A RADAR STUDY OF THE THERMOSPHERE

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Submitted to the Department of Meteorology on 14 January 1975 in partial fulfillment of the requirements for the degree of Master of Science.

ABSTRACT

Measurements obtained by the incoherent scatter radar at Millstone Hill (42.6°N, 71.5°W) are used in a study of the mid-latitude thermosphere. The measurements of plasma temperatures, densities, and drift velocities are used in conjunction with neutral composition data from the OGO-6 quadrupole mass spectrometer to obtain estimates of the neutral temperatures and drifts along the magnetic field line. A least squares fitting procedure is then used in a model of the neutral particles to find the zonal and meridional gradients that would lead to the observed temperatures and drifts. Two equinox days are analyzed.

The harmonic coefficients of the meridional temperature gradient were found to be important to the fourth harmonic and to be of the order of 50 K/rad. The gradient shows peaks after sunrise and sunset and an unexplained peak after noon. The mean gradient can be of either sign. Exospheric temperatures within 10° of latitude are calculated from the diurnal variations in the temperature and its meridional gradient at Millstone Hill.

The derived neutral zonal wind is generally eastwards from mid-afternoon to after midnight and then westwards with wind speeds of up to 200 or 300 m/sec. The meridional wind is of the order of 25 or 50 m/sec polewards during the day. At night, the velocities are equatorwards and are usually larger than the daytime velocities because of the reduced electron densities and hence reduced ion drag. Velocities at night are of the order of 50 to 200 m/sec equatorwards depending on the electron densities.

The technique used to derive these winds and meridional temperature gradients is uncertain to about ±30%.
The major uncertainties are in the ion-neutral collision cross-sections, in the electric fields present, in statistical uncertainties of the experimental measurements leading to neutral drift estimates, and in the absence of the north-south non-linear term in the equations of motion.

The results of the present study are compared to other studies, theory and results in order that a better understanding of the technique can evolve.

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1. INTRODUCTION

The thermosphere begins above the mesopause around 80 km. It is a region of low air density, the number density at 300 km being about $10^9 \text{ cm}^{-3}$, while that at the surface is about $10^{19} \text{ cm}^{-3}$. The sun's influence in this region is very strong, heating and ionizing the air during the daylight hours. The maximum ionized number density is about $10^6 \text{ cm}^{-3}$ and occurs at an altitude of around 300 km. This peak is in the F region of the ionosphere. The kinetic temperature in this region is of the order of 1000$^0\text{K}$ or more. At night, the peak number density of the ionized species in the F region can fall by a factor of 5, while the temperature can drop by 200 or 300$^0\text{K}$. It is this heat imbalance between night and day, along with latitudinal differences, which drives the thermospheric winds.

The major neutral constituents are $N_2$, $O_2$, and $O$, which are distributed in diffusive equilibrium according to their scale heights. The major ionized species are $O_2^+$, NO$^+$, and O$^+$. Between 200 and 1000 km, $O$ is the dominant neutral species and $O^+$ the dominant ionized species.

The presence of ions and electrons in the thermosphere makes it possible to study this region through the use of the incoherent scatter radar technique.
(Evans, 1969). This technique measures charged particle densities, temperatures, and drift velocities. Since the neutrals and ions affect each other through collisions, this information on the charged species can yield information on the neutrals as well.

The purpose of the present study is to show that, given observations of charged particle number densities, temperatures, and drifts, and assuming electric field strengths and approximate neutral number densities, it is possible to deduce the diurnal variation of neutral winds and temperatures in the thermosphere. In particular, it is possible to deduce the north-south temperature gradients near the observing station.

Chapter 2 briefly outlines the experimental data taken at the Millstone Hill incoherent scatter facility (42.6°N, 71.5°W) and the information on the neutrals that can be derived therefrom. Chapter 3 discusses the numerical model for the neutrals, and Chapter 4 contains a description of a least squares fitting procedure which fits experimentally derived neutral winds to those derived from the numerical model. The results of this fitting procedure for two equinox days are presented and discussed in Chapter 5. Finally, Chapter 6 summarizes the conclusions and the uncertainties of the technique, and makes suggestions for future work.
2. EXPERIMENTAL MEASUREMENTS

2.1 Data Collection

The parameters that are directly determined from the radar backscatter measurements are electron densities \( N \) \((\text{cm}^{-3})\), electron and ion temperatures \( T_e \) and \( T_i \) \((\text{K})\), and the vertical plasma drift velocity \( V_{iz} \) \((\text{m/sec})\). \( N \) is sampled between 150 and 900 km and has a height resolution of 15 km, which is the vertical distance illuminated by the radar pulse. \( T_e \), \( T_i \), and \( V_{iz} \) are measured using two different, longer pulses, their height resolution being 75 km below about 450 km and 150 km above. The centers of the scattering volumes are called the nominal heights of the measurements and are located at 225, 300, 375, 450, 600, 750, 900, and 1050 km for \( T_e \), \( T_i \), and \( V_{iz} \). The signal back-scattered from these volumes is dependent on the electron density which varies with height, as well as on a triangular weighting imposed by the finite pulse length and the filter response. Therefore, the "true" centers tend to be shifted towards the electron density peak, which is usually near 300 km. These shifted centers are referred to as equivalent heights and should correspond to the true height in the atmosphere.

The electron density is lowest at night so the backscattered signal is weakest then; often results taken at 225 km are unreliable at night for this
reason. The time resolution for the measurements is about 30 minutes, this being the time it takes to complete one cycle of measurements (or run). Observations are generally carried out continuously for 24 hours.

The quantities $N$, $T_e$, $T_i$, and $V_{iz}$ are smoothed by a polynomial fitting routine called INSCON, devised by J. M. Holt of Millstone Hill. This provides a smooth history of the parameters in height and in time. The heights can be nominal or equivalent depending on what is desired. The following subsections describe how these parameters can be used with other information to estimate the neutral temperature, the plasma diffusion velocity, and the neutral wind component along the magnetic field lines.

2.2 Exospheric Temperature

2.2.1 Form of the Neutral Temperature

Above about 200 km, the neutral temperature $T_n$ increases exponentially to a limiting value around 450 km. This temperature is called the exospheric temperature, $T_\infty$. The exponential increase is modeled in the Bates (1959) form

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

(2.1)

where $z$ is the height in km, $T_n(z)$ is the neutral temperature in $^\circ$K at height $z$, $T_{120}$ is the neutral temperature at 120 km, and $s$ is the shape parameter. $T_{120}$ was
chosen to be $355^\circ$K and $s$ to be $0.020$ km$^{-1}$. Changing $T_{120}$ by $\pm 100^\circ$K introduces less than $\pm 1\%$ error in $T_\infty$, and varying $s$ between $0.015$ and $0.030$ km$^{-1}$ introduces less than $\pm 2\%$ error in $T_\infty$ (Salah and Evans, 1973). Knowing values for $T_n$, one can then estimate a value for $T_\infty$.

2.2.2 Heat Balance Equation

The neutral temperature $T_n$ is found through the ion heat balance equation. Above about 200 km, the temperatures of the electrons $T_e$, ions $T_i$, and neutrals $T_n$ are different. At low energies when other excitation methods become ineffective, the photoelectron kinetic energy is dissipated in the electron gas, increasing $T_e$. Some of this energy is transferred to the ions through elastic collisions with electrons. In turn, some of this is passed on to the neutrals through ion-neutral collisions. The ion thermal energy balance requires that the heat transferred from the electrons to the ions equal the heat transferred from the ions to the neutrals. This is written as (Salah and Evans, 1973)

$$
4.82 \times 10^7 N n(0^+) \frac{T_e - T_i}{T_e^{3/2}} = \left[ 6.6 n(N_2) + 5.8 n(O_2) + 0.21 n(O)(T_i + T_n)^{1/2} \right] n(0^+)(T_i - T_n) \quad (2.2)
$$

This equation assumes that the major ion is $O^+$. It
further assumes that the major neutrals are O, N₂, and O₂. These assumptions are valid roughly between 250 and 600 km. At 300 km, \( n(N₂)/n(O) \approx 0.2-0.3 \) and \( n(O₂)/n(O) \approx 0.008-0.016 \). These ratios decrease with altitude.

The values for N, Tₑ, and Tᵢ are taken from the polynomial fit. The values for n(O), n(N₂), and n(O₂) are unknown and must be taken from a different source, usually a model. Two available models are the Jacchia (1971) model based on satellite drag data, and the model of Hedin et. al. (1974) based on mass spectrometer measurements of the OGO-6 satellite. Hedin et. al. compared the two model densities and found that the shape of the diurnal density curves at 45⁰ latitude and 450 km (mostly n(O)) are in quite good agreement although the OGO-6 magnitudes are about 17% higher. Since \( (Tᵢ-Tₑ) \leq 100^⁰K \) at 300 km and \( Tₑ \) is of the order of 1000⁰K, a 20% error in n(O) will result in a 20⁰K or 2% error in \( Tₑ \). The present study uses Jacchia (1971) model densities in the determination of the exospheric temperature.

Ignoring systematic errors in n(O), the total uncertainty in the exospheric temperature is estimated to be \( \pm 50^⁰K \) or 5% from one observation to another (Salah and Evans, 1973). The INSCON program smooths the results in time as well as height, and reduces this
uncertainty to about \( \pm 25^\circ \text{K} \). Figure 2.1 shows the exospheric temperature for March 23-24, 1970 using smoothing in height (Salah and Evans, 1973), and smoothing with INSCON.

2.3 Plasma Diffusion Velocity

In the region above about 180 km, the electron and ion gyrofrequencies \( \omega_{e(i)} = eB/m_{e(i)} \), are much larger than collision frequencies \( \nu_{\text{coll}} \). Here \( e = \) electronic charge, \( B = \) magnetic field strength, and \( m_{e(i)} = \) electron (ion) mass. In the absence of electric fields, particles are confined to travel along the magnetic field lines. With this assumption, a plasma diffusion velocity \( V_{D\|} \text{(m/sec)} \) can be defined. This is the velocity the plasma would have if it were not in diffusive equilibrium. It is defined positive upwards along field lines and can be written as (Schunk and Walker, 1970)

\[
V_{D\|} = -D_a \sin I \left[ \frac{1}{N} \frac{\partial N}{\partial \theta} + \frac{1}{T_e + T_i} \frac{\partial (T_e + T_i)}{\partial \theta} \right] + \frac{38}{T_i + T_h} \frac{\partial (T_e + T_i)}{\partial \theta} + \frac{m_i q}{k (T_i + T_e)} \tag{2.3}
\]

where \( D_a \) is the ambipolar diffusion coefficient (Stubbe, 1968)

\[
D_a = \frac{k(T_e + T_i)}{\nu_{\text{in}} m_{\text{in}}} = \frac{k(T_e + T_i)}{(\nu_{O+O} + \nu_{O+N_2} + \nu_{N_2N_2} + \nu_{O_2O_2} + \nu_{O_2N_2} + \nu_{N_2O_2})} \tag{2.4}
\]

and \( I = \) magnetic dip angle (72° for Millstone Hill),
\( z = \text{altitude}, \) \( g = \text{gravity}, \) \( k = \text{Boltzmann's constant}, \)
\( \mu_{i,n} = m_i m_n / (m_i + m_n) = \text{reduced particle mass}, \)
\( m_n = \text{neutral particle mass}, \) and \( \nu_{i,n} (\propto n(\text{neutral})) = \text{ion-neutral collision frequency}. \) The values used for the ion-neutral collision frequencies were taken from Stubbe (1968) and Banks (1966).

It is assumed that \( n(O^+) = N. \) Values for \( N, T_o, T_i, \) and their height derivatives are found through the polynomial fit of INSCON. \( T_n \) and its height derivative are found from \( T_{\infty} \) and the Bates (1959) profile described in section 2.2.1. Values for \( n(O), n(N_2), \) and \( n(O_2) \) were taken from the model by Hedin et. at. (1974). This represents a small inconsistency in the data analysis, but accurate values for \( n(O) \) are much more important in the determination of \( V_{D_H} \) than they are in the determination of \( T_{\infty}. \) Since \( V_{D_H} \propto 1/n(O) \), a 20% error in \( n(O) \) would create a 20% error in \( V_{D_H} \) compared to a 2% error in \( T_{\infty}. \)

Equation 2.3 is useful between about 250 km and 400 km. Below 250 km, there are no measurements of temperature and above 400 km, the magnification of errors caused by the exponential decrease with height of \( n(O) \) becomes important. Also, hydrostatic balance is more nearly attained, leaving the determination of the diffusion velocity dependent on the poorly estimated temperature derivatives with height. Neglecting
any systematic errors in \( n(0) \) or in the ion-neutral collision frequencies, the errors in \( V_{DI} \) are estimated to be about 1-3 m/sec around the electron density peak at 300 km (Salah and Holt, 1974).

2.3.1 Ion-Neutral Collision Frequencies

Experimental data on ion-neutral collision frequencies is scarce and is estimated to be only \( \pm 25\% \) accurate at temperatures of ionospheric interest (Mason, 1970). This is assuming the mechanism involved in the collision process is understood. For the case of ions in their parent gases, this mechanism at ionospheric temperatures is resonant charge-exchange. For ions in unlike gases, the mechanism depends on the short-range force between ions and neutrals which can be repulsive, or attractive. In the region between about 250 and 600 km, practically the only ion is \( O^+ \), while the dominant neutral is \( O \). In the lower part of this region, \( N_2 \) and \( O_2 \) can be important. The mechanism between \( O^+ - O \) is charge-exchange, and that between \( O^+ - O_2 \) is attractive forces. It is unclear whether the force between \( O^+ - N_2 \) is attractive, as is normally assumed, or repulsive. If it were repulsive, \( V_{O^+ + N_2} \) would be reduced to nearly half the value it would have with an attractive force in the temperature range of interest.

Banks (1966) has used the experimental values and theory to find the collision frequencies. Stubbe
(1968) assumed that the long-range attractive forces added to the charge-exchange mechanism, increasing \( \nu_{o-o} \) by about 25%. Most of the present study was carried out using Stubbe's (1968) values, although one case was carried out using Banks' (1966) values. These are hereafter referred to as the Stubbe and Banks cases.

Through confusion about the definition of the ambipolar diffusion coefficient \( D_a \), the Stubbe case in the present study incorrectly expressed \( D_a = k(T_e+T_i)^2/\nu_{in} \mu_{in} (T_i+T_n) \). The difference between \( T_e \) and \( T_i \) at 300 km is of the order of 100°K at night, 500°K during the day, and 700-1000°K after dawn for the two days analyzed in the present study. This will lead to increases in \( D_a \) of about 5% at night, 25% during the day, and 35 to 50% at dawn. Fortunately, this can be considered to be within the experimental error in \( \nu_{in} \), since Stubbe's \( \nu_{o-o} \) makes \( D_a \) 25% smaller to begin with. The form of \( D_a \) used for the Banks case was \( D_a = k(T_e+T_i)^2 T_i/\nu_{in} \mu_{in} (T_i+T_n) \). This is nearly the correct form since \( T_i \approx T_n \) near 300 km, which is where the diffusion velocity can be most accurately estimated from experimental measurements. The \( D_a \) in the Stubbe case was later changed to that in the Banks case in order that a valid comparison between the cases could be made.

2.3.2 Comparison with Previous Work

Salah and Holt (1970) studied two equinox days
where $V_{D\parallel}$ was calculated. They used Dalgarno's (1964) value for the collision frequency, which is similar to Banks (1966). They assumed only $O^+ - O$ collisions at the nominal height of 300 km, and found $n(0)$ through the Jacchia (1971) model. Figure 2.2 shows $V_{D\parallel}$ at the nominal height of 300 km for March 23-24, 1970 calculated by Salah and Holt (1974) and calculated in this paper for the Banks and (incorrect) Stubbe cases. The diffusion velocity is downward along the magnetic field line, being about 55 m/sec at night, 15 m/sec during the day, and about 3 m/sec after dawn in the Salah and Holt (1974) calculations. It is about 33 m/sec at night, 10 m/sec during the day, and about 5 m/sec after dawn for the Banks case of the present study.

The inclusion of $N_2$ and $O_2$ collisions with $O^+$ increases $V_{in}$ by about 12% at 300 km. (This increase would be about 7% if the $O^+ - N_2$ collision were repulsive instead of attractive). The use of OGO-6 model densities increases $n(0)$ by about 17%. The net result of these two differences would be to decrease $V_{D\parallel}$ by about 25% in the Banks case compared to the Salah and Holt (1974) calculations. A look at Figure 2.2 reveals that the decrease is about 40% at night, about 25% during the day, and becomes an increase after dawn of about 30% when the diffusion velocity is the smallest. The
balance of these differences is attributable to the use of INSCON smoothed values in the present study.

The largest terms in the summation part of equation 2.3 at 300 km are the electron density gradient term and the scale height term. The \( (T_e + T_i) \) gradient term is of secondary importance, while the \( (T_i + T_n) \) gradient term is the least significant. The electron density gradient term is usually smaller using INSCON values except during the midday peak of electron density. This term was negative just after dawn so the effect would be to increase the summation part of equation 2.3 between dawn and about mid-afternoon. The effects are largest when the summation term is smallest, which is just after dawn. At night the effect of using INSCON values is to decrease the summation term by about 20%. This coupled with the 25% decrease due to the \( n(0) \) model source and the inclusion of \( \text{N}_2 \) and \( \text{O}_2 \), gives the observed 40% decrease in \( V_{DH} \) at night between the Salah and Holt (1974) calculation and the Banks case. INSCON values also usually find \( (T_e + T_i) \) gradients smaller, but this is of secondary importance.

The electron density on March 23-24, 1970 was larger than usual. Since the diffusion velocity is usually downward during the day, this means that \( V_{DH} \) for this day was smaller than usual except after dawn when \( V_{DH} \) was negative.

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Correcting the form of Da in the Stubbe case would only decrease the diffusion velocity by 1 to 3 m/sec throughout the day. The decrease would be about 1.3 m/sec at night (5% of 25 m/sec), about 1.7 m/sec after dawn (35% of 5 m/sec), and about 2.5 m/sec during the day (25% of 10 m/sec).

2.4 Neutral Wind Component

The ion drift velocity $\mathbf{v}_i$ can be divided into three parts (Salah and Holt, 1974)

$$\mathbf{v}_i = \mathbf{v}_{D/} + \mathbf{v}_{n/} + \mathbf{v}_l$$

(2.5)

One part is due to diffusion $\mathbf{v}_{D/}$, another to neutral winds $\mathbf{v}_{n/}$, and the third to electric fields $\mathbf{v}_l$.

The components due to diffusion and neutral winds are parallel to the magnetic field line since ion gyrofrequencies are much larger than ion-neutral collision frequencies above about 180 km. The ion drift due to electric fields is $\mathbf{E} \times \mathbf{B}/B^2$ and is perpendicular to both electric and magnetic fields. Here $\mathbf{E}$ = electric field strength and $\mathbf{B}$ = magnetic field strength.

The neutral wind component $\mathbf{v}_{n/}$ defined positive upwards along the field line can be written as (Salah and Holt, 1974)

$$\mathbf{v}_{n/} = - \mathbf{v}_n \cdot \hat{\mathbf{B}}/B = -u \sin D \cos I - v \cos D \cos I + w \sin I$$

$$\equiv - (u \sin D + v \cos D) \cos I = -S_0 \cos I$$

(2.6)

where $D$ = magnetic declination ($-14^0$ for Millstone Hill), $u$ = east-west wind velocity defined positive
eastward, \( v \) = north-south wind velocity defined positive northward, \( w \) = vertical velocity defined positive upward, and \( S_D = u \sin D + v \cos D \) (\( \propto v \) for Millstone Hill) = the horizontal part of the neutral wind component parallel to the magnetic field line. This quantity \( S_D \) can be written in terms of the measured vertical ion drift \( V_{\|} \), the diffusion velocity, and electric fields

\[
S_D = \frac{-V_{\|}}{\sin I \cos I} + \frac{V_{\perp}}{\cos I} + \frac{1}{\sin I} \left( \frac{E_x}{\sin D} \cos D - \frac{E_y}{\sin D} \sin D \right) \quad (2.7)
\]

The measured vertical plasma drifts at the nominal height of 300 km for March 23-24, 1970 are shown in Figure 2.3. The Salah and Holt (1974) results as well as the present INSCON smoothed results are shown. The drift is downward at all times and varies around a value of about 17 m/sec. The errors in the experimental measurements at this altitude are estimated to be ±5 to 10 m/sec. The time and height smoothing in the INSCON values reduces this statistical uncertainty to ±2 to 4 m/sec. Neglecting any systematic errors in \( V_{\|} \), the experimental error in \( S_D \) comes mainly from the error in \( V_{\|} \). Because of the factor \( \cos I \sin I \), this is estimated to be about ±20 to 25 m/sec at 300 km. This is reduced by the INSCON program to about ±12 m/sec.

Figure 2.4 shows the deduced horizontal neutral
wind component \( S_D \) for March 23-24, 1970 at the nominal height of 300 km. The previous analysis by Salah and Holt (1974) and the Banks and (incorrect) Stubbe cases of the present analysis are shown ignoring electric fields in equation 2.7. The (incorrect) Stubbe case is also shown with an electric field model discussed in the next section. Correcting the form of \( D_a \) for the Stubbe cases would only result in shifting the neutral velocity northward by about 5 m/sec, since the average decrease in \( V_{D//} \) using the correct form is about 1.5 m/sec \((1.5\text{ m-sec}^{-1}/\cos 72^\circ \approx 5\text{ m/sec})\). The shift would have been larger if the electron densities had been smaller this day.

The smoothing in \( V_{iz} \) smooths out \( S_D \), and the reduction in \( V_{D//} \) reduces \( S_D \) as well. The largest reduction between the Banks case and the Salah and Holt (1974) analysis, is about 60 m/sec near midnight and about 40 m/sec on the average at night. During the day when the major contribution to \( S_D \) is from \( V_{iz} \), the correspondence is much closer. For the Banks case, magnitudes are of the order of 60 m/sec southward at night, and between about 25 and 50 m/sec northward during the day. Electric fields can change these magnitudes and an exact calculation of the horizontal aligned neutral wind component must include any imposed fields that are present.
2.5 Electric Fields

Electric fields can be deduced from plasma drifts measured in two directions perpendicular to the magnetic field. Such measurements were made at Millstone Hill by Evans (1972) and by Kirchhoff and Carpenter (1975) using a steerable radar. Similar measurements carried out in the U. K. have been reported by Taylor (1974).

In general, nighttime measurements were very difficult to make because of the low signal-to-noise ratio. In addition, large fluctuations in the drifts were not uncommon. Consequently, nighttime measurements are very few and are in considerable doubt. The measurements made at Millstone Hill for 14 relatively quiet days were combined by Kirchhoff and Carpenter (1975) into a model. The electric-field-induced plasma drifts predicted by this model are shown in Figure 2.5, which was taken from their paper. During the day, the electric fields are thought to be of dynamo origin and produce drifts of 10 to 20 m/sec toward the north in the morning and toward the south in the afternoon. This pattern seems reliably established. At night, the fields are probably of magnetospheric origin and are larger, causing northward drifts of up to 50 or 60 m/sec. These values are uncertain and may not be representative of quiet days. This electric field model
was used in some parts of the present analysis. In Figure 2.4, its effect is to decrease $S_D$ by 20 to 50 m/sec at night. During the day, the effect is small.
Figure 2.1 The exospheric temperature on March 23-24, 1970 at Millstone Hill.
Figure 2.2 The diffusion velocity calculated at the nominal height of 300 km for Millstone Hill on March 23-24, 1970.
Figure 2.3 The vertical ion drift at the nominal height of 300 km measured at Millstone Hill on March 23-24, 1970.
Figure 2.4 The horizontal component of the neutral wind in the magnetic meridian calculated at the nominal height of 300 km for Millstone Hill on March 23-24, 1970.
Figure 2.5 Ion drifts induced throughout the day by the Kirchhoff and Carpenter (1975) electric field model for Millstone Hill. The figure is reproduced from their paper.
3. DYNAMIC MODEL

3.1 Equations of Motion

The horizontal equation of motion for neutral particles in the thermosphere can be written as

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + f v + \frac{\mu}{\rho} \frac{\partial^2 u}{\partial y^2} + \lambda(u_{ion} - u)
\]

\[
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} = -\frac{1}{\rho} \frac{\partial p}{\partial y} + f u + \frac{\mu}{\rho} \frac{\partial^2 v}{\partial x^2} + \lambda(v_{ion} - v)
\]

(3.1)

where \(x, y,\) and \(z\) are directed east, north, and up; \(u, v, u_{ion},\) and \(v_{ion}\) are respectively the eastward and northward neutral and ion velocities; \(\lambda = n_i \mu_{in} v_{in} / \rho\) is the ion drag parameter; \(n_i, \mu_{in},\) and \(v_{in}\) are the ion number density, the ion-neutral reduced particle mass, and the ion-neutral collision frequencies taken from Stubbe (1968) or Banks (1966); \(\rho\) and \(p\) are the neutral density and pressure; \(f = 2\Omega \sin \phi\) the Coriolis parameter \((\Omega = \text{Earth's rotation rate, } \phi = \text{latitude});\) and \(\mu\) is the coefficient of molecular viscosity for atomic oxygen (Dalgarno and Smith, 1962).

The only non-linear term that has been included in 3.1 is the east-west gradient term \(u \partial(u, v) / \partial x\), which was assumed to be adequately represented by the time gradient \((24 h_0 / 2\pi) u \partial(u, v) / \partial t\). This interchange of time and longitude is possible because of the strong dominance of solar radiation in thermospheric diurnal variations. The north-south gradient term cannot be
included in a model using data from one station only. The east-west non-linear term is largest at dawn when changes are most rapid. The effect of adding the north-south non-linear term as well, has been studied by Rüster and Dudeney (1972). This term adds in the same sense as the east-west non-linear term, and its effects are likewise largest at dawn. The effect on the winds of adding both terms at mid-latitudes is 10% or less (Blum and Harris, 1973).

The horizontal ion velocity components can be written as (Salah and Holt, 1974)

\[
\begin{align*}
    u_{\text{ion}} &= (u \sin D + v \cos D) \sin D \cos I - V_{D,11} \cos D \cos I + U_{\text{ielect}} \\
    v_{\text{ion}} &= (u \sin D + v \cos D) \cos D \cos I - V_{D,11} \cos D \cos I + V_{\text{ielect}}
\end{align*}
\]

where \( u_{\text{ielect}} \) and \( v_{\text{ielect}} \) are the components of \( \mathbf{E} \times \mathbf{B} / B^2 \) in the eastward and northward directions.

The present model operates between 120 and 600 km with boundary conditions of zero velocity at 120 km and constant velocity (or zero height gradient) at 600 km. The Crank-Nicholson method was used to solve the equations with a time step of 10 minutes and a height step of 10 km.

3.2 Ion Composition

Between 270 and 600 km, the only ion is assumed to be \( \text{O}^+ \), so its number density is the same as the
measured electron density. Between 120 and 260 km, the ions are $O_2^+$, NO$^+$, and $O^+$. Day and night seasonal profiles of their relative abundances were computed by averaging profiles in the literature (e.g. Holmes et al. (1965), Cox and Evans (1970)) for the same approximate latitude, season, and time of day. A transition was then made over a period of one hour and 40 minutes around sunrise and sunset. Figure 3.1 shows the percent of $O^+$ used in the present model. For simplicity, the ratio of NO$^+/(NO^+ + O_2^+)$ was assumed to be constant. This constant was chosen to be valid above about 200 km. Since we are mainly concerned with results above 250 km, this inclusion of NO$^+$ and $O_2^+$ was probably unnecessary.

3.3 Neutral Temperature and Number Density

Assuming hydrostatic equilibrium, the temperature and number density in the dynamic model were assumed to be of a Bates (1959) - Walker (1965) form

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

(3.3)

\[ n_j(z) = n_{j120} \left( \frac{T_{120}}{T(z)} \right)^{1+\alpha_j + \gamma_j} \exp \left[ -\sigma z \right] \]

(3.4)

where $T$ is the neutral temperature, $T_{120}$ the temperature at 120 km ($355^0$K), $s$ the shape factor (0.020 km$^{-1}$), $\sigma = s + 1/(R_E+120)$, $R_E$ the Earth radius, $\xi = (z-120)^*$
the geopotential height, \( n_j \) the number density in cm\(^{-3}\) of species \( j \), \( n_{j120} \) the number density at 120 km (assumed constant), \( \beta_j = \frac{M_j g_{120}}{\sigma RT_\infty} \), \( M_j \) the molecular weight of species \( j \), \( R \) = universal gas constant, and \( \alpha_j \) the thermal diffusion factor. The thermal diffusion of \( n(O) \), \( n(N_2) \) and \( n(O_2) \) through a gas chiefly made up of \( n(O) \) were assumed to be zero.

Equations 3.3 and 3.4 assume the temperature and density are in phase. This is not actually true. Above about 200 km, the maximum in the density of the neutral particles (chiefly atomic oxygen), has been found to precede the maximum in temperature by about 0.7 hours (Reber et. al., 1973). Mayr and Volland (1973) suggest that this phase difference is a result of diffusion effects. For the sake of simplicity, any phase difference is ignored in the present model.

With the temperature and number density in the form of equation 3.3 and 3.4 and assuming fixed boundary conditions, their horizontal derivatives can be written as

\[
\frac{dT_{\text{horiz}}}{x} = \left(1 - e^{\phi_- (\xi)}\right) \frac{dT_{\text{horiz}}}{y} \\
\frac{dn_{j\text{horiz}}}{x} = n_j \left[ \left( \frac{\ln \frac{T_{120}}{T}}{T} - \alpha_j \right) \frac{dT_{\text{horiz}}}{y} \\
- \left( \frac{1 + \beta_j + \alpha_j}{T} \right) \frac{dT_{\text{horiz}}}{y} \right] - 34
\]
Through the gas law \( p = nkT \), the horizontal pressure derivatives are also directly proportional to the horizontal exospheric temperature derivatives.
Figure 3.1 Percent of atomic oxygen ion concentrations used in the present study and the ratios of NO⁺/(NO⁺+O₂⁺).
4. DATA ANALYSIS

4.1 Introduction

The temperature of the upper atmosphere, and therefore other parameters such as the pressure and winds, is a function of latitude, longitude, and altitude. An incoherent scatter radar at one station can determine the longitude (time) and altitude variations of temperature in the thermosphere, but cannot directly determine latitude variations. However, a single station can estimate the neutral wind component along the magnetic field line and thereby, through the linkage of the equations of motion, provide an indirect determination of the latitude variation near the station. It is this indirect determination of north-south variations which is accomplished by the fitting procedure to be described. Essentially, the east-west pressure gradient is obtained from the observed east-west exospheric temperature variations, and the north-south variation is adjusted such that the computed value of $S_D ( = u \sin D + v \cos D)$, the horizontal part of the neutral wind component along the magnetic field line, fits the observations.

4.2 Exospheric Temperature Structure

It is assumed that the latitudinal and longitudinal variations in the exospheric temperature can be
represented by a truncated Fourier-Taylor series of the form (Roble et al., 1974)

\[
T_{\infty} (\varphi, t) = \sum_{n=0}^{n_{\text{harm}}} \left[ (a_n + \partial a_n / \partial \varphi (\varphi - \varphi_s)) \cos \frac{n \pi t}{12} 
+ (b_n + \partial b_n / \partial \varphi (\varphi - \varphi_s)) \sin \frac{n \pi t}{12} \right] 
= \sum_{n=0}^{n_{\text{harm}}} \left( a_n \cos \frac{n \pi t}{12} + b_n \sin \frac{n \pi t}{12} \right) 
+ (\varphi - \varphi_s) \sum_{n=1}^{n_{\text{harm}}+1} A_n f_n (t) \tag{4.1}
\]

where \( n_{\text{harm}} \) is the number of harmonics searched, \( \varphi_s \) is the latitude of the observing station, \( t \) is the time in hours, \( A_1 = \partial a_0 / \partial \varphi, A_2 = \partial a_1 / \partial \varphi, A_3 = \partial b_1 / \partial \varphi, \ldots \) etc., and \( f_1 = 1, f_2 = \cos \frac{n \pi t}{12}, \ldots \) etc. The zero harmonic refers to the average temperature or meridional temperature gradient of the day under study, and the first, second, third etc. harmonics refer to the diurnal, semi-diurnal, terdiurnal, etc. terms. The east-west coefficients, \( a_n \) and \( b_n \), are found in terms of amplitudes and phases through a harmonic analysis of the observed exospheric temperature.

The form of 4.1 assumes that the temperature is the same at the beginning and end of a 24 hour period. This assumption sometimes necessitates the removal of data points near the beginning or end of a 24 hour observation period to insure that the temperature at the start of the period is approximately the same as the temperature at the end. Similarly, other
experimental data must be blended near the endpoints to satisfy this criterion of continuity from day to day.

4.3 Fitting Procedure

Because the horizontal pressure gradients are proportional to the horizontal exospheric temperature gradients, the velocities \( u \) and \( v \) and thus the neutral wind component along the magnetic field line, are also functions of these gradients. Provided that any non-linear or electric field term is small, a least squares fit to \( S_D \) can be performed to find the north-south temperature coefficients. (This is because \( S_D \) is largely controlled by the meridional pressure gradient).

The quantity which is minimized in the fit is

\[
S = \sum_{i=1}^{n_{data}} \frac{1}{(\sigma_{iso})^2} \left[ S_{Di} (\sum_{j} (A_j + \Delta A_j)) - S_{Di} (\text{data}) \right]^2
\]  

(4.2)

Minimization requires that \( \partial S/\partial A_k = 0 \) or that

\[
\sum_{i=1}^{n_{data}} \frac{1}{(\sigma_{iso})^2} \left[ S_{Di} (\text{data}) - S_{Di} (\sum_{j} (A_j)) \right] \frac{\partial S_{Di} (\sum_{j} (A_j))}{\partial A_k} = \sum_{i=1}^{n_{data}} \sigma_{iso}^2 \sum_{j=1}^{n_{data}} \frac{\partial S_{Di} (\sum_{j} (A_j))}{\partial A_j} \frac{\partial S_{Di} (\sum_{j} (A_j))}{\partial A_k} \Delta A_j 
\]

(4.3)

where the approximation

\[
f (\sum_{j} (A_j + \Delta A_j)) = f (\sum_{j} (A_j)) + \sum_{j} \Delta A_j \frac{\partial f (\sum_{j} (A_j))}{\partial A_j}
\]

(4.4)

has been used (Bevington, 1969). Here \( n_{data} \) is the number of data points to be fitted, \( \sigma_{iso} \) is the
estimated uncertainty in the data points $S_D(data)$, and the $A_j$ are estimates of the unknown meridional coefficients. $S_D(\hat{\xi} A_j)$ is the $S_D$ computed with these estimates and
\[ \frac{\partial S_D(\xi A_j)}{\partial A_k} = \frac{(S_D(\xi A_{j'} + \xi A_k) - S_D(\xi A_j))}{\Delta A_k} \]
is a measure of the effect on $S_D$ of changing the $k$th coefficient by an amount $\Delta A_k$. The $\Delta A_j$ are to be found by the fit. An average uncertainty was used for $S_D$, and as discussed in section 2.4, is between 10 and 15 m/sec depending on the smoothness of the data.

The procedure is to make a first estimate of the $A_j$, and then vary each $A_j$ separately to determine the effect of this change on $S_D$. A fit is then made to determine the $\Delta A_j$ which are to be added to the initial $A_j$ to provide a new estimate of the $A_j$. This procedure is repeated and usually converges to a final set of $A_j$. Normally, it takes between 2 and 7 iterations to converge to a set of coefficients that do not vary by more than 2 or $3^\circ K/rad$ from their previous values. A decrease in the standard deviation between $S_D(fit)$ (or $S_D(\hat{\xi} (A_j + \Delta A_j))$ and $S_D(data)$ is not always achieved at every iteration. Cyclical oscillations are not uncommon, even though the procedure usually results in convergence to a final set of $A_j$. The standard deviation between $S_D(fit)$ and $S_D(data)$ for the final set of $A_j$ is among the lowest achieved in the entire process. Other difficulties can arise if the non-linear terms become
too large, creating instabilities in the numerical scheme that solves the equations of motion 3.1.

4.4 Alternate Procedures

The $S_D$ found by the least squares fitting routine is not the same as the $S_D$ from the equations of motion 3.1, even though the temperature coefficients are the same. This is because the velocities are not strictly linear in their relation to the zonal and meridional exospheric temperature coefficients. Because of this, it may be desirable to improve the fit by finding the difference between the observed data and the $S_D$ computed by solving 3.1. The fit should then be redone, fitting not to the observed data, but to the observed data plus this difference. This technique will be referred to hereafter as the second fit procedure.

Another procedure does not require that the east-west coefficients be fixed beforehand. Instead, except for the zero harmonic, they are calculated simultaneously with the north-south coefficients, by fitting $S_D$ and the longitudinal temperature derivative $\partial T_\lambda / \partial \lambda$. Here the equation corresponding to 4.1 would be

$$T_\infty (\nu, t) = a_0 + \sum_{j=1}^{ncoll} A_j f_j (\nu, t)$$

(4.5)

where $A_0 = \partial a_0 / \partial \nu$, $f_1 = (\nu - \nu_0)$, $A_2 = a_1$, $f_2 = \cos \frac{\nu \pi}{\nu_0}$.
... etc.. The equation corresponding to 4.3 would be

\[
\sum_{i=1}^{n_{data}} \left\{ \frac{1}{(\sigma_{o_i})^2} \left[ S_{o_i}(\text{data}) - S_{o_i}(\Sigma A_j) \right] \frac{\partial S_{o_i}(\Sigma A_j)}{\partial A_k} + \frac{1}{(\sigma_{\lambda_i})^2} \left[ \frac{\partial}{\partial A_i} \Sigma A_j \right] \frac{\partial (\partial A_i / \partial \lambda_i)(\Sigma A_j)}{\partial A_k} \right\} \right. \\
\left. + \frac{1}{\sigma_A \Sigma A_j} \left( \frac{\partial}{\partial A_i} \Sigma A_j \right) \frac{\partial (\partial A_i / \partial \lambda_i)(\Sigma A_j)}{\partial A_k} \right\} \Delta A_i
\]

(4.6)

where the uncertainties \( \sigma_{\partial A_i / \partial \lambda} \) are assumed to vary with the harmonic being searched rather than with the data points.
5. RESULTS AND DISCUSSION

5.1 Temperature Gradients

5.1.1 Results from Present Study

Two equinox days, March 23-24 and October 5-6, 1970 were analyzed. The daily equivalent planetary amplitude of magnetic activity, Ap, was 2.6 (Kp between 0+ and 1-) for the March day and 5.7 (Kp between 1+ and 2-) for the October day. Tables 5.1 and 5.2 show the east-west and north-south exospheric temperature coefficients in terms of amplitudes and phases for the two days. Table 5.1 contains the harmonic analysis of the temperatures for the two days (the data) and also shows the effect of a simultaneous fit to $S_D (= u \sin D + v \cos D)$ and to the east-west exospheric temperature gradient $\partial T_e / \partial \lambda$ (case b1). Both tables contain the coefficients found from the Jacchia (1971) model (case f), and from the OGO-6 model (case g). The results of a previous analysis of March 23-24, 1970 by Roble et al. (1974) are included in the tables as case h.

Table 5.2 shows the effects of (i) performing a simultaneous fit to $S_D$ and $\partial T_e / \partial \lambda$ (case b1), (ii) finding more than 3 harmonics (case c1), (iii) doing a second fit for greater accuracy as described in section 4.4 (case c2), (iv) using the correct form of the ambipolar diffusion coefficient, $D_a$, in the Stubbe case.
(case c₂), (v) using Banks' (1966) values for the collision frequencies in place of Stubbe's (1968) (case c₄), (vi) eliminating the east-west non-linear term in the equations of motion 3.1 (cases d₁ and d₂), and (vii) including the electric field model of Kirchhoff and Carpenter (1975) (case e₂). The standard deviations between the fit and the data for these cases are included as well. The effect of changing the ionic composition model (summer, winter, or only n(O⁺)) was found to be negligible, and so was not included. Not all cases were computed for both days.

The present study began by using Stubbe's (1968) values for the collision frequencies. After most of the analysis was complete, it was decided that it would have been better to have used Banks' (1966) values because his results are based more squarely on the available experimental data (Mason, 1970). This led to the inclusion of case c₄ in Table 5.2. About the same time, it was discovered that the form of the ambipolar diffusion coefficient, Da, in the Stubbe cases was wrong. This necessitated the inclusion of case c₃ in Table 5.2. However, the previous analyses are still valid in comparing the effects brought about by electric fields etc.. The differences between Banks (1966) and Stubbe (1968) is about 25%, which is about the size of the error bar on the laboratory measurements of the
collision crosssections. Therefore, a comparison between the Banks and Stubbe cases can be looked upon as a comparison of results within the limits of the error bar on the collision frequencies.

The harmonic analyses of the exospheric temperature for both days given in Table 5.1, show a zero harmonic (i.e. average temperature) of about 1000°K. The diurnal term (first harmonic) is about 100°K with a maximum around 1400 LT. The higher harmonics have amplitudes of up to about 25°K, the largest one being either the second or third harmonic. The temperature variations on the two days under study are shown in Figures 5.1a and b.

The most accurate computation of the north-south temperature coefficients are probably those in case c_4. This is the case where Banks' (1966) values were used for the collision frequencies. No electric field was included, partly because the electric field for those days was not known, and also because it is suspected that the drifts on very quiet days, such as the March day, are less than the drifts in the Kirchhoff and Carpenter (1975) model. The east-west non-linear term was included because results with the one term that can be determined should be better than those with neither term included (Rüster and Dudeney, 1972).

For this case, as well as for the others, the
zero and first four harmonics of the north-south temperature coefficients were found to be the most important. For October 5-6, 1970, sunrise at 300 km was around 0440 LT and sunset around 1900 LT. For case $c_4$, the zero harmonic is about $15^\circ$K/rad, signifying a small mean temperature increase toward the pole. This term is negative for the other cases. The positive value in case $c_4$ is consistent with the fact that the velocities in the Banks case exhibit a larger southward component. The velocities to which fits are made are shown in Figures 5.2a and b. The amplitude of the first harmonic for case $c_4$ is about $60^\circ$K/rad and peaks around 0415 LT. The second harmonic is small. The third harmonic is the largest term, about $85^\circ$K/rad and peaks around 0545 LT. The fourth harmonic is about $20^\circ$K/rad and peaks around 0600 LT.

For March 23-24, 1970, sunrise at 300 km was around 0430 LT and sunset around 1940 LT. The zero harmonic in case $c_4$ is about $-55^\circ$K/rad, and is opposite in sign from the October day. The first harmonic is around $30^\circ$K/rad and peaks around 0630 LT. The second harmonic is smaller, around $25^\circ$K/rad and peaks around 0930 LT. The third and fourth harmonics are both around $40^\circ$K/rad and peak around 0630 and 0600 LT respectively. The relative magnitudes of the terms are approximately the same for the other cases as well.
The diurnal variations in the north-south exospheric temperature gradient for both days are shown in Figures 5.3a and b. The exospheric temperatures for both days in the latitude region of the Millstone Hill facility (42.6°N) as determined from these coefficients are shown in Figures 5.4a and b.

5.1.2 Comparison with Different Cases

A summary of this section is provided in Chapter 6 (section 6.1) for those who are not interested in a detailed comparison.

Simultaneously fitting $S_D$ and $\partial T_e/\partial \lambda$ takes approximately twice as long as fitting $S_D$ alone. It was desired to see if the improvement in the fit was worth the extra computational time. The changes in the east-west temperature coefficients (case $b_1$ and data) were up to 4K in amplitude and about 10 minutes in phase, which is not very significant. For the north-south coefficients, the decrease in the standard deviation of $(S_D(\text{data})-S_D(\text{fit}))$ between case $a_1$ and $b_1$ is about 15%. None of these changes are really large enough to justify the need for a simultaneous fit of the east-west and north-south coefficients.

On the other hand, adding more than 3 harmonics for the north-south variation results in important effects. This is demonstrated by comparing case $a_1$ and case $c_1$. For October, adding a fourth harmonic
reduces the standard deviation of \((S_D(\text{data})-S_D(\text{fit}))\) by 25%. In March, the addition of the fourth, fifth, and sixth harmonics yielded a reduction of 75%. Most of this was probably due to the addition of the fourth harmonic which was as large in amplitude as the third harmonic, and larger than the first and second harmonics. On both the March and October days, the lower harmonics \((0, 1, 2, \text{ and } 3)\), were not changed significantly.

The effect of carrying out a second fit to account for the non-linearity of the problem appears to be important only when the non-linear terms or the electric field terms become large. The range of velocities fitted was at least twice as large for the October day as compared with the March day. The sunrise and sunset periods were also times of more rapid change on October. As a result, one might expect the east-west non-linear term to be more important on the October day than on the March day. This is indirectly confirmed by noting that the effect of performing a second fit on October (case \(c_2\) versus \(c_1\)), reduced the standard deviation of \((S_D(\text{data})-S_D(\text{fit}))\) by 25%. The reduction for March was only 10%. Since the electric field model induces drifts that are large at night, it was decided that it would be best to do a second fit when these were included as well. The reduction in standard deviation for the linear case \((d_2\ \text{versus } d_1)\)
was only 14% in October.

Comparing cases $c_2$ and $c_3$ in October, one can see that the major effect of using the correct form for Da (discussed in section 2.3.1), is to increase the zero harmonic by about $10^0 K/\text{rad}$, making the temperature decrease toward the pole more pronounced. This increase would have been smaller for March because of the smaller diffusion velocities present that day.

Cases $c_3$ (or $c_2$ remembering the change in the zero harmonic) and $c_4$ show the effects of using Stubbe's (1968) or Banks' (1966) values for the collision frequencies. These come into the calculations of the $S_D(\text{data})$ (through $V_{D//}$) and into the ion drag term in the equations of motion 3.1. The largest effect is on the zero harmonic, changing it by about $30^0 K/\text{rad}$ and decreasing the temperature gradient toward the pole. As mentioned before, this term actually changed sign in the Banks case ($c_4$) for October. In March, the change in the magnitude of the first harmonic was not significant, although the change in phase was about an hour. For October, the phase change was about half an hour, and the magnitude change was about $10^0 K/\text{rad}$ or about 17%. The changes in the higher harmonics was not significant.

The only non-linear term that could be included in the equations of motion 3.1 was the east-west term.
If it is assumed that the north-south and east-west non-linear terms have approximately the same magnitude, an estimate of the effect of including the north-south non-linear term can be obtained by dropping the east-west term in the equations of motion (case d). Case d$_2$ can be compared to case c$_2$ (or case d$_1$ to c$_1$). The effect on the zero harmonic is small, being between 2 and 7.6$^\circ$K/rad. This is only about twice the normal uncertainty. The effects on the first through fourth harmonics are similar to one another, changes in amplitudes varying between 2 and 20$^\circ$K/rad and changes in phase between 0 and 1 hours. The average change in amplitude is about 12.5$^\circ$K/rad, which can be between 15% and 75% of the total amplitude for different harmonics. Changes in phase are usually less than an hour, except for cases when the amplitude is small, as is true for the second harmonic on both days. The average percentage change in amplitude for the first, third, and fourth harmonics is 23%. The changes in the fifth and sixth harmonics in March are within the uncertainty of the technique. It is somewhat surprising that the changes on the March day are as large as they are. One might have expected them to be smaller because the percent decrease in the standard deviation of $(S_D(data)-S_D(fit))$ between case c$_1$ and c$_2$ was smaller for that day than for October.
The effect of electric-field-induced drifts can be important if the fields are large enough. The electric field model of Kirchhoff and Carpenter (1975) was included in case e₂. This model produced ion drifts at night of up to 60 m/sec. Such large drifts may not be present on very quiet days, and are probably exceeded on the more disturbed days. However, such drifts are representative of what might occur, and so can give a crude estimate of the importance of the electric field. Case e₂ can be compared to case c₂. The effect on the zero harmonic is small, being up to 5°K/rad, slightly intensifying the decrease in temperature toward the pole. The change in amplitude of the first four harmonics varies from 1 to 9°K/rad, the average being about 5.5°K/rad. Excluding the second harmonic, the average percent change in amplitude is 14%. Phase changes are between 0.1 and 1.3 hours, the phase change in the first harmonic being about three quarters of an hour.

5.1.3 Comparison with Theory and Other Studies and Models

The temperature coefficients derived from the Jacchia (1971) and OGO-6 (Hedin et. al. (1974)) models were included in Tables 5.1 and 5.2 for comparison. The east-west coefficients show a general similarity to the experimental values, except that the amplitudes for the second and higher harmonics are usually less in the
models. This is probably due to the greater averaging required to obtain the models. The phases for the higher harmonics are also different. In the Jacchia model, the zero harmonic (mean) is about $1000^\circ$K, and the amplitude of the first harmonic is about $100^\circ$K with a peak at about 1430 LT. For the OGO-6 model, the zero harmonic is about $1075^\circ$K, with a first harmonic of about $150^\circ$K which peaks around 1500 LT. Previous studies comparing Millstone temperature data with the OGO-6 model (Salah and Evans, 1973), have found that OGO-6 temperatures are about 7% higher on the average and show good agreement in the diurnal variations except for the early morning hours in winter.

The east-west coefficients are also similar to other experimental data taken at St. Santin. Alcayde (1974) reported on two long observation periods of 3 to 5 days. There, diurnal and semi-diurnal components were found with amplitudes of about 10% and 2.5% of the zero harmonic. The diurnal component peaked around 1500 LT and the semi-diurnal component peaked around 1400 LT in the winter and 0700 LT in the summer. No higher harmonics appeared clearly. These findings are also in good agreement with findings from Millstone Hill (Salah and Evans, 1973).

The four main heat sources for the thermosphere are the extreme ultra-violet (EUV) portion of the solar
spectrum, the dissipation of waves that propagate from the lower atmosphere, joule heating and particle precipitation. These last two are important mainly in the auroral zones at night (Volland and Mayr (1972), Banks and Siren (1974)). As a first approximation, one would expect the highest and lowest temperatures on the globe to be at the latitudes of the sub-solar and anti-solar points, respectively. At any location on the globe, the maximum temperatures would be expected to occur just prior to sunset and the minimum temperatures just prior to sunrise since the atmosphere should continue to be heated as long as the sun is up. In fact, maximum temperatures are usually found in the late afternoon and the minimum just before dawn. From this picture, one would expect the diurnal term of the meridional temperature gradient to be significant and to peak around sunrise. If heating were due to EUV radiation alone, the zero harmonic of the meridional temperature gradient would be negative, indicating a net temperature decrease toward the poles. However, the heat input at the auroral zones may change the sign of this term for mid-latitudes. The higher harmonics should adjust themselves to fit to the pattern set up by the sunrise and sunset times.

The changes in temperature are rapid just after sunset and especially just after sunrise, so one might
expect the meridional temperature gradients to be largest at these times. The sun rises earlier and sets later in higher latitudes at 300 km during equinox. Therefore, at Millstone Hill during equinox conditions when there is about 9 hours of darkness at 300 km, one might expect the third harmonic to be very important in fitting the peaks after sunrise and sunset.

Figures 5.3c and d are the north-south temperature gradients at the latitude of Millstone Hill (\( \varphi = 42.6^\circ \)N) for March 23-24, 1970 derived from the Jacchia and OGO-6 models. These models both show a small average increase in temperature toward the poles and have the general picture of a maximum temperature in late afternoon and a minimum before dawn. The OGO-6 model even has the suggestion of a peak after sunrise and sunset, although this is absent in the Jacchia model, probably because of the greater averaging involved.

The meridional gradients from the least squares fit shown in Figures 5.3a and b, both reveal peaks after sunrise and sunset, with the sharpest peak after sunrise. There also seems to be a third peak around 1200 or 1300 LT. The physical process to explain this peak is presently unknown. The coefficients in Table 5.2 found by the fit indicate that the third harmonic is important for both days and peaks around 0600 LT. The diurnal component is also important, peaking at
0500 LT. The March day shows a relatively large decrease in temperature toward the pole. In October, there is a small decrease toward the pole for the Stubbe cases, and a small increase toward the pole for the Banks case.

5.2 Winds

5.2.1 Average Velocities from Other Studies

Satellite drag studies of the superrotation of the upper atmosphere (King-Hele (1972), King-Hele and Walker (1973)), suggest an average zonal velocity at 30°N and S at 300 km to be between 20 and 100 m/sec eastward. Radar studies of east-west ion drifts near the magnetic equator (Woodman, 1972), suggest an average neutral zonal velocity of about 50 m/sec eastward. This is low latitude data. If the mechanism behind superrotation is polarization fields at night, then the superrotation is greatest at low latitudes. If the superrotation is due to electric fields associated with magnetic substorms, then the effect is greatest at high latitudes (Rishbeth, 1972).

A study of eddy mixing and circulation in the ionosphere (Johnson and Gottlieb, 1970), indicates that average meridional velocities should be directed toward the winter pole. In a radar study at St. Santin (44.6°N, 2.2°E) using Stubbe (1968) and covering the years
1971-1972 (Amayenc, 1974), the average neutral meridional velocity at 300 km was found to be 22 m/sec southward during the spring (March and April), 24 m/sec southward during the summer (May through August), 10 m/sec southward during the fall (September and October), and 10 m/sec northward during the winter (November through February).

5.2.2 Results from Present Study

Figures 5.5a and b show the fit in $S_D$, the horizontal component of the neutral wind in the magnetic meridian, for the Banks case $c_4$ on both days that were analyzed. Figures 5.6a and b and Figures 5.7a and b show the zonal velocity, $u$, and meridional velocity, $v$, where $u$ is positive eastward and $v$ is positive northward. Table 5.3 summarizes the averages and the ranges of the $u$ and $v$ velocities at 300 km obtained in a few of the different cases which were tested.

The zonal winds are generally eastward from mid-afternoon until after midnight, with a maximum about two hours after sunset. They are then westwards, with another peak about two hours after sunrise. The meridional winds are generally equatorwards at night and polewards during the day. The highest velocities are usually found around 200 km, above which there is a decrease imposed by ion drag where the ion density starts to become large.
For the Banks case (c₄) at 300 km on October 5-6, 1970, the average zonal velocity was about 11 m/sec westward and actual values ranged from 195 m/sec at 2130 LT to -340 m/sec at 0500 LT. The meridional velocity was about half the zonal velocity, with an average of -34 m/sec and a range from 45 m/sec at 1800 LT to -210 m/sec at 0020 LT. This average southward velocity is consistent with the average increase in temperature toward the pole for this case.

For the Banks case (c₄) on March 23-24, 1970, the average zonal velocity was 26 m/sec eastward and ranged from 195 m/sec at 2120 LT to -190 m/sec at 0620 LT. The average meridional velocity v, was 9 m/sec and ranged from 40 m/sec at 1620 and 0900 LT to -60 m/sec at 0020 LT. This average northward velocity would have been expected from the large average north-south temperature decrease toward the poles. The velocities for both days were generally larger at night when the influence of the ion drag was smaller. The winds in October were generally stronger than in March. This could have been expected by noting that the $S_D$ in Figures 5.2a and b was about 4 times larger (and generally more southward) for October than for March. The larger velocities on October are a result of the smaller electron densities for that day.

The average zonal velocity on the March day
falls within the lower bounds of the estimates of super-rotation at 30°N and S. The average on the October day for the Banks case indicates subrotation. Neither number is very significant by itself. A statistical study of many days should be made before any conclusions are drawn. It would also be useful if the results of such a study could be compared to other mid-latitude data.

The winter pole, as evidenced by the presence of a polar vortex in the upper stratosphere and mesosphere, usually changes its position between hemispheres around early September and early April. Therefore, chances are that the winter pole was in the Northern Hemisphere on the two days under investigation. The March day shows an average poleward wind, but the October day shows an equatorward wind. However, the magnetic activity indices were larger for the October day. Therefore, heating in the auroral zone may have played a part on the October day. Also, there may be a time lag between the circulation system in the upper stratosphere and mesosphere, and that in the thermosphere. As well, the equinoxes are times of transition, so one cannot really say for sure which way the winds ought to blow. Again, no conclusions should be drawn until a statistical study of many days throughout the year has been completed.
5.2.3 Comparison with Different Cases

This section, along with section 5.1.2, is summarized in Chapter 6 (section 6.1) for those who are not interested in the details of the comparisons.

Changing the collision frequencies from Banks' (1966) values to Stubbe's (1968), resulted in an $S_D$ which is more northward on the average. Also, since the ion drag term is larger, the range of velocities should be smaller. Both of these tendencies are manifest in the comparison between cases $c_4$ and $c_3$ of the October day. The average $v$ velocity is now $-14$ m/sec, which is still an average southward velocity. This would not have been predicted beforehand, since there is a mean temperature decrease toward the pole in the Stubbe case. However, this decrease is small, so that a southward average velocity could arise. The range in velocities is reduced by about 50 m/sec in $v$ and 80 m/sec in $u$.

The new ranges on the October day for $u$ are from 185 m/sec eastward at 2150 LT to 256 m/sec westward at 0530 LT. The new range for $v$ is from 50 m/sec northward at 1740 LT to 145 m/sec southward at 0020 LT. There is also a change in the average zonal velocity from $-11$ m/sec to 6 m/sec. The decrease in the range of velocities is about 20\% for the zonal velocities and 30\% for the meridional velocities.

The change between the correct and the incorrect
forms of the ambipolar diffusion coefficient (cases c\textsubscript{3} and c\textsubscript{2}) in the Stubbe case, is essentially a larger northward velocity in the correct formulation. For October, this increase is 5 m/sec. (The increase for March would probably be smaller because of the larger electron densities). There is also a change in the average zonal velocity from 5.6 m/sec to 1.5 m/sec. The velocity ranges are approximately the same. The comparison between the (correct) Banks and the (incorrect) Stubbe cases (c\textsubscript{4} and c\textsubscript{2}) in March, show the general picture of an increased average northward velocity and a smaller range of velocities. The average zonal velocity remained about the same at 26 m/sec eastward.

The east-west non-linear term acts to increase the westward velocity and decrease the eastward velocity. Since \( u \) is generally positive when \( \partial v / \partial x \) is negative, and vice versa, the east-west non-linear term increases the northward velocities and decreases the southward ones. All of these effects except for the increased northward velocity are apparent from a comparison of cases d and c\textsubscript{2} in Table 5.3. However, the non-linear term is very small at the time of the maximum northward velocity. The average meridional velocity is more northward for the non-linear case as predicted, but the average zonal velocity is more eastward even though the range of velocities has been
shifted to the west. The average percent change in the magnitudes of the velocities is about 10%. The study by Rüster and Dudeney (1972) suggests that the north-south non-linear term would act in the same sense as the east-west term.

The major effect of the electric field model (case e2) is to induce northward drifts of the ions at night. This drift is communicated to the neutrals through collisions and results in a more northward average meridional velocity. Comparing cases c2 and e2 in Table 5.3, the average northward increase is about 10 m/sec for both days. The electric field model also induces large westward drifts between about 2000 and 2400 LT. This will reduce the eastward maximum around 2200 LT and will make the average zonal velocity more westward. For October, the average is 2.5 m/sec more westward, whereas for March, the average is 7 m/sec more westward. The percent changes in the range of velocities is of the order of 15%.

5.3 Comparison with Previous Analysis of March 23–24, 1970

The March day was previously analyzed in a paper by Roble et al. (1974) using the temperature and $S_D$ presented in the Salah and Holt (1974) paper (see Figures 2.1 and 2.4). This analysis did not include non-linear terms. A relatively small semi-diurnal electric field was used (Salah and Holt, 1974), and velocities
due to this field alone were calculated separately. Once the effects due to the electric field were removed, it was assumed that the velocities were linearly related to the exospheric temperature gradients. This is not strictly true as can be seen from equation 3.6. Also, it takes 2 iterations before the coefficients stop changing by $10^0\text{K/rad}$ or more. The results of this previous study are included in Tables 5.1, 5.2, and 5.3 as case h.

The east-west temperature coefficients used in this previous study are in very good agreement with the mean term and the diurnal term of the present study. The semi-diurnal term is larger in the Roble et. al. (1974) study and peaks sooner. The terdiurnal term is smaller and also peaks sooner. Such differences can be explained on the basis that the present analysis used INSCON smoothed values while the other did not.

On the other hand, there is less agreement between the north-south temperature coefficients found in the Roble et. al. (1974) study (case h) and the present study. The zero harmonic for case h is $-4.6^0\text{K/rad}$, which is about $50^0\text{K/rad}$ smaller than case c4. One might have expected it to be positive, since the $S_D$ which is fitted (Salah and Holt, 1974) exhibits a large southward component. The diurnal term for case h is larger than in case c4, because the range in the $S_D$
being fitted is larger. The terdiurnal term is not as important as it is in the present analysis, and the semi-diurnal term is more important. The phases do not coincide with the present analysis either.

The average meridional velocity is 55 m/sec southward, which, as expected, is larger than the 9 m/sec northward of the present analysis (case c4). The average zonal velocity is about 50 m/sec more westward, although the range in zonal velocities appears to be about the same.
TABLE 5.1 East-West Exospheric Temperature Coefficients for March 23-24 and October 5-6, 1970

i) harmonic analysis of data  
b) simultaneous fitting of $S_D$ and $\partial T / \partial \lambda$  
f) Jacchia (1971) model  
g) OGO-6 model (Hedin et al., 1974)  
h) Roble et al. (1974)

March 23-24, 1970

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TABLE 5.1 continued

October 5-6, 1970

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TABLE 5.2 North-South Exospheric Temperature Coefficients for March 23-24 and October 5-6, 1970

a1) fit SD only, harmonics 0-3, non-linear, no E
b1) same as a1, only fit $\Delta T_{20}/\Delta \lambda$, too
c1) fit SD only (more harmonics), non-linear, no E
c2) same as c1, only second fit procedure, too
c3) same as c2, only correct form for Da
c4) same as c3, only Banks (1966)
d1) same as c1, only linear
d2) same as c2, only linear
e2) same as c2, only K-C (1975) E field
f) Jacchia (1971) model
g) OGO-6 model (Hedin et al., 1974)
h) Roble et al. (1974)

March 23-24, 1970

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s. d. in (SD) | * | * | * | * |

(s. d. in S_D - S_D(fit)) | 0.600 | 0.158 | 0.142 | 0.161 | 0.179
### March 23-24, 1970 continued

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### October 5-6, 1970

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October 5-6, 1970 continued

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s. d. in (S_D) (data)-S_D(fit) | 0.826 | 0.751 | 0.643 | 0.861 |

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TABLE 5.3  Velocity Averages and Ranges for March 23-24 and October 5-6, 1970

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March 23-24, 1970

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- 69 -
TABLE 5.3 continued

October 5-6, 1970

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Figure 5.1a The exospheric temperature at Millstone Hill on March 23-24, 1970 and its representation in 6 harmonics.
Figure 5.1b The exospheric temperature at Millstone Hill on October 5-6, 1970 and its representation in 4 harmonics.
Figure 5.2a The horizontal component of the neutral wind in the magnetic meridian at 300 km on March 23-24, 1970 at Millstone Hill which is to be fitted in cases $c_2$, $c_4$, and $e_2$ (see text).
Figure 5.2b The horizontal component of the neutral wind in the magnetic meridian at 300 km on October 5-6, 1970 at Millstone Hill which is to be fitted in cases $c_2$, $c_4$, and $e_2$ (see text).
Figure 5.3a The meridional exospheric temperature gradient at Millstone Hill for case $c_4$ (see text) on March 23-24, 1970. Positive values are an increase in temperature toward the pole.
Figure 5.3b The meridional exospheric temperature gradient at Millstone Hill for case c₄ (see text) on October 5-6, 1970. Positive values are an increase in temperature toward the pole.
Figure 5.3c The meridional exospheric temperature gradient at Millstone Hill from the Jacchia (1971) model on March 23-24, 1970. Positive values are an increase in temperature toward the pole.
Figure 5.3d The meridional exospheric temperature gradient at Millstone Hill from the OGO-6 model (Hedin et. al., 1974) on March 23-24, 1970. Positive values are an increase in temperature toward the pole.
Figure 5.4a Latitudinal and longitudinal (time) variations in the exospheric temperature within ±10° of Millstone Hill (42.6°N) for case c4 (see text) on March 23-24, 1970.
Figure 5.4b  Latitudinal and longitudinal (time) variations in the exospheric temperature within ±10° of Millstone Hill (42.6°N) for case c4 (see text) on October 5-6, 1970.
Figure 5.5a  Fit in $S_D$ ( = $u \sin D + v \cos D$) at 300 km for case $c_4$ (see text) at Millstone Hill on March 23-24, 1970.
Figure 5.5b Fit in $S_D = u \sin D + v \cos D$ at 300 km for case $c_4$ (see text) at Millstone Hill on October 5-6, 1970.
Figure 5.6a Zonal winds (positive is eastward) at Millstone Hill for case c₄ (see text) on March 23-24, 1970.
Figure 5.6b  Zonal winds (positive is eastward) at Millstone Hill for case $c_4$ (see text) on October 5-6, 1970.
Figure 5.7a Meridional winds (positive is northward) at Millstone Hill for case c_{4} (see text) on March 23-24, 1970.
Figure 5.7b Meridional winds (positive is northward) at Millstone Hill for case $c_4$ (see text) on October 5-6, 1970.
6. CONCLUSIONS

6.1 Summary of Results and Uncertainties

The previous chapters have described a method of deducing north-south temperature variations and a description of the neutral wind in the thermosphere near the Millstone Hill incoherent radar station where data on the ionized particles in the F layer were obtained. In general, the zero and first four harmonics of the north-south exospheric temperature gradient were most important, being on the order of 50°K/rad in amplitude. For the two equinox days studied, there were 3 peaks in this gradient. The peaks after sunrise and sunset can be explained on the basis of different sunrise and sunset times at different latitudes, but the peak after noon has not been explained.

The zonal winds are generally eastwards from mid-afternoon until after midnight, and then westwards with wind speeds of up to 200 to 300 m/sec. The meridional winds are generally polewards during the day with speeds up to 40 or 50 m/sec. During the night the winds are equatorwards with speeds up to 50 or 200 m/sec depending on the ion drag and thus on electron density concentrations.

There are several uncertainties involved in the method. In general, the method provides the most reliable results around 300 km. The quantities at other
heights are constrained by the model restrictions and may be quite different from what is calculated. This is especially true for heights less than 200 km where tidal winds, which were ignored, become important.

The basic uncertainty in the meridional temperature coefficients is about $3^\circ$K/rad in amplitude and about half an hour for the first harmonic. There are also uncertainties in the ion-neutral collision frequencies, in the neutral densities used in the ambipolar diffusion coefficient, in the electric field, in the absence of the north-south non-linear term, and in the statistical uncertainties in the temperature, diffusion velocity, and vertical ion drift.

A 25% increase in the collision frequency can decrease the magnitude of the winds by up to 30% and make the average meridional velocity 20 m/sec more northward. This average increase in the northward wind is accompanied by an average decrease of $30^\circ$K/rad in the zero harmonic of the meridional temperature gradient, implying a shift to decreasing the temperature toward the pole. The significant changes in the higher harmonics are limited to the first harmonic, where changes in amplitude of up to 15% and phase changes of up to an hour can occur.

The electric field is used in the experimental determination of $S_D ( = u \sin D + v \cos D)$ as well as
in the equations of motion 3.1. This provides a sort of feedback mechanism whereby average meridional velocities can change by 10 m/sec, but a corresponding change in the north-south zero harmonic is considerably damped. Changes in the zero harmonic are only about 40K/rad while changes in the amplitudes of the higher harmonics is of the order of 15%. The phase change in the first harmonic is up to an hour. The effect on the neutral winds can be up to 20%. If the electric field is known, then the known field should be used in the calculations. This particular model is only intended as an estimate of the possible effects a field might have on the winds and meridional temperature gradients.

Removing the east-west non-linear term from the equations of motion had the largest effect on the higher harmonics of the meridional temperature coefficients. The changes averaged 12.50K/rad in amplitude or 25% for the first four harmonics, but changes of up to 200K/rad or 35% were not uncommon. However, changes in the winds were only of the order of 10%. If it were possible to include the north-south non-linear term as well, it would probably act in the same sense as the east-west term (Rüster and Dudeney, 1972).

The use of a model neutral number density in the determination of the ambipolar diffusion coefficient
introduces a further uncertainty. If the OGO-6 values are consistently too high or too low, then we can expect a change in the average meridional velocity and in the zero harmonic of the north-south temperature coefficients. The effect on the higher harmonics is unknown, but is probably not more than the uncertainties introduced by the electric field case.

The statistical uncertainty in $S_D$ of about 12 m/sec (see section 2.4) provides another uncertainty in the winds. On the March day when the wind speeds were smaller than usual, such an uncertainty is about 25% of the meridional wind. On October, it is less than 10% of the meridional wind at night.

In summary, it appears that the uncertainty in the winds is about 30%, while the uncertainty in the first four harmonics of the meridional temperature coefficients is between 25 and 35%. The phase of the first harmonic could be off by an hour and the zero harmonic off by 30°K/rad. The average meridional velocity is also uncertain to 25 m/sec and the zonal velocity to about 15 m/sec. All this implies that the general technique is uncertain to about ±30%.

However, some of the uncertainty in the winds and in the zero harmonic is due to uncertainties in the collision frequency. This uncertainty would be the same from day to day, and is more in the nature of
a bias than a statistical uncertainty. Further, if the electric field is known, or can be reliably guessed at, then this removes all but about a 15% statistical uncertainty in the winds. This remaining uncertainty is mostly due to the statistical uncertainty in the experimental $S_D$, and to the absence of the north-south nonlinear term in the equations of motion 3.1. The statistical uncertainties in the first four harmonics of the temperature coefficients would still remain large at about 25% on average.

6.2 Suggestions for Future Work

Studies of seasonal variations in the winds or in the average north-south temperature gradient could be carried out, especially for days when electric fields are known. Although a bias may exist in the results, the percent changes throughout the year should still be valid. The results of such studies may confirm the presence of different circulation patterns in summer and winter as suggested in the study of Salah et. al. (1974). Such a seasonal study may also determine the migration of the pressure bulge to higher latitudes during solstice conditions as suggested by the Hedin et. al. (1974) model.

The technique can estimate the temperatures within about 5° latitude of the incoherent scatter facility through the north-south exospheric temperature
coefficients. If the technique were applied to data from other facilities as well, then a map of temperatures such as in Figures 5.4a and b could be constructed over a larger portion of the globe to determine whether or not they fit together in a reasonable manner. This could be a possible method of determining the correct collision frequencies. However, the other uncertainties involved, such as the statistical uncertainty in $S_D$, the use of INSCON smoothed data, or model atomic oxygen densities, may make it very difficult to get anything better than the present $\pm 25\%$ error bar on the $0^+ - 0$ collision frequency. It is also quite possible that a reasonable fit in the temperatures cannot be made within the limits of the uncertainty in the collision frequencies. In which case, we can be relatively certain that something else is amiss.

Data from other radar facilities could also be used to estimate the north-south non-linear term. The facility at Malvern is probably best situated for this kind of cooperative study with Millstone Hill since the St. Santin facility may be located too close in latitude.
ACKNOWLEDGEMENTS

I am very grateful to Dr. J. V. Evans for the use of the facilities at the Millstone Hill incoherent scatter observatory, and for his assistance in the project. I would like to thank Drs. R. E. Newell and J. E. Salah for their comments and support, and Dr. J. M. Holt for his assistance in mathematical and computer programming matters. I am grateful to the Advanced Study Program at the National Center for Atmospheric Research for the opportunity to be there during the summer of 1973. This thesis topic began at that time. I wish to thank Dr. R. G. Roble for his advice and support during my stay there, and for his continued interest thereafter. I also wish to thank Dr. P. B. Hays for helpful discussions and Dr. L. A. Carpenter for his electric field data. Financial support during the project came from a National Science Foundation Fellowship.
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