Radar Measurements of High-Latitude Ion Composition between 140 and 300 km Altitude

J. D. Kelly

Vincent B. Wickwar

Utah State University

Follow this and additional works at: https://digitalcommons.usu.edu/physics_facpub

Recommended Citation

Radar Measurements of High-Latitude Ion Composition Between 140 and 300 km Altitude

JOHN D. KELLY AND VINCENT B. WICKWAR

SRI International, Menlo Park, California 94025

The Chatanika radar has been used to measure the ratio of atomic (O⁺) ions to molecular (O₂⁺, NO⁺) ions in the high-latitude ionosphere. The radar results agreed well with simultaneous in situ rocket data, giving confidence in the radar method of deducing ion composition. Measurements made over long periods of time show seasonal variations, diurnal variations, and variations due to auroral processes. The transition altitude, where the number densities of atomic and molecular ions are equal, is a convenient parameter for describing the composition variation with altitude or 'composition altitude profile.' The transition altitude occurs at ~190 km at night and ~170 km during the day, in agreement with midlatitude results. During the winter the daytime transition altitude is 15 km lower than in summer, a seasonal variation similar to that at midlatitudes. Energetic particle precipitation results in the lowering of the transition altitude by 10 km in one case when energetic particles deposited ~20 ergs/cm² s in the atmosphere. The largest variations in ion composition were found during periods of large joule heat input resulting from electric fields on the order of 50 mV/m. The transition altitude increased by 50 km in a case where the joule heat input rate was 30 ergs/cm² s. These observations were compared to calculations from a simple steady state model involving the principal constituents and reactions. The results indicate that the transition altitude during particle precipitation is most influenced by the increased ion production. There do not appear to be significant effects from possible increases of N₂ vibrational temperature. A number of interrelated effects contribute to the increase in transition altitude during joule heating. The most important effect is the electric field contribution in raising the effective ion temperature. In addition, it appears that increased N₂ density is also required to account for the observed change.

1. INTRODUCTION

Theoretical studies of the high-latitude ionosphere predict that there should be large changes in ion composition as a result of energy and momentum input from the magnetosphere during auroral activity [Schunk et al., 1975, 1976; Jones and Rees, 1973]. Rocket and satellite measurements have indeed shown that during magnetic activity the relative abundance of light (O⁺) and heavy (NO⁺ and O₂⁺) ions can be very different from that of the midlatitude ionosphere. One of the most striking variations was measured by the ISIS 2 satellite during the solar proton event of August 1972. During the peak of that magnetic storm, the satellite measured large concentrations of NO⁺ at 1400 km, as compared to the trace amounts that are normally measured at that altitude [Hoffman et al., 1974].

Incoherent scatter radars at midlatitudes have proven very useful for the study of the altitude and time variation of ion composition [Evans, 1967; Evans and Cox, 1970; Wand and Perkins, 1970; Alcyade et al., 1974]. In this paper we adapt and extend the radar analysis techniques of ion composition used previously at midlatitudes, to the auroral region. As a result, using the Chatanika radar (described by Leadabrand et al. [1972]), we are able to make the first long-term measurements of ion composition in the auroral region. The results found in this study are consistent with both theory and previous experiments. We expect that future, more extensive studies of the composition can be used to enable us to understand the variations of the high-latitude neutral atmosphere on both short (substorm) and long (seasonal) time scales.

2. TECHNIQUE

Ground-based incoherent scatter radar is well suited for determining both temporal and spatial variations of ionospheric parameters such as the electron density, ion velocity, electron and ion temperatures, and the ion mass, all as functions of altitude. The electron density is derived from the received backscattered power, the ion drift velocity from the Doppler shift, and the electron-to-ion-temperature ratio (T_e/T_i) and the ion-temperature-to-ion-mass ratio (T_i/m) from the autocorrelation function (ACF) of the received signal [Farley, 1969; Dougherty and Farley, 1960; Hagfors, 1961; Moorcroft, 1964; Waldteufel, 1971; Evans, 1969].

The temperatures and composition are found by means of a least squares fitting of theoretical ACF's with the measured ACF's [Wickwar, 1974; Waldteufel, 1971]. When one examines the autocorrelation functions computed from the backscattered signals, one finds that, since the ACF functional dependences on temperature and ion composition are correlated, there are a number of temperatures that could be inferred depending on the ion composition. As a result, one of the three variables (T_e, T_i, m) must be independently measured or assumed before the other two can be determined. In the altitude region covered by this study, the ions considered are atomic (O⁺) and molecular (NO⁺ and O₂⁺).

Fortunately, altitude regimes exist where the ion composition is known and unchanging, e.g., between 85 and 130 km molecular ions dominate and above much of the F region atomic oxygen ions dominate. It is in the transition region, where T_e, T_i, and m, are all changing, that difficulties arise in the data analysis.

Midlatitude Case

In the transition region, the technique used by Evans [1967] for midlatitude studies requires modeling the T profiles. Evans made the following assumptions based on rocket and satellite data: (1) at low altitudes (<130 km) the dominant ions are molecular and at high altitudes (>250 km) the domi-
nant ion is $O^+$; (2) at altitudes below ~300 km the ions are in thermal equilibrium with the neutrals.

The model used by Evans for $T_e$ corresponds to a neutral temperature profile from Cira (1965). The particular $T_e$ profile is chosen by using values of $T_e$ measured at altitudes (130 and 250 km) where one can safely assume the ion composition to be entirely molecular and atomic, respectively. Composition profiles can then be determined because $T_e (T_i = T_e)$ is known throughout the altitude region where the ion composition is changing from molecular to atomic. The ion composition results are, therefore, somewhat dependent on the neutral temperature model chosen.

**High-Latitude Case**

At high latitudes, the magnetic field lines couple the ionosphere to the magnetosphere, giving rise to energy input into the ionosphere from energetic particle precipitation and from joule heating by the convection electric field. Theory and observation [Walker and Rees, 1968; Rees and Walker, 1968] show that particle precipitation leads to an increase of the electron temperature, and joule heating leads to an increase of the ion temperature. Recent observations [Schlegel and St. Maurice, 1981; Wickwar et al., 1981] show that coincident with joule heating, there is an increase of the $E$-region but not of the $F$ region electron temperature.

The midlatitude technique for finding the ion composition in the transition region can be applied when the ions are not heated above the neutral temperature. Therefore, the technique can still be applied to high-latitude data during periods when the effects of joule heating are small. Indeed, we have confirmed this with a comparison to a simultaneous rocket measurement of in situ ion composition (section #3). However, the midlatitude technique cannot be applied during joule heating events that are large enough to significantly increase the ion temperature above the neutral temperature. For this situation we have developed another technique. Instead of constraining $T_e$ to a model $T_e$ profile, we constrain $T_e$ at each altitude to have a value consistent with the values measured before and after the heating event. Then the least squares fitting procedure is used to solve for $T_e$ and $m_i$.

The rationale for adopting this procedure is based on the following:

1. When $F$ region temperatures are deduced under the assumption of unchanging $O^+$ composition, a large apparent decrease in $T_e$ is found during periods of joule heating. This apparent decrease is an artifact of the constant $O^+$ composition assumption. The assumption of atomic ions rather than a mixture of atomic and molecular ions results in an underestimate of both the temperature ratio, $T_e/T_i$, and the ion temperatures. As a result, $T_e$ is substantially underestimated.

2. Real decreases in the electron temperature can arise from four principal sources: decrease of the solar EUV, decrease of low-energy particle precipitation, increase of the electron density, and an increase of neutral density. The first three sources can be monitored and their effects evaluated; the solar EUV through knowledge of the solar zenith angle, the soft particle flux through the particle energy deposition, and the electron density through direct measurement. In fact, the effects of the second and third sources are in opposite directions. For example, a decrease in soft particle flux will decrease the thermal energy input and also have the effect of reducing the electron density, which, in turn, leads to a reduction in the electron-neutral loss rate. As a result, the change in $T_e$ is small. Although an increase in neutral density will increase the loss rate, in the cases we observed, neutral density increases were modest and did not compensate for the terms that would cause heating. Constraining $T_e$ to vary slowly and smoothly between its preheating and postheating event values accounts to first order for these possible real sources of $T_e$ decrease.

In fact, one would expect joule heating to result in increased electron temperature in most cases and thus our analysis technique results in a lower bound in both electron and ion temperatures and the percentage of molecular ions. To evaluate the effects of joule heating on $T_e$, calculations were performed under a number of different conditions using the energy loss and input rates from Schunk and Nagy [1978], Stubbe and Varnum [1972], Prasad and Furman [1973], and Hoegy [1976], and assumed that the external energy sources—low energy particle precipitation or solar EUV—were constant.

The changes during periods of joule heating that would increase the electron temperature are: decreased electron density, increased ion temperature, and conversion of atomic ions to molecular ions. A decrease of electron density by up to a factor of 2 is often observed. This has a significant effect because it reduces all electron-neutral loss rates by this amount, tending to increase $T_e$. An increase in the ion temperature of 300 to 1000 K will greatly reduce the energy loss from electrons to ions and will often lead to energy input to the electrons from hotter ions. This effect is much more important at the higher altitudes and for the higher electron densities. Finally, the conversion of atomic to molecular ions is comparatively unimportant.

The joule heating-induced change that would decrease the electron temperature is increased neutral density. The effect of an increase in neutral density is considered because it has often been reported in connection with the auroral region [e.g., Prölls, 1980]. Even a fourfold increase in the $N_2$ density would not decrease the electron temperature sufficiently to fully compensate for the temperature increase sources considered above. Increases in the $O_2$ density would further increase the electron-neutral losses, but are much less important. Increases in the $O$ density would be important, but have not been observed. Thus, in general, the electron temperature should, in theory, increase during joule heating and our assumption of a constant $T_e$ does provide a lower bound for the proportion of molecular ions and for the transition altitude.

The procedures we follow to determine the ion composition at high latitudes and the choice between procedures are summarized in the flowchart in Figure 1.

3. Observations

The measurements were made with the Chatanika radar [Leadabrand et al., 1972] that is located near Fairbanks, Alaska, at 65°N geographic and geomagnetic latitude. The autocorrelation function measurements were made using both a 320-μs and a 160-μs transmitted pulse at 1290 MHz together with hardware correlators. The system is described and references given in Kofman and Wickwar [1980]. The antenna was cycled between three positions or kept scanning in azimuth during the observations so as to determine the vector ion velocities (electric fields) and the joule heating rate [Brekke and Rino, 1978] in addition to $T_e$, $T_i$, and ion composition.

The ion composition varies from a mixture of $NO^+$ and $O_2^+$...
in the lower E region to O\(^+\) in the upper F region. We usually describe it in terms of the fraction that is made up of atomic ions. In addition, a single parameter that succinctly describes the composition is the transition altitude, the altitude where the ion composition is half atomic and half molecular. The variation of this parameter with time, season, and auroral activity (both particle precipitation and joule heating) was studied.

The transmitted pulses, particularly the 320-\(\mu\)s pulse, are longer than optimum at the low-altitude end of the altitude scale, although they do have the distinct advantage of yielding good signal-to-noise ratios. There is, however, the possibility of small, systematic errors in the altitude because the temperature averaged over the sampled region may not be the temperature at the midpoint of the pulse. The difference depends on the gradients of the measured parameters (\(T_e\), \(T_n\), and composition), the weighting effect of the electron density gradient, the pulse length, and the details of the measurement technique. Using gradients measured during day and night conditions, this source of error was evaluated and the resulting bias in the transition altitude calculation was found to be less than 5 km. We consider this acceptable because the geophysical effects that we are observing vary the transition altitude by 20 to 50 km.

A set of three 24-hour periods supplemented by one shorter period were chosen to study the variation of composition with time, season, and auroral activity. The days examined are given in Table 1 with local magnetic activity indices and the solar zenith angle extremes.

**Comparison With In Situ Ion Composition Measurements**

On February 28, 1976, a Nike Tomahawk rocket carrying a University of Texas at Dallas (UTD) payload was launched from Poker Flat, Alaska, 3 km from the Chatanika radar. The payload instrumentation included probes for measuring ion and electron densities and a positive ion mass spectrometer. The densities were measured on the upleg and the ion composition on the downleg.

The radar was operated before, during, and after the flight in order to obtain background as well as simultaneous data. The antenna was operated in the three-position mode described earlier and the pulse length used was 160 \(\mu\)s, allowing a 24-km range resolution. The background data include both joule and particle energy input rates, shown in Figure 2. During the rocket flight, the joule heat input rate (3 ergs/cm\(^2\) s) was relatively small, causing an enhancement in \(T_e\) of less than 100 K. Hence, the \(T_i = T_n\) method was used to compute composition.

The radar beam was pointed along the rocket upleg trajectory from 1216 to 1219 UT. The electron density profile measured during that period agreed well with the profile measured by the rocket. The ion temperature data were averaged over those four minutes and an exospheric temperature of 900 K was computed. The radar data were analyzed for composition using two \(T_n\) profiles, one from Cira (1972) and one from Banks and Kockarts [1973]. The composition profiles and the one measured by the rocket on the downleg are shown in Figure 3. Because of the similar density profiles observed by the radar at three azimuths and the uniform postsubstorm glow in the all-sky-camera photographs, we believe that the auroral conditions were sufficiently homogeneous to allow us to compare the radar and rocket ion compositions.

Good agreement is evident between the ion composition profiles measured by the rocket and radar techniques (Figure 3). Although the agreement is better when Cira (1972) is used instead of Banks and Kockarts for the radar analysis, the results are not particularly sensitive to the assumed neutral model. The maximum differences occur below the 185-km transition altitude where the two model temperature profiles differ the most. The systematic difference in the transition alti-

![Fig. 1. Technique for the determination of ion composition profiles.](image-url)
tude due to the choice of neutral model is less than 10 km for this determination, with the Cira (1972) model giving the lower altitude. For the remainder of the study, the Cira (1972) neutral model is used.

Quiet Periods

The diurnal variations in the transition altitude were examined during quiet periods of October 15 and May 13. The percentage of atomic ions was evaluated at altitudes corresponding to the midpoint of each range gate and a composition profile was constructed by fitting a curve through those data points. For the October data the radar pulse length was 320 $\mu$s, providing a 48-km range resolution. Analysis was not attempted at altitudes below about 160 km because the pulse length was large relative to the scale height. For the May data the pulse length was 160 $\mu$s. These data were analyzed down to 140 km altitude.

The October day shown in Figure 4 contained the most pronounced diurnal variation in transition altitude. This is expected since the solar zenith angle in October varies between 74$^\circ$ and 123$^\circ$—i.e., from a sunlit to a nighttime situation. The transition altitude was a maximum during the night—approximately 190 to 200 km—and a minimum during the day—approximately 175 km at 1400 local time (0000 UT). During the May experiment, the energy deposition and joule heat input rates were minimal. There was very little variation in the transition altitude (Figure 5). The altitude was consistently between 190 and 200 km. Pronounced diurnal variation is not expected because the solar zenith angle never exceeded 96$^\circ$ on that day.

Seasonal variations were examined by comparing data in May and October with the same solar zenith angles. For a solar zenith angle of 76$^\circ$ the transition altitude in May was about 195 km—while in October (at 1300 local) it was about 175 km. This difference, 20 km, is the maximum observed. For larger solar zenith angles in October, the transition altitude increases and approaches the relatively constant May value. This observed variation is consistent with the seasonal variation observed at midlatitudes [Evans and Cox, 1970; Oliver, 1975] and with the seasonal variation in the $[O]/[N_2]$ ratio [Roble, 1977] that will be discussed in a later section.

Active Periods

Particle precipitation. We have examined the effect of particle precipitation on the transition altitude. A typical event took place in the early portion of the October data already presented. Between about 1100 and 1200 UT, moderate particle precipitation occurred. The particle energy input rate varied from 5 to 20 ergs/cm$^2$ s, as shown in Figure 6. During this time, the electric fields (and the joule heat input) were minimal (<3 ergs/cm$^2$ s), and consequently $T_i = T_e$ was assumed. We have already examined the transition altitude for October during the quiet period and we have found that the transition altitude during the night was ~190 km or greater. During the particle precipitation event, however, it can be seen from Figure 4 that the transition altitude decreased to ~180 km. At the peak of the event the composition profiles are quite similar to daytime profiles previously discussed.

Joule heating. The largest variation in ion composition was observed when the joule heating rate was large. Such was the case on August 13, 1975, at about 1400 UT. The temperatures, electric fields, and the joule heat input rates were computed for the full 24-hour period. As can be seen in Figure 7, the joule heating was large at 1400 UT and moderate at 1200 UT.

The electron and ion temperatures (computed with the assumption of 100 percent O$^+$ ions) at 277 km altitude are shown in Figure 8. Early in the experiment, 0000 to 0600 UT, the ion temperature is fairly constant with a value of 900 K. The electron temperature shows the effect of the changing solar zenith angle. Initially, during sunlit conditions, the temperature is about 2200 K and it gradually decreases with increasing solar zenith angle (from 0430 to 0800 UT). At 0900 UT, the temperature is at a minimum of about 1400 K and gradually increases again as the solar zenith angle decreases.

---

<table>
<thead>
<tr>
<th>UT Date</th>
<th>Min. SZA</th>
<th>Max. SZA</th>
<th>00-03</th>
<th>03-06</th>
<th>06-09</th>
<th>09-12</th>
<th>12-15</th>
<th>15-18</th>
<th>18-21</th>
<th>21-24</th>
<th>$\Sigma$</th>
<th>$S_{107}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aug. 13, 1975</td>
<td>50</td>
<td>100</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>4</td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>0</td>
<td>1</td>
<td>14</td>
</tr>
<tr>
<td>Oct. 15, 1975</td>
<td>74</td>
<td>123</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>80</td>
</tr>
<tr>
<td>Feb. 28, 1976</td>
<td>3</td>
<td>5</td>
<td>3</td>
<td>69.2</td>
<td>82.1</td>
<td>72.1</td>
<td>82.1</td>
<td>69.2</td>
<td>82.1</td>
<td>72.1</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

---

**Fig. 2.** Energy deposition rates measured by the Chatanika radar during rocket flight on February 28, 1976. The rocket collected data from 1219 to 1221 UT when the joule heat input rate was negligible.

**Fig. 3.** Comparison of ion composition profiles from rocket and radar data. The radar-determined profiles were computed using two neutral models for an exospheric temperature of 900 K.
Fig. 4. Ion composition contours for October 15, 1975. The transition altitude lowers as the solar zenith angle decreases. The decrease between 1000 and 1200 UT is due to particle precipitation.

There is little change in the ion temperature until about 1000 UT, when the joule heat input begins. There are obvious enhancements in ion temperature at 1000, 1200, and 1400 UT. The electron temperature around 1400 UT appears to decrease suggesting that the 100 percent O+ assumption is no longer valid. The apparent dip in \( T_e \) for three altitudes is shown in Figure 9. The dip at 277 km is larger than at 218 km because of a larger relative change from atomic to molecular ions. While the dip is small at 335 km, it does indicate the presence of a significant percentage (20%) of molecular ions during the joule heating event. During this period (1330-1430 UT) the method of constraining \( T_e \) was used to compute the ion composition.

The composition contours are indicated in Figure 10. The transition altitude is seen to be near 190 km at the beginning of the observations. However, at the peak of joule heating, 1400 UT, this altitude increased by 50 km—molecular ions are predominant up to 240 km.

During this heating period, the electron density was also seen to decrease. The maximum \( F \) region electron density 30 min prior to the depletion was about \( 2 \times 10^6/cm^3 \). During the heating, the density dropped to \( 10^5/cm^3 \), and 30 min later it had recovered to \( 1.8 \times 10^5/cm^3 \). In addition to transport effects, a change in the dominant \( F \) region ion from atomic to molecular should cause a decay in the electron density since molecular ions, \( O_2^+ \) and \( NO^+ \), recombine much more rapidly. Electron density profiles measured before, during, and after the heating event are shown in Figure 11.

The \( E \) region electron density was also seen to decrease during the period when the electric field was large. This decrease must be due to a reduction, or softening, of the particle precipitation and not due to a change in ion composition. At \( E \) region altitudes, the predominant ions are molecular regardless of the auroral activity.

4. DISCUSSION

To better understand the ion composition variations deduced from our data, model calculations were performed using the measured physical parameters (\( N_e \) and temperature) and the neutral model atmosphere parameters. The factors that cause variations in the ion composition can be examined with a simple numerical model involving the major constituents, the most important reactions, and their associated reaction rates:

\[
\begin{align*}
O^+ + O_2 &\rightarrow O_2^+ + O \\
O_2^+ + e &\rightarrow O + O \\
O^+ + N_2 &\rightarrow NO^+ + N \\
NO^+ + e &\rightarrow N + O \\
N_2^+ + O &\rightarrow NO^+ + N \\
N_2^+ + e &\rightarrow N + N
\end{align*}
\]

where

\[
\begin{align*}
\gamma_1 &= 2 \times 10^{-11} \left( \frac{300}{T_e} \right)^{1/2} \text{ cm}^3/\text{s} \\
\alpha_1 &= 2.2 \times 10^{-7} \left( \frac{300}{T_e} \right) \text{ cm}^3/\text{s}
\end{align*}
\]
Fig. 8. Electron and ion temperatures computed for 100% O⁺ at 277 km altitude on August 13, 1975. A diurnal variation in $T_e$ can be seen. During the period of joule heating (1400 UT) the ion temperature is enhanced coincident with an anomalous decrease in electron temperature indicating a change in the ion composition.

$$\gamma_2 = 1 \times 10^{-12} \left( \frac{300}{T_i} \right) \text{cm}^3/\text{s}$$  \hspace{0.5cm} (2c)

$$\alpha_2 = 4.1 \times 10^{-7} \left( \frac{298}{T_i} \right) \text{cm}^3/\text{s}$$  \hspace{0.5cm} (2d)

$$\gamma_3 = 1.4 \times 10^{-19} \text{cm}^3/\text{s}$$  \hspace{0.5cm} (2e)

$$\alpha_3 = 2.9 \times 10^{-7} \left( \frac{300}{T_i} \right)^{1/3} \text{cm}^3/\text{s}$$  \hspace{0.5cm} (2f)

The reaction rate coefficients are taken from Jones and Rees [1973].

The continuity equation for each species is expressed as

$$\frac{\partial [X^+]}{\partial t} = q(X^+) - L$$  \hspace{0.5cm} (3)

where $[X^+]$ is the number density of species $X^+$, $q(X^+)$ is the production rate of species $X^+$, and $L$ is the loss rate, when dynamical factors are unimportant, as we will assume for this analysis. For steady state conditions and charge neutrality ($N_e = [O^+] + [NO^+] + [O_2^+]$), the continuity equations can be combined and solved for the fraction of atomic ions

$$\frac{[O^+]}{N_e} = \left\{ 1 + \frac{\gamma_1 [O_2]}{\alpha_1 N_e} + \frac{\gamma_2 [N_2]}{\alpha_2 N_e} \right\}^{-1}$$  \hspace{0.5cm} (4)

where

$$q(O^+) = \frac{0.56[O_2\eta]}{1.15[N_2] + 1.5[O_2] + 0.56[O]}$$  \hspace{0.5cm} (5a)

$$q(O_2^+) = \frac{[O_2\eta]}{1.15[N_2] + 1.5[O_2] + 0.56[O]}$$  \hspace{0.5cm} (5b)
TABLE 2. Ionospheric Data and Model Atmosphere for Quiet Day and Quiet Night Comparison

<table>
<thead>
<tr>
<th>Altitude, km</th>
<th>$N_{e}$, cm$^{-3}$</th>
<th>$T_e$, K</th>
<th>$T_m$, K</th>
<th>$[O_2]_m$, cm$^{-3}$</th>
<th>$[O]$, cm$^{-3}$</th>
<th>$[N_2]_m$, cm$^{-3}$</th>
<th>$N_{e}$, cm$^{-3}$</th>
<th>$T_e$, K</th>
<th>$T_m$, K</th>
</tr>
</thead>
<tbody>
<tr>
<td>140</td>
<td>7.5E4</td>
<td>596</td>
<td>650</td>
<td>6.22E9</td>
<td>3.78E10</td>
<td>5.37E10</td>
<td>8.6E4</td>
<td>596</td>
<td>596</td>
</tr>
<tr>
<td>160</td>
<td>1.0E5</td>
<td>765</td>
<td>1000</td>
<td>1.55E8</td>
<td>9.10E9</td>
<td>5.89E9</td>
<td>7.0E4</td>
<td>765</td>
<td>765</td>
</tr>
<tr>
<td>180</td>
<td>1.2E5</td>
<td>861</td>
<td>1600</td>
<td>5.26E8</td>
<td>5.51E8</td>
<td>2.59E9</td>
<td>5.6E4</td>
<td>861</td>
<td>861</td>
</tr>
<tr>
<td>200</td>
<td>1.5E5</td>
<td>916</td>
<td>1800</td>
<td>2.07E8</td>
<td>9.10E9</td>
<td>5.89E9</td>
<td>5.0E4</td>
<td>916</td>
<td>916</td>
</tr>
<tr>
<td>220</td>
<td>2.0E5</td>
<td>948</td>
<td>2000</td>
<td>8.87E7</td>
<td>3.52E9</td>
<td>1.22E9</td>
<td>3.4E4</td>
<td>948</td>
<td>948</td>
</tr>
<tr>
<td>240</td>
<td>3.0E5</td>
<td>967</td>
<td>2000</td>
<td>3.99E7</td>
<td>2.33E9</td>
<td>6.04E8</td>
<td>3.0E4</td>
<td>967</td>
<td>967</td>
</tr>
<tr>
<td>260</td>
<td>2.5E5</td>
<td>979</td>
<td>2000</td>
<td>1.86E7</td>
<td>1.57E9</td>
<td>3.09E8</td>
<td>2.8E4</td>
<td>979</td>
<td>979</td>
</tr>
<tr>
<td>280</td>
<td>2.45E5</td>
<td>986</td>
<td>2000</td>
<td>8.85E6</td>
<td>1.07E9</td>
<td>1.61E8</td>
<td>2.6E4</td>
<td>986</td>
<td>986</td>
</tr>
<tr>
<td>300</td>
<td>2.0E5</td>
<td>991</td>
<td>2000</td>
<td>4.30E6</td>
<td>7.45E8</td>
<td>8.55E7</td>
<td>2.4E4</td>
<td>991</td>
<td>991</td>
</tr>
</tbody>
</table>

The ionospheric data are from February 18, 1976. The neutral data are from the Cira (1972) model for an exospheric temperature of 1000 K.

$$q(N_2^+) = \frac{0.92[N_2]_\eta}{1.15[N_2]_\eta + 1.50[O_2]_\eta + 0.56[O]}$$  \hspace{1cm} (5c)

$$[N_2^+] = \frac{q(N_2^+)}{\gamma_3[O] + \omega N_e}$$  \hspace{1cm} (6)

and $\eta$ is the total ion production rate. These production rates are from Jones and Rees [1973].

Using (4) for the ion composition, the effects on the transition altitude of varying the atmospheric density, electron density, production rates, and temperatures can be assessed.

**Diurnal Variation**

Our observations indicate that at night the transition altitude increases. In the absence of auroral activity, the ionospheric parameters most affected by a change in solar zenith angle from day to night are the electron density and electron temperature. The expected diurnal variation in ion composition can be calculated by solving (4) using measured ionospheric data for $N_e$, $T_e$, $T_m$, and a model atmosphere for $[N_2]$, $[O_2]$, and $[O]$ (Cira, 1972). The values used are given in Table 2. The total ion production rate was adjusted during the daytime to fit the observed electron density profiles. The production rate was decreased to match measured nighttime profiles. The resultant ion composition profiles are shown in Figure 12. The difference in the profiles is most pronounced above the transition altitude where the percentage change in electron densities is greatest. In the model calculations, the transition altitude increases by 20 km at night, which is consistent with the measurements shown in section 3.

**Seasonal Variation**

A number of factors could be responsible for the observed seasonal variation in the transition altitude (as much as 20 km higher in May than in October): variations in the solar EUV flux and variations in the neutral atmospheric density (specifically the $[O]/[N_2]$ ratio). As shown in Table 1, there is very little change in the solar radiation as represented by the $S_{10.7}$ parameter (<10%). However, an examination of the peak electron density in the $F$ region shows that the October day had more than twice as much ionization present. This is consistent with expected seasonal variations in the $[O]/[N_2]$ ratio being smaller in summer than in winter [Roble, 1977]. Enhancing the $[N_2]$ at a given altitude relative to $[O]$ increases the $O^+$ loss rate, thereby decreasing the electron density and increasing the transition altitude. The model calculations indicate a small increase in transition altitude. The model calculations indicate a small increase in transition altitude (~5 km) when the $[O]$ to $[N_2]$ ratio is decreased by small amounts above 160 km to a factor of 2 at 300 km (approximated by using model values for a higher exospheric temperature). A larger $N_2$ variation would be needed to match the observed variation in the transition altitude.

**Particle Precipitation**

During enhanced particle precipitation we observe a lowering of the transition altitude. The variation can be similar in magnitude and sense to the 20-km decrease in going from night to day.

In the $F$ region, whether the ionizing source is solar EUV or auroral particle precipitation, the effects are similar. As the source strength increases, both molecular and atomic ions are created and the electron density increases. Because the $O^+$...
TABLE 3. Ionospheric Data for Particle Precipitation Event

<table>
<thead>
<tr>
<th>Altitude, km</th>
<th>( N _e ), cm(^{-3} )</th>
<th>( T_e ), K</th>
<th>( T_m ), K</th>
</tr>
</thead>
<tbody>
<tr>
<td>140</td>
<td>2.0E5</td>
<td>596</td>
<td>750</td>
</tr>
<tr>
<td>160</td>
<td>1.3E5</td>
<td>765</td>
<td>1120</td>
</tr>
<tr>
<td>180</td>
<td>1.5E5</td>
<td>861</td>
<td>1300</td>
</tr>
<tr>
<td>200</td>
<td>1.8E5</td>
<td>916</td>
<td>1800</td>
</tr>
<tr>
<td>220</td>
<td>1.9E5</td>
<td>948</td>
<td>2000</td>
</tr>
<tr>
<td>240</td>
<td>2.0E5</td>
<td>967</td>
<td>2000</td>
</tr>
<tr>
<td>260</td>
<td>1.9E5</td>
<td>979</td>
<td>2000</td>
</tr>
<tr>
<td>280</td>
<td>1.6E5</td>
<td>986</td>
<td>2000</td>
</tr>
<tr>
<td>300</td>
<td>1.4E5</td>
<td>991</td>
<td>2000</td>
</tr>
</tbody>
</table>

The data are from February 18, 1976.

The lifetime is long compared to that of NO\(^+\), O\(^+\) dominates to lower altitudes and the transition altitude lowers.

Another point relative to the production of NO\(^+\) is the increase in the reaction rate for O\(^+\) \( + \) N\(_2\) \( \rightarrow \) NO\(^+\) \( + \) N, when N\(_2\) becomes vibrationally excited [Schunk and Banks, 1975; Newton et al., 1977] During auroral activity N\(_2\) could become vibrationally excited N\(_2\)\(^*\) through the process [Schunk and Banks, 1975]

\[
e + N\_2 \rightarrow N\_2\^* + e
\]

\[
O('D) + N\_2 \rightarrow N\_2\^* + O('P)
\]

where e is an energetic electron. The effect would be to increase the number density of NO\(^+\) relative to O\(^+\), thereby raising the transition altitude.

Apparently, the increase in ionization rate is more important because in the cases that we have observed (e.g., October 15, 1975), there has always been a decrease in the transition altitude.

The model calculations show a similar decrease of the transition altitude. The effects of electron precipitation considered are enhancements of electron density, electron temperature, and N\(_2\) vibrational temperature. The ionospheric parameters are given in Table 3. The temperature and electron density profiles used are from data taken on February 18, 1976, and the reaction rate coefficient for N\(_2\) \( + \) O\(^+\) was enhanced to account for an N\(_2\) vibrational temperature of 1400 K. The particle energy input was approximately 20 ergs/cm\(^2\) s. The model composition profiles shown in Figure 13 indicate that relative to the background profile (quiet night) the transition altitude is lower by 25 km, in agreement with the data. This indicates that the effects of enhanced N\(_2\) vibrational temperature are less important than the increase in ion production rate in altering the transition altitude.

**Joule Heating**

The measurements indicate that during periods of large joule heating the transition altitude increases substantially. As a result of this change in ion composition, the F region ionization rapidly decays because of enhanced recombination of the molecular ions as compared to O\(^+\).

The observed behavior is expected both from theoretical arguments and satellite measurements. The following processes contribute to increasing the NO\(^+\) abundance [Banks et al., 1974; Schunk et al., 1975, 1976; Schunk and Banks, 1975]: (1) enhanced [N\(_2\)], (2) enhanced ion temperatures, (3) relative motion of the ions and neutrals.

Enhanced [N\(_2\)] has been observed by satellites in the auroral region [e.g., Prölls, 1980]. It is expected to occur because, during periods of large joule heating, the neutral atmosphere should be heated by the ion-neutral collisions. A heated volume element of neutral particles will rise with a vertical velocity that increases with altitude [Hays et al., 1973]. This expansion will cause an upwelling of N\(_2\) molecules, thereby increasing the number density of N\(_2\) at higher altitudes. The number density of NO\(^+\) will be enhanced because more N\(_2\) is available to either react with O\(^+\) to form NO\(^+\), or be ionized to form N\(_2\)\(^+\), which in turn reacts with O to form NO\(^+\).

The second and third factors contributing to the enhancement of NO\(^+\) ions during heating events are the dependence of the reaction rate \( \gamma_2 \) for O\(^+\) \( + \) N\(_2\) \( \rightarrow \) NO\(^+\) \( + \) N on ion temperature, and relative motion. From Banks et al. [1974]:

\[
\gamma_2 = 1.2 \times 10^{-12} \left( \frac{300}{T_{\text{eff}}} \right) \text{ cm}^3/\text{s} \quad T_{\text{eff}} < 750 \quad (8a)
\]

\[
\gamma_2 = 8.0 \times 10^{-14} \left( \frac{T_{\text{eff}}}{300} \right)^2 \text{ cm}^3/\text{s} \quad T_{\text{eff}} > 750 \quad (8b)
\]

where \( T_{\text{eff}} = T_i + 0.329 E_B^2 \), and \( E_B \) is the orthogonal \( E \) field (mV/m) measured in the neutral-wind frame. This is a simplified form assuming \( v_i / \Omega_e \ll 1 \) and \( B = 0.5 \) gauss (polar region). As an example, for an \( E \) field of 50 m V/m, and \( T_e = 900 \) K, \( T_{\text{eff}} \) will be increased by a factor of 1.9, as compared to \( T_{\text{eff}} \) for no \( E \) field. This in turn increases \( \gamma_2 \) by a factor of 3.7. The result is an enhancement in the loss rate of atomic ions and production rate of NO\(^+\) and, hence, an enhancement in the ratio of molecular to atomic ions. Furthermore, if for any reason there were an increase in the electron temperature, then the reaction rate \( \alpha_2 \) for dissociative recombination of NO\(^+\) would decrease [Walls and Dunn, 1974]. The net result would be an even larger increase in molecular ions relative to...
atomic ions and, indeed, an increase in the number density of molecular ions.

The atmospheric model and ionospheric data used to numerically model the effect of joule heating are given in Table 4. The ionospheric data are from August 13, 1975. The total ion production rates were adjusted to have the ion composition match the observation prior to the joule heating event. These rates were kept constant through the joule heating period. Due to the enhanced electric field and elevated ion temperatures observed during the event, \( N_b \) was considerably increased. The resultant model ion composition profile (Figure 14) indicates that more molecular ions should exist at all altitudes as compared to the quiet period. However, the transition altitude is only increased by 20 km. If we further increase the \( N_b \) density by a factor of two at 160 km (corresponding to an increase in neutral temperature of \( \sim 250^\circ \)), the transition altitude is increased by a total of 40 km, in better agreement with the observation.

5. SUMMARY AND CONCLUSIONS

We have described the incoherent scatter radar techniques by which variations of ion composition in the high-latitude ionosphere can be determined. The midlatitude technique based on the equality of ion and neutral temperatures between 130 and 300 km [Evans, 1967] can be applied under quiet conditions, under conditions of energetic particle precipitation, and under conditions of joule heating when the heating rate is low—less than 5 ergs/cm\(^2\) s. Confidence in the applicability of this technique has been gained by a comparison of radar-derived ion composition and simultaneous in situ rocket measurements of ion composition. We have developed another technique for conditions with considerable joule heating. It is based on the realistic assumption that during such periods the electron temperature will not decrease. If we assume the electron temperature is constant, we can determine a lower bound to the fraction of ionization that is due to molecular ions (NO\(^+\) and O\(^+\)) rather than atomic ions (O\(^+\)).

The existence of a ground-based radar technique enables us to examine the long-term behavior of the desired parameter, ion composition. In addition, the other parameters that can be determined from the radar measurements can be used to understand the variations in the ion composition. For simplicity of discussion we examine in detail only the time variation of the transition altitude, the altitude at which the composition is half molecular ions and half atomic. During the winter when there is a definite day and a definite night, the transition altitude increased from 175 km during the day to 190 km at night. In the summer, when the sun never sets on the ionosphere, the transition altitude remained consistently near 190 km. These two quiet time situations are very similar to the diurnal variation and the daytime winter-summer variation observed at midlatitude. During winter-nighttime particle precipitation events, the transition altitude decreases on the order of 10 to 20 km, depending on the energy input. The most dramatic changes occur during joule heating events when the transition altitude increases sharply—e.g., 50 km for one 30 erg/cm\(^2\) s event.

We compared our data with a simple steady state model of the ion chemistry using the important reactions to examine these variations in the ion composition. The observed diurnal and seasonal variations are similar to those seen at midlatitudes and are presumably due to the same causes. Indeed, the simple model calculations do show that the diurnal variation can be accounted for by the changes in photoionization with varying solar zenith angles, and that the seasonal variation is consistent with an increasing [O] to [N\(_2\)] ratio from summer to winter.

During periods of particle precipitation three things are expected to happen: ion production rates increase, electron temperatures increase, and N\(_2\) becomes vibrationally excited. The first would cause the transition altitude to decrease; the other two would cause it to increase. The modeling shows that the effect in transition altitude of an increased ion production rate is much more important than the effect of an increased electron temperature. The fact that we observe a decrease in the transition altitude shows that the increased ion production rate is also more important than vibrational excitation of N\(_2\). For this to be the case, the modeling shows that either the vibrational temperature has to be low or the vibrationally enhanced reaction rates are in error. Increased N\(_2\) density appears to be unimportant.

**TABLE 4.** Ionospheric Data and Model Atmosphere for Period of Joule Heating

<table>
<thead>
<tr>
<th>Altitude, km</th>
<th>( N_a ) cm(^{-3})</th>
<th>( T_e ) K</th>
<th>( T_m ) K</th>
<th>([O_2]_a) cm(^{-3})</th>
<th>([O]_a) cm(^{-3})</th>
<th>([N_2]_a) cm(^{-3})</th>
</tr>
</thead>
<tbody>
<tr>
<td>140</td>
<td>7.5E4</td>
<td>1590</td>
<td>600</td>
<td>6.22E9</td>
<td>3.78E10</td>
<td>5.37E10</td>
</tr>
<tr>
<td>160</td>
<td>1.0E5</td>
<td>1780</td>
<td>780</td>
<td>1.55E9</td>
<td>1.66E10</td>
<td>1.54E10</td>
</tr>
<tr>
<td>180</td>
<td>1.2E5</td>
<td>1910</td>
<td>1050</td>
<td>5.26E8</td>
<td>9.10E9</td>
<td>5.89E9</td>
</tr>
<tr>
<td>200</td>
<td>1.0E5</td>
<td>2000</td>
<td>1300</td>
<td>2.07E8</td>
<td>5.51E9</td>
<td>2.59E9</td>
</tr>
<tr>
<td>220</td>
<td>9.0E4</td>
<td>2050</td>
<td>1550</td>
<td>8.87E7</td>
<td>3.52E9</td>
<td>1.22E9</td>
</tr>
<tr>
<td>240</td>
<td>9.2E4</td>
<td>2100</td>
<td>1750</td>
<td>3.99E7</td>
<td>2.33E9</td>
<td>6.04E8</td>
</tr>
<tr>
<td>260</td>
<td>1.0E5</td>
<td>2130</td>
<td>1900</td>
<td>1.66E7</td>
<td>1.57E9</td>
<td>3.09E8</td>
</tr>
<tr>
<td>280</td>
<td>1.2E5</td>
<td>2150</td>
<td>2050</td>
<td>8.65E6</td>
<td>1.07E9</td>
<td>1.61E8</td>
</tr>
<tr>
<td>300</td>
<td>1.2E5</td>
<td>2200</td>
<td>2250</td>
<td>4.30E6</td>
<td>7.43E8</td>
<td>8.53E7</td>
</tr>
</tbody>
</table>

The ionospheric data come from August 13, 1975.
During periods of substantial joule heating there are again three things that we expect: \( N_2 \) density increase, ion temperature enhancement, and large relative ion-neutral velocities. All three would lead to increases in the transition altitude by increasing the rate of conversion of \( O^+ \) to \( NO^+ \). For the case examined here the modeling shows that the relative ion-neutral velocity can account for 40\% of the change in the transition altitude. The change in ion temperature has a smaller effect. This implies that there must be an increase in the \( N_2 \) density—e.g., a factor of 2.5 at 277 km—to help account for the change in transition altitude. The importance of an \( N_2 \) density increase for the ion composition variation during joule heating is very different from the particle precipitation situation.

Thus, for the first time we have examined the ion composition in the auroral region using incoherent scatter. Under quiet conditions, the behavior is similar to that observed at midlatitudes. Under active conditions, both particle precipitation and joule heating produce significant changes in ion composition. By comparison of observation and model, we have shown that for particle precipitation the most important factor contributing to a decrease in transition altitude is an enhanced ion production rate. For joule heating, the two important factors contributing to an increase in transition altitude are a large differential ion-neutral velocity and an increase in the \( N_2 \) density.

Acknowledgments. We would like to thank A. B. Christensen, formerly of the University of Texas at Dallas, now of Aerospace Corporation, for permission to use the rocket data prior to publication. We would also like to thank M. H. Rees of the Geophysical Institute, University of Alaska, for numerous valuable discussions. We would finally like to thank numerous members of the staff of the Radio Physics Laboratory, SRI International, who are involved in the operation of the Chaticana radar and the reduction of the data. In particular, we would like to acknowledge the invaluable assistance of Mary McCreary in all phases of the data reduction and analysis. This research was carried out under contract DNA001-77-C-0042 from the Defense Nuclear Agency and several grants from the Division of Atmospheric Sciences, National Science Foundation. The Chaticana radar is operated by SRI International under a grant from the Atmospheric Program, National Science Foundation.

The Editor thanks R. H. Wand for his assistance in evaluating this paper.

REFERENCES


Schunk, R. W., W. J. Raitt, and P. M. Banks, Effect of electric fields on the daytime high-latitude \( E \) and \( F \) regions, J. Geophys. Res., 80, 3121-3130, 1975.

Schunk, R. W., P. M. Banks, and W. J. Raitt, Effects of electric fields and other processes upon the nighttime high-latitude \( F \) layer, J. Geophys. Res., 81, 3271-3282, 1976.


(Received December 17, 1980; revised April 6, 1981; accepted April 22, 1981.)