Indian Journal of Radio & Space Physics
Vol. 22, April 1993, pp. 122-127

122-7

Seasonal and latitudinal variations of stratospheric small ion density and conductivity*

N Srinivas & B S N Prasad
Department of Studies in Physics, University of Mysore, Manasagangotri, Mysore 570 006

Received 3 December 1992; accepted 9 February 1993

The seasonal and latitudinal variations of stratospheric small ion density and conductivity are studied using a simplified ion-aerosol model which includes the charged aerosol-ion recombination and charged aerosol-aerosol recombination, in addition to the usual ion-ion recombination and ion-aerosol attachment processes. The charged aerosol-ion recombination coefficient is computed from the model and the desirability of such a computation is discussed. The model computations in this study assume the background stratospheric aerosols (or Hake distribution), but the model is particularly useful in the study of stratospheric ion density and conductivity under conditions of enhanced stratospheric aerosols resulting from volcanic eruptions. The model results show seasonal and latitudinal variations of both small ion density $N_\pm$ and conductivity $\sigma_\pm$. It has been shown that the latitudinal variation of $\sigma_\pm$ is primarily controlled by the corresponding variation in $N_\pm$, whereas, the seasonal and height variations of $\sigma_\pm$ are largely governed by the corresponding variations in the ion mobility $b_\pm$. The available experimental profiles for low, middle and high latitudes show good agreement with those from model prediction.

1814

1 Introduction

Small ions consisting of aggregates of a few molecules determine the stratospheric conductivity. The small ion density is controlled by the ionizing mechanism for the production of ions and electrons and the loss processes for these charged particles. Ion production in the stratosphere is chiefly due to galactic cosmic rays (ionizing O_2 and N_2) and the resulting electrons and positive ions undergo several hydration reactions which lead to the formation of negative and positive molecular ion clusters referred to as small ions. These singly charged ions have mobilities large enough to move considerably under the influence of the electric field and thus determine the stratospheric electrical conductivity.

The equilibrium density of ions (positive or negative) is governed by the equations of continuity for the production and loss of the ions, where the gain and loss of ions due to transport is neglected. The ion-ion recombination of the molecular ions is greatly altered in the presence of stratospheric aerosols or particulates. The attachment of small ions to aerosols produces the charged aerosols referred to as large ions which are less mobile than the smaller molecular ions. The subsequent recombination of charged aerosols with ions as well as the charged aerosol-aerosol recombination would result in the depletion of small ion concentration more rapidly than in the absence of aerosols. Thus, the formation of the less mobile aerosol ions and the reduction of more mobile molecular ions alter the stratospheric ion conductivity and hence the electrical nature of the stratosphere. The ever present background aerosol concentration, referred to as Hake distribution[1], will be vastly enhanced following volcanic eruptions.

Stratospheric ion density and conductivity have been measured by balloon-borne instruments by several research groups covering equatorial, middle and high latitude locations for normal (background) aerosol conditions and following volcanic eruptions.[2-8] Simplified ion-aerosol models have been used[9-12] by several research workers for validating the experimental measurements of stratospheric ion density and conductivity and also to predict changes in these parameters due to enhanced aerosol loading.[4] However, in these models, the loss of molecular ions is considered as solely due to (i) ion-ion recombination (α_i recombination coefficient) and (ii) ion attachment to aerosols (β attachment coefficient). The other two types of losses arising from are: (i) the recombination of molecular ion with the oppositely charged aerosols (α_s recombination

*This paper was presented at the National Space Science Symposium held during 11-14 March 1992 at Physical Research Laboratory, Ahmedabad 380 009.

coefficient) and (ii) charged aerosol-aerosol recombination (α_2 recombination coefficient) have been incorporated in the model study by Datta *et al.*[13] Of these two losses, the first one directly affects the equilibrium density of molecular ions, whereas the second loss process, resulting in the liberation of neutral aerosol particles which in turn enter the reaction chain (for attachment loss of molecular ions), also depletes the molecular ions. Obviously, addition of these two ion loss processes in the ion-aerosol models results in a more realistic ion aerosol reaction scheme. However, the model computation of ion depletion due to background aerosols (Hake distribution[1]) remains unaltered with or without the inclusion of the above mentioned additional recombination coefficient α_s and α_2. It is found that the background aerosols produce less than 10% ion depletion at stratospheric heights.[4,9,13] For the model prediction of equilibrium ion density and stratospheric conductivity for enhanced aerosol conditions (where ion depletion can be larger than 10%),it is essential that all the four ion loss processes mentioned earlier should be used. Results of the present analysis for large ion depletion conditions show the restrictions and limitations of the range of the values of α_2 and α_s that can be used in ion-aerosol model studies. These and our model results on latitudinal and seasonal variations of the stratospheric ion density and conductivity for the background aerosol concentration are presented in this paper.

2 Simplified ion-aerosol model, small ion density and stratospheric conductivity

Figure 1 shows the ion-aerosol model used in this study. In this model, the detailed reaction paths for the formation of individual cluster ions are not considered and hence, the individual ion densities are not considered. The total number densities of positive and negative ions are taken to be equal from the charge neutrality criterion. The cluster ions are grouped together and the equilibrium ion density is computed from the equations of continuity which

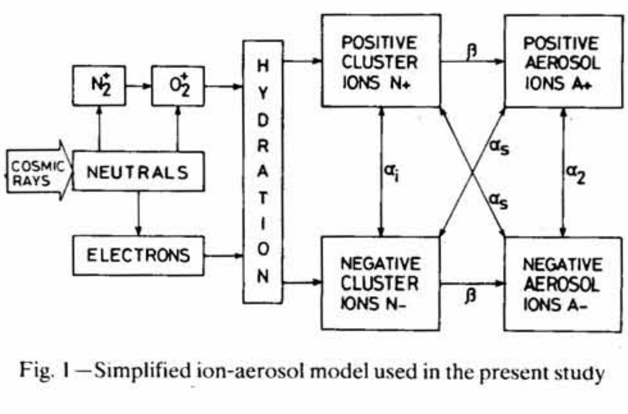

Fig. 1—Simplified ion-aerosol model used in the present study

include the four types of ion losses (Fig. 1). The positively and negatively charged aerosols formed due to positive and negative ion attachments, respectively, are also considered to be equal from the charge neutrality criterion. The recombination coefficient for the positively and negatively charged molecular ions with the oppositely charged aerosol ions are taken to be equal in our model. The equations of continuity for (i) small ions in the absence and in the presence of aerosols and (ii) the charged aerosols are:

$$dN_0/dt = q - \alpha_i N_0^2 \quad \ldots (1)$$

$$dN_\pm/dt = q - \alpha_i N_\pm^2 - \beta Z N_\pm - \alpha_s A_\pm N_\pm \quad \ldots (2)$$

$$dA_\pm/dt = \beta Z N_\pm - \alpha_s A_\pm N_\pm - \alpha_2 A_\pm^2 \quad \ldots (3)$$

where,

- q Cosmic ray ion production rate
- N_0 Positive/negative ion density in the absence of aerosols
- $N_\pm$ Positive/negative ion density in the presence of aerosols
- $A_\pm$ Positively/negatively charged aerosol density
- Z Aerosol number density
- β Aerosol-ion attachment coefficient
- α_i Ion-ion recombination coefficient
- α_s Charged aerosol-ion recombination coefficient
- α_2 Charged aerosol-aerosol recombination coefficient.

Under steady state conditions Eqs (1), (2) and (3) reduce to:

$$N_0 = (q/\alpha_i)^{1/2} \quad \ldots (4)$$

$$q - \alpha_i N_\pm^2 - \beta Z N_\pm - \alpha_s A_\pm N_\pm = 0 \quad \ldots (5)$$

$$\beta Z N_\pm - \alpha_s A_\pm N_\pm - \alpha_2 A_\pm^2 = 0 \quad \ldots (6)$$

By solving Eqs (4), (5) and (6) for $N_\pm$, with $A_\pm = N_0 - N_\pm$ [see Eq. (10)], we get:

$$N_\pm = [R \pm (B^2 - CD)^{1/2}]/C \quad \ldots (7)$$

and also from Eqs (5) and (6), we have

$$\beta Z = (q - \alpha_i N_\pm^2 + \alpha_2 A_\pm^2)/(2 N_\pm) \quad \ldots (8)$$

where,

$$R = \alpha_2 N_0 + \beta Z$$

$$B = \alpha_2 N_0 - \beta Z$$

$$C = \alpha_2 - \alpha_i$$

$$D = q + \alpha_2 N_0^2$$

Again, by simultaneously solving Eqs (5) and (6) for α_s (by eliminating βZ term) we get:

$$\alpha_s = (q - \alpha_i N_\pm^2 - \alpha_2 A_\pm^2)/(2 A_\pm N_\pm) \quad \ldots (9)$$

The ion depletion ΔN, i.e., loss of molecular ions in the presence of aerosols is:

$$\Delta N = N_0 - N_\pm \qquad \ldots (10)$$

This is also equal to the charged aerosol density $A_\pm$ The fractional depletion $\eta (= \Delta N/N_0)$ of small ions is also used in the computations. We thus have

$$\Delta N = A_\pm \,;\, \eta = (\Delta N/N_0) = (A_\pm/N_0) \qquad \ldots (11)$$

Using Eq. (11) and Eq. (4), we can write Eqs (8) and (9), respectively, as:

$$\beta Z = N_0 \eta \{\alpha_i (2-\eta) + \alpha_2 \eta\}/\{2(1-\eta)\} \qquad \ldots (12)$$

$$\alpha_s = \{\alpha_i (2-\eta) - \alpha_2 \eta\}/\{2(1-\eta)\} \qquad \ldots (13)$$

The conductivity of the stratosphere is related to the small ion density and ion mobility. Defining σ_0 and $\sigma_\pm$ as the conductivities in the absence and presence of aerosols, respectively, we have

$$\sigma_0 = N_0 e b_\pm \,;\, \sigma_\pm = N_\pm e b_\pm \qquad \ldots (14)$$

where,

e Electronic charge
$b_\pm$ Ionic mobility

Taking $\Delta\sigma$ as the reduction in conductivity due to depletion of small ions, we have

$$\Delta\sigma = \sigma_0 - \sigma_\pm$$

The ionic mobility $b_\pm$ is expressed in terms of the reduced mobility b_0 at NTP and the ambient temperature T and pressure P (Ref. 14) as follows:

$$b_\pm = b_0 P_0 T/(T_0 P) \qquad \ldots (15)$$

The value of $\Delta\sigma$ can be obtained in terms of the model parameters α_i, α_s, βZ etc. from Eqs (5)-(15). Following the method of Gringel *et al.*[2] we get

$$\Delta\sigma = \beta Z e b_\pm / [2\alpha_i \{f(\Delta\sigma)\} - \alpha_s] \qquad \ldots (16)$$

Here, $f(\Delta\sigma) = 1 + \{\Delta\sigma/(2\sigma_\pm)\}$

It may be noted that the denominator in Eq. (16) contains α_s as the extra term (from our model) as compared to the results of Gringel *et al.*[2] The above equation [Eq. (16)] is quadratic in $\Delta\sigma$ and solving this we get:

$$\Delta\sigma = [\{\pm(F^2 - 4EG)^{1/2}\} - F]/(2E) \qquad \ldots (17)$$

where,

$$E = \alpha_s - \alpha_i$$

$$F = (2\alpha_i - \alpha_s)\sigma_0 + \beta Z e b_\pm$$

$$G = -\sigma_0 \beta Z e b_\pm$$

For the case of aerosol free environment (i.e. $\beta Z = 0$) we have $\Delta\sigma = 0$, and hence in Eq. (17) the positive sign inside the curly bracket is appropriate. In the presence of aerosols and depending on the magnitudes of the quantity βZ and relative magnitudes of α_i and α_s, we get a range of values for $\Delta\sigma$ with $\pm$ sign in Eq. (17). For a realistic situation (where $0 < N_\pm < N_0$ and $0 < \sigma_\pm < \sigma_0$), we consider only the values of $\Delta\sigma$ which satisfy $0 < \Delta\sigma < \sigma_0$. In our model study, we compute $N_\pm$ and $\Delta\sigma$ for the model inputs on q, α_i, α_2, α_s and βZ, and hence obtain $\sigma_\pm = (\sigma_0 - \Delta\sigma)$. Since βZ can be expressed in terms of the ion depletion η [Eq. (12)], our model can also predict the ion density $N_\pm$ and conductivity $\sigma_\pm$ for an assumed aerosol concentration Z and size r (through the dependence of β on r).

We have used the available parametrization formulae for computing the latitudinally and seasonally varying height dependent values of q (Ref. 15) (using the neutral density $[M]$ data from the reference atmosphere of Cole and Kantor[16]) and α_i (Ref. 17). The temperature and pressure data are used for computing $b_\pm$ [Eq. (15)] with $b_0 = 1.3$ cm^{2}V^{-1}s^{-1} (Ref. 2). The aerosol concentration Z is considered for two different radii $r = 0.04$ μm and 0.4 μm (Ref. 1) and the corresponding attachment coefficients β are taken from Zikmunda and Mohnen.[18] The values of α_s and α_2 for our model computations are discussed in Sec. 3.

3 Methodology

We can compute $N_\pm$ and $\sigma_\pm$ from the Eqs (7), (10), (11), (14), (16) and (17), for the input data on q, α_i, etc. and α_s and α_2 from parametric formulae.[13] The derived height profiles of $\sigma_\pm$ are shown in Fig. 2 for the background aerosols with $r = 0.04$ and 0.4 μm (βZ values are shown in Fig. 3). Also shown in Fig. 2 are the experimental conductivity profiles for 39° and 69°N (Refs 3 and 5). We note that the experimental profiles show a monotonic increase with height, whereas the model profiles exhibit this property only up to 32 km and, thus, the nature of height variation for the experimental and model $\sigma_\pm$ pro-

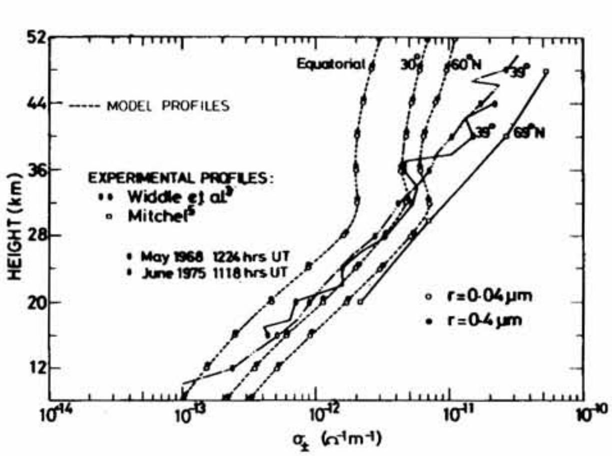

Fig. 2—Comparison of experimental conductivity profiles with the model conductivity profiles generated using parametric formulae[13] for α_2 and α_s

files is similar only at lower heights (below 32 km). This aspect is observed for larger aerosol concentrations also, where βZ is computed for ion depletions $\eta > 10\%$ [Eq. (12)]. For higher η values we find $\Delta\sigma > \sigma_0$, which arises because of large α_s and α_2-values that go into the model. To avoid this unrealistic situation, we compute α_s from our model for an assumed α_2 (or, vice versa). For this, Eqs (8) and (9) [or Eqs (12) and (13)] are used in our model. The computations are made as follows:

To start with, a large value of α_2 is assumed and α_s is determined from Eq. (9) [for Hake distribution[1] or any other assumed aerosol concentration or ion depletion, Eq. (13)]. This α_s should be positive. If not, α_2 is reduced in steps to get a positive value of α_s keeping the other input parameters unaltered. With this α_s value, the quantity $\Delta\sigma$ is determined [using Eq. (17)] which, for the realistic condition, should be $0 < \Delta\sigma < \sigma_0$. These computations are carried out with an iterative procedure to arrive at the realistic values of α_s, α_2 and $\Delta\sigma$, and hence $\sigma_\pm$. We find that $\alpha_2 = 2\times10^{-6}$ cm^{3}s^{-1} (Ref. 13) satisfies the above mentioned requirements and the derived α_s profiles are discussed in Sec. 4.

4 Results and discussion

Although our model computations cover a wide range of locations (0-75°N in steps of 15°), we are presenting the results for 0, 30 and 60° as representative of equatorial, middle and high latitudes. The computations are also made with $[M]$, T and P for January and July[16] for determining seasonal variation in $N_\pm$ and $\sigma_\pm$. As mentioned earlier, the model results are obtained for the background aerosol distributions with $r = 0.04$ and 0.4 μms (Ref. 1).

Figure 3 shows the height profiles of α_i, α_2 and βZ used in our model and the derived height profiles of α_s from the present study. The height variation of α_i and α_s are similar, and α_s exhibits an inverse dependence on aerosol size. For comparison, we have shown the α_s profile from the parametric formula[13]. At higher heights, these α_s values[13] are very large compared to the values derived from our model. Figure 4 shows the computed profiles of $N_\pm$ and $A_\pm$ where latitudinal and seasonal variations are seen only in $N_\pm$ and not in $A_\pm$. The seasonal variation in $N_\pm$ (Fig. 4) is more obvious at higher heights and higher latitudes because of the significant variation in P and T with height and latitude. In Fig. 5, the height variation of mobility $b_\pm$ for the latitudes 0, 30 and 60°N and for January and July are similar with monotonic increase with height. We note an inverse dependence of $b_\pm$ on pressure [Eq. (15)].

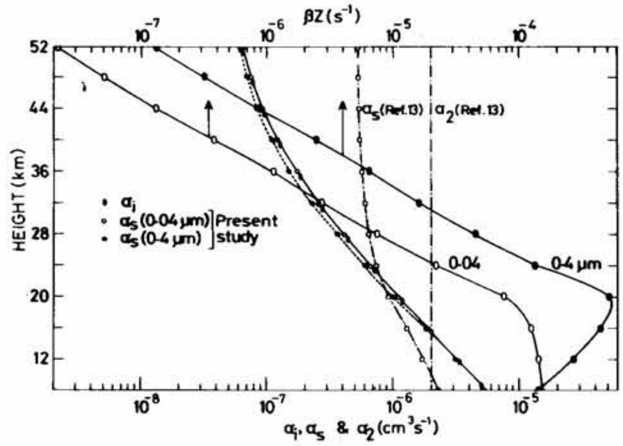

Fig. 3—Height profiles of α_i, α_s and α_2 and βZ for Hake distribution of aerosols

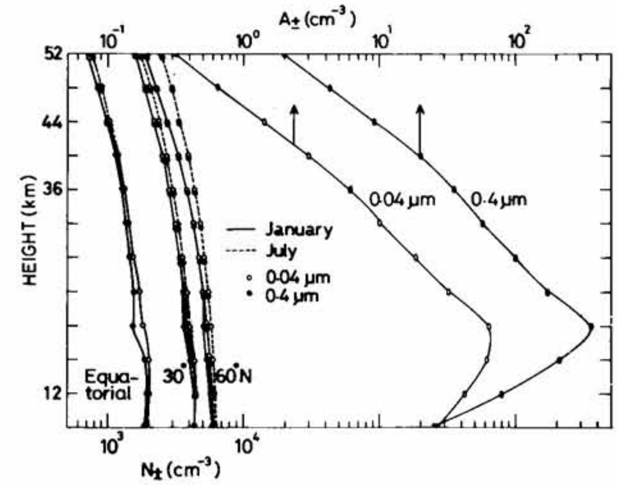

Fig. 4—Model profiles showing seasonal and latitudinal variations of $N_\pm$ and the variation of $A_\pm$ for Hake distribution of aerosols

Figure 6 shows the model profiles of $\sigma_\pm$ and a few of the experimental profiles for the three latitudes. We note that the model $\sigma_\pm$ varies with height and season similar to the corresponding variation in $b_\pm$ (Fig. 5), whereas the latitudinal variation of $\sigma_\pm$ follows closely the corresponding variation in $N_\pm$ (Fig. 4). The experimental profiles for the equatorial, middle and high latitudes show close agreement with the corresponding model profiles, particularly, in respect of height variation. This agreement in Fig. 6 compared with Fig. 2 brings out the necessity of using α_s from our model computation based on α_2 and βZ, rather than using a parametric formula.[13] It is also possible to derive the height variation of α_2 for an assumed value of α_s and βZ, for which case

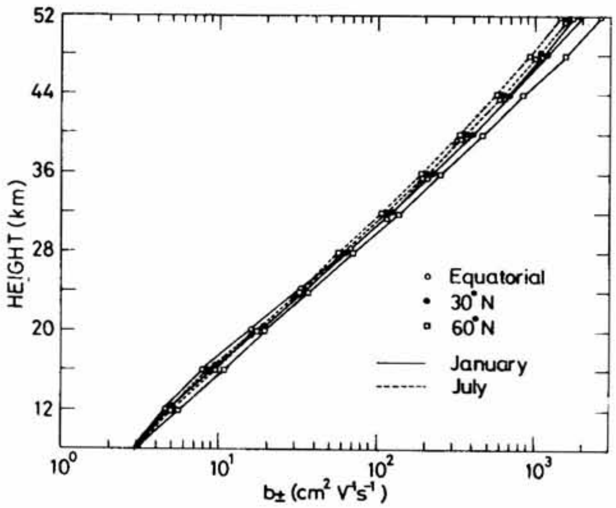

Fig. 5—Seasonal and latitudinal variations of small ion mobility

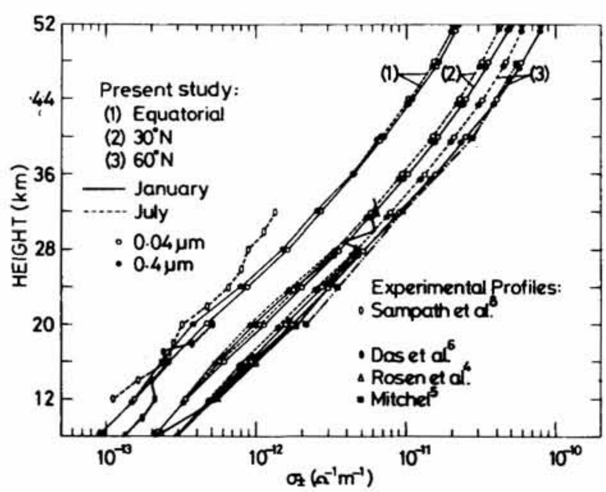

Fig. 6—Model conductivity profiles of the present study (for Hake distribution) showing seasonal and latitudinal variations and the experimental conductivity profiles for equatorial, middle and high latitudes

also we find a close and similar agreement in the height variation of $\sigma_\pm$ from the model and experimental studies (Fig. 6).

We have also obtained model values of $\Delta\sigma$ and hence $\sigma_\pm$ for small ion depletions larger than 10% [Eqs (12) and (17)], which would be the case for enhanced aerosol environment such as volcanic eruption effects. For these conditions we find that with the α_s values from the parametric formula[13] in the model would result in $\Delta\sigma > \sigma_0$, which is unrealistic. This is not so when we use α_s value from our model as discussed above. Thus, it is essential that the parameter α_s is derived for an assumed value of α_2 and βZ (or η). Alternatively, we can derive α_2 for an assumed α_s and βZ. Thus, it is necessary to be cautious in model computations of $N_\pm$ and $\sigma_\pm$ where the ion-charged aerosol recombinations is incorporated. This is particularly so in the presence of enhanced aerosol environment. It is needless to say that the models using α_i, βZ, α_s and α_2 are more realistic than the earlier models incorporating only α_i and βZ.

5 Conclusion

The previously used models[2,6,8-12] for predicting the height variation of stratospheric conductivity do not incorporate the recombination loss of oppositely charged aerosols with one another or the loss of charged aerosol with ions. However, the use of the available parametric formulae[13] for these recombination coefficients in the model leads to the predictions of height variation of stratospheric conductivity which do not agree with the experimental measurements, particularly, at higher heights. Our model study shows the desirability of deriving the above mentioned recombination coefficients analytically. The measured conductivity profiles show a good agreement with those from our model prediction.

Acknowledgement

The authors are thankful to the University Grants Commission (UGC), New Delhi, India, for the financial assistance under the Indian Middle Atmosphere Programme (IMAP). One of them (NS) is grateful to the UGC for the grant of teacher fellowship under the Faculty Improvement Programme.

References

1 Hake R D (Jr), Pierce E T & Viezee W, *Stratospheric Electricity, Final report, SRI project* 1724, Stanford Research Institute, Manlopark, California, USA, 1973.
2 Gringel W, Kaselau K H & Muleisen R, *Pure & Appl Geophys (Switzerland)*, 116 (1978) 1101.
3 Widdle H U, Rose G & Borchers R, *J Atmos & Terr Phys (GB)*, 41 (1979) 1141.
4 Rosen J M, Hofmann D J & Gringel W, *J Geophys Res (USA)*, 90 (1985) 5876.
5 Mitchell J D, *Hand Book for MAP*, edited by R A Goldberg (SCOSTEP Secretariat, University, Illinois, Illinois, USA), Vol. 19, 1986, p. 155.
6 Das M S, Sasikumar V & Sampath S, *Indian J Radio & Space Phys*, 16 (1987) 215.
7 Gupta S P & Narayan A, *Planet & Space Sci (GB)*, 35 (1987) 439.
8 Sampath S, Murali Das S & Sasikumar V, *J Atmos & Terr Phys (GB)*, 51 (1989) 533.
9 Prasad B S N, Srinivas N & Chandramma S, *Indian J Radio & Space Phys*, 20 (1991) 304.
10 Uchino O & Hirono M, *J Geomagn & Geoelectr (Japan)*, 27 (1975) 201.

11 Kawamoto H & Ogawa T, *Planet & Spaçe Sci (GB)*, 32 (1984) 1223.
12 Hoppel W A, *J Geophys Res (USA)*, 85 (1985) 5917.
13 Datta J, Revankar C P, Chakravarty S C & Mitra A P, *Phys Scr (Sweden)*, 36 (1987) 705.
14 Meyerott R E, Reagan J B & Joiner R G, *J Geophys Res (USA)*, 85 (1980) 1273.
15 Swider W, *Adv Space Res (GB)*, 2 (1983) 213.
16 Cole A E & Kantor A J, Airforce Reference Atmospheres, *AFGL-TR*-78-0051, Airforce Geophys Lab, Honscom AFB, Massachusetts, USA, 1978.
17 Smith D & Adams N D, *J Geophys Res (USA)*, 9 (1982) 1085.
18 Zikmunda J & Mohnen V A, *Meteorol Rundsch (Germany)*, 25 (1972) 1101.