Ann. Geophysicae 16, 602–608 (1998) © EGS – Springer-Verlag 1998

Annales Geophysicae

Physical mechanism of strong negative storm effects in the daytime ionospheric F2 region observed with EISCAT

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Received: 14 January 1997 / Revised: 2 December 1997 / Accepted: 12 December 1997

Abstract. A self-consistent method for daytime F-region modelling was applied to EISCAT observations during two periods comprising the very disturbed days 3 April 1992 and 10 April 1990. The observed strong N_e decrease at F2-layer heights originated from different physical mechanisms in the two cases. The negative F2-layer storm effect with an N_mF2 decrease by a factor of 6.4 on 3 April 1992 was produced by enhanced electric fields ($\boldsymbol{E} \approx 85$ mV/m) and strong downward plasma drifts, but without any noticeable changes in thermospheric parameters. The increase of the $O^+ + N_2$ reaction rate resulted in a strong enrichment of the ionosphere with molecular ions even at F2-layer heights. The enhanced electric field produced a wide mid-latitude daytime trough on 03 April 1992 not usually observed during similar polarization jet events. The other strong negative storm effect on 10 April 1990 with a complete disappearance of the F2-layer maximum at the usual heights was attributed mainly to changes in neutral composition and temperature. A small value for the shape parameter S in the neutral temperature profile and a low neutral temperature at 120 km indicate strong cooling of the lower thermosphere. We propose that this cooling is due to increased nitric oxide concentration usually observed at these heights during geomagnetic storms.

Key words. Atmospheric composition and structure · Thermosphere · Ionosphere · Ion chemistry and composition · Atmosphere interactions

1 Introduction

Ionospheric F2-layer negative storms have been studied for years and general mechanisms of their origin are well known. A comprehensive review on this problem can be found in Prölss (1995). So far mid- and low-latitude F2-layer storm effects have been studied more extensively than high-latitude ones. This is partly due to the difficulties with ground-based ionosonde observations during geomagnetically disturbed periods. In addition, the high-latitude F2 region is very variable, being strongly influenced by magnetospheric processes; in particular, substantial electric fields are usually present during geomagnetic storms. These electric fields and the corresponding horizontal $\boldsymbol{E} \times \boldsymbol{B}$ plasma drifts can strongly perturb the electron density distribution at F2-layer heights. Neutral composition and temperature changes are the other source of negative F2-layer storm effects. The aim of the present work is to study the physical mechanism and to estimate the contribution of various processes to the observed storm-time $N_e(h)$ changes. Two periods of EISCAT observations comprising the very disturbed days 03 April 1992 and 10 April 1990 with a very strong electron concentration depletion are analyzed in the paper.

These cases were chosen to demonstrate different physical mechanisms of the observed negative storm effect. The first case presents an isolated geomagnetic storm with a moderate Ap = 32 but a sharp and very strong upsurge in the westward $\boldsymbol{E} \times \boldsymbol{B}$ plasma drift around noon followed by a N_mF2 decrease of a factor of 6.4 relative to the quiet prestorm day 02 April 1992. The other case of 10 April 1990 refers to an extremely severe geomagnetic storm with Ap = 124 when the F2-layer maximum practically disappeared at the usual heights and an electron concentration peak around 200 km constituted the ionospheric maximum. Similar effects were observed in the American longitudinal sector with the Millstone Hill incoherent-scatter radar on 10 April 1990 (Buonsanto *et al*., 1992): The daytime ionospheric densities were extremely low over the probed latitude range, and since the peak height was in the molecular ion-dominated region below 200 km, the F2-layer maximum was in fact a F1-layer peak. The storm effects at Millstone Hill were studied by Mikhailov and Foster (1997).

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The EISCAT observations are analyzed in this paper with the help of the self-consistent method developed by Mikhailov and Schlegel (1997). It provides the set of main aeronomic parameters ([O], [O_2], [N_2], vertical plasma drift W, exospheric temperature T_{ex}, and the shape parameter S for the neutral temperature height profile, which are altogether responsible for the formation of the $N_e(h)$ profile) and yields information about the physical mechanisms of the observed F2-layer storm effects.

The main aim of Mikhailov and Schlegel (1997) was to demonstrate that our method is very effective and yields valuable information about the ionosphere and thermosphere. In this investigation we apply our model to two specific examples of ionospheric storm events. Moreover, as a new feature in this study, neutral temperature at 120 km height was included to the set of fitted parameters, as it may differ from the values provided by the MSIS model during severe storm periods.

Our original method can be used when the F2-layer maximum is well defined. In this case some of the parameters can be found from fitting theoretical N_mF2 and h_mF2 values to the observed ones. On 10 April 1990 when the F2-layer maximum was absent, a different approach was used to find the solution. In this case all six aeronomic parameters T_{ex}, S, T_{120}, [O], [O_2], [N_2] were varied to fit the observed $N_e(h)$ profile. The vertical plasma drift was calculated as in the previous method using the observed $V_l(h)$ and other parameters calculated in a self-consistent way in the $N_e(h)$ fitting procedure.

2 Observations and model calculations

A list of the available observations along with solar and magnetic indices used in the study is given in Table 1. The 3-month average solar index F10.7 was 151.7 and 187.9 for the two periods in question.

The April 1992 period is very suitable for the analysis as it includes three quiet days preceding the storm, so 2 April can be used as a quiet reference day. For the second period we do not have a similar quiet reference, but Millstone Hill radar observations for the same period (Mikhailov and Foster, 1997) show that the F2 layer was only moderately disturbed on 9 April 1990 with N_mF2 reduced by a factor of 2 with respect to the nearest quiet day 7 April.

Table 1. Dates and geophysical parameters of the periods used in the study

date	time (UT)	Ap	F10.7 (day/day-1)	$\boldsymbol{E} \times \boldsymbol{B}$ (m/s)
31 March 1992	1300–1400	14	191.4/182.4	250
1 April 1992	1300–1400	13	186.1/191.4	150
2 April 1992	1300–1400	6	161.2/186.1	50
3 April 1992	1400–1500	32	159.7/161.2	1700
9 April 1990	1400–1500	34	146.3/151.9	500
10 April 1990	1315–1415	124	148.7/146.3	700

The EISCAT CP-1 program provides range profiles of N_e, T_e, T_i and V_l with the antenna beam directed along the local geomagnetic field line. We used the long-pulse data for our purpose in the form as distributed by EISCAT (integration time 5 minutes, standard profile of [O^+]/N_e). We calculated median profiles over about an hour (12–13 values, see Table 1) for each of the four parameters with standard deviations at each height. Data with finer height resolution from an alternating code pulse scheme for $\boldsymbol{E}$ and lower F regions were analyzed as well, but they were not used in the study (see Sect. 3).

On 3 April 1992 a steep upsurge of the westward $\boldsymbol{E} \times \boldsymbol{B}$ drift started after 1000 UT, peaked around 1200–1400 UT at values of about 2000 m/s, and then subsided down to 1000 m/s. Therefore we accepted a mean value of 1700 m/s ($\boldsymbol{E} \approx 85$ mV/m) for the analyzed period 1400–1500 UT. On 10 April 1990, strong $\boldsymbol{E} \times \boldsymbol{B}$ drifts were observed during the whole day. A northwest drift of 700 m/s ($\boldsymbol{E} \approx 35$ mV/m) was observed during the analyzed period 1315–1415 UT.

In Figs. 1 and 2 we present observed N_mF2 and h_mF2 variations during daytime hours for the two periods. The disturbed day 3 April is characterized by an increase in h_mF2 and a strong decrease in N_mF2 in comparison with the previous quiet days. Such h_mF2 and N_mF2 variations are similar to mid-latitude negative F2-layer storm effects resulting from changes of neutral composition and temperature (see for instance Prölss, 1995). In

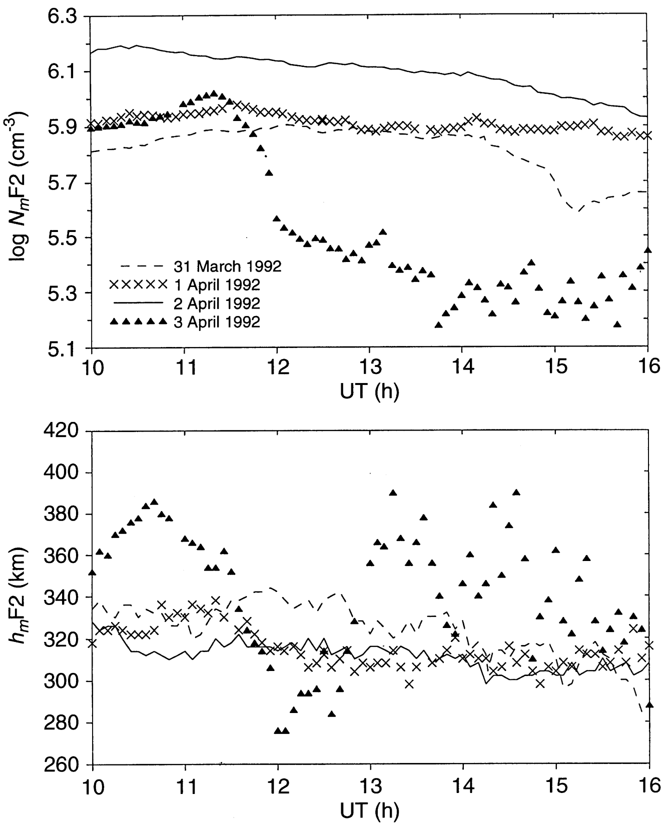

Fig. 1. Observed N_mF2 and h_mF2 variations for 31 March to 3 April 1992 during daytime hours

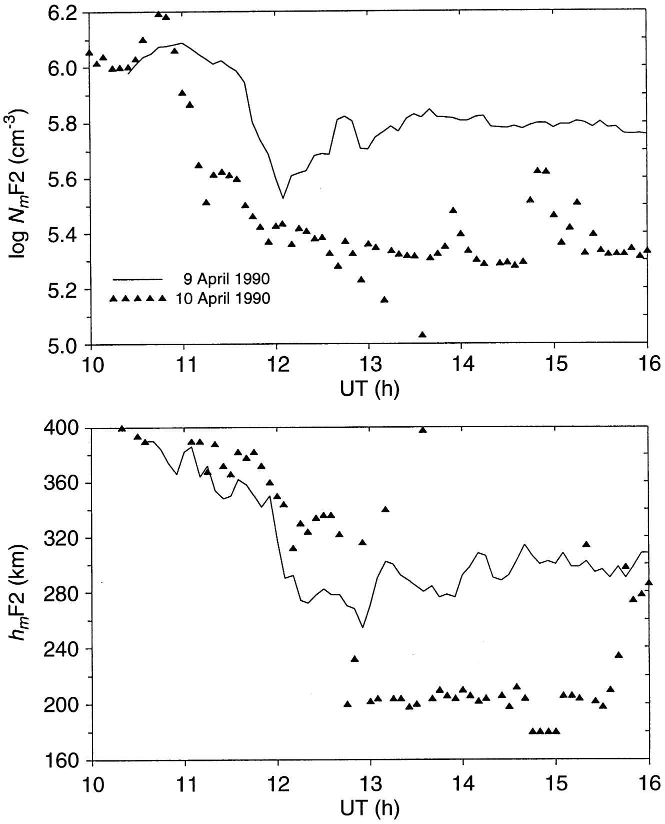

Fig. 2. Same as Fig. 1 but for the period 9–10 April 1990

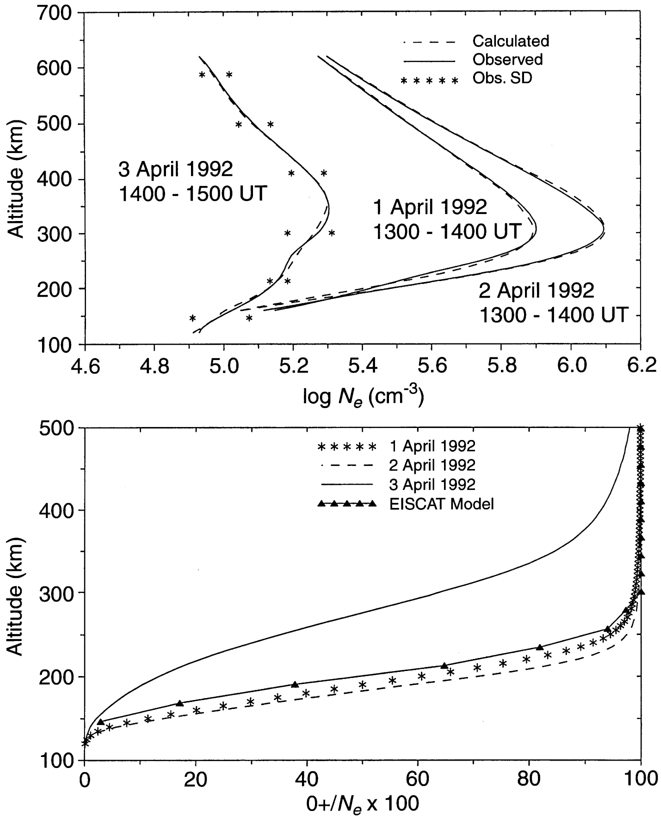

Fig. 3. *Top panel*: observed (together with a band of standard deviations) and calculated $N_e(h)$ profiles for two geomagnetically quiet (1, 2 April) and disturbed (3 April 1992) days. *Bottom panel*: calculated and EISCAT model $[O^+]/N_e$ height variations for the 3 days

this case N_mF2 was reduced by a factor of 6.4, and h_mF2 increased by about 40 km. On 10 April 1990 a large decrease in both N_mF2 and h_mF2 was observed. Figure 2 shows a normal negatively disturbed F2 layer around 1230 UT which is then split into two layers. After 1300 UT the ionosphere was characterized by one layer with a peak around 200 km height.

Electron density profiles resulting from our model calculations in comparison with observed $N_e(h)$ profiles for the two periods in question are displayed in Figs. 3 and 4 (upper panels). The lower panels show a comparison of the calculated $[O^+]/N_e$ ratio with the EISCAT ion composition model used for the routine incoherent data reduction. The calculated height variations for $[O^+]/N_e$ are close to the standard EISCAT ion composition model for quiet days (1, 2 April 1992), and therefore no correction was applied for the measured $N_e(h)$, $T_e(h)$, and $T_i(h)$ profiles. For disturbed days, however, the calculated ion composition differs significantly from the standard EISCAT model. A strong enrichment of the ionosphere with molecular ions even at heights of the F2 layer required a correction of the measured $N_e(h)$, $T_e(h)$, and $T_i(h)$ profiles for these days as outlined in Mikhailov and Schlegel (1997).

3 Discussion

Figures 3 and 4 show that our model calculations describe the observed $N_e(h)$ profiles with good accuracy for quiet as well as for disturbed conditions. So we can conclude that the considered set of main physical processes in the model is sufficient for the description of the sunlit auroral F2 region even during such disturbed conditions. These are the same processes commonly used for mid-latitude F2-layer modelling. Similar conclusions were obtained by Farmer *et al.* (1984) and Lathuillere and Brekke (1985) analyzing EISCAT observations.

The observed negative storm effect on 3 April 1992 looks much like mid-latitude events with reduced N_mF2 and increased h_mF2 values, but the mechanism causing it turns out to be different. The calculated relative variations of the neutral composition during the 31 March - 3 April period (Fig. 5, left panel) are not strong. This may be attributed to the fact that the considered period (14–15 UT) was relatively close to the beginning of the geomagnetic storm at about 10 UT. Much more time is usually required for changes in the neutral composition. Nevertheless a pronounced difference in the shape of the $T_n(h)$ profile can be already seen for this period (Fig. 6, left panel): The calculated shape parameter is $S = 0.035$ km^{-1}, while MSIS-83 gives $S = 0.017$ km^{-1} as for the previous quiet days. The increased value of S results from the strong Joule heating which took place for the period in question. Figure 6 shows two $T_n(h)$ profiles together with the observed $T_i(h)$. The unusual $T_i(h)$

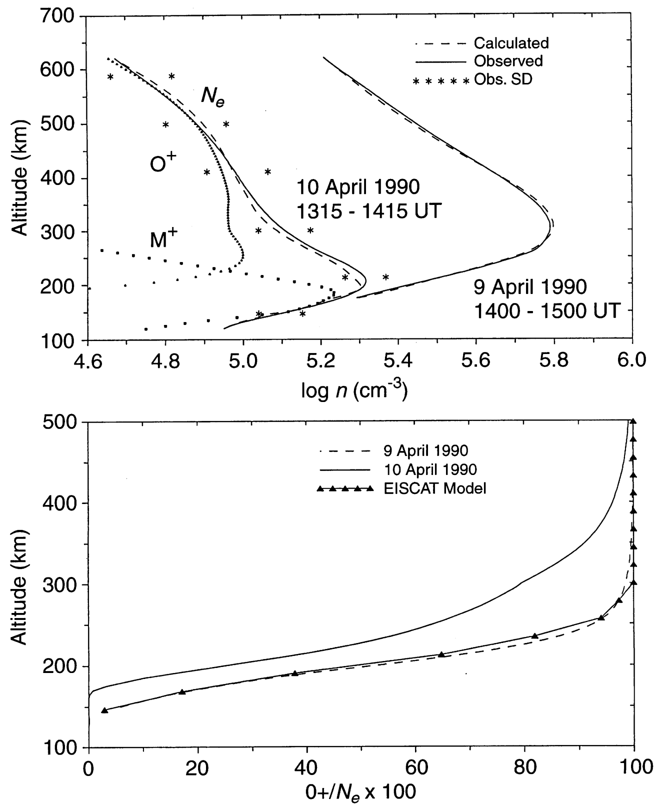

Fig. 4. Same as Fig. 3 but for 9–10 April 1990. Calculated atomic oxygen and molecular ion concentrations are shown as well (in the *upper panel*)

profile below 200 km clearly indicates the strong frictional heating in this height range. Our method is indeed sensitive enough to reproduce such changes in the shape of $T_n(h)$.

The observed strong N_mF2 reduction from 2 to 3 April 1992 (by a factor of up to 6.4) does not result from neutral composition changes: the O/N_2 ratio equals 5 on 2 April and 4.7 on 3 April at the h_mF2 height. Also, Fig. 5 reveals only small neutral composition changes at 300 km altitude. On the other hand, the linear loss coefficient $\beta = \gamma_1[N_2] + \gamma_2[O_2]$ for O^+ ions increased by a factor of 5 at the F2-layer maximum height. This results from a large increase in γ_1 by more than a factor of 10 due to its dependence on the electric field (Schunk *et al.*, 1975). Therefore, the very large negative F2-layer storm effect on 3 April 1992 is mainly a consequence of the strong electric field ($\boldsymbol{E} \approx 85$ mV/m), rather than of changes in neutral composition. In principle, an additional contribution to the N_mF2 reduction can originate from a strong downward plasma drift (Fig. 5, lower left panel). Under normal conditions the vertical plasma drift can be fully attributed to the meridional thermospheric wind V_{nx} (Farmer *et al.*, 1984), as the magnetic field line declination at Tromso is only 1.24° and the zonal wind component, V_{ny} is not efficient at least for near-noon hours when it is small. The calculated strong downward drift of 40 m/s on 3 April (Fig. 5, lower left panel) however, can hardly be attributed to a northward wind of ≈ 180 m/s, but may partly be explained by a geometrical effect proposed by Schlegel (1986): if the line-of-sight velocity measured by the radar and interpreted as V_l is not exactly parallel to $\boldsymbol{B}$ on a local scale (especially during geomagnetic disturbances), a component of the perpendicular ion velocity, V_{perp}, can map into the line-of-sight velocity. Even if the deviation from the parallel direction is only of the order of 1°, a large perpendicular ion velocity of 1700 m/s (as we have for 3 April) can easily give a contribution of several 10 m/s to the measured line-of-sight velocity, falsely assumed to be parallel to $\boldsymbol{B}$. Therefore we assume that the calculated strong downward drift on 3 April 1992 may have – at least partly – a geometrical rather than a geophysical explanation.

The observed steep upsurge of the electric field produced not only a negative disturbance at the latitudes of EISCAT, but also a daytime trough in the latitudinal distribution of N_mF2. Figure 7 shows the observed latitudinal variation of N_mF2 along a chain of European ionosonde stations for the same time-interval. The disturbance due to the high-latitude electric field is seen to extended down to latitudes of Uppsala ($\Phi = 58°$), and a small N_mF2 reduction is seen even at the mid-latitude station Kaliningrad. Therefore 3 April provides an excellent example of a daytime trough formation due to an enhanced electric field. The observed effect resembles a polarization jet first reported by Galperin *et al.* (1973) but is not identical to so-called SAID events [subauroral ion drift, see Anderson *et al.* (1991) for details].

The considered case of 3 April 1992 demonstrates a pure effect of increased electric field without any noticeable changes in thermospheric parameters, as the analyzed period was just at the beginning of the geomagnetic storm. The case of 10 April 1990 already represents the second day of a severe geomagnetic storm with strong worldwide F2-layer negative storm effects (e.g. Buonsanto *et al.*, 1992). A complete disappearance of the F2-layer maximum at the usual heights implies strong changes in the thermosphere, as was shown by Mikhailov and Foster (1997) using Millstone Hill observations for that day.

Figure 4 (top panel) shows that the observed ionospheric maximum around 200 km results from the superposition of two layers: a normal F1 layer at 175 km composed of molecular ions and a strongly reduced and distorted F2 layer composed of O^+ ions. Similar to the previous case, the ionosphere on 10 April 1990 is enriched with molecular ions even at F2-layer heights. As we do not have a real quiet reference day, we can only compare 10 April with 9 April. The latter looks like a normal day with undisturbed ion composition (Fig. 4, bottom), although in fact it was a disturbed day as well (Table 1). The analysis has revealed that the main reason for the strong F2-layer depletion is the decrease in atomic oxygen concentration by a factor of 4.3 at 300 km altitude, accompanied by increased $[N_2]$ and $[O_2]$ (Fig. 5, top right). Both, the increase in $[N_2]$ and $[O_2]$ and in γ_1 due to enhanced temperature, and a high electric field result in an increment in the linear loss coefficient $\beta = \gamma_1[N_2] + \gamma_2[O_2]$ on 10 April with respect

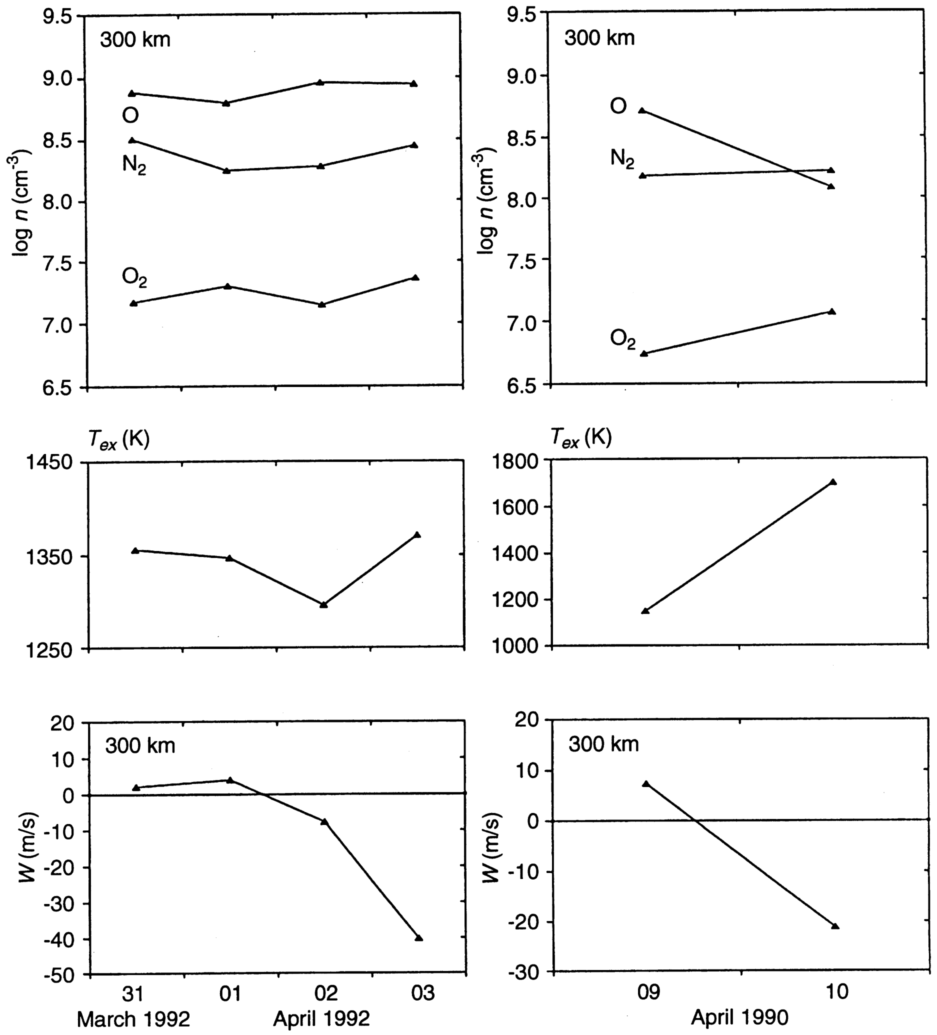

Fig. 5. Calculated neutral composition (at 300 km) and exospheric temperature T_{ex} (*top* and *middle panels*) and calculated vertical plasma drift (*bottom panel*) for 31 March–3 April 1992 (*left*) and 09–10 April 1990 (*right*)

to 9 April (by a factor of 1.6). This type of neutral composition changes is typical of the auroral zone where Joule heating results in atmospheric upwelling followed by a decrease in light atmospheric species and an increase in heavy species (e.g., Prölss and Fricke,1976; Trinks *et al.*, 1976; Prölss, 1991). An additional reduction of [O^+] can be caused again by a downward plasma drift. As in the 10 April 1990 case the calculated downward drift of 21 m/s hardly can be fully attributed to a $\approx$ 100 m/s northward thermospheric wind, since usually a southward circulation is expected for disturbed days. Again the observed northwest ion drift (700 m/s) can easily give a contribution of several m/s to the measured downward line-of-sight plasma velocity.

Both the height of the $N_e(h)$ maximum around 200 km, and the value of the electron concentration in the peak are determined by the height distribution of neutral gas concentrations. Large changes in the $T_n(h)$ profile are required to fit the observed $N_e(h)$ profile. A very low value of the shape parameter, S = 0.0055 km^{-1} instead of 0.016 km^{-1} predicted by MSIS-83, and a reduced neutral temperature at 120 km (T_{120} = 385 K instead of 515 K from MSIS-83) resulted from the $N_e(h)$ fitting procedure, while T_{ex} is close to the MSIS-83 prediction (Fig. 6, right panel). The observed T_i(h) has a similar slope around 200 km height, confirming the obtained result of the low S value. Such a $T_n(h)$ profile results in a moderate [O_2] and [N_2] increase at 300 km height despite the 600-K increase in T_{ex} (Fig. 5, right panel).

A small value of S (0.011–0.013 km^{-1} with respect to the MSIS-83 predicted S = 0.017 km^{-1}) was obtained for the same day, 10 April 1990 from the Millstone Hill data analysis (Mikhailov and Foster, 1997). Therefore the results of our calculations show a pronounced decrease in the neutral temperature in the lower thermosphere for this severe geomagnetic storm. We propose that this T_n decrease results from the enhanced atmospheric cooling trough nitric oxide (Richards *et al.*,1982; Gerard and Roble, 1988). The increase of nitric oxide concentration during disturbed periods is well documented (e.g., Rush, 1973; Barth, 1989; Siskind *et al.*,1989 a,b). It should be mentioned that the MSIS-83 model shows a tendency for the S parameter to decrease with growing geomagnetic activity level represented by the Ap index. The 10 April 1990 storm was a unique event with Ap = 124 which cannot be properly described by an empirical model like MSIS-83.

We have solely used the long-pulse EISCAT results in this paper. A comparison of these data and CP1

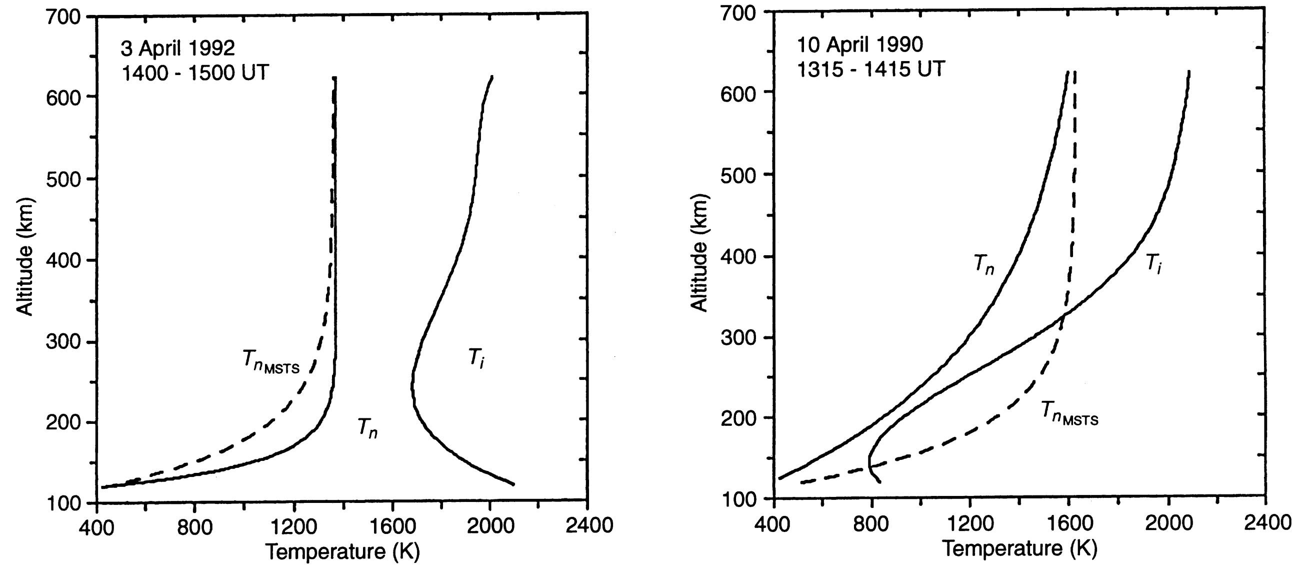

Fig. 6. Calculated and MSIS-83 $T_n(h)$ profiles together with the observed $T_i(h)$ profile for the disturbed days 3 April 1992 (*left*) and 10 April 1990 (*right*)

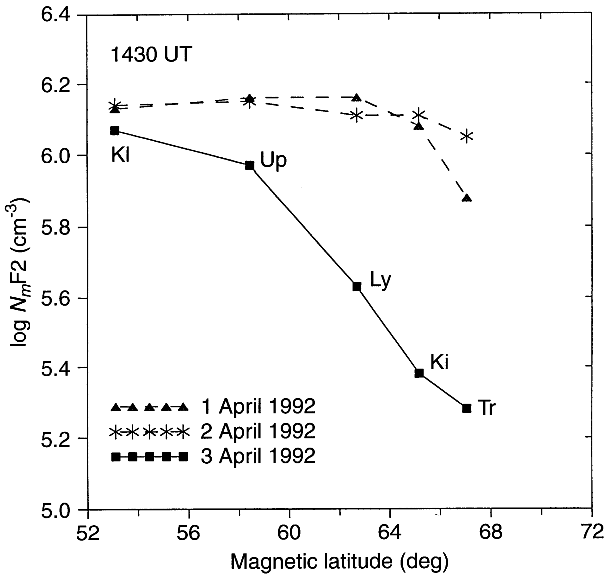

Fig. 7. Observed N_mF2 latitudinal variation for the chain of European ionosonde stations: Kaliningrad, Uppsala, Lycksele, Kiruna and Tromso (EISCAT location)

multipulse data with their finer height resolution below 260 km has shown a noticeable difference between the two sets of results. A systematic study of the cause of these differences is beyond the scope of this paper. The fact is that the multipulse data above 200 km often suffer from bad signal-to-noise ratio. The long-pulse data on the other hand may be unreliable below about 180 km because the pulse length (corresponding to about 40 km) does not adequately match the scale height of E-region structures. Since in our model the parameters $N_e(h)$, $T_i(h)$, $T_e(h)$, and $V_I(h)$ are most important at F2-region heights, we restricted our analysis to the long-pulse data.

It should also be noted that non-Maxwellian effects in the ion velocity distribution which become important in the presence of large electric fields have not been considered in our study. It is well known that these effects give rise to systematic errors in the interpretation of incoherent-scatter spectra [see Hubert *et al.* (1996) for a recent review]. For field-aligned measurements as in the case of CP-1, however, these effects should be small even at electric fields of 85 mV/m (3 April 1992). The resulting uncertainties are within the overall error limits of our method which are discussed in Mikhailov and Schlegel (1997).

4 Summary

A self-consistent method for daytime F-region modelling has been applied to EISCAT observations during two disturbed periods with a strong N_e decrease at F2-layer heights. The main results of our analysis can be summarized as follows:

1. The large negative F2-layer storm effect on 3 April 1992 with an N_mF2 reduction by a factor of 6.4 compared to the previous quiet period, is mostly due to strong electric field ($\boldsymbol{E} \approx 85$ mV/m), rather than to changes in neutral composition. This is attributed to the fact that the considered period was only about 4 h later than the storm onset. Additional contribution to the N_mF2 reduction can result from a strong downward plasma drift, although it is uncertain how much of this is due to geometrical effects.This is different from mid-latitude F2-layer storms where the vertical plasma drift is usually more positive due to enhanced equatorward thermospheric wind.

2. Unlike the 3 April 1992 case, the strong negative storm effect on 10 April 1990 with a complete disappearance of the F2-layer maximum at usual heights is

mainly attributed to changes in neutral composition and temperature. Three processes are responsible for the observed F2-layer depletion: a reduction in [O] (by a factor of 4.3) causes a corresponding reduction of the O^+ production rate; the increase in $[O_2]$, $[N_2]$, T_{ex}, and electric field provide a substantial growth in the linear loss coefficient $\beta = \gamma_1[N_2] + \gamma_2[O_2]$ for O^+ ions and finally a downward plasma drift.

3. The appearance of the ionospheric maximum around 200 km height is the result of the superposition of two layers: a normal F1 layer at 175 km and a depleted and distorted F2 layer with a maximum around 260 km height. The positions of both maxima are strongly controlled by the neutral composition, and a very small value of the shape parameter ($S = 0.0055\ \mathrm{km}^{-1}$) for the neutral temperature profile was required to fit the observed $N_e(h)$.

4. A small value of the shape parameter S and a low neutral temperature at 120 km ($T_{120} = 385$ K instead of 515 K predicted by MSIS-83) indicate strong cooling of the lower thermosphere. We propose that this cooling is due to enhanced nitric oxide concentration usually observed at these heights during geomagnetic storms.

Acknowledgements. The authors thank the Director and the staff of EISCAT for running the radar and providing the data. The EISCAT Scientific Association is funded by scientific agencies of Finland (SA), France (CNRS), Germany (MPG), Japan (NIPR), Norway (NF), Sweden(NFR), and the United Kingdom(PPARC). A.M. is grateful to the Max-Planck-Gesellschaft for a research stipend at the MPAE.

Topical Editor D. Alcaydé thanks U. Lovhaug and another referee for their help in evaluating this paper.

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