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IONOSPHERIC RESEARCH

Scientific Report 431

NEUTRAL WINDS AND ELECTRIC FIELDS FROM MODEL STUDIES USING REDUCED IONOGRAMS

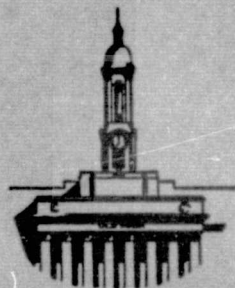
by

Daniel Edward Baran

October 10, 1974

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IONOSPHERE RESEARCH LABORATORY



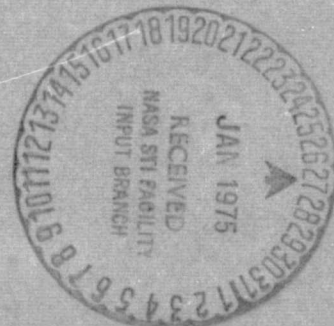
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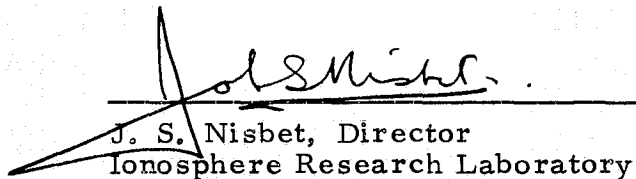
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LIST OF SYMBOLS

In this text the following symbols are used. The coordinate system is cartesian with x pointing to geographic south, y pointing to geographic east and z pointing upwards. The subscripts n, i and e are commonly used to refer to neutral particles, ions and electrons, respectively.

$\vec{B}$	geomagnetic induction vector
B_0	magnetic intensity (weber-m ⁻²)
C_i	charge number of the i-th ion species
d	diffusion rate
e	charge
$\vec{E}$	electron field vector
$\vec{E}_s$	electrostatic field vector
$\vec{E}_t$	total electric field vector (refers to the electric field generated by the combined action of the E region and magnetosphere dynamos)
$\vec{g}$	acceleration due to gravity
k	Boltzmann's constant
L_i	loss rate of the i-th ion species
m	mass
m_{in}	reduced mass for the i-th ion and n-th neutral species
n_i	number density of the i-th ion species
n_n	number density of the n-th neutral species
N_e	number density of electrons
NmF2	electron density at the maximum of the F2 layer
p	pressure = nkT

P_i	production rate of the i-th ion constituent
$P_i^{(p)}$	photoionization contribution of the photoionization rate
Q	production rate of electrons
$\overline{R}$	radius vector from the center of the earth
S_q	magnetic variations observed during quiet solar conditions
t	local time
T	temperature
T_e	electron temperature
T_i	ion temperature
$\overline{U}$	neutral wind velocity vector
U_M	equivalent wind in the direction of the magnetic meridian
$\overline{V}_i$	ion species velocity vector
z	altitude
z_m	reduced height
β	recombination rate coefficient
λ	wavelength of the solar radiation
μ	coefficient of viscosity
ν	collision frequency
ν_{in}	collision frequency of ions with neutrals
ρ	density
ρ_M	correlation coefficient of the equivalent winds in the direction of the magnetic meridian
$\sigma_n^{(a)}$	absorption cross section of the n-th neutral constituent
$\sigma_n^{(i)}$	ionization cross section of the n-th neutral constituent
ϕ	latitude
Φ_∞	unabsorbed photon flux

χ	solar zenith angle
ω	radian frequency
$\overline{\Omega}$	earth's angular rotation vector

ABSTRACT

A relationship between the vertical component of the ion velocity and electron density profiles derived from reduced ionograms is developed. Methods for determining the horizontal components of the neutral winds and electric fields by using this relationship and making use of the variations of the inclinations and declinations of the earth's magnetic field are presented. The effects that electric fields have on the neutral wind calculations are estimated to be small but not second order. Seasonal and latitudinal variations of the calculated neutral winds are given and it is shown that these are in general agreement with results from current theoretical models and from measurements. In particular, the midnight reversal of the meridional wind at Arecibo is observed from these calculations.

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CHAPTER I

INTRODUCTION

1.1 General Statement of the Problem

In the upper atmosphere temperature changes caused by the variation of the solar zenith angle with time and latitude give rise to pressure gradients which in turn cause neutral winds to flow. Dynamo action in the E region of the ionosphere and in the magnetosphere produces electric fields. The neutral wind system is important because it is necessary for the understanding of the energy budget of the thermosphere, the redistribution of neutral constituents and the dynamics of the F region of the ionosphere. The electric field system is important because electric fields propagate along the magnetic field lines and cause ionospheric motions which in turn affect the neutral winds.

Experimental and theoretical investigations concerning neutral winds and electric fields have been made by many authors. Experimentally, the incoherent backscatter technique has proven to be a very powerful tool for determining both neutral winds and electric fields. Unfortunately, the number of incoherent backscatter stations is severely limited by the high cost of constructing and maintaining such facilities and thus the data produced represents only local conditions. On the other hand, the complex coupling between the ionospheric and neutral atmospheric processes renders a complete theoretical solution to the problems impossible. Present theoretical neutral wind and electric field calculations are made by assuming a neutral atmospheric model and calculating the ionospheric model or

vice versa. In either case the assumed models are usually experimentally determined from limited or low resolution data sources.

An adequate resolution of the neutral wind system and electric field system requires measurements at all times of day, season and solar cycle and at all latitudes and longitudes. Ionogram data can produce these resolution requirements. It is the purpose of this study to determine the extent to which ionosonde data can be used to derive neutral winds and electric fields and to compare these quantities with currently accepted values.

1.2 Previous Related Studies

1.2.1 Methods for Calculating Vertical Drift Velocities from Electron Density Profiles

For simplified conditions with only diffusion and recombination considered Martyn (1956) showed that the shape of a Chapman layer is unaffected by diffusion and that this type of layer decays uniformly without change of shape and height. To test the stability of the Chapman layer a number of initial electron densities were assumed. It was shown that in all cases the initial distributions became redistributed to form a Chapman layer. Martyn showed that at the F2 peak

$$\beta = g \sin^2 I / 2H\nu \quad (1.1)$$

where

β is the recombination rate coefficient

g is the acceleration due to gravity

I is the inclination of the earth's magnetic field

H is the scale height

ν is the ion-neutral collision frequency

Because the recombination coefficient β and the ion-neutral collision frequency both decrease with altitude, these heights can be readily calculated. The Martyn study was extended by Duncan (1956) who showed that the Chapman layer drifts bodily towards an equilibrium height which can be altered by a uniform vertical "tidal" drift. The amount that the height is perturbed is given by

$$\Delta z_m = \operatorname{arcsinh} \left[\frac{V_z v_0}{2g \sin^2 I} \right] \quad (1.2)$$

where

z_m is the reduced height of the electron density maximum

V_z is the vertical drift velocity

v_0 is the ion-neutral collision frequency at the unperturbed equilibrium height

g is the acceleration due to gravity

I is the inclination of the earth's magnetic field

The equilibrium continuity equation was solved by Rishbeth and Barron (1960) with the help of a digital computer. Various mixing ratios of nitrogen and oxygen were employed. Through empirical considerations the authors concluded that:

- 1) the electron density peak occurs where the recombination rate is approximately equal to the diffusion rate (i.e., $\beta \approx d$);
- 2) at the peak and below it the electron density is approximately given by the ratio of production to loss (i.e., $N_e \approx Q/\beta$);
- 3) at levels greater than one scale height above the peak the electron distribution is controlled by diffusion and takes an exponential form;

- 4) a vertical drift, V_z , alters the height of the peak of the relationship $\Delta z_m \approx -.9V_z/Hd$ where d is the diffusion rate;
- 5) the above results are not greatly affected if the scale height of the atmosphere varies with height.

In a later paper, Rishbeth (1972) showed that $V_z = -U_m \sin I \cos I$.

Chandra et al. (1960) rearranged the continuity equation in the form

$$V_z = \frac{N_{e0} V_{z0}}{N_{ez}(z)} - \frac{1}{N_{ez}(z)} \int_{z_0}^z \left[\frac{\partial N_e}{\partial t} + \beta N_e - Q \right] dz \quad (1.3)$$

where

V_z is the vertical drift velocity

N_e is the electron density

z is altitude

t is time

β is the recombination rate coefficient

Q is the production rate

The subscript zero refers to the value at a reference level.

Electron density profiles obtained from ionograms were used to calculate vertical drifts as a function of height. Because the drift velocity constant at the lower boundary was unknown, a series of V_{z0} values from $+20 \text{ m-sec}^{-1}$ to -20 m-sec^{-1} were assumed in order to obtain some sort of bracketing effect. Ionogram data from three low latitude and one mid-latitude station were used. The calculations showed a predominant downward velocity at night and an upward velocity during the day both in summer and winter at all stations except at Washington, D.C., where the velocities were downward in winter during the day.

In these calculations the authors assumed a recombination coefficient which was constant with time. It is now known that the recombination coefficient is a function of the molecular oxygen and molecular nitrogen densities which show a large diurnal variation.

Doupnik (1967) employed the technique of Chandra et al. to determine vertical drifts at Arecibo. Because electron density profiles from the Arecibo incoherent scatter radar were available high above the F2 peak, Doupnik was able to assume that the vertical fluxes at an upper altitude of 700 km were negligible when compared to fluxes at lower altitudes. In this way the indeterminate constant of integration was determined. The vertical drifts in an altitude range from 250 km to 400 km as calculated for each of six incoherent scatter experiments ranged from -40 m-sec^{-1} to 20 m-sec^{-1} . The diurnal electron density peak variations were shown to be correlated with and to lag the vertical drift velocity changes by about an hour.

1.2.2 Ion Drift and Neutral Wind Measurements

In 1967 it was first suggested and demonstrated by the French group (Carru et al., 1967) that it was possible to measure bulk ion drift velocities in the ionosphere by means of an incoherent scatter radar system. Ion drift measurements are now made by all of the incoherent scatter stations. The basic techniques and principles applied in such measurements were reviewed by Evans (1972a). Primarily, the method involves detecting the doppler shift of the ion power spectrum from the radar frequency. Since the ratio of this doppler shift to the overall doppler broadening of the power spectrum is small, the drift measurement is difficult to make.

Some of the more common sources of error in drift measurements are transmitter chirp (characteristic of pulse radar systems), system frequency response, statistical errors and asymmetrical power spectra. The extent of these errors is a function of the incoherent scattering system, the ionospheric parameters to be measured (i.e., T_e , T_i and ion composition) and the data taking procedure. Various authors (Evans et al., 1970; Carpenter, 1971; Vasseur, 1969; Woodman and Hagfors, 1969; Woodman, 1970; Behnke, 1970; and Harper, 1971) estimated the extent of the above types of error for their particular systems. The error estimates given by these authors for vertical drift measurements were in the range of $\pm 1 \text{ m-sec}^{-1}$ to $\pm 15 \text{ m-sec}^{-1}$, and typical errors were of the order of $\pm 5 \text{ m-sec}^{-1}$.

Vasseur (1969) reported on the vertical drift observations made at CNET. (The drifts were measured in a direction parallel to the magnetic field. The results presented were projections of the measured drifts onto the vertical and thus do not account for motions caused by electric fields.) Seasonal observations for the years 1966-1967 were presented and the drifts were harmonically analyzed. The general features of the drift behavior were that the ions moved downward during the day, with speeds ranging between -20 m-sec^{-1} and -40 m-sec^{-1} , and at night the flow was near zero, with speeds ranging from $+10 \text{ m-sec}^{-1}$ and -10 m-sec^{-1} . Some of the more striking seasonal features were that a clear maximum was observed around 500 hours followed by a morning decrease in the summertime; for the winter the daytime drifts were strongly negative; and for equinox a weak diurnal amplitude variation was observed. The CNET measurements were averaged for the altitude range of 200-400 km.

Results on observations made at CNET for the years 1968-1969 were presented by Amayenc and Vasseur (1972). In this study the authors measured the drifts parallel to the magnetic field line and then, after calculating the diffusion velocity, deduced the horizontal component of the neutral wind in the plane of the magnetic field. The neutral winds were calculated for summer, equinox and winter conditions. The summer neutral winds were shown to be less negative (northward) than the winter neutral winds during the day and the change in wind direction from southward to northward occurred earlier in the summer. The equinox neutral winds results were more like the summer results than the winter results.

At Millstone Hill, Evans et al. (1970) and Carpenter (1971) performed drift measurements for the altitude range 450-900 km. For the year 1968 vertical drifts for equinox conditions showed little diurnal variation and the average values were approximately zero. Winter observations were more negative and showed greater diurnal variations. For both seasons a sharp upward drift was observed for the period after sunrise.

Vertical drifts were calculated by Taylor (1974) using data from the new multistatic CW incoherent scatter radar at Malvern. For two days of observation the author showed that the vertical drifts were in the range of $-15 \pm 15 \text{ m-sec}^{-1}$.

At Arecibo, in the 250-400 km range, Behnke (1970) made vertical drift measurements for the winter of 1969-1970, and Harper (1971) made the same type of measurements for equinox and winter conditions during 1971. The winter observations showed a significant diurnal amplitude variation with the values being mostly downward except for a two-hour

period occurring some time between 2200 and 400 hours. Because of the scarcity of data, little was said about this nighttime reversal. As in the case of the CNET and Millstone Hill measurements, the vertical drifts showed very little diurnal variations for equinox conditions. The equinox vertical drifts as measured by Harper were downward with speeds of about -15 m-sec^{-1} .

Harper made three dimensional drift measurements and was able to calculate the drift parallel to the magnetic field line. After the diffusion velocity was determined, the horizontal neutral wind in the plane of the magnetic field was calculated. The distinguishing feature of this wind was that it reflected the nighttime reversal as was seen in the vertical drift measurements. Subsequent measurements (Behnke and Harper, 1973; Harper, 1973) have indicated that this nighttime reversal is a regular feature at Arecibo.

Woodman (1970) reported on data consisting of nearly two years of vertical drift observations made at Jicamarca for the years 1968-1969. Measurements were made over the altitude range 150-450 km with the measurements between 300-450 km being dubbed most accurate. No significant drift variations with altitude were observed over this range. In general, the drifts were upward during the day ($+20 \text{ m-sec}^{-1}$) and downward at night (-20 m-sec^{-1}) with the reversal times being one to two hours after sunset and within one hour of sunrise. The spread of velocities was large even on magnetically quiet days; however, during magnetically disturbed conditions larger than average fluctuations were observed. The diurnal variation was not sinusoidal.

1.2.3 Dynamo Generated Electric Fields

It was first suggested by Stewart in 1882 that dynamo action in the ionosphere was responsible for the diurnal magnetic variations observed on the ground. In the ionosphere the driving force for the dynamo is the neutral wind which induces electromotive forces by moving electrons and ions perpendicular to the geomagnetic field. The motion of the charged particles is dependent upon collisional forces and hence the conductivity. In the E region of the ionosphere the conductivity is sufficient for current flow, but because the conductivity is anisotropic, the flow is restricted by polarization charges which are set up. It may be viewed that the polarization charges produce an electrostatic field which prevents the current flow from being divergent. A simple mathematical formulation of this mechanism is given as

$$\bar{J} = \sigma_c \bar{E}_t = \sigma_c (\bar{E}_s + \bar{U} \times \bar{B}) \quad (1.4)$$

where

$\bar{J}$ is the current density vector

σ_c is the conductivity tensor

$\bar{E}_t$ is the total electric field vector

$\bar{E}_s$ is the electrostatic field vector

$\bar{U}$ is the neutral wind velocity vector

$\bar{B}$ is the magnetic flux density vector

From observed magnetic variations it is possible to deduce the current density. A world-wide current density system was so derived by several investigators using data from a series of magnetic stations and the results were presented by Chapman and Bartels (1940). With a current density system as an input, Equation 1.4 was solved by many authors

after making assumptions concerning the location of the currents, the conductivity and the nature of the neutral wind field.

Maeda (1955) considered a conductivity model, which was a function of the solar zenith angle, and assumed that the neutral air motion was horizontal and non-rotational. The neutral winds could then be written as gradients of a scalar which after equation manipulation was the only unknown in the resulting differential equation. Current densities were obtained from measurements. Upon solving this differential equation, Maeda was able to deduce a wind system and the electrostatic fields.

A continuation of Maeda's work was presented by Kato (1956) who removed the condition of a non-rotational wind system. This complicated the solution of the problem by making it necessary to solve the equation of motion of the atmosphere simultaneously with the dynamo equation. The more significant differences in the results of the two studies were that the diurnal pressure maximum occurred almost ten hours earlier and that the semi-diurnal pressure amplification was less by a factor of two in the Kato study.

Other investigators (Maeda and Murata, 1968; Matsushita, 1969) assumed neutral wind and conductivity models and then compared the calculated current systems with the ones estimated from geomagnetic variations. The calculated current systems were shown to agree qualitatively with current systems derived from measurements.

Electrostatic fields are also generated in the magnetosphere as a result of the interaction of energy and momentum from the solar wind at the magnetospheric boundary. The method of interaction is dependent upon the magnetospheric model used (Piddington, 1960; Axford, 1969;

Dungey, 1961). Magnetospheric electrostatic fields can be mapped down to the ionosphere by means of the equipotential magnetic field lines. The reverse is also true. Farley (1959, 1960) studied the factors governing the amount of attenuation in such mapping processes.

The concept of a solar-wind dynamo (magnetosphere) and an ionospheric wind dynamo (E region) was reviewed by Maeda (1966). It was shown that by means of coupling along the magnetic field lines the two dynamos interact with each other. Maeda indicated that the task of separating out the effects of each dynamo was very complicated and concluded that only by the simultaneous solution of Maxwell's equations, the current equation, the continuity equation, the equation of motion and the heat equation could the dynamo theory be properly formulated.

Matsushita and Tarpley (1970) and Matsushita (1971) considered the effect of the E region dynamo processes on the magnetosphere. In the former reference the authors showed that, within reasonable limits, changes in conductivity and wind models did not drastically alter the electrostatic field distributions. These dynamo electrostatic fields are mapped into the magnetosphere where they cause plasma drifts. The drift velocities were shown to increase with distance from the earth because the magnetic field falls off more quickly than does the (mapped) electrostatic field which is reduced by spreading of the field lines. In the latter reference Matsushita showed what effects plasmaspheric motions, which are in turn caused by weak magnetospheric motions, could have on the E region dynamo current system. The author pointed out that current belief had been that geomagnetic variations were primarily caused by E region dynamo currents and the induced currents

in the ground and that magnetospheric currents were responsible for only a minor part of these variations. In this study Matsushita computed the dynamo region total electric field, which is a combination of the electrostatic field and the induced field, that should come down from the plasmasphere to produce the observed S_q current system and conductivities. Matsushita concluded that direct measurements of the winds in the lower ionosphere were necessary in order to determine the relative effects of the two current generation mechanisms, i.e., E region and magnetospheric dynamo action.

Stening (1973) employed an equivalent circuit method to determine the ionospheric current flow and the (1, -2) mode tidal winds as the driving force and calculated electrostatic fields in the ionosphere for different seasons and longitudinal zones. It was shown that the longitudinal and seasonal variations were considerable but that these variations could be reproduced by variations in the conductivities.

1.2.4 F Region Electric Field Measurements

By using incoherent scatter doppler detection techniques to measure the bulk plasma drift velocities in directions perpendicular to the magnetic field line, F region electric fields can be deduced. At Jicamarca, a magnetic equatorial station, Woodman (1970) in nearly two years of observations measured electric fields which are typically .5 mV/m eastward during the day and .5 mV/m westward at night. The spread of the electric field measurements was large for any one time at different days, even for magnetically quiet conditions. On magnetically disturbed days, the magnitude of the field could be as high as 2 mV/m. The reversal times were usually 1-2 hours after sunset and within an hour of sunrise, but the diurnal variation was not sinusoidal.

At Arecibo, a 32°N magnetic latitude station, Behnke and Harper (1973) measured electric fields for winter conditions. The morning electric fields were directed eastward and changed to westward at 1400 hours. The maximum magnitude of the daytime field was about 1.1 mV/m. A large eastward field of about 3mV/m occurred regularly at 0400 hours. The overall field pattern suggested an eight hour tidal mode in the dynamo region.

Evans (1972b) made measurements of horizontal drifts in the E and F regions at Millstone Hill, a 58°N magnetic latitude station. Because of lack of signal strength only daytime measurements were made. The electric fields derived by Evans exhibited amplitudes of approximately 1 mV/m and rotated in a clockwise progression through 360° in 12 hours. Because the E region drift measurements appeared similar to but lagged behind the F region drifts, Evans suggested that the drifts result from electrostatic fields generated by dynamo action by means of a solar semi-diurnal (2, 2) tidal oscillation. On one particular day the direction of the drift reversed close to the time of a magnetic storm. Evans indicated that in this instance the electric fields were of magnetospheric origin.

By carefully choosing the elevation angle of the incoherent scatter radar, Carpenter and Kirchhoff (1974) and Kirchhoff (1974) showed that nighttime drift velocity measurements could be made at Millstone Hill. However, because the root mean square deviations of the measurements were large, a series of daily measurements had to be averaged in order to derive a reasonable electric field model (Kirchhoff and Carpenter, 1974). This model indicated that the diurnal component was the dominant mode. The West-East electric fields were

directed eastward during most of the day except during the afternoon when small westward directed fields were observed. The eastward field reached a maximum of about 2 mV/m at about 2200 hours while a maximum westward field of about .5 mV/m occurred at 1400 hours. The North-South electric fields were larger than the West-East fields and had a maximum magnitude of nearly 3 mV/m. The authors pointed out that their experimentally derived model was in better agreement with Matsushita's theoretical model for the total electric field than other dynamo theories. This indicated that the F region fields measured at Millstone Hill originated predominantly from plasmaspheric motions.

Taylor (1974) presented results from data taken at the new multi-static CW incoherent scatter radar system at Malvern. The West-East electric fields calculated from averaged data showed magnitudes of approximately 1 mV/m but these magnitudes were as great as 4 mV/m on individual days. The author indicated that the variation of the averaged West-East field was similar to the variation of the total field as given by Matsushita (1971) but pointed out that the F region drifts (produced by electric fields) were like those given by Maeda (1963).

Doupnik et al. (1972) and Banks et al. (1973) measured electric fields at Chatanika, Alaska, a 68°N magnetic latitude station. These fields were very dependent upon magnetospheric activity and had magnitudes ranging from less than 1 mV/m at times of low magnetic activity to 40 mV/m and more for periods of increased activity. In general, the North-South electric fields were larger by a factor of 3-10 times the West-East fields.

1.2.5 Effects of Electric Fields on the Mid-Latitude F Region of the Ionosphere

The effects of electric fields perpendicular to the field line on ion motions in the ionosphere have been investigated extensively. For mid-latitudes Bramley (1969) solved the F region continuity equation by assuming no horizontal variations and employing neutral winds obtained from Geisler (1967) and electric fields derived by Maeda (1963) from dynamo theory. Though neutral winds were shown to be the dominant factor in producing HmF2 and NmF2 variations, electric fields did cause significant NmF2 variations at night. Thus, Bramley concluded that electric fields should be included in F region calculations.

Having set the North-South pressure gradients to zero, Stubbe and Chandra (1970) showed that mid-latitude West-East electric fields ranging from 7 mV/m eastward at 0200 hours to 5 mV/m westward at 1600 hours would be necessary to duplicate the results produced with the North-South pressure gradient in place. In this study, Jacchia's (1964) model was employed for neutral atmosphere parameters. It was pointed out that both the electric field strength and its time derivative are important. The reason is that the initial vertical electrodynamic drift velocity is nullified after a time by an opposing neutral gas motion which is built up by means of the frictional force. For the F region this time is typically from 15 minutes to 3 hours. From their calculations and from plasma drift measurements made at CNET, the authors deduced that the West-East electric field should not exceed 1-2 mV/m at night.

The neutral temperature of Jacchia's model was altered in order to agree with the results obtained from incoherent scatter data. This

caused an increase in the North-South directed neutral wind which in turn led to an unrealistic rise of the F layer. West-East electric fields were introduced to reduce this effect. Though these fields were similar in phase to those driven by Matsushita (1969) their amplitudes were much larger than Matsushita's amplitudes and also much larger than those which were deduced previously by the authors.

Bramley and Rüster (1971) using Jacchia's (1965) model with the particle densities reduced by 40 percent, showed that NmF2 is significantly affected by Matsushita's (1969) electric fields at night when ion drag is small. By day, ion drag substantially reduces the effect of the electric fields. Electric field induced hmF2 variations were of the order of 10 km at night. It was shown by Rüster (1971) that changes in neutral atmospheric models, within the range of their uncertainties, have far greater effects upon the ionosphere than do electric fields.

1.2.6 Neutral Wind Calculations

With Hanson and Patterson's (1964) suggestion that neutral winds, causing a vertical ion drift, might be responsible for the maintenance of the nighttime F layer, a series of neutral wind calculations began. King and Kohl (1965) showed that atmospheric pressure gradients derived from satellite drag data were large enough to cause atmospheric winds which produce important ionospheric drifts. Estimates of the neutral wind were made by equating the ion drag to the driving force from the pressure gradient.

Early neutral wind calculations (Lindzen, 1966; Geisler, 1966; Kohl and King, 1967) employed simplified forms of the equation of motion for neutrals. The ion drag, viscous and/or non-linear terms

were frequently neglected at first but then included as the development of neutral wind calculations proceeded. Kohl et al. (1968) recognized that the dependence of the neutral wind velocity and ionization upon each other was important and that the simultaneous solution of the continuity equation for ionization and equation of motion for neutral air was necessary for a satisfactory solution to the F region dynamics problem.

Kohl et al. (1968), Bailey et al. (1969), Cho and Yeh (1970) and Stubbe (1970) all made ion density and neutral wind calculations by simultaneously solving the ion continuity equation and the momentum equations under various conditions. In general, the authors showed that their results were in qualitative agreement with the current measured values. However, it should be mentioned that the degree of agreement was highly dependent upon the assumptions which were imposed. When discrepancies between calculated and measured results did occur, the authors frequently altered some of the terms in order to demonstrate their effect. A summary of the results achieved by the various authors with regard to neutral wind and related calculations was given by Rishbeth (1972).

In this review some of Rishbeth's more important conclusions were:

- 1) The neutral pressure gradient drives the winds and the ion drag and Coriolis forces, with some modification due to inertia, control the wind's direction.
- 2) At and above the F2 peak viscosity is important. At these altitudes the viscous force smooths out the height variation of the wind speed.

- 3) The magnetic meridional neutral wind produces some interesting effects with regard to the diurnal variations of H_mF2 and N_mF2 . The preservation of the F2 layer is very dependent on the local time at which this wind changes from poleward to equatorward. When the variations of the declination and inclination of the earth's magnetic field are taken into account, many of the geographic F2 layer variations can be explained.

Under outstanding problems the author emphasized that the great uncertainties that exist in the parameters of neutral atmospheres limited the accuracy of current neutral wind calculations. In particular it was pointed out that current neutral models are too highly averaged for the necessary temporal and spatial variations of pressure, density and temperature and that the fixing of the lower boundary is a major source of the uncertainties.

1.2.7 Neutral Atmospheric Models

In construction atmospheric models, the physics of the neutral air dictates that it be divided into two regions: a lower region which is well mixed by turbulence and an upper region in which diffusive separation is established. Since the upper region is the main interest in this study, this discussion of neutral models is aimed at this region with reference to the lower region being made through the introduction of boundary conditions. Two types of models of the upper region have been developed: static diffusive models and theoretical models. For each type a set of boundary conditions in the 120 km altitude range is assumed. The densities calculated from the models were compared with densities derived from satellite drag data and thus these models take on a semi-empirical nature.

Nicolet (1961) developed a family of static diffusion models in which the thermosphere was allowed to cool from a high temperature. It was found that under these circumstances the temperature profile for any given exospheric temperature was almost independent of the initial temperature profile. Jacchia (1961) used these temperature profiles and, on assuming diffusive equilibrium for all species, developed a model of the neutral atmosphere to fit the satellite orbital decay densities. It should be noted, however, that the one parameter series of profiles produced in this way can give very different results for temperatures and other parameters because the boundary conditions may vary with latitude, longitude and season, because the temperature profiles may differ greatly from the cooling profiles assumed and because relative transport of the various constituents may cause major departures from diffusive equilibrium.

Subsequent static diffusive models (U.S. Standard Atmosphere, 1962, 1966; Jacchia, 1964, 1965, 1970, and 1971; CIRA, 1972) were developed based on the procedure described above. The assumptions upon which these models were based did not change drastically during their evolutionary period; the amount of satellite data increased.

In fitting the satellite drag measurements to the mathematical relationships of the static models, a certain amount of smoothing was introduced. Barlier et al. (1972) determined the residuals that existed between the real data and the values given by the Jacchia (1971) static model. Though these residuals rarely exceeded 20 percent, they did not show a random variation. The authors reanalyzed the satellite drag data on a finer scale and showed that differences between

the newly derived neutral densities and especially the neutral density gradients and those calculated by the Jacchia model were substantial.

The temperature obtained from static neutral atmospheric models such as the CIRA and Jacchia models is not a true measure of the real neutral atmospheric temperature. It is merely a mathematical parameter which is used to define the neutral atmospheric densities. Nisbet (1967) showed that temperatures derived from incoherent scatter data taken at Arecibo differed from those calculated from the Jacchia (1964) neutral model. More recently Blamont et al. (1974) presented global temperature distributions measured by the 6300Å airglow photometer on the OGO-6 satellite. The results showed that the location of the global maximum and the diurnal variation at all latitudes were very different than those determined from static neutral atmospheric models.

Hedin et al. (1974) derived an empirical model of the global thermospheric temperatures and the N_2 , O and H_e densities based on data from the OGO-6 quadrupole mass spectrometer. Two model schemes were used for representing the densities: one extrapolated the densities to 450 km and the other extrapolated the densities to 120 km. The temperatures were obtained by using a method similar to that employed by Jacchia (1965) except that only the N_2 densities were used (Jacchia used the total density). The time, solar, geographic and magnetic variations were introduced by fitting the data to a spherical harmonic expansion by using a least squares technique.

One of the first theoretical neutral models was developed by Harris and Priester (1962). These authors numerically solved an expression for the time-dependent energy balance in a one-dimensional system. It was necessary to introduce a second heat source which

was out of phase with the primary heat source, provided by the absorption of the solar EUV radiation, in order to obtain general agreement between calculated densities and densities observed from the reduction of satellite drag data.

It was realized that the introduction of transport mechanisms was a means for overcoming some of the shortcomings of the Harris-Priester model and that ion drag was an important factor in controlling the transport. To account for these effects two- and three-dimensional models were developed by a number of researchers, some of whom are Volland and Mayr (1968, 1972a, 1972b, 1972c, 1973), Friedman (1970) and Vest (1973a). An excellent review of the evolution of theoretical models was given by Blum et al. (1972).

1.2.8 Boundary Condition Effects

It was mentioned in Section 1.2.6 that the lower boundary conditions used by neutral models were inaccurate. These values were obtained by way of measurements (mass spectrometers, vapor trails, etc.). Because the difficulties in making the measurements are great (Von Zahn, 1967) and observations are scarce, an adequate global model or even a local model of the temperature and density variations at the lower boundary is not yet available. For this reason the values used are somewhat arbitrary and variations of the lower boundary with time, location and solar activity are rarely introduced in thermospheric models.

Harris and Priester (1965) studied the effects that various alterations of the lower boundary had on their neutral model. By changing the level of turbopause diurnally, Harris and Priester were able to get good agreement between calculated and observed densities at

600 km but they were not able to get good agreement at lower altitudes simultaneously. When the atomic to molecular oxygen ratio and temperature were varied as a function of time, no significant effects on the upper thermospheric neutral densities and temperatures were observed. Consequently, Harris and Priester concluded that diurnal variations at the lower boundary cannot account for phase changes of the upper thermospheric densities.

Blum (1969) showed that by artificially changing the temperature gradient at the lower boundary the proper density variation in the upper thermosphere could be reproduced.

Chandra and Stubbe (1970) sinusoidally varied the densities and temperature at the lower boundary. They found that by proper choice of the amplitude and phase factors of the sinusoidal variations at the lower boundary, the temperature and densities at 300 km could be made to agree with density measurements from satellites and temperature measurements from incoherent scatter. Thus the authors suggested that in addition to altering upper thermospheric behavior, the variations at the lower boundary could also resolve the upper atmospheric densities and temperature anomaly.

In his later static models, Jacchia (1971) tried to compensate for variable boundary conditions at 125 km by moving his constant boundary condition down to 90 km. Empirical density and temperature variations were derived for the 90 km-125 km altitude range. The temperature at 125 km is an empirical function of the exospheric temperature. As a result of Jacchia's efforts, density variations at 125 km are of the order of ± 1 percent and temperature variations are typically $\pm 10^\circ\text{K}$. By empirical mathematical manipulation Jacchia

attempted to derive a model showing the best possible agreement with densities measured by way of satellite drag, mass spectrometer, EUV absorption, etc.

Stubbe (1973) derived a relationship between $n(0)$, $n(O_2)$ and $n(N_2)$ at 125 km and arranged it so that $n(O_2)$ and $n(N_2)$ were functions of $n(0)$. Stubbe's (1970) ionospheric program was run and the calculated NmF2 and HmF2 were compared with their respective measured values for noontime conditions. The measured values were monthly averages obtained from reduced ionograms for seven northern hemisphere stations for the years 1958-1959 and 1963-1964. By adjusting the $n(0)$ density at 125 km (and hence implicitly the $n(O_2)$ and $n(N_2)$ densities) and the component of the neutral wind in the plane of the magnetic field in an iterative manner, Stubbe was able to make the calculated and measured values of the peak height and peak densities agree. During this process the total neutral density at 350 km was held at the value given by the Jacchia (1971) model by simultaneously altering the exospheric neutral temperature. The result of the study was that an empirical relationship between the densities at 125 km and latitude, day number and solar activity was obtained. This new atmospheric model, which is actually Jacchia's model with modified lower boundary conditions and modified exospheric temperatures, seemed to improve the photochemical aspects of F region modeling, but as far as the dynamic aspects were concerned both models were equivalent.

1.2.9 Accuracy of the Ionospheric Parameters HmF2 and NmF2

The ionospheric parameters HmF2 and NmF2 that are used in this study were obtained from ionosonde measurements. Because the group velocity of radio waves in the ionosphere varies with the electron

density, the ionograms, which are records of the group delay as a function of frequency, require processing to obtain the electron density profiles as a function of height. An extensive discussion of the problems encountered in such processing was presented by Wright, Smith, Paul, Titheridge, Becker and a number of other authors in the special issue of Radio Science in October, 1967. The most significant problem appeared to be the determination of the correction for underlying ionization. Underlying ionization is the term used to refer to the ionization which exists at the lower levels of the ionosphere but which is not recorded on the ionograms because it corresponds to plasma frequencies below the operating range of the ionsonde. These corrections are largest at night.

Bennett (1974) tested the more commonly used ionogram reduction programs including the Wright program used to reduce the data for this study, by using both theoretical and real data for a number of conditions. Virtual height profiles calculated theoretically from analytic electron density profiles were used and various studies were made in which the data was truncated and in which random errors were added to the data set in order to simulate measurement errors. In all cases Bennett showed that the error in determining the peak height, H_mF_2 , was of the order of 5 km and the error in determining the peak density was negligible. By using real data for nighttime conditions, it was estimated that the corrections for underlying ionization were of the order of 5 km at the peak. These corrections were shown to be much greater at other altitudes. The results obtained by Bennett were in agreement with those presented in the earlier studies.

1.3 Specific Statement of the Problem

The problems to be investigated in this study are the following:

- 1) To develop and evaluate a method for using electron density profiles derived from ionograms to calculate the vertical component of the ion drift velocity and an equivalent meridional wind.
- 2) To calculate neutral winds and electric fields from the equivalent meridional winds, to compare these results with measurements and theoretically calculated values and to determine the significance of the electric fields.
- 3) To determine what pressure distributions are required to give the observed wind velocities and to develop a set of sensitivity factors which relate the pressure gradients and the neutral winds.

CHAPTER II

DEVELOPMENT OF EXPRESSIONS RELATING THE ELECTRON DENSITY PROFILE TO THE WINDS AND THE ELECTRIC FIELDS

2.1 Procedure

The vertical motion of ions and electrons in the F2 region of the ionosphere is caused by diffusion, neutral winds, and electric fields which are perpendicular to the magnetic field. In the following sections relationships are derived between the electron density profile and the vertical component of the ion velocity and between the vertical component of the ion velocity and an equivalent wind in the magnetic meridional direction.

The equivalent wind in the magnetic meridional direction has been used by Rishbeth (1972) and Stubbe (1973). It is convenient because it represents an equivalent horizontal wind in the plane of the magnetic meridian which would duplicate the effects of the neutral winds and electric fields. If only a series of electron density profiles from one station is available, then this gives all the information obtainable about the winds and electric fields. If a large number of stations are available with different declinations and inclinations, the equivalent winds in the magnetic meridional directions can be used to provide information about the zonal and meridional winds. This topic will be discussed in Chapter III.

2.2 Assumptions and Derivation of the Relationship Between the Electron Density Profile and the Vertical Component of the Ion Velocity

The continuity equation for ions is

$$\frac{\partial n_i}{\partial t} = P_i - L_i - \text{div} (n_i \bar{V}_i) \quad (2.1)$$

where

n_i is the number density of the i -th ion species

P_i is the production rate of the i -th ion species

L_i is the loss rate of the i -th ion species

$\bar{V}_i$ is the velocity vector of the i -th ion species

Production results from photoionization of neutrals and by charge exchange due to chemical reactions. The photoionization, $P_i^{(p)}$, contribution of P_i is given by

$$P_i^{(p)}(z) = n_n(z) \int_0^\infty \left[\sigma_n^{(i)}(\lambda) \Phi_\infty \exp \left[-\sec \chi \sum_n \sigma_n^{(a)}(\lambda) \int_z^\infty n_n dz \right] \right] d\lambda \quad (2.2)$$

where

λ is the wavelength of the radiation

$\sigma_n^{(i)}, \sigma_n^{(a)}$ are the ionization and absorption cross sections of the n -th neutral constituent in the wavelength range $\lambda - (\lambda - d\lambda)$

Φ_∞ is the unabsorbed photon flux

χ is the solar zenith angle

Loss results solely from chemical reactions. The photon fluxes, ionization and absorption cross sections, reactions, and production and loss rates that are employed in this study are given in Appendix A. Because of charge neutrality the number density for electrons is given by

$$N_e = \sum_i c_i n_i \quad (2.3)$$

where

N_e is the electron number density

c_i is the charge number of the i -th ion species

For the ion continuity equation it is assumed that horizontal flux gradients are negligible and Equation 2.1 reduces to

$$\frac{\partial n_i}{\partial t} = P_i - L_i - \frac{d(n_i V_{zi})}{dz} \quad (2.4)$$

where

V_{zi} is the vertical velocity of the i -th ion species

z refers to the vertical direction, upwards being positive

Equation 2.4 can be rearranged and integrated to yield

$$n_i V_{zi} \Big|_1^u = \int_1^u \left[P_i - L_i - \frac{\partial n_i}{\partial t} \right] dz \quad (2.5)$$

The vertical velocity at the lower boundary is then

$$V_{zil} = - \frac{1}{n_{il}} \int_1^u \left[P_i - L_i - \frac{\partial n_i}{\partial t} \right] dz + \frac{(n_i V_{zi})_u}{n_{il}} \quad (2.6)$$

where

u refers to the upper boundary

1 refers to the lower boundary

In this way, the vertical velocity of the i -th ionic constituent at the lower boundary is given as a function of the density of the i -th ion at the lower boundary, the production, loss and time derivative function of the i -th ion as a function of height and the flux of the i -th ion at the upper boundary. For this study the flux at the upper boundary, which is taken to be 1000 km, is considered negligible and Equation 2.6 reduces to

$$V_{zil} = - \frac{1}{n_{il}} \int_1^{1000\text{km}} \left[P_i - L_i - \frac{\partial n_i}{\partial t} \right] dz \quad (2.7)$$

This assumption is investigated in Section 2.6. If the density profile for the i -th ion from the lower to upper boundary can be specified, the remaining terms can be obtained from theoretical considerations and the neutral model and Equation 2.7 can be solved for the vertical drift of the i -th ion.

In the F2 region of the ionosphere atomic oxygen ions are dominant and therefore the ionospheric parameters H_mF2 and N_mF2 are a very good approximation to the height and density of the atomic oxygen peak. As an example of the proximity of the electron density and atomic oxygen peaks, their heights are presented in Figure 1 and densities are presented in Figure 2 as calculated by using the Stubbe ionospheric model for a typical case. The general description is that the height of the atomic oxygen peak is at most a few kilometers higher than the height of the electron density peak. The density of the atomic oxygen peak is at most a few percent less than the electron density peak. By examining the right hand side of Equation 2.7 it can be seen that the errors caused by these approximations are in the opposite sense and hence tend to cancel. These differences are most distinguishable during the day when the layer is lower.

By choosing the i -th ion to be an atomic oxygen ion and the lower boundary to be the F2 electron density peak, H_mF2 , Equation 2.7 can be solved if the shape of the atomic oxygen profile is known. The shapes of these profiles are considered in Sections 2.4 and 2.5.

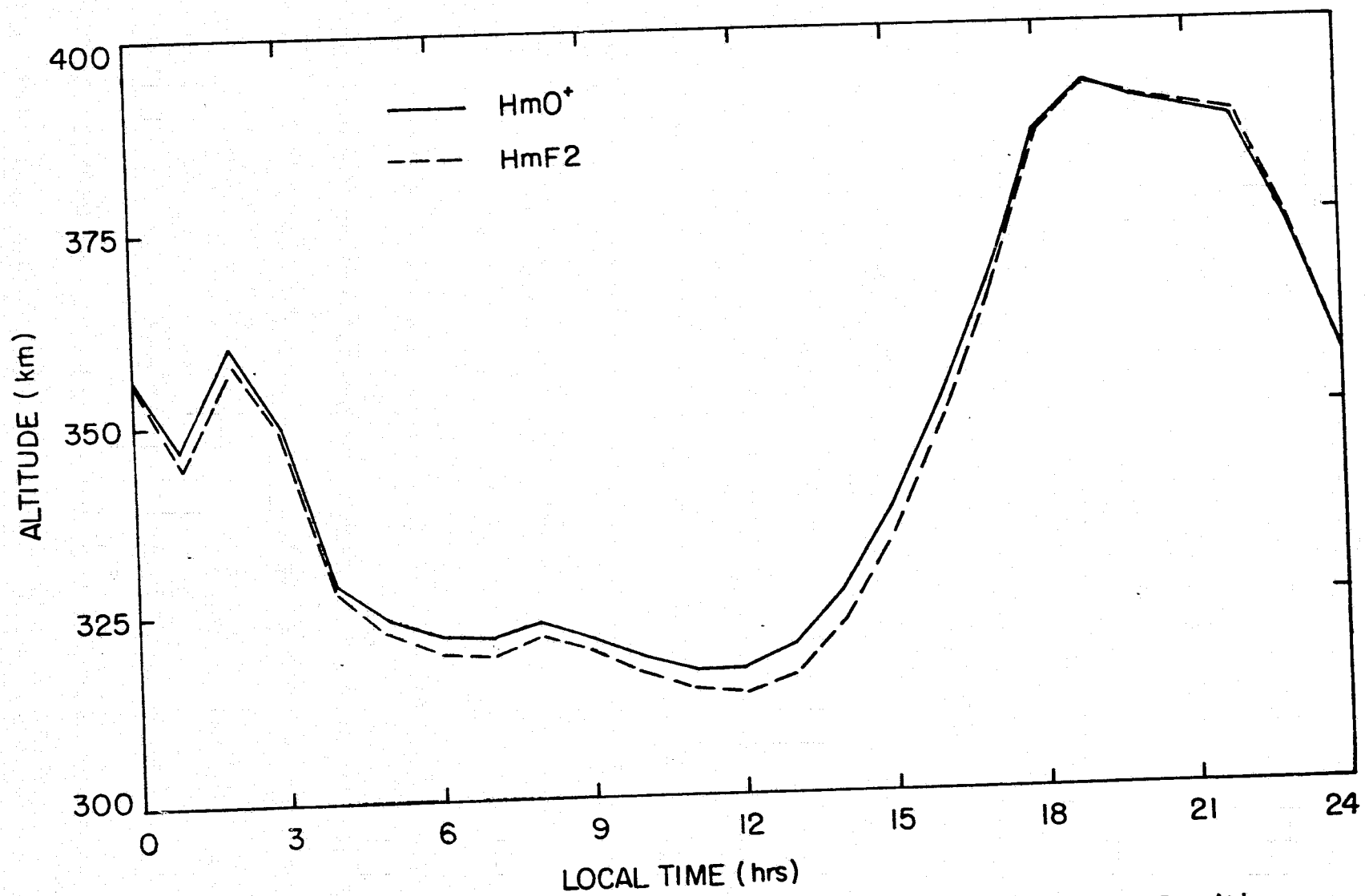


Figure 1. Comparison of the Peak Heights of the Electron and Atomic Oxygen Ion Densities

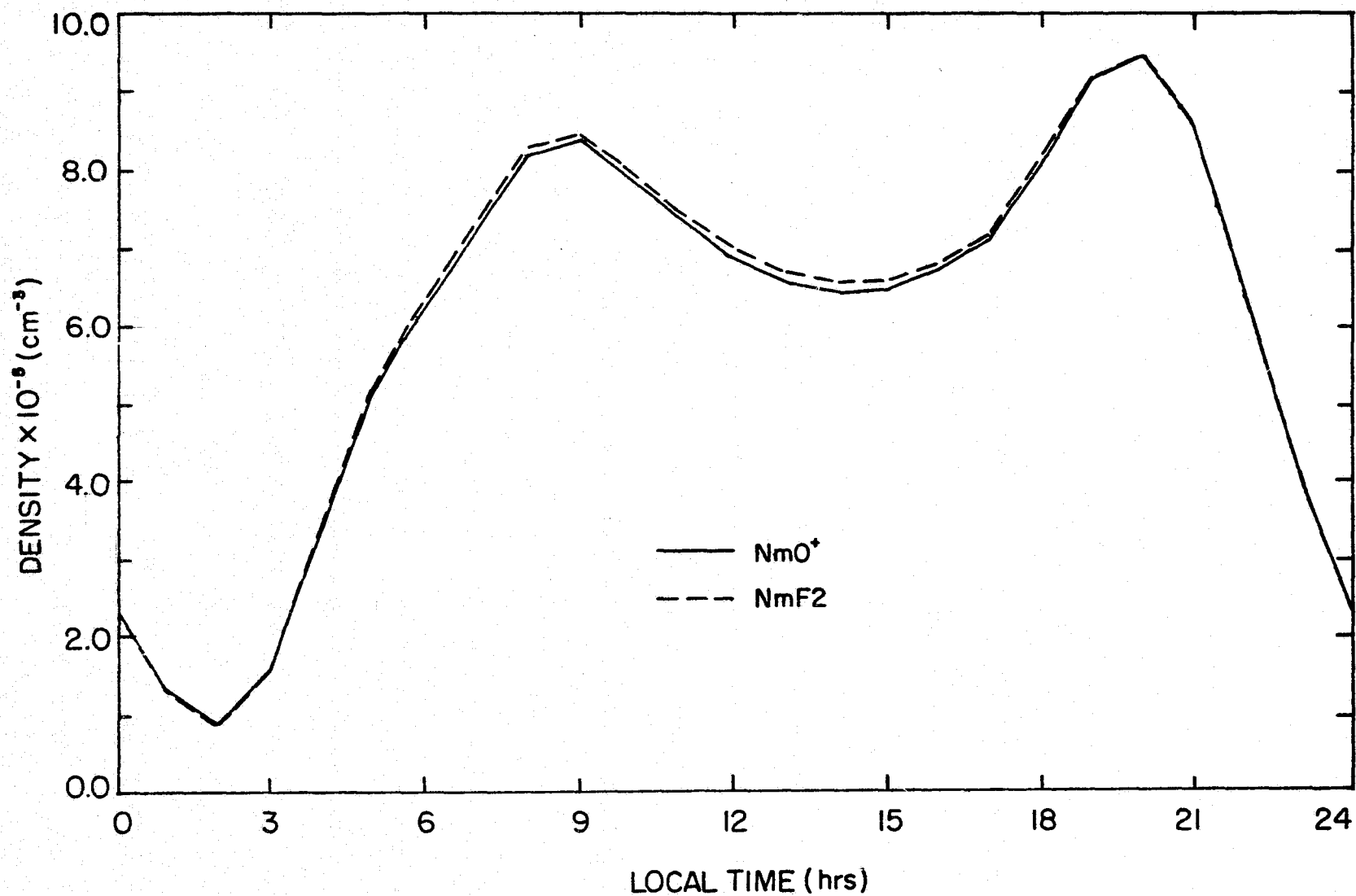


Figure 2. Comparison of the Peak Density of Electrons and Atomic Oxygen Ions

2.3 Assumptions and Derivation of the Relationship Between the Vertical Component of the Ion Velocity and an Equivalent Wind in the Direction of the Magnetic Meridian

The equation of motion for the ions is

$$\begin{aligned}
 \frac{d\bar{v}_i}{dt} + \bar{\Omega} \times (\bar{\Omega} \times \bar{R}) + 2(\bar{\Omega} \times \bar{v}_i) = & -\frac{1}{m_i} (\bar{v}_i - \bar{U}) \sum_n v_{in} m_{in} \\
 & - \frac{1}{m_i} \sum_{k=i} v_{ik} m_{ik} (\bar{v}_i - \bar{v}_k) - \frac{1}{\rho_i} \text{grad}(p_i) + \frac{\mu_i}{\rho_i} \nabla^2 \bar{v}_i + \bar{g} \\
 & + \frac{e}{m_i} (\bar{E} + \bar{v}_i \times \bar{B})
 \end{aligned} \tag{2.8}$$

and the equation of motion for the electrons is

$$\begin{aligned}
 \frac{d\bar{v}_e}{dt} + \bar{\Omega} \times (\bar{\Omega} \times \bar{R}) + 2(\bar{\Omega} \times \bar{v}_e) = & -\frac{1}{m_e} (\bar{v}_e - \bar{U}) \sum_n v_{en} m_{en} \\
 & - \frac{1}{m_e} \sum_i v_{ei} m_{ei} (\bar{v}_e - \bar{v}_i) - \frac{1}{\rho_e} \text{grad}(p_e) + \frac{\mu_e}{\rho_e} \nabla^2 \bar{v}_e + \bar{g} \\
 & - \frac{e}{m_e} (\bar{E} + \bar{v}_e \times \bar{B})
 \end{aligned} \tag{2.9}$$

where

$\bar{v}$ is the velocity vector

$\bar{\Omega}$ is the earth's angular rotation vector

$\bar{R}$ is the radius vector from the earth's center

m is the mass

$\bar{U}$ is the neutral wind velocity vector

v is the collision frequency

m_{in} is the reduced mass for the i -th ion and n -th neutral species

ρ is the density

μ is the coefficient of viscosity

$\bar{g}$ is the acceleration due to gravity

e is the charge

$\bar{E}$ is the electric field vector

$\bar{B}$ is the magnetic field vector

The subscripts n, i and e refer to neutrals, ions and electrons, respectively.

The acceleration, Coriolis, centripetal and viscous forces are neglected in both the equations of motion for ions and electrons. Because the electron mass is so much smaller than the ion mass, the gravity and drag (friction) terms are also neglected in the equation of motion for electrons. With these assumptions the momentum equations for ions and electrons are greatly simplified and are given respectively as

$$0 = \frac{-1}{m_i} (\bar{V}_i - \bar{U}) \sum_n v_{in} m_{in} - \frac{1}{m_i} \sum_k v_{ik} m_{ik} (\bar{V}_i - \bar{V}_k) - \frac{1}{\rho_i} \text{grad} (p_i) + \bar{g} + \frac{e}{m_i} (\bar{E} + \bar{V}_i \times \bar{B}) \quad (2.10)$$

and

$$0 = - \frac{1}{\rho_e} \text{grad} (p_e) - \frac{e}{m_e} (\bar{E} + \bar{V}_e \times \bar{B}) \quad (2.11)$$

Assuming that polarization fields due to charge separation of ions and electrons are negligible, that horizontal gradients are negligible and that the divergence of the current density is zero, Equations 2.10 and 2.11 can be written in terms of scalar components. Because of the

summations involved in the ion drag terms, this becomes quite complicated and it is convenient to divide the ionosphere into regions over which different terms, such as the ion-ion collision frequency, ion-neutral collision frequency and gyro frequency, can in turn predominate. A complete treatment of this has been given by Stubbe (1973, 1970).

For the region of interest to this study, namely the F2 region, the ion-ion collision force can be neglected and the ion-neutral collision frequency can be considered small compared to the gyro frequency. Under these assumptions and by noting that $p = nkT$, Equations 2.10 and 2.11 can be resolved into their scalar form and then solved to yield the component of the ion velocity in the vertical direction.

$$\begin{aligned}
 V_{zi} = & \frac{E_x}{B_0} \cos I \sin D + \frac{E_y}{B_0} \cos I \cos D \\
 & + U_x \sin I \cos I \cos D - U_y \sin I \cos I \sin D \\
 & - \left[\frac{\sin^2 I}{\sum_n v_{in} m_{in}} \left(\frac{k}{n_i} \left(\frac{\partial n_i (T_e + T_i)}{\partial z} \right) + m_i g \right) \right] \quad (2.12)
 \end{aligned}$$

where

I is the inclination of the earth's magnetic field

D is the declination of the earth's magnetic field

k is Boltzmann's constant

T is temperature

The last term in this equation is referred to as the diffusion term.

By definition the equivalent wind in the magnetic meridional direction,

U_M , is given as

$$U_M = U_x \cos D - U_y \sin D + \frac{E_x}{B_0} \frac{\sin D}{\sin I} + \frac{E_y}{B_0} \frac{\cos D}{\sin I} \quad (2.13)$$

By combining Equations 2.7, 2.12 and 2.13 an expression for U_M is obtained

$$U_M = \frac{1}{\sin I \cos I} \left[-\frac{1}{n_{i1}} \int_1^{1000} \left[P_i - L_i - \frac{\partial n_i}{\partial t} \right] dz \right. \\ \left. + \frac{\sin^2 I}{\sum_n v_{in} m_{in}} \left(\frac{k}{n_i} \left(\frac{n_i (T_i + T_e)}{\partial z} \right) + m_i g \right) \right] \quad (2.14)$$

In order to solve Equation 2.14, only the ion density profile and its time derivative can be supplied from reduced ionograms. The remaining terms have to be obtained using theoretical considerations.

2.4 Development of the Density Profiles

As shown in Sections 2.2 and 2.3 the solution for vertical drifts and neutral winds requires that the ion density profile be known. Two schemes for the estimation of the density profile are presented here. The first is the simple Chapman C_α layer approximation which is given as

$$n_i(z) = NmF2 \exp \left[.5 \left[1. - \frac{z - HmF2}{H} - \exp \left(- \frac{z - HmF2}{H} \right) \right] \right] \quad (2.15)$$

where

$H = kT/m_i g$ is the scale height of atomic oxygen

As indicated by Martyn (1956) and Duncan (1956), the Chapman C_α layer provides an excellent approximation to the true profile at night when diffusion, recombination and uniform vertical tidal drifts need be considered. During the day, the complicated production of molecular ions renders the Chapman approximation less valid. Nevertheless, as will be shown in Section 2.5, the Chapman approximation can be useful in this study.

In order to improve upon the daytime approximation of the Chapman layer, a second method for estimating the density profile was developed. This method is referred to as the "model method." As an initial condition a Chapman layer is used. Midnight is chosen because at this time the Chapman approximation is most accurate. The ion continuity equation is then solved for the next time data is available and a new ion density profile is produced. Because the vertical drift is the unknown, the new ion density profile exhibits an incorrect peak height and peak density. But, because the H_mF2 and N_mF2 values are known, the calculated ion density profile can be normalized to these quantities. The normalized profile then serves as the input for the next time iteration. This procedure is repeated until solutions for an entire 24-hour period are obtained. In this way profiles whose shapes are dependent upon production, loss and diffusion are generated. Implicit in such a treatment is the assumption that the wind induced vertical drifts do not cause variations in the shape of the profile. This, of course, is only approximate; nevertheless, as will be shown in Section 2.5, the profiles generated by the "model method" are an improvement over the simple Chapman approximation (see Appendix B for details).

2.5 Comparison of the Various Methods for Calculations of the Equivalent Wind in the Magnetic Meridional Direction

In this section three methods for estimating the wind are compared. These are:

- 1) the Rishbeth method described in Section 1.2.1,
- 2) the method derived from the continuity equation (Section 2.4) using the Chapman layer approximation, and

- 3) the method derived from the continuity equation (Section 2.4) using the "model method" for the profile shape.

To ensure that the conclusions reached in this comparison study are not unduly biased by a particular set of geographical and seasonal conditions, the comparisons have been made for summer and winter conditions for the stations Adak and Puerto Rico. These conditions represent the extremes of the data base used in this study. As a basis for the evaluation procedure, Stubbe's ionospheric model was employed in the following manner. The model was run for a series of conditions which bracket the extremities of this study. The products of the model's calculations were HmF2, NmF2 and the equivalent wind in the magnetic meridional direction. By using the parameters HmF2 and NmF2 as inputs, the derived expressions were evaluated on their ability to reproduce the equivalent wind.

The Rishbeth method requires HmF2 as an input while the methods employing the ion continuity equation require HmF2 and NmF2 data as inputs.

In Figures 3, 4, 5 and 6 the results of the comparisons are presented. In each figure the solid line represents the equivalent wind in the magnetic meridional direction supplied by the Stubbe program. This is the standard with which the results of the other models are to be compared. From Figures 3 and 5 it is apparent that the model method is best followed by the Chapman layer method. In Figures 4 and 6, for winter conditions the distinction between the model and Chapman layer methods is less and both methods are comparable. In all four cases the Rishbeth method produces the poorest agreement.

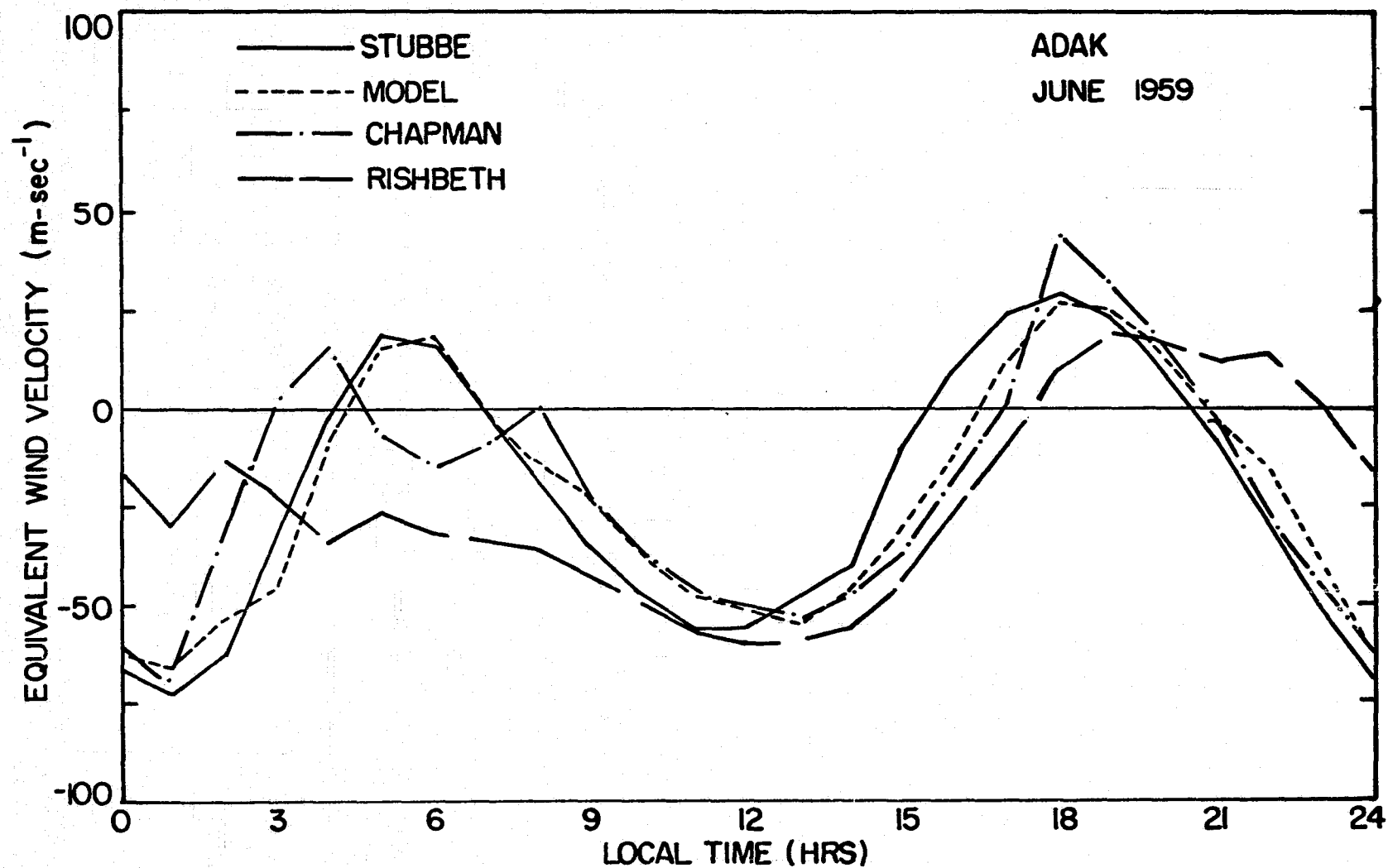


Figure 3. Comparisons of Methods for Estimating the Equivalent Magnetic Meridional Wind for Summer Conditions at Adak

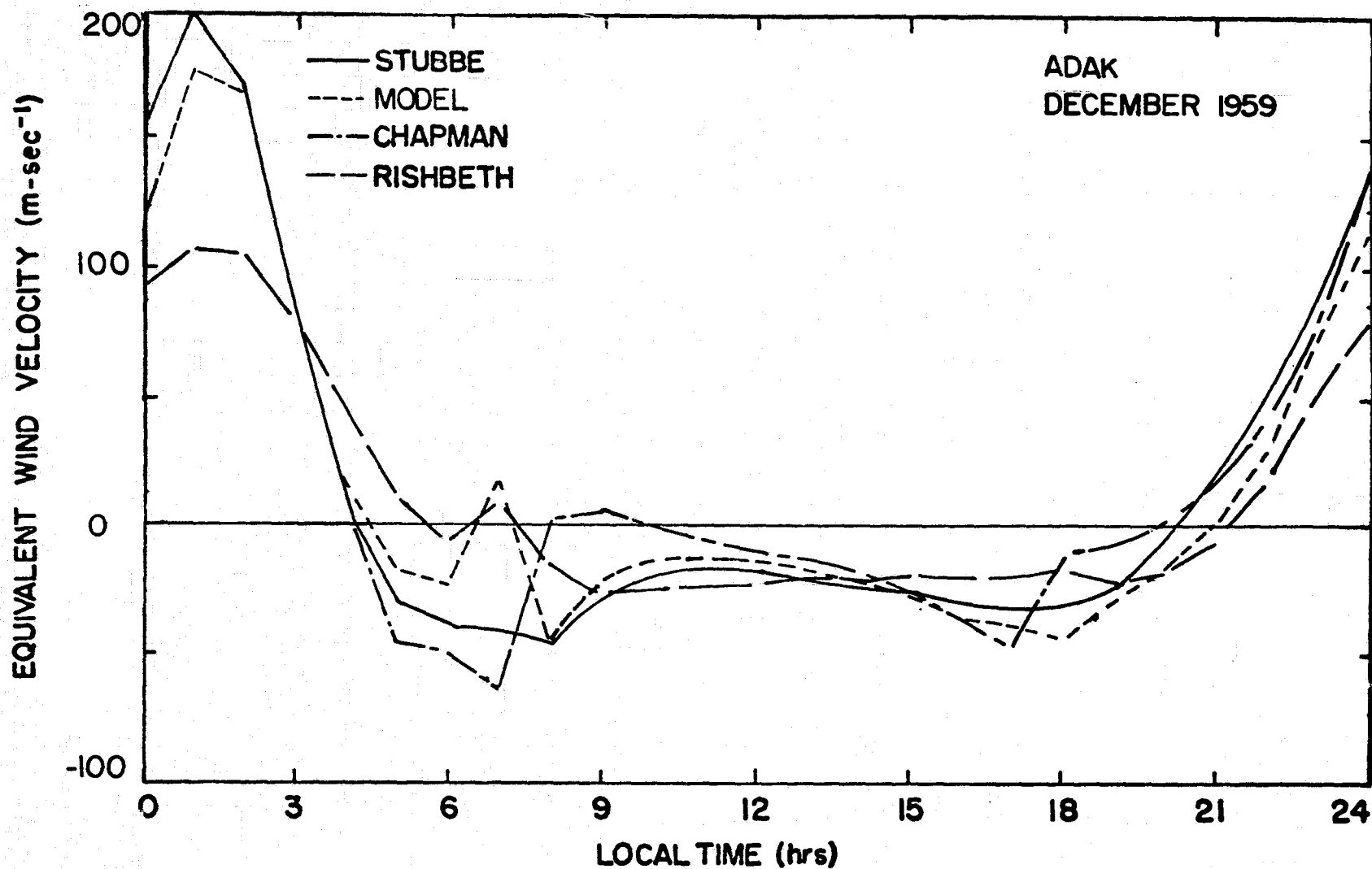


Figure 4. Comparisons of Methods for Estimating the Equivalent Magnetic Meridional Wind for Winter Conditions at Adak

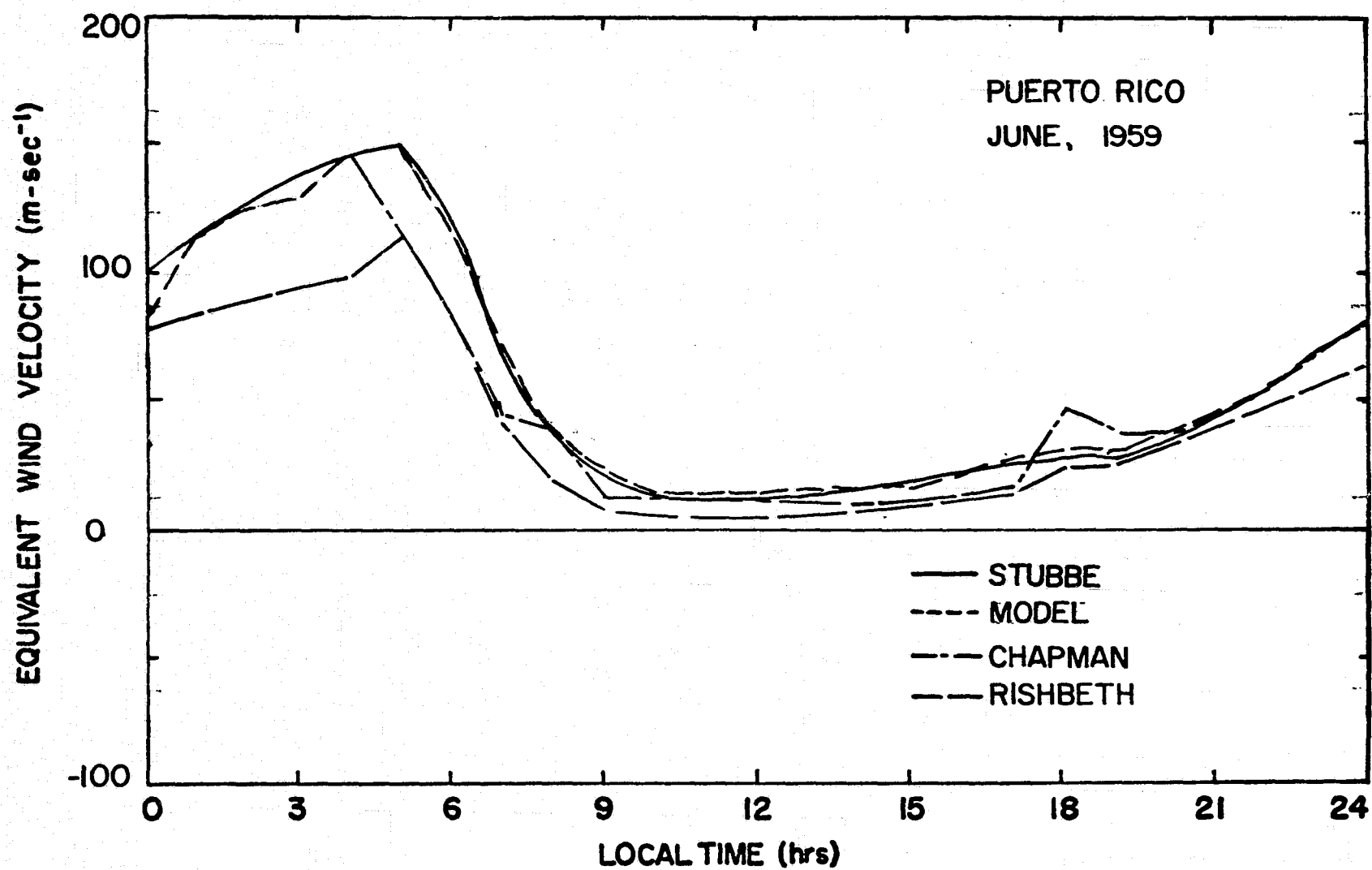


Figure 5. Comparisons of Methods for Estimating the Equivalent Magnetic Meridional Wind for Summer Conditions at Puerto Rico

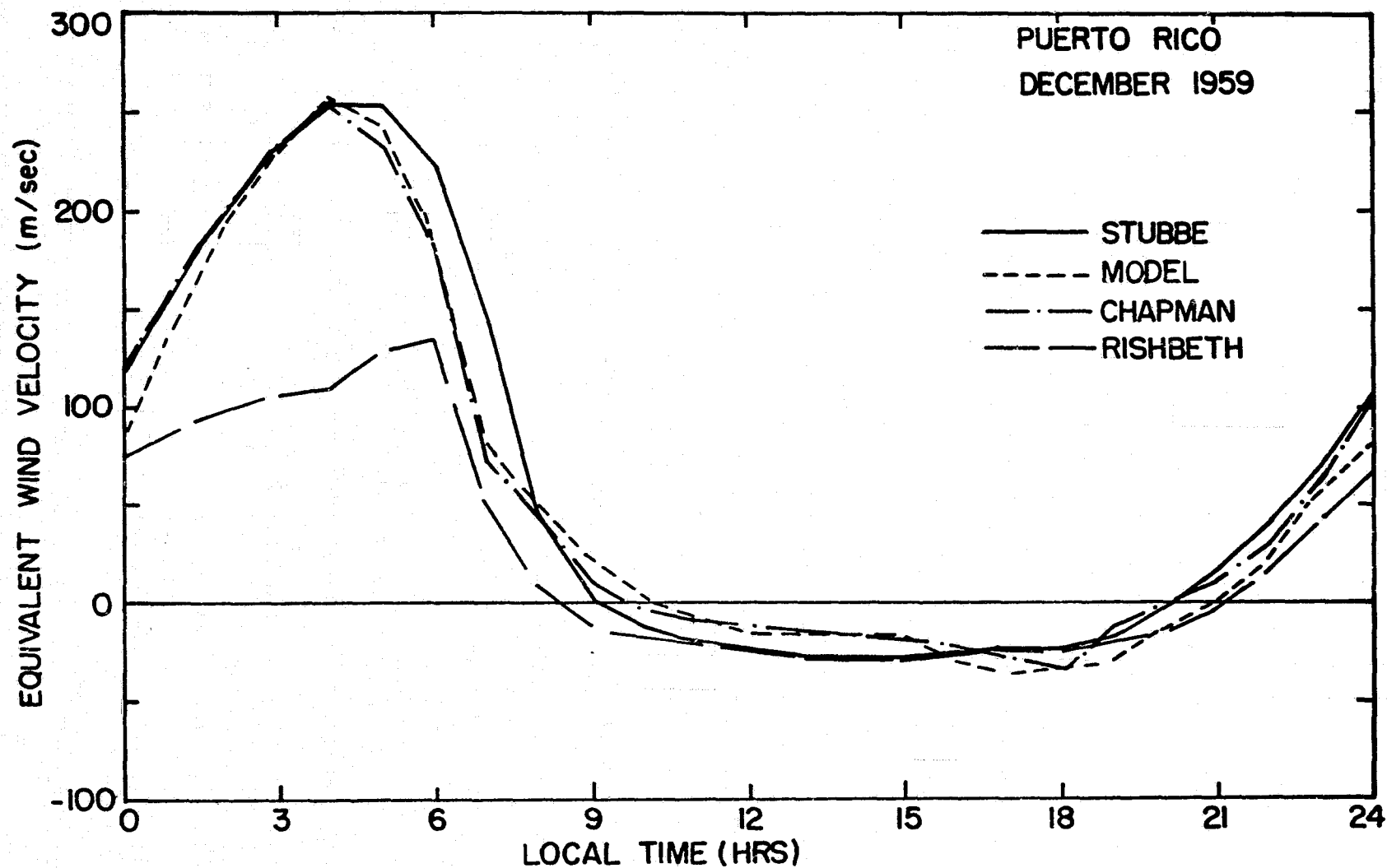


Figure 6. Comparisons of Methods for Estimating the Equivalent Magnetic Meridional Wind for Winter Conditions at Puerto Rico

To evaluate these results quantitatively, the root mean square errors for the results in Figures 3 through 6 are shown in Table 1. This shows that the "model method" reproduces the theoretical equivalent wind most accurately.

Table 1
Evaluation of the Various Methods for
Calculating the Equivalent Meridional Wind

Station	Time	Root Mean Square Error (m-sec^{-1})		
		Rishbeth	Chapman	Model
Adak	6/59	31.5	17.9	9.8
Adak	12/59	33.6	16.4	16.7
Puerto Rico	6/59	21.1	13.2	3.1
Puerto Rico	12/59	61.2	19.7	20.1
Average		36.9	16.8	12.4

It should be pointed out that the time differentials used in the "model method" equivalent meridional wind calculations have increments of one hour. As a result of this, the error can increase at times of rapid variation of the input parameters $HmF2$ and/or $NmF2$. As evidence of this note the sudden increase in the equivalent meridional wind velocity at 700 hours in Figure 4. This sudden increase coincides with a 500 percent rise in the value of $NmF2$ from 700 to 800 hours.

Because the vertical drift calculations are part of the equivalent meridional wind calculations and because errors in the diffusion velocity calculations are small, the comparisons of the methods presented above also refer to the vertical drift calculations.

2.6 Relative Significance of the Terms which Contribute to the Equivalent Meridional Wind Determination

In Section 2.2 an expression for vertical drift was derived in terms of production, loss, the ion density time derivative and the ion flux at 1000 km (see Equation 2.7). An expression for the equivalent meridional wind was then derived in terms of the vertical drift velocity and the diffusion velocity (see Section 2.3). It was shown in Section 2.5 that, by applying this equivalent meridional wind expression and HmF2 and NmF2 data together with an expression which suitably approximates the ion density profile, equivalent meridional winds calculated at the F2 peak agreed rather well with those calculations using the much more complicated Stubbe ionospheric program.

In order to determine the range and conditions over which the approximations used above are valid, the contributions from individual terms will be discussed in this section. It should be noted that except for results shown in Figures 7 and 11 all vertical drifts, equivalent meridional winds and their component terms are calculated at the F2 peak. This is done because, in processing large quantities of data, these parameters are simplest to use and are less affected by underlying ionization errors than densities at lower heights. The results given in Figures 7 and 11 represent a special case in which the entire electron density profile was available by means of the Millstone incoherent scatter radar system.

In Figure 7 the contributions from the production, loss, time derivative and 1000 km flux terms are presented as a function of altitude. It can be seen that at lower altitudes the contributions from the production and loss terms are dominant but that these

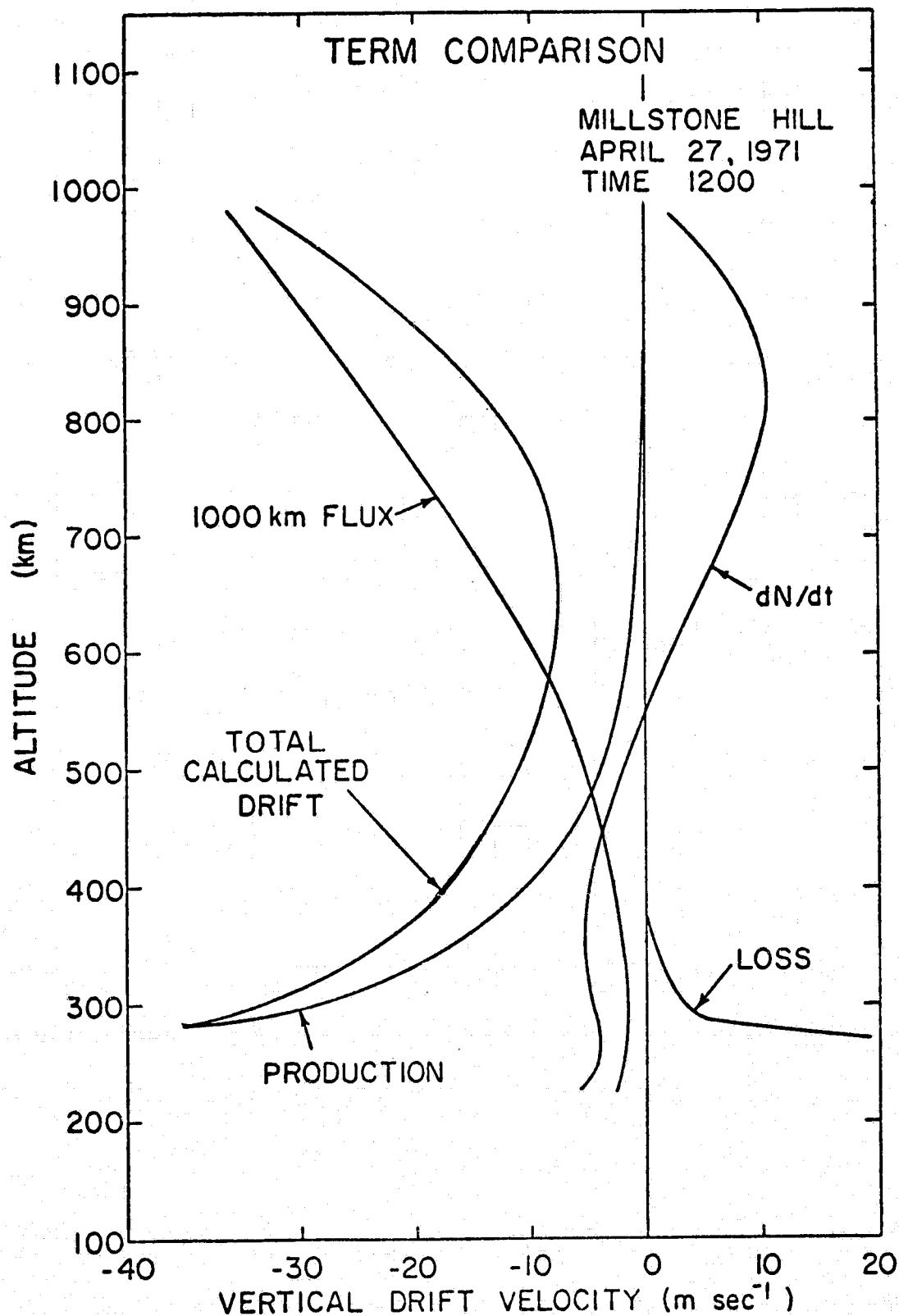


Figure 7. Altitude Dependence of the Terms Entering into the Vertical Drift Velocity Calculation

contributions decrease exponentially with altitude. On the other hand, the 1000 km flux makes little contribution at the F2 peak but increases exponentially to dominate at higher altitudes. Because of this it has been assumed that the flux contribution is negligible for the wind calculations. The time derivative term does not appear to vary significantly with altitude.

The diffusion velocity term is a function of gravity, the ion density, the ion and electron temperature height derivatives, and the ion collision frequency (see Equation 2.12). If the diffusion velocity is calculated at the F2 peak, the ion density height derivative is zero and the temperature height derivatives are small. The diffusion velocity then becomes a function of gravity and the ion collision frequency.

At the F2 peak the contributions from the production, loss and time derivative terms as a function of time are presented in Figure 8. As is to be expected the production term is significant only during the day. Because the F2 layer typically rises at night to an altitude at which little loss occurs, the loss term is also large only during the daytime. The time derivative term is most important at sunrise and sunset, the times of most rapid change.

The contributions of the vertical velocities and diffusion velocities to the equivalent winds are presented as a function of time in Figure 9. As in Figure 8 these values were calculated at the F2 peak. From Figure 9 it can be seen that the diffusion velocity is dominant at night whereas the vertical drift dominates during the day. While this is true in general, it can be shown that the times at which a given term dominates can vary greatly with season and location.

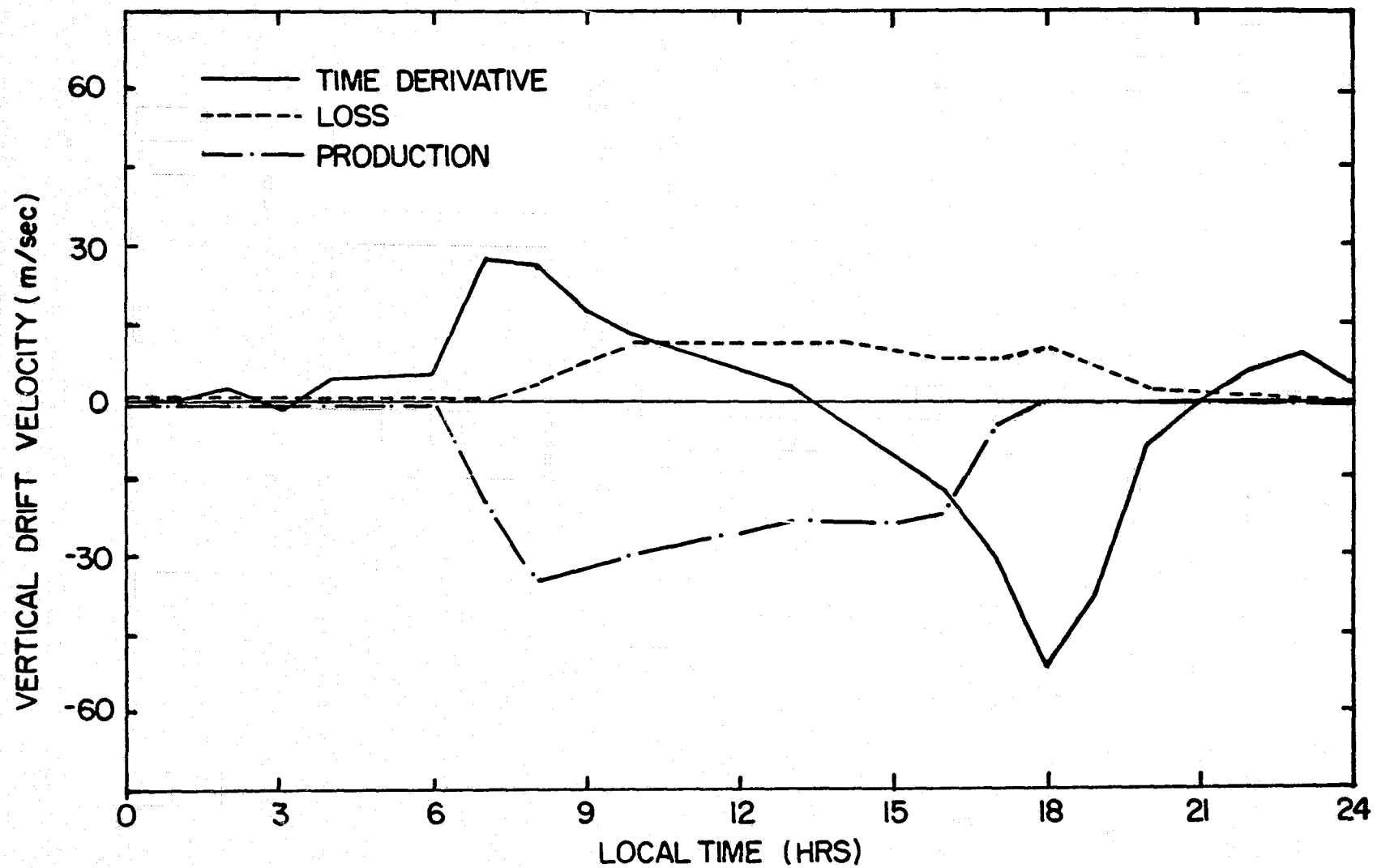


Figure 8. Diurnal Variation of the Production, Loss and Time Derivative Terms at the F2 Peak

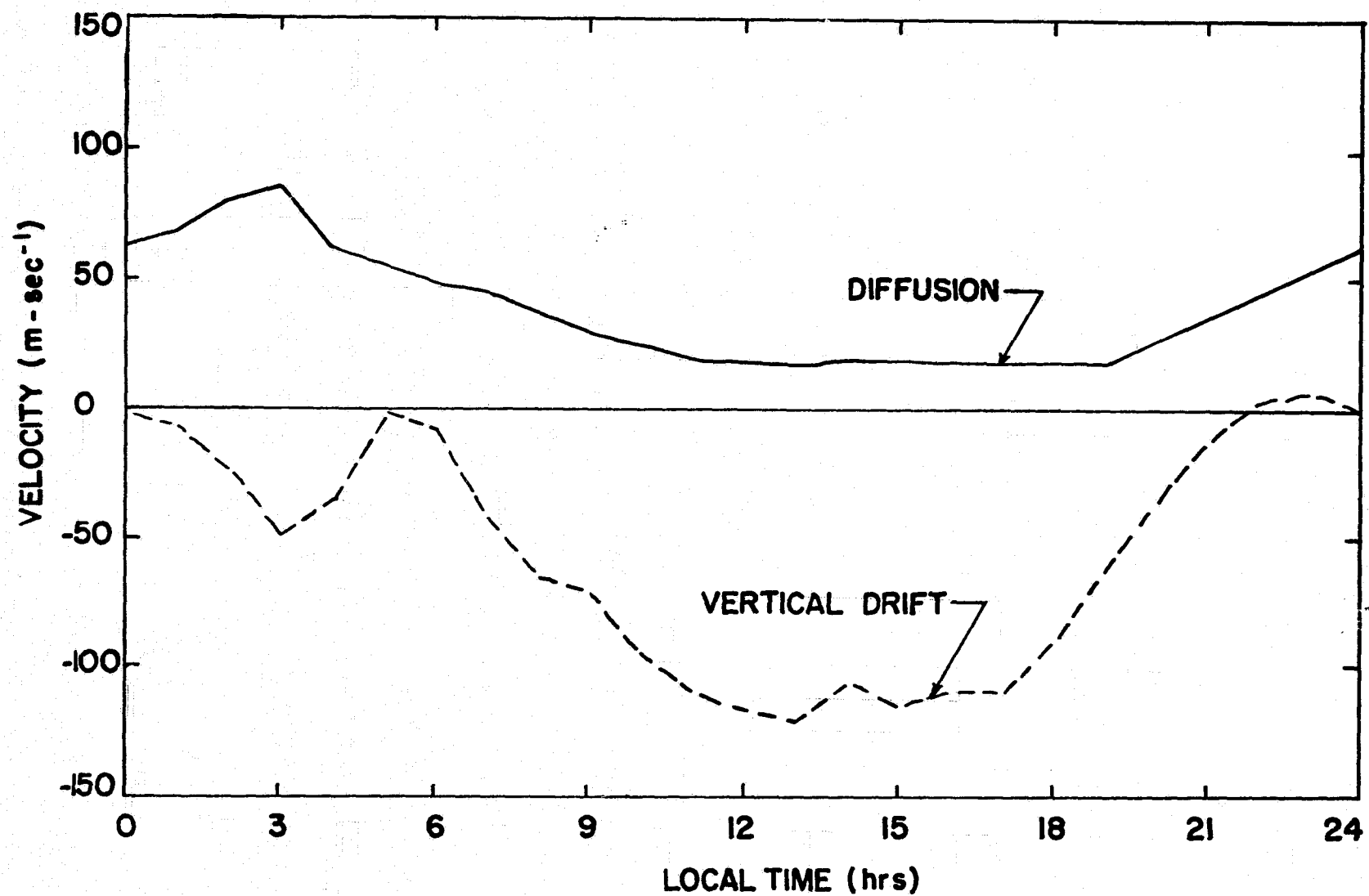


Figure 9. Diurnal Variation of the Vertical Drift Velocity and Diffusion Velocity at the F2 Peak

From the analysis above, it is concluded that for satisfactory equivalent wind calculations the daytime production and loss term and the nighttime time derivative and diffusion terms must be known accurately.

2.7 Discussion of the Errors in the Calculated Equivalent Winds in the Direction of the Magnetic Meridian

In the calculation of equivalent winds from HmF2 and NmF2 data three sources of errors are apparent. The first is inaccuracies in the measured parameters HmF2 and NmF2. It was shown in Section 1.2.9 that the error in determining HmF2 is of the order of 5 km whereas the error in determining NmF2 is negligible. To determine the effect of the errors in the peak height on equivalent wind calculations, random errors were introduced into the calculations. The results showed that a 5 km error in peak height produced an error of the order of 5 m-sec^{-1} in the equivalent wind.

The second source of error results from the approximations made in the calculation procedure itself such as the shape of the profile, the substitution of the peak electron density parameters for peak ion density parameters, etc. Errors introduced by these approximations were already discussed in Section 2.5 where it was shown that the average error for the "model method" was 12.4 m-sec^{-1} .

The third error results when the equivalent winds are applied to additional F2 region calculations. It was noted in the previous sections that the calculated equivalent winds were most accurate at the F2 peak because the parameters which were used in the estimation of the density profiles were supplied only at the peak. This is not

true when the entire density profile is available as from incoherent scatter measurements. For some calculations, presented in Chapter III, it is necessary to assume that the equivalent wind is height independent in the vicinity of the F2 peak. In order to test the accuracy of this assumption, equivalent winds above and below the F2 peak were calculated by the Stubbe program and these were compared with the equivalent winds calculated at the peak. It was shown that the assumption introduced errors of 7 m-sec^{-1} for differences of 50 km from the F2 peak. It should be noted that equivalent winds calculated by the "model method" can introduce errors of the order of 14 m-sec^{-1} at height differences of 50 km from the F2 peak.

In the error calculations and comparisons discussed above, the errors presented are root mean square values. However, the individual errors, which do occur, are very dependent upon location, altitude, time of day, etc., and as a result possess systematic as well as random characteristics. For the purpose of determining the total equivalent wind error, it will however be assumed that the individual errors are random and uncorrelated. The total error is then of the order of 15 m-sec^{-1} on the average and varies from 10 m-sec^{-1} to 20 m-sec^{-1} .

2.8 Comparison with Incoherent Scatter Drift Measurements

As was shown in Section 2.3 the vertical drift calculation is intermediate to the determination of the equivalent meridional wind. In this section calculated vertical drifts are compared with vertical drifts measured at the Millstone Hill and CNET incoherent scatter facilities.

In Figure 10 the calculated drifts are compared with the F2 peak vertical drifts measured at Millstone. The agreement is very good except for a three-hour period in the morning when the calculated and measured quantities vary in an opposite sense. An examination of the input data showed that the peak heights at the time of this discrepancy were abnormally low. In the previous section it was shown that the contributions to the vertical velocity drift calculation by the production and loss terms increased very rapidly with decreasing altitude. It was implied that inaccuracies in these terms at low altitudes would cause serious discrepancies in the vertical drift calculation. In order to test the possibility that this was the case, vertical drifts were calculated at 300 km and these were compared with drifts measured at 300 km. In this instance it was possible to perform these calculations without introducing density profile errors because the Millstone Hill data provides the entire profile. The results of this comparison are presented in Figure 11. It can be clearly seen that the agreement is greatly improved. This appears to verify the implication made about errors for low altitude vertical drift calculations.

A comparison of calculated drifts and drifts measured at CNET is shown in Figure 12. The drifts compared in this figure refer to the F2 peak. The agreement is of the order of 5 m-sec^{-1} rms with the most serious discrepancies of up to 27 m-sec^{-1} occurring in the late morning period. It should be noted that the CNET drifts are obtained by resolving drifts measured parallel to the magnetic field lines into the vertical direction. Therefore, drifts perpendicular to the magnetic field lines, as can be caused by an East-West electric field,

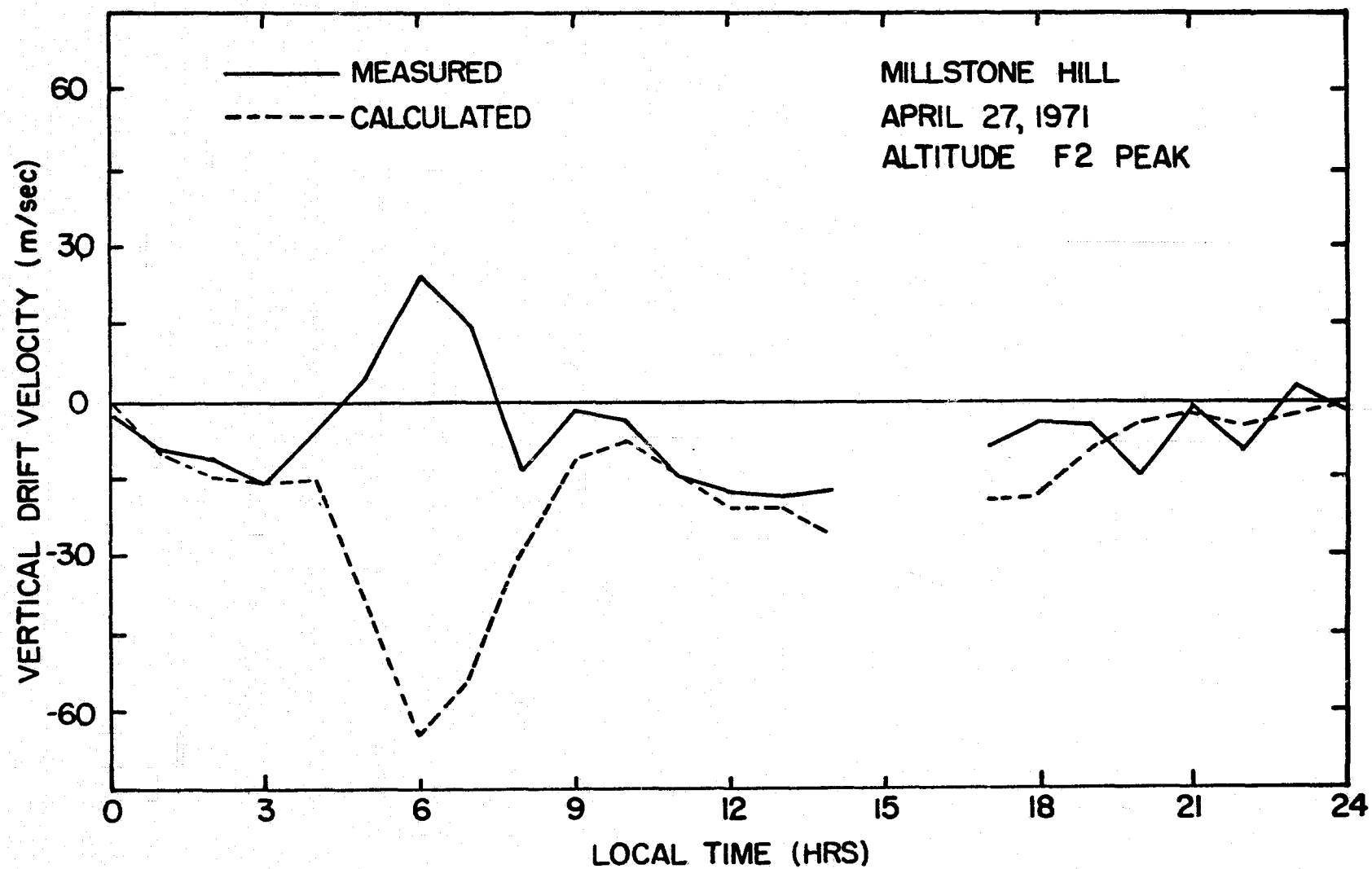


Figure 10. Comparison of Vertical Drifts Calculated at the F2 Peak with F2 Peak Vertical Drifts Measured by the Incoherent Scatter Radar at Millstone Hill

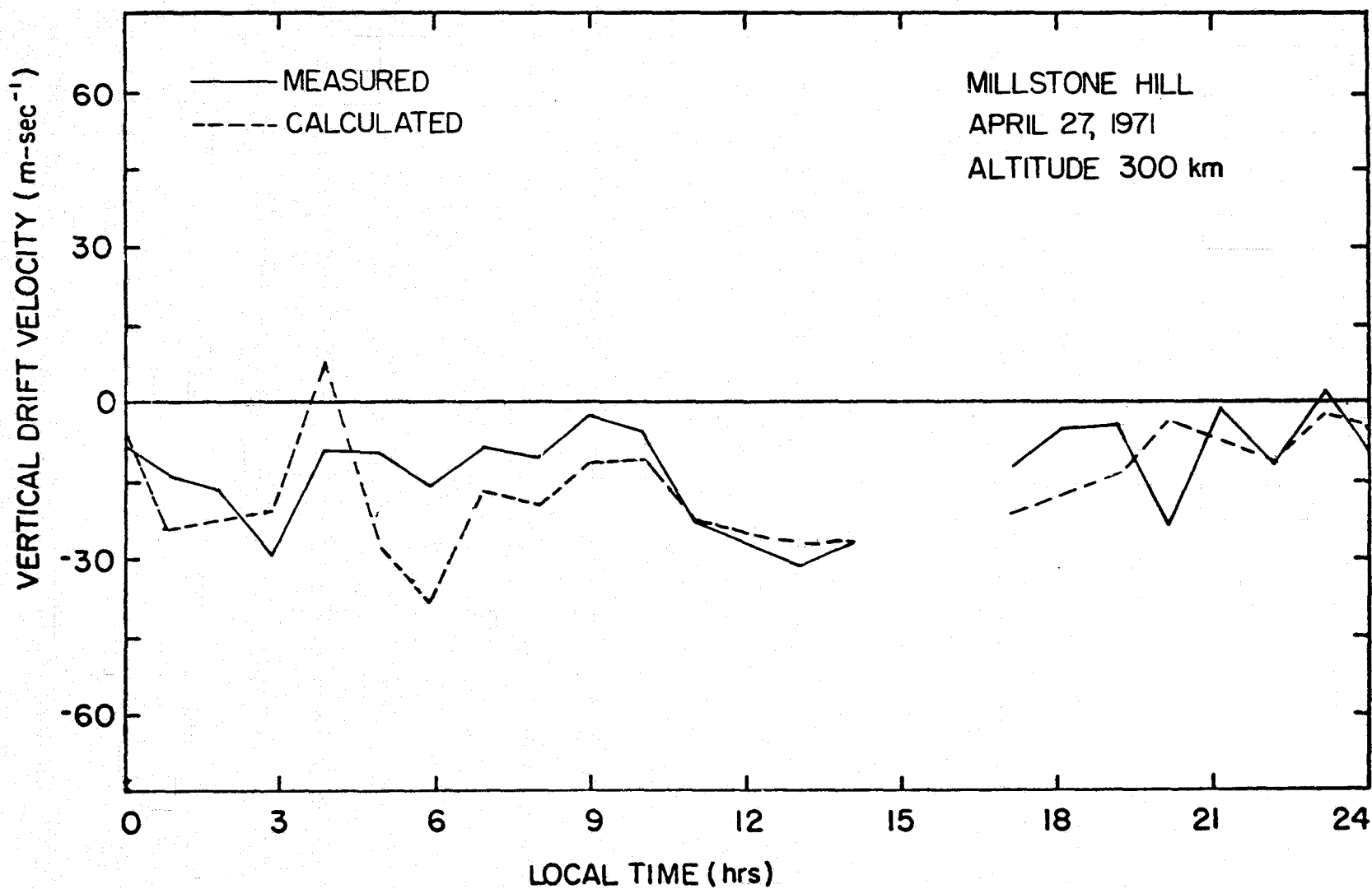


Figure 11. Comparison of Vertical Drifts Calculated at 300 km with Vertical Drifts Measured at 300 km at Millstone Hill

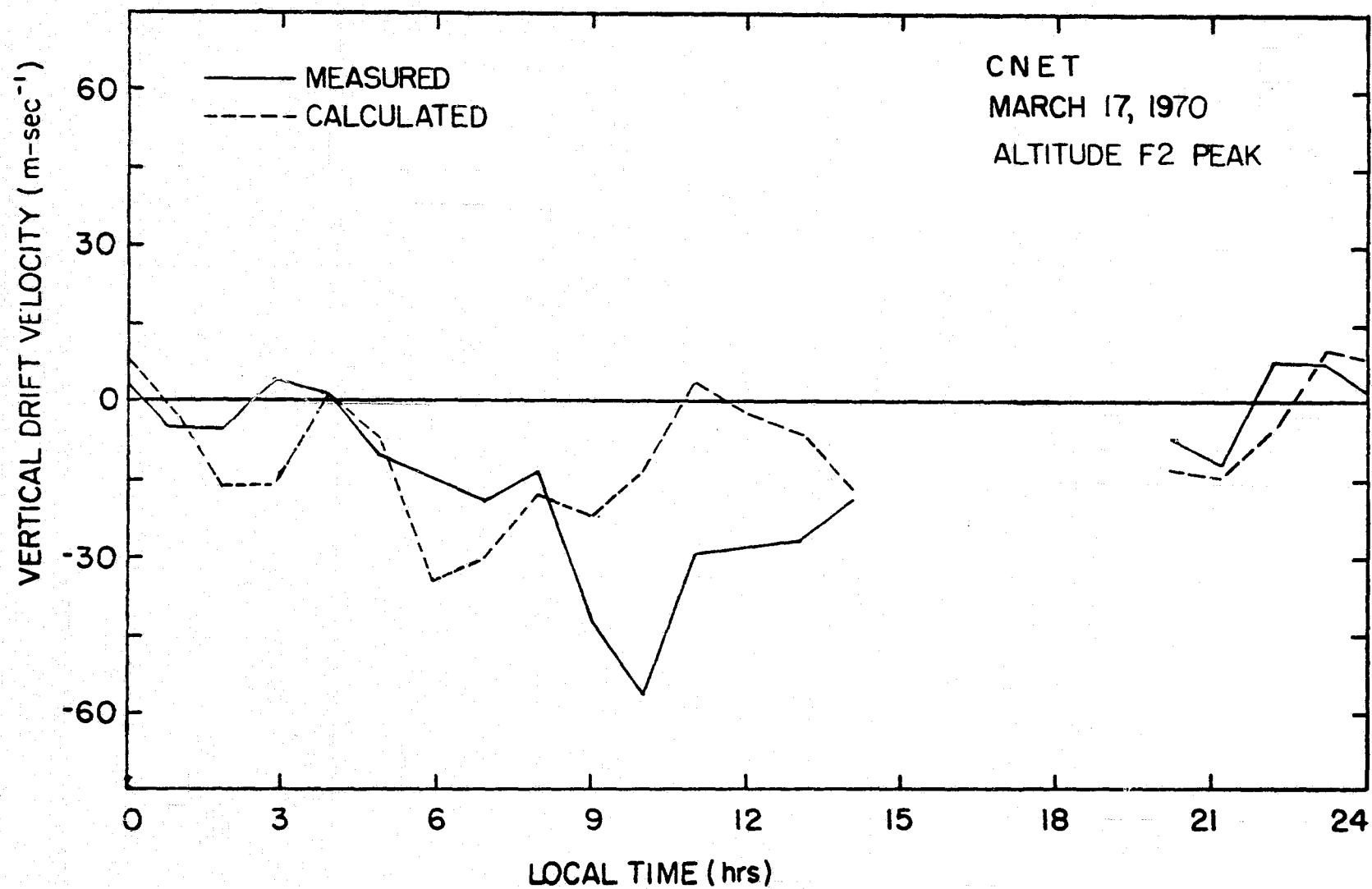


Figure 12. Comparison of Vertical Drifts Calculated at the F2 Peak with F2 Peak Vertical Drifts Measured by the Incoherent Scatter Radar at CNET

are not accounted for in the vertical drift measurements. For the magnetic field configuration at CNET, a 1 mV/m East-West electric field can induce a 10 m-sec^{-1} vertical drift. Since electric fields of this order of magnitude are now measured at CNET, it is possible that some of the discrepancies in the vertical drift comparison are attributable to these electric fields. From the comparisons presented above, it can be seen that for the most part calculated and measured results agree within about 5 m-sec^{-1} but that care must be taken when the layer is low and when the electron density is changing rapidly.

2.9 Discussion and Summary of the Results

In Section 2.3 an expression for determining the equivalent wind in the magnetic meridional direction was derived. It was shown that proper use of this expression requires that the production, loss and time derivative terms of the ion continuity equation be known as a function of altitude and that the diffusion term be known at the altitude at which the wind calculation is being made. All of the parameters necessary for the calculation of these terms can be supplied by neutral atmospheric models except for the ion density profile which must be supplied from measurements. Since ion density profile measurements above the peak are not very abundant, but measurements of the peak parameters H_mF2 and N_mF2 are, methods were derived for approximating the ion density profile from these peak parameters. The two methods, "model" and Chapman, derived for this purpose were compared by employing theoretical data calculated by means of the Stubbe ionospheric program. The "model method" produced better agreement with the theoretical equivalent winds than did the Chapman method, and both methods were far

superior to a method developed by Rishbeth. For the conditions tested the average root mean square error for the "model method" was 12.4 m-sec^{-1} .

When employing such approximation schemes, the most accurate equivalent winds and vertical drift calculations are made at the altitude at which the measured quantities are supplied, i.e., the F2 peak. However, this is only true for theoretical cases or for experimental results which are adequately described by the neutral model. It was shown in Section 2.6 that the magnitudes of the production and loss terms increase with decreasing altitude. If the neutral model is inaccurate, and the peak height is at a low altitude, the terms, such as production and loss, which enter into the vertical drift and equivalent wind calculations are larger and substantial errors can occur. This was shown in the comparison of measured and calculated vertical drifts for the Millstone Hill data (see Figures 10 and 11).

With respect to the problem of neutral model accuracies, it is noted that the static models employed in this study are most accurate in the altitude range of the F2 peak. The reason for this is that most of the data which were used in the construction of these models were measured in this altitude range. In order to extrapolate these measurements down to lower altitudes assumptions about the lower boundary conditions were made. The accuracy of the lower boundary conditions was improved by Stubbe (1973), but the accuracy of the lower boundary is still not adequate. Because in the solution of the theoretical ionospheric models, the conditions occurring in the lower atmosphere are propagated up, the entire neutral atmospheric model must be accurate if reasonable theoretical results are to be obtained.

If large amounts of data are to be employed for the purpose of determining the equivalent winds, economical factors may have to be considered. On an IBM model 370-168 computer, the "model method" takes approximately five times as much time as the Chapman or Rishbeth methods. However, the iterative approach as described by Stubbe (1973) would take at least six times as much time as the "model method." Note that the economy of Stubbe's iterative technique is greatly increased if the investigator can make accurate estimates.

CHAPTER III

ZONAL AND MERIDIONAL NEUTRAL WINDS AND NORMAL ELECTRIC FIELD CALCULATIONS

3.1 Procedure

In Chapter II it was shown how the diurnal variation in the electron density profile at one station can be used to derive the equivalent horizontal neutral winds in the direction of the magnetic meridian and how these are related to the actual winds and electric fields. In this chapter the problem of taking these equivalent winds for a number of stations and using them to calculate winds in the geographical zonal and meridional directions and the normal electric fields is investigated.

3.2 Station Selection

In this study the equivalent winds in the direction of the magnetic meridian are first calculated for each station. Because of the non-uniformity of the earth's magnetic field, stations are available with different declinations; and if a suitable distribution of stations is available, both the zonal and meridional components of the neutral wind can be determined. The magnetic inclination is also a function of latitude and this can be used to provide information on the electric fields if suitable stations are available. During IGY and IQSY there were approximately 150 ionosonde stations which made ionograms on a routine basis; however, only a minute fraction of these data have been reduced to electron density height profiles. A large project is underway to reduce more of the data but this has not been completed. The present selection of stations is based on the

availability of the data and an attempt will be made to demonstrate the feasibility and capability of the techniques so that they will be available for use with the complete data set. A list of the stations and their descriptions is given in Table 2. All data used in this study represent average monthly data for 1959, a year of very high solar activity.

3.3 Method of Calculating the Zonal and Meridional Winds

It may be assumed that the zonal and meridional winds can be expressed in terms of spherical harmonics. Since only a limited data base is available for the present study, a simplified set of equations has been assumed for the neutral wind calculations. Each component is represented by a dc. term and the first three harmonics of the diurnal variation and each term has been assumed to vary linearly with geographic latitude. This gives

$$U_x = \sum_{n=0}^3 (A_{xn} + B_{xn}\phi) \cos n\omega t + (C_{xn} + D_{xn}\phi) \sin n\omega t \quad (3.1)$$

$$U_y = \sum_{n=0}^3 (A_{yn} + B_{yn}\phi) \cos n\omega t + (C_{yn} + D_{yn}\phi) \sin n\omega t \quad (3.2)$$

where

ϕ refers to the latitude

These equations were solved for the 28 unknowns by using a least squares technique with the hourly values from the available stations.

It is difficult to obtain estimates of the errors associated with the solution of a large matrix by least square techniques and their relation to the station parameters. It was therefore decided

Table 2
Description of Ionosonde Stations

Station	Latitude (+ North)	Longitude (+ East)	Declination (+ East)	Inclination (+ Downwards)	Magnetic Intensity (weber-m ⁻² x 10 ⁴)
Adak	51.9	183.4	9.6	63.8	.427
St. John's	47.6	307.3	-26.5	71.9	.467
Fort Monmouth	40.4	285.9	- 9.9	71.5	.489
White Sands	32.3	253.5	11.8	60.0	.445
Grand Bahama I	26.6	281.8	- .8	59.8	.435
Okinawa	26.3	127.8	- 2.4	36.9	.379
Puerto Rico	18.5	292.8	- 6.8	51.9	.383

to study the errors which would result in situations where two stations with the same geographic latitude but different declinations are available and where the electric field effects can be neglected. Under these circumstances Equations 3.1 and 3.2 become

$$U_{M1} = U_x \cos D_1 + U_y \sin D_1 \quad (3.3)$$

$$U_{M2} = U_x \cos D_2 + U_y \sin D_2 \quad (3.4)$$

The solution of Equations 3.3 and 3.4 yields the horizontal neutral wind components

$$U_x = \frac{U_{M1} \sin D_2 - U_{M2} \sin D_1}{\sin (D_2 - D_1)} \quad (3.5)$$

$$U_y = \frac{U_{M1} \cos D_2 - U_{M2} \cos D_1}{\sin (D_2 - D_1)} \quad (3.6)$$

In order to determine the accuracy of the neutral wind components as a function of the declination angles, the following assumptions are made:

- 1) Equivalent meridional winds are random variables that have means U_{M1} and U_{M2} and variances $\sigma_{M1}^2 = \sigma_{M2}^2 = \sigma^2$.
- 2) The random variables are not correlated, i.e., $\rho_M = 0$.

It is recognized that the conditions for fulfilling the above assumptions are probably never completely met. The purpose for making such simplified assumptions is to obtain an order of magnitude estimate of the errors. From the above assumptions the deviations for the horizontal neutral wind components are

$$\sigma_{U_x} = \sigma \frac{\sqrt{\sin^2 D_1 + \sin^2 D_2}}{\sin (D_2 - D_1)} \quad (3.7)$$

$$\sigma_{U_y} = \sigma \frac{\sqrt{\cos^2 D_1 + \cos^2 D_2}}{\sin (D_2 - D_1)} \quad (3.8)$$

Equations 3.7 and 3.8 were applied to the declinations of the available station pairs for this study and the results are presented in Table 3.

Table 3
Accuracies of the Horizontal Neutral Wind
Components Derived from Pairs of Ionosonde Stations

Station Pair	Latitudes	Declinations	m-sec ⁻¹	
			σ_{U_x}	σ_{U_y}
Adak/St. John's	51.9/47.6	9.6/-26.5	12.1	33.9
St. John's/Ft. Monmouth	47.6/40.4	-26.5/- 9.9	25.0	69.7
Ft. Monmouth/White Sands	40.4/32.3	- 9.9/ 11.8	10.8	56.2
White Sands/Grand Bahama I	32.3/26.6	11.8/- .8	14.1	96.1
Grand Bahama I/Okinawa	26.6/26.8	- .8/- 2.4	23.7	75.43
Okinawa/Puerto Rico	26.3/18.5	- 2.4/- 6.8	24.6	275.4

From Table 3 it is seen that the uncertainties in determining the North-South component of the neutral winds are of the order of 10-25 m-sec⁻¹ whereas the uncertainties in determining the West-East neutral wind component range from approximately 34 m-sec⁻¹ to more than 100 m-sec⁻¹. The uncertainties in determining the West-East component are large because there is not much variation in the declinations of the

station pairs available for this study especially in the lower latitude regions. Estimates of the accuracies that could be attained if the reduced ionosonde data from the entire network of ionosonde stations were available are presented in Table 4.

Table 4

Estimate of the Accuracies of the Horizontal Neutral
Wind Components from the Entire Ionogram Data Set

Region	m - sec ⁻¹	
	σ_{U_x}	σ_{U_y}
High Northern Latitudes	10-20	15-25
Middle Northern Latitudes	5-15	25-35
Low Latitudes	10-20	40-50
Middle Southern Latitudes	10-20	25-35
High Southern Latitudes	20-30	20-30

3.4 Calculations of the Horizontal Neutral Winds

3.4.1 Electric Field Effects

As was shown in Section 2.2, only the vertical ion velocity can be determined and in Equations 2.7, 2.11 and 2.12 relationships were developed between the horizontal winds in the direction of the magnetic meridian, the electric field normal to the magnetic field, and the shape of the F layer. For convenience up to this point, equivalent horizontal wind velocities have been calculated to account for both effects and in this section the electric field component will be considered. To examine the effects of electric fields the

theoretical model of Matsushita (1971) and the Millstone Hill measurements of Kirchhoff and Carpenter (1974) are used. In the Matsushita study two mechanisms for the generation of electric fields are given: the E region dynamo and the magnetospheric dynamo. Electric fields which result solely from E region dynamo action are referred to as electrostatic fields while those resulting from the combined action of the E region and magnetospheric dynamos are referred to as the total fields. Because these two models provide a measure of current uncertainty in the absolute values of the electric field, the effects that each of these field models have on the neutral wind calculations are considered.

In Figures 13 and 14 comparisons are made between neutral winds which were calculated by neglecting the electric field effects and those calculated by including the theoretical fields given by Matsushita (1971) and the measured fields given by Kirchhoff and Carpenter (1974). From these figures it can be seen that the neglect of the total field can cause uncertainties of the order of 50 m-sec^{-1} in the neutral wind calculations for either component whereas the neglect of the electrostatic field can produce uncertainties of the order of 40 m-sec^{-1} for the North-South component and 25 m-sec^{-1} for the West-East component. On the other hand, the neglect of the measured field produces inaccuracies of the order of 15 m-sec^{-1} and 25 m-sec^{-1} for the North-South and West-East components respectively. On comparing the differences between the results calculated by employing Matsushita's (1971) theoretical total field and the measured fields of Kirchhoff and Carpenter (1974), it can be estimated that errors due to

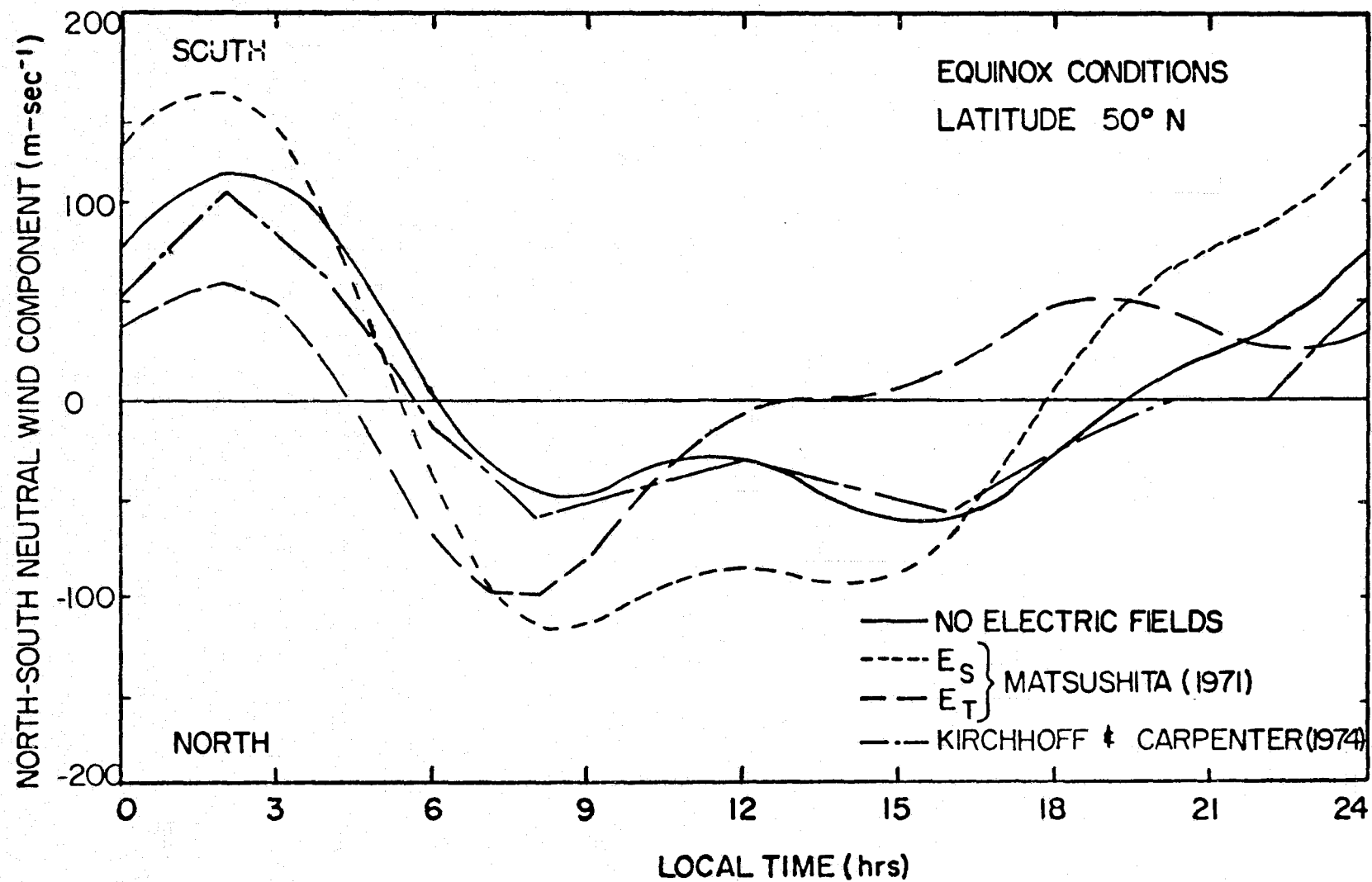


Figure 13. The Effect of Electric Fields on the Calculation of the North-South Component of the Neutral Wind at 50°N Latitude

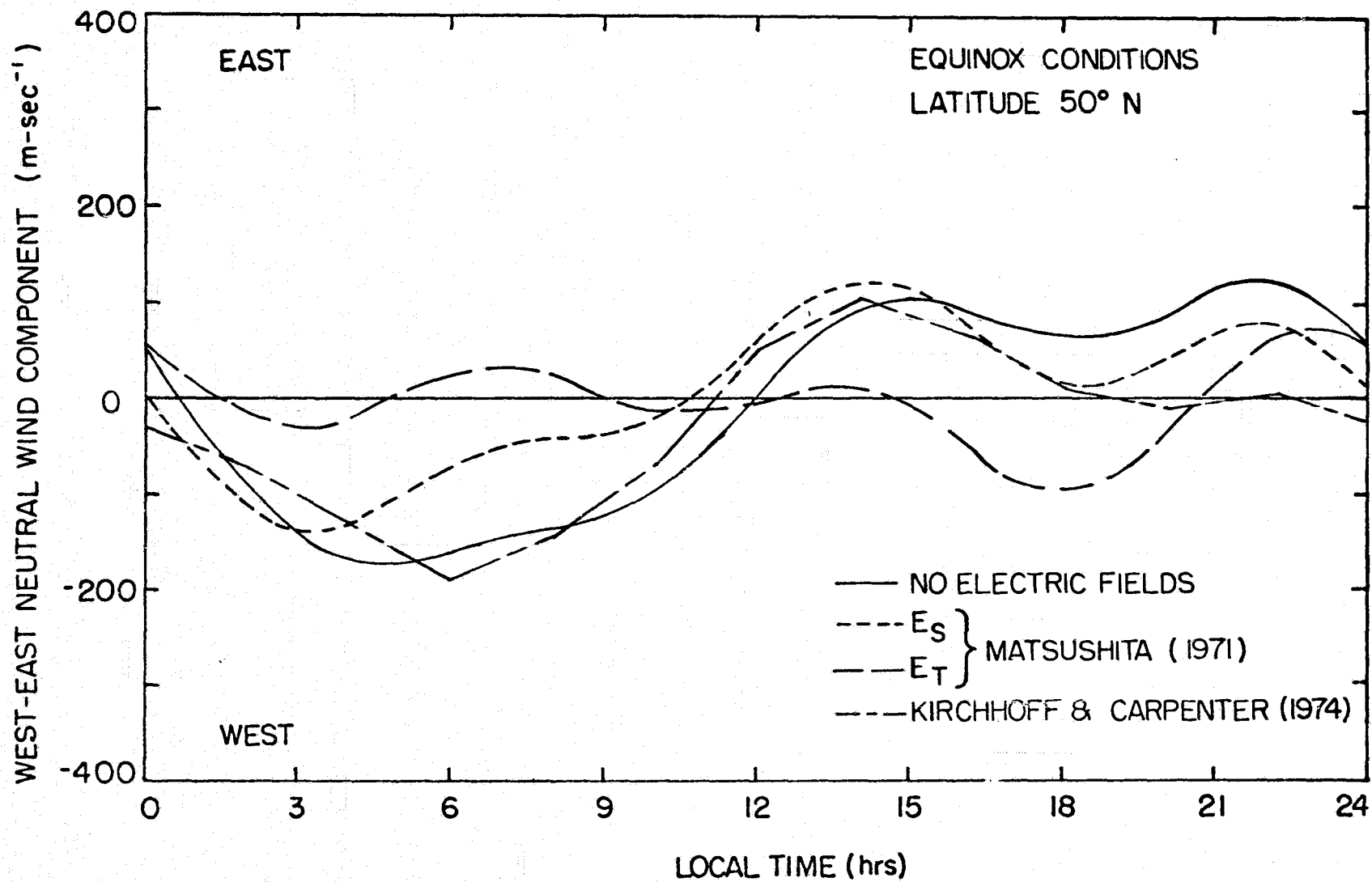


Figure 14. The Effect of Electric Fields on the Calculation of the West-East Component of the Neutral Wind at 50°N Latitude

the inaccuracies of the electric fields themselves are of the order of 30 m-sec^{-1} .

3.4.2 Method for Determining Electric Fields Directly from the Data

In Section 3.3 a method was presented for the calculation of the horizontal neutral wind components by neglecting electric field effects and by employing equivalent meridional winds from pairs of stations having different magnetic declinations. Since the equivalent meridional neutral winds are functions of the electric field (see Equation 2.13), it is implicit in the procedures outlined in Section 3.3 that electric fields must be obtained by other means for the proper solution of the wind problem. This, in fact, is what was done in Section 3.4.1.

In this section a method is presented for the direct determination of the electric fields. This method, which is analogous to the method described in Section 3.3, requires equivalent meridional winds from four stations having different magnetic inclinations and declinations. Equation 2.13 is then rewritten in the form

$$\frac{E_x}{B_1} \frac{\sin D_1}{\sin I_1} + \frac{E_y}{B_1} \frac{\cos D_1}{\sin I_1} + U_x \cos D_1 - U_y \sin D_1 = U_{M1} \quad (3.9)$$

$$\frac{E_x}{B_2} \frac{\sin D_2}{\sin I_2} + \frac{E_y}{B_2} \frac{\cos D_2}{\sin I_2} + U_x \cos D_2 - U_y \sin D_2 = U_{M2} \quad (3.10)$$

$$\frac{E_x}{B_3} \frac{\sin D_3}{\sin I_3} + \frac{E_y}{B_3} \frac{\cos D_3}{\sin I_3} + U_x \cos D_3 - U_y \sin D_3 = U_{M3} \quad (3.11)$$

$$\frac{E_x}{B_4} \frac{\sin D_4}{\sin I_4} + \frac{E_y}{B_4} \frac{\cos D_4}{\sin I_4} + U_x \cos D_4 - U_y \sin D_4 = U_{M4} \quad (3.12)$$

The simultaneous solution of Equations 3.9, 3.10, 3.11 and 3.12 yields the horizontal neutral wind components as well as the horizontal electric field components. Though the solution of these equations is straightforward, the results contain too many terms for presentation here.

An error analysis similar to the one used in Section 3.3 was performed and the results are given in Table 5. For the seven stations available to this study, it is possible to generate 35 combinations taking four stations at a time. However, in keeping with our assumption that the latitudinal separation between stations be small, only the combinations shown in Table 5 are discussed. Note that the errors given for the electric field components are in units of mV/m. For the regions defined by these configurations current electric field theories estimate that the West-East electric fields are of the order of -10mV/m to 10mV/m and that the North-South electric fields are approximately three times the West-East fields. When the order of magnitudes of the electric field components is taken into account, it can be seen from Table 5 that the last two station configurations listed could be used to derive reasonable West-East electric fields. These results also indicate that the possibility of making accurate neutral wind and electric field calculations simultaneously is rather poor when only a limited number of stations are available.

It was shown in Sections 1.2.3, 1.2.4, and 1.2.5 that there are many theories and experimental techniques for the description and measurement of electric fields. Though the orders of magnitudes of the various results are in general agreement, the individual results can vary substantially. With this in mind the West-East electric fields calculated from the White Sands/Grand Bahama I/Okinawa/

Table 5

Accuracies of Horizontal Electric Field and Neutral Wind
Components Derived from a Quadruple of Ionosonde Stations

Station Configuration	σ_{U_x} m/sec	σ_{U_y} m/sec	σ_{E_x} mV/m	σ_{E_y} mV/m
Adak/St. John's/Ft. Monmouth/White Sands	282.3	5446.6	234.0	15.1
St. John's/Ft. Monmouth/White Sands/Grand Bahama I	202.3	1070.5	43.7	7.5
Ft. Monmouth/White Sands/Grand Bahama I/Okinawa	111.7	1700.2	67.9	3.0
White Sands/Grand Bahama I/Okinawa/Puerto Rico	39.2	955.8	34.7	1.5

Puerto Rico station configuration are presented in Figures 15 and 16. These stations represent a latitude range from 18.5°N to 32.3°N and the average latitude of the four stations is 25.9°N .

In Figure 15 West-East electric fields calculated by the method outlined in the study are compared with the West-East electrostatic fields calculated by Stening (1973) for winter conditions. It can be seen that the agreement is remarkably good except for the early morning periods where it appears that a two-hour phase shift exists. Also shown in this figure are West-East electric fields derived from incoherent scatter measurements made at Arecibo in February, 1973 (Behnke and Hagfors, 1974). The measured electric fields are smaller than the calculated fields and are opposite in direction at night. It is to be noted that the electric fields derived from this study are for the year 1959, and for a slightly higher latitude than Arecibo and that the Arecibo results are for the year 1973.

In Figure 16 the West-East electric fields are compared for summer conditions. The order of magnitude of the Stening fields and this study's fields is the same but the agreement is not as good as it was for the winter conditions.

For this study the sample of stations available is quite insufficient to make both field and wind measurements. These comparisons are presented to show that electric fields calculated by this method are in general agreement with current electric field theories. For a complete evaluation of this method, more station combinations with suitable magnetic configurations are necessary.

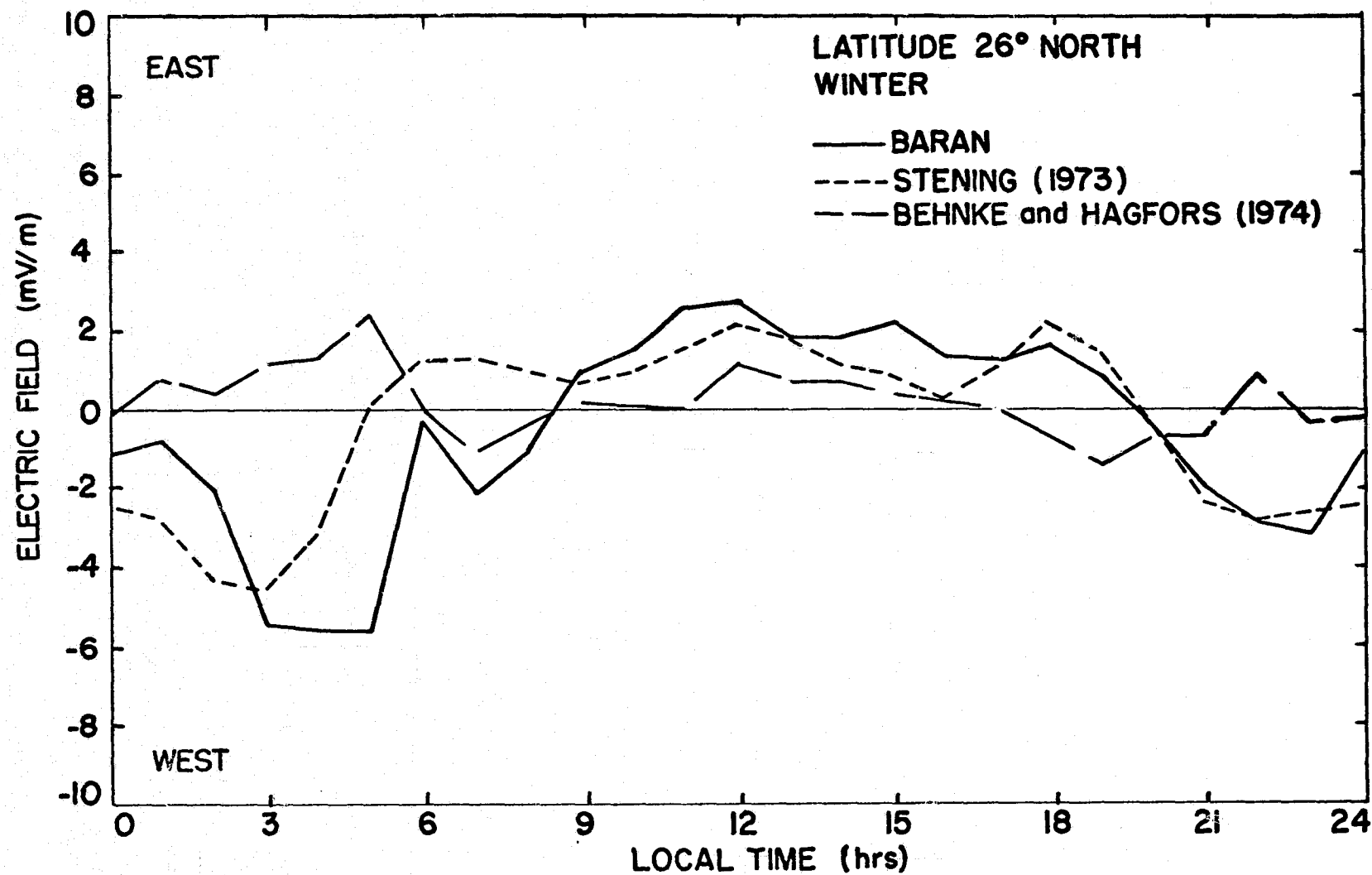


Figure 15. Comparisons of Electric Field Calculations and Measurements for Winter Conditions

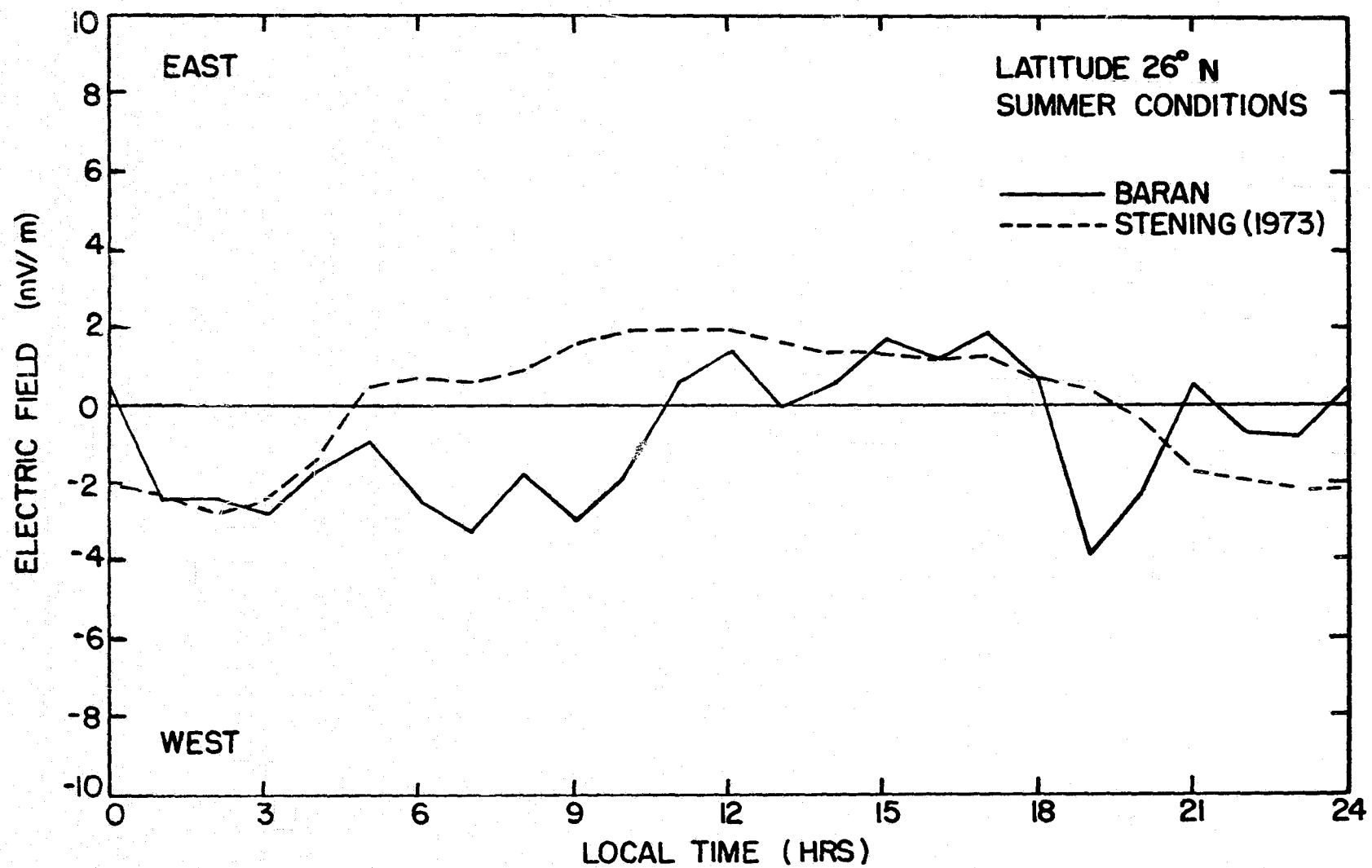


Figure 16. Comparisons of Electric Field Calculations for Summer Conditions

3.4.3 Seasonal-Latitudinal Variations of the Neutral Winds

In the previous sections the effects of electric fields on neutral wind calculations were considered. Theoretically derived fields were shown to cause variations of the order of $40\text{-}50 \text{ m-sec}^{-1}$ and measured fields were shown to cause variations of $15\text{-}25 \text{ m-sec}^{-1}$. Fields calculated directly by methods employed by this study were shown to be in general agreement with the theoretical electrostatic fields of Stening (1973). Because the uncertainties in the theoretical and measured fields are of the order of magnitude of the fields themselves, electric fields are neglected in the following calculations.

In this section neutral winds calculated from June, September and December data from the year 1959 are presented. The solar activity for this time period was relatively high with $S_{10.7} \approx 200$ units. The seasonal variations for the North-South neutral wind component for the latitude of 50°N are shown in Figure 17. The magnitudes of the extrema are seen to be comparable for all seasons but the significance of the various modes shows a seasonal dependence. For all seasons the North-South winds are dominated by the diurnal mode. For equinox and winter the significance of the semi-diurnal and terdiurnal modes is seen to increase with the higher order modes being most significant at equinox.

Seasonal variations for the West-East neutral wind component are shown in Figure 18. Unlike the North-South component, the magnitude of this component is shown to vary greatly with season. The West-East winds are largest in the summer and smallest in the winter. It is apparent that the diurnal, semi-diurnal and terdiurnal modes are present for all seasons and that the diurnal mode is dominant.

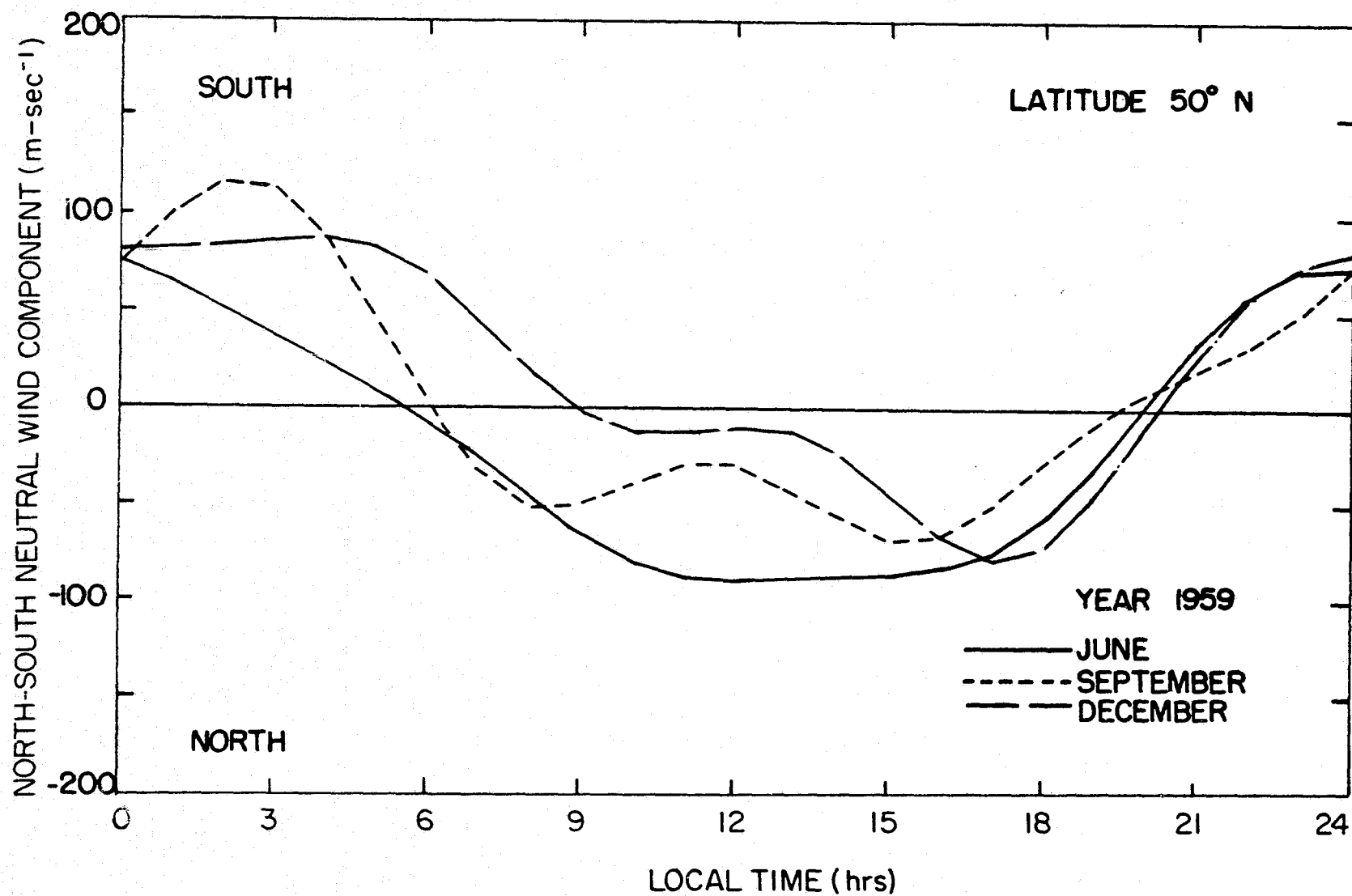


Figure 17. Seasonal Variations of the North-South Neutral Wind Component at 50°N Latitude

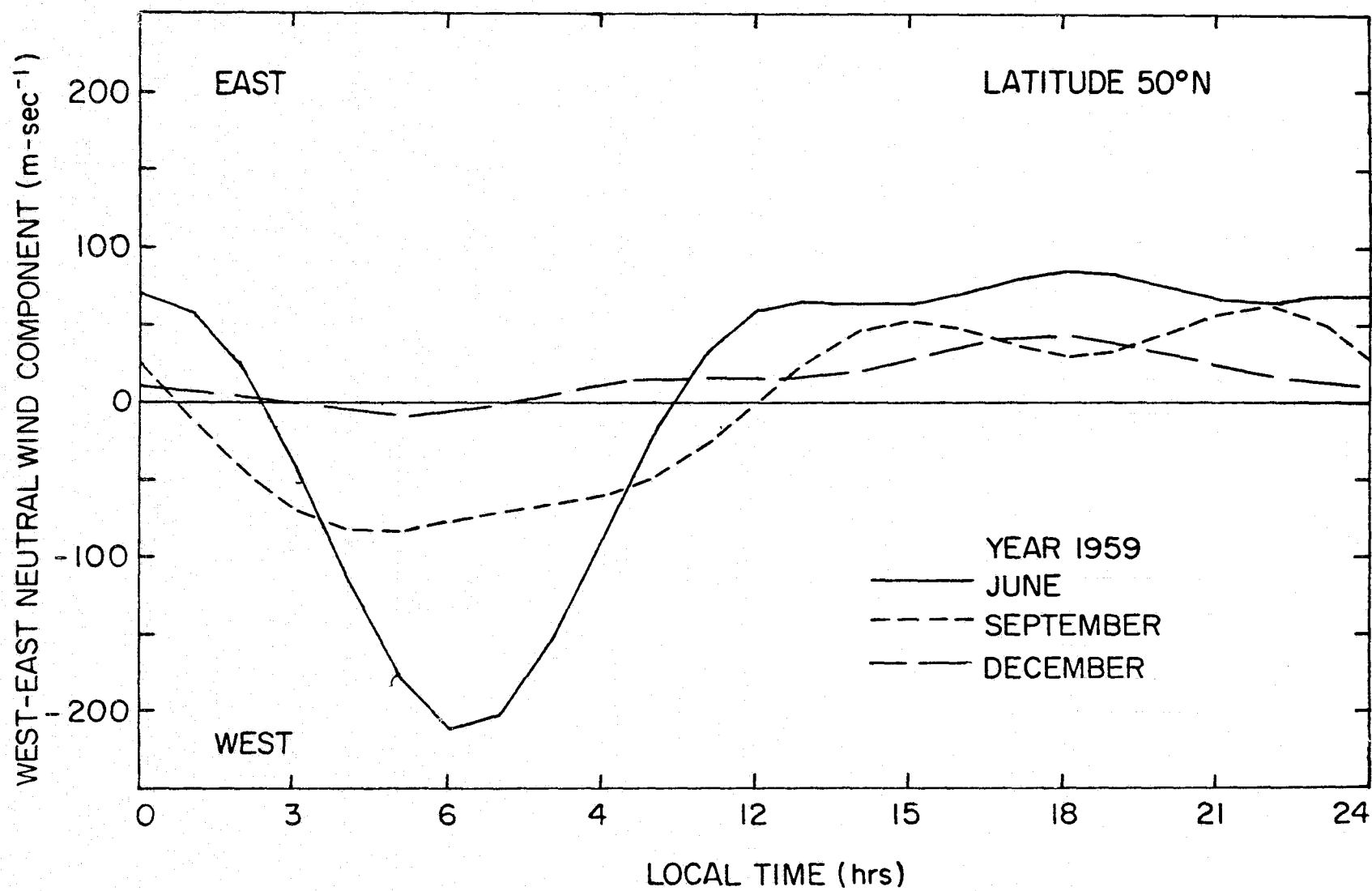


Figure 18. Seasonal Variations of the West-East Neutral Wind Component at 50°N Latitude

However, in contrast to the results presented for the North-South component, the significance of the higher order modes is greatest in summer.

For a latitude of 20°N the seasonal variations of the North-South neutral wind components are presented in Figure 19. For all seasons the magnitude of the diurnal mode is seen to be much less than it is at higher latitudes. In fact, for summer conditions the diurnal mode is practically non-existent and the semi-diurnal mode dominates. For equinox, the diurnal, semi-diurnal and terdiurnal modes are equally significant, but the magnitudes of these modes are small. For winter conditions, the diurnal mode dominates but the semi-diurnal and terdiurnal modes are very significant.

The latitudinal variations of the North-South neutral wind components are more readily seen in Figures 20, 21 and 22 for summer, equinox and winter conditions respectively. For all seasons the diurnal mode is dominant at the high latitudes. As latitude decreases, the magnitude of the diurnal mode is seen to decrease and semi-diurnal and terdiurnal modes become more prominent.

3.5 Comparisons with Contemporary Neutral Wind Calculations

Currently, there are three means by which neutral winds are determined and these are from

- 1) direct velocity measurements such as incoherent scatter and vapor trails, etc.;
- 2) theoretical neutral atmospheric models; and
- 3) theoretical ionospheric models.

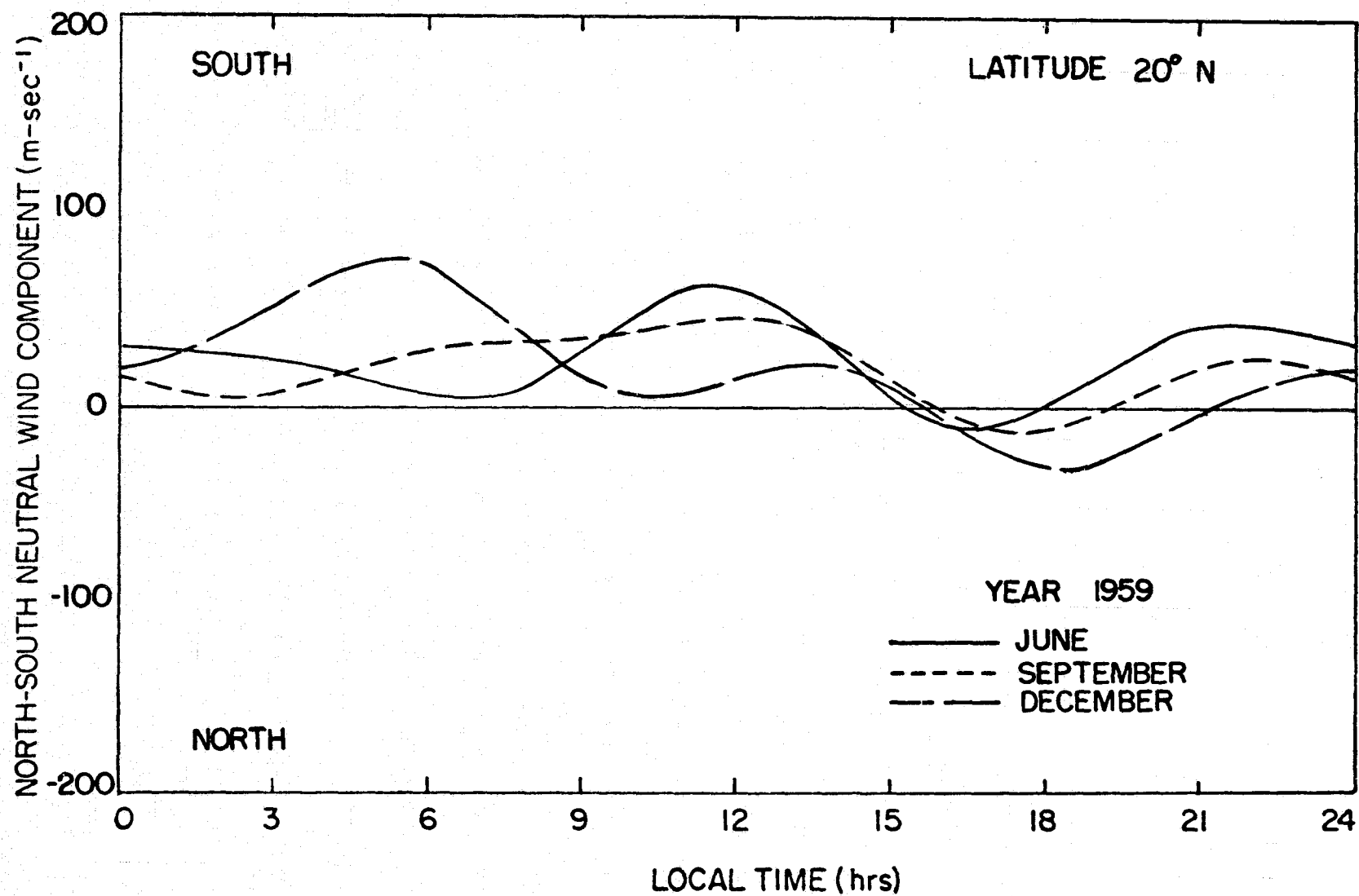


Figure 19. Seasonal Variations of the North-South Neutral Wind Component at 20°N Latitude

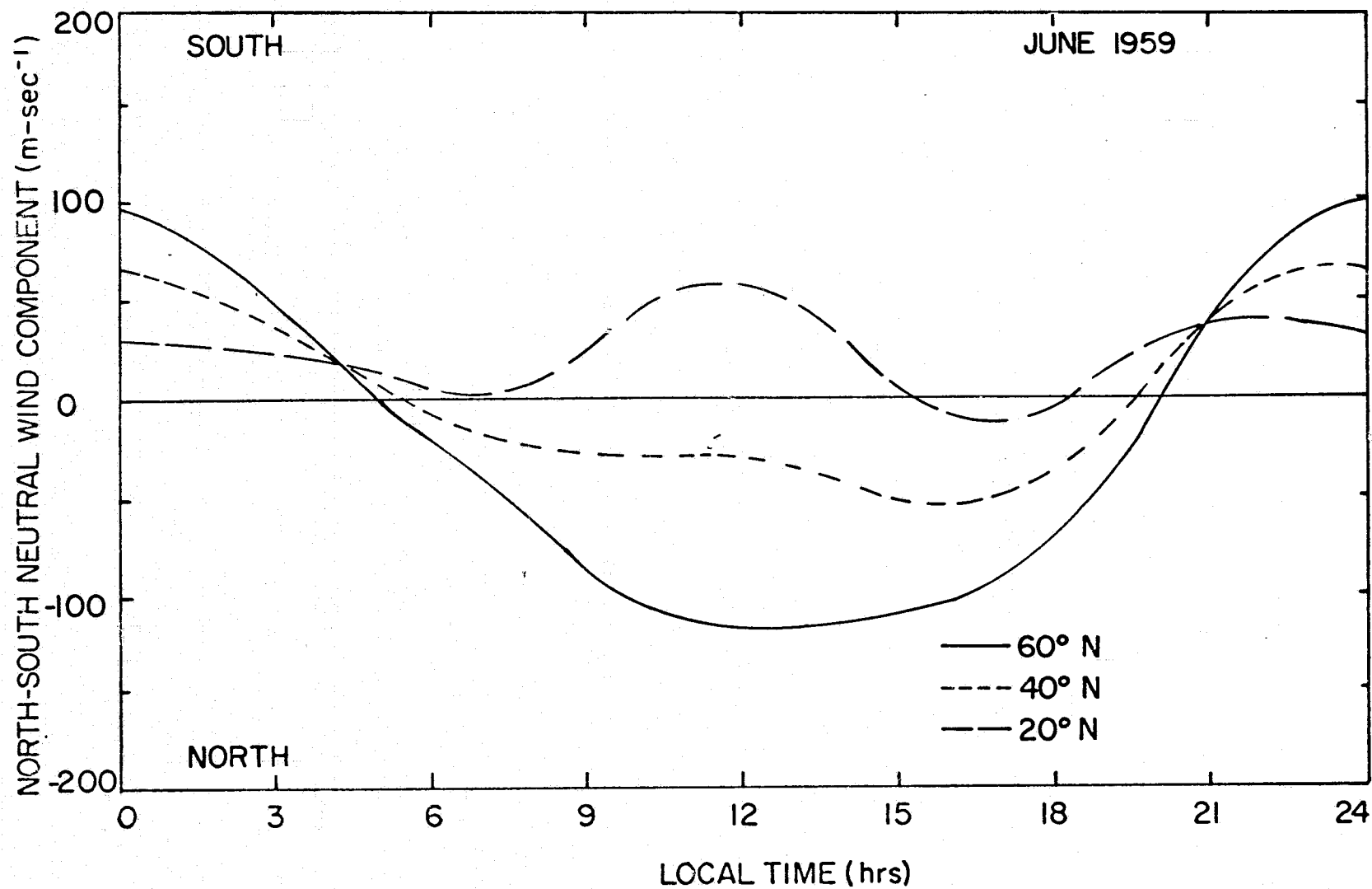


Figure 20. Latitudinal Variations of the North-South Neutral Wind Component for Summer Conditions

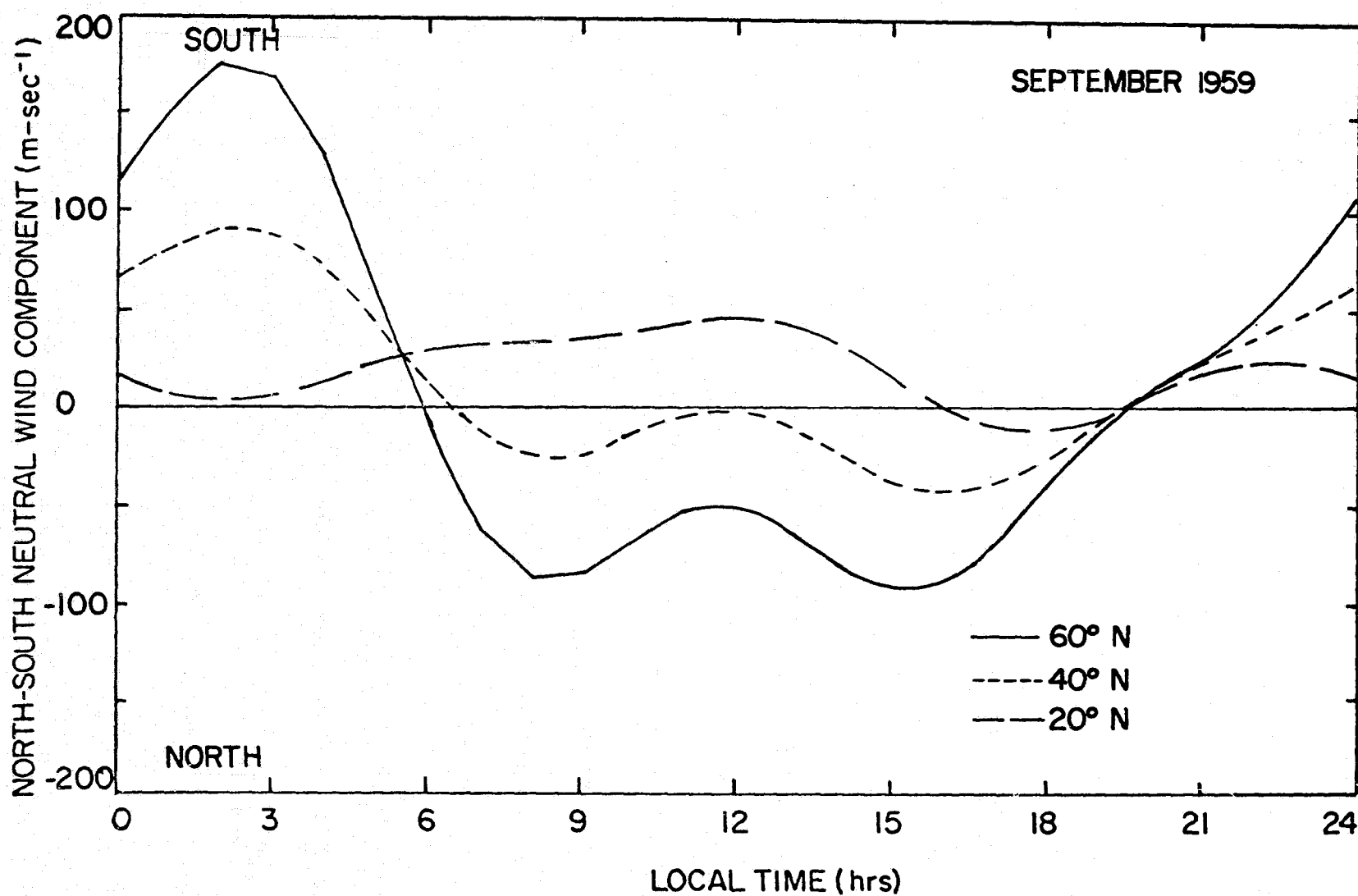


Figure 21. Latitudinal Variations of the North-South Neutral Wind Component for Equinox Conditions

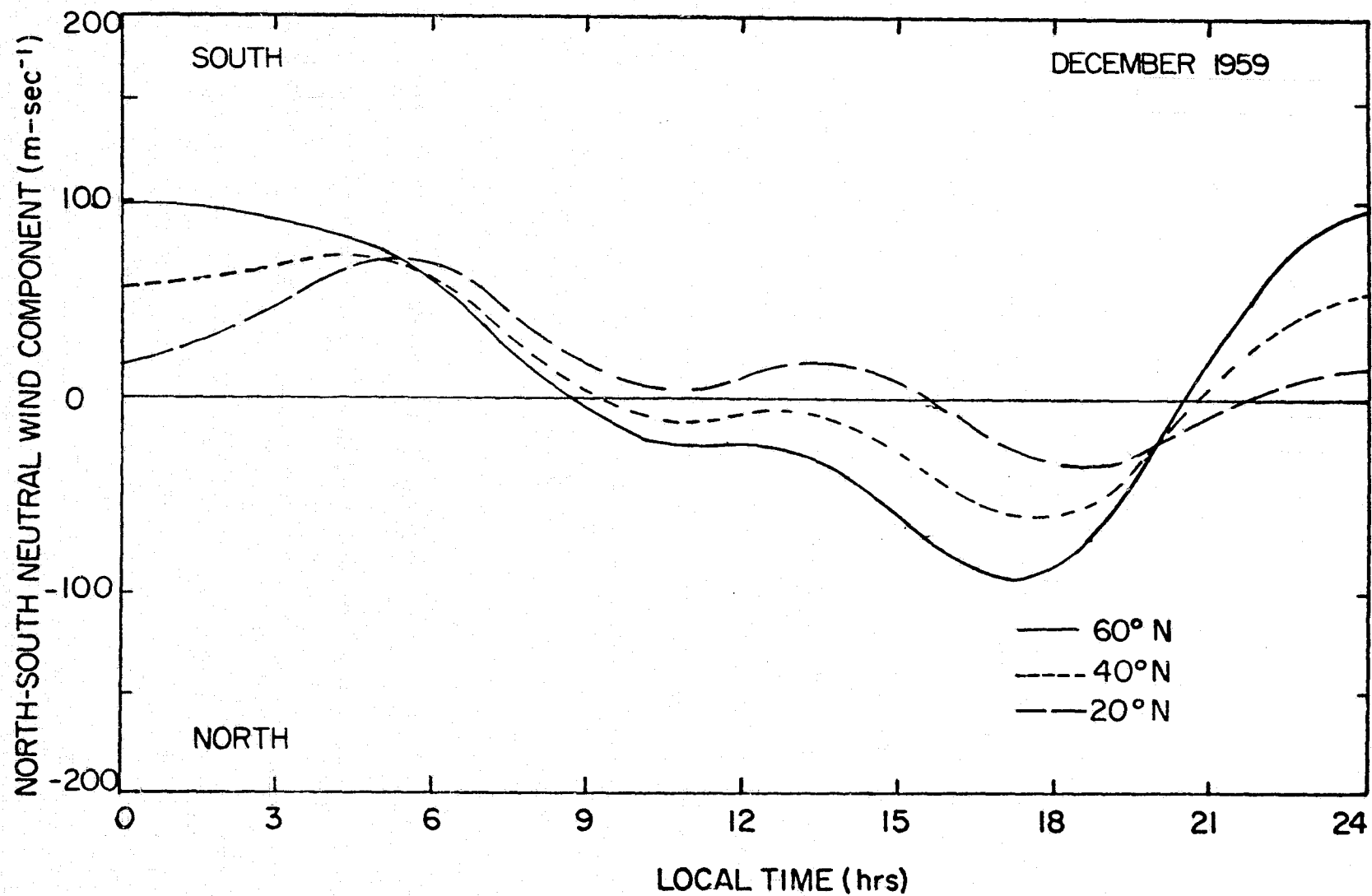


Figure 22. Latitudinal Variations of the North-South Neutral Wind Component for Winter Conditions

The incoherent scatter technique is an indirect method for neutral wind determination because it is the ion velocity which is actually measured. By measuring the drift velocities parallel to the magnetic field line the ion drifts induced by any electric fields are filtered out. The ion drift is then a function of the neutral wind velocity in the direction of the magnetic meridian and the diffusion velocity which is calculated by employing an atmospheric model.

Theoretical neutral atmospheric models are obtained by solving the mass, energy and momentum conservation equations for neutrals and the momentum equations for ions in coupled form or a subset of these equations with various assumptions. Because the ion drag term in the momentum equation requires the knowledge of the ion density, it must be supplied by an ionospheric model. Theoretical ionospheric models are obtained by solving the mass, energy and conservation equations for ions and the momentum equation for neutrals in coupled form. In this case the neutral density which is required for the determination of the ion drag term has usually been supplied by a static neutral atmospheric model.

For both theoretical atmospheric neutral models and theoretical ionospheric models electric fields are supplied either directly or indirectly from outside sources. Because the knowledge of electric fields is rather uncertain and because the effect of electric fields is assumed to be small at mid-latitudes, many authors have not included electric field effects in their calculations.

In Figures 23, 24, 25 and 26, comparisons of the North-South and West-East neutral wind components determined by this study are compared with those calculated by Stubbe's theoretical ionospheric model

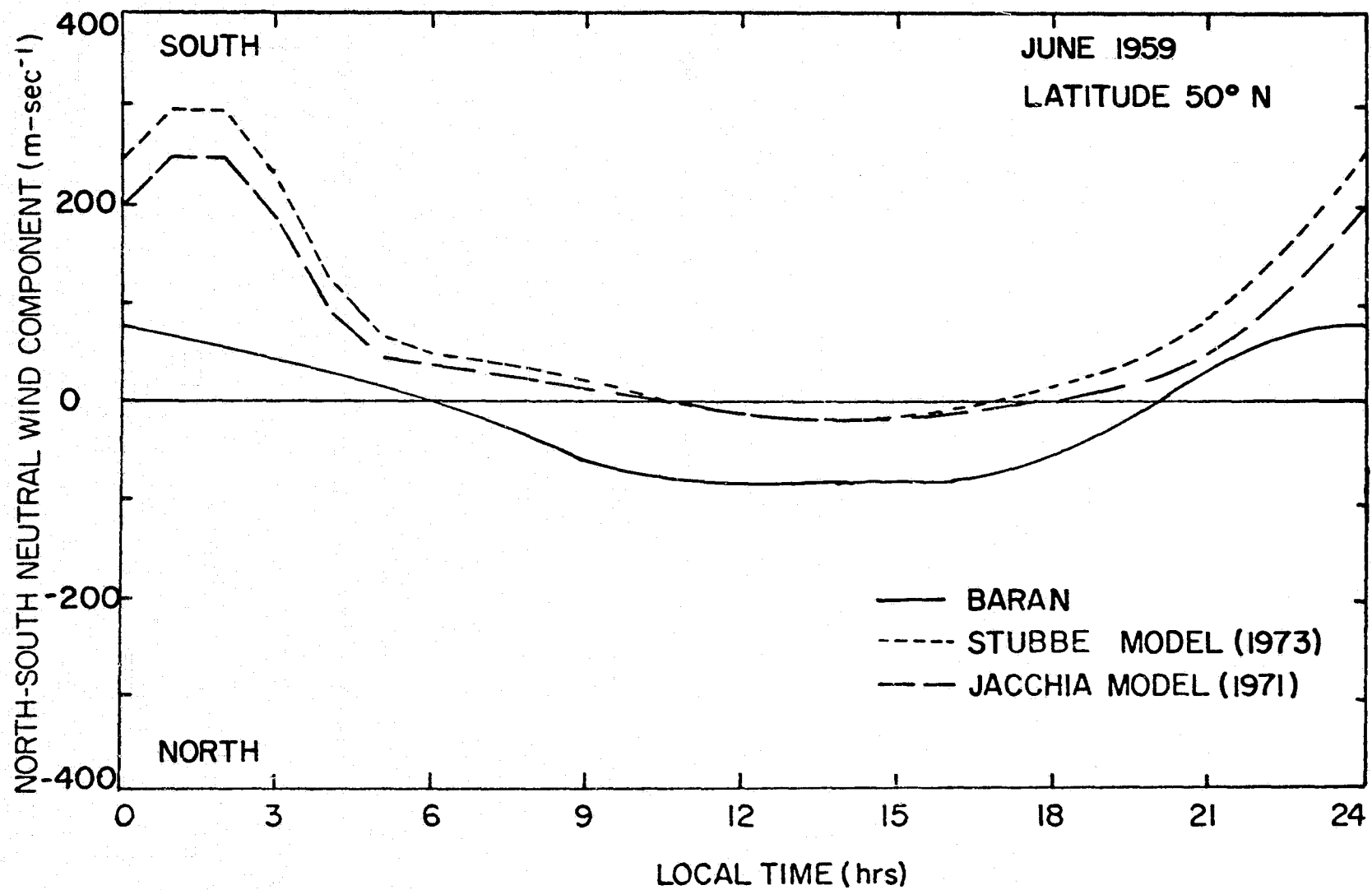


Figure 23. Comparisons of the Calculated North-South Neutral Wind with the Results of Stubbe for Summer Conditions

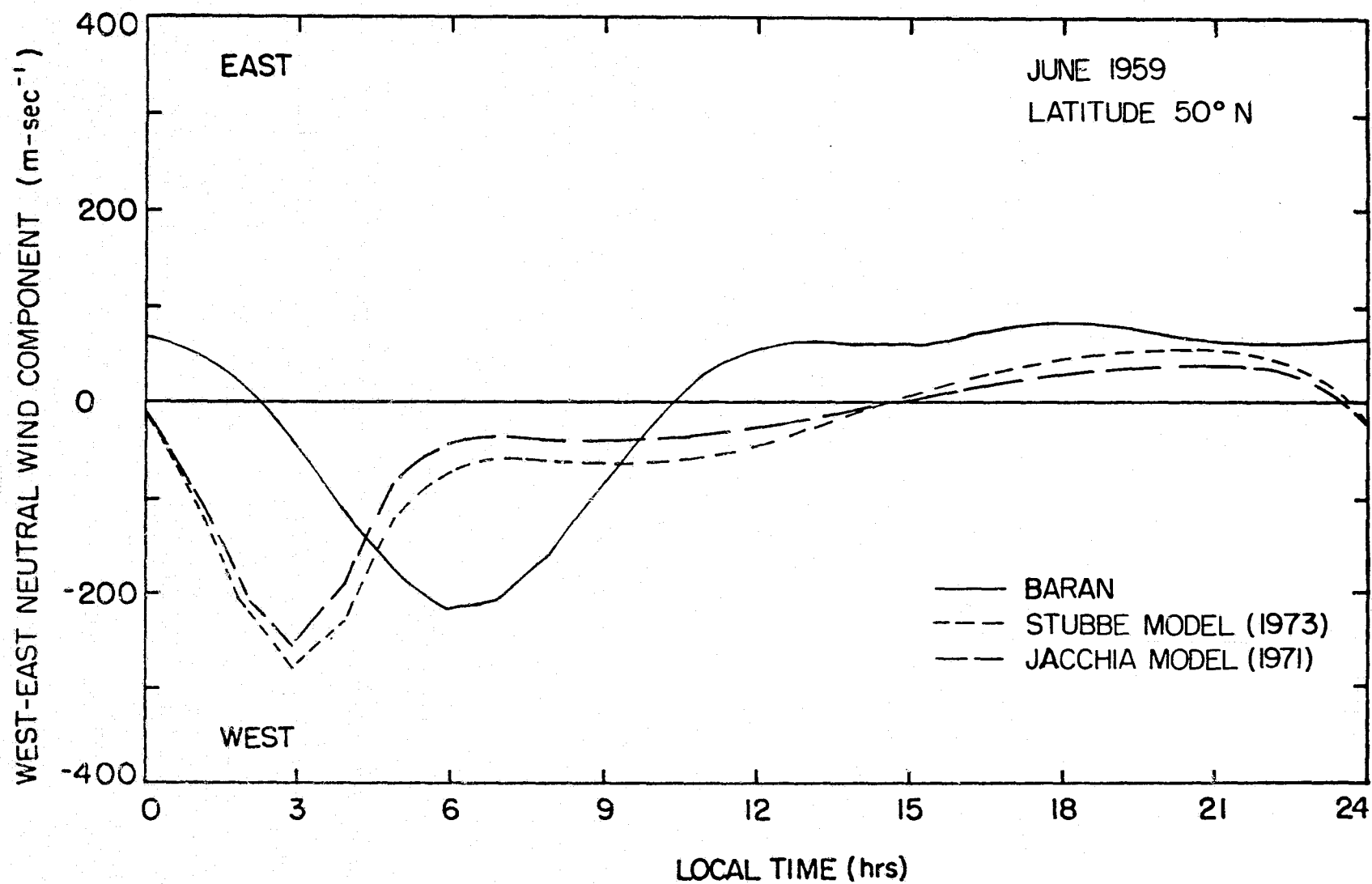


Figure 24. Comparisons of the Calculated West-East Neutral Wind with the Results of Stubbe for Summer Conditions

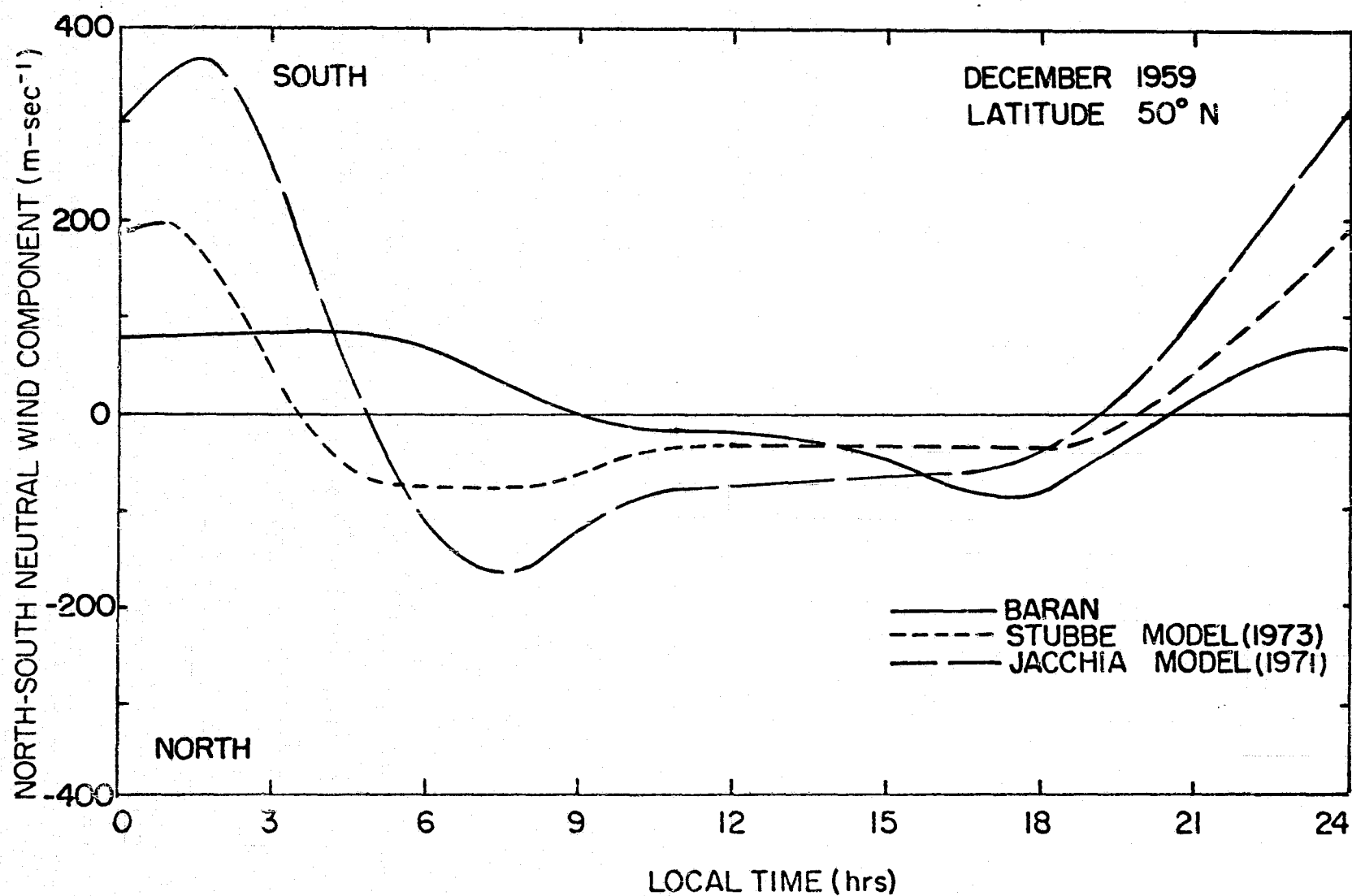


Figure 25. Comparisons of the Calculated North-South Neutral Wind with the Results of Stubbe for Winter Conditions

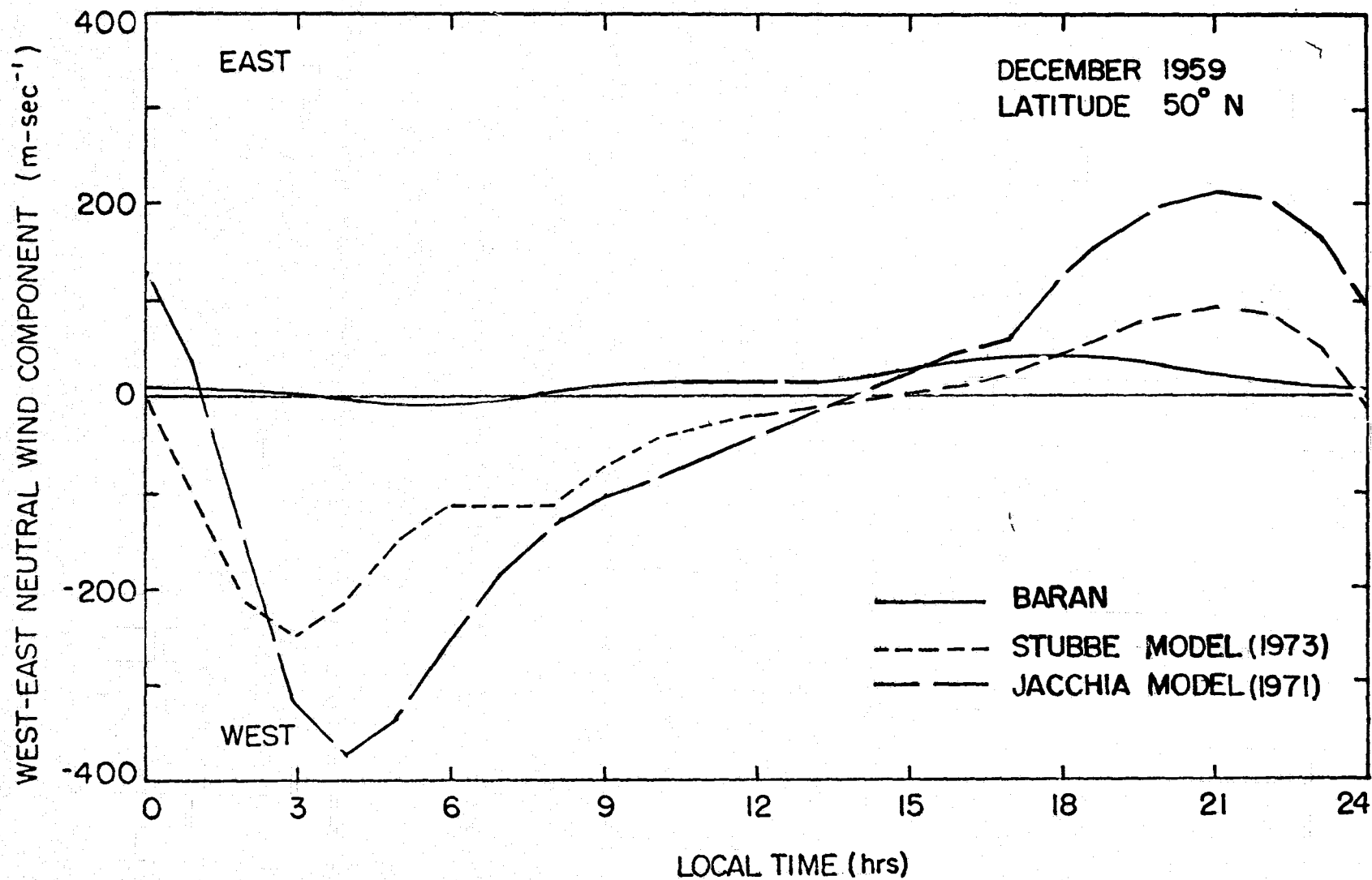


Figure 26. Comparisons of the Calculated West-East Neutral Wind with the Results of Stubbe for Winter Conditions

(1973, 1974) for summer and winter conditions. The data base used in the two sets of calculations is the same but Stubbe has assumed the Jacchia (1971) density gradients. The theoretical winds calculated by Stubbe are given for two cases: in one the Stubbe (1973) modified Jacchia static neutral atmospheric model was employed and in the other the Jacchia (1971) static atmospheric model was employed (see Section 1.2.8). From these figures it is clearly seen that a very large difference exists between the results calculated by this study and the theoretical results calculated by Stubbe for either static neutral atmospheric model. The fact that these theoretical neutral winds do not produce the proper ionospheric behavior indicates that the pressure gradients of these static neutral atmospheric models are incorrect. Note that the inclusion of the electric fields could not account for discrepancies observed in the above results. Also, it is pointed out that the theoretical results of Stubbe are in general agreement with the results of many other investigators who employ theoretical ionospheric models for determining the neutral winds. This is not surprising because, as was shown in Section 1.2.7, the Jacchia (1971) temperatures have quite different gradients from those observed in practice.

For equinox conditions at a latitude of approximately 50°N , the North-South neutral wind components calculated by this study are compared with results from incoherent scatter measurements made at CNET (Amayenc and Vasseur, 1972) and with theoretical results calculated by Volland and Mayr (1973) and Vest (1973b) in Figure 27. The incoherent scatter results are for the year 1968. Though the solar activity was approaching its maximum at this time, the solar fluxes

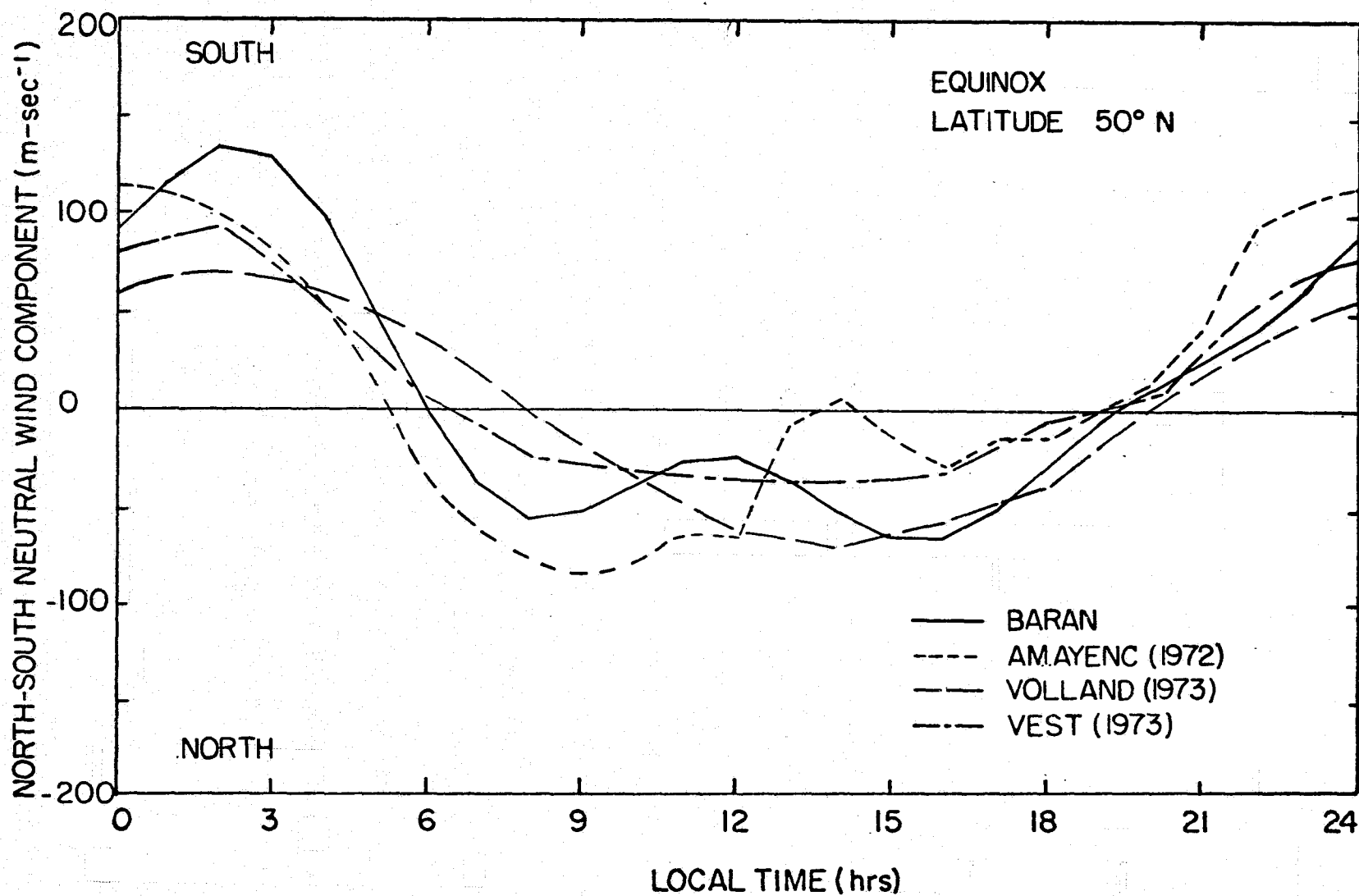


Figure 27. Comparisons of the Calculated North-South Neutral Wind with Theoretical and Measured Results at 50°N Latitude

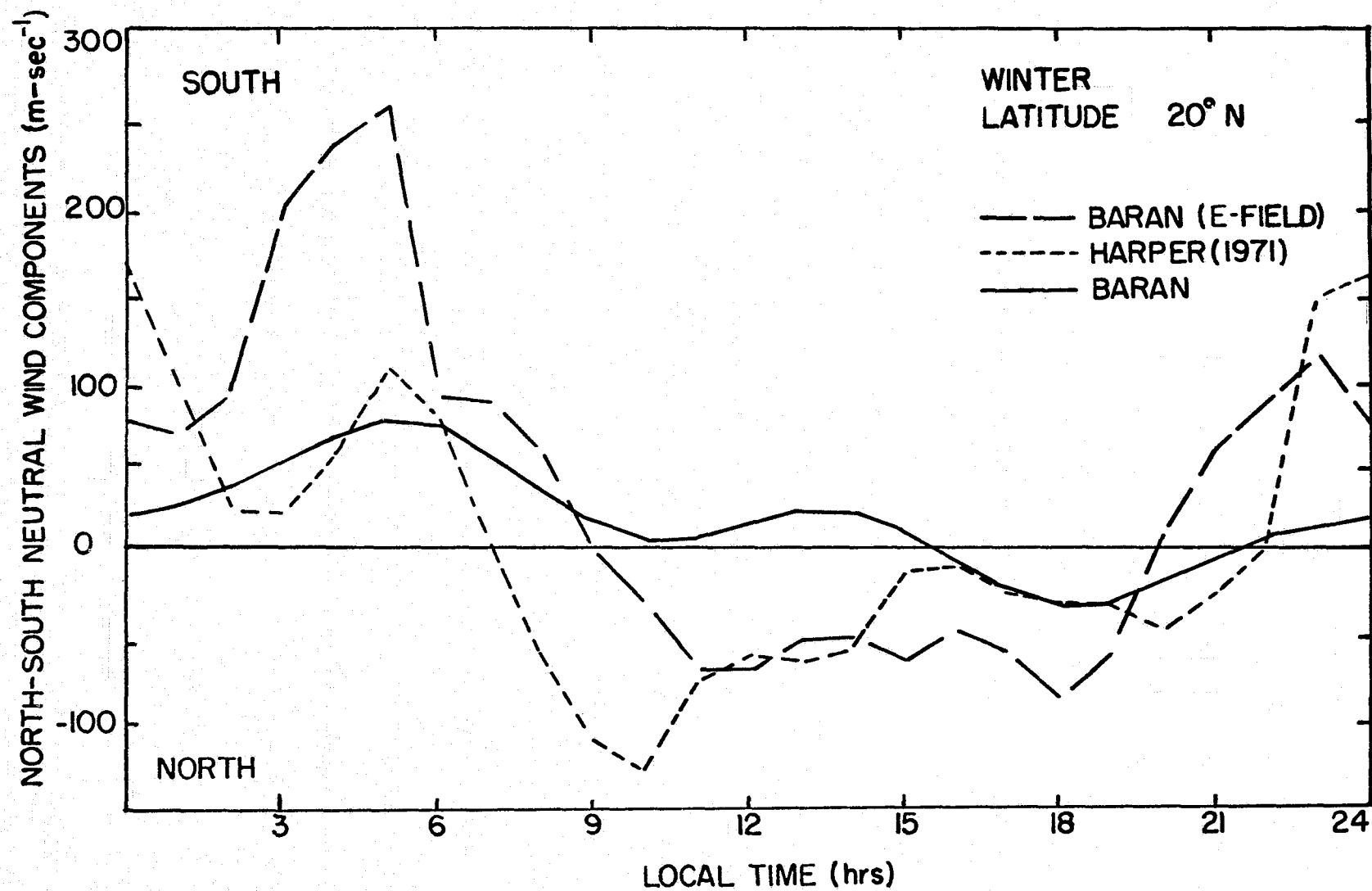


Figure 28. Comparisons of the Calculated North-South Neutral Wind with Theoretical and Measured Results at 20°N Latitude

were not as high as they were during the 1958 period. Consequently, this comparison is made under similar but not identical conditions. In addition, the incoherent scatter measurements are affected differently by electric fields. Nevertheless, it is seen that there is general agreement between the results calculated by this study and those calculated by incoherent scatter measurements. The winds calculated by Volland and Mayr (1973) are those produced by the fundamental symmetric diurnal mode. The agreement with theoretical winds is also reasonable.

In Figure 28 the North-South neutral wind components calculated by this study for two cases are compared with those calculated from incoherent scatter measurements made at Arecibo (Harper, 1971) for winter conditions. These Arecibo measurements are for the year 1971. The agreement is seen to be rather poor for the case in which no electric fields are included. However, for the case in which the electric field effects are included, the comparison is rather striking in that the "midnight reversal," which is a regular feature observed at Arecibo, is reproduced. The effect of the electric fields was included by employing the technique described in Section 3.4.2. It is pointed out that when employing this technique the error in determining the North-South neutral wind component is of the order of 40 m-sec^{-1} .

From the comparisons presented in this section it can be concluded that neutral winds calculated by methods developed by this study show little agreement with neutral winds calculated from static atmospheric models. The agreement with neutral winds calculated from incoherent scatter measurements and dynamic neutral atmospheric models

can be described as being good and the inclusion of electric field effects can improve this agreement.

It is thus apparent that wind fields calculated from ionograms can provide information obtainable in no other way on a global basis. The accuracy of the technique is currently limited by the number of stations for which reduced ionograms are available and the accuracy with which the electron fields are known. Both of these problems are receiving considerable attention at the present time.

CHAPTER IV

DETERMINATION OF THE HORIZONTAL PRESSURE GRADIENTS AND SENSITIVITY FACTORS

4.1 Procedure

After generating the neutral wind field, the final step is to generate a new set of pressure gradients which, when used in the theoretical ionospheric models with the densities from the static neutral atmospheric models, would reproduce the desired wind field. The new pressure gradients are compared with pressure gradients calculated from various static neutral atmospheric models. A system of sensitivity factors which relate the output neutral wind harmonics to the input pressure gradient harmonics is also developed and discussed. These sensitivity factors are of great importance as they allow the coupling between the different modal components to be estimated. This coupling between the zonal and meridional components and between the harmonics of each component is particularly important because of the attempts being made to develop models of the neutral atmosphere based on linearized expansions of densities and temperatures in terms of spherical harmonics.

4.2 Horizontal Pressure Gradients Calculated from the Neutral Winds

It was suggested in Section 3.5 that the horizontal pressure gradients derived from static neutral atmospheric models are in error. Previous studies (Barlier, et al., 1972; Blamont, et al., 1973) have cited possible causes for pressure gradient discrepancies. In this section the pressure gradients which are necessary to produce the neutral winds which were presented in Section 3.4.3 are calculated.

This calculation is performed by employing the momentum equations for neutrals and ions. In a coordinate system fixed with respect to the rotating earth the momentum equation for neutral gas is given as

$$\begin{aligned} \frac{1}{\rho} \text{grad } (p) = & - \frac{\partial \bar{U}}{\partial t} - (\bar{U} \cdot \bar{V}) \bar{U} - \bar{\Omega} \times (\bar{\Omega} \times \bar{R}) \\ & - 2\bar{\Omega} \times \bar{U} - \frac{1}{\rho} \sum_i (\bar{U} - \bar{V}_i) \sum_n n_n v_{ni} m_{ni} \\ & + \frac{\mu}{\rho} \nabla^2 \bar{U} + \bar{g} \end{aligned} \quad (4.1)$$

With regard to the equation of motion for neutrals itself the centripetal acceleration and electron drag terms are neglected. Because the vertical velocity is several orders of magnitude less than the horizontal velocity, terms involving the vertical velocity are neglected. Horizontal neutral wind gradients are neglected when compared to the vertical gradients, but when they stand alone as in the non-linear advection term, the horizontal neutral wind gradients are estimated by interchanging geographic longitude and local time (Stubbe, 1973; Rüster and Dudeney, 1972). This interchange is shown by the relationship

$$\frac{\partial}{\partial y} = f(\phi) \frac{\partial}{\partial t} \quad (4.2)$$

where

$$f(\phi) = \frac{2.16 \times 10^{-3} \text{ m}^{-1} \text{-sec}}{\cos \phi}$$

By using these assumptions and employing the momentum equation for ions (Equation 2.10) for the ion velocities, the neutral momentum equation is written in its component form

$$\begin{aligned}
\frac{1}{\rho} \frac{\partial p}{\partial x} = & - \frac{\partial U_x}{\partial t} (1 + 2.16 \times 10^{-3} U_y / \cos \phi) + 2\Omega U_y \sin \phi + \frac{\mu}{\rho} \frac{\partial^2 U_x}{\partial z^2} \\
& - \frac{v_{in}}{2} \left[U_x - U_x \cos^2 I \cos^2 D + U_y \cos^2 I \sin D \cos D \right. \\
& + \frac{k}{v_{in}^m} \left(\frac{\partial (T_i + T_e)}{\partial z} + \frac{2g}{v_{in}} \sin I \cos I \cos D \right) \\
& \left. - \frac{E_x}{B} \cos I \cot I \sin D \cos D + \frac{E_y}{B} (\sin I + \sin I \cot I \sin^2 D) \right]
\end{aligned} \tag{4.3}$$

$$\begin{aligned}
\frac{1}{\rho} \frac{\partial p}{\partial y} = & - \frac{\partial U_y}{\partial t} (1 + 2.16 \times 10^{-3} U_x / \cos \phi) - 2\Omega U_x \sin \phi + \frac{\mu}{\rho} \frac{\partial^2 U_y}{\partial z^2} \\
& - \frac{v_{in}}{2} \left[U_y + U_x \cos^2 I \sin D \cos D - U_y \cos^2 I \sin^2 D \right. \\
& - \frac{k}{v_{in}^m} \left(\frac{\partial (T_i + T_e)}{\partial z} + \frac{2g}{v_{in}} \sin I \cos I \sin D \right) \\
& \left. - \frac{E_x}{B} (\sin I + \cos I \cot I \cos^2 D) + \frac{E_y}{B} \cos I \cot I \sin D \cos D \right]
\end{aligned} \tag{4.4}$$

All the terms in Equations 4.3 and 4.4 are readily available except the viscous term and the electric field term. The effects of these terms on the pressure gradient calculations are estimated and then these terms are neglected in succeeding calculations.

Because the force of the North-South pressure gradient is coupled into the West-East directions and vice versa by means of the Coriolis and ion drag terms, both components of the neutral winds must be supplied simultaneously. For this study both neutral wind components are only available for the 50°N latitude and therefore the pressure gradient calculations were made only for that latitude. The altitude for which the calculated values are given is 300 km. In

addition, it is seen that the calculated pressure gradients are divided by the density. This was done because the pressure gradient/density has the units of an acceleration which is a more convenient unit with which to work. The viscous term is important in calculations in which the differential momentum equation is solved because it causes the vertical gradient in the wind velocity to become small. The neglect of this term can cause the solution to blow up. However, if the values of the various terms in the momentum equation are known, as they are for theoretical studies, the algebraic solution of the momentum equation yields a value for the viscous term which typically ranges from -1 cm-sec^{-2} to 1 cm-sec^{-2} .

In order to determine the effects of the neglect of the electric fields, the pressure gradient/density terms are calculated for three cases:

- 1) no electric fields,
- 2) total electric fields from Matsushita (1971), and
- 3) electrostatic fields from Matsushita (1971).

These electric field models were employed earlier in Section 3.4.1. The measured electric fields of Kirchhoff and Carpenter (1974) are not used in these calculations because it was shown in Section 3.4.1 that they produce much smaller effects than do theoretical fields. Comparisons of the three cases are shown in Figures 29 and 30 for the North-South and West-East pressure gradient/density components. For the North-South component the introduction of the total electric field can alter the values of the no field case by as much as 4 cm-sec^{-2} for the extreme condition, but typical variations are approximately 2.5 cm-sec^{-2} . The variations caused by the introduction of the

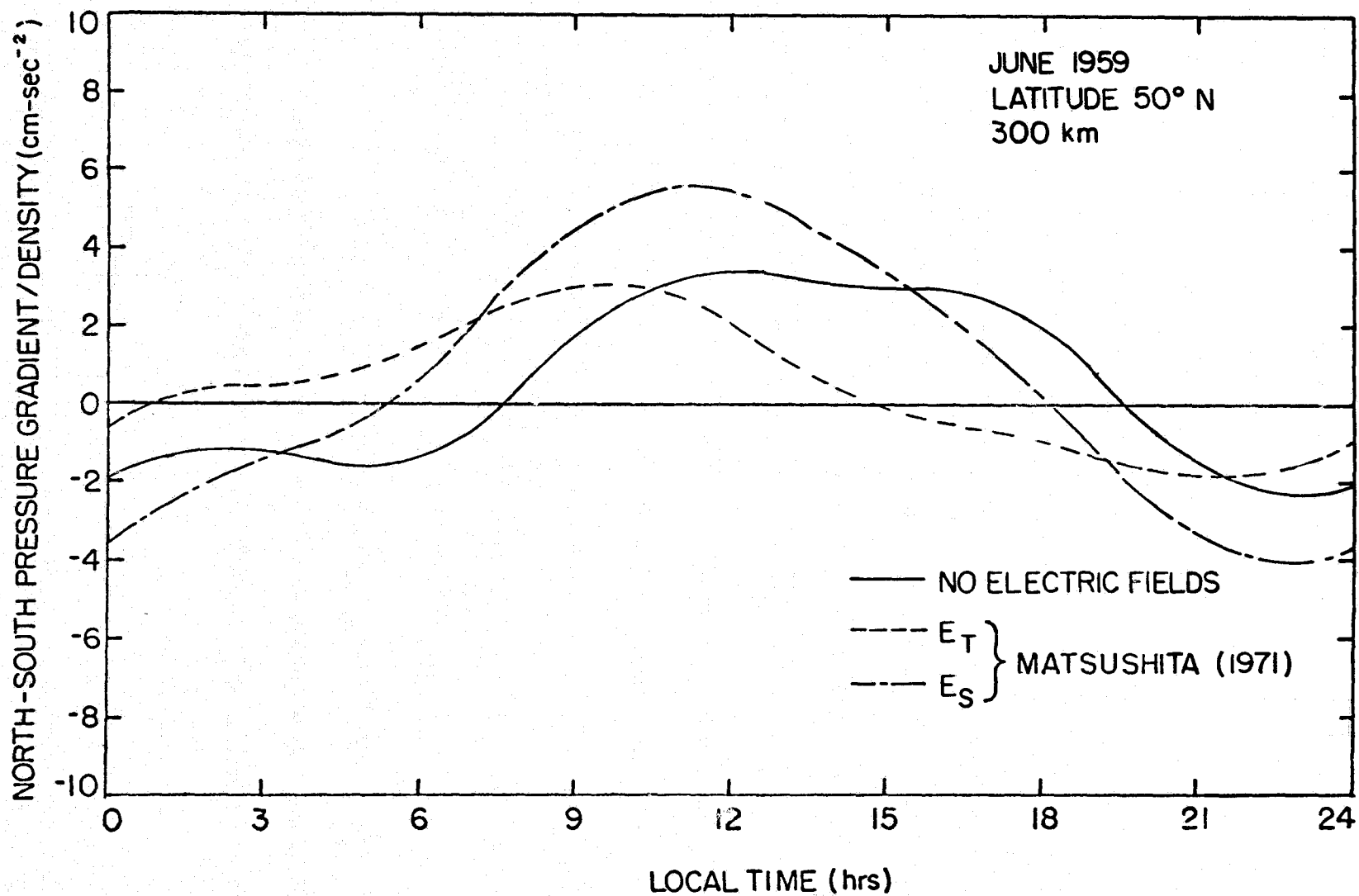


Figure 29. Effect of Electric Fields on the Calculations of the North-South Pressure Gradient/Density Term

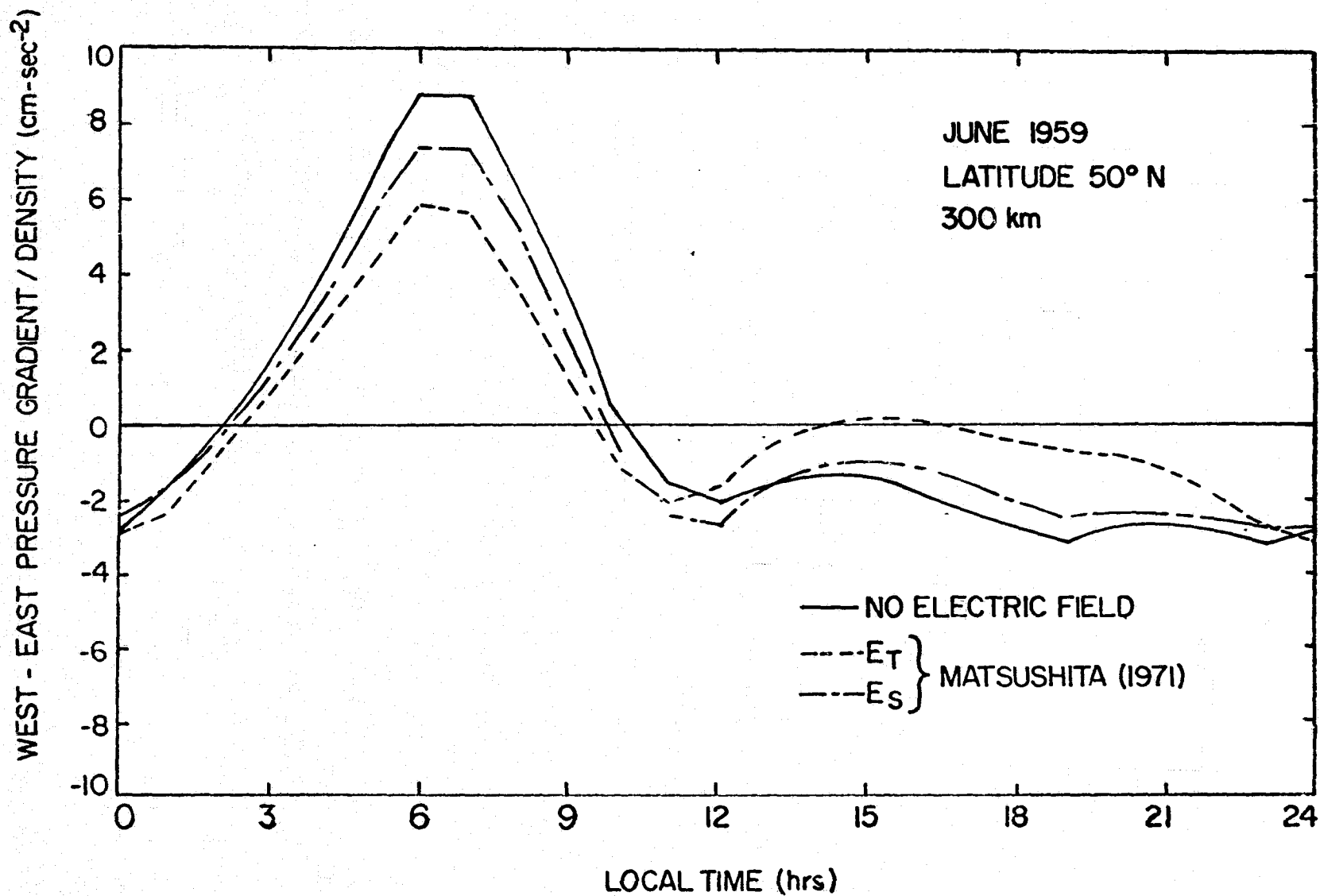


Figure 30. Effect of Electric Fields on the Calculations of the West-East Pressure Gradient/Density Term

electrostatic fields are not as great. For the most part these variations are less than 2 cm-sec^{-2} . The directions of the variations caused by the electric fields are dependent upon the field model assumed; hence, no statement about the systematic effects of electric fields in general can be made.

On the other hand, the effects of the electric field on the West-East pressure gradient/density component are more readily categorized (see Figure 30). The introduction of either field type tends to decrease the calculated pressure gradient/density in the morning and increase it in the afternoon, and the total electric field produces the larger variations. The magnitudes of the electric field induced variations on the West-East component are much less than those induced on the North-South component. Notice that the variations caused by the introduction of the electric fields on the pressure gradient/density calculations parallel those field induced variations caused on the neutral wind calculations (see Section 3.4.1).

After establishing an approximate range of the variations of the calculated pressure gradient/density terms caused by the neglect of the viscous term and electric fields, the North-South and West-East components of the pressure gradient/density terms are presented for summer and winter conditions in Figures 31, 32, 33, and 34. These calculated values are compared with values given by the Stubbe (1973) modified Jacchia model, and those given by two schemes of the Goddard OGO VI model (Hedin et al., 1974). Values calculated from the Jacchia model are not presented because these are almost identical to those of the Stubbe modified Jacchia model.

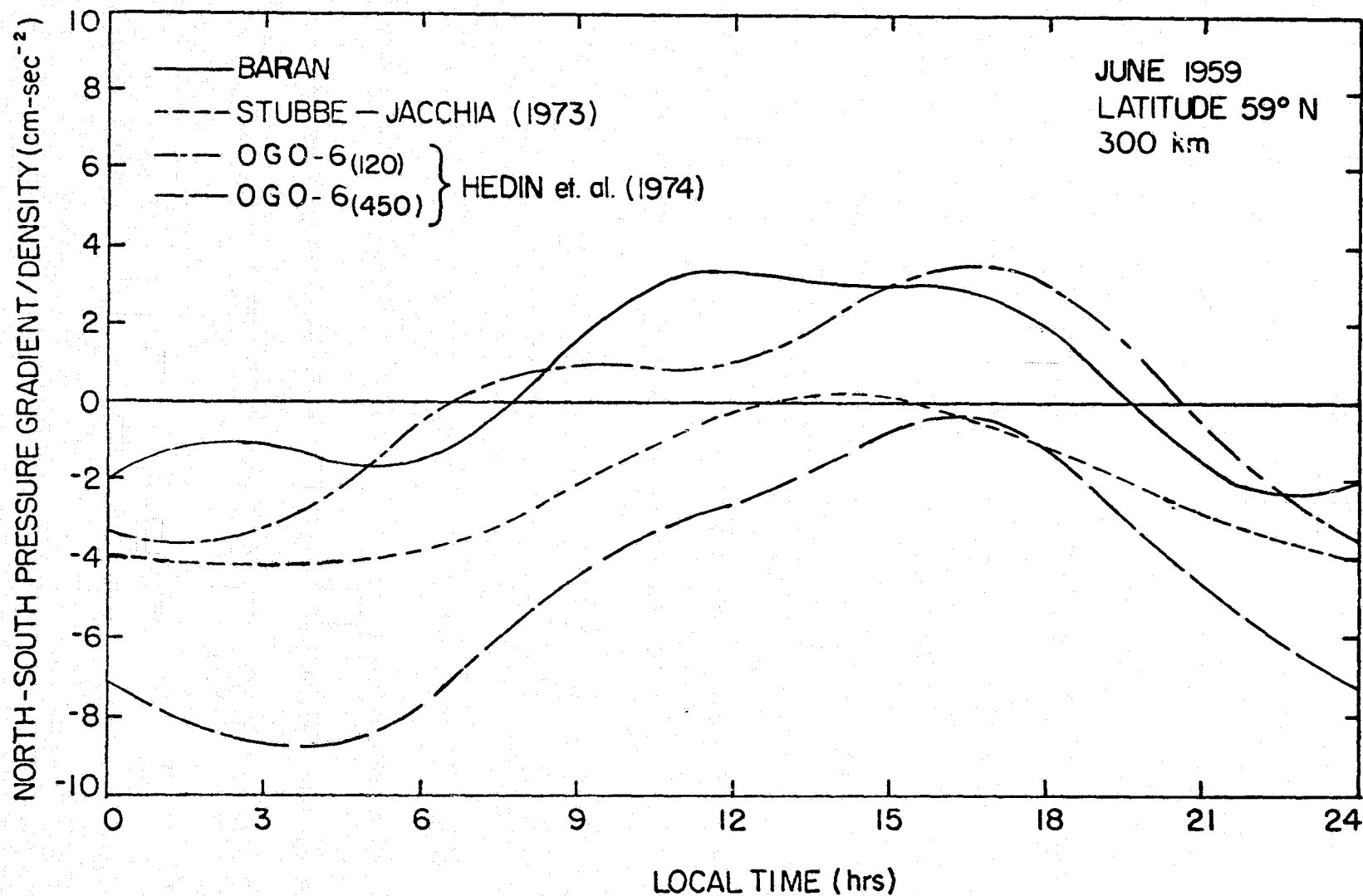


Figure 31. Comparisons of the Calculated North-South Pressure Gradient/Density with Results from Static Neutral Atmospheric Models for Summer Conditions

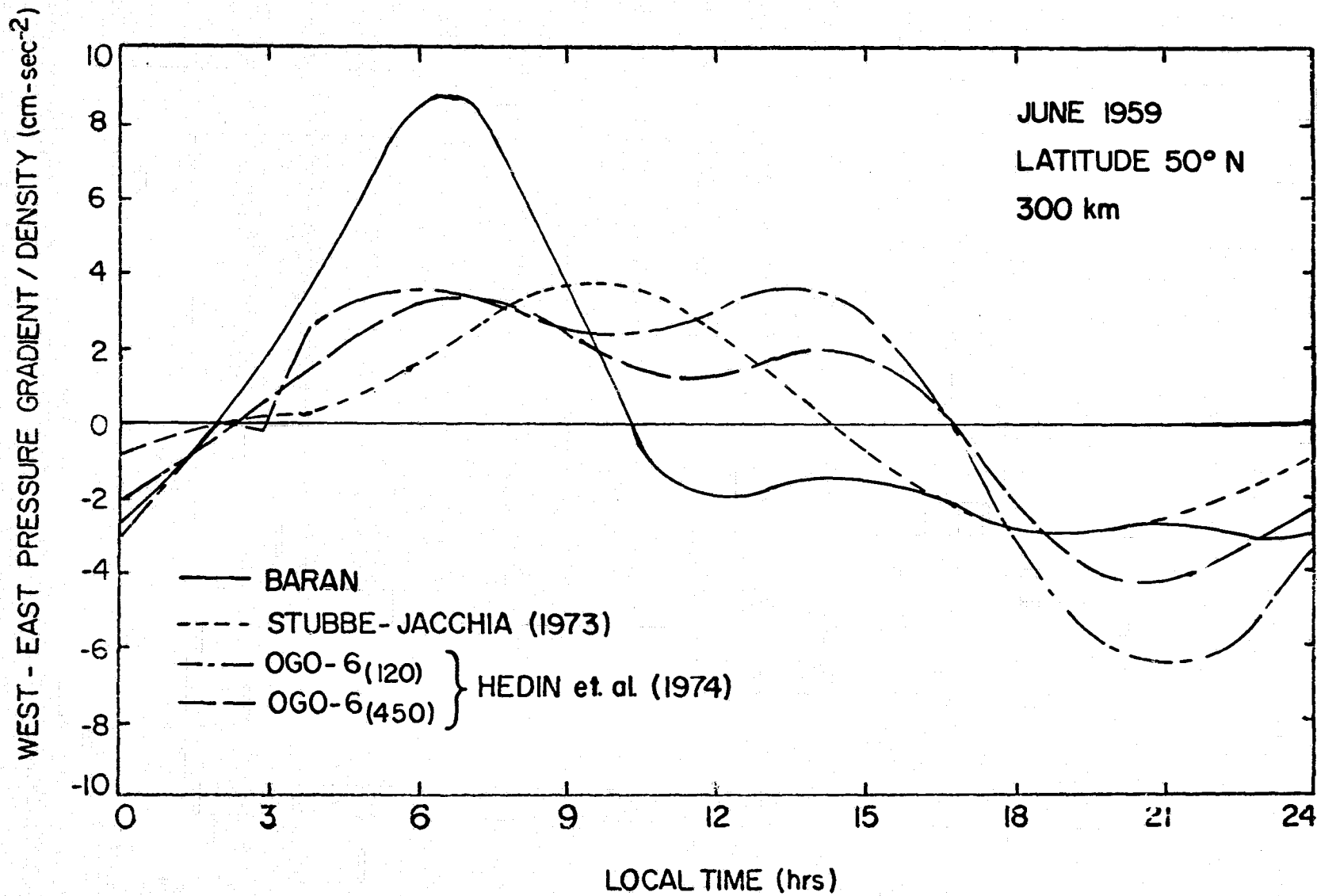


Figure 32. Comparisons of the Calculated West-East Pressure Gradient/Density with Results from Static Neutral Atmospheric Models for Summer Conditions

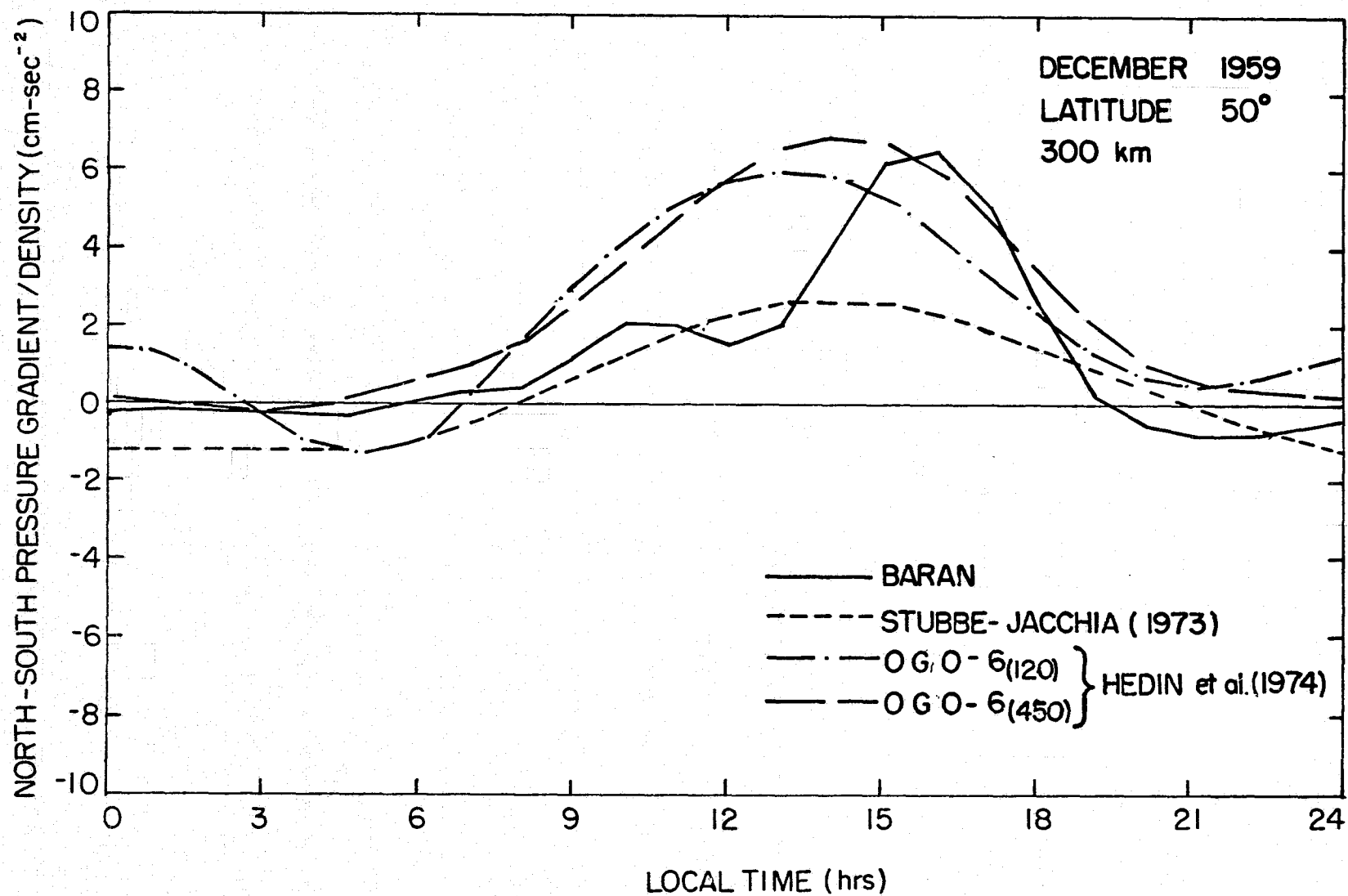


Figure 33. Comparisons of the Calculated North-South Pressure Gradient/Density with Results from Static Neutral Atmospheric Models for Winter Conditions

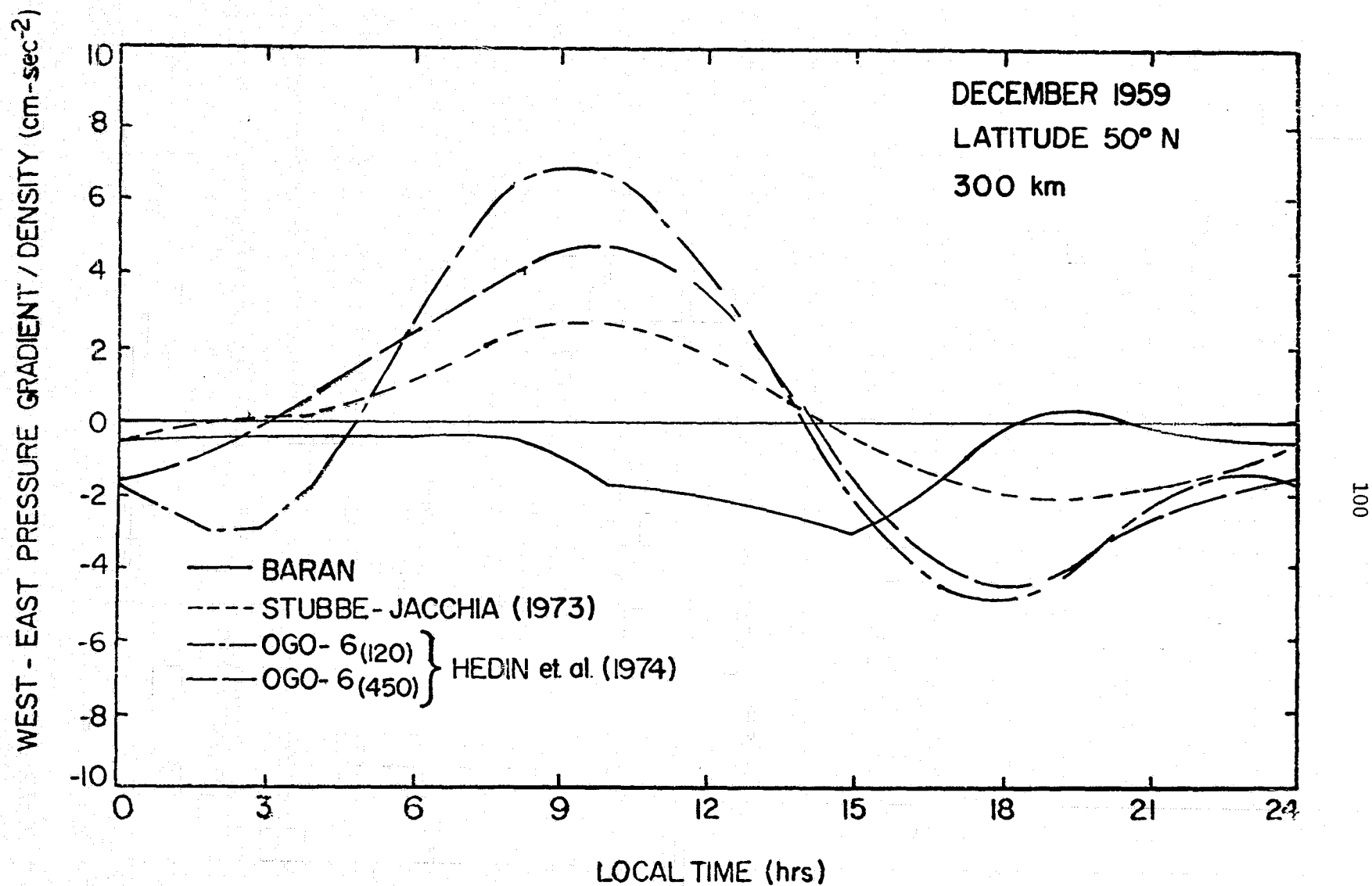


Figure 34. Comparisons of the Calculated West-East Pressure Gradient/Density with Results from Static Neutral Atmospheric Models for Winter Conditions

From Figure 31 for June conditions, it is seen that the North-South component of the pressure gradient/density calculated by this study agrees best with those values given by the Goddard OGO VI₍₁₂₀₎ model. In addition to this generalized agreement, it is apparent that this study's values and the OGO VI₍₁₂₀₎ values both contain higher order harmonics (though out of phase) while the values from the other models primarily show only a diurnal variation. It is also seen that the differences among the various models are quite large especially with regard to the constant term.

The West-East components of the pressure gradient/density for June conditions are presented in Figure 32. In general, the values calculated from the various models agree with each other whereas those values calculated by this study differ greatly from the other models' values at 0600 hours and noon. Higher order modes are apparent in this study's values and both OGO IV representations. The Stubbe modified Jacchia model primarily shows only a diurnal variation.

In Figure 33 the variations of the North-South pressure gradient/density components for December are given. During the nighttime the values calculated from this study and those calculated from all of the models are in reasonable agreement. The results from both OGO VI modeling schemes are similar and these differ greatly from the Stubbe modified Jacchia results only during the day. The results from this study agree best with the Stubbe-Jacchia results during the morning and after noon are more similar to the OGO VI results.

From Figure 34 the results for the West-East pressure gradient/density calculations are seen. The differences among all of the calculated values are substantial particularly in the late morning

hours and the results calculated by this study are not similar to the results calculated from any of the models.

To obtain a more quantitative feeling for the results presented in the last four figures, a harmonic analysis of the pressure gradient/density terms is given in Tables 6 and 7 for summer and winter conditions respectively. Because the Jacchia (1971) model has been adopted as the CIRA 1972 reference atmosphere, the Stubbe (1973) modified Jacchia model has been taken as the standard and the percentage variation of the magnitude of the harmonics of the other models and this study's values are given in terms of it. From these tables the important points to notice are:

- 1) The magnitudes of the harmonics of the OGO VI results and this study's results are generally greater than those of the Jacchia model.
- 2) The percentage change is greater for the higher order harmonics.
- 3) The variations suggested by this study are of the same order of magnitude as the variations between the models.

In summary, the results of this section indicate that the horizontal neutral wind derived from ionogram data can be used to obtain as a first approximation a new system of horizontal pressure gradients. This new system of horizontal pressure gradients is well within the range of the uncertainties of the pressure gradients calculated from current neutral atmospheric models. Though the effects of electric fields are not second order and must be considered, they cannot account for the variability of the pressure gradients derived by the various methods.

Table 6

Percentage Change of the Magnitudes of the Harmonics of the
Pressure Gradient/Density for Summer Conditions

Harmonic	Modified Jacchia (1971) Stubbe (1973) Magnitude (cm-sec ⁻²)	Magnitude Change (%)		
		Hedin et al. (1974) OGO VI (450)	Hedin et al. (1974) OGO VI (120)	Baran
N-S DC	.19	1078	921	415
N-S 1	2.00	66	40	21
N-S 2	.27	274	441	381
N-S 3	.05	300	720	1960
W-E DC	- .01	600	700	10600
W-E 1	2.09	90	122	- 53
W-E 2	.61	123	334	21
W-E 3	.14	107	264	114

Table 7

Percentage Change of the Magnitudes of the Harmonics of the
Pressure Gradient/Density for Winter Conditions

Harmonic	Modified Jacchia (1971) Stubbe (1973) Magnitude (cm-sec ⁻²)	Magnitude Change (%)		
		Hedin et al. (1974) OGO VI (45)	Hedin et al. (1974) OGO VI (120)	Baran
N-S DC	-2.25	107	-104	-119
N-S 1	2.24	78	33	25
N-S 2	.28	- 14	329	82
N-S 3	.06	617	333	1050
W-E DC	- .01	-200	100	-2400
W-E 1	2.94	8	55	50
W-E 2	.83	42	161	213
W-E 3	.21	22	86	519

4.3 Mode Coupling and Sensitivity Factors

The calculations of the horizontal pressure gradients presented in the previous section employed the momentum equations for ions and neutrals. Because in this study it is assumed that the horizontal neutral winds are independent of height, the viscous term was neglected. In order to include the effects of viscosity in the pressure gradient calculations a method of sensitivity factors was developed. This method employed the Stubbe ionospheric model in the following manner. For a given set of conditions the Stubbe model was run in its entirety except that the harmonics of the pressure gradient/density driving force were supplied externally. (Normally these harmonics are implicitly supplied from the neutral atmospheric model.) This technique is compatible with one of the assumptions of this study; namely, that the total densities supplied by the static neutral atmospheric models are quite accurate but the pressure gradients are incorrect. The input pressure gradient/density harmonics were varied and a set of sensitivity factors was obtained by relating changes in the harmonics of the neutral horizontal winds to changes in the harmonics of the pressure gradient/density. By using a set of sensitivity factors such as these and the harmonics of the horizontal neutral winds as given in Section 3.4.3, the horizontal pressure gradients could be calculated. Pressure gradients calculated in this way would include the effects of the viscous term.

Though the technique described above is very straightforward in theory, its application is not very practical. This lack of practicality results from the non-linearities in the ionospheric system, which in turn cause the sensitivity factors to be non-linear and the

harmonic modes to be coupled. The degree of the non-linearity and mode coupling would require the entire Stubbe ionospheric model or an equivalent model to be run many times. For the small improvement to be gained on pressure gradient accuracies by this method, at present it is not economically feasible to do this. It should be pointed out, however, that an accurate initial estimate of the pressure gradients, which could be obtained by the method described in Section 4.2, would reduce the degree of the non-linearity.

For economical reasons the sensitivity factors were not used to calculate pressure gradients. However, because mode coupling is of great interest to a number of thermospheric studies which employ spherical harmonics and perturbation theory (Volland and Mayr, 1972a, 1973), a set of sensitivity factors is presented to demonstrate the degree of mode coupling. The constant term, first harmonic and second harmonic of the North-South and West-East components of the pressure gradient/density terms were individually altered and the first four Fourier components of the output horizontal neutral wind velocities were calculated. This analysis was done for a latitude of 51°N , for an altitude of 300 km for June, 1959. A three-dimensional representation of the changes in magnitude of output neutral wind velocities per change in magnitude of the input pressure gradients/density is shown in Figure 35. If the ionospheric system were linear, no mode coupling would be present and only terms on the output West-East-2 diagonal would be present. This diagonal is indicated by the arrow. The terminology is that zero represents the constant term, one the first harmonic, etc. From this figure it can be seen that the cross directional coupling and harmonic generation effects are substantial.

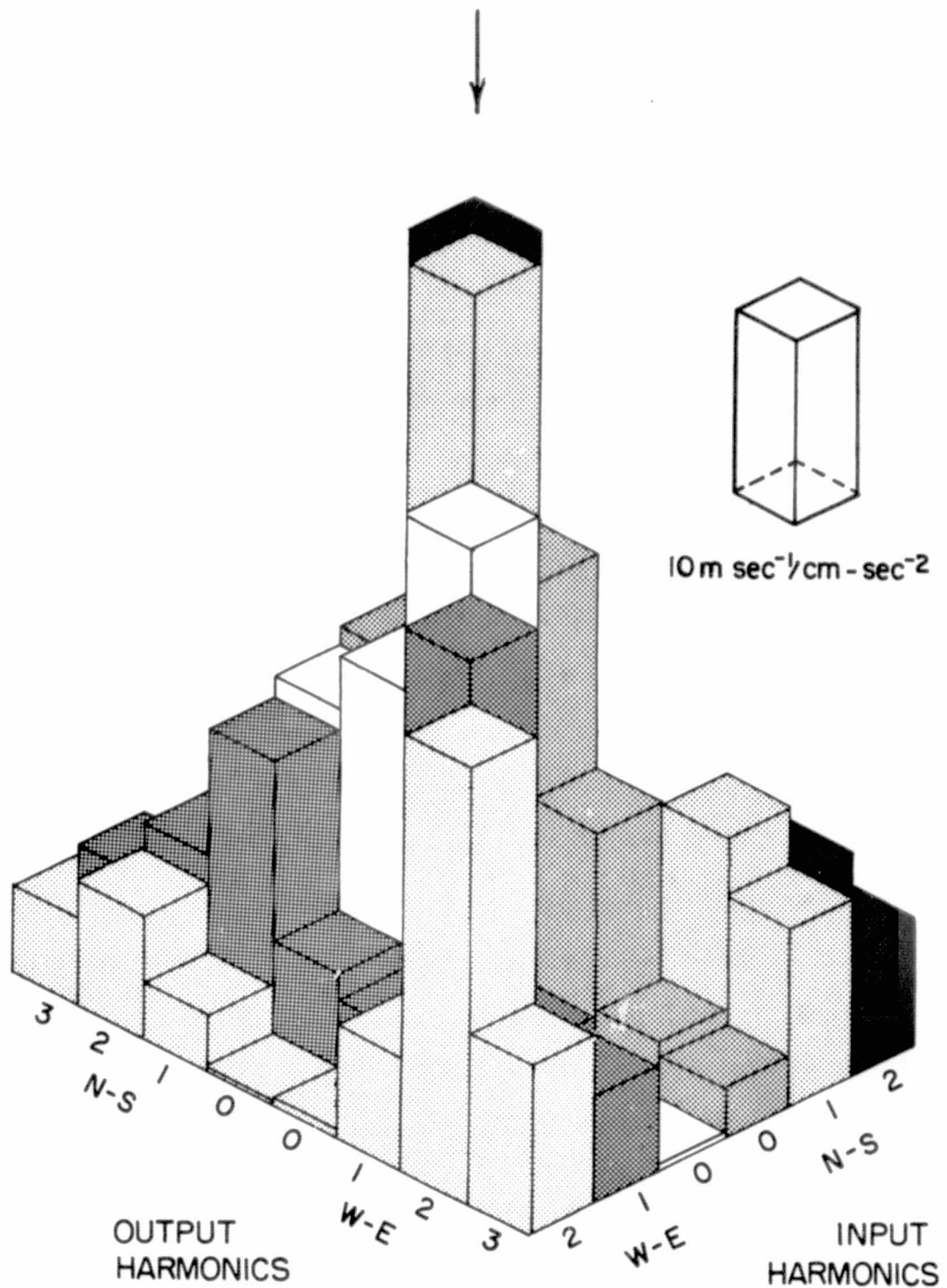


Figure 35. Magnitude Changes in the Fourier Components of the Neutral Wind Velocity per Magnitude Change in the Fourier Components of the Pressure Gradient/Density at 300 km

Though these effects are largest for the constant terms, cross directional coupling and harmonic generation are commonly greater than 20 percent for the first and second harmonic inputs. The maximum sensitivity is caused by the first harmonic component of the North-South term. This sensitivity shows a change of 35.3 m-sec^{-1} in the first harmonic of the North-South neutral wind component for a change of 1 cm-sec^{-2} in first harmonic component of the North-South pressure gradient/density term.

In Figure 36 the sensitivities of the magnitude changes in the Fourier components of the neutral wind components per phase change in the Fourier components of the pressure gradient/density terms are presented. The results of Figure 35 are superimposed on this figure. It is seen that input phase changes produce cross directional coupling and harmonic generation which can be factors of two to three times greater than the harmonic which would be normally altered in a linear system. The first harmonic West-East input causes the largest sensitivities, producing a cross-directional coupling into the first harmonic of the North-South output neutral velocity with a sensitivity of 10.3 m-sec^{-1} per change of one hour in phase. This term also produces large effects on both the North-South and West-East constant neutral wind terms.

The results presented here very definitely support Volland and Mayr's (1973) opinion that mode coupling is a significant problem. Volland and Mayr pointed out that the mode coupling problem is not yet solved. Recently, Harris and Mayr (1974) included mode coupling and some of the non-linear terms in their analysis and showed that their effects can be substantial, especially at F2 region heights.

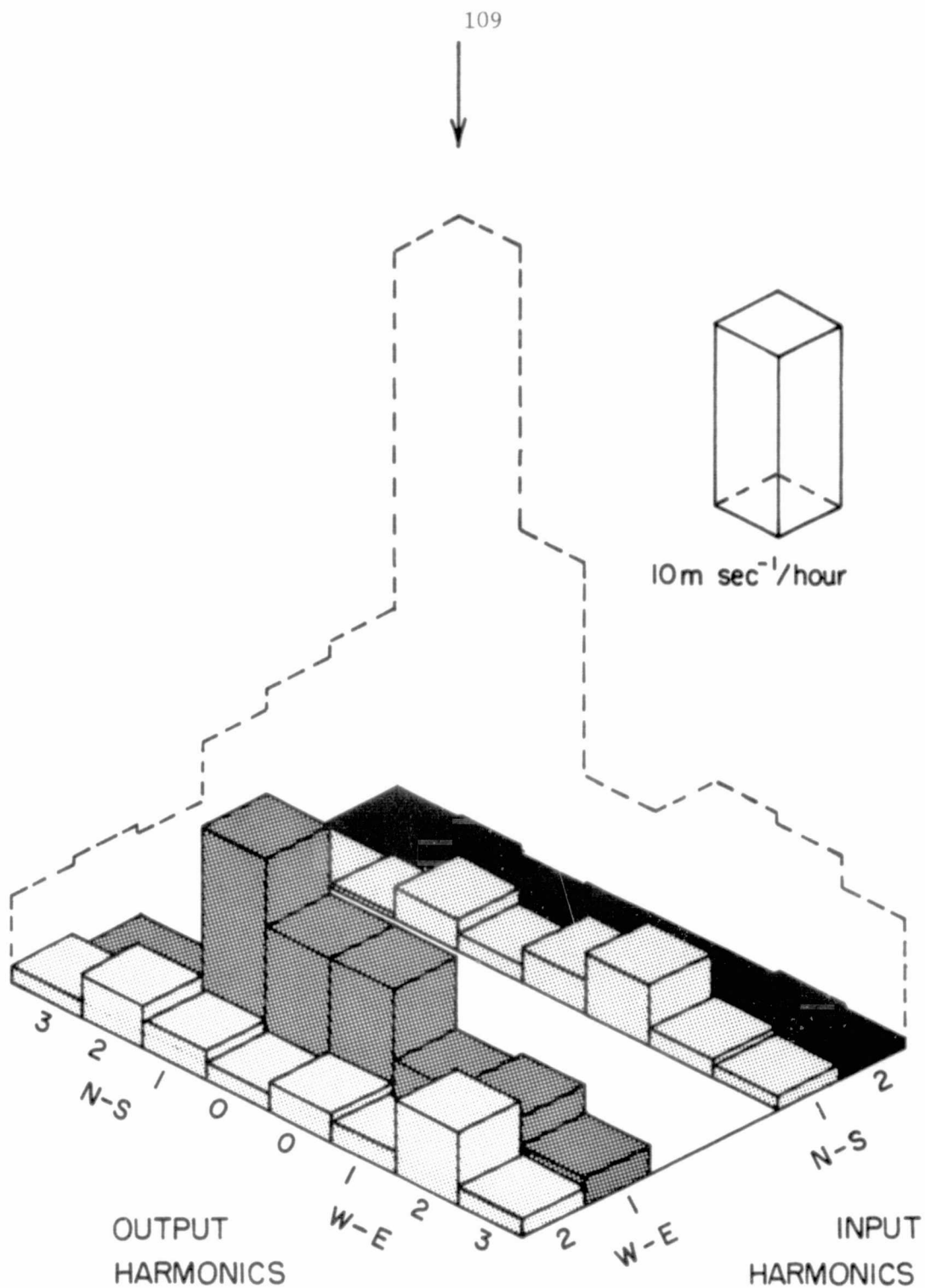


Figure 36. Magnitude Changes in the Fourier Components of the Neutral Wind Velocity per Phase Change in the Fourier Components of the Pressure Gradient/Density at 300 km

It is suggested that the sensitivity factors determined by this study be used by studies which employ spherical harmonics and perturbation theory to sort out the more important harmonic generation and cross directional coupling effects.

Though the sensitivity factors calculated by this study are only valid for a limited range of temperatures and densities and for specific geographical conditions, they are useful in determining the generalized effects that the pressure gradients calculated from the various neutral atmospheric models (see Section 4.2) have on the horizontal neutral winds. Calculations using these sensitivity factors and the pressure gradient/density from the various models were made and the results show neutral wind variations of hundreds of meters per second. These results are in agreement with those given in Section 4.2.

CHAPTER V

CONCLUSIONS

5.1 General Discussion

Methods were developed for relating the electron density profile to the vertical component of the ion velocity and the vertical component of the ion velocity to the equivalent meridional wind. In these developments the number density continuity equation for ions and the momentum equations for electrons and ions were employed. It was shown that these relationships were functions of the production, loss, time derivative and diffusion terms. These terms could be calculated by employing a static neutral atmospheric model and the electron density profile. Because the data available for the study provided information only at the F2 region peak, it was necessary to derive a method for estimating the density profile from HmF2 and NmF2 data.

Several methods for estimating the shape of the electron density were tested and it was shown that the "model method" reasonably reproduced results obtained from a sophisticated theoretical ionospheric model. This "model method" was also used to calculate vertical ion drifts and these were shown to be in good agreement with results from incoherent scatter measurements.

In analyzing the terms which contribute to the vertical drift and equivalent meridional wind calculations it was suggested that the accuracy of the neutral atmospheric model is important especially with regard to the production and loss terms. The magnitude of this effect increases with decreasing altitude because the magnitudes of these terms increase and because the uncertainties at the lower

boundary are substantial. This fact was demonstrated in comparisons with results from the Millstone Hill incoherent scatter facility.

The equivalent meridional winds determined from ionosonde stations having different magnetic declinations and inclinations were used to calculate the horizontal components of the neutral winds and the electric fields. The accuracy with which these components could be resolved was dependent upon the magnetic field geometry of the station configurations used. Because data from a limited number of ionogram stations were available, only certain components of the neutral winds and electric fields could be calculated with a reasonable degree of accuracy.

By assuming that the electric fields were zero or known from other sources, some interesting results concerning the neutral wind components were obtained. In general, the neutral winds were shown to be much smaller than those calculated by the Stubbe ionospheric model but of the same order as winds determined from incoherent scatter measurements and those calculated from Volland and Mayr's (1973) theoretical neutral atmospheric model. It should be noted, however, that the magnitudes of winds calculated from theoretical neutral atmospheric models are very dependent upon the ion density profiles which are employed by these models for the purpose of evaluating the ion drag term.

The effect of electric fields on the neutral wind calculations was determined by employing two models of theoretical electric fields calculated by Matsushita (1971) and measured electric fields by Kirchhoff and Carpenter (1974). Though these theoretical fields altered the neutral wind calculations by $\pm 50 \text{ m-sec}^{-1}$, they could not

account for the differences between the winds calculated by the methods of this study and those calculated from the current international reference atmosphere CIRA (1972). The measured electric fields were shown to alter the neutral wind calculation by $\pm 20 \text{ m-sec}^{-1}$.

For the White Sands/Grand Bahama I/Okinawa/Puerto Rico configuration the magnetic field geometry was such that reasonably accurate North-South neutral winds and West-East electric fields could be calculated simultaneously. The North-South winds were shown to exhibit the midnight reversal effect which is a distinguishing feature of the North-South wind measured by the Arecibo incoherent scatter facility. The West-East electric fields agreed rather well with the theoretical fields of Stening (1973).

After generating the neutral winds, new sets of pressure gradients which, when used in the theoretical ionospheric models with the densities from the static neutral atmospheric models, would reproduce the desired wind field were determined. These sets of pressure gradients were compared with pressure gradients calculated from various static neutral models and it was shown that the variations between the new sets of pressure gradients and those determined from the models were of the same order of magnitude as the variations between pressure gradients calculated from various neutral static models. This is not surprising since the spatial resolution of these models is known to be inadequate and the temperature distributions are not correct.

A set of sensitivity factors was calculated by relating the harmonics of the neutral winds to the harmonics of the pressure gradient/density driving force. At first it was intended to use these

sensitivity factors to produce more accurate pressure gradient calculations. However, the large coupling coefficients which were calculated made it impractical to do so.

The sensitivity factors showed both harmonic generation and cross directional coupling effects. These effects are important to studies which employ linearized perturbation theory and spherical harmonics.

5.2 Specific Conclusions

The results of this study are as follows:

- 1) The vertical component of the ion velocity was obtained by integrating the continuity equation for ions and the equivalent wind in the direction of the magnetic meridian was determined by using the momentum equations for ions and electrons and the vertical ion velocity. The electron density profile which is necessary for the use of these relationships was estimated by solving the ion continuity equation in an iterative manner and normalizing the ion density profiles to $HmF2$ and $NmF2$ values obtained from ionograms after each iteration. Equivalent winds in the direction of the magnetic meridian calculated in this manner were shown to have an accuracy of the order of $\pm 15 \text{ m-sec}^{-1}$.
- 2) Neutral winds and electric fields were calculated by employing the equivalent wind in the direction of the magnetic meridian at a number of locations and by making use of the variability of the declination and inclination of the earth's magnetic field. The accuracies of the neutral wind and electric field

components are dependent on the geometries of the magnetic field at the ionosonde stations. For the stations used in this study the North-South neutral wind could be determined to an accuracy of less than $\pm 15 \text{ m-sec}^{-1}$ and the West-East neutral wind could at best be determined to an accuracy of $\pm 35 \text{ m-sec}^{-1}$. The neglect of electric fields in the neutral wind calculations can introduce uncertainties of the order of $\pm 50 \text{ m-sec}^{-1}$ if the electric fields are of the order of magnitude of the theoretical values given by Matsushita (1971) and $\pm 20 \text{ m-sec}^{-1}$ if the electric fields are of the order of magnitude of the measured fields given by Kirchhoff and Carpenter (1974). For the most part the results calculated by this study are in general agreement with results from measurements and results from theoretical models.

- 3) Pressure gradients calculated by this study were shown to differ with, but be of the same order of magnitude as, those calculated from various static neutral atmospheric models. In view of the uncertainties in the models themselves, it is suggested that pressure gradients calculated by methods shown in this study can be used to improve the accuracies of the static neutral atmospheric models. A set of sensitivity factors relating the pressure gradients to the neutral winds were calculated and these were shown to exhibit significant harmonic generation and cross directional coupling effects.

5.3 Contributions to Ionospheric Research

The Stubbe ionospheric model forms the basis for many of the theoretical calculations performed in this study. It was shown by Stubbe (1973) that the lower boundary conditions of the CIRA 1972 reference atmosphere and the F region equivalent winds could be adjusted to make calculated HmF2 and NmF2 values agree with those obtained from measurements. In this study a method was developed for determining the equivalent winds directly from HmF2 and NmF2 measurements by approximating some of the procedures used in the Stubbe ionospheric model (see Section 2.3 and Appendix B). Because this method requires approximately six times less computational time than the iterative approach used by Stubbe and yields results which are equivalent to those obtained by Stubbe, it is now economically feasible to calculate equivalent winds on a world-wide basis by employing ionosonde data.

By using the equivalent winds at a number of locations having different magnetic inclinations and declinations, new methods for determining the horizontal components of the neutral winds and electric fields were developed. These methods are very important to studies in aeronomy because they demonstrate a means for studying neutral winds and electric fields on a global basis. Although the results obtained by this study were presented to demonstrate the feasibility of the methods, the uncertainties associated with the neutral winds and pressure gradients calculated at 50°N latitude are small and these results yield believable geophysical information.

5.4 Suggestions for Future Research

- 1) More ionosonde data must be reduced so that the methods developed in this study can be applied on a world-wide basis. Currently ionograms from approximately 50 ionosonde stations having the desired magnetic field properties are being reduced at The Pennsylvania State University.
- 2) The lower boundary conditions of the neutral atmospheric models must be altered on a diurnal basis. Stubbe (1973) adjusted the equivalent horizontal wind and the atomic oxygen boundary at 125 km in an iterative manner in order to match measured and calculated HmF2 and NmF2 values at 1200 hours. The employment of the Stubbe modified Jacchia model, which resulted from this procedure, improves the predictions of theoretical ionospheric models at times near 1200 hours, but at other times the theoretical predictions are still not adequate. In determining the diurnal variation of the lower boundary, the methods described in Chapter II to determine the equivalent horizontal wind are more economical than the manual method employed by Stubbe.
- 3) The "model method" developed by this study for the estimation of the ion density profile resulted from a theoretical solution of the ion continuity equation and from the employment of HmF2 and NmF2 data from ionograms. Since ionograms provide electron density information up to the F2 peak, an effort should be made to utilize this data in order to calculate a more accurate electron density profile. Presently, the lack

of an adequate underlying ionization correction severely limits the accuracy of the electron densities calculated below the peak by ionogram reduction programs.

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APPENDIX A

PHOTON FLUXES, IONIZATION AND ABSORPTION CROSS SECTIONS, REACTIONS, AND PRODUCTION AND LOSS RATES

The photon fluxes, ionization and absorption cross sections, reactions, and production and loss rates given here are those used by Stubbe (1973). The asterisks signify the production and loss rates which are characteristic of the approximations made in this study.

The values shown are in units $\text{\AA}$ for λ , $10^9 \text{ photons-cm}^{-2} \text{ sec}^{-1}$ for Φ_∞ and 10^{-18} cm^2 for σ .

Table 8
Photon Fluxes

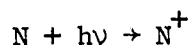
λ	Φ_∞^{150}	$\sigma^{(a)}(O_2)$	$\sigma^{(a)}(N_2)$	$\sigma^{(a)}(O)$	$\sigma^{(i)}(O_2)$	$\sigma^{(i)}(N_2)$	$\sigma^{(i)}(O)$
31- 165	1.85	1.1	.46	.55	1.1	.46	.55
165- 205	3.7	6.8	2.2	3.4	6.8	2.2	3.4
205- 280	4.5	13.4	3.6	6.7	13.4	3.6	6.7
280- 370	10.3	21.0	7.7	9.3	21.0	7.7	9.3
370- 460	.63	24.0	17.0	11.1	23.0	17.0	11.1
460- 630	4.7	29.0	24.0	12.9	26.0	22.0	12.9
630- 796	2.4	28.0	25.0	3.4	16.8	17.4	3.4
796- 911	8.3	12.6	5.0	3.2	7.0	0	3.2
911-1027	11.6	5.5	5.0	0	3.0	0	0

The photon flux dependence on solar activity is given by

$$\Phi_\infty = \Phi_\infty^{150} \left[1 + 5 \times 10^{-13} (S_{10.7} - 150) \right]$$

The reaction or production rates are as follows:

Reaction	Reaction or Production Rate in Unit $\text{cm}^3 \text{sec}^{-1}$
$\text{O}_2^+ + \text{NO} \rightarrow \text{NO}^+ + \text{O}_2$	$k_r(1,1) = 8 \times 10^{-10}$
$\text{O}_2^+ + e^- \rightarrow \text{O} + \text{O}$	$k_r(1,2) = 2.2 \times 10^{-7} (300/T_e)^{0.65}$
$\text{O}_2 + h\nu \rightarrow \text{O}_2^+$	$P_{\text{O}_2}^{(p)}$ see Equation 2.2
$\text{N}_2^+ + \text{O}_2 \rightarrow \text{O}_2^+ + \text{N}_2$	$k_r(2,1) = 5 \times 10^{-11}$
$\text{N}_2^+ + \text{O} \rightarrow \text{NO}^+ + \text{N}$	$k_r(2,2) = 1.5 \times 10^{-10}$
$\text{N}_2^+ + e^- \rightarrow \text{N} + \text{N}$	$k_r(2,3) = 2.0 \times 10^{-7} (300/T_e)^{0.33}$
$\text{N}_2 + h\nu \rightarrow \text{N}_2^+$	$P_{\text{N}_2}^{(p)}$ see Equation 2.2
$\text{O}^+ + \text{O}_2 \rightarrow \text{O}_2^+ + \text{O}$	$k_r(3,1) = 1.6 \times 10^{-11}$
$\text{O}^+ + \text{N}_2 \rightarrow \text{NO}^+ + \text{N}$	$k_r(3,2) = 6 \times 10^{-12}$
$\text{O}^+ + \text{H} \rightarrow \text{H}^+ + \text{O}$	$k_r(3,3) = 2 \times 10^{-9}$
$\text{O} + h\nu \rightarrow \text{O}^+$	$P_{\text{O}}^{(p)}$ see Equation 2.2
$\text{H}^+ + \text{O} \rightarrow \text{O}^+ + \text{H}$	$k_r(4,1) = (8/9) \times 2 \times 10^{-9}$
$\text{He}^+ + \text{N}_2 \rightarrow \text{He} + \text{N}^+ + \text{N}$	$k_r(5,1) = 7.5 \times 10^{-10}$
$\text{He}^+ + \text{N}_2 \rightarrow \text{He} + \text{N}_2^+$	$k_r(5,2) = 7.5 \times 10^{-10}$
$\text{He}^+ + \text{O}_2 \rightarrow \text{He} + \text{O}^+ + \text{O}$	$k_r(5,3) = 1.0 \times 10^{-9}$
$\text{He} + h\nu \rightarrow \text{He}^+$	$P_{\text{He}}^{(p)} = 4 \times 10^{-8} n(\text{He})$
$\text{NO}^+ + e^- \rightarrow \text{N} + \text{O}$	$k_r(6,1) = 4 \times 10^{-7} (300/T_e)$
$\text{N}_2^+ + \text{O} \rightarrow \text{NO}^+ + \text{N}$	$k_r(7,1) = 2.5 \times 10^{-10}$
$\text{N}_2^+ + \text{O}_2 \rightarrow \text{O}_2^+ + \text{N}_2$	$k_r(7,2) = 3.5 \times 10^{-10}$



$$P_N^{(p)} = .21 P(N_2)$$

The production and loss rates for the individual ion species are as follows:

$$P(O_2^+) = P_{O_2}^{(p)} + \left[k_r(2,1)n(N_2^+) + k_r(3,1)n(O^+) + k_r(7,2)n(N^+) \right] n(O_2) \quad (A1)$$

$$L(O_2^+) = \left[k_r(k,1)n(NO) + k_r(1,2)n_e \right] n(O_2^+) \quad (A2)$$

$$P(N_2^+) = P_{N_2}^{(p)} + k_r(5,2)n(He^+)n(N_2) \quad (A3)$$

$$L(N_2^+) = \left[k_r(2,1)n(O_2) + k_r(2,2)n(O) + k_r(2,3)n_e \right] n(N_2^+) \quad (A4)$$

$$P(O^+) = P_O^{(p)} + k_r(4,1)n(H^+)n(O) + k_r(5,3)n(He^+)n(O_2) \quad (A5)$$

$$*P(O^+) = P_O^{(p)} \quad (A6)$$

$$L(O^+) = \left[k_r(3,1)n(O_2) + k_r(3,2)n(N_2) + k_r(3,3)n(H) \right] n(O^+) \quad (A7)$$

$$*L(O^+) = \left[k_r(3,1)n(O_2) + k_r(3,2)n(N_2) \right] n(O^+) \quad (A8)$$

$$P(H^+) = k_r(3,3)n(O^+)n(H) \quad (A9)$$

$$L(H^+) = k_r(4,1)n(O)n(H^+) \quad (A10)$$

$$P(He^+) = P_{He}^{(p)} \quad (A11)$$

$$L(He) = \left[(k_r(5,1) + k_r(5,2))n(N_2) + k_r(5,3)n(O_2) \right] n(He^+) \quad (A12)$$

$$P(NO^+) = k_r(1,1)n(O_2^+)n(NO) + k_r(2,2)n(N_2^+)n(O) + k_r(3,2)n(O^+)n(N_2) + k_r(7,1)n(N^+)n(O_2) \quad (A13)$$

$$L(\text{NO}^+) = k_r(6,1)n_e n(\text{NO}^+) \quad (\text{A14})$$

$$P(\text{N}^+) = P_N^{(p)} + k_r(5,1)n(\text{He}^+)n(\text{N}_2) \quad (\text{A15})$$

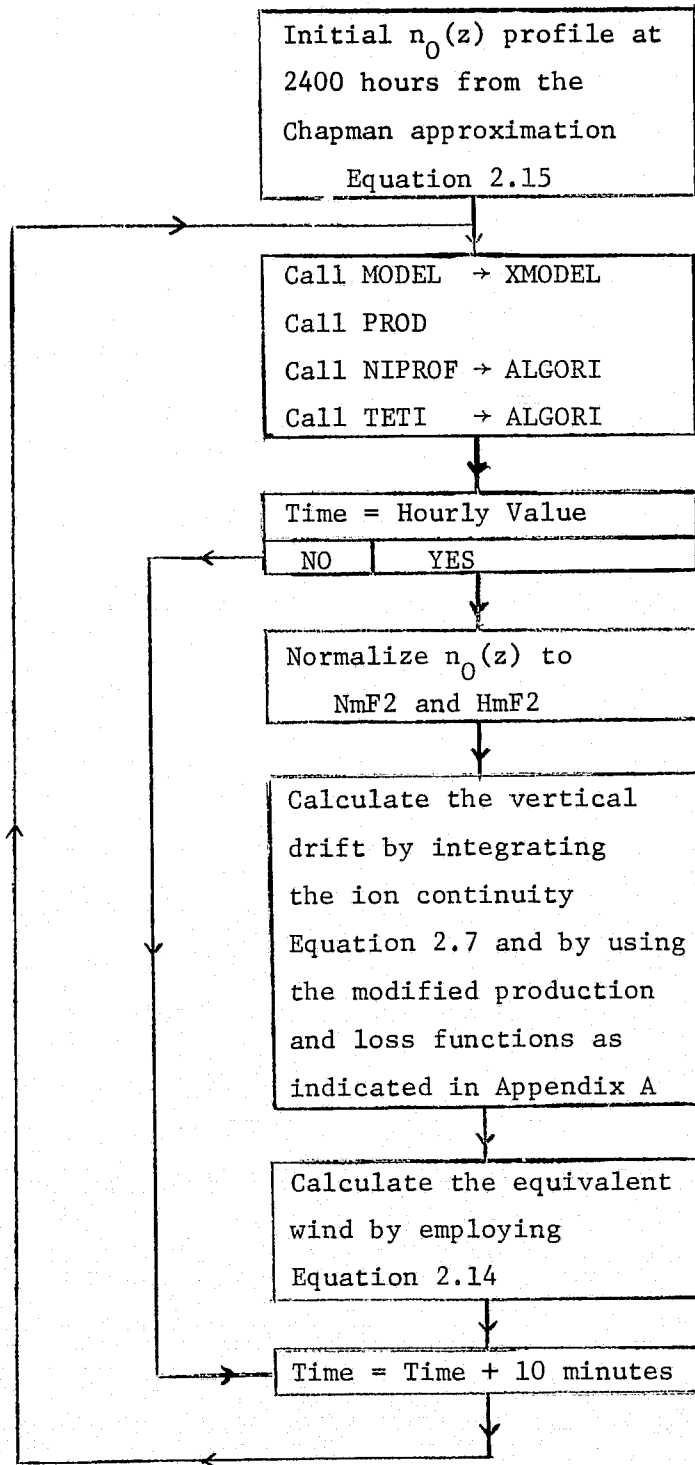
$$L(\text{N}^+) = \left[k_r(7,1) + k_r(7,2) \right] n(\text{O}_2) n(\text{N}^+) \quad (\text{A16})$$

APPENDIX B

DESCRIPTION OF THE STUBBE MODEL AND THE "MODEL METHOD"

The Stubbe (1973) ionospheric program and subprograms form the basis for many of the theoretical calculations performed in this study. This ionospheric model employs the static neutral atmospheric model of Jacchia (1971) or the Stubbe (1973) modified Jacchia model and solves the equations of motions for neutrals, ions and electrons, the number density continuity equations for ions and electrons, and the heat conduction equation for ions and electrons. Because this system of equations is coupled and non-linear, the solutions are obtained by numerically solving in succession the differential equations governing each ionospheric process and by utilizing the results from the previous solution. The assumptions governing the equations and boundary values have been given by Stubbe (1970, 1973). The model utilizes the neutral constituents O_2 , N_2 , O , H , H_2 , NO and N and calculates the ionic constituents O_2^+ , N_2^+ , O^+ , H^+ , He^+ , NO^+ and N^+ , the electron densities, horizontal neutral winds, and electron and ion temperatures as a function of altitude and local time. Vertical ion velocities for O^+ and H^+ ions are also calculated and it is assumed that the electron velocity is a weighted average of these ion velocities.

This study utilizes the MODEL, XMODEL, PROD, NI3457, TETI and ALGORI subprograms (see Appendix C) of the Stubbe model to calculate the atomic oxygen ion profile, vertical drifts and equivalent winds and this procedure is shown in the flow chart on the following page.



APPENDIX C

DESCRIPTION OF THE SUBPROGRAMS

1. MODEL calculates the neutral North-South and West-East pressure gradients and calls XMODEL.
2. XMODEL calculates the neutral atmospheric densities and temperature from either the Jacchia (1971) static neutral atmosphere or the Stubbe (1973) modified Jacchia model which is basically the Jacchia model with modified boundary conditions (see Section 1.2.8). The inputs which must be supplied to XMODEL are latitude, day of the year, time, the daily and monthly average of the $S_{10.7}$ index and the K_p index.
3. PROD calculates the photoionization production of O^+ , O_2^+ and N_2^+ ions by using Equation 2.2, the neutral densities calculated by XMODEL and the photon fluxes given in Appendix A.
4. NI3457 calculates the coefficients and calls ALGORI in order to solve the differential ion continuity equations for O^+ , H^+ , He^+ and N^+ ions.
5. NIPROF, a modified version of NI3457, calculates the atomic oxygen ion density profiles. The production and loss functions are modified as shown in Appendix A and the vertical drift which is induced by neutral winds or electric fields is set equal to zero.
6. TETI calculates the coefficients for the heat conduction differential equations for T_e and T_i and calls ALGORI which performs the numerical solution.
7. ALGORI uses finite differences to perform the numerical operations which are necessary to solve the first order in time and second order in space differential equations.