https://ntrs.nasa.gov/search.jsp?R=19710026008 2020-03-11T22:28:04+00:00Z

General Disclaimer

One or more of the Following Statements may affect this Document

- This document has been reproduced from the best copy furnished by the organizational source. It is being released in the interest of making available as much information as possible.

- This document may contain data, which exceeds the sheet parameters. It was furnished in this condition by the organizational source and is the best copy available.

- This document may contain tone-on-tone or color graphs, charts and/or pictures, which have been reproduced in black and white.

- This document is paginated as submitted by the original source.

- Portions of this document are not fully legible due to the historical nature of some of the material. However, it is the best reproduction available from the original submission.

Produced by the NASA Center for Aerospace Information (CASI)

NIOK-00
CR #

RSD-64

SCIENTIFIC REPORT NO.64

IONOSPHERIC ESTIMATES OF ATOMIC OXYGEN CONCENTRATION FROM CHARGED PARTICLE MEASUREMENTS

K.K. Mahajan

APRIL 1971

FACILITY FORM 602

N71-35484 (ACCESSION NUMBER)	(THRU)
21 (PAGES)	H2 (CODE)
CR-122640 (NASA CR OR TMX OR AD NUMBER)	13 (CATEGORY)

RADIO SCIENCE DIVISION
NATIONAL PHYSICAL LABORATORY,
NEW DELHI-12, INDIA

RSD-64 PREPRINT

RADIO SCIENCE DIVISION

Scientific Report No.64

IONOSPHERIC ESTIMATES OF ATOMIC OXYGEN CONCENTRATION FROM CHARGED PARTICLE MEASUREMENTS

K.K. Mahajan[1]

April 1971

Approved by : A. P. Mitra

A.P. Mitra, Head, Radio Science Division

NATIONAL PHYSICAL LABORATORATORY, DELHI -12, INDIA.

---

[1] Part of this work was done when the author was a NRC-NASA resident research associate (1967-69) at Goddard Space Flight Center, Greenbelt, Md., U.S.A.

(To appear in the Journal of Geophysical Research, 1971)

CONTENTS

ABSTRACT

Radar backscatter and the rocket probe measurements of electron and ion temperatures and electron concentration are used to solve the thermal energy equation for the O^+ ions. By equating the heat gained by the O^+ ions from the hotter thermal electrons, to the heat lost to the cooler neutrals, the concentration of the oxygen atoms, n(O), is deduced. Significant seasonal changes in n(O) are identified. The calculated values of oxygen concentration are found to be consistently smaller than the CIRA (1965) values throughout the latest rising phase (1964-1967) of solar activity. This disagreement is greatest at low solar activity and reduces with the increase in activity. There is, however, general agreement on the time of diurnal maximum of n(O) between the calculated and model values.

1. INTRODUCTION

At present there are two major techniques for the direct measurement of atomic oxygen in the thermosphere. One is based upon the mass spectrometer measurements (Nier et al. 1964; Reber, 1964; Schaefer and Nichols, 1964; Reber and Nicolet, 1965; Hedin and Nier, 1966; Mauersberger et al, 1968; Schaefer, 1968; Gross et al, 1968; Kasprazk et al, 1968; Krankowsky et al, 1968) and the other upon the solar EUV extinction observations (Hall et al, 1963, 1965, 1967; Hinteregger and Hall, 1969). In this paper we shall estimate the concentration of atomic oxygen n(O), by another method which makes use of the measurements of the concentrations and temperatures of the electrons and the ions. It is shown that fairly reliable estimates of n(O) can be made if one concentrates on the regions where there is significant difference between the electron and ion temperatures and the ion and neutral temperatures. Since data from various rocket soundings and radar backscatter measurements is now available through the latest rising phase of solar cycle (1964-1967),the solar activity changes in n(O) are also examined. Monthly mean values of electron and ion temperatures and electron concentration published by Evans (1967) for low solar activity are used to identify seasonal changes in n(O).

2. THE METHOD

The relations between the electron, ion and neutral temperatures have been developed following Hanson and Johnson (1961), by Hanson (1963); Dalgarno et al, (1963, 1967);Geisler and Bowhill (1965); Banks (1966, 1967) and more recently by Herman and Chandra (1969). The high energy photo-electrons, created during the photo-ionization of the neutral atmosphere by the solar EUV, form the major heat source. These photo-electrons heat the thermal electrons which in turn loose their energy to the ions and the neutrals. The relevant equations describing these processes, in the ionospheric region where O^+ is the predominant ion, are (Banks, 1967):

$$Q_{ei} = 4.8 \times 10^{-7} \times N_e^2 (T_e - T_i)\, T_e^{-3/2} \text{ eV cm}^{-3}\text{sec}^{-1} \tag{1}$$

$$Q_{in} = 2.1 \times 10^{-15} \times N_e\, n(O)(T_i - T_n)(T_i + T_n)^{+1/2} \text{eV cm}^{-3}\text{sec}^{-1} \tag{2}$$

where T_n = Neutral temperature,

T_e = Electron temperature,

T_i = Ion temperature,

N_e = Electron concentration

Q_{ei} is the heat gained by the ions from the thermal electrons and Q_{in} the heat lost by the ions to the neutral gas. Heat losses to atomic oxygen alone have been considered, as the losses are negligibly small for the O_2 and N_2 molecules. Since the heat transfer in the ions by bulk transport or by conduction can be neglected (Nisbet, 1967; Banks, 1967) one could equate Q_{ei} and Q_{in} to get :

$$n(0) = 2.29 \times 10^8 \times N_e \frac{(T_e - T_i)}{(T_i - T_n)} (T_i + T_n)^{-1/2} T_e^{-3/2} \qquad (3)$$

Equation (3) is the basic relation to obtain atomic oxygen concentration from charged particle measurements. As the quantity n(0) is dependent on the ratio $(T_e - T_i)/(T_i - T_n)$, one has to restrict to heights where both the quantities, $(T_e - T_i)$ and $(T_i - T_n)$, are appreciable. At altitudes below 300 km, for examples, $T_i - T_n$ will only be a few degrees while $T_e - T_i$ may be hundreds of degrees. At altitudes well above 700 km, while $T_i - T_n$ will be large, $T_e - T_i$ may be very small. Thus one has to leave out such heights, because the accuracy of n(0) calculated this way, will be greatly influenced by the accuracy of $T_i - T_n$ and $T_e - T_i$ measurements. The ideal height range is around 500 km where T_i is mid-way between T_e and T_n (see also Bowhill, 1967).

An unknown parameter which exists in Equation (3) is the neutral temperature. It should be permissible to use T_i at 250 km as the temperature of the neutrals, since $T_i - T_n$ is only a few degrees (about 10°K) at this height (Nisbet, 1967).

An error of a few degrees will only have minor effect in the quantity $T_i - T_n$ at higher altitudes, because $T_i - T_n$ is several hundred degrees for heights above 400 km. This is seen in Fig.1 where observed values of T_e and T_i at Arecibo are shown for one of the days. The T_n profile is an exponential extrapolation of the T_i value observed at 250 km and is of the form given by Jacchia (1965) :

$$T = T_\infty - (T_\infty - T_{120}) \exp\left[-s(z - 120)\right] \qquad (4)$$

where T_{120} is the temperature at 120 km, T_∞ the asymptotic (exospheric) temperature, z the height in kilometers and s a constant, known as the shape parameter. For constructing the T_n profile in Fig.1, T_{120} was taken as 355°K from CIRA (1965) and s as 0.02, a value very close to the daytime CIRA model.

The method is suitable for day hours alone when appreciable differences exist between T_e and T_i and T_i and T_n. During the night T_e, T_i and T_n are nearly equal and so this method cannot be applied for obtaining n(O) during the night hours.

A modified form of this method has recently been used by Bauer et al, (1970) for a simultaneous determination of the exospheric temperature, T_∞ , the atomic oxygen concentration, n(O) and the shape parameter, s. The measured values of T_e and N_e were used in Equations (1) and (2) to

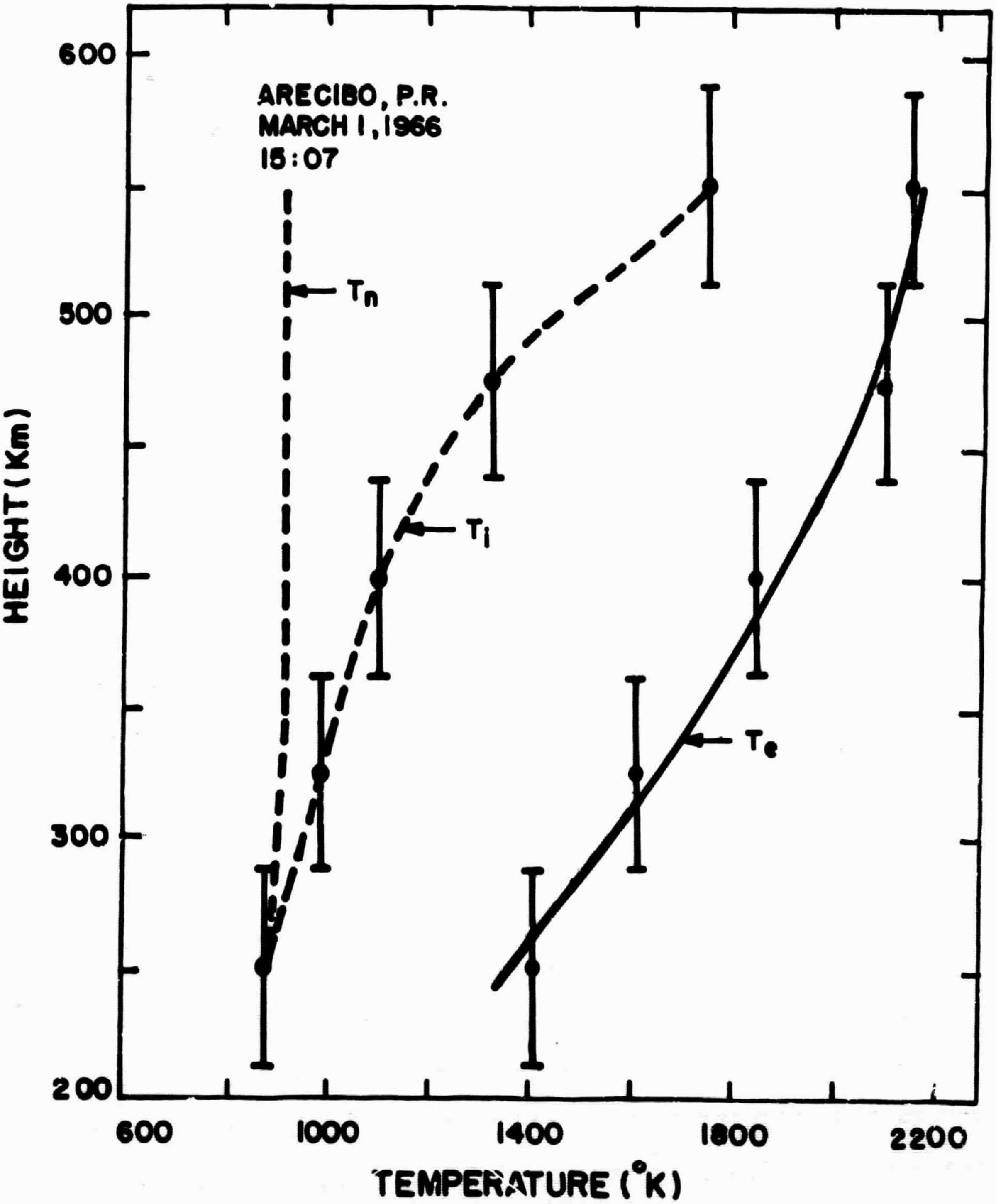

Fig.1 Sample electron and ion temperature profiles at Arecibo during sunspot minimum. The vertical bars indicate the pulse length used for collecting the spectral information.

find a set of self consistent values of T_∞, n(0) and s. This self consistency was achieved by calculating a T_i profile from Equations (1) and (2), which fitted best to the experimentally observed ion temperature profile.

3. THE RESULTS

We have used the above method to examine the diurnal, seasonal and solar activity changes in the atomic oxygen concentration by employing T_e, T_i and N_e measurements obtained from various sources. Mostly radar-back scatter data from Arecibo (18.4°N, 66.8°W) have been used, although data from Millstone Hill (42.6°N, 71.5°W) published by Evans (1967) and St. Santin (44.6°N, 2.2°E) published by Petit (1968) have also been used. Probe measurements of T_e, T_i and N_e reported by Maier (1969) and Hanson et al (1969) have also been employed.

Diurnal Variations :

Figure 2 shows n(0) values calculated for one of the days during a period of low solar activity. The error bars have been calculated by assigning $\pm 50^\circ K$ accuracy in T_i and ± 0.1 accuracy in T_e/T_i measurements (Perkins and Wand, 1965). It can be noted that there is a sharp increase in the n(0) values from morning to noon and then a steady decrease towards the evening. There is some evidence of a maximum around 1400 hours. This behaviour seems consistent with the diurnal maximum evolved from the satellite drag measurements of atmospheric

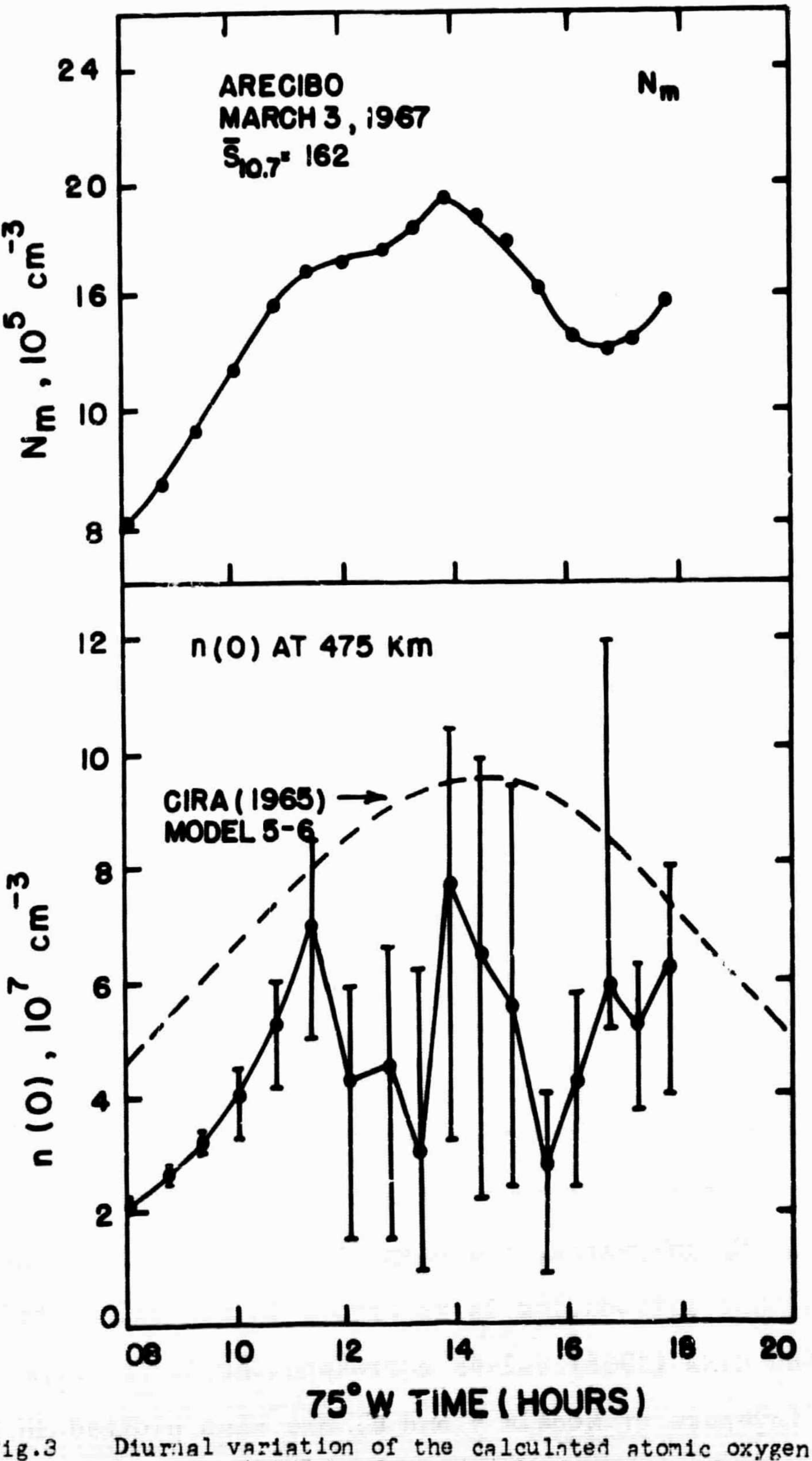

Fig.3 Diurnal variation of the calculated atomic oxygen concentration at 475 km for a day during high solar activity. The hour-to-hour fluctuations may not be real and are probably due to large error bars in the calculated n(O).

density (see e.g. Priester et al, 1967 for a complete review). We have also compared the calculated n(0) values with CIRA (1965). The model values correspond to a 10.7 cm solar radio flux of 75 units (Model No.2) and have been scaled down by a factor of 4 to force the agreement near the time of diurnal maximum. It is clear from Figure 2 that the model values are larger by a factor of 4 for the noontime and by almost an order of magnitude for the forenoon hours.

We have also plotted the peak electron concentration of the F2-layer (N_m) corresponding to the time of n(0) values in Figure 2. There is a good deal of similarity between the diurnal variations of n(0) and N_m, although there is a time delay of about 1.5 hour between the two. This time delay seems to be due to the sluggishness of the ionosphere and is consistent with the electron loss coefficient values for heights around the F2 peak.

The calculated n(0) values for a day during high solar activity are shown in Figure 3. There is apparently a large hour-to-hour fluctuation in the n(0) values. We, however, think that such fluctuations are not real. During high solar activity as N_e increases, the quantities (T_i-T_n) and (T_e-T_i) decrease, thus introducing large errors in the calculated n(0) values. The CIRA (1965) values corresponding to the same solar activity (average of Models 5 and 6) are also plotted in Fig.3,

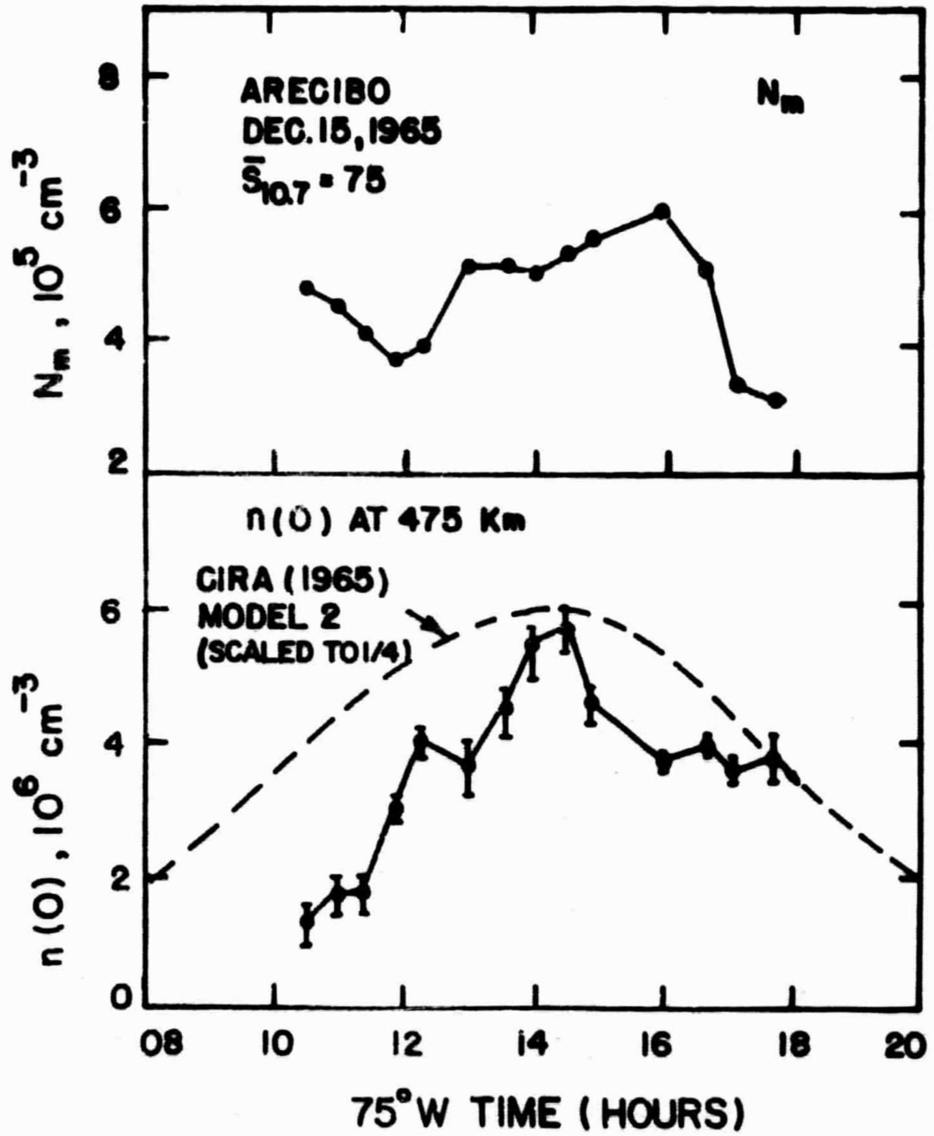

Fig.2 Diurnal variation of the calculated atomic oxygen concentration at 475 km for a day during low solar activity. CIRA (1965) values for Model 2 (Flux =75 units) are also plotted for comparison and have been scaled down by a factor of 4.

for comparison. The disagrement between the model and the calculated values exists at the high activity also, although it is not as large as during low solar activity.

We have analysed diurnal variation of n(0) on several other days during the period Oct. 1965 through March 1967. The n(0) values have generally shown a diurnal maximum between 13.00 and 15.00 L.T. for cases when electron concentration values were low. Due to low N_e, the T_e and T_i distributions were of the type whown in Figure 1, and $(T_e-T_i)/(T_i-T_n)$ could be estimated relatively more accurately. For days with high N_e, the accuracy of n(0) determination was very poor and thus it was not possible to identify the time of diurnal maximum. It is to be noted that this is not a short coming of the method, but simply the accuracy of T_e and T_i measurements, which determines the accuracy of n(0).

Seasonal Variations :

Evans (1967) has published monthly mean hourly values of T_e, T_i and N_e for Millstone Hill for the year 1964, a period at the minimum in the sunspot cycle. These means are derived from the 30-hour period observations taken every two weeks. We have used the average daytime (0900 to 1500 hours) values for the electron and ion temperatures and the noontime (1200 hrs) values for the electron density (Figures 15a, 16a and 13c respectively of Evans, 1967). The electron and ion temperature profiles at Millstone Hill behave differently from those at Arecibo, at least in the sense, that T_i does not fall middle of T_e and T_n untill a height of 600 km. Thus, for better accuracy, we have selected a height of 600 km for studying seasonal changes in n(0) at Millstone Hill.

Figure 4 is a plot of calculated n(0) at 600 km. We have also plotted the observed values of N_m and N_e at 600 km in the figure for comparison. A semi-annual effect in n(0), as well as in N_e at 600 km, can be identified. A semi-annual effect in n(0)/n(O_2) has also been observed at a height of 120 km by Mayr and Mahajan (1971) from various neutral mass spectrometer measurements. This would imply a semi-annual effect in n(0) at 120 km, provided T_∞ did not show any seasonal change. While this semi-annual effect satisfactorily explains the semi-annual variations seen in the F2 peak at all low latitudes and at all levels of solar activity (Mayr and Mahajan, 1971), this fails to explain the winter anomaly in N_m at Millstone.

The semi-annual effect in n(0) at 600 km and the winter anomaly at Millstone, could possibly by explained by a combination of two effects - an annual effect in n(0) in the lower thermosphere, with maximum in winter and an annual effect in T_∞ , with maximum in summer. Radar measurements at St. Santin, infact, give some evidence to this possibility. Waldteufel(1970),

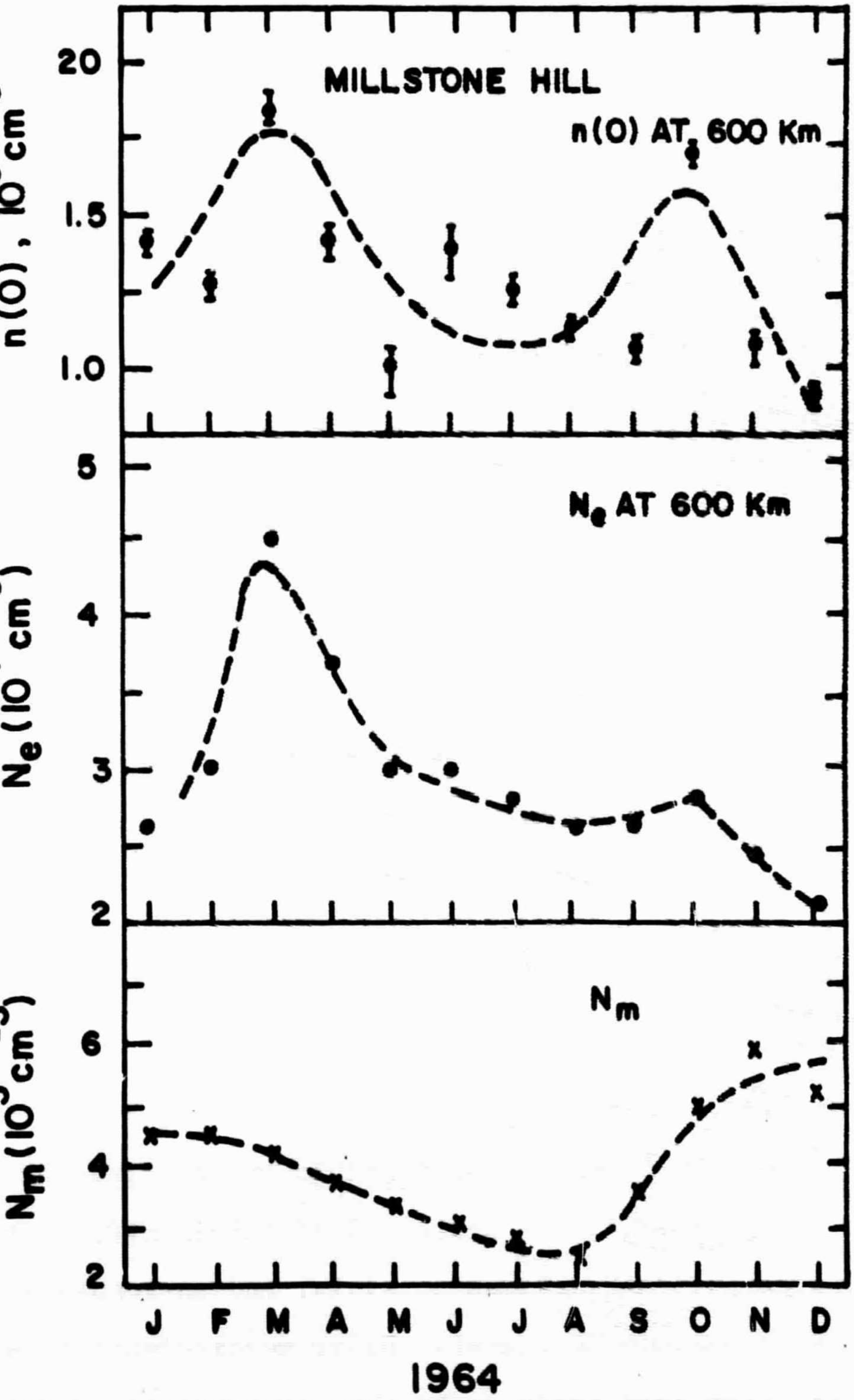

Fig.4 Seasonal variations in the calculated n(O) at 600 km ove Millstone Hill. The radar-backscatter data published by Evans (1967) have been used.

for example, from measurements in the lower thermosphere has inferred an annual effect in n(O) with a maximum in winter. Carru and Waldteufel (1969) have seen an annual variation in T_∞ with an amplitude of $\pm 60^\circ$K and with the maximum occuring in summer. The winter anomaly, however, does not occur in the topside, as evident from Millstone Hill data and the Alouette-1 satellite data (King et al, 1967). This could possibly be explained due tc low peak heights during winter caused by poleward winds, as suggested by Mayr and Mahajan (1971; see also Vasseur, 1970).

Solar Activity Variations :

We have plotted the calculated n(O) values at 475 km against the 27-day average 10.7 cm solar radio flux ($\overline{S}_{10.7}$) in Figure 5. The data used are summarised in Table 1 and mostly correspond to 14.00 Local time. No distinction has been made for the various seasons, because the seasonal changes in n(O) are much smaller than the solar activity changes at 475 km (see e.g. CIRA, 1965). We have also plotted the n(O) values from CIRA (1965) model for comparison. It can be noted that the calculated n(O) values are lower than the CIRA values. The disagreement is greatest at the solar minimum and reduces with increase in solar activity.

The discrepancy between the calculated n(O) and the model values has also been observed by Bauer et al (1970). They have,

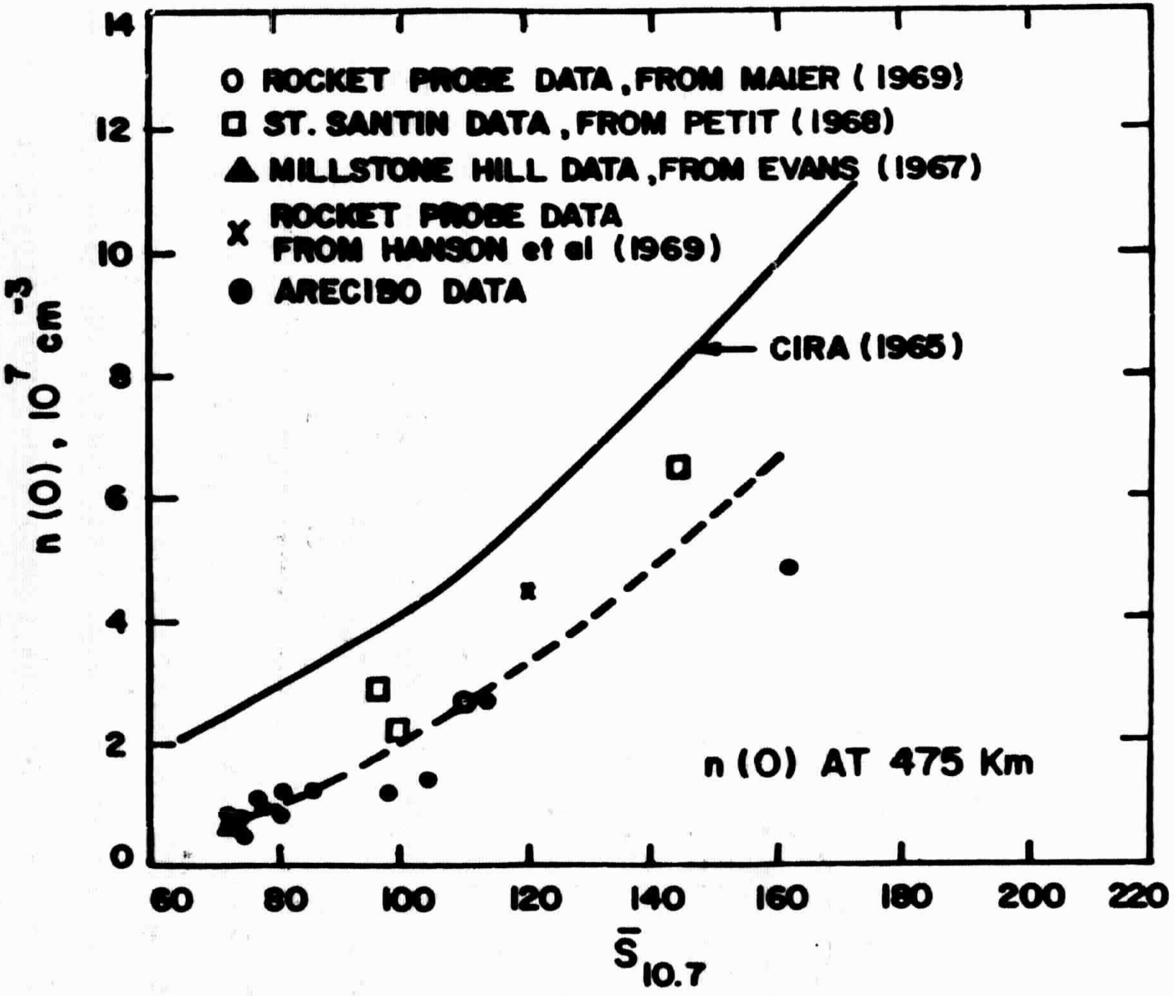

Fig.5 Solar activity changes in the calculated n(O) for 14:00 hour at 475 km. The data used are summarized in Table 1. CIRA (1965) values are also plotted for comparison.

TABLE 1: Summary of the data used in Fig.5

Station	Date and Time	Source
Arecibo (18.4°N, 66.8°W)	1965 Oct. 6, Oct.28, Dec. 1, Dec. 15, Dec. 29. 1966 Feb. 1, Mar. 1, Apr. 2, Aug. 17, Nov. 13. 1967 March 4. Average $n(0)$ for 12 to 16 hours	Arecibo Ionosphere Group
Millstone Hill (42.6°N, 71.5°W)	Monthly averages for 1964. Noontime $n(0)$ averaged for the whole year.	Evans (1967)
St. Santin (44.6°N, 2.2°E)	April 9,1966 11:00 to 13:20 June 30,1966 11:08 to 12:37 Mar. 31,1967 11:58 to 13:58	Petit (1968)
Wallops Island (37.9°N, 75.5°W)	Oct. 6, 1966 15:30	Maier (1969)
Wallops Island (37.9°N, 75.5°W)	June 2, 1967 14:00	Hanson et al (1969)

however, found that this disagreement tends to disappear at lower altitudes and have suggested the difference between the model T_∞ and the radar T_∞ as the cause of the discrepancy. As a matter of fact, the Arecibo T_i values at 250 km (where $T_i \approx T_n$), published by Mahajan (1967), are found to be systematically lower than the CIRA (1965) values by more than 100°K, throughout the latest rising phase of solar activity. Thus we also believe that the disagreement at 475 km between the calculated n(0) and the CIRA (1965) values is due to high T_∞ used in the CIRA model.

4. CONCLUSION

We have seen that there is a general agreement on the time of diurnal maximum for n(0) between the calculated and the model values. This, thus helps in resolving some conflict between the satellite drag and the radar measurements (see e.g. Carru et al, 1967; Nisbet, 1967; Mahajan 1969 and McClure, 1969), in the sense that there is some lag between the times of diurnal maximum of the neutral density and the neutral temperature and that it may not be entirely correct to deduce T_∞ from density measurements from the static diffusion models. We have also seen that the calculated values of n(0) are smaller than the CIRA values and that this disagreement is greatest at solar minimum and during the early morning hours when the T_∞ values are low. This disagreement, however, tends to disappear if one uses T_∞ values from radar data (the CIRA values are systematically higher than the radar values by more than 100°K at all levels of solar activity). In view of this, the atmospheric density values deduced from satellite drag and the T_∞ values obtained from radar measurements, should be combined to develope more realistic models of the neutral atmosphere.

Acknowledgements :

I am thankful to Dr. B.C.Narasinga Rao for helpful comments. Part of this work was done while I was NRC-NASA Resident Research Associate (1967-69) at the Goddard Space Flight Center, Greenbelt, Md. and am grateful to the U.S. National Academy of Sciences for awarding the associateship. The Arecibo radar data was gathered jointly with all of the members of the then resident ionospheric group of the Arecibo observatory under a regular observation program. The Arecibo observatory is operated by Cornell University with the support of the Advanced Research Project Agency and the National Science Foundation.

P.C. Rana

REFERENCES

Banks, P.M., "Charged particle temperatures and electron thermal conductivity in the upper atmosphere", Ann. Geophys., 22, 577, 1966.

Banks, P.M., "Ion temperature in the upper atmosphere", J. Geophys. Res., 72, 3365, 1967.

Bauer, P., P. Waldteufel and D. Alcayde, "Diurnal variations of the atomic oxygen density and temperature determined from incoherent scatter measurements in the ionospheric F-region", J. Geophys. Res., 75, 4825, 1970.

Bowhill, S.A., "Panel discussion on the next generation Thomson Scatter radar" in Thomson scatter studies of the ionosphere an informal conference record, Aeronomy Report No.19, Electrical Engineering Dept., Univ. of Illinois, Urbana, Illinois.

Carru, H., M. Petit and P. Waldteufel, "On the diurnal variation of the thermopause temperature", Planet. Space Sci., 15, 944, 1967.

Carru, H. and P. Waldteufel, "Etude par diffusion de Thomson des variation de la temperature exospherioue", Ann. Geophys., 25, 433, 1969.

CIRA, COSPAR International Reference Atmosphere, 1965, North-Holland, Amsterdam.

Dalgarno, A., M.B. McElroy and R.J. Moffet, "Electron temperatures in the ionosphere", Planet. Space Sci., 11, 463, 1963.

Dalgarno, A., M.B. McElroy and J.C.G. Walker, "The diurnal variation of ionospheric temperatures", Planet. Space Sci., 15, 331, 1967.

Evans, J.V., "Millstone Hill Thomson-Scatter results for 1964", Technical Report 430, Lincoln Laboratory, Massachussetts Institute of Technology, Lexington, Mass., 15 Nov., 1967.

Geisler, J.E. and S.A. Bowhill, "Ionospheric temperatures at solar minimum", J. Atmosph. Terr. Phys., 27, 457, 1965.

Gross, J., D. Offermann and U. Von Zahn, "Neutral particle densities in the lower thermosphere as measured by mass spectrometers above Fort Churchill and Sardinia, Space Res., 8, 920, 1968.

Hall, L.A., W. Schweizer and H.E. Hinteregger, "Diurnal variation of the atmosphere around 190 kilometers derived from solar extreme ultraviolet absorption measurements", J. Geophys. Res., 68, 6413, 1963.

Hall, L.A., W. Schweizer and H.E. Hinteregger, "Improved extreme ultraviolet absorption measurements in the upper atmosphere", J. Geophys. Res., 70, 105, 1965.

Hall, L.A., C.W. Chagnon and H.E. Hinteregger, "Daytime variations in the composition of the upper atmosphere", J. Geophys. Res., 72, 3425, 1967.

Hanson, W.b. and F.S. Johnson, "Electron temperatures in the ionosphere", Mem. Soc. Roy. Sci. Liege, Ser. 5, 4, 390, 1961.

Hanson, W.B., "Electron temperatures in the upper atmosphere", Space Res., 3, 282, 1963.

Hanson, W.B., S. Sanatani, L.H. Brace and J.A. Findlay, "Thermal structure of an Alouette-2 topside profile as deduced from rocket measurements", J. Geophys. Res., 74, 2229, 1969.

Hedin, A.E. and A.O. Nier, "A determination of the neutral composition number density, and temperature of the upper atmosphere from 120-200 kilometers with rocket borne mass spectrometers", J. Geophys. Res., 17, 4121, 1966.

Herman, J.R. and S. Chandra, "The influence of varying solar flux on ionospheric temperatures and densities - a theoretical study", Planet. Space Sci., 17, 815, 1969.

Hinteregger, H.E. and L.A. Hall, "Thermospheric densities and temperatures from EUV absorption measurements by OSO-III Space Res., 9, 619, 1969.

Jacchia, L.G., "Static diffusion models of the upper atmosphere with empirical temperature profiles", Smithsonian Astrophys. Obs. Spec. Rept. 170, 1964.

Kasprzak, W.T., D. Krankowsky and A.O. Nier, "A study of day-night variations in the neutral composition of the lower thermosphere", J. Geophys. Res., 73, 6765, 196?.

Krankowsky, D., W.T. Kasprzak and A.O. Nier, "Mass spectrometer studies of the composition of the lower thermosphere during summer 1967", J. Geophys. Res., 73, 7291, 1968.

Mahajan, K.K., "10.7 cm - solar radio flux and ionospheric temperature," J. Atmosph. Terrest. Phys., 29, 1153, 1967.

Mahajan, K.K., "Diurnal variation of the ion temperature", J. Atmosph. Terrest. Phys., 31, 93, 1969.

Maier, E.J.R., "Sounding rocket measurements of ion composition and charged particle temperatures in the topside ionosphere," J. Geophys. Res., 74, 815, 1969.

Mauersberger, K., D. Muller, D. Offerman and U. Von Zahn, "Neutral constituents of the upper atmosphere in the altitude range of 110 to 160 km above Sardinia", Space Res., 7, 1150, 1967.

Mayr, H.G. and K.K. Mahajan, "Seasonal variations in the F2-region", J. Geophys. Res., 1971 (in press).

McClure, J.P., "Diurnal variation of neutral and charged particle temperatures in the equatorial F-region", J.Geophys. Res., 74, 279, 1969.

Nier, A.P., J.H. Hoffman, C.Y. Johnson and J.C. Holmes, "Neutral composition of the atmosphere in the 100-200 km range", J. Geophys. Res., 69, 979, 1964.

Nisbet, J.S., "Neutral atmospheric temperatures from incoherent scatter observations", J. Atmosph. Sci., 24, 586, 1967.

Perkins, F.W. and R. Wand, "Analysis of the ionospheric incoherent scatter signal by digital methods, CRSR Rept. 207, Cornell University Ithaca, N.Y., 1965.

Petit, M., "Measures de temperatures, de densite electronique et de composition ionique dans l'ionosphere par diffusion de Thomson, Etude du desequilibre thermodynamique dans l' ionosphere diurne", Ann. Geophys., 24, 1, 1968.