Global convection electric field and current: Comparisons between model’s predictions and data from STARE, SAINT-SANTIN and magnetometers
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Ionospheric electric field and ground magnetic field measurements at high, middle and low latitudes are used to study the global pattern of the convection electric field on March 26, 1979. The IMF $B_z$ component was southward and relatively steady over a prolonged period of time. The data set is thus representative of a steady magnetospheric convection. A semi-analytical time-dependent convection model is used to interpret the observations. During the daytime the model steady state results roughly reproduce the mean observed convection pattern. Fluctuations of the observed electric field from the model predictions are interpreted as resulting from (1) the transient responses of the ionosphere to small time variations of the IMF, (2) substorm effects, (3) neutral wind disturbances. Only the first type of fluctuations can be reproduced by the initial state prediction of the convection model.

1. INTRODUCTION

Axford and Hines [1961] and Dungey [1961] first suggested that solar wind/magnetosphere viscous interaction and magnetic reconnection transfer solar wind energy and momentum to the magnetosphere and ionosphere. This coupling creates a large-scale circulation of the magnetospheric plasma (current convection cells). Since then, it has gradually been recognized that the ionosphere also plays an important role in determining this plasma flow. Different methods have been developed to study the ionospheric electric current and field generated by the solar wind-magnetosphere interactions.

First, the observations of the ground magnetic variations led to the establishment of the $S_p$ and $D_p$ equivalent current systems [Nagata and Kokubun, 1962; Nishida, 1968]. Both current systems are composed of two cells. They differ only in their latitudinal extension: the $S_p$ current system remains confined at high latitudes, while the $D_p$ one extends to low latitudes. Nishida was the first to interpret these current systems as having their origin in the interaction between the solar wind and the magnetosphere.

Second, observations of the ionospheric electric fields led to the recognition that the convection electric fields generated by the solar wind/magnetosphere dynamo penetrate to middle and low latitudes. This is contrary to an earlier belief that the only ionospheric region fed by the energy available at the solar wind/magnetosphere interface is the auroral oval. Incoherent scatter radars have played an essential role in the observations made by both coherent and incoherent radars [Brekke et al., 1974; Zi and Nielsen, 1981, 1982; Mazaudier et al., 1984].

Third, these observations were interpreted by using theoretical models which considered the effects on the electric field distribution of ionospheric conductivities, ring current and interplanetary magnetic field (IMF). The influence of the magnetospheric ring current on the ionospheric electric field was investigated in analytical studies [Vasyliunas, 1970; 1972; Pellat and Laval, 1972; Southwood and Wolf, 1977] semi-analytical studies [Senior and Blanc, 1984], and numerical methods [Jaggi and Wolf, 1973; Harel et al., 1981a, b; Spiro et al., 1981; Wolf et al., 1982; Chen et al., 1982; Spiro and Wolf, 1984].

To progress further in the experimental determination of the global convection patterns, multi-instrument studies are essential. We have made observations of convection electric fields by means of coherent and incoherent radars (Scandinavian twin auroral radar experiment (STARE) and Saint-Santin radars), and equivalent current systems (Tromsø magnetometer), and compared these observations with model predictions in order to understand the solar wind, magnetosphere, and ionosphere interactions.

In a companion paper, Mazaudier et al. [1984] carried out an analysis of this kind, using magnetic and radar observations together with convection model predictions. This first study investigated the transient response of ionospheric electric fields to an increase of magnetospheric convection due to a sudden southward turning of the IMF $B_z$ component. The present paper is concerned with steady state convection, i.e., with the global convection pattern observed during a prolonged period of southward directed IMF $B_z$. We present observations made during an 18-hour period during which the IMF was southward and showed small magnitude fluctuations in time. In the following section, we present the geomagnetic conditions characterizing this time interval. We then relate electric field and magnetic observations at high, middle, and low latitudes to the Senior and Blanc [1984] model predic-
Tromsø's magnetic indices; (d) H component of the ground magnetic field at Tromsø, 26, 1979. (a) B\textsubscript{z} component of the IMF measured by the IMP 8 satellite; (b) equatorial magnetic Dst index; (c) AU and AL auroral magnetic indices; (d) H component of the ground magnetic field at Tromsø.

This linear time-dependent model estimates electric field due to the sole physical process of direct penetration of magnetospheric convection.

2. GEOMAGNETIC CONDITIONS ON MARCH 26, 1979

March 26, 1979, is a magnetically disturbed day (\(\Sigma Kp = 31, \Sigma Kn = 27\)). The amplitude of the IMF as measured by the IMP-8 satellite (located near the magnetopause in the solar wind) is quasi steady during the entire day (around 20 nT); the \(B_x\) component remains positive (around 18 nT). The \(B_y\) component, shown in the first panel of Figure 1, is negative centered around the mean value of \(-3\) nT from 0000 to 1800 UT. After 1800 UT, its magnitude decreases and \(B_y\) turns northward at about 2230 UT.

The Dst index (Figure 1b) shows two major negative excursions at 1200 and 1800 UT. This index remains below \(-25\) nT. The AU and AL auroral indices are shown in the third panel. These curves reveal sustained activity during the entire day with two clear substorm onsets shown with arrows. The most intense substorm begins around 1700 UT with a maximum AL of about 1200 nT at 1800 UT. At 2230 UT, when the IMF turns northward, the auroral activity stops. In the European sector, the Tromsø H component (Figure 1d) remains undisturbed from 0500 to 1100 UT, while the Dst is quasi steady. From 0000 to 0500 UT, the Tromsø station is under the westward electrojet, and from 1200 to 1700 UT under the eastward one. At the time of the 1700 UT substorm, the Tromsø H trace shows westward directed currents for a short period, before being again dominated by the eastward electrojet. From 2030 to 2230 UT Tromsø moves back into the westward electrojet. At 2200 UT, the auroral activity at Tromsø ends.

3. OBSERVATIONS AND RESULTS

In this section, we present separately the IMF measurements in the solar wind, and the electric field at high, middle and low latitudes and we compare some of these data to the outputs of the convection model of Senior and Blanc [1984].

The total cross polar cap potential drop \(\Phi_0\), and the local time of the maximum potential at the polar cap boundary are taken, respectively, as 70 kV and 0400 LT for the simulation. These values are chosen to fit the best the high latitude data from the STARE radar. The value of 70 kV is smaller than the estimation of \(\sim 110\) kV that is calculated from the empirical formula relating the IMF \(B_y\) component to \(\Phi_0\) given by Reiff et al. [1985].

3.1. High Latitudes

Figure 2 shows a polar plot of the \(E \times B\) drift as measured by the STARE coherent radar system, [Nielsen and Whitehead, 1983; Nielsen and Schlegel, 1983, 1985]. It illustrates the typical convection pattern at high latitude, with westward ion drift in the afternoon sector and eastward in the morning sector. Two reversals of the drift occur around 2100 and 2200 MLT (1930 and 2030 UT). Around 1200 MLT (1030 UT), the amplitude of the auroral electric field was probably smaller than 15 mV/m, the threshold at which the STARE radar can measure electric fields. The convection pattern observed on March 26, 1979, corresponds to the average convection pattern obtained by Zl and Nielsen [1982].

In Figure 3 we have superimposed the STARE southward electric field measured at the latitude of 69°N extracted from the Figure 2 and the \(B_x\) component of the interplanetary magnetic field. There is some correlation between the maxima of the electric field and the southward intensification of the \(B_x\) component. This suggests that variations of the \(B_x\) component of the IMF is an important parameter governing the ionospheric electric fields.

On Figure 4 we superimposed the diurnal variation of the \(H\) component of the ground-magnetic perturbation for March 26, 1979, at Tromsø (same data as Figure 1d) and the averaged \(Sd\) variation of \(H\) over Tromsø obtained by Mayaud [1965]. The averaged \(Sd\) variation is the mean value of the five most disturbed days (following the \(Kp\) index) of each month. This \(Sd\) curve gives the magnetic variation that would be observed by a station if it were rotating under a system of two convection current cells of fixed intensity and location. Hence the
difference between the two curves on Figure 4 can be interpreted as due to variations in intensity and/or location of the current cells. The best agreement between the two curves is obtained on the dayside between 0500 and 1800 UT. It shows that during that time interval, the European sector moves under a steady convection pattern. Before 0700 UT and after 1700 UT the $H$ component at Tromsø fluctuates and reaches maxima which correspond in general to the averaged $S_d$ value except during the main substorm around 1800 UT.

According to Chapman and Bartels [1940] the variation of the ground magnetic field is related to the height-integrated ionospheric current density by

$$|\Delta B| = \frac{2\pi}{10} f \cdot J$$  \hspace{1cm} (1a)

for an infinite current sheet above a plane earth.

In expression (1a), $B$ is in $\gamma$, $J$ is in amperes per kilometer, and $f$ is equal to 0.6 [Kamide and Brekke, 1975]. An horizontal ionospheric eastward $J_y$ current induces a variation in the horizontal component of the magnetic field

$$\Delta H \approx J_y$$  \hspace{1cm} (1b)

Following Ohm's law, and assuming homogeneous conductivity [see Wilkinson et al., 1986] the eastward current density is related to the ionospheric electric field by

$$J_y \approx \Sigma_H E_x$$  \hspace{1cm} (2)

where $\Sigma_H$ is the height-integrated Hall conductivity, $E_x$ is the northward electric field in the plane perpendicular to the magnetic field. From expressions (1b) and (2) we can derive

$$\Delta H \approx \Sigma_H E_x$$  \hspace{1cm} (3)

Figure 5 illustrates the relation between the ground magnetic perturbation and the southward electric field. In this figure the STARE southward electric field measured in the region over Tromsø is superimposed on the $H$ component at

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Fig. 3. Southward component of the STARE electric field and $B_z$ component of the IMF. The southward/northward electric field intensifications in the auroral zone are generally associated with southward increases of $B_z$.

Fig. 4. The $H$ component of the ground magnetic field observed at Tromsø on March 26, 1979 (solid line), is superimposed on the mean $S_d$ value of the $H$ component estimated by Mayaud [1965] (dashed line). This $S_d$ variation is the mean value over a solar cycle of the five most magnetically disturbed days of each month.
A very good agreement between the variations of $\Delta H$ and $E_z$ is found except from 1700 to 2000 UT, during the substorm event. The ratio between the $H$ component and the southward electric field gives an estimate for the height-integrated Hall ionospheric conductivity in the auroral zone. For this data set the Hall conductivity varies between 5 and 12 mhos. These two extremas have been selected for simulations.

The convection model of Senior and Blanc [1984], which considers the direct penetration of convection to midlatitudes (see the appendix), has been used to determine the steady state ionospheric electric field. The two simulations developed in this study are described in Table 1. Some of the input parameters have been derived from our observations. The electric potential $\Phi_0$ along the polar cap/auroral zone boundary has been adjusted to fit the STARE northward electric field component. Its amplitude has been chosen as 70 kV, with a maximum at 0400 LT.

The auroral conductivity has been selected by comparing the Tromsø $H$ component and the northward STARE electric field. In Figure 6a the model southward electric field is superimposed on our STARE data. The model reproduces the mean diurnal variation observed except during the 1800 UT substorm event. The model obviously does not reproduce the fluctuations of the electric field related to the $B_z$ fluctuations. This is because we consider the convection electric field to be induced by a steady potential drop over the polar cap. The agreement between model predictions and data for the eastward component of the electric field is rather poor (Figure 6b). This is in fact not very surprising since this component is known to be very weak and to show little systematic magnetic activity dependence [Senior and Blanc, 1984; Mazaudier et al.,

![Fig. 5. The southward STARE electric field and ground magnetic $H$ component at Tromsø are superimposed. A good correlation is found between the southward/northward intensifications of the electric field and the northward/southward increases of the magnetic field, except during the 1800–2000 UT substorm event.](image)

TABLE 1. Inputs of the Convection Model for the Two Simulations S1 and S2

<table>
<thead>
<tr>
<th>Latitudinal Range</th>
<th>S1</th>
<th>S2</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\lambda = 72^\circ$</td>
<td>$\Phi_0 = 70$ kV*</td>
<td>$\Phi_0 = 70$ kV*</td>
</tr>
<tr>
<td>Cross Polar Cap Potential</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>$\Sigma_n = 5$ mhos</td>
<td>$\Sigma_n = 5$ mhos</td>
</tr>
<tr>
<td></td>
<td>$\Sigma_n/\Sigma_p = 1.6$</td>
<td>$\Sigma_n/\Sigma_p = 1.6$</td>
</tr>
<tr>
<td>Conductivities</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Auroral zone $67^\circ \leq \lambda \leq 72^\circ$</td>
<td>$\Sigma_n/\Sigma_p = 2$</td>
<td>$\Sigma_n/\Sigma_p = 2$</td>
</tr>
<tr>
<td>Subauroral zone $64^\circ \leq \lambda \leq 67^\circ$</td>
<td>$\Sigma_n = 5$ mhos</td>
<td>$\Sigma_n = 5$ mhos</td>
</tr>
<tr>
<td>Midlatitudes $10^\circ \leq \lambda \leq 64^\circ$</td>
<td>$\Sigma_n = 5$ mhos daytime, $\Sigma_n = 5$ mhos nighttime</td>
<td>$\Sigma_n = 5$ mhos daytime, $\Sigma_n = 5$ mhos nighttime</td>
</tr>
<tr>
<td></td>
<td>$\Sigma_n/\Sigma_p = 1.6$</td>
<td>$\Sigma_n/\Sigma_p = 1.6$</td>
</tr>
<tr>
<td>Equatorial region $\lambda \leq 10^\circ$</td>
<td>equatorial noontime conductance $S_0$</td>
<td>same as S1</td>
</tr>
<tr>
<td></td>
<td>$\delta_0 = 3.83 \times 10^5 \Omega \text{m}$</td>
<td>same as S1</td>
</tr>
</tbody>
</table>

*Maximum at 0400 LT.
Fig. 6a. The southward component of the STARE electric field on March 26, 1979 (solid line), is superimposed on the southward convection electric field obtained from two distinct numerical simulations of the Senior and Blanc [1984] model. The model predictions correspond to the steady state of magnetospheric convection.

Fig. 6b. Similar to Figure 3b for the eastward component of the auroral electric field.
Santin), the steady state convection model give a rough estimate of the southward electric field variation observed on March 26, 1979. The same agreement is obtained for the eastward component (Figure 7b) on the dayside. It should be noticed however that, for midlatitudes, the convection model predicts similar convection patterns on the dayside at the initial and steady states. On the nightside strong discrepancies are observed between model and data for both components. Those occurring on the morningside (from 0000 to 0700 UT) can be attributed to neutral wind disturbances. Those occurring on the eveningside (after 1700 UT) can be explained by the substorm that affects the Tromsø area from 1800 to 2000 UT, and by the northward turning of $B_z$ around 2230 UT, both effects that are not taken into account in the model.

### 3.3. Equatorial Latitudes

Figure 8a is composed of two panels, the upper one for Addis-Ababa (geographic latitude 9.03°N, geographic longitude 38°77E, LT = UT + 3) and the bottom one for Huancayo (12.05°S, 284°67E, LT = UT − 5). On these two panels we have superimposed the $H$ component of the terrestrial magnetic field on March 26, 1979 (dashed line), and the same component observed on March 21, 1979; March 21 was a magnetically quiet day, that we use as a reference. It represents the usual equatorial electrojet of the quiet ionospheric dynamo. Following Akasofu and Chapman [1961] the ground magnetic variation can be expressed by

$$\Delta H = DR + DP + DCF + DT$$

(4a)

in this expression $DR$, $DP$, $DCF$, and $DT$ correspond to ring, ionospheric, magnetopause and tail currents, respectively.

On both March 21 and 26, due to the fact that there is no magnetic storm, the equatorial variation reduced to [Fukushima and Kamide, 1973]

$$\Delta H = DR + DP$$

(4b)

The symmetric part of the $DR$ field is approximated by $Dst \cos \lambda$, where $Dst$ is the magnetic equatorial index and $\lambda$ the latitude of the station.

To emphasize the $DP$ disturbance on March 26, 1979, it is necessary to have a quiet reference for the ionospheric $DP$ current system. This reference is represented by the circled crosses on Figure 8a. It is obtained by subtracting the $DR$ field ($DR = Dst \cos \lambda$) from the $H$ component of the ground magnetic variations of the quiet reference day (March 21). Therefore, the difference between the $H$ component of ground magnetic field on March 26 and the circled crosses is a measure of the ionospheric $DP$ disturbance on that day.
At Addis-Ababa, the equatorial DP disturbance on the day-side observed on March 26, 1979, is southward, and corresponds to a westward equivalent current. It occurs from 1100 to 1500 LT (0800 to 1200 UT). At Huancayo the equatorial DP disturbance fluctuates and shows strong eastward intensifications of the electrojet from 0700 to 1300 LT (1200 to 1800 UT).

In Figure 8b the northward component of the Stare electric field and the DP disturbance at Huancayo are superimposed. It is clear that the northward intensifications of the auroral zone electric field correspond to the eastward equatorial electrojet intensifications. These fluctuations are also associated to southward increases of $B_z$ and are the signature of the $Dp_2$ equivalent current system.

The convection model of Senior and Blanc [1984] which includes a rough description of the electrojet region through its low-latitude boundary condition (see appendix and Table 1), predicts the total equatorial currents at initial state and steady state (see Figure 9). On the dayside, at the initial state, the total electrojet increase is eastward. This variation is similar to the $H$ component fluctuations observed at Huancayo. The total current in the equatorial electrojet was estimated from the Huancayo data by assuming the jet to be 10° latitude wide in the northern hemisphere as in the model, and to have uniform electric current density of 100 A/km (100 nT from Figure 8b). The total current in the equatorial electrojet is then 100 kA; this order of magnitude corresponds to the initial state model predictions.

At Addis-Ababa (Figure 8a upper panel), the daytime electrojet disturbance is westward with an intensity of about 100 kA. This is not in good agreement with the predictions of the convection model. It is therefore probable that another mechanism operates locally at Addis-Ababa. In fact, Blanc and Richmond [1980] have shown that the ionospheric disturbance dynamo has a negative effect on the equatorial electrojet, i.e. a westward increase of the current during the daytime. Study of the $F$ region neutral winds at Saint-Santin reveal the presence of neutral wind disturbances on the nightside until 0700 UT. It is thus possible that these disturbances propagate south-eastward toward Addis-Ababa with a time constant of several hours, characteristic of the ionospheric dynamo process. The result would be westward electrojet disturbance as observed at Addis-Ababa. Huancayo which is located west of Saint-Santin, would not see these perturbations.

4. DISCUSSION AND CONCLUSION

In this paper we have combined magnetic data with coherent and incoherent radar data, and compare these data to the predictions of a convection model. This enables us to analyze quantitatively the large-scale convection electric field pattern on March 26, 1979.

On this particular day, (1) the IMF $B_z$ component was southward and relatively steady over a prolonged period of time from 0000 to 1900 UT, (2) there was no neutral wind disturbance at midlatitudes in the dynamo region from 0700 to 1800 UT. This made it possible to study the prolonged action of the magnetospheric convection process from 0700 to 1800 UT. The convection model of Senior and Blanc [1984] was used to calculate the ionospheric electric field. The steady state model prediction was used to establish the mean diurnal variation of the electric field due to the mechanism of direct penetration of magnetospheric convection.

![Figure 8b](image-url)
This paper shows that on the dayside the steady state of the Senior and Blanc convection model gives a rough estimate of the ionospheric convection pattern in the latitude range from the auroral zone to midlatitude.

At auroral latitudes, the main diurnal variation of the southward STARE auroral electric field is roughly reproduced by the steady state model as well as the $S\theta$ variation of the $H$ component observed at Tromsø [Mayaud, 1965].

The southward/northward observed electric field intensifications, usually associated with southward $B_z$ increases, are not reproduced by the steady state model. They were interpreted as due to transient responses of the ionosphere to the southward $B_z$ fluctuations and reproduced by initial state of the convection model.

At middle latitudes, the daytime variations of the two electric field components are well reproduced by the convection model, both the initial and steady states of the model (which are close to each other on the dayside).

At Huancayo, close to the equator, the $H$ component of the ground magnetic field showed strong eastward intensifications of the equatorial electrojet associated with increases of the northward STARE auroral electric field. These observations were not reproduced by the steady state model, which predicts a small westward electrojet intensification associated with northward electric field penetration as due to the transient response of the ionosphere (initial state model) to the IMF $B_z$ fluctuations ($\Phi_o$ cross polar cap potential). The discrepancies between convection model and data at Addis-Ababa has been interpreted as a local effect due to the ionospheric disturbance dynamo process.

A previous companion paper [Mazaudier et al., 1984] illustrated the initial state of convection electric field induced by rapid change in the solar wind magnetosphere dynamo (southward turnings of $B_z$). This paper showed that even during a time of prolonged southward IMF $B_z$, as on March 26, 1979, it is difficult to find a perfect steady state of ionospheric convection electric field in the whole ionosphere. Rather, the data can be analyzed in terms of a mean diurnal variation corresponding to the steady state model, and of short-term fluctuations produced by transient changes of the source potential due to IMF variations. It seems therefore that a linear model of the convection process is sufficient to interpret quantitatively the electric field observations when there is no substorm or neutral wind disturbance operating simultaneously to the convection process.

**Appendix: Description of the Senior and Blanc [1984] Model**

This linear time-dependent and self-consistent model calculates the response of the magnetosphere ionosphere circuit to a step function of the source, i.e., the cross polar cap potential. It takes into account the latitudinal and longitudinal gradients of the midlatitude ionospheric conductivities created by solar radiations.

This model keeps the simplicity (and also indeed the limitations) of its linearity and has freely adjustable values of the high latitude conductivities. The conductivity in the auroral zone is simulated with two uniform conductivity rings: the auroral zone from colatitudes $\theta_0$ to $\theta_2$ and a subauroral zone from $\theta_2$ to $\theta_1$. For $\theta$ greater than $\theta_1$ (middle and low latitudes) the integrated Pedersen conductivity varies during the daytime as

$$\Sigma_p = \Sigma_p^0 \cos \chi \cdot \frac{B_0}{B}$$

where $\chi$ is the solar zenith angle, $B$ is the local value of the magnetic field, $B_0$ and $\Sigma_p^0$ are the values of the equatorial magnetic field and conductivity at local noon. During the night the midlatitude conductivities are small and uniform in local time:

$$\Sigma_p = \frac{\Sigma_p^0 - 2 B_0}{30} \cdot \frac{B_0}{B}$$

The subsolar Pedersen conductance has been taken as 20 Mhos and a smooth transition between the day and night values, modeled as a parabolic function of $\chi$, is imposed on the dawn and dusk sectors over 20° of the parameter. Finally, the midlatitude Hall to Pedersen conductivity ratio is 1.7 and the magnetic field is dipolar with an equatorial value of 0.28 $\times 10^{-4}$ T.

The equatorial electrojet region is included in the model as a boundary condition. It is a small conducting ring though which the meridional currents close. The equatorial conductance is 3.83 $\times 10^4$ mhos at noon and has the same local time variations as the midlatitude conductivities.

The inner boundary of the equatorial ring current is assumed to map dawn on the circle of colatitude $\theta_1$, which represents the equatorial boundary of the auroral zone. The model yields self-consistent variations of the inner boundary in the magnetospheric equatorial plane and of the LT distribution of Birkeland currents generated along this boundary. The colatitudes $\theta_0$ and $\theta_2$ and conductivities in each zone are free parameters of the model and can be adjusted to fit the observations.

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**References**


Mayaud, P. M., A programme de la variation journalière Sg du champ magnétique terrestre par la variation journalière SD et d'un type spécial de perturbations contribuant au SD d'été, Ann. Geophys., 21, 219, 1965.


