CHAPTER IV NEUTRAL COMPOSITION MEASUREMENTS AT THE DIP EQUATOR

4.1 Introduction:

The thermosphere is defined as that region of the earth's upper atmosphere which lies above approximately 85 km. The lower thermosphere is one of the most interesting regions of the upper atmosphere. It is here that the atmospheric parameters undergo their greatest changes; atomic oxygen begins to exist as an independent entity. The temperature rises steeply, convective equilibrium ends and diffusive equilibrium sets in. The level of separation is known as the turbopause ($h_t$). Usually the absolute densities at all higher altitudes are determined by the turbopause level and the exospheric temperature ($T_\infty$).

Reliable composition measurements are one of the important inputs to the modelling of the upper atmosphere. These measurements are used either as boundary conditions or as a check to the photochemical calculations. Information on atmospheric total density alone may yield a completely misleading picture, since large changes of two or more constituents may nearly cancel each other and thus only a small density effects may be exhibited. (Mayr and Volland 1972). Almost all our knowledge about thermospheric composition below 200 km, barring a few recent satellite measurements, stems from rocket experiments. VonZahn (1970) has collected the results of a number of such measurements.
at 150 km altitude. In the same lines Offerman (1974) has updated the measurements and has tried to evolve a composite picture of the different constituents and their possible variations, for three different altitudes namely 120, 150 and 200 km. Almost all the experiments referred above had been performed at middle and high latitudes. There had been very few neutral density measurements from low, in particular from dip equatorial latitudes. In the present investigation some of the measurements made from THUMBA, the dip equatorial station are presented and comparisons are made with other prevailing low latitude results.

4.2 Data used:

Almost all the data used in the study were obtained from the Soviet ion/neutral switching mass spectrometer referred earlier. Though the observations are spaced over a period of seven years, the use of one and the same technique, excludes the systematic errors to a great extent. It becomes necessary to consider these errors when the data that is being compared are obtained by different types of mass spectrometers.

The most important constituents measured in the lower thermosphere could be listed as $N_2$, $O_2$, $O$ and $Ar$. Other minor species like Nitric Oxide (NO) and atomic Nitrogen would generally be beyond the sensitivity of the instruments and different techniques are to be adopted in evaluating their densities.
As the neutral and ionized parts of the atmosphere are closely inter-linked it may not be absolutely correct to deal with them separately but it becomes necessary to do so in order to bring out some important features. This aspect would become evident in the following sections. In the upper atmosphere which is never static, very complex processes are always in progress. Photoionization, ion-molecule reactions, effects of electric and magnetic fields, winds etc. to name a few, are some of the important processes.

4.3 Importance of the relative composition changes:

In the upper atmosphere, photoionization due to the solar UV radiation is the major source of ionization. The ions thus produced chemically react with the prevailing neutral molecules (ion-molecule reactions), and the rates of such different reactions eventually decide the resultant ion densities and their variations. These aspects are brought out clearly by the following equations (Danilov 1970).

The concentrations of the principal ionospheric ions could be represented under photochemical equilibrium condition as,

\[
\begin{align*}
[N_2^+] &= \frac{2[N_2] + [N]^+ [N_0]}{[N_0] \alpha + [O]^+ + [O_2] \alpha + [N] \alpha} \\
[O^+] &= \frac{3[O] + [N_2]^+ [O]}{[N_2] \alpha + [O_2] \alpha + [N_0] \alpha}
\end{align*}
\]
\[
\begin{align*}
[\text{NO}^+] & = \frac{-109-}{[N_e] \alpha} \left\{ [O^+] [N_2] \alpha + [O^+] [\text{NO}] \alpha + \\
& + [O_2^+] [N] \alpha + [O_2^+] [N_2] \alpha + [N]^+ [O_2^-] \alpha + \\
& + [\text{NO}] [N^+] \alpha + [\text{NO}] \right\} \quad (4.3)
\end{align*}
\]
\[
\begin{align*}
[O_2^+] & = \frac{[O_2^+] \frac{\dot{N}}{[N_c]} + [O_2^-] [O^+] \alpha + [O_2] [N_2^+] \alpha + [\text{NO}] [\text{O}^+] \alpha}{[N_e] \alpha + [N] \alpha + [N_2] \alpha} \quad (4.4)
\end{align*}
\]
\[
\begin{align*}
[N^+] & = \frac{\frac{\dot{N}}{[N_c]} [N_2] + \frac{\dot{N}}{[N_c]} [N] + N [N_2^+] \alpha}{[O_2^-] \alpha + [\text{NO}] \alpha} \quad (4.5)
\end{align*}
\]

Where \( \dot{N} \) stands for the rate coefficient for direct production, like for example photoionization and \( \alpha \) the dissociative recombination or the reaction coefficient, representing the loss process, by the reactions between the resultant ion and the species referred.

These equations take into account actually all the processes. In reality the relative role of different processes is different and these expressions could be simplified.

1) As \([\text{NO}]\) is a minor constituent less by a few orders, in concentration and also the photoionization cross section being very small all the terms containing \([\text{NO}]\) could be neglected.
The terms containing \([N^+]\) could also be dropped as there are no measurable \([N^+]\) below 200 km.

Taking these into consideration, the simplified equations could be written as (Danilov 1970),

\[
\begin{align*}
[N_2^+] &= \frac{[N_2]^j}{\alpha[N_e] + \alpha[O]} \\
[C_2^+] &= \frac{[C_2]^j + [O_2][O^+]}{\alpha[N_e] + [N_e]} \\
[O^+] &= \frac{[O]^j + [N_2^+]}{\alpha[N_2] + \alpha[C_2]} \\
[NO^+] &= \frac{[O^+][N_2] + [O_2][N_2]}{\alpha[N_e]} \\
[N^+] &= \frac{[N_2]^j + [N]^j + [N][N_2^+] \alpha}{[O_2] \alpha}
\end{align*}
\]
The relative importance of the different neutral species are brought out by these equations. For example the \([O^+]\) and \([N_2^+]\) densities have a relatively simple dependence on the ratio of concentrations of atomic and molecular species. Below 250 km \(\propto [N_2^+]/[N_2]\) being small can be neglected. Therefore \([N_2^+]\) is inversely proportional to the \([O]/[N_2]\) ratio. The equation governing \([O^+]\) densities normally decides the electron densities in the F-region. \([N_2^+]\) being a minor constituent can be neglected as an unimportant parameter. Hence the F-region electron density is decided by the ratios \([O]/[N_2]\) and \([O]/[O_2]\) and their variations.

So, at any time the distribution of the ionic species are governed by, the distribution, the relative concentration and the reaction rate constants of the major neutral species. In the present work therefore the data pertaining to the neutral composition are given only in their relative form.

4.4 \([O]/[O_2]\) ratio:

The atmospheric mixing and diffusion play an important role in determining the \([O]\) and \([O_2]\) distributions. Results of different rocket measurements show a large discrepancy mainly as a result of the different instrumentation rather than from variations in the atmosphere. The atomic oxygen being a highly
reactive species mainly reacts with the walls of the analyzer in particular the ion source walls. (Lake and Nier 1973; Nier et al 1972). The resultant density obtained is felt to be less by a factor of 2.5 as compared to the actual densities (Offerman, 1974). Different source geometry of the ion sources had been tried like open sources, semi-open and cryo cooled sources etc., with some success and larger value of atomic oxygen densities were obtained.

In the author's measurements when calculating the relative concentrations of \( [O^+] \) and \( [O_2] \) from the ion current ratio, the differences of the ionization cross sections for these gases are taken into consideration and also the mass discrimination inside the analyzer is also considered (Pokhunkov, 1974). The relative ionization cross section of \(^0\) as compared with that of \( ^{16}O_2 \) for ionization energy of 50 eV is taken to be 0.85 (Fite and Brackman, 1959). Further the instrument relative efficiency for the light ions i.e. for \( O^+ \) in relation to \( O_2^+ \) is equal to 0.55 according to the laboratory calibration data. Another coefficient due to the influx of molecules of different mass of \(^0\) and \(^{16}O_2\) varies between 0.67 to 0.77, for the altitude interval of 120-160 km. From the above considerations the sensitivity of the instrument for \(^0\) as compared with \(^{16}O_2\) equals 0.313-0.360. Account has also been taken of the loss of atomic oxygen in the walls of the chamber and at the cathode of the ion source. Also it is assumed that
44\textsuperscript{+}(CO\textsubscript{2}) and 18\textsuperscript{+}(H\textsubscript{2}O) peaks are entirely due to the reaction of ambient 'O' atoms with 'C' and 'H' inside the mass tube (Pokhunkov 1972).

In figure 4.1 (P.113a) is depicted the various \([O]/[O\textsubscript{2}]\) profiles obtained from THUMBA on various occasions. Also included are the low latitude measurements from Atlantic ocean for comparison.

4.5 Day to night variability:

Profiles labelled 1 and 2 in 4.1 a were obtained in the study of the night to day variation of the atmospheric constituents by the author (Danilov et al, 1980). The results reveal that the night time values of \([O]/[O\textsubscript{2}]\) are consistently larger than the morning values. The time difference between these two flights being 3 hr 45 mins; the night-time flight preceded the early moving one. The night-time values of \([O]/[O\textsubscript{2}]\) are seen to be larger by a factor of 1.2 to 1.6. These data are in disagreement with the earlier low latitude measurements conducted from Atlantic Ocean (Pokhunkov, 1972; Schaffer, 1969) which agree among themselves fairly well. In figure 4.1b profiles 2 and 3 represent the measurements by Pokhunkov (1972) where the ratio of \([O]/[O\textsubscript{2}]\)\textsubscript{night} to \([O]/[O\textsubscript{2}]\)\textsubscript{day} turns out to be about 0.6 all through. As the technique is one and the same for all the measurements the experimental uncertainty could be taken as small. This aspect is discussed in detail in the later section.
Fig 4.1: a) \([O_1]/[O_2]\) ratio as obtained from
the equator.

b) Same ratio as obtained from
low latitude Atlantic ocean
measurements.
4.6 Day to day variability:

In order to study the day to day behaviour of the $\left[ O \right]/\left[ O_2 \right]$ profiles, measurements made around the same time but on different days Danilov et al. (1982) are grouped by the author. They are included in figure 4.1. In figure 4.1(a) the results of the morning flights from THUMBA are presented. As such there are three sunrise time profiles presented there in. Interestingly the profiles 3 and 4 which were obtained on consecutive days match well though the values are nearly 2.5 times less than the nighttime values. These are also lower as compared to profile 2 obtained on a different period around the same time by a factor of 1.8-2.

$\left[ O \right]/\left[ O_2 \right]$ ratios obtained from shipboard measurements ($00^\circ$N and $21^\circ$N geographic latitudes) at the same local time on different days which are depicted in figure 4.1b by profiles 1 and 2 also reveal such large day to day variability (Pâkhunkov, 1972). The difference in ratio is larger at higher altitudes varying from 1.3-5. Also included for comparison are the values obtained from Thumba at noon on a different occasion. It is interesting to note that the values obtained at Thumba agree well with profile 1 at lower altitudes ($\leq 130$ km) and with profile 2 higher above. The six examples presented above indicate that the day to day variability is important and probably more important than the day to night variability.
The ratio $\frac{[O]}{[N_2]}$ is accepted to be an important factor in the theory of F-region because it is this ratio that controls the electron density (Shimazaki 1965).

It has been shown that if the rate coefficients of the reaction

$$N_2^+ + 0 \rightarrow O^+ + N_2 \quad \cdots \quad (4.11)$$

and

$$N_2^+ + 0 \rightarrow N_0^+ + N \quad \cdots \quad (4.12)$$

are less than $10^{-11}$ cm$^3$ sec$^{-1}$ the equilibrium electron density is represented to a good approximation by

$$[N_e] = \frac{j[O]}{\alpha_1[O_2] + \alpha_2[N_2]} \quad \cdots \quad (4.13)$$

Where $j$ is the photo ionization rate of 'O', $\alpha_1$ and $\alpha_2$ are the rate coefficients of the ion-molecule interchange reactions

$$O^+ + O_2 \xrightarrow{\alpha_1} O_2^+ + O \quad \cdots \quad (4.14)$$

$$O^+ + N_2 \xrightarrow{\alpha_2} NO^+ + N \quad \cdots \quad (4.15)$$
respectively. Normally the electron density \( [N_e] \sim O^+ \) in the F-region altitudes.

Because \( \alpha [O_2] \) and \( \alpha [N_2] \) are of the same order of magnitude around 220 km, it may be safely said that '\( N_e \)' varies approximately in proportion to \( [O]/[N_2] \).

Though the relevance of \( [O]/[N_2] \) is higher above-in the F-region, their density distributions, under diffusive equilibrium conditions, are governed by their distributions in the lower thermosphere, especially at and above the turbopause level. Hence the measurements of this parameter under varying conditions in the lower thermosphere becomes important.

4.8 \( [O]/[N_2] \) Morning to night variation:

Similar to \( [O]/[O_2] \) measurements of \( [O]/[N_2] \) are grouped to study the day to night variation. Profiles labelled 1 and 2 in figure 4.2a(page 116a) represent the values obtained during midnight and early morning hours at Thumba. The midnight values are larger, above 110 km and the difference increases with altitude. At 150 km it is larger by a factor of 2. The parallel case of Atlantic measurements in figure 4.2b (profiles 1 and 3) reveals that the night-time values are less by about 1.7. The difference gets reduced at 150 km. It is clear from the figures that in the case of \( [O]/[N_2] \) the day to night variability is significant.
Fig. 4.2: a) Similar study of $[O]/[N_2]$ at equator and

b) at Atlantic Ocean
4.9 Day to day variability:

Returning to figure 4.2a the profiles labelled 2, 3 and 4 refer to the early morning measurements on different occasions. Profiles 3 and 4 were obtained on consecutive days. They seem to match well, suggesting little day-day variation. When these two profiles are compared with profile 2 it is found that profile 2 values are larger by a factor of 1.5 below 130 km and by a factor of 1.3 higher above. Similar exercise when performed on the Atlantic profiles 2 and 3 of figure 4.2b it is found that the probable variation is by a factor of 1.4 all through. The measurements from Thumba on an entirely different occasion is seen to agree within a factor of 1.2 of the low latitude Atlantic measurements. On the whole it could be concluded that the day-day variations of the \( \frac{[O]}{[N_2]} \) ratio is around by a factor of 1.3.

4.10 \( \frac{[O_2]}{[N_2]} \) Day to night variations:

In the earlier relative concentrations, where an uncertainty in the determination of atomic oxygen exists and also that of atomic oxygen itself is a result of the dissociation of molecular oxygen, the ratios showed large variabilities. The \( \frac{[O_2]}{[N_2]} \) ratio, devoid of such problems is one of the most stable parameters. The total sensitivity coefficient of the instrument could be taken as unity for both the species under consideration.
The values obtained in different measurements are plotted in the same fashion as in the earlier cases in figures (4.3a) and (4.3b) (page 118a).

Profiles 1 and 2 in figure 4.3a, obtained from Thumba, reveal that there is hardly any day to night variation. This aspect is further confirmed from the Atlantic ocean measurements depicted in same figure. The profiles labelled 5 and 6 represent the morning and night values of the Atlantic measurements respectively. The agreement is very good as obtained by the author. Another set, of low latitude shipboard measurement by Schaffer (1969) also reveals similar effects, i.e. the day to night variation of $\frac{[O_2]}{[N_2]}$ is extremely small.

4.11 $\frac{[O_2]}{[N_2]}$ Day to day variability:

The day-to-day variability of $\frac{[O_2]}{[N_2]}$ ratio is also brought out in figure 4.3a and 4.3b as obtained from Thumba and Atlantic Ocean, measurements. Profiles 2, 3 and 4 of figure 4.3a represent the early morning values, on different occasions. The values vary by a factor of 1.2 almost all through the altitude of interest. Similar measurements from Atlantic ocean representing day time conditions reveal that the ratio $\frac{[O_2]}{[N_2]}$ changes by 1.2 and they are depicted by profiles 2 and 3 in figure 4.3b. Also plotted are the noon time values as obtained from Thumba labelled as 1. The Thumba values agree well and are similar to one of the Atlantic values.
Fig. 4.3: The ratio \( \frac{[O_2]}{[N_2]} \) as obtained on different occasions at (a) Thumba and (b) Atlantic Ocean.
4.12 Discussion:

The availability of a large amount of data from satellites have enabled investigators to formulate empirical models of the upper atmosphere. Such empirical models provide increasingly accurate picture of thermospheric variations in usable form and have provided many of the basic quantities needed for ionospheric, dynamical, and aeronomic calculations and confirmation of theoretical predictions. Also they provide means to intercompare various data sets which are rarely obtained under identical conditions. The different models range from summaries of data from individual satellite to synthesis of data from many satellites or instruments.

Mass spectrometric data from five satellites and exospheric temperature from four incoherent scatter radars were combined into a global model (MSIS) of thermospheric temperature and composition (N\textsubscript{2}, O, He, Ar, H, O\textsubscript{2}) above 120 km by Hedin et al (1977). The basic method is a spherical harmonic expansion of exospheric temperature in a latitude local time frame with allowance for variations with the day of the year and solar and magnetic activity. Corresponding expressions for temperature and density of individual species at 120 km are available. Phase difference among the species in the diurnal, annual and other variations are discussed later.
are thereby introduced realistically. The model is seen to depict the observed features fairly well.

Complementing this synthesis of several data sets, were individual models of the ESRO-4 composition measurements (Von Zahn et al, 1977) and OGO-6 Fabrey-Perot temperature measurements (Thullier et al, 1977). In general major features of these models are in fairly good agreement with the MSIS model.

Jacchia (1977) produced a revised model on the basis of satellite drag. The now well known changes in phase of the daily variations with height are introduced through a pseudo-temperature for each species and composition variations during magnetic storms are introduced via changes in turbopause height.

Alcayde et al (1978) compared several models and demonstrated the difficulties encountered when inferring temperature from density measurements and composition from satellite drag.


Also formulated is yet another model (C-model) by combing data from a larger number of satellites and in the same lines as the MSIS model (Kohnlein 1980).
4.12.1 Composition variation as represented by the models:

The empirical models reflect positive correlation between the temperature and density, with the solar radio flux index in the 10.7 cm wave length. The radio flux is assumed to track variation in EUV flux from the sun thus reflecting variations in the thermospheric energy input. Nevertheless, the determination of, even qualitative relationship between $F_{10.7}$ and EUV measurements eludes the investigators (Hinteregger et al 1978). Joule heating & particle precipitation are also treated as potential energy inputs (Straus 1978).

The diurnal and semidiurnal tidal variations in the thermosphere and the systematic diurnal phase and amplitude variations with altitude are fairly well represented by MSIS and other models (Mayr and Harris 1977; Forbes 1978). The lighter elements are known to be phase shifted to earlier local times as probably due to global transport and wind induced diffusion.

Magnetic storm time variation as represented by various models, variation of individual species like $O_2$ and other minor constituents are discussed by Hedin (1979) in a comprehensive review.

Also major thermospheric composition variations are interpreted in the theoretical studies by Mayr et al (1978) and by Mayr and Harris (1979). They have increasingly concentrated on global circulation patterns driven by solar
radiation, magnetospheric energy and momentum sources and wave sources in the lower atmosphere as the causative phenomena for the density fluctuations and the accompanying redistribution of minor species. It was expressed by them that changes in eddy diffusion and turbopause are probably minor perturbations in the general picture, particularly for variations with time scales of the order of a day or less.

Though the different models so far discussed above represent some of the prominent features in the variation of the upper atmospheric composition it should be mentioned that they are basically statistical and can never completely substitute for the original measurements, particularly when studying phenomena with short time and spatial scales.

This aspect becomes evident in the representation of the relative composition by the models corresponding to those occasions when insitu measurements are available and presented through figures 4.1 to 4.3. The OGO.VI model, the formulation of which was readily available is made use of in this study. It is taken as a representation of other models; the differences if any could probably be minor in nature. In figures 4.4 and 4.5 are presented the ratios of the different neutral constituents viz. \([\text{O}] / [\text{O}_2]\), \([\text{O}] / [\text{N}_2]\). Hardly any variation is predicted in the \([\text{O}_2] / [\text{N}_2]\) ratio either in the day to night or in the day to day variations and therefore is not represented by a figure.
Fig. 4.4: Representation of the \( \frac{[O]}{[O_2]} \) ratio corresponding to the actual measurements i.e. both for Thumba and the Atlantic ocean measurements (OGO VI model).

Fig. 4.5: OGO VI model representation of the \( \frac{[O]}{[N_2]} \) ratios for equatorial and low latitude stations.
The predicted day to night and day to day variations for the coordinates of Thumba and Atlantic Ocean are depicted in figure 4.4. The night time values are smaller than the day time ratios which is contrary to the measurements. The profiles running parallel to each other, which is typical of a model, show lesser fluctuations and lesser variability on the whole than the measured profiles both for Thumba and Atlantic Coordinates. The results of a similar exercise performed for obtaining $\frac{[O]}{[N_2]}$ ratio are presented in figure 4.5. The night-time values are smaller than the day time values, a result contradicting the actual measurements. Similar to $\frac{[O]}{[O_2]}$ ratio the $\frac{[O]}{[N_2]}$ model profiles show lesser variations; let be day to night or day to day.

The above presentation brings out clearly the deficiencies of the statistical models calling for more refinements. The possible factors that may result in these discrepancies are discussed later.

With this background some of the interesting results in the relative composition variation as obtained from Thumba are discussed.

The molecular nitrogen $N_2$ and oxygen $O_2$ are considered as the relatively stable species and their measurements could be made to accuracies better than 10%. The abundance of $N_2$ makes it close to the mean mass at least in the lower thermosphere. Further it is also considered to
represent the neutral temperature more truly than any other
species (Hedin et al, 1977a). The neutral temperature
profile is usually derived from the scale height distri-
bution of \( N_2 \). The relative concentration of \( \frac{[O_2]}{[N_2]} \)
reveals hardly any day to night changes. Therefore only
day to day variations are important as far as these species
are concerned. As such the day-day changes could be due to
either changes in \( [O_2] \) or in \( [N_2] \) densities or both.
Changes in \( [N_2] \) densities if any would have reflected
in the temperature profile. Using Jacchia 77 model the
exospheric temperature \( T_{eo} \) was obtained for the different
occasions.

The values corresponding to different measurements
are tabulated in Table (4.1).

<table>
<thead>
<tr>
<th>Flight launch site</th>
<th>exospheric temperature ( T_{eo} ) °K</th>
<th>Time IST</th>
<th>Coordinates</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Geographic</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Lat.  Long.</td>
</tr>
<tr>
<td>05.22 21.4.75 Thumba</td>
<td>765</td>
<td>1330</td>
<td>8.53°N 76.95 E</td>
</tr>
<tr>
<td>05.50 20.12.78 &quot;</td>
<td>965</td>
<td>0230</td>
<td>&quot;</td>
</tr>
<tr>
<td>05.57 21.12.78 &quot;</td>
<td>937</td>
<td>0615</td>
<td>&quot;</td>
</tr>
<tr>
<td>05.69 27.4.80 &quot;</td>
<td>1086</td>
<td>0645</td>
<td>&quot;</td>
</tr>
<tr>
<td>05.70 28.4.80 &quot;</td>
<td>1094</td>
<td>0620</td>
<td>&quot;</td>
</tr>
<tr>
<td>12.5.68 Atlantic</td>
<td>1240 LT 1400</td>
<td>10°N 35° W</td>
<td>1</td>
</tr>
<tr>
<td>21.5.68 &quot;</td>
<td>1370 LT 1400</td>
<td>21.6°N 20.16° W</td>
<td>1</td>
</tr>
<tr>
<td>22.5.68 &quot;</td>
<td>1100 LT 0050</td>
<td>24.3°N 20° W</td>
<td>1</td>
</tr>
</tbody>
</table>
If the ratio changes are purely due to \( \left[ N_2 \right] \) changes, then it should have reflected in the day to night profiles also. For example, in the Atlantic measurements the \( T_0 \) gets reduced by 270° from day to night. The corresponding changes in \( \left[ N_2 \right] \) values must have yielded a different \( \left[ O_2 \right]/\left[ N_2 \right] \) ratio in the night. With the repeatability of the results that less or no day to night variation in their ratios, one may conclude \( \left[ O_2 \right] \) values also change nearly by the same factor and in the same sense. There must then be a different mechanism by which such a large day to day variability is obtained. The variation in \( \left[ O_2 \right] \), mainly due to photo-dissociation and chemistry apart from the level of the turbopause may be the cause for the same.

4.12.2 \( \left[ O \right]/\left[ N_2 \right] \) ratio:

It had been pointed out that the \( \left[ O \right]/\left[ N_2 \right] \) ratio has significant day to night as well as day to day variations. The day to night variation being larger than the day to day variations by a factor of 1.6. Though the day to day variations cannot be completely ignored as insignificant the dominant variation of this ratio is day to night. As the day to day variation could then be systematic one could evolve the possible seasonal variation as had been shown in connection with the F-region electron density variation (Ruster and King, 1973).
The day to night variability of \( [\text{O}] / [\text{O}_2] \) ratio, though significant, turns out to be smaller than the day to day variability. Sometimes day to day variations turn out to be so high that they can completely mask the diurnal values, particularly when the cause of these changes are fast enough (within a few hours). The large discrepancies observed in the measurements of this ratio on different occasions could be due to the large day to day variability. For example, the contradicting results of Schaffer (1969) and Kasprazek (1968) in the diurnal variability of \( [\text{O}] / [\text{O}_2] \) at midlatitudes and the measurements at Thumba and Atlantic could be explained on the basis of the day to day variability of this ratio.

Offerman (1974) while compiling the composition measured by different techniques has discussed the possible seasonal variation of \( [\text{O}] / [\text{O}_2] \). The \( [\text{O}_2] \) densities did not show any seasonal variation and all the variations were attributed to the variations in \( [\text{C}] \) only. The data points presented were scattered and therefore they were grouped at 60 day intervals and the mean value of each group was calculated and a possible seasonal variation by a factor of 2 around 150 km in \( [\text{O}] / [\text{O}_2] \) with a summer minimum was obtained. These measurements were also compared with OGO-6 measurements. In the context of the present investigation, as had been clearly indicated by the same technique,
that the day to day variability is very large some times as large as a factor of 5 around 150 km, the sort of 60 day averaging of the values obtained by different techniques, to evolve a seasonal variation may have to be revised. No doubt, this averaging technique has yielded fairly agreeable results with the assumed variations of Ruster and King (1973) in order to explain the F-region ionospheric electron densities. Wherever discrepancies are observed, may be due to the day to day variations not being given their due weightage.

When we analyse the sort of variations that are obtained more critically, the larger day to day variation seems to be caused by the molecular oxygen, as evident from the qualitative representation in the table 4.2. From the altitude variation profiles it is seen that the trend in the variation of \( \frac{[O]}{[O_2]} \) is complementary, conforming with the conclusion that it is the molecular oxygen that governs the variation in the ratios. In earlier studies by Offerman (1974) it was suggested that it might be the atomic oxygen which probably governs the variations. However 'O' variations seem to be significant in the day-to-night variation of \( \frac{[O]}{[N_2]} \) ratio.

<table>
<thead>
<tr>
<th>Ratio</th>
<th>day to night</th>
<th>day to day</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \frac{[O]}{[O_2]} )</td>
<td>smaller</td>
<td>larger</td>
</tr>
<tr>
<td>( \frac{[O]}{[N_2]} )</td>
<td>larger</td>
<td>smaller</td>
</tr>
<tr>
<td>( \frac{[O_2]}{[N_2]} )</td>
<td>very small</td>
<td>larger</td>
</tr>
</tbody>
</table>
4.13 TURBOPAUSE AND ITS IMPLICATIONS:

The major constituents of the atmosphere are known to have been well mixed up to about 100 km, above which diffusive separation of the individual constituents sets in. The dissociation of oxygen has been studied in detail by Nicolet and Mange (1954). They had pointed out that even though 'O₂' molecules are dissociated by solar radiation, notably Lyman α (1216 Å) and Schumann Runge continuum (<1750 Å), the lifetime of O₂ molecule above 100 km is some days. However 'O₂' concentration gets replenished in a time shorter than this by diffusion maintaining hydrostatic equilibrium in [O₂]. Meadows and Townsend (1960); Pokhunkov (1963); Scaffer and Nicholas (1964), have shown that [Ar]/[N₂] ratio falls off with altitude, as is to be expected if the constituents are in diffusive equilibrium. Hedin and Nier (1965) and Krankowsky et al (1968) have treated the individual constituents quantitatively to arrive at the same conclusion. The question of diffusive separation is linked to that of turbulence, which is believed to be absent at heights above about 100 km. Vapour cloud experiments by Blamont and de Jager (1961) Ananda Rao et al (1978); Desai et al (1975) and Rosenberg (1964) have recorded the level where turbulence ceases, to be between 91 to 115 km.
Hines (1963, 1971) has reviewed the problem of mixing of major constituents with special reference to turbulence. If mixing were to suddenly cease, the gases would separate with a relative velocity which is found by balancing the resistance to motion due to collisions, against the buoyancy force which tends to separate the gases. Up to about 100 km the power available from turbulence can maintain the mixing. The precise level of the 'turbopause' where turbulence ceases is also dependent on the strength of the mechanism causing turbulence.

Unless some other mechanism for opposing separation is effective at greater heights, diffusive separation of atmospheric gases must be established not far above the turbopause. If the level of the turbopause should change the atmospheric composition at higher levels would be affected.

Recently much interest has been shown in the problems of mesospheric turbulence and in particular the turbopause and its variations (Zimmerman and Murphy, 1977; Vonzahn and Herwigs, 1977; Blum and Schuchardt, 1977; Schuchardt and Blum, 1977; Danelcv et al, 1979) have made use of a large number of rocket flight data where the $[\text{Ar}] / [\text{N}_2]$ ratio was used as an indicator of turbopause level. The level where the ratio deviates from its ground value is considered as the turbopause level. They have tried to bring out some of the possible systematic
variations of this important parameter both for high and equatorial latitudes. As had been shown by Vonzahn and Herwig (1977) the turbopause height $h_e$ is equally important for atmospheric modelling.

4.13.1 $\frac{[\text{Ar}]}{[\text{N}_2]}$ at equatorial latitudes:

Argon and Nitrogen both being relatively stable species and fairly well separated in the mass scale, are chosen as ideal indicators of turbulence. When thorough mixing prevails the relative composition of these species remains same, up to the point where the diffusion effects start dominating. From then on, $[\text{Ar}]$ the heavier of the two starts falling off faster with altitude as compared to $[\text{N}_2]$, the difference in decrease being proportional to their mass difference. The turbopause level as derived by the above method for the equatorial flights, are listed in Table (4.3).

**TABLE 4.3**

<table>
<thead>
<tr>
<th>Flight no.</th>
<th>Date</th>
<th>Time IST</th>
<th>$h_t$ turbopause level km</th>
</tr>
</thead>
<tbody>
<tr>
<td>05.22</td>
<td>21.4.75</td>
<td>1330</td>
<td>107</td>
</tr>
<tr>
<td>05.50</td>
<td>21.12.78</td>
<td>0230</td>
<td>112</td>
</tr>
<tr>
<td>05.51</td>
<td>21.12.78</td>
<td>0615</td>
<td>102</td>
</tr>
<tr>
<td>05.69</td>
<td>27.4.80</td>
<td>0645</td>
<td>119</td>
</tr>
<tr>
<td>05.70</td>
<td>26.4.80</td>
<td>0620</td>
<td>101</td>
</tr>
<tr>
<td></td>
<td>22.2.68</td>
<td>0110</td>
<td>109 Golišnev et al (1969)</td>
</tr>
<tr>
<td></td>
<td>22.2.68</td>
<td>1655</td>
<td>112 }</td>
</tr>
</tbody>
</table>
In figure 4.6 (p. 131 a), the variations of the $\frac{[Ar]}{[N_2]}$ with altitude are depicted, as obtained from different measurements made at Thumba. Out of these the profiles corresponding to the early morning hours 1, 2 & 3 reveal the $h_t$ level to vary from 100-120 km. Profile No. 2 shows the turbopause level to be very high (119 km). Such a large value of turbopause has not been observed so far at equatorial latitudes. It seems that during this flight some unusual conditions must have been prevalent leading to increased turbulence in the lower thermosphere.

The turbulence regime around $h_t$ is dependent on temperature at that altitude. The altitude profiles of $\frac{[Ar]}{[N_2]}$ or $\frac{[O]}{[N_2]}$ which are sensitive indicators of temperature, do not show any peculiarities between profiles 2 and 1. Therefore the turbulence increase is not definitely due to pure temperature. The peculiar oscillatory structure as observed in the $[NO^+]$ and $[O_2^+]$ profiles (figure 3.15 "p. 101a") measured simultaneously suggests, that on this occasion probably some sort of a wave like disturbance (internal gravity wave, acoustic gravity wave type) might have arrived from below dissipating their energy and thus increasing the turbulence.

Gravity waves, either directly or through a mechanism of turbulence generation, enhance the mixing ability of the atmosphere in a region where molecular diffusion is a competitive process (Hines 1965). As such
Fig. 4.6: The turbopause level as obtained by the deviation of the $[\text{Ar}]/[\text{N}_2]$ ratio from the ground level value for the various flight.
the origin of such disturbances are uncertain calling for more such early morning measurements.

As shown in Table 4.3 the $h_t$ level (turbopause level) has a wide variation diurnally, as well as day to day, indicating that the parameter which controls the turbulence and hence the turbopause level, either has no systematic variations or it is a function of more than one independent parameter. Mass spectrometric data obtained under varying geophysical conditions on different days were compiled by Danilov et al (1980a) in order to obtain systematic variations for equatorial latitudes. On many occasions the lowest altitude where $\left[\frac{\text{Ar}}{\text{N}_2}\right]$ ratios were obtained, happened to be above the turbopause level and the values had to be extrapolated. In order to make this extrapolation the temperature profile obtained in each case from the same extrapolated $\left[\frac{\text{Ar}}{\text{N}_2}\right]$ height profile was used. In general the accuracy of the turbopause level determination is $\pm 1$ km.

Another interesting observation is the wavy structure in the $(\text{Ar})/(\text{N}_2)$ variation with altitude. Under diffusive equilibrium conditions the profiles are normally expected to run parallel as will be discussed later. Some sort of a bump in the altitude range of 130-135 km also calls for an explanation. Though the profiles are obtained under varying conditions and at different seasons, the altitude of the bump seems to remain unaffected. This aspect needs further investigation.
Though the observations made from Thumba are insufficient to draw any definite conclusion on the variation of the turbopause with geomagnetic activity or neutral temperature, or with season etc., Danilov et al (1980a) in a compilation of equatorial and low latitude data have shown a definite trend in the variation of \( h_t \) level. Figure 4.7 and 4.8 (page 133a) depict the variation of \( h_t \) with \( A_p \) and with neutral temperature \( T_n \), at 120 km for both equatorial and polar latitudes. The reverse behaviour of the \( h_t \) level with respect to \( A_p \) is interesting. As for the temperature dependence is concerned the variations are similar, although the data points are very much scattered.

4.13.2 Discussion:

The turbopause level is assuming an increasingly important role in the studies of upper atmosphere because of its possible effect in deciding the distribution of different species higher above in the thermosphere. It is becoming clearer now that the turbopause height \( h_t \) is a parameter no less important than temperature for upper atmospheric modelling. Variations in the upper atmospheric structure i.e. in the density, pressure and temperature are known to have their origin in the rotation of the earth on its axis (diurnal), revolution of the earth about the sun (seasonal) and the sunspot cycle (solar cycle variation). In addition physical properties depend upon geographic and magnetic latitudes also.
Fig. 4.7a: Variation of the turbopause level ($h_t$) with $A_p$ (Danilov et al 1979, 1980)
The cause of the variations discussed above is being thought of due to the i) atmospheric dynamics (winds) thus disrupting diffusive equilibrium and ii) maintaining diffusive equilibrium and changing the turbopause level \( (h_t) \). Some efforts had also been made where the effects of both winds and \( h_t \) level are taken into account (Blum and Schuchuardt 1980).

It is so far believed that the turbopause level, usually decided by the strength of turbulence could be described by a single parameter namely the Eddy diffusion coefficient \( 'K' \). Detailed observational data on the height variations of \( 'K' \) and also it is variations with time and location are lacking. Zimmerman et al (1972) using grenade data and Danilov et al (1979, 1980a) arrive at the conclusion that the turbulence is higher during winter than summer while Incoherent Scatter data (Alcayde et al 1979), Partical reflection data (Schelegal et al 1977) and Chemical release data (Teitelbaum and Blamont, 1977) reveal increased turbulence during summer i.e. larger the temperature, more the turbopause height. These contradictory results show the lack of understanding of the phenomena of turbulence in the upper atmosphere.

4.13.3 Effect of \( h_t \) on composition:

Upto the \( h_t \) level all the species, high or low mass relative to \( N_2 \) are kept tied down to the \( 'N_2' \)
distribution. The scale height of all the species remain the same as that of \( 'N_2' \), purely due to thorough mixing. Above turbopause each of the species is let to redistribute on its own and they have their own scale height. Diffusive equilibrium prevails. If due to some reason the \( h_t \) level is increased, upto this new level \( 'N_2' \) will govern the distribution. The lighter species would have decreased faster with altitude while the heavier ones would have decreased slower up to this level. This essentially means that more number of heavier species and less number of lighter species are available for redistribution higher above. The altitude profiles of these species under diffusive equilibrium would have run parallel to the original profile. Figure 4.9 (p.136a) illustrates this aspect. At a particular altitude the heavier species get enhanced and the lighter ones get depleted. The differential treatment depending upon the species related to the mean mass is brought out in this illustration. Sinha and Chandra (1974) have tried to explain the seasonal and magnetic storm related changes theoretically by assuming an effective reduction in the winter time eddy diffusion coefficient by an order of magnitude. Jacchia et al (1976) in trying to obtain a relation between the turbopause level and neutral temperature have started that \( \frac{dz_H}{dT} \) i.e. the variation in \( h_t \) level due to temperature may vary from 30 m - 50 m/°K. This simple illustration in figure 4.9 seems to explain, some of the observed features like winter \( 'He' \) bulge (Keating, 1968) and summer Argon
bulge (Von Zahn, 1973) in midlatitudes. In actual practice the interpretation of the measured \( h_t \) levels does not turn out to be that simple. The eddy diffusion coefficient must then be a function of at least two mutually independent variables. In this context some of the measured composition would be discussed below.

4.14 Variation of \( \frac{[O_2]}{[N_2]} \) with respect to turbopause \( (h_t) \) and exospheric temperature \( (T_{\infty}) \)

To start with, the stable species 'O\(_2\)' and 'N\(_2\)' are chosen from some of our own measurements. Though 'O\(_2\)' gets dissociated, as had been discussed earlier, it gets replenished by diffusion thus allowing this species to be in diffusive equilibrium. In the lower thermosphere since 'N\(_2\)' is close to mean mass it does not get affected by turbulence or turbopause level. Only temperature changes affect its distribution. Representative altitudes of 125 km and 150 km are chosen to study their variation.

As seen from table 4.4 (page 137), there does not seem to be any direct correlation between the level of \( h_t \) and \( \frac{[O_2]}{[N_2]} \) ratios. Figure 4.10 (page 136) brings out the variation of this ratio with turbopause level and the exospheric temperature independently. As regards to variation with respect to \( h_t \) the \( \frac{[O_2]}{[N_2]} \) ratio appears to have a minimum around 103.5 km increasing at higher and lower altitudes. However, with respect to \( T_{\infty} \) a trend
Fig. 4.9: The effect of $h_e$ on the distribution of the different species higher above (Blum and Schuchardt 1978).
of increasing $\left[ \frac{O_2}{N_2} \right]$ ratio with increasing $T_e$ at any particular height is obtained. But the data points are very much scattered.

### TABLE 4.4

<table>
<thead>
<tr>
<th>Flight</th>
<th>ALT 125 km</th>
<th>ALT 150 km</th>
<th>$h_t$ km</th>
<th>$T_e^{\circ}$K</th>
</tr>
</thead>
<tbody>
<tr>
<td>05.69</td>
<td>0.165</td>
<td>0.142</td>
<td>119</td>
<td>1085</td>
</tr>
<tr>
<td>05.70</td>
<td>0.155</td>
<td>0.13</td>
<td>101</td>
<td>1094</td>
</tr>
<tr>
<td>05.50</td>
<td>0.15</td>
<td>0.12</td>
<td>112</td>
<td>965</td>
</tr>
<tr>
<td>05.51</td>
<td>0.14</td>
<td>0.118</td>
<td>102</td>
<td>937</td>
</tr>
<tr>
<td>05.22</td>
<td>0.124</td>
<td>0.115</td>
<td>107</td>
<td>765</td>
</tr>
</tbody>
</table>

Instead, if the exospheric temperature ($T_e$) which represents the total heat input and $h_t$ are combined, the product of $h_t \times T_e$ shows a direct correlation to the values of $\left[ \frac{O_2}{N_2} \right]$ at any altitude, as depicted in Fig.4.11 (page 138a). The scatter in the data points is far reduced than the previous plots and the plots corresponding to different altitudes fall in straight lines parallel to each other.

The results of the above analysis indicate clearly that the neutral composition at any altitude is governed by two mutually independent parameters namely the turbopause and the exospheric temperature. One of the most remarkable aspect is that the data inspite of being obtained
Fig. 4.10: a, b) \( \frac{[O_2]}{[N_2]} \) as obtained from the flights from Thumba Vg the turbopause level \( h_g \) at two representative altitudes viz 125 km and 150 km. c, d) \( \frac{[O_2]}{[N_2]} \) Vg model exospheric temperature \( T_{\infty} \) values for the occasion.
on different days, times, seasons and other geophysical conditions bear a simple relation to the product of \( h_t \) and \( T_\infty \). As such the precise level of turbopause, is dependent on the strength of the mechanism that causes turbulence. As had been discussed earlier gravity waves and tides either directly or indirectly affect the turbopause. Therefore local disturbances, like waves depositing energy, are likely to have an independent control on the turbopause level and hence on the neutral composition itself. So, it becomes essential to incorporate this parameter \( h_t \) also in evolving systematic variations in neutral composition, in particular by atmospheric modelling. The present day models either take the level of \( h_t \) at fixed altitudes or hypothesize its variation only during magnetically disturbed periods.

Similar exercises to get relation between the variation of \( [O]/[N_2] \) and \( [O]/[O_2] \) with \( h_t \) and \( T_\infty \) are difficult because of 'O' being a product of dissociation of 'O_2' by solar EUV. One more unknown parameter in the form of the dissociating radiation complicates the problems. As such increase in the flux of dissociating radiation would increase 'O' density. But if the eddy mixing increases with temperature this would increase the downward flow of atomic oxygen, i.e. decreasing 'O'. The net effect of these two mechanisms must be taken into account while explaining the variation of either \( [O]/[N_2] \) or \( [O]/[O_2] \) ratios.
Fig. 4.11 \( \frac{[O_2]}{[N_2]} \) as measured at two representative altitudes. An empirical relation could be derived.