). The total integral Hall and Pedersen conductivities are composed of the conductivities resulting from both ionization by solar radiation and particle precipitation. The solar UV input produces height-integrated conductivities that vary as the cosine of the solar zenith angle (METHA, 1979; ROBINSON and VONDRAK, 1984). This conductivity source represents a background upon which the contribution from auroral charged particles should be added (WALLIS and BUDZINSKI, 1981; SPIRO *et al.*, 1982). The models of high-latitude height-integrated ionospheric conductivities proposed elsewhere were discussed and compared in (REIFF, 1984; FELDSTEIN and LEVITIN, 1984). Different conductance models are used as input to solve the equation (4) numerically.

KAMIDE and RICHMOND (1982) calculated electric fields and ionospheric and field-aligned currents on the basis of ground-based magnetic records used with drastically different models for ionospheric conductivity. It has been shown that the ionospheric current patterns are in practice independent of, the field-aligned current patterns are weakly dependent on, and the electric field patterns depend strongly on the choice of the conductivity. Our calculations have proved to be in complete agreement with these conclusions. They also substantiate the use of the uniform ionospheric conductivity model to calculate the summer pattern of the field-aligned currents from the ground-based magnetic variation data in the pioneer studies of such kind.

The application of more realistic conductivity models will little change the j_i distribution shown in Fig. 3. The uniform ionospheric conductivity approximation seems to be sufficient in summer to calculate the estimates of the ionospheric and field-aligned currents from the ground-based data. The experimental data of magnetic and electric fields at low altitudes do not contradict this conclusion (WILHJELM *et al.*, 1978; PRIMDAHL and SPANGSLEV, 1977; PRIMDAHL *et al.*, 1979). The analysis of the combined measurements of magnetic and electric fields from the low-altitude satellites suggests that the summer ionospheric conductivity may be treated as nearly uniform. In the following sections we describe the results of model calculations of the electrostatic potential and the electric fields from the magnetic field variations on the Earth's surface and the inhomogeneous ionospheric conductivity models. As a rule, for the corpuscular part we use the WALLIS and BUDZINSKI (1981)

model. These calculations fail to introduce essential corrections in the results for all of summer and part of equinox seasons. In winter, when solar UV induced conductivity is practically absent, the situation changes. The necessity arises to have more accurate information on the high-latitude distribution of the corpuscular conductivity which is extremely variable and may differ noticeably in the particular cases from the mean model distributions.

6. The Electrostatic Potential and Convection in the High-Latitude Ionosphere as a Function of Interplanetary Conditions

6.1 The main features of the spatial distribution of $\phi^M(\Phi, \text{MLT})$ in different seasons

Numerical model calculations (AFONINA *et al.*, 1983; FELDSTEIN *et al.*, 1984a, b; LEVITIN *et al.*, 1984), of the potential ϕ^M at the observation point (Φ, MLT) can be represented by

$$\phi^M(\Phi, \text{MLT}, B_y, B_z) = \phi_y^M(\Phi, \text{MLT})B_y + \phi_z^M(\Phi, \text{MLT})B_z + \phi_0^M(\Phi, \text{MLT}) \quad (5)$$

where B_y and B_z are the hourly means of the IMF vector components at the moment when ϕ is determined. The model contains different sets of the coefficients $\phi_y^M, \phi_z^M, \phi_0^M$ relevant to the calculations for the $B_z \geq 0$ component (use is made of the WALLIS and BUDZINSKI (1981) model for the height-integrated ionospheric conductivity at $K_p < 3$) and at the $B_z < 0$ component (use is made of the height-integrated conductivity model at $K_p > 3$). The distributions of $\phi_y^M(\Phi, \text{MLT})$, $\phi_z^M(\Phi, \text{MLT})$, and $\phi_0^M(\Phi, \text{MLT})$ show how the electric field in the high-latitude ionosphere responds to the IMF components $B_y = \pm 1$ nT, $B_z = \pm 1$ nT, and $B_z = B_y = 0$, respectively. As a result we can obtain the distribution of $\phi^M(\Phi, \text{MLT}, B_y, B_z)$ under a particular IMF situations.

Figure 5 shows ϕ_y^M, ϕ_z^M and ϕ_0^M in the Φ -MLT coordinate. The isolines present the equal values of the electrostatic potential in kV. Each panel shows the IMF conditions situations, the season, the universal time for calculating the distribution of the height-integrated ionospheric conductivity due to solar UV radiation from the Sun (UT_m corresponds to the coincidence between the geomagnetic and geographic axes), the K_p -index characterizing the high-latitude ionospheric conductivity due to the corpuscular precipitations in terms of the WALLIS-BUDZINSKI (1981) model.

These "elementary" potential distributions are characterized by the following features. Two convective vortices exist in summer when $B_z = B_y = 0$, i.e., in case of quasiviscous interaction of the solar wind plasma with the magnetosphere (the ϕ^M contour lines represent convective flow paths of magnetospheric plasma with $\phi^M > 0$ characterizing the counterclockwise convection). The centers of the vortices are located at $\Phi \sim 77^\circ$ for 0300–0400 MLT (the dawn vortex) and at $\Phi \sim 80^\circ$ for 1600–1700 MLT (the dusk vortex). The potential is 12.6 kV at the dawn vortex center and 16.8 kV at the dusk vortex center. In its spatial dimensions the dusk vortex exceeds the dawn vortex.

The B_y component produces a single convective vortex centered about the corrected geomagnetic pole. The convection direction in the vortex depends on the sign of the IMF azimuthal component namely for $B_y > 0$ ($B_y < 0$) the convection in the northern hemisphere is clockwise (counterclockwise). The inverse relationship between the B_y and convection directions is observed in the southern hemisphere.

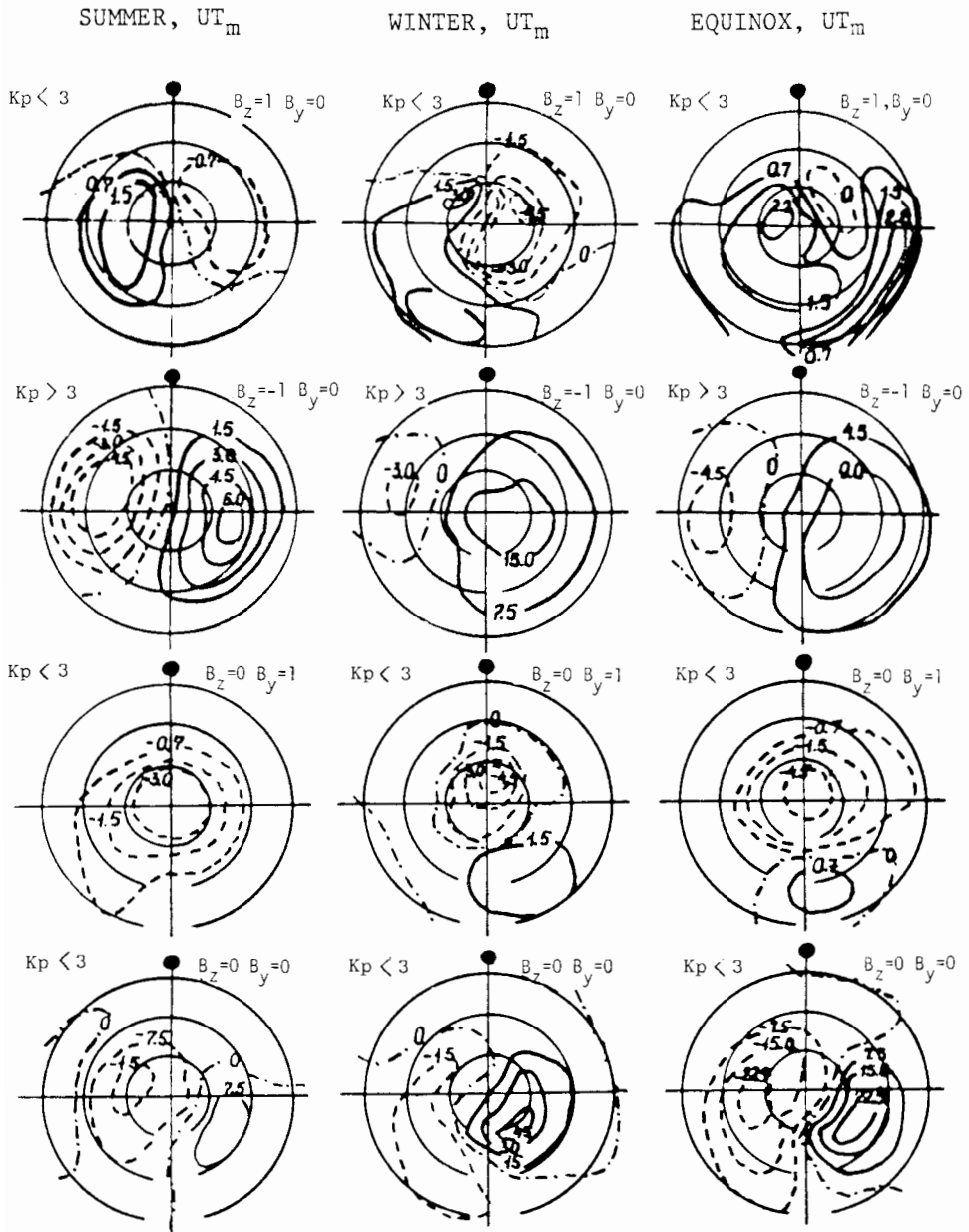


Fig. 5. The isolines of the electrostatic potential ϕ^M in kV during three seasons in various interplanetary medium conditions. The geographic axis coincides with the geomagnetic axis (UT_m). The corrected geomagnetic latitude 80°, 70° and 60°-local geomagnetic time coordinates are used (LEVITIN *et al.*, 1984).

Like the quasiviscous interaction, the southward IMF vector component produces two convective vortices of the same direction. The $B_z < 0$ condition enhances both dawn and dusk vortices which are permanently present in the

high-latitude ionosphere due to the continuous interaction of the solar wind with the magnetosphere.

In summer, the center of the dawn vortex produced by the $B_z = -1$ nT component is located at $\Phi \sim 75^\circ$ for ~ 0500 MLT; where $\phi^M \sim 7.5$ kV. The dusk vortex center is located at $\Phi \sim 75^\circ$ for ~ 1700 MLT; where $\phi^M \sim -6$ kV. In this case the spatial dimensions of the dawn vortex exceed those of the dusk vortex.

The northward IMF vector component produces two convective vortices with the convection directions opposite to those in the vortices produced by the viscous interaction and the $B_z < 0$ component. In their dimensions, these vortices are much smaller and are located in the dayside polar cap sector. In summer the dawn vortex center is located at $\Phi \sim 82^\circ$ for ~ 0900 MLT (for where $\phi^M \sim -1.5$ kV), and the dusk vortex center at $\Phi \sim 82^\circ$ for ~ 1500 MLT ($\phi^M \sim 1.8$ kV).

In winter, the vortices are rearranged. At $B_y = B_z = 0$, the dawn vortex center is located at $\Phi \sim 80^\circ$ for ~ 0200 – 0300 MLT (where $\phi^M \sim 54$ kV), and the dusk vortex center is at $\Phi \sim 72^\circ$ for ~ 2000 MLT ($\phi^M \sim -21$ kV). The dawn vortex produced by the $B_z < 0$ component is spatially larger than of the dusk vortex and covers almost the entire high-latitude region (the maximum value of ϕ^M is ~ 21 kV). The B_y component-produced main vortex gets narrow, and an additional vortex appears at night hours. At the equinoxes, the distributions of ϕ_y^M , ϕ_z^M and ϕ_0^M are intermediate between summer and winter.

The representation of the ionospheric convection pattern by the sum of elementary cells coincide with the three types of magnetospheric convection cells suggested by BURCH *et al.* (1985), and REIFF and BURCH (1985), namely, the merging cells ($B_z < 0$ in our case, see Fig. 5(b)), the viscous cells (at $B_z = B_y = 0$ in our case, see Fig. 5(d)) and the lobe cells (controlled by B_y in our case, see Fig. 5(c), and by $B_z > 0$, see Fig. 5(a)). In Fig. 5, however, the ionospheric manifestation of the viscous cells cover also exist in the polar cap, while the lobe cells are extended to the auroral zone. Therefore, a more significant overlapping of the magnetospheric cells may occur. As a result, the ionospheric systems are somewhat different than the magnetospheric systems predicted by BURCH *et al.* (1985) and REIFF and BURCH (1985). However, the coexistence of three types of convection cells as elements of single convection model (BURCH *et al.*, 1985) is completely in agreement with our model.

Under the real conditions, a particular $\phi^M(\Phi, \text{MLT})$ pattern depends on the IMF orientation and may be much more complicated. To illustrate the total effect of the B_y and B_z components on the potential distribution pattern. Figure 6 shows the summer distributions of $\phi^M(\Phi, \text{MLT})$ for $B_z = 0$ and $B_y = \pm 6$ nT, $B_y = 0$ and $B_z = \pm 4$ nT, $B_z = -4$ nT and $B_y = \pm 6$ nT.

The model distributions of ϕ_y^M , ϕ_z^M and ϕ_0^M make it possible to estimate the potential difference, $\Delta\phi$, across the polar cap. In the situation with $B_y = B_z = 0$ (the case of quasiviscous interaction), $\Delta\phi_0^M \approx 29$ kV but varies somewhat depending on the conductivity model applied. According to AFONINA *et al.* (1979) and SERGEEV and KUZNETSOV (1981), the value of $\Delta\phi_0$ is proportional to the square of the solar wind velocity. When the southward IMF increases by 1 nT the value of $\Delta\phi^M$ is 13.5 kV, and when the northward IMF increases by 1 nT the potential difference is 3.3 kV and is directed inversely from dusk to dawn and appears in the polar cap day sector. TROSHICHEV (1984) obtained approximately the same value of $\Delta\phi^M$ on the assumption of uniform height-integrated ionospheric conductivity in summer with

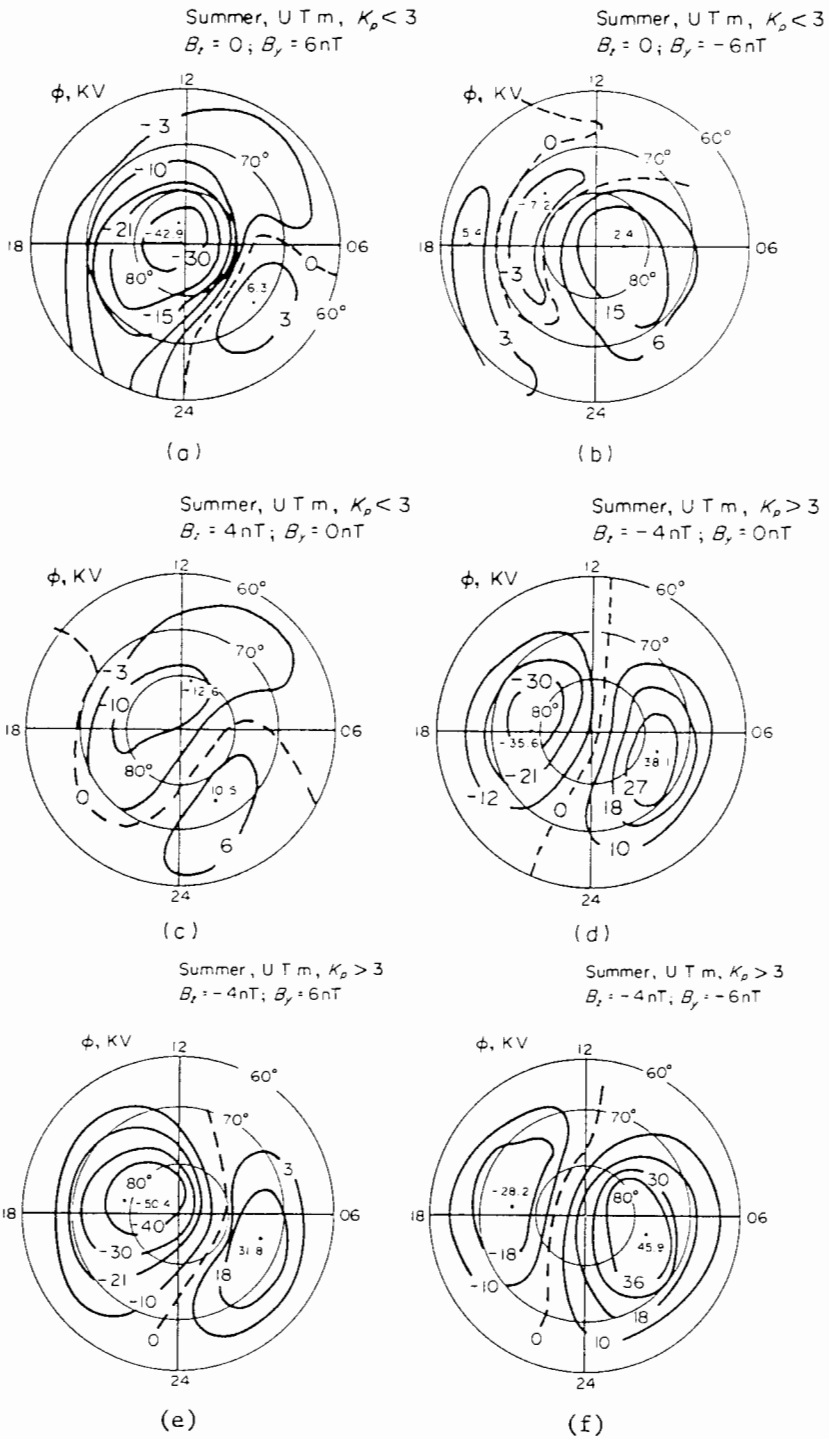


Fig. 6. The isolines of the electrostatic potential ϕ^M in kV in summer for various interplanetary conditions (FELDSTEIN *et al.*, 1984a).

$\Sigma_H = 10$. The model value of $\Delta\phi_0^M \approx 29$ kV in the $B_z = B_y = 0$ situation agrees with the results of the direct measurements over the polar cap from the low-orbiting satellites AE-C and D (REIFF *et al.*, 1981), S3-2 (DOYLE and BURKE, 1983), and S3-3 (WYGANT *et al.*, 1983). They report a ~ 30 kV IMF-independent potential difference over the polar cap. However, this value is much larger than the $\Delta\phi$ value measured by MOZER (1984) in the magnetospheric boundary region which was ~ 5 kV in case of viscous interaction. Further satellite and radar measurements to be taken to establish the ultimate relation between the electric field induced by viscous interaction and merging of the IMF and the magnetospheric magnetic field on the magnetopause and the value of $\Delta\phi$ across the polar cap.

LYATSKY (1978), CROOKER (1979), COWLEY (1982, 1984), and TROSHICHEV (1984) have analyzed the IMF effect on the high-latitude convection. The B_z component controls the merging at the magnetopause giving rise to a decrease or increase of the potential difference across the polar cap. The B_y component gives rise to an increase of the convection intensity in the dawn (dusk) sector of the northern hemisphere when $B_y > 0$ ($B_y < 0$). This feature of the convection behaviour was first noted by HEPNER (1972) and confirmed afterward by numerous studies including JØRGENSEN (1981) and CLAUER *et al.* (1984). GALPERIN *et al.* (1978) have demonstrated the dependence of the convection direction in the dayside cusp on the B_y component sign. The possible effects of the IMF components on the magnetospheric convection were discussed by STERN (1983), COWLEY (1981a, 1982), and ALEXEEV and BELENKAYA (1983a, b) using a model for the superposition of a homogeneous IMF on the dipolar geomagnetic field. Despite its simplicity, the models account for numerous fine scale features of the convective motion distribution in the magnetosphere. It should be noted that the convection model at $B_z > 0$ presented in Fig. 6 differs from the analogous schemes proposed by CROOKER (1979), TROSHICHEV (1984), and others.

The model calculations have shown that the concept of narrow (± 1 hour near noon) throat proposed in (HEELIS, 1984 and the references therein) for the convection pattern on the polar cap day side fails to be reflected in the $\phi^M(\Phi, \text{MLT})$ distributions inferred from the ground-based magnetic observations.

The transition from sunward convection in the dawn and dusk sectors to the antisunward pattern in the polar cap takes the form of a rotational reversal. The change from the sunward velocity direction to antisunward in Fig. 6 is due to the rotation of the convection velocity vector with a moderate increase in velocity. These results agree with the radar experiments (EVANS *et al.*, 1980; FOSTER *et al.*, 1981; JØRGENSEN *et al.*, 1984) which show that the convection in the polar cap occurs over a broad time interval from 0800 to 1600 MLT. The measurements (FOSTER *et al.*, 1985) have shown that plasma enters the polar latitudes across a rotational convection reversal which span at least 3 hours of local time in the noon sector. Something like a throat, but much more extended than ± 1 hour near noon, appears at high values of B_y and $B_z < 0$ (FELDSTEIN *et al.*, 1984a).

6.2 Intercomparison of the ϕ^M patterns inferred from the ground-based magnetic field variations

It is presently possible to compare two independent calculations of ϕ^M using the magnetogram-inversion schemes. FELDSTEIN *et al.* (1984a) used hourly data from the

standard high-latitude observatories for 1968, the FAERMARK (1980) algorithm, and the METHA (1979) and WALLIS-BUDZINSKI (1981) models for height-integrated ionospheric conductivity. FRIIS-CHRISTENSEN *et al.* (1985) used the 20-min data from the chain of magnetometers along the west coast of Greenland for the summer seasons of 1972 and 1973, the KAMIDE *et al.* (1981) algorithm, and the height-integrated ionospheric conductivity models similar to those used by KAMIDE and MATSUSHITA (1979). In both cases, the magnetic field variations on the Earth's surface controlled by the B_z and B_y components and independent of IMF were separated using the same method.

Figure 7 presents the results of FRIIS-CHRISTENSEN *et al.* (1985) for the various IMF orientations. The ϕ (Φ , MLT) patterns for $B_z = B_y = 0$ is shown in the center. Eight other patterns for various combinations of B_z and B_y are displayed in the B_y - B_z coordinate system, with $\sqrt{B_z^2 + B_y^2} = 5$ nT. Although the calculations were made using different algorithms and different input data, the general features of the potential distribution proved with several exception turnings to be in a good agreement with FELDSTEIN *et al.* (1984a) shown in Fig. 6.

The two-cell convection pattern has been obtained for the $B_z = B_y = 0$ condition with the zero potential isoline near the geomagnetic pole and a ~ 30 – 35 kV potential difference across the polar cap. The $B_z > 0$, $B_y = 0$ condition is characterized by a decrease of the dawn vortex, by a shift of its focus near to midnight, and by a clockwise rotation of the distribution as a whole. The two-cell convection pattern is preserved, although the electric field direction may change from dawn-dusk to dusk-dawn in a limited latitude region of the day sector. In Fig. 7 it is seen from the isolines character in the day sector at $\Phi \sim 85^\circ$, and in Fig. 6 isoline surge is absent due to the spatial smoothing. However, the two studies do not confirm the occurrence of the four vortex convection system in high latitudes during the $B_z > 0$ intervals as proposed by BURKE *et al.* (1979). The $B_y > 0$ ($B_y < 0$), $B_z = 0$ situations are accompanied by an enhanced convection in the dawn (dusk) sectors.

KAMIDE *et al.* (1983) and MISHIN *et al.* (1984) calculated the high-latitude potential distributions using their magnetogram inversion techniques. The magnetic field variations at 107 magnetic observatories during the CDAW-6 period for March 22, 1979 were used as input data. WOLF and SPIRO (1984) have found a reasonably good agreement between the results of the two studies by comparing visually between the potential distribution plots at 10.30 UT, 11.10 UT and 11.40 UT on March 22, 1979. Figure 8 is a quantitative comparison between the distributions of ϕ^M along the meridional cross-section separated by three hours of MLT. The respective KAMIDE *et al.* (1983) meridional cross-sections were averaged for the 10 minutes which fell within the 30–45 minutes intervals used by MISHIN *et al.* (1984). Noticeable discrepancies exist during all the substorm phases in the character of the latitude variations of ϕ^M along fixed meridian. Since the values of magnetic field variations used as input data were identical, the discrepancies must be unexpected. They cannot be ascribed to differences in the conductivity models used in calculations. The allowance for the difference in the boundary conditions also fails to significantly improve the agreement between the results of the two groups of researchers. It is desirable to find the reasons for the disagreement, which is necessary for further perfection of the method of potential restoration from the ground-based magnetograms. BAKER and KAMIDE (1985) also revealed some discrepancies for the CDAW-6 period at the comparison of

ELECTRIC POTENTIAL

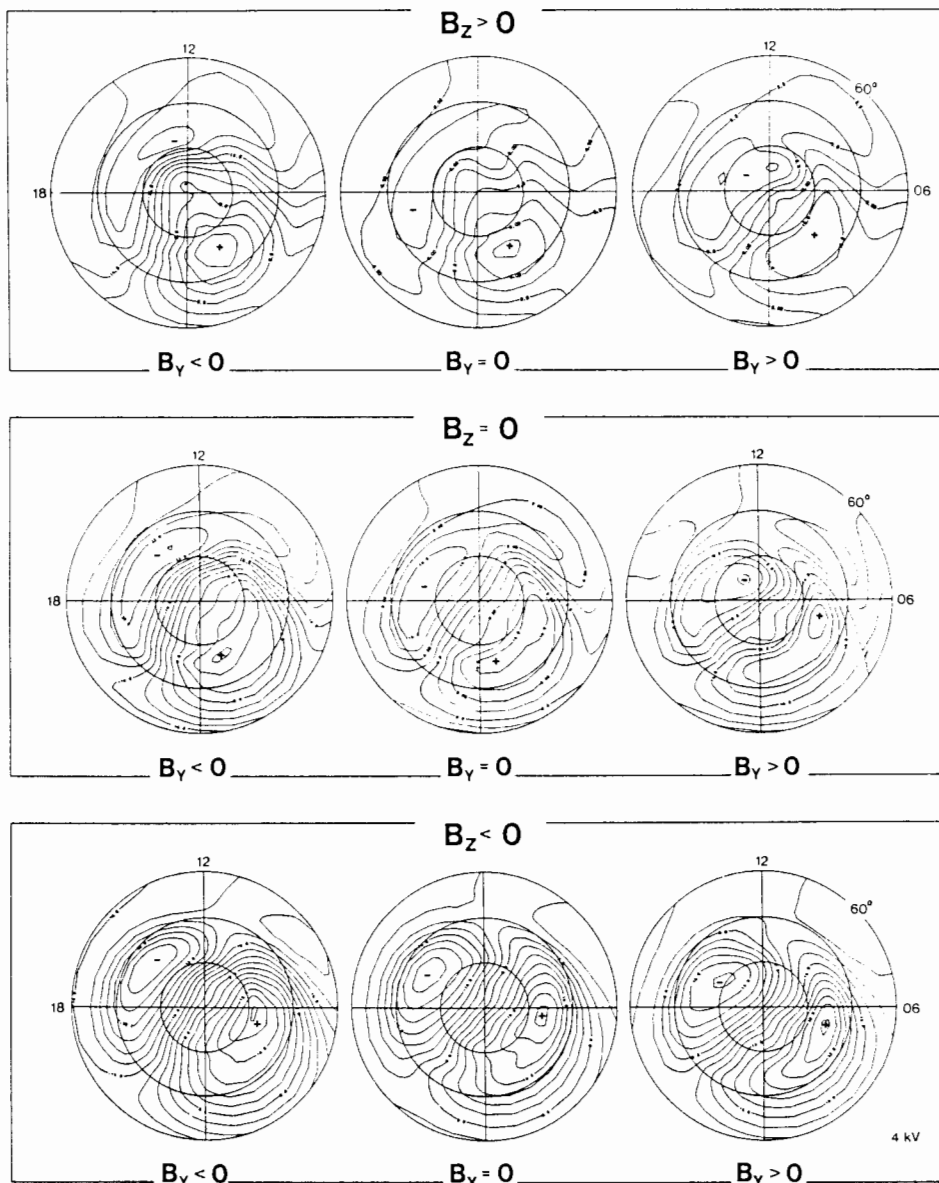


Fig. 7. The electrostatic potential distribution for various IMF conditions from FRIIS-CHRISTENSEN *et al.* (1985). The isolines are spaced 4 kV apart. The 0 kV isoline runs through the pole.

radar electric fields with the KRM output.

The electric field model obtained by statistical processing of the incoherent radar data (FOSTER, 1983) was scaled to the model for the potential ϕ (Φ , MLT) at northern-hemisphere auroral latitudes (FOSTER, 1984). Figure 9 shows the potential

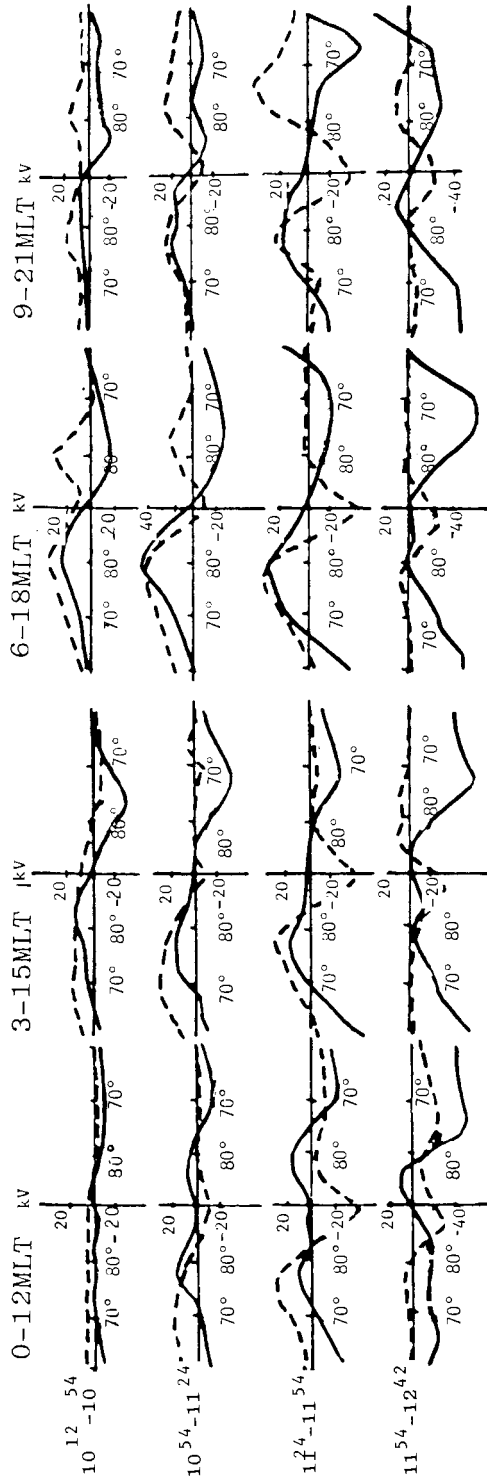


Fig. 8. Comparison between the meridional profiles of the electrostatic potential calculated for individual intervals of the March 22, 1979 magnetic disturbance by the magnetogram inversion technique (KAMIDE *et al.*, 1983, the solid lines; MISHIN *et al.*, 1983, the dashed lines). The horizontal lines characterize ϕ^M during the 1012-1052, 1054-1124, 1124-1154, 1154-1242 UT intervals. The vertical columns are for the meridional profiles at 0000-1200, 0300-1500, 0600-1800, and 0900-2100 MLT with the start time on the left.

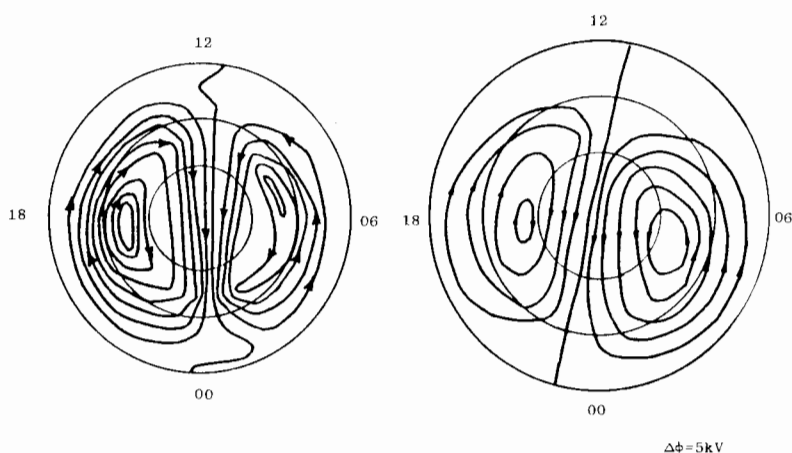


Fig. 9. The space-time distribution of the electrostatic potential in summer in terms of the FOSTER (1984) model (on the left) and the LEVITIN *et al.* (1984) model (on the right) in the $B_z = -2$ nT, $B_y = -3$ nT situation. The isolines are spaced 5 kV apart.

isolines in this model (at the left) and in the model LEVITIN *et al.* (1984) (at the right) for the IMF situation with $B_z = -2$ nT, $B_y = -3$ nT. The results of comparing between the models indicate that the average potential distribution restored from the geomagnetic data reflects the global features of the electric field potential ϕ (Φ , MLT) inferred from direct measurements.

HEELIS (1984), POTEIRA *et al.* (1984), and REIFF and BURCH (1985) have proposed the models of high-latitude plasma convection dependent on the B_y component. HEELIS (1984) has analyzed the convection distribution along the AE-C orbits and constructed a detailed model for the convection on the day side of the polar cap for $B_z < 0$ and various B_y directions. Figure 10 shows the vectors of the convection velocity V_d along the satellite trajectories (to the left) and the model V_d (Φ , MLT, B_y , B_z) distributions calculated in terms of the model LEVITIN *et al.* (1984). The model describes adequately the large-scale features of the experimental data, in particular the changes of the convection directions in the respective IMF situations. It may be expected, therefore, that the model is close to reality throughout the high-latitude region. Figure 11 shows the model convection distributions (HEELIS, 1984) for the dayside hemisphere at $\Phi \geq 60^\circ$ to the left and the convection velocity vector distributions in terms of the electric field model (LEVITIN *et al.*, 1984) for the same region (at the right). The values $B_z < 0$ and $B_y \geq 0$, $B_y \leq 0$ adopted by Heelis correspond to $B_z = -2$ nT, $B_y = \pm 3$ nT, $B_y = \pm 10$ nT in the LEVITIN *et al.* (1984) model. In the $B_y = +10$ nT situation, plasma flows to the near-pole region in the dawn sector at an increased velocity at $\Phi \sim 80^\circ$ HEELIS *et al.* (1976) called the throat. In the $B_y = -10$ nT IMF situation, the encounter of the sunward convective streams occurs near noon at $65^\circ < \Phi < 75^\circ$ and the plasma flows into the near-pole region at the highest velocity through the dusk sector. Thus, the plasma flows into the near pole region mainly through the dawn (dusk) sector at $B_y > 0$ ($B_y < 0$). However, the flux change over from sunward to antisunward direction is different in the HEELIS (1984) and LEVITIN *et al.* (1984) models. Perhaps the local convective whirlings seen in Fig.

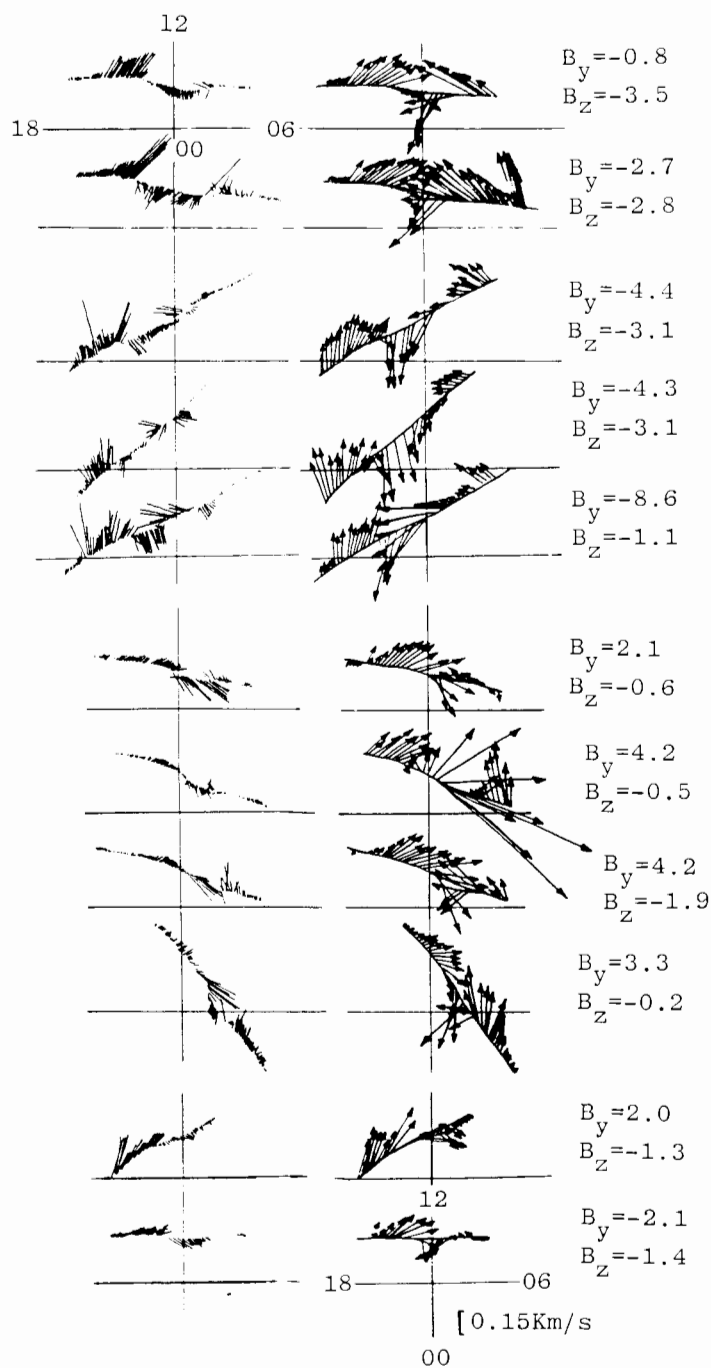


Fig. 10. The distribution of the convection velocity V_d along the AE-C orbits. The experimental data (HEELIS, 1984) are shown on the left and the model data (LEVITIN *et al.*, 1984) on the right for the IMF situations indicated.

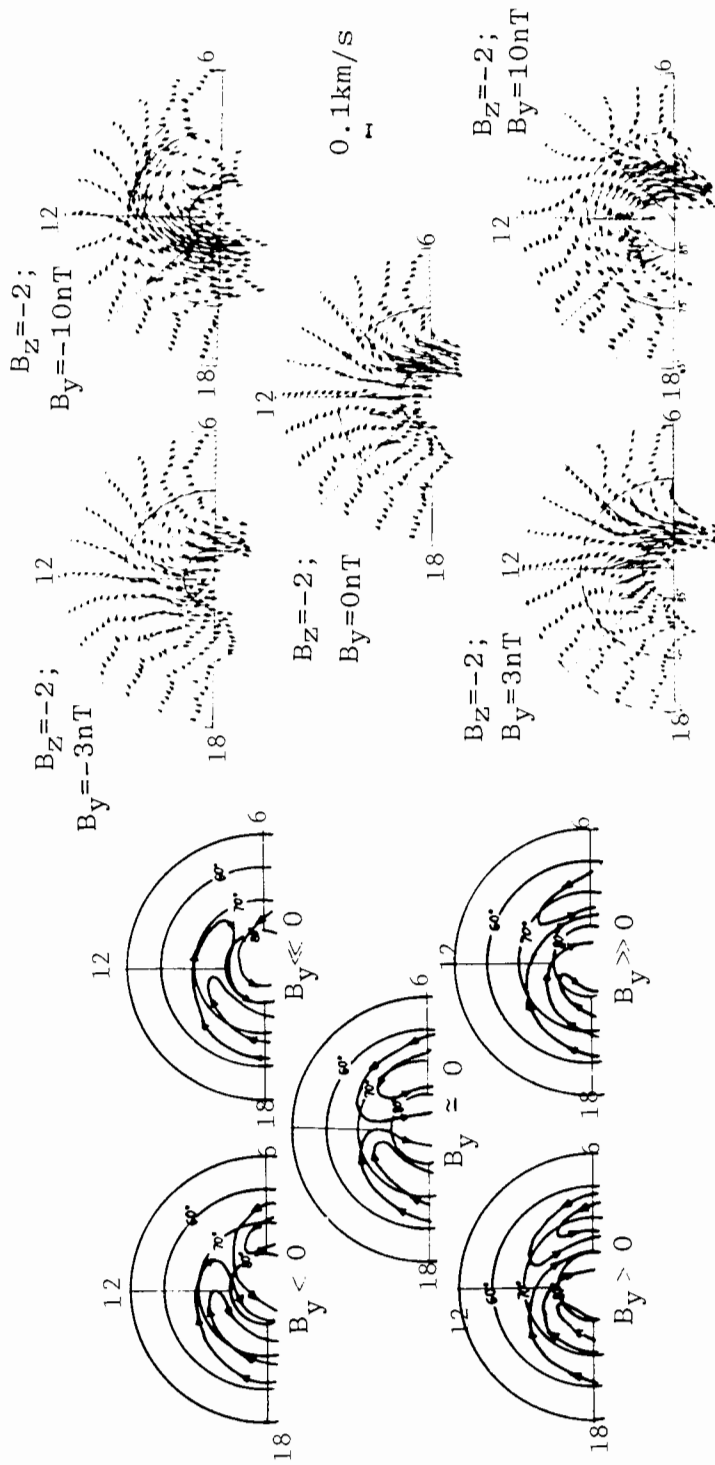


Fig. 11. The model distributions of convection in various interplanetary medium conditions (According to HFFLIS (1984) on the left and according to LEVITIN *et al.* (1984) on the right).

II at the right in higher latitudes may be interpreted in terms of the throat, namely at $B_z = 10$ nT, 0600–0800 MLT on $\Phi \sim 75\text{--}78^\circ$ and at $B_z = -10$ nT, 0600–0800 MLT at $\Phi \sim 75\text{--}83^\circ$.

The presence of the inverse (sunward) convection in the polar cap during the $B_z > 0$ intervals, which was noted first in the geomagnetic data (MAEZAWA, 1976; KUZNETSOV and TROSHICHEV, 1977; and many other works) and then in the satellite data (BURKE *et al.*, 1979), has motivated researchers to set forth the hypothesis that four convective vortices exist in the high-latitude ionosphere during such intervals (BURKE *et al.*, 1979; REZHENOV *et al.*, 1980; REZHENOV, 1981; AKASOFU and ROEDERER, 1983; TROSHICHEV, 1984; REIFF and BURCH, 1985 and the references therein).

The model (LEVITIN *et al.*, 1984) predicts a complicated spatial pattern of convection depending on the IMF which is considerably different from such idealized configuration probably because the satellite observations yield true information on the convection distribution only along satellite orbits. The extrapolation to the entire high-latitude region depends in a certain manner on the researcher's intuition. Therefore, such extrapolation should be treated with caution.

Figure 12 shows the isolines of the potential $\phi(\Phi, \text{MLT})$ for IMF conditions with $B_z = 6, 8, 15$ nT and $B_y = 0$ in accordance with the LEVITIN *et al.* (1984) model. These distributions demonstrate that even for large positive values of B_z the region with the dusk-dawn electric field direction covers but a limited sector of the dayside polar cap and does not extend to the night side. An interesting fact should be noted, namely the clockwise potential rotation of the potential pattern with increasing B_z while the conductivity remains constant. The convection pattern which is symmetric with respect to the noon-midnight meridian at $B_z < 0$ turns to the dawn-dusk meridian for $B_z \approx 8$ nT. The same rotation was described by YASUHARA *et al.* (1983) relevant to the conductivity variations in the auroral oval.

The effect of B_y IMF on high-latitude convection during northward IMF was discussed in POTEIRA *et al.* (1984), REIFF and BURCH (1985) and the references therein. Figure 13 shows the appropriate schematic convection diagrams according to Potemra *et al.* (the upper row), Reiff and Burch (the middle row) and Levitin *et al.* (the lower row) for the $B_z = 10$ nT, $B_y = 0, \pm 10$ nT situations. Significant differences in the proposed convection models can be seen. We expect that the recent satellite and radar observations will make it clear to what extent the proposed models are able to simulate the real situation.

7. The Dependence of the Electric Field in the High-Latitude Ionosphere on Interplanetary Parameters

Numerical model calculations (AFONINA *et al.*, 1983; LEVITIN *et al.*, 1984) of the electric field E (the meridional component is positive directed towards the corrected geomagnetic pole, the azimuthal component is positive directed towards the magnetic east) at the observation point (Φ, MLT) can be represented by

$$E(\Phi, \text{MLT}, B_y, B_z) = E_y(\Phi, \text{MLT})B_y + E_z(\Phi, \text{MLT})B_z + E_0(\Phi, \text{MLT}) \quad (6)$$

and is determined by the structure of the potential $\phi(\Phi, \text{MLT}, B_y, B_z)$ which is

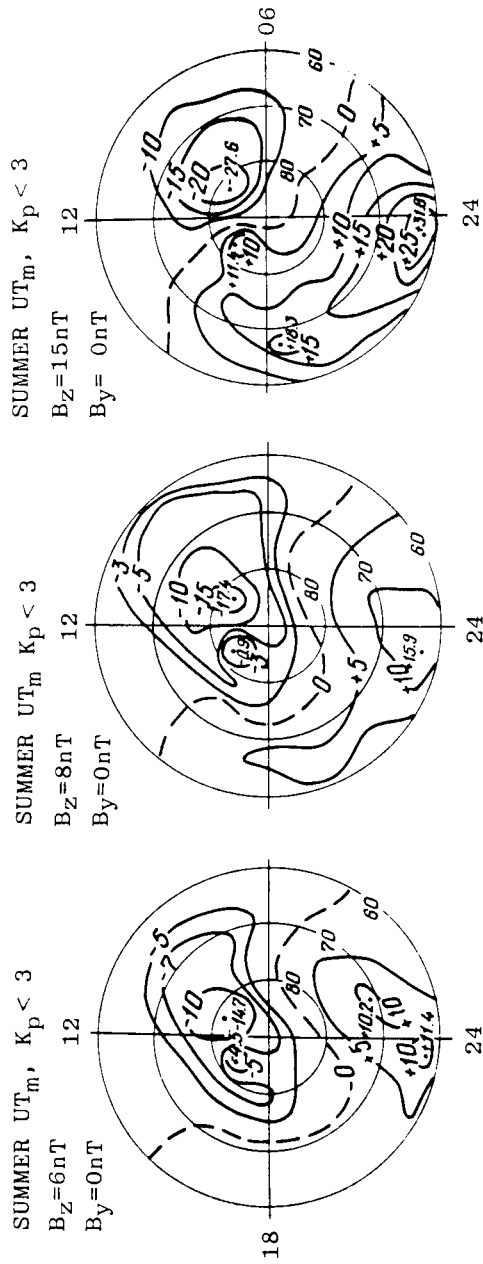


Fig. 12. The model (LAVITTIN *et al.*, 1984) distributions of the potential isolines at various values of the $B_z > 0$ component. (a) $B_z = 6 \text{ nT}$, $B_y = 0$; (b) $B_z = 8 \text{ nT}$, $B_y = 0$; (c) $B_z = 15 \text{ nT}$, $B_y = 0$.

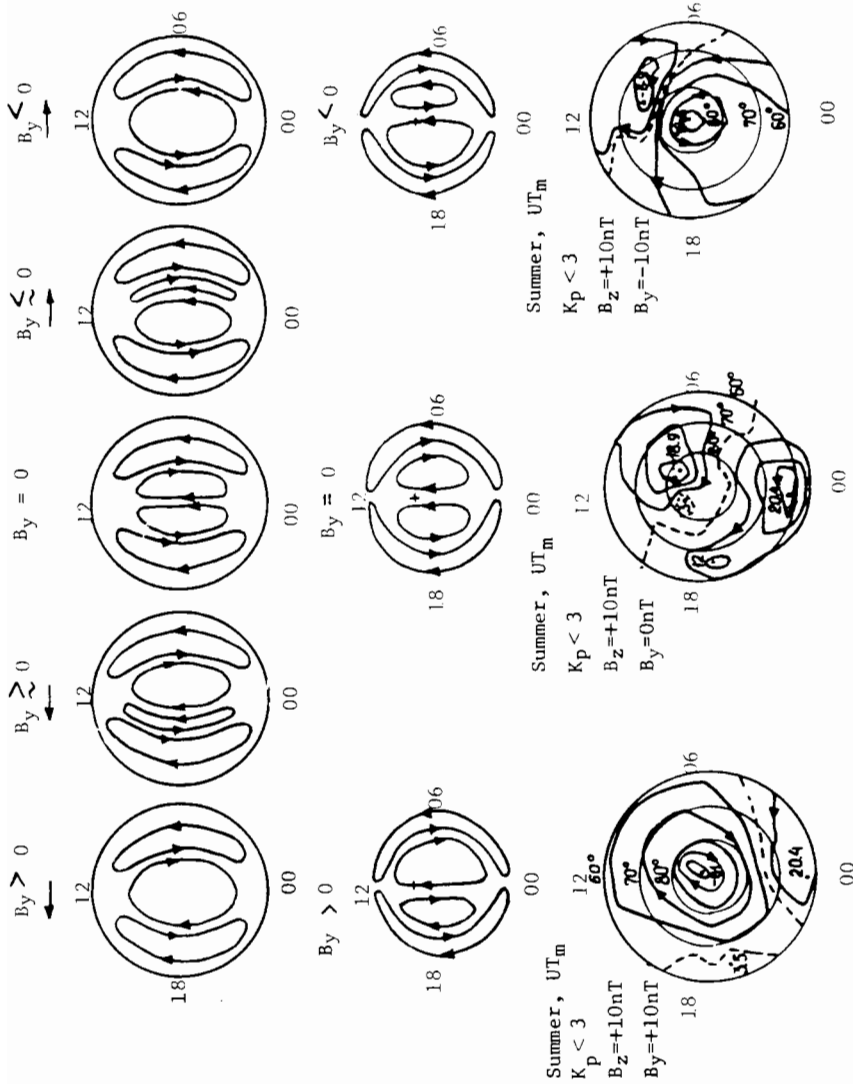


Fig. 13. The schematic diagram of magnetospheric convection in the north polar region for positive values of B_y . The azimuthal component B_z ranges from strongly positive values (at the left) to zero values (at the center) and strongly negative values (at the right). The upper row is from POTTEMRA *et al.* (1984), the middle row is from REIFF and BURCH (1985), and the lower row is from LEVITIN *et al.* (1984).

associated with \mathbf{E} by the relation $\mathbf{E} = -\nabla\phi$.

The distributions of \mathbf{E}_y , \mathbf{E}_z and \mathbf{E}_0 are determined by the respective distributions of ϕ_1^M , ϕ_2^M and ϕ_0^M and reflect the response of the electric field in the high-latitude ionosphere to the variations of $B_y = 1$ nT, $B_z = \pm 1$ nT, and $B_z = B_y = 0$.

The model distributions of \mathbf{E}_1 , \mathbf{E}_2 , \mathbf{E}_0 predict a complicated pattern of the spatial structure of the total electric field which can only describe the general electric field distribution due to the assumptions adopted (for example, VOLIAND 1975a; KAWASAKI, 1975; STERN, 1977; IVANOV and EVDOKIMOVA, 1975).

This circumstance should be borne in mind when analyzing the satellite data. The comparison between the results obtained on close, but not identical, orbits through the $\Phi > 60^\circ$ region should be made very cautiously. Sufficiently pronounced spatial variations of the electric field must also be taken into account when interpreting the complex experiments, like those of HEPNER *et al.* (1971) when the electromagnetic measurements are taken simultaneously but the Earth observations point spaced several degrees of latitude apart (FELDSTEIN *et al.*, 1984a).

7.1 Comparison of the model calculations of $\mathbf{E}(\Phi, MLT)$ with the HEPNER (1972) observations on the dawn-dusk meridian

The OGO-6 observations were used in HEPNER (1972; 1977) to obtain the typical configurations (labelled A, B, C, D, E, F, G, H, I, K, SC, RC) of the meridional electric field component in the northern high-latitude region in summer along the dawn-dusk meridian. Figure 14 presents the latitudinal profiles of the meridional component E_m on the dawn-dusk meridian in summer (a) and winter (b). In all the profiles, the positive values of E_m correspond to the dawn-to-dusk field. Each figure shows the $E_{mz}(\Phi)$ distributions for $B_z > 0$ and $B_z < 0$ (above), $E_{m0}(\Phi)$ for $B_z = B_y = 0$ (in the middle), and $E_{my}(\Phi)$ for $B_y = 1$ nT, $B_z < 0$ and $B_y = -1$ nT, $B_z > 0$. In the LEVITIN *et al.* (1984) model, $E_{my}(B_z < 0) = -E_{my}(B_z > 0)$ and, therefore, two plots at the bottom of the figure are sufficient for all the possible situations with $B_z \geq 0$, $B_y \geq 0$ to be described. In the $B_z = B_y = 0$ IMF condition, as follows from Fig. 14, the polar cap electric field is directed from dawn to dusk. In the auroral zone, the field sign reverses. The permanent existence of the dawn to dusk field and its relatively high intensity compared with the rest of the terms in Eq. (6) provide for the predominant dawn-to-dusk direction of E_m in the polar cap. However, this electric field is inhomogeneous (is stronger in the dawn sector). Correspondingly, Heppner's configuration B must be observed not only at $B_z = B_y = 0$, but also in case of low B_z and B_y values.

The electric field $E_{mz}(B_z < 0)$ configuration is of the Heppner type A. When that field is added to the field E_{m0} the field $E_{mz}(B_z < 0)$ gives rise to an increased field intensity in the polar cap and in the auroral zone. The electric field controlled by the northward IMF vector component $E_{mz}(B_z > 0)$ composed with E_{m0} will, on the contrary, give rise to decreased field intensity in the polar cap and in the auroral zone. This latter condition may yield the Heppner type I configuration with high-intensity $B_z > 0$ fields.

The electric field $E_{my}(B_y > 0)$ gives rise to a decreased field E_{m0} in the dusk sector and to an increased E_{m0} in the dawn sector. The direction of the field $E_{my}(B_y > 0)$ is the same as the $E_{mz}(B_z > 0)$ direction in the dusk sector on $\Phi > 84^\circ$ and is opposite in the dawn sector on $\Phi < 87^\circ$. The E_{my} distributions at $B_z > 0$ and $B_z < 0$ differ little from each other. Thus, the model distributions shown in Fig. 14 may account for all the

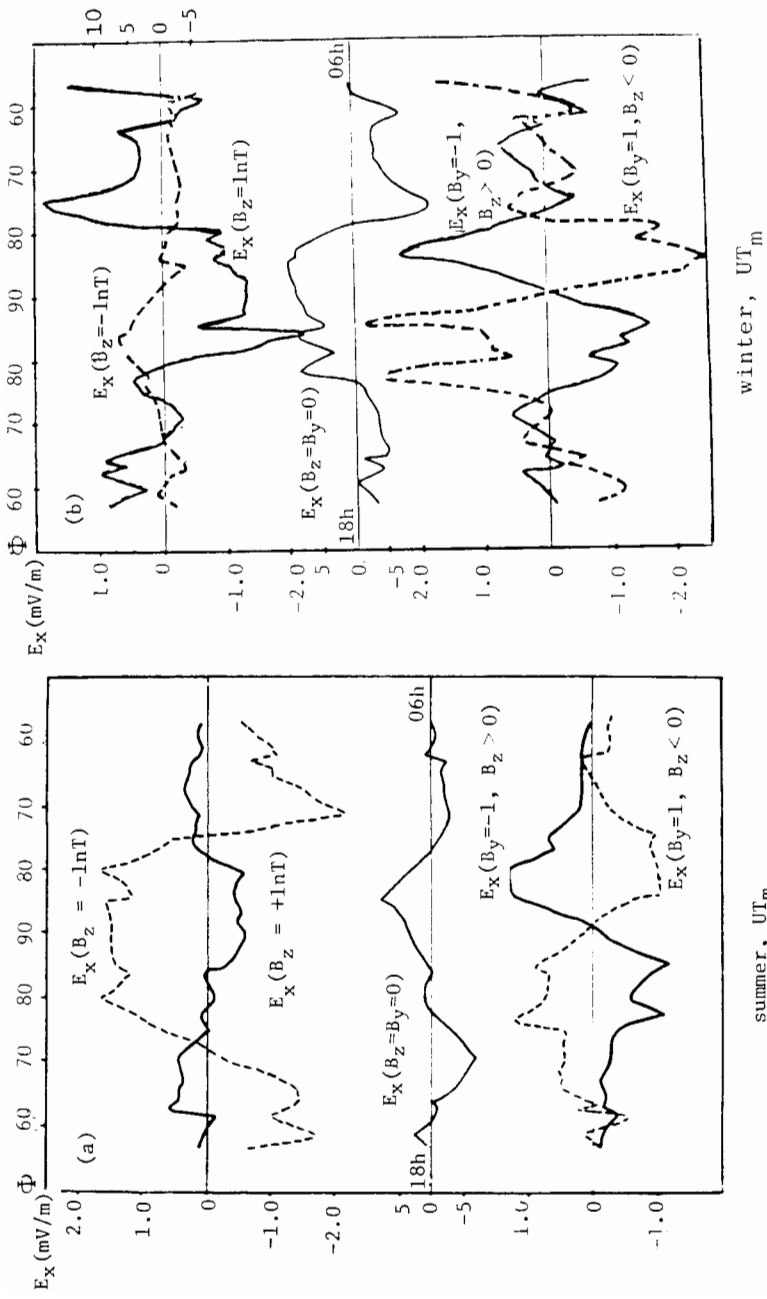


Fig. 14. The latitudinal profiles of the meridional electric field component taken along the dawn-dusk meridian in summer (a) and in winter (b) according to LEVITIN *et al.* (1984). The height-integrated ionospheric conductivity was taken for UT_m and for the respective magnetic activity levels according to WALLIS and BUDZINSKI (1981). The situations $B_z = 1$ nT, $B_z = 0$ and $B_z = -1$ nT, $B_z = 0$ are shown at the top, the situation $B_z = 0$ is shown in the middle, and the situations $B_z = -1$ nT, $B_z > 0$ and $B_z = 1$ nT, $B_z < 0$ are shown at the bottom. The scale for the $B_z = -1$ nT, $B_z = 0$ situation in winter is shown on the ordinate axis at the right.

configurations described by HEPPNER (1972) when analyzing the OGO-6 data.

The model distributions of $E_m(\Phi)$ in winter (Fig. 14(b)) not only preserve the main features of the summer distributions but also exhibit their own characteristic features. The field $E_m(B_z < 0)$ in the polar cap (the scale for this configuration is shown in Fig. 14(b) along the ordinate axis to the right) is much different from the Heppner type A structure. The field controlled by the $B_z < 0$ is inhomogeneous in the polar cap and is directed mainly from dawn to dusk in the dusk sector where it takes its maximum values. The configuration of the electric field E_m in polar cap in winter is closer to the Heppner configuration A than in summer.

Thus, it follows from the proposed model for $E(\Phi, \text{MLT})$, that the polar cap electric field along the dawn-dusk meridian is highly inhomogeneous and depends strongly on the IMF orientation. The model predicts a decrease of the polar cap electric field with values of positive B_z . According to the model, the potential difference along the dawn-dusk meridian through the polar cap cannot be treated as an unambiguous index of the intensity of the processes in the Earth's magnetosphere.

First, the highest potential difference through the polar cap is not located in all cases on the 0600-1800 MLT meridian (in certain IMF conditions such difference may be observed, for example, on the 1200-2400 MLT meridian, see Fig. 13). Second, even the zero potential difference does not indicate that the solar wind-magnetosphere interaction has ceased. In terms of the proposed model, the potential difference existing permanently in the magnetosphere is superposed on an additional field of opposite direction. As a result, the potential difference decreases. It should be borne in mind that the B_r component-controlled current system is produced by the polar cap electric field whose potential difference across the polar cap is almost zero. By no means, however, does this fact imply that the processes of the electric field generation in the magnetosphere have ceased.

8. Conclusions

Due to the efforts of recent years the quantitative concept of interplanetary medium connection with the electric and magnetic fields, field-aligned and ionospheric currents in the Earth's magnetosphere has been created. Ground observations combined with radar, rocket and polar satellite data together with numerical methods of solving the equation for the electrostatic potential have made it possible to find the large-scale features of the electromagnetic weather in the Earth's environment resulting from the changes in the interplanetary medium. Some results of such efforts are described above. They show the importance of ground-based observations of the geomagnetic field variations for the purposes of continuous detection of the changing conditions in the space. Conversely, the conditions in the space make it possible to appraise the electromagnetic conditions in the Earth's magnetosphere. The development of the proposed methods will be an extremely important advance toward quantitative understanding of large-scale magnetosphere-ionosphere coupling. The improvement of the network of ground-based magnetic stations, especially in the polar cap, more exact and detailed data about ionospheric conductivity will permit the local features of the electromagnetic state of the magnetosphere and their detailed structure to be studied further.

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