INTERNATIONAL JOURNAL OF GEOMAGNETISM AND AERONOMY
VOL. 2, NO. 3, PAGES 209–224, DECEMBER 2001

Effects of geomagnetic storms on the ionosphere and atmosphere

A. D. Danilov

Institute of Applied Geophysics, Moscow, Russia

J. Lastovička

Institute of Atmospheric Physics, Prague, Czech Republic

Abstract. A geomagnetic storm is a complex process: its various features act at different heights. In the $F2$ layer the midlatitude effect is basically an ionospheric response to storm-induced changes in the neutral atmosphere, which are primarily a consequence of a strong Joule heating in the auroral thermosphere. At lower heights the role of ionization and photochemical processes increases due to shorter electron lifetimes. At the base of the $F1$ layer (160–170 km) the storm effect is almost absent. At E-region maximum a complex action of several factors results in a slight decrease of f_oF2, even though below and above, the electron density increases. Farther down, in the lower ionosphere, a strong increase of the electron density is observed as a consequence of a very strong enhancement of particle precipitation. In the neutral upper middle atmosphere, the effects of enhanced precipitation weaken with decreasing altitude and become insignificant and/or absent in the stratosphere. The effect of geomagnetic storms reappear in the lower atmosphere but as an effect of different morphology and origin.

1. Introduction

Geomagnetic storms are probably the most important phenomenon among those related to solar wind and high-energy particles. They produce large and global disturbances in the ionosphere, but they affect also the neutral atmosphere, including the middle atmosphere and troposphere [e.g., *Lastovička*, 1996].

The geomagnetic storm is a complex process of solar wind/magnetospheric origin. Various features of this complex process act at various altitudes in the ionosphere and neutral atmosphere. This paper presents an experiment construct a "vertical profile" and related scenarios of the geomagnetic storm effects on the Earth's atmosphere and ionosphere starting from the $F2$-layer maximum down to the troposphere. It is in no case a full review of scientific results in this field. The paper concentrates predominantly (even though not only) on the northern hemisphere middle latitudes and on the results of East European authors, which are often less familiar to the scientific community.

The effects of geomagnetic storms at different altitudes and latitudes differ in development in time and in intensity. They reflect different features of geomagnetic storms, therefore their mechanisms are different. This makes construction of the vertical profile of effects and responsible mechanisms difficult.

The geomagnetic storm should be called magnetospheric storm, because the observed changes of geomagnetic field are essentially a consequence of strong and rapid magnetospheric processes and changes under solar wind action. The name "geomagnetic storm" is traditional, because storms had been observed first as changes of geomagnetic activity/field, and they have been monitored until now by geomagnetic activity measurements.

Paper number GAI99312.
CCC: 1524–4423/2000/0203–0312$18.00

The online version of this paper was published January 25, 2001.
URL: http://ijga.agu.org/v02/gai99312/gai99312.htm
Print companion issued December 2001.

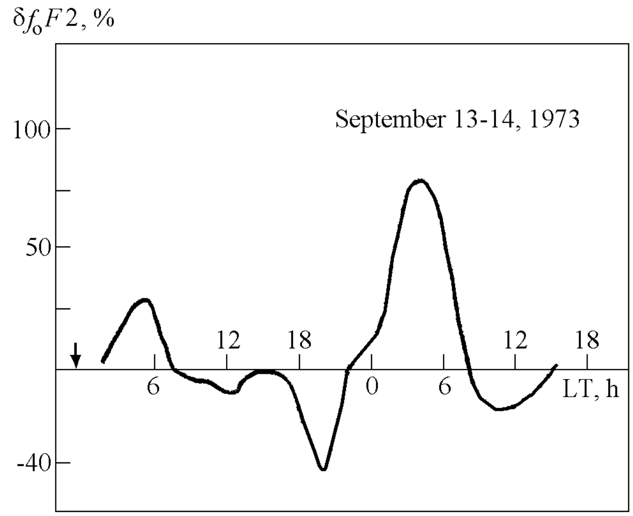

Figure 1. The $\delta f_o F2$ variations during the ionospheric storm of September 13–14, 1973 in Leningrad. The arrow indicates the storm sudden commencement (SSC) moment.

2. $F2$ Layer

2.1. Morphology

The response of the ionospheric $F2$ layer to a magnetospheric disturbance is different from that of the lower ionosphere described below. The difference is due to the differences in physical mechanisms responsible for the changes of the electron concentration. While in the E and D regions the primary reason of the electron concentration changes is the variation of the ionization rate because of corpuscular intrusions, in the $F2$ layer the electron concentration variations are predominantly due to such factors as neutral composition changes and dynamical processes.

There is one more difference in the electron concentration behavior in various ionospheric regions. All "disturbances" in the lower ionosphere (poststorm effects, auroral absorption, polar cap absorption (PCA)) are actually an increase of electron concentration above some quiet-time background level. Contrary to that, the $F2$-layer response to a geomagnetic disturbance (so called ionospheric storm) consists of effects of both signs. Both a depletion and an increase of the electron concentration relative to a background level are observed during those storms and are called "positive" and "negative" phases of the storm, respectively.

Table 1. Delay and amplitude of the negative phase maximum [after *Mebagishvili and Khochalava*, 1977]

Station	Φ, deg	ΔT, h	$\delta f_o F2$, %
Leningrad	56.3	6	20
Moscow	50.8	8	16
Rostov	42.5	13	11
Alma-Ata	33.5	20	8

The morphology of ionospheric storms is rather complicated. The response of the ionosphere as seen at different ionospheric stations may be quite different during the same storm depending on the station coordinates, local time of the geomagnetic storm onset, and some other parameters. The global distribution of ionospheric storm effects is also rather complicated and differs considerably from one storm to another.

Here we cannot provide a detailed description of this rather complicated situation, referring the readers to a detailed review by *Prölss* [1995] and the references therein and citing some recent publications. We are going to present a rather "smoothed" picture aimed mainly at a comparison of the $F2$-layer and thermospheric response to geomagnetic disturbances with that of the lower ionosphere.

An ionospheric storm in the $F2$ layer is usually manifested in variations of the electron concentration in a wide altitude range from about 200 km to 800–1000 km. The principal features of such storms are also manifested in the behavior of the total electron content (TEC) as observed from the satellite-to-ground radio wave propagation.

The $F2$-layer response to a geomagnetic disturbance is in the most convenient way described in terms of $\delta f_o F2$, i.e., in deviations of the critical vertical reflection frequency $f_o F2$ of the $F2$-layer maximum from a quiet-time median:

$$\delta f_o F2 = (f_o F2_{\rm obs} - f_o F2_{\rm med})/f_o F2_{\rm med}$$

The temporal behavior of $\delta f_o F2$ during an ionospheric storm usually consists of periods with positive values (positive phases of the storm) and negative values (negative phases). An example of the $\delta f_o F2$ behavior during a typical storm is shown in Figure 1. The morphology of negative and positive phases of an ionospheric storm is different because the principal physical mechanisms responsible for their formation are believed to be different. Below, we consider the two phases separately and their interaction to produce a storm as a whole.

One of the significant features of the negative phase is its equatorward shift during the storm from auroral latitudes to middle latitudes. The amplitude of the effect (the maximum absolute value of $\delta f_o F2$) decreases during this shift. This is illustrated in Table 1 on the basic of averaging the ionospheric storms of 1957–1964. Table 1 shows that the lower the station geomagnetic latitude Φ, the larger the delay ΔT of the negative phase beginning relative to the beginning of the magnetic disturbance (storm sudden commencement (SSC)) and the lower the maximum value of $\delta f_o F2$. The velocity of the negative phase equatorward "drift" is, according to various estimates, about 50–300 m s^{-1} [*Danilov and Belik*, 1991].

The equatorward penetration of negative phases demonstrates a seasonal behavior. In the summer hemisphere, during the same storm, the negative phase is better developed and drifts to lower latitudes than in the winter hemisphere [*Prölss*, 1995].

The negative phase in most cases demonstrates a well-pronounced dependence of the intensity $I_{\max}$ (the maximum

δf_oF2 in percent) on the AE index. Figure 2 shows such dependence for 10 selected storms according to *Danilov and Belik* [1991]. It is worth noting that the dependence of AE is equally well pronounced at least for three harmonics (changes from negative to positive phase and back. Dependencies of $I^-_{\rm max}$ on the Kp, Dst, and Ap indices have been also reported (see the references in the *Danilov and Belik* [1991] work).

The morphology of the positive phase is even more complicated [*Zevakina*, 1971]. This fact points to more complicated physical processes of its formation. However, some features may be indicated. Positive phases are sometimes observed several hours before the beginning of the magnetic disturbance, which is considered to be a reason of this particular ionospheric storm [*Danilov and Belik*, 1991; *Szuszczewicz et al.,* 1998]. Sometimes the entire storm consists of a single durable positive phase, δf_oF2 never becoming negative. Examples of these two features are presented in Figure 3. Also, the existence of so called "dead zones" should be mentioned when in the middle of a storm for several hours the deviation from the median does not exceed 10% [*Danilov and Belik*, 1991]. All of these features are important for understanding the physical mechanisms operating in the thermosphere and ionosphere during a magnetic disturbance.

While negative phases are almost always observed at high latitudes and more often than positive phases at middle latitudes, positive phases tend to occur at middle and low latitudes. The positive phase is observed in most cases (90%) at equatorial latitudes during magnetic disturbances [*Adeniyi*, 1986; *Mikhailov et al.,* 1994]. However, during prominent disturbances, the negative phase may also be observed [*Adeniyi*, 1986; *Turunen and Rao*, 1980]. As for the seasonal preference, negative phases occur in all seasons, less in winter, when positive phases show the a maximum of occurrence.

2.2. Physical Mechanisms of Ionospheric Storm Formation

2.2.1. Negative phase. Since the global picture of δf_oF2 distribution during ionospheric storms is usually rather complicated, the role of various physical mechanisms is

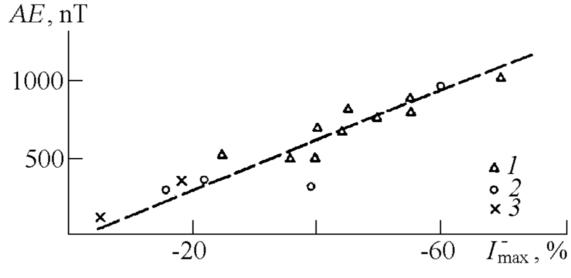

Figure 2. Negative phase amplitude $I^-_{\rm max}$ dependence on the geomagnetic AE index according to *Danilov and Belik* [1991] (1, 2, and 3 are the first, the second, and the third day of the storm, respectively).

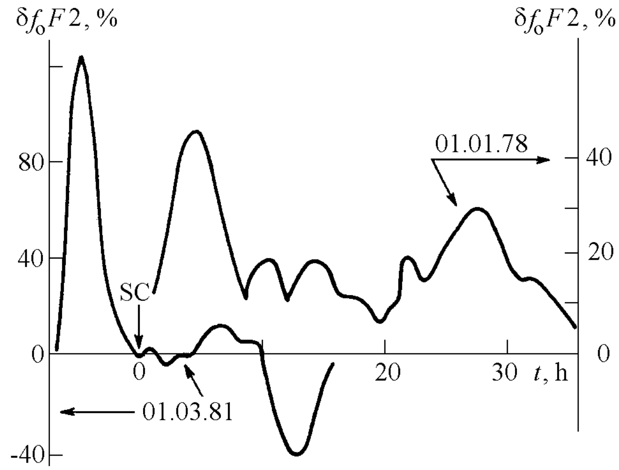

Figure 3. Examples of occurrence of the positive phase before the SC (the storm on March 1, 1981) and purely positive disturbance (the storm on January 1, 1978) according to *Danilov and Belik* [1991].

usually considered on the basis of case studies in which the δf_oF2 (or $\delta[e]$) behavior for a particular storm is studied for one particular location (for example, an incoherent scatter facility), or for several stations of one particular region [e.g., *Buonsanto*, 1995; *Mikhailov and Förster*, 1997; *Mikhailov and Schlegel*, 1998; *Mikhailov et al.,* 1994, 1995].

Since the first suggestion by *Seaton* [1956], it was believed that the negative phase is caused by the changes of the thermospheric composition due to heating of the thermosphere during geomagnetic disturbances. The electron concentration is, roughly speaking, directly proportional to the $[O]/[N_2]$ ratio at the $F2$-layer maximum height [*Rishbeth and Barron*, 1960]. This proportionality is true for stationary daytime conditions without vertical drifts. We refer the readers for more exact relations to *Mikhailov et al.* [1989, 1995]. This means that under other unchanged conditions we should have a depletion of electron concentration (a negative phase) in all the regions where $[O]/[N_2]$ has been decreased at F-region heights.

Prölss and von Zahn [1974] and *Prölss* [1980] found a close relation between the $[O]/[N_2]$ ratio measured by the ESRO 4 satellite and the electron concentration in the $F2$ peak measured at several ionospheric stations. Figure 4 from *Prölss and von Zahn* [1974] shows that the variations of electron concentration and $[O]/[N_2]$ at each station are very similar, though different for different stations.

The reason of the thermospheric composition changes ($[O]/[N_2]$ decrease) is fairly obvious. During a geomagnetic disturbance there is a heating of the lower part of the thermosphere (100–140 km) in the auroral region. The main source of this heating is the Joule dissipation of electric currents, but some energy may be deposited also by precipitating particles [*Prölss*, 1995].

The heating should lead to a significant decrease of the atoms-to-molecules ratio throughout the entire thermosphere in the high-latitude region. If the thermospheric dynamical regime stayed unchanged during magnetic disturbances,

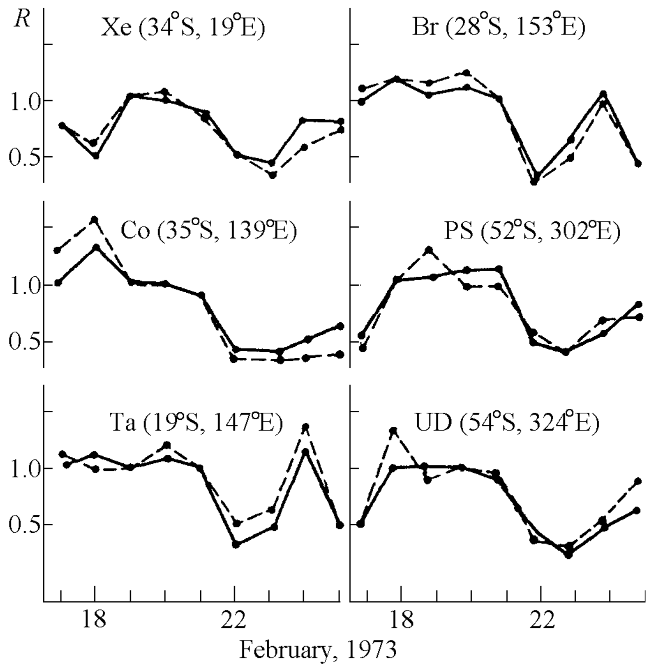

Figure 4. Variations of the parameter R (the disturbed-to-quiet ratio) for [O]/[N_2] (solid lines) and electron concentration (dashed lines) during the ionospheric storm of February 1973 for six ionospheric stations (Hermanus, Saulisbury, Townsville, Brisbane, Port Stanley, and South Georgia) according to *Prölss and von Zahn* [1974].

the zone of depleted [O]/[N_2] (and so electron concentration and f_oF2) would be limited by the high-latitude ionosphere (approximately by the auroral oval). However, the heating induces its own circulation (which may conflict with the "regular" circulation; see below), which at the $F2$-layer heights tends to bring the air equatorward to lower latitudes [*Duncan*, 1969]. That, in the simplest approximation, leads to the aforementioned drift of the negative phase to lower latitudes.

It should be noted here that the heated gas with depleted [O]/[N_2] ratio has a higher temperature throughout the thermosphere. The increase of temperature leads at the F-region heights to an increase of the linear recombination coefficient and thus to a further decrease of the electron concentration [*Mikhailov et al.,* 1995]. Thus, actually, the negative phase is formed in the heated thermospheric gas due to two factors: the depleted [O]/[N_2] and the increased recombination due to increased temperature [*Mikhailov and Förster,* 1997].

The latitudinal distribution of the negative phase according to the simple scheme described should be different in the sunlit and nighttime sectors of the winter ionosphere. The storm-induced circulation is directed equatorward. In winter it is opposite to the daytime background thermospheric circulation, which is directed poleward. This leads to the effect of stopping the negative phase equatorward drift, the region of the negative phase development thus being confined to high latitudes.

During the nighttime period the two circulations (the background and the storm-induced ones) coincide (both are directed equatorward), and so the air with the disturbed (depleted) [O]/[N_2] ratio spreads out to much lower latitudes than in the daytime. This leads to a rather frequent occurrence of the negative phase at middle latitudes at night in winter.

In summer the poleward background (quiet) circulation is reduced as compared with winter. It is equatorward the most part of the day and thus coincides with the stormin-duced circulation. This is favorable for penetration of the air with depleted [O]/[N_2] (and increased temperature) to middle latitudes, and so the negative phase in summer is observed at middle latitudes both in the daytime and in the night time.

There was some discussion on the role of vibrationally excited nitrogen in forming the negative phase. The matter is that due to several factors, including the temperature increase, there may occur an increase of the N_2 vibrational temperature $T(N_2^{\#})$ at $F2$-layer heights. The increase should lead to strong intensification of the $O^+ + N_2$ reaction and so to significant reduction of the electron concentration. Some authors [e.g., *Pavlov,* 1994; *Pavlov and Buonsanto*, 1997] believe that this process plays a significant role in negative phase formation. However, there is no direct proof that the increase of $T(N_2^{\#})$ actually occurs. On the other hand, in many case studies [*Mikhailov and Foster*, 1997; *Mikhailov and Schlegel*, 1998; *Mikhailov et al.,* 1994, 1995] the principal features of many storms (including prominent ones) were successfully explained without any assumption on the $T(N_2^{\#})$ increase. *Loewe and Prölss* [1998] claim that $N_2^{\#}$ is not very important in general, but it may play some role under high solar activity conditions.

2.2.2. Positive phase. Several mechanisms have been considered as a probable source of the ionospheric storm positive phase [e.g., *Danilov and Belik,* 1991; *Prölss*, 1995]. These are as follows: the $F2$-layer uplifting due to vertical drift, plasma fluxes from the plasmasphere, and downwelling of the gas as a result of the storm-induced thermospheric circulation.

The $F2$-layer drift may be caused by two principal mechanisms: an increase of the electric fields of magnetospheric origin and the equatorward horizontal thermospheric circulation which lifts up the $F2$-layer plasma along the inclined field lines. The question of strong electric fields penetrating the F region from the magnetosphere is still open, and there is no direct proof that this mechanism contributes significantly to the $F2$-layer behavior during ionospheric storms at middle latitudes [*Prölss*, 1995]. Neither are there any indications to the significant role of plasma fluxes from the plasmasphere. As for the circulation-induced drift and thermospheric gas downwelling, they are believed to participate essentially in a positive phase formation and will be considered below.

The above described proportionality of the electron concentration in the $F2$-layer maximum to the [O]/[N_2] ratio was considered for decades to be enough to explain the negative phase by the atoms-to-molecules ratio variations. In the recent years, more sophisticated analyses have shown that actually, the $[e]_{\max}$ dependence on [O] and [N_2] is more complicated [*Mikhailov et al.,* 1989, 1995]. The electron concentration is proportional to $[O]^n$. In a general case, $n = 4/3$. At lower latitudes where $[O]/[N_2]$ does not change, we have purely a compensated dependence with $n = (0.7 - 0.85)$ [*Mikhailov et al.,* 1989], and hence the simultaneous variation of both [O] and $[N_2]$ with the $[O]/[N_2]$ ratio kept unchanged still leads to an increase of the electron concentration. This is what sometimes happens during ionospheric storms at low latitudes. The storm-induced circulation leads to a downwelling of the thermospheric gas at low latitudes and so to an increase of [O] and $[N_2]$ without any significant changes of the $[O]/[N_2]$ ratio [*Mikhailov et al.,* 1995]. Satellite (for example, ESRO 4) measurements show a constancy of the $[O]/[N_2]$ ratio in the downwelling zone during some storms. Attempts to reproduce this effect in a three-dimensional approximation were successful [*Förster et al.,* 1999]. This is the nature of the majority of the ionospheric storms observed at latitudes below 30°–40°.

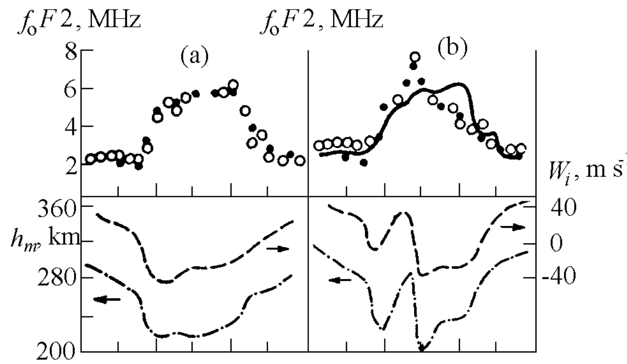

Figure 5. Variations of the $F2$-layer parameters: f_oF2 (circles), the vertical drift velocity w_i (dashed curve), and layer peak altitude h_mF2 (dashed-doted curve) on (a) January 24 (quiet day) and (b) January 25 (disturbed day), 1974 at the Boulder station according to *Mikhailov et al.* [1989]. The mean diurnal variation of f_oF2 observed on January 24 is shown by the solid line in the panel (b).

2.2.3. Common features. The vertical plasma drift is a very important factor influencing the state of the $F2$ layer. In quiet conditions the drift is caused by the horizontal circulation. Equatorward and poleward horizontal winds lead to upward and downward drift, respectively, due to the inclination of the magnetic field lines at middle latitudes. In quiet steady state conditions the drift is balanced with other processes. However, when a storm-induced circulation is present, it enhances or weakens the quiettime circulation, and so enhances or weakens the vertical drift.

In the midlatitude winter ionosphere in the daytime the zone of the decreased $[O]/[N_2]$ ratio is "closed" at high latitudes [*Duncan*, 1969] because the normal circulation prevents the heated gas penetration to middle latitudes. At the same time the quiet circulation is weakened by the opposite storm-induced circulation, so the downward drift of the plasma is weaker. It leads to new equilibrium conditions in the $F2$-layer maximum with its height h_mF2 increasing, and so increasing the electron concentration. This is the reason of the positive phase predominating in these conditions. An example of f_oF2, drift velocity, and h_mF2 variations during a quiet and a storm day (January 24–25, 1974) according to *Mikhailov et al.* [1998] is shown in Figure 5.

In summer the quiet-time and storm-induced circulations coincide (both are equatorward). In this case the upward vertical drift increases, and again, we should have a positive phase. However, it is not always the case because the storm-induced circulation brings the gas with the depleted $[O]/[N_2]$ ratio, and this tends to reduce electron concentration in the $F2$ peak. The rivaling of the two factors explains why in summer both the positive and the negative phases are observed. It explains also the existence of the dead zones, when during several hours the δf_oF2 modulus does not exceed 10%.

The above presented physical picture explains the most typical features of the ionospheric storm morphology. It is generally consistent with the scenario presented by *Rees* [1995]. However, to explain some peculiarities of the storms, especially the ones during pronounced magnetic events, some other mechanisms are considered.

In the case of positive phases, which appear sometimes before the beginning of the corresponding magnetic disturbance, the above scheme does not work because there is still neither the depleted $[O]/[N_2]$ nor the storm-induced circulation. It was suggested by *Danilov and Belik* [1991, 1992] that the effect may be related to some other channel of penetration of the disturbed solar wind energy to ionospheric heights rather than auroral precipitation. The channel was assumed to be soft particle precipitation in the region of the dayside cusp. The cusp is the only formation which starts to respond to the coming geomagnetic disturbances before any geomagnetic index does: the cusp begins to move equatorward a few hours before the beginning of the *Dst* depletion [*Danilov and Belik,* 1992].

The reason of the $F2$-peak vertical drift may be also electric fields penetrating from the magnetosphere. There is some evidence [*Mikhailov and Foster*, 1997] that strong enough electric fields (up to 70–100 mV m^{-1}) may sometimes penetrate to middle latitudes to account for the positive phase as has been suggested by some authors. However, the electric field plays an important role in maintaining the $F2$ layer at equatorial latitudes and a dominant role in geomagnetic storm effects at equatorial latitudes.

In quiet daytime conditions the E_y component of the electric field, which participates in the $\mathbf{E} \times \mathbf{B}$ drift, is eastward. In most cases during magnetic disturbances, E_y decreases in the equatorial zone, and this leads to a reduction of the upward drift and so to the positive phase [*Mikhailov et al.,* 1994; *Prölss*, 1995]. However, during severe disturbances there may happen an increase of E_y with a corresponding increase of the upward drift and corresponding negative phase [*Adeniyi*, 1986; *Danilov and Belik*, 1991; *Turunen and Rao*, 1980]. Evidently, it was the case with the September 23, 1973, storm at the equator considered in detail by *Mikhailov et al.* [1994].

Strong magnetic fields are often observed in polar regions in the zone of Joule heating and field-aligned current dissipation. Due to the magnetic field geometry, they do not produce such strong vertical drift as at low latitudes, but they are able to influence the $F2$-layer behavior via the recombination coefficient. The rate of the $O^+ + N_2$ reaction, which provides an important step in the recombination chain, depends strongly on the electric field [*Schunk et al.,* 1975], and thus an increase of the field should lead to a depletion of electron concentration, i.e., to the negative phase. This, evidently, was the case over the EISCAT installation during the strong magnetic disturbance on April 3, 1990, when a strong electric field (85 mV m^{-1}) was observed [*Mikhailov and Schlegel,* 1998].

Some positive effects in the electron concentration during geomagnetic disturbances may be due to the so-called travelling atmospheric disturbances (TAD) [*Prölss*, 1995, *Mikhailov et al.,* 1995]. The energy injected into the upper atmosphere during magnetic storms may generate TADs, which are pulse-like atmospheric perturbations, which propagate equatorward with the velocity of a few hundred meters per second and may exceed by a factor of 3–5 the velocity of moderate equatorward winds (about 150 m s^{-1}). At middle latitudes the TADs would produce an increase of the vertical drift and so the positive phase in the same way as the "smooth" storm-induced circulation but of much shorter duration.

Summarizing, we may draw the following picture of the ionosphere and thermosphere response at $F2$-layer heights to geomagnetic disturbances:

1. During such disturbances a large amount of energy is deposited into the thermosphere at high latitudes. This leads, first of all, to an increase of the neutral gas temperature T and variations of the neutral composition (decrease of the atoms-to-molecules ratio). Both factors contribute to a decrease of the electron concentration (the negative phase of an ionospheric storm) in the high-latitude ionosphere.

2. The energy deposition produces also a strong enough storm-induced circulation which is directed equatorward and may coincide or conflict with the quiet-time circulation. Depending on this, the gas with depleted $[O]/[N_2]$ is either brought toward low latitudes (and so the negative phase extends equatorward) or "locked" in the higherlatitude thermosphere.

3. The storm-induced circulation increases the plasma upward vertical drift in the $F2$ layer and so leads to an uplifting of the layer and an increase of the electron concentration. This is the most frequent cause of the positive phase at middle latitudes. Sometimes (especially at lower latitudes), the downwelling of the circulation leads to an increase of the atomic oxygen concentration and so to the positive phase.

4. Strong electric fields of magnetospheric origin may penetrate to F-region heights during geomagnetic disturbances. These fields evidently contribute substantially to strong positive and negative phases (depending on the sign of the E_y zonal component) of ionospheric storms observed in the equatorial ionosphere. Moreover, such fields, during prominent magnetic events, may be responsible for strong negative phases observed immediately after the magnetic disturbance SC.

5. Due to the energy injection into the thermosphere, travelling atmospheric disturbances are generated in the high-latitude thermosphere and propagate equatorward with the velocity much higher than the velocity of the regular meridional circulation. They may be responsible for short-period positive phases observed in the daytime at middle latitudes.

2.3. Modeling and Predictions

The modeling and predictions of geomagnetic storm effects on the F-region ionosphere develop in two ways: (1) improvement of physical, semiempirical, and empirical models and (2) exploitation of real-time data from the present-day and future solar-terrestrial monitoring systems, which among others allows application of neural networks for predictions [e.g., *Cander and Mihajlovic*, 1998].

A comparison of several physical models with experimental data [*Anderson et al.,* 1998] showed that all models generally described well basic features of the $F2$ layer ionosphere in quiet conditions, but each had some deficiency. Moreover, the main problem of modeling, and particularly predicting the geomagnetic storm effects, is insufficient knowledge and/or accuracy of observational inputs into physical models, namely of various forms of incoming external energy [e.g., *Codrescu et al.,* 1997]. Thus physical models provide insight into physical processes associated with geomagnetic storm effects but do not provide sufficiently accurate predictions. One more problem is that various ionospheric storms may have significantly different mechanisms, as shown for instance by EISCAT observations [*Mikhailov and Schlegel*, 1998]. Empirical or semiempirical models do not provide sufficiently accurate predictions as well.

The very recent development of prediction models is based on a new idea: first, to establish the background variation from monthly medians or other median or average values, either from physical models or from statistical or other models, and then to scale these values by an empirical model to storm-time conditions [*Fuller-Rowell et al.,* 1998; *Muhtarov et al.,* 1998]. Such an approach, among others, makes it possible to overcome the problem of somewhat different ionospheric effects of geomagnetic storms in different solar cycles [e.g., *Field and Rishbeth*, 1997].

3. $F1$ Layer

When we go down from the $F2$-layer maximum, the effect of geomagnetic storms in the neutral thermosphere becomes less dramatic. The influence of ionization and photochemistry processes on the ionospheric storm becomes more important, primarily due to shorter lifetime of free electrons in the more dense atmosphere. The $F1$ layer ionosphere is a region where both the neutral atmosphere changes (dominant in the $F2$ layer) and the ionization rate and photochemistry changes (dominant in the lower ionosphere) play an important role.

The thermosphere remains substantially disturbed. For instance, measurements of the total mass densities near

200 km by the SETA satellite reveal an increase of the total neutral density under daytime conditions at latitudes 60°–80° by about 50–70% with a penetration of a substantial increase almost to the equator in the summer hemisphere, when geomagnetic activity increases from quiet ($Kp \sim 1-2$) to disturbed ($Kp \sim 4-7$). This increase is about one half in the winter hemisphere with much less equatorward penetration, mainly due to the prevailing summer-to-winter background meridional wind [*Forbes et al.,* 1996].

The $F1$-layer ionosphere response to geomagnetic storms has been much less studied than that in the $F2$ layer. The electron density in the F/E region (∼100–200 km) is known generally to increase during magnetically disturbed nights [e.g., *Rowe*, 1974], but this increase substantially depends on altitude, somewhere the effect being almost none or even slightly opposite. An interesting result is that a remarkable spring/autumn asymmetry has been found, while in autumn in the altitude range 160–190 km, the effect of geomagnetic storms is not detectable, in spring it is observed in a well-pronounced form down to about 180 km, and only near and below 170 km it becomes nondetectable [*Buresova and Mosert de Gonzalez*, 1999].

An interesting effect at middle latitudes is filling in the valley between the E and the F regions, as it is illustrated, for instance, by rocket measurements at Wallops Island [*Geller et al.,* 1975] or ground-based measurements at Arecibo [*Rowe,* 1974].

Recently, *Bauske et al.* [1997] demonstrated effects of another source contributing to enhanced nighttime ionization in the E and $F1$ layers during geomagnetic storms at low and middle latitudes, precipitation of neutralized ring current particles (mainly hydrogen) in the energy range 1–100 keV.

4. *E* Layer

The ionospheric E region is a region where both the neutral atmosphere changes and the ionization rate and photochemistry changes play an important role in the response to geomagnetic storms. Their relative importance depends on altitude and latitude. The electron density in the E region at high latitudes generally increases as a response to geomagnetic storms, but at middle and low latitudes, the effect is opposite. The storm effects at high latitudes are remarkably better developed at night, when the corpuscular ionization source competes with much weaker noncorpuscular ionization sources than during daytime.

4.1. High Latitudes

The high-latitude ionosphere is a rather complicated and changeable formation. The ionospheric behavior in each particular point depends on many factors, such as its position, local and universal time, magnetic activity, the sign of various components of the interplanetary magnetic field etc. Any attempt to describe even briefly this problem would be out of the scope of this paper. We just state that all said above about the polar ionosphere variability on the whole is true for the high-latitude E region in particular.

There are several formations or subregions within the high-latitude ionosphere. However, for the purposes of this paper it is sufficient to consider two principal regions in which the E-region response to geomagnetic disturbances is different.

Fluxes of energetic electrons precipitating from the Earth's radiation belts are typical for the auroral oval. The intensity of these fluxes during magnetic storms and substorms may increase by several orders of magnitude [*Avakyan et al.,* 1994, *Hardy et al.,* 1985]. This results in an increase of the electron concentration in the entire E region. Even during moderate substorms, the increase in the peak of the E layer may be by a factor of 2–2.5, and during strong enough magnetic storms, it may exceed an order of magnitude [*Troshichev*, 1986]. For example, according to direct rocket measurements of *Zhlud'ko and Chasovitin* [1983] at Heiss Island under solar zenith angle of 96°, the value of $[e]$ at 120 km increased from 1×10^4 cm^{-3} under moderate disturbance with $Kp = 3+$ to 1.5×10^5 cm^{-3} during magnetic storm with $Kp = 6+$. In some cases the electron concentration in the E-region peak may reach $(4-6) \times 10^5$ cm^{-3} [*Anderson et al.,* 1995; *Troshichev,* 1986].

An increase of geomagnetic activity leads to intensification of the irregular structure of the auroral E region [e.g., *Blagoveshchensky et al.*, 1983; *Ogawa and Igarashi*, 1982]. These irregularities also are clearly seen at the electron concentration vertical profiles obtained by rockets or incoherent scatter installations [*Troshichev,* 1986].

Sporadic E_s layers in the high-latitude ionosphere are a special problem. The wind shear mechanism responsible for formation of E_s layers at middle latitudes is less effective at high geomagnetic latitudes. However, the occurrence frequency of E_s in the auroral oval, is much higher than at middle latitudes and increases with geomagnetic activity. There are several particular types of E_s characteristic for auroral oval, but often they are just called "auroral E_s" [*Troshichev*, 1986]. The occurrence of auroral E_s correlates with the intensity of aurora, because it is associated with penetration of particles of similar energies. The region of maximum probability of auroral E_s occurrence has a crescent-type shape and is located approximately in the 2000–0600 MLT sector of the auroral oval. The real picture of the auroral E_s occurrence distribution is rather complicated and depends on many factors, including the solar activity. Following the auroral oval, the region of the highest probability of E_s occurrence expands during geomagnetic disturbances to lower geomagnetic latitudes. The intensity of auroral E_s may be very high, and the corresponding screening frequencies may reach 5 to 7 MHz.

In contrast to the situation with the auroral oval, the response of the polar cap E layer to geomagnetic disturbances is relatively weak. The shape of electron concentration vertical profiles is very similar to that in the midlatitude E layer. Neither is there any significant increase in the probability of E_s layer occurrence. The screening frequency of this layer (so-called "plain" type contrary to the "slant" types typical for the auroral oval) is within 1.0–1.4 MHz both in quiet

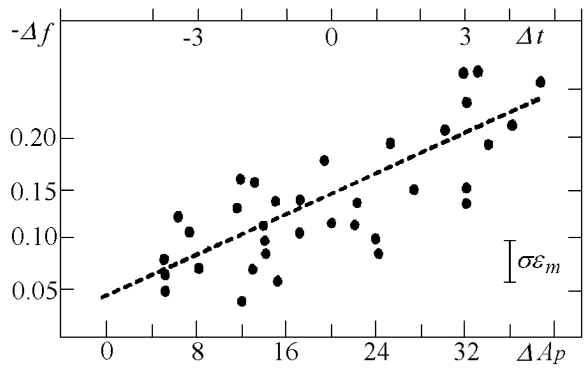

Figure 6. Depletion of the E-region critical frequency Δf (Moscow station) versus the increase ΔAp of the Ap index during the 3 days studied as compared with the previous 3 days [from *Antonova et al.,* 1996].

and in disturbed conditions. Moreover, there are some indications [*Troshichev*, 1986] to the presence in the polar cap of a negative relation between E_s occurrence frequency and geomagnetic activity.

The main cause of the above described differences in the E region response to geomagnetic activity in the auroral oval and the polar cap is the difference in precipitating particle fluxes. The electrons intruding the auroral region have energies above 1 keV, and the most efficient in the E region are those with the energy of 1.2–1.4 keV. They produce the extra ionization in the entire E region and may create also narrow (sporadic) layers of ionization if the fluxes are rather monoenergetic. Since the intensity of the fluxes increases strongly with magnetic activity [*Avakyan et al.,* 1994; *Hardy et al.,* 1985], there is a strong response of the E layer to magnetic disturbances.

In the polar cap the energy of the precipitating electrons is much lower (below 600 eV), and they cannot penetrate the atmosphere as deep as to the E layer. This explains the relative week response of the polar cap E region to geomagnetic disturbances. Formation of sporadic E_s layers in the polar cap is evidently controlled by the magnetospheric convection [*Besprozvannaya et al.,* 1983; *Troshichev*, 1986], the low velocities of the convection being more favorable for E_s occurrence.

The increased ionization in the auroral E region leads to an intensification of nitric oxide formation. As a result, [NO] in the auroral E region may be by 1 or 2 orders of magnitude higher than at the same altitudes in the midlatitude E region. The latter fact has been confirmed experimentally [e.g., *Solomon and Barth*, 1999; *Torr et al.*, 1995] and is also manifested in variations of ion composition. The $[NO^+]/[O_2^+]$ ratio in the auroral E region is sometimes an order of magnitude higher than at middle latitudes [*Danilov*, 1994]. During and after geomagnetic storms the excess of NO molecules is brought by the horizontal winds to middle latitudes and may lead there to the increase of the effective recombination coefficient and small decrease of $[e]$ in the E region, as described below.

4.2. Middle and Low Latitudes

The most pronounced effect in the midlatitude E layer is the depletion of the electron concentration in its maximum (or, which is the same, of f_oE). The effect was for the first time described in detail by *Beynon and Brown* [1959] and then discussed by many authors (for references, see the monograph by *Antonova et al.* [1996]). The magnitude of the effect is usually small (about 8–10%), so special procedures (smoothing, running means, regression equations) are usually applied to reveal the effect reliably. *Antonova et al.* [1994] analyzed 37 years of f_oE observations at the Moscow station. They have found that the mean value of the correlation coefficient R_A between the specially calculated and filtered index of the E-layer I_E and the Ap geomagnetic index was -0.38 ± 0.02 at the entire temporal interval considered with $\sigma(R) < 0.09$. The mean value of the coefficient K in the regression equation

$$I_E = KAp + L$$

was found to be -0.57 ± 0.03. Here L is a constant which is independent of the Ap index but may change with season.

An analysis of the time delay of the ionospheric response relative to geomagnetic disturbances was based on the position of the maximum of the cross-correlation function between I_E and Ap. The delay was found to be 0.74 ± 0.07 days. This means that the maximum effect in f_oE is observed by about 18 hours after the corresponding maximum in Ap.

Analyzing the reaction of f_oE to individual geomagnetic disturbances, *Ivanov-Kholodny et al.* [1991] showed that at least 80–90% of geomagnetic disturbances with the magnitude of $\Delta Ap > 15$ (ΔAp is the total excess in daily Ap during the 3 days analyzed relatively to the previous 3 days) are clearly manifested in the f_oE behavior (see Figure 6). Using the superposed epoch method, *Ivanov-Kholodny et al.* [1991] found that there is a depletion of f_oE after geomagnetic disturbances. The average amplitude of the depletion is 0.07 ± 0.02 MHz, and its maximum lags the maximum of the disturbance (maximum in Ap) by 0.5 days. This result qualitatively agrees with the first estimates by *Beynon and Brown* [1959].

There is a latitudinal effect in the E-region response to geomagnetic disturbances [*Antonova et al.,* 1996]. The magnitude of the negative effect in f_oE (or $[e]$) increases with geographic latitude φ. The mean depletion of $[e]$ in the $\varphi < 40^\circ$ zone is $2.0 \pm 0.3\%$, and outside the zone it is $3.3 \pm 0.4\%$. The difference in the magnitudes exceeds 2σ and is significant. The magnitude of the ionospheric response at $\varphi > 60^\circ$ is about twice that at $\varphi > 20^\circ$ [*Antonova et al.,* 1996].

There is no common agreement on the physical mechanisms responsible for the aforementioned depletion of f_oE and $[e]$ after magnetic disturbances. Initially, two mechanisms were proposed: variations of the neutral density and changes of the nitric oxide concentration. The density variation required is by a factor of 2–3, which seems unreal. The increase of [NO] should enhance the $[NO^+]/[O_2^+]$ ratio and the effective recombination coefficient α_r which would result in a depletion of $[e]$. To provide the observed decrease of f_oE and [e] after geomagnetic disturbances, one needs an increase of [NO] by a factor of 1.5–2. Such an increase does not seem unreal, because there are indications of an increase of [NO] in the E region in geomagnetically disturbed conditions [*Cravens and Stewart,* 1978]. The advection of the nitric oxide molecules from the high-latitude ionosphere (where extra amount of NO is formed as a result of corpuscular ionization) may be the source of these increases [*Ivanov-Kholodny and Nusinov*, 1979]. Some role in the effective recombination coefficient increase may also be played by vibrationally excited NO^+ ions, the concentration of which may increase during disturbances because of the temperature increase [*Antonova and Ivanov-Kholodny,* 1989].

Morton and Matthews [1993] observed at Arecibo a strong effect of geomagnetic storms on sporadic E layers of the tidal ion layer type. During a very strong storm they found even complete disruption of tidal ion layers in the whole E region.

5. Lower Ionosphere

The lower ionosphere (h < 100 km) responds very dramatically to geomagnetic storms [e.g., *Lastovička*, 1988, 1996]. Its electron concentration is considerably enhanced, particularly in the auroral zone, which results in a large increase of radio wave absorption and, eventually, in the disappearance of radio signal in MF/HF ranges. This enhancement of electron density is caused by a strong increase of precipitation of energetic particles, mainly electrons of energies of tens to hundreds kiloelectron volts.

There are two different types of the lower ionosphere response to geomagnetic storms, as shown in Figure 7 in terms of the response of radio wave absorption or $f_{\min}$ ($f_{\min}$ is a minimum reflected frequency at vertical ionospheric sounding used as an indirect measure of absorption). The first type, observed at high latitudes (Kiruna) and marked PSE I (poststorm effect, phase I), consists in a large increase of electron concentration and thus of radio wave absorption, which is coincident with the geomagnetic storm and caused by direct energetic electron injections from the magnetosphere into the auroral ionosphere. At middle latitudes (A3-MF, A3-LF data) we observe both phases of the PSE, the direct phase PSE I and the delayed phase PSE II (aftereffect). The PSE I phase, coincident with the geomagnetic storm, is considerably weaker than at auroral latitudes. The delayed phase PSE II, which is usually well separated from PSE I, may last up to 10 days and is usually the dominant phase of the PSE (although the difference between PSE I and PSE II magnitude is usually much less pronounced than in Figure 7). Subauroral latitudes (Uppsala) represent a transition zone from the auroral-type effect to the midlatitude type of the effect.

The PSE II is caused by a considerable increase of the pitch-angle diffusion, which forces the trapped energetic electrons, injected during the PSE I, to move into the loss cone and to precipitate. This mechanism works in the slot region between outer and inner radiation belts. The increase of the pitch-angle diffusion after the storm is caused by a

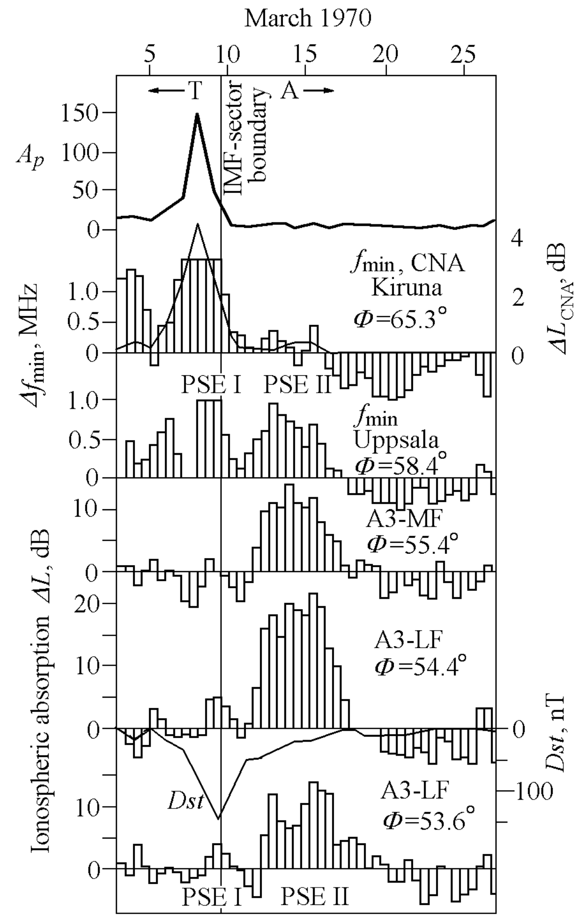

Figure 7. Effect of a geomagnetic storm (Ap, Dst) in March 1970 on the lower ionosphere ($\Delta f_{\min}$, $\Delta L - A3$ radio wave absorption) at auroral and middle latitudes in Europe [after *Lauter and Bremer*, 1983].

considerable increase of the ELF-VLF wave activity (mainly emissions such as plasmaspheric hiss) after the storm and motion of electrons into the loss cone via wave-particle interactions.

The low-latitude boundary of the geomagnetic storm effects on the lower ionosphere seems to be about 35°–37°N, but the effects mostly do not reach latitudes much below 50°N (geomagnetic coordinates). The latitudinal boundary between the high-latitude direct effect and the midlatitude PSE II occurrence, which roughly coincides with the equatorward boundary of the auroral oval, varies substantially during geomagnetic storm periods. It had been used for tracking the motion of the equatorward boundary of the auroral oval by *Wagner and Ranta* [1983].

Geomagnetic storms affect considerably the lowermost part of the ionosphere and thus the ELF/VLF radio propagation in the Earth–ionosphere waveguide. Geomagnetic storms improve ELF/VLF radio propagation [e.g., *Satori*, 1991].

It should be mentioned that there is a region of anomalously intense particle precipitation, the South Atlantic mag-

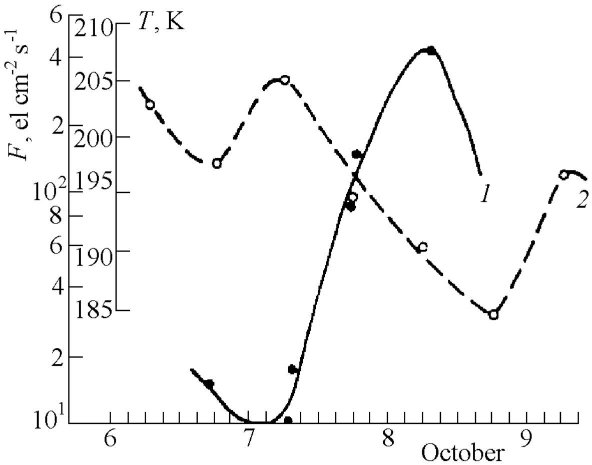

Figure 8. Opposite variations of the energetic electron flux (1, solid line) and the middle mesosphere temperature (2, dashed line) on October 6–9, 1971, as measured by rockets at Volgograd [after *Butko et al.*, 1974].

netic anomaly. The effects of geomagnetic storms and particle precipitation in the South Atlantic anomaly are stronger than those at respective middle and moderate latitudes of the northern hemisphere.

There are also some changes in atmospheric electricity at lower ionosphere heights during geomagnetic storms. According to rocket measurements by *Zadorozhny et al.* [1994], the mesospheric maximum of electric field intensifies and its altitude increases with increasing geomagnetic activity at high latitudes, while there is little change at middle latitudes.

6. Middle Atmosphere

The energy of geomagnetic storm-related precipitating particles is lost not solely through ionization. It causes also excitation, heating and dissociation processes. As a consequence of this energy deposition, various effects have to exist in the neutral middle atmosphere. Such effects occur in temperature, composition, wind field, atmospheric emissions (particularly auroras), the tropopause height and temperature, and in the turbopause. *Lastovička* [1988, 1989, 1996] briefly reviewed these effects. The most important ones, those in the temperature, composition and winds, are shortly described in the following sections.

6.1. Temperature

Any energy deposition should result in increasing temperature. *Lastovička* [1988, 1989] summarized various results of rocket soundings at Heiss Island, Volgograd, and Wallops Island and results of some other experiments. The strongest effect of geomagnetic storms was observed at high latitudes. However, the storm effect pattern does not differ qualitatively at high and middle latitudes, the difference is only quantitative. The results make it possible to suggest the following scheme of the geomagnetic storm effects on temperature in the middle atmosphere: (1) lower thermosphere and upper mesosphere – heating; (2) middle mesosphere (~70 km), cooling; (3) lower mesosphere (~60 km), moderate heating; and (4) upper stratosphere, positive but marginal correlation.

Figure 8 shows an evident, well-developed anticorrelation of the energetic electron flux and temperature in the middle mesosphere as measured by rockets at Volgograd. The anticorrelation means cooling of the middle mesosphere during geomagnetic storms. This anticorrelation is caused by changes in atmospheric composition. The overproduction of NO_y and HO_x during geomagnetic storms as a consequence of enhanced particle precipitation results in the production

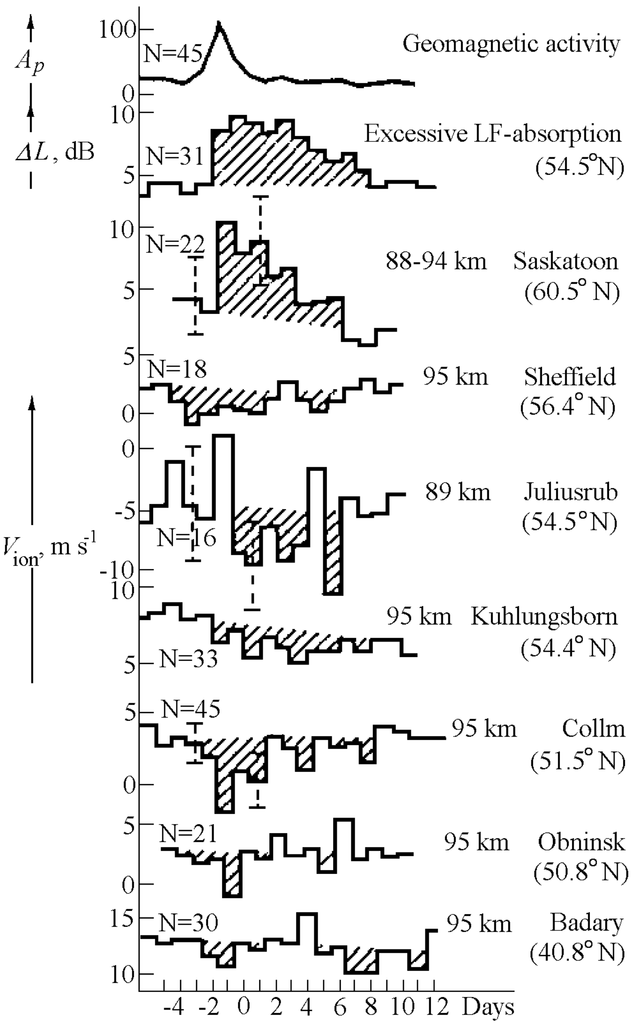

Figure 9. Superposed-epoch analysis of the zonal wind field during geomagnetic disturbances (day zero corresponds to the maximum of *Ap*). Stations near latitude 50°N are arranged according to their geomagnetic latitude [after *Singer et al.*, 1994].

of odd nitrogen and odd hydrogen, which destroy ozone. The ozone loss results in a decrease of solar radiation absorption with a consequent cooling of the middle mesosphere, which is stronger than the direct heating by precipitating particles. This temperature decrease in the middle mesosphere means that the downward propagation of heat from the lower thermosphere/upper mesosphere to the stratosphere is inhibited and cannot contribute to geomagnetic storm effects in the troposphere/lower stratosphere.

6.2. Wind

Lastovička [1988, 1996] summarized older results on the geomagnetic storm/activity effects on winds in the upper mesosphere and lower thermosphere of high and higher middle latitudes. The existence of such effects at lower altitudes is questionable, if they occur at all. The effects occur in the prevailing wind (including zonal wind reversal in some events) as well as in tidal components. There is an apparent difference between North America (weaker effects) and Europe (stronger effects).

Singer et al. [1994] analyzed long-term wind measurements at several stations near 50°N. Figure 9 shows a westward wind acceleration of zonal wind near 95 km for all stations in Europe (geomagnetic latitude $\Phi \sim 50°$N), whereas the wind acceleration is eastward at Saskatoon (Canada, $\Phi \sim 60°$N) as a response to strong geomagnetic storms. This confirms the difference between Europe and North America, which has tentatively been interpreted in terms of different geomagnetic latitudes for the same geographic latitudes. The sign of the effect in Saskatoon did not change in the whole height range studied (~80–100 km). *Singer et al.* [1994] found an even stronger wind response in summer than in winter, partly due to a larger masking by internal atmospheric disturbances (e.g., by stratospheric warmings). The observed effects in tidal components and in the prevailing meridional wind were markedly smaller than those in the prevailing zonal wind. *Fahrutdinova et al.* [1998] observed at Kazan an evident dependence of the geomagnetic storm effect on altitude with a general tendency to westward and northward acceleration. *Salah et al.* [1996] found no effect of a geomagnetic storm of June 1991 on winds measured at 80–100 km over Millstone Hill and Durham (both 43°N), maybe as a consequence of lower latitude.

At high latitudes, *Price and Jacka* [1991] and *Price et al.* [1991] found in Antarctica a statistically significant correlation of the horizontal wind with a local K index down to 86 km. The horizontal wind and its oscillations in the ranges of 1–3 and 2–6 hours were stronger and accompanied by a rather strong upward vertical wind under high geomagnetic activity conditions. Meteor radar wind measurements at the South Pole by *Portnyagin et al.* [1997] confirm the upwelling as a response to strong geomagnetic storms.

6.3. Composition

An important part of the upper middle atmosphere response to geomagnetic storms are changes of minor component composition, particularly of NO, as a consequence of energetic particle impact on production of atomic nitrogen. Due to the quasi-continuous particle penetration in the auroral zone as a result of magnetospheric activity (substorms, etc.), the NO concentration in the upper middle atmosphere is, similarly to the lower thermosphere (see Section 4.2), remarkably higher at high latitudes than at middle latitudes, and it further decreases toward low latitudes. This was confirmed by rocket and satellite observations, for example references summarized by *Rusch and Clancy* [1987] and *Lastovička* [1988]. The vertical and horizontal transport affect the NO distribution and partly diminishes the difference between high and mid-latitude NO concentration. A reduction of ozone concentration in the upper mesosphere is expected. This is much better pronounced and well documented for the solar proton events.

7. Lower Atmosphere

When we go down from the upper and middle stratosphere into the troposphere and partly into the lower stratosphere, the effect of geomagnetic storms reappears but as an effect of different morphology and origin (see Section 7.4.), which is not caused by particle precipitation. The most promising mechanism at these altitudes seems to be at present the Tinsley's hypothesis of "electrofreezing," based on the geomagnetic storm modulation of the cosmic ray flux with subsequent modification of global electric circuit and intracloud electric and nucleation microprocesses (see Section 7.3).

7.1. Troposphere

The field of Sun-weather relationships, and particularly, possible relationships between geomagnetic storms activity and tropospheric processes, has been controversial for many decades. Researchers have been divided into "believers" and "nonbelievers." Typical arguments of non-believers may be found, for instance, in the works of *Pittock* [1978] or *Salby and Shea* [1991]. A comprehensive analysis of the reliability of methods of investigation and results was carried out by *Taylor* [1986] for the interplanetary magnetic field (IMF) sector boundary effects in the vorticity area index (VAI). Further on, we present some of the observational results (mainly correlations) of believers in support of reality of the effects of geomagnetic storms in the troposphere.

A review of earlier observational correlations was made by *King* [1975]. *Roberts and Olson* [1973] observed an effect of geomagnetic storms on the wintertime 300 hPa trough development in the North Pacific-North America area. A decrease of surface pressure after strong sporadic geomagnetic storms, developed particularly in the northern Atlantic/European and eastern Siberian/Aleutian sectors, was reported in several papers of the group of Mustel [e.g., *Mustel et al.,* 1977]. These two regions were confirmed by *Smirnov* [1984] to be the most sensitive areas of the northern hemisphere troposphere to solar-geomagnetic forcing. *Bucha* [1991] observed

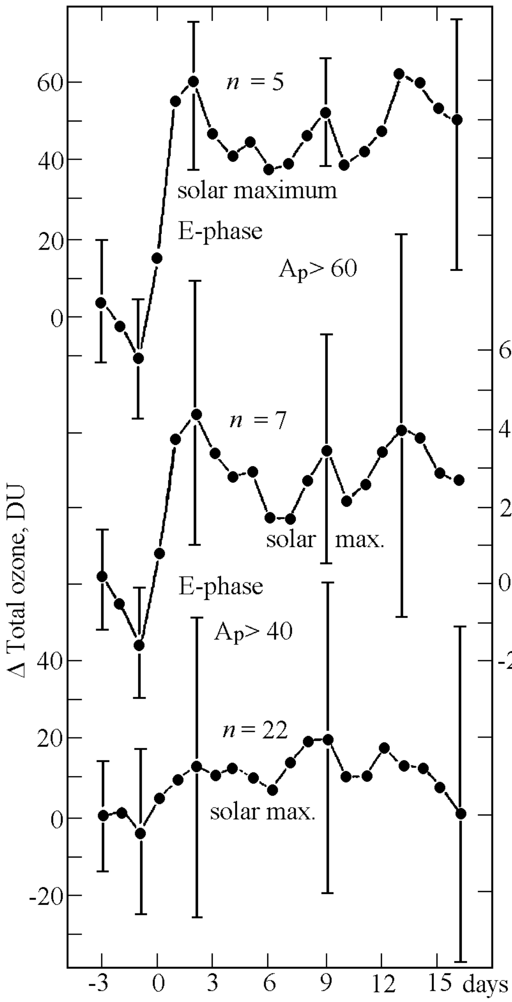

Figure 10. Total ozone (Europe, midlatitudes) deviations from average level over days −6 to zero for winter major geomagnetic storms under high solar activity conditions. All events (bottom curve), events under the E phase of QBO (middle curve), very strong storms ($Ap > 60$) under the E phase of QBO (top curve). Vertical lines show error bars; n is the number of events. Period studied is 1963–1988 (25 years) [after *Lastovička et al.*, 1992].

a decrease of surface air pressure in the northern Atlantic, deepening of the Icelandic low, a considerable zonalization of the 500 hPa circulation, and related changes of temperature over the northern Atlantic and Europe as a consequence of geomagnetic storms; the effect was developed evidently better in winter than in summer. *Bucha and Bucha* [1998] summarized results of many Bucha's papers and specified geomagnetic storm effects on the surface temperature in various areas of the northern hemisphere middle and high latitudes, which differ even in their sign as a consequence of pressure field and circulation changes. *Padgaonkar and Arora* [1981] found an evident geomagnetic storm effect in the 500 hPa tropospheric vorticity area index, which is to some extent a measure of the degree of disturbed state of the troposphere.

Bochnicek et al. [1996] looked for regions of significant positive or negative deviations of winter temperatures in the northern hemisphere from long-term averages as caused by geomagnetic storms. They found such deviations to be principally dependent on the phase of the quasi-biennial oscillation (QBO). Their distribution was very macroregional: there were regions of continental size or ever larger which displayed either positive or negative deviations. These macroregions were not organized according to latitude or longitude. *Bochnicek et al.* [1998] obtained similar results also for wind fields in the troposphere.

We can establish three typical features of the effects of geomagnetic storms on the northern hemisphere troposphere.

1. Tropospheric effects are very (macroregional) regional, the nature of this being probably the principal role of circulation changes and orography. The two most sensitive regions appear to be the north Atlantic/European region and the east Siberian/Aleutian region.
2. Tropospheric effects are much better developed in winter than in summer half of the year, i.e., in the period of lower direct solar energy input and less stable atmosphere.
3. Tropospheric effects in winter appear to be significantly dependent on the phase of QBO.

7.2. Total Ozone

Lastovička et al. [1992] and *Lastovička and Mlch* [1999] summarized various results of other authors on the effects of geomagnetic storms upon the total ozone content and found that these results did not provide a consistent pattern of the effect. In order to clarify this apparent inconsistency and to obtain a more global pattern of the total ozone response, *Lastovička et al.* [1992], *Mlch* [1994], and *Mlch and Lastovička* [1995] investigated the total ozone response to geomagnetic storms at higher midlatitudes of the northern hemisphere. These investigations resulted in a consistent pattern of the response.

Figure 10 shows how the effect in Europe near 50°N becomes more pronounced with more favorable conditions. If we take all major geomagnetic storms under high solar activity together, no significant effect is found in total ozone. If we select only events under the E-max conditions (solar activity maximum and the east QBO phase), the peak value on day +2 reaches 45 DU (Dobson units), and it begins to be statistically significant. After turning to solely very strong storms ($Ap > 60$), the effect of storms in total ozone in midlatitudinal Europe becomes statistically quite significant, it peaks at 60 DU on day +2 (values after days +6/+7 are "artificially" increased by the occurrence of other major storms). Very strong geomagnetic storms in winter under E-max conditions occur very rarely. On the other hand, all such events show a very persistent general pattern, quite close to the average one shown in Figure 10.

First, results of European ground-based measurements over 25 years were studied [*Lastovička et al.,* 1992]. Then TOMS measurements along 40°, 50°, and 60°N and related behavior of circulation parameters/indices were analyzed [*Mlch*, 1994; *Mlch and Lastovička*, 1995]. The results of *Lastovička et al.* [1992], *Mlch* [1994], *Mlch and Lastovička*

[1995], and *Lastovička and Mlch* [1999] may be summarized as follows:

1. There is an effect of geomagnetic storms in total ozone at northern higher middle latitudes. However, it occurs very rarely, only under specific conditions. A significant and persistent response of total ozone near 50°N to geomagnetic storms has been observed only for strong storms, in winter, and under solar activity maximum/E-QBO conditions. The last condition is equivalent to high solar activity/no major stratospheric warming condition, which is probably physically more relevant. No significant and persistent effect has been found under other conditions and/or near 60°N and 40°N.

2. Just after a strong storm, we observe a significant increase of total ozone in the European sector due to a substantial smoothing of longitudinal variation of total ozone. Such a variation is very weak in summer, and thus there is no effect in summer. For a similar reason there is no effect in zonal mean total ozone.

3. There are two sectors sensitive to geomagnetic storms both in total ozone and in tropospheric parameters: (1) Northeastern Atlantic plus European sector, and (2) Eastern Siberian plus Aleutian sector.

4. The total ozone response to major geomagnetic storms is probably caused by storminduced changes in circulation. Changes of the circulation pattern agree (at least qualitatively) with changes in total ozone.

7.3. Mechanisms

There is no generally accepted mechanism for tropospheric effects of geomagnetic storms. The observed cooling in the middle mesosphere [e.g., *Lastovička*, 1996] and absence of the geomagnetic storm effect in the middle stratospheric temperature exclude the possibility of direct downward transport of heat from the greatly heated auroral thermosphere down to the troposphere. The agent responsible for the tropospheric effects must basically skip across the stratosphere. Only two agents fulfill this request, the galactic cosmic ray flux modulated by the geomagnetic storm and the global electric circuit and/or atmospheric electricity affected by in situ changes of conductivity and by ionospheric/magnetospheric electric fields and currents. The mechanism must include triggering and/or amplification phenomena, because the energy of storm-related atmospheric processes is by several orders of magnitude higher than the input energy of solar wind/geomagnetic storm origin.

No hypothesis based on direct changes of atmospheric electric field (e.g., that by *Markson* [1979]) has passed successfully through quantitative tests and calculations to show that such a mechanism can explain observations.

Recently, Tinsley [*Tinsley*, 1996, 1997; *Tinsley and Dean*, 1991] developed the electrofreezing-based hypothesis: solar wind changes, which cause also geomagnetic storms, modulate the cosmic ray flux; the modulated galactic cosmic ray flux changes the (upper) tropospheric ion production and thus the conductivity; simultaneously, the electric potential difference between the ionosphere and the Earth is changed due to the IMF forcing at high latitudes; the vertical air-Earth current is changed; in the presence of large-scale clouds this current determines the rate of polarization charging of clouds via accumulation of a positive electrostatic charge in droplets near cloud tops; this electrostatic charge influences substantially the rate of ice nucleation (electrofreezing); the rate of precipitation is changed; the net latent heat release and vertical motions are changed; atmospheric vorticity is changed; general circulation is changed. A large energy amplification is quite possible in this mechanism; for example, in the supercooled cloud tops it seems to be sufficient to transform a very small fraction of water molecules into ice configuration to freeze the whole droplet. This amplification might be up to more than 11 orders of magnitude [*Tinsley*, 1997]. The electrofreezing mechanism can be significantly affected by the presence of volcanic (or other) aerosols in the atmosphere. However, this promising hypothesis needs further laboratory and model support and observational evidence.

In this context it may be of some interest that *Veretenenko and Pudovkin* [1997] found observationally an evident influence of the galactic cosmic ray flux on the solar radiation input into the lower stratosphere at higher latitudes (about 60°–80°) via changes of high level clouds (cirrus).

7.4. Difference Between the Effects in the Troposphere and in the Upper Middle Atmosphere

When we go down from the upper middle atmosphere (upper mesosphere and lower thermosphere), the effect of geomagnetic storms weakens, becomes statistically insignificant, and can be hardly traced in the middle stratosphere (at least at middle latitudes). Farther down, it reappears in the troposphere and partly in the lower stratosphere. However, this is a different effect. These differences in geomagnetic storm effects were studied by *Lastovička* [1997] with the following results: (1) essentially, no regionality, only dependence on (geomagnetic) latitude in the upper middle atmosphere versus strong regionality in the troposphere; (2) little seasonal dependence in the upper middle atmosphere versus strong seasonal dependence in the troposphere; (3) no significant dependence in the upper middle atmosphere on the QBO versus a strong dependence on the QBO in the troposphere (at least in winter); (4) the effect in the upper middle atmosphere is caused primarily by the energetic particle precipitation; in the troposphere the origin is not reliably known, energetic particles are not responsible for the observed effects, possible modulation of galactic cosmic rays and atmospheric electricity plays a role; (5) triggering/amplification phenomena has to play a principal role in the troposphere, while they are unnecessary to account for the effects in the upper middle atmosphere.

Thus we can summarize that the effects of geomagnetic storms in the upper middle atmosphere and the troposphere are of different morphology, origin, and nature.

8. Conclusion: "Height Profile" of the Effects of Geomagnetic Storms

In the $F2$ layer of the ionosphere the midlatitude response (both its positive and negative phase) to a geomagnetic storm is basically an ionospheric response to storm-induced changes in the neutral atmosphere. These changes, namely changes in composition and winds, are primarily a consequence of a strong Joule heating of the (auroral) thermosphere by storm-related electric currents. At lower heights, such a neutral atmosphere effect becomes less important, mainly due to shorter electron lifetime, which increases the role of ionization and photochemistry processes. At the base of the $F1$ layer ($h \sim 160-170$ km) the storm effect becomes very weak, sometimes being absent. Farther down, the most pronounced feature is the filling in of the valley between the E and the F regions, which is caused predominantly by particle precipitation. At E-region maximum and below, the nitric oxide concentration increases due to enhanced production by precipitating particles, and the temperature increases as well. There is a slight decrease of critical frequency f_oE at middle latitudes, even though below and above the E-layer maximum, the electron density increases. In the high-latitude ionosphere a significant enhancement of the electron concentration is observed.

Farther down, in the lower ionosphere, a large increase of electron density, particularly at night, is observed at auroral and higher middle latitudes as a consequence of a very large increase of energetic particle precipitation. In the neutral lower thermosphere and mesosphere we observe some effects of enhanced particle precipitation, but these effects weaken with decreasing altitude. They become insignificant in the upper stratosphere and absent in the middle stratosphere. Farther down, in the troposphere and partly in the lower stratosphere, the effect of geomagnetic storms reappears but as an effect of different morphology and origin, which is not caused by energetic particle precipitation. The most promising mechanism at these altitudes seems to be at present the Tinsley's hypothesis of "electrofreezing," based on the geomagnetic storm modulation of the cosmic ray flux with subsequent modification of global electric circuit and intracloud electric and nucleation microprocesses.

Thus we find at least three altitudinal regions of distinctly different geomagnetic storm-related processes responsible for observed effects:

1. $F2$ layer, where the ionospheric effect is basically a response to storm-time changes of the neutral thermosphere caused primarily by Joule heating.
2. The lower ionosphere and upper middle atmosphere, where the effects are caused by storm-related injections and precipitation of energetic particles of magnetospheric origin.
3. The lower atmosphere, where the effect has different morphology as well as mechanisms, which is possibly related to changes in galactic cosmic ray flux and atmospheric electricity.

The effects of a geomagnetic storm are generally strongest in the auroral zone, their amplitude weakens toward middle latitudes, some of them disappear at low latitudes, but some of them reappear or strengthen near the geomagnetic equator, namely effects in the F region.

References

Adeniyi, J. O., Magnetic storm effect on the morphology of the equatorial $F2$ layer, *J. Atmos. Sol. Terr. Phys., 48*, 695, 1986.

Anderson, P. C., et al., The ARIA I rocket campaign, *J. Geophys. Res., 100* (A9), 17,265, 1995.

Anderson, D. N., et al., Intercomparison of physical models and observations of the ionosphere, *J. Geophys. Res., 103*, 2179, 1998.

Antonova, L. A., and G. S. Ivanov-Kholodny, *Solar Activity and Ionosphere (at Altitudes 100–200 km)* (in Russian), 168 pp., Nauka, Moscow, 1989.

Antonova, L. A., G. S. Ivanov-Kholodny, and V. E. Chertoprud, Manifestation of solar and geomagnetic effects in long series of observations of the f_oE critical frequency, *Geomagn. Aeron.* (in Russian), *34*, 103, 1994.

Antonova, L. A., G. S. Ivanov-Kholodny, and V. E. Chertoprud, *Aeronomy of the E layer* (in Russian), 166 pp., Yanus, Moscow, 1996.

Avakyan, S. V., A. I. Vdovin, and V. F. Pustarnakov, *Ionizing and Penetrating Radiation in the Near-Earth Space* (in Russian), 501 pp., Gidrometeoizdat, St. Petersburg, Russia, 1994,

Bauske, R., S. Noel, and G. W. Prölss, Ionospheric storm effects in the nighttime E region caused by neutralized ring current particles, *Ann. Geophys., 15*, 300, 1997.

Besprozvannaya, A. S., V. M. Uvarov, A. V. Shirochkov, and T. I. Shchuka, Anomalous events in the ionospheric E region in the polar caps, *Ionos. Issled.* (in Russian), *37*, 91, 1983.

Beynon, W. J. G., and G. M. Brown, Geomagnetic distortions of the E region, *J. Atmos. Sol. Terr. Phys., 14,* 50, 1959.

Blagoveshchensky, D. V., N. F. Blagoveshchenskaya, and O. M. Pirog, Dynamics of parameters of the irregular structure in the auroral ionosphere during substorms, *Ionos. Issled.* (in Russian), *37*, 80, 1983.

Bochnicek, J., V. Bucha, P. Hejda, and J. Pycha, Relation between Northern Hemisphere winter temperatures and geomagnetic or solar activity at different QBO phases, *J. Atmos. Sol. Terr. Phys., 58*, 883, 1996.

Bochnicek, J., P. Hejda, V. Bucha, and J. Pycha, Temperature-pressure deviations and prevailing winds in the NH winter troposphere in years of high and low geomagnetic activity, *Ann. Geophys., Suppl., 16,* C887, 1998.

Bucha, V., Solar and geomagnetic variability and changes of weather and climate, *J. Atmos. Sol. Terr. Phys., 53*, 1161, 1991.

Bucha, V., and V. Bucha Jr., Geomagnetic forcing of changes in climate and in the atmospheric circulation, *J. Atmos. Sol. Terr. Phys., 60*, 145, 1998.

Buonsanto, M., Millstone Hill incoherent scatter F region observation during the disturbances of June 1991, *J. Geophys. Res., 100*, 5743, 1995.

Buresova, D., and M. Mosert de Gonzalez, $F1$ region electron density during geomagnetic storms as measured at Pruhonice, in *Proceedings of the IRI Task Force Activity 1998,* Int. Center Theor. Phys. (ICTP), Trieste, 1999.

Butko, A. S., I. N. Ivanova, and G. A. Kokin, Some features of thermal and circulation regime of the atmosphere, in *Solnechno-Atmosfernye Svyazi* (in Russian), pp. 125–131, Gidrometeoizdat, Leningrad, 1974.

Cander, L. R., and S. J. Mihajlovic, Forecasting ionospheric structure during the great geomagnetic storms, *J. Geophys. Res., 103*, 391, 1998.

Codrescu, M. V., T. J. Fuller-Rowell, and I. S. Kutiev, Modeling the F layer during specific geomagnetic storms, *J. Geophys. Res., 102*, 14,315, 1997.

Cravens, T. E., and A. I. Stewart, Global morphology of nitric oxide in the lower E region, *J. Geophys. Res., 83*, 2446, 1978.

Danilov, A. D., Ion chemistry of the lower thermosphere, *J. Atmos. Sol. Terr. Phys., 56,* 1213, 1994.

Danilov, A. D., and L. D. Belik, Thermospheric-ionospheric interaction during ionospheric storms, *Geomagn. Aeron.* (in Russian), *31* (2), 209, 1991, Danilov, A. D., and L. D. Belik, Thermospheric composition and the positive phase of an ionospheric storm, *Adv. Space Res., 12*(10), 257, 1992.

Danilov, A. D., and L. D. Morozova, Ionospheric storms in the $F2$ region, The morphology and physics (a review), *Geomagn. Aeron.* (in Russian), *25*, 705, 1985,

Duncan, R. A., F-region seasonal and magnetic storm behavior, *J. Atmos. Sol. Terr. Phys., 31*, 59, 1969.

Fahrutdinova, A. N., O. N. Sherstyukov, and S. V. Maksyutin, The effect of geomagnetic activity on the dynamics of upper mesosphere-lower thermosphere and parameters of E layer, paper presented at the 32nd COSPAR Scientific Assembly, Comm. on Space Res., Nagoya, Japan, 1998.

Field, P. R., and H. Rishbeth, The response of the ionospheric $F2$ layer to geomagnetic activity: An analysis of worldwide data, *J. Atmos. Sol. Terr. Phys., 59*, 163, 1997.

Förster, M., A. A. Namgaladze, and R. Y. Yurik, Thermospheric composition changes deduced from geomagnetic storm modeling, *Geophys. Res. Lett., 26*, 2625, 1999.

Forbes, J. M., R. Gonzalez, F. A. Marcos, D. Revelle, and H. Parish, Magnetic storm response of lower thermosphere density, *J. Geophys. Res., 101*, 2313, 1996.

Fuller-Rowell, T. J., E. Araujo-Pradere, and M. V. Codrescu, An empirical ionospheric storm-time correction model, paper presented at the 32nd COSPAR Scientific Assembly, Comm. on Space Res., Nagoya, Japan, 1998.

Geller, M. A., L. G. Smith, and H. D. Voss, Analysis of nighttime E-region winds and ionization production, *Radio Sci., 10*, 335, 1975.

Hardy, D. A., M. S. Gussenhoven, and E. A. Holeman, A statistical model of auroral electron precipitation, *J. Geophys. Res., 90*, 4229, 1985.

Ivanov-Kholodny, G. S., and A. A. Nusinov, *Formation and Dynamics of the Daytime Midlatitude Ionospheric E layer* (in Russian), 128 pp., Gidrometeoizdat, Moscow, 1979.

Ivanov-Kholodny, G. S. , G . S. Stankevich, and V. E. Chertoprud, Solar and geomagnetic activity effects in observations of the f_oE critical frequency, *Geomagn. Aeron.* (in Russian), *31*, 848, 1991,

King, J. W., Solar-weather relationships, *Aeron. Astronaut., 13*, 10, 1975.

Lastovička, J., A review of solar wind and high energy particle influence on the middle atmosphere, *Ann. Geophys., 6*, 401, 1988.

Lastovička, J., Solar wind and high energy particle effects in the middle atmosphere, *Handb. MAP, 29*, 119, 1989.

Lastovička, J., Effects of geomagnetic storms in the lower ionosphere, middle atmosphere and troposphere, *J. Atmos. Sol. Terr. Phys., 58*, 831, 1996.

Lastovička, J., Effects of geomagnetic storms – different morphology and origin in the upper middle atmosphere and the troposphere, *Stud. Geophys. Geod., 41*, 73, 1997.

Lastovička, J., and P. Mlch, Is ozone affected by geomagnetic storms?, *Adv. Space Res., 24*(5), 631, 1999.

Lastovička, J., J. Bremer, and M. Gil, Ozone response to major geomagnetic storms, *Ann. Geophys., 10,* 683, 1992.

Lauter, E. A., and J. Bremer, Post-storm effects, *HHI-STP Rep.*, No. 17, pp. 65–70, Henrich Hertz Inst., Berlin, 1983.

Loewe, C. A., and G. W. Prölss, The importance of vibrationally excited molecular nitrogen for the negative phase of ionospheric storms, paper presented at the 32nd COSPAR General Assembly, Comm. on Space Res., Nagoya, Japan, 1998.

Markson, R., Atmospheric electricity and the sun-weather problem, in *Solar-Terrestrial Influences on Weather and Climate*, edited by B. M. McCormick and T. A. Seliga, pp. 215–232, D. Reidel, Norwell, Mass., 1979.

Mebagishvili, N. N., and G. Khochalava, Negative stage of ionospheric storms, *Proc. Georgian Acad. Sci.* (in Russian), *75*, 77, 1974,

Mikhailov, A. V., and J. C. Foster, Daytime thermosphere above Millstone Hill during severe geomagnetic storms, *J. Geophys. Res., 102*, 17,275, 1997.

Mikhailov, A. V., and M. Förster, Day-to-day thermosphere parameter variation as deduced from Millstone Hill incoherent scatter radar observations during March 16–22, 1990 magnetic storm period, *Ann. Geophys., 15*, 1429, 1997.

Mikhailov, A., and K. Schlegel, Physical mechanism of strong negative storm effect in the daytime ionospheric $F2$ region observed with EISCAT, *Ann. Geophys., 16*, 602, 1998.

Mikhailov, A. V., M. Förster, and M. G. Skoblin, Neutral gas composition changes and $\mathbf{E} \times \mathbf{B}$ vertical plasma drift contribution to the daytime equatorial $F2$ region storm effects, *Ann. Geophys., 12*, 226, 1994.

Mikhailov, A. V., M. G. Skloblin, and M. Förster, Daytime $F2$ layer positive storm effect at middle and lower latitudes, *Ann. Geophys. 13*, 532, 1995.

Mikhailov, A. V., Yu. L. Terekhin, and V. V. Mikhailov, Does the $F2$ layer follow variations of the constant pressure level?, *Geomagn. Aeron.* (in Russian), *29*, 1035, 1989.

Mlch, P., Total ozone response to major geomagnetic storms during non-winter period, *Stud. Geophys. Geod., 38*, 423, 1994.

Mlch, P., and J. Lastovička, Total ozone response to major geomagnetic storms and changes in meteorological situations, *Stud. Geophys. Geod., 39*, 189, 1995.

Morton, Y. T., and J. D. Mathews, Effects of the 13–14 March 1989 geomagnetic storm on the E region tidal ion layer structure at Arecibo during AIDA, *J. Atmos. Sol. Terr. Phys., 55*, 467, 1993.

Muhtarov, P., L. Cander, M. Levy, and I. Kutiev, Application of the geomagnetically correlated statistical model to short-term forecasting of f_oF2, in *Proceedings of the 2nd COST 251 Workshop, Algorithms and Models for COST 251 Final Product, 30–31 March 1998, Side, Turkey,* p. 241, Rutherford Appleton Lab., U. K., 1998.

Mustel, E. R., V. E. Chertoprud, V. A. Kovedeliani, Comparison of changes of the field of surface air pressure in the periods of high and low geomagnetic activity, *Astron. Zh.* (in Russian), *54,* 682, 1977,

Ogawa, T., and K. Igarashi, VHF radar observations of auroral E region irregularities associated with moving arcs, in *Proceedings of the Fourth Symposium on Coordinated Observations of the Ionosphere and Magnetosphere in the Polar Regions,* p. 125, Nat. Inst. of Polar Res., Tokyo, 1982.

Padgaonkar, A. D., and B. R. Arora, Tropospheric vorticity responses to the solar magnetic sector structure and geomagnetic disturbances, *Pure Appl. Geophys., 119*, 893, 1981.

Pavlov, A. V., The role of vibrationally excited nitrogen in the formation of the midlatitude negative ionospheric storms, *Ann. Geophys., 12*, 554, 1994.

Pavlov, A. V., and M. J. Buonsanto, Comparison of model electron densities and temperatures with Millstone Hill observations during undisturbed period and the geomagnetic storms of 16–23 March and 6–12 April 1990, *Ann. Geophys., 15*, 327, 1997.

Pittock, A. B., A critical look at long-term sun-weather relationships, *Rev. Geophys.,16,* 400, 1978.

Portnyagin, Y. I., J. M. Forbes, and N. A. Makarov, Unusual characteristics of lower thermosphere prevailing winds at South Pole, *Geophys. Res. Lett., 24*, 81, 1997.

Price, G. D., and F. Jacka, The influence of geomagnetic activity on the upper mesosphere/lower thermosphere in the auroral zone, I, Vertical winds, *J. Atmos. Sol. Terr. Phys., 53*, 909, 1991.

Price, G. D., F. Jacka, R. A. Vincent, and G. B. Burns, The influence of geomagnetic activity on the upper mesosphere/lower thermosphere in the auroral zone, II, Horizontal winds, *J. Atmos. Sol. Terr. Phys., 53*, 923, 1991.

Prölss, G. W., Magnetic storm associated perturbations of the upper atmosphere: Recent results obtained by satellite-borne gas analyzers, *Rev. Geophys., 18*, 183, 1980

Prölss, G., Ionospheric F-region storms, in *Handbook of Atmospheric Electrodynamics,* vol. 2, edited by H. Volland, pp. 195–248, CRC Press, Boca Raton, F/a., 1995.

Prölss, G. W., and U. von Zahn, ESRO 4 gas analyzer results, 2, Direct measurements of changes in the neutral composition during an ionospheric storm, *J. Geophys. Res., 79,* 2335, 1974.

Rees, D., Observations and modeling of ionospheric and thermospheric disturbances during major geomagnetic storms: A review, *J. Atmos. Sol. Terr., Phys., 57*, 1433, 1995.

Rishbeth, H., and D. W. Barron, Equilibrium electron distribution in the ionospheric $F2$ layer, *J. Atmos. Sol. Terr. Phys., 18*, 234, 1960.

Roberts, W. O., and R. H. Olson, Geomagnetic storms and wintertime 300 mb trough development in the North Pacific–North America area, *J. Atmos. Sci., 30*, 135, 1973.

Rowe, F. J., Jr., Magnetic activity variations of the nighttime E region at Arecibo, *Radio Sci., 9*, 175, 1974.

Rusch, D. W., and R. T. Clancy, Minor constituents in the upper stratosphere and mesosphere, *Rev. Geophys., 25*, 479, 1987.

Salah, J. E., W. Deng, and R. R. Clark, Observed response of the Earth's lower thermosphere to a major geomagnetic storm, *Geophys. Res. Lett., 23*, 575, 1996.

Salby, M. L., and D. J. Shea, Correlations between solar activity and the atmosphere: An unphysical explanation, *J. Geophys. Res., 96*, 22,579, 1991.

Satori, G., Combined ionospheric effect due to Forbush decreases and magnetospheric high-energy particles at mid-latitudes, *J. Atmos. Sol. Terr. Phys., 53*, 325, 1991.

Schunk, R. W., W. J. Raitt, and P. M. Banks, Effects of electric fields on the daytime high-latitude E and F regions, *J. Geophys. Res., 80,* 3121, 1975.

Seaton, M. J., A possible explanation of the drop in F region critical densities accompanying major ionospheric storms, *J. Atmos. Sol. Terr. Phys., 8*, 122, 1956.

Singer, W., et al., Geomagnetic influences upon tides-winds from MLT radars, *J. Atmos. Sol. Terr. Phys., 56*, 1301, 1994.

Smirnov, R. V., Spatial dependence of impacts of solar activity on the troposphere, *Astron. Zh.* (in Russian), *61*, 1168, 1984.

Solomon, S. C., and C. A. Barth, Auroral production of nitric oxide measured by the SNOE satellite, *Geophys. Res. Lett., 26*, 1259, 1999.

Szuszczewicz, E. P., M. Lester, P. Wilkinson, P. Blanchard, M. Abdu, R. Hanbaba, K. Igarashi, S. Pulinets, and B. M. Reddy, A comparative study of global ionospheric responses to intense magnetic storm conditions, *J. Geophys. Res., 103,* 11,665, 1998.

Taylor, H. A., Selective factors in sun-weather research, *Rev. Geophys., 24*, 329, 1986.

Tinsley, B. A., Solar wind modulation of the global electric circuit and apparent effects on cloud microphysics, latent heat release, and tropospheric dynamics, *J. Geomagn. Geoelectr., 48*, 165, 1996.

Tinsley, B. A., Do effects of global atmospheric electricity on cloud cause climate change?, *Eos Trans. AGU, 78*, 341, 1997.

Tinsley, B. A., and G. W. Dean, Apparent tropospheric response to MeV–GeV particle flux variations: A connection via electronucleation of ice in high-level clouds?, *J. Geophys. Res., 96,* 22,283, 1991.

Torr, M. R., D. G. Torr, T. Chang, P. Richards, W. Swift, and N. Li, Thermospheric nitric oxide from ATLAS 1 and Spacelab 1 missions, *J. Geophys. Res., 100,* (A9), 17,389, 1995.

Troshichev, O. A., *Ionospheric-Magnetospheric Disturbances at High Latitudes* (in Russian), 256 pp., Gidrometeoizdat, Moscow, 1986.

Turunen, T., and M. N. Rao, Examples of the influence of strong magnetic storms on the equatorial F layer, *J. Atmos. Sol. Terr. Phys., 42*, 323, 1980.

Veretenenko, S. V., and M. I. Pudovkin, Effect of the galactic cosmic ray variations on the solar radiation input in the lower atmosphere, *J. Atmos. Sol. Terr. Phys., 59,* 1739, 1997.

Wagner, C.-U., and H. Ranta, The midlatitude post-storm electron precipitation belt, *J. Atmos. Sol. Terr. Phys., 45,* 811, 1983.

Zadorozhny, A. M., A. A. Tyutin, O. A. Bragin, and V. N. Kikhtenko, Recent measurements of atmospheric electric fields and related parameters, *J. Atmos. Sol. Terr. Phys., 56*, 321, 1994.

Zevakina, R. A., Ionospheric disturbances, in *Ionospheric Disturbances and Their Impact on Radio Communication* (in Russian), p. 3, Nauka, Moscow, 1971.

Zhlud'ko, A. D., and Yu. K. Chasovitin, Coordinated rocket studies of the ionosphere at various latitudes, *Proc. Inst. of Exp. Meteorol.* (in Russian), *15,* 72, 1982.

A. D. Danilov, Institute of Applied Geophysics, 9 Rostokinskaya Str., Moscow 129128, Russia. (danilov@wdcb.ru)

J. Lastovička, Institute of Atmospheric Physics, Prague, Czech Republic. (jla@ufa.cas.cz)

(Received July 5, 1999; revised July 28, 2000;
accepted November 21, 2000)