THE VISIBLE AIRGLOW EXPERIMENT—A REVIEW*

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Abstract—Contributions of the Visible Airglow Experiment (VAE) to the understanding of thermospheric aeronomy are summarized and discussed. The importance of instrumental design and operation is considered and particular attention is given to the relationship between observation and inversion for optical measurements of light emissions from a diffuse medium. The specific influence of the VAE investigation on the present understanding of the physical mechanisms which produce and destroy the excited species, $O^+(^2P)$, $O(^1D)$, $O(^1S)$, $N(^2D)$, $N_2^+(B^2\Sigma_u^+)$, $N_2(C^3\Pi_u)$ and $Mg^+(^2P_{1/2})$, is discussed in detail.

1. INTRODUCTION

The Visible Airglow Experiment (VAE) was one of 14 interrelated instruments which were flown on the *Atmosphere Explorer* (*AE*)-C, -D and -E satellites (Dalgarno et al., 1973) in the 1970s. The goal of this unique cooperative experiment was to provide the mission science team with the information required to quantify the state of the thermosphere and ionosphere of the Earth. This process has been highly successful and our knowledge of the thermosphere and ionosphere today rests in large part on the database that was collected and analyzed as a part of the *Atmosphere Explorer* investigation. It is of interest to note that Alex Dalgarno was one of the originators of this unique NASA mission, and the Science Team leader during the early phase of analysis of results from the satellites.

This review will concentrate on the understanding of various airglow and auroral processes that were clarified from the information obtained with the Visible Airglow Experiment; however, it should be noted that in many of these studies information was derived from all of the instruments operating on the specified *AE* spacecraft at the time. The VAE was flown to provide the *AE* science team with detailed profiles of the distribution of excited states of atoms and molecules that contribute to or are bi-products of the chemistry induced by the absorption of solar ultraviolet radiation or auroral particles. The major species of interest to the airglow instrument team that will be discussed here are $O^+(^2P)$, $O(^1D)$, $O(^1S)$, $N(^2D)$, $N_2^+(B^2\Sigma_u^+)$, $N_2(C^3\Pi_u)$ and $Mg^+(^2P_{1/2})$. We note that a number of hydrogen emissions were recorded, but due to the low signal-to-noise ratio have yet to be

analyzed in detail. The technique of determining the density of these excited species involved first measuring the brightness of the emission feature with a photometer, VAE, and then inverting an integral equation to determine the local emission rate or species density. This relationship between observation and inversion is inherent in interpreting measurements of light emissions from any diffuse medium, particularly in planetary atmospheres (Hays and Roble, 1968a; Solomon, 1987).

2. INSTRUMENT AND INVERSIONS

The VAE instrument is in principle a very elementary device (Hays et al., 1973), consisting of a two channel filter wheel photometer with carefully baffled telescopes. The two separate optical channels have different fields of view which allow high angular resolution in channel one for horizon and auroral observations, and high sensitivity in channel two for nightglow and other weak-emission features. These channels were oriented at $90°$ from each other with the narrow channel pointed directly aft of the spacecraft and the sensitive channel either pointed up or down depending on orientation of the sun with respect to the orbital plane. This arrangement of the optical channels is convenient when the spacecraft is oriented, allowing the narrow channel to utilize the Chapman enhancement of the horizon while the sensitive channel views the airglow vertically at various altitudes. When the spacecraft is spinning the two channels scan through all zenith angles within the orbital plane, allowing two separate horizon scans to be recorded on each spin of the satellite. Most of the results discussed in this review were taken in this dynamic scanning mode from the narrow angle channel of VAE. A typical two horizon scan sequence is shown in Fig. 1 where raw counts are recorded for *AE-C* orbit 531 on
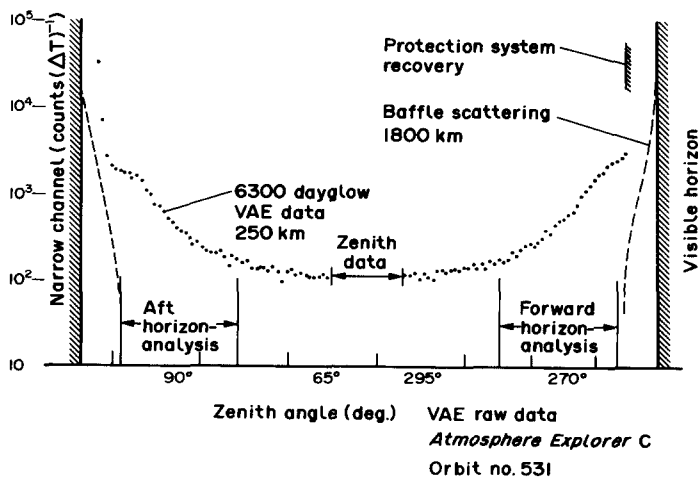

FIG. 1. RAW DATA FROM THE VISIBLE AIRGLOW EXPERIMENT DURING A SINGLE SPIN OF THE SPACECRAFT.
(From Hays et al., 1978.)

the narrow angle channel with a 6300 Å filter in the light beam. The atmospheric brightness of the individual emission features was obtained by removing the galactic contribution to the observed brightness and scaling the resulting signal with calibration constants obtained in the laboratory before flight. We note that there is no indication of a significant variation in sensitivity of these instruments during their lifetime in orbit, a 5 y period for *AE-C* and *-E*.

The requirement for inversion of optical observations of an extended diffuse object is related to the fact that the observed quantity is always the surface brightness, the integral of the emission rate along the line of sight (Hays and Roble, 1968a). There are a multitude of specific observational geometries which change the detailed nature of the inversion process, but even in the most elementary case, for instance linear sounding from a rocket, the brightness must be differentiated to obtain the volume emission rate. The first truly remote sensing of airglow emissions by VAE from limb scans were inverted using the Abel inversion (Hays and Roble, 1968b; Roble and Hays, 1972; Hays et al., 1973) under the assumption that the airglow did not vary in the horizontal direction. It was recognized soon that although this is a good assumption during the day and at midlatitudes at night, it is not valid at twilight, in the equatorial regions at night, and in the auroral zone where horizontal structure is always evident. Several techniques have been developed to deal with these more complex situations. The first general treatment of VAE data involved the use of the algebraic reconstruction technique (Hays et al., 1978) using an iterative relaxation method to solve the linear equations generated from

the observational set, a method first used with data from the *OGO*-6 satellite (Thomas and Donahue, 1972). A modified "onion peeling" technique (Fesen and Hays, 1982a) was the next step in the evolution toward modern two-dimensional generalized inversions (Solomon et al., 1984, 1985; Solomon, 1987) developed from classical X-ray tomography (Radon, 1917; Cormack, 1963). This recent work allows VAE airglow and auroral observations to be interpreted even in highly structured situations.

The remainder of this paper is devoted to reviewing the contributions of the observations taken by Visible Airglow Experiment to our present understanding of the physical processes that control the airglow and auroral emissions from $O^+(^2P)$, $O(^1D)$, $O(^1S)$, $N(^2D)$, $N_2^+(B^2\Sigma_u^+)$, $N_2(C^3\Pi_u)$ and $Mg^+(^2P_{1/2})$.

3. AURORAL TRANSITION OF OII, 7320 Å

The metastable transition between excited ^2P and ^2D states of the atomic oxygen ion results in the thermospheric emission at 7320–30 Å. The $O^+(^2P)$ ion can radiate either at 7320–30 Å in the Auroral transition to $O^+(^2D)$ or at 2470 Å in the Transauroral transition to $O^+(^4S)$ (Chamberlain, 1961). The theoretical transition probabilities of Seaton and Osterbrock (1957) yield a radiative lifetime of 4.57 s for $O^+(^2P)$ ions and a branching ratio of 0.781 for the emission of radiation in the multiplet at 7320 Å.

Dalgarno and McElroy (1965) were the first to show that 20% of the O^+ ions formed by photon and photoelectron ionization are in the $O^+(^2P)$ state. Auroral and twilight observations of these doublet emissions have indicated that this metastable is heav-

ily quenched in the thermosphere (Chamberlain, 1961; Carlson and Suzuki, 1974). Using the rate coefficient for electron quenching evaluated by Henry (1970) together with VAE measurements, Walker et al. (1975) demonstrated that $O^+(^2P)$ is strongly quenched by collisions with neutral particles below 300 km at a rate close to the gas kinetic rate. Rusch et al. (1977) later analyzed a larger VAE data set and derived quenching rate coefficients for $O^+(^2P)$ by O and N_2 of 5.2×10^{-11} cm^3 s^{-1} and 4.8×10^{-10} cm^3 s^{-1}, respectively. Figure 2, adopted from the studies of Rusch et al., shows the 7320 Å volume emission rate profile deduced from the VAE observed surface brightness along with the calculated results using various rate coefficients of quenching by O and N_2. A morphological study of the emission analyzing all the VAE measurements taken between 1974 and 1979 was conducted and found to be in excellent agreement with a theoretical model based on the photochemistry of $O^+(^2P)$ (Yee et al., 1981).

In the absence of quenching, the rate of emission of 7320 Å radiation is 0.781 times the rate of production of $O^+(^2P)$ ions. Thus, measurements of the intensity of the doublet at 7320–30 Å serve as a direct monitor of the rate of ionization of atomic oxygen (Hays et al., 1973; Rusch et al., 1977). Abreu et al. (1980) later established that the $F10.7$ index and ionization rate vary in a similar fashion and the deduced ionization frequency is found to be in good agreement with the theoretical calculation by using the observed solar flux and the calculated cross-section of Kirby-Docken et al. (1979).
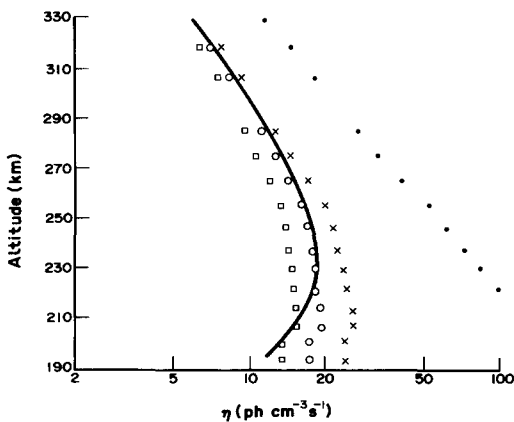

FIG. 2. VOLUME EMISSION RATE OF OII (7320 Å) DAYGLOW OBTAINED ON AE-C ORBIT 669 (SOLID LINE) COMPARED WITH THE UNQUENCHED SOURCE (DOTS) AND WITH THE THEORY FOR VARIOUS ATOMIC OXYGEN QUENCHING RATES WITH $K_{N_2} = 0$. Dots represent $K_O = 0$ and $K_e = 0$; crosses, $K_O = 2 \times 10^{-10}$ cm^3 s^{-1}; circles, $K_O = 3 \times 10^{-10}$ cm^3 s^{-1}; and squares, $K_O = 6 \times 10^{-10}$ cm^3 s^{-1}. (From Rusch et al., 1977.)

Concerning the production of $O^+(^2P)$ due to ionization by photoelectrons, controversy arose after a set of ground-based and rocket observations of auroral 7320 Å emission were published by Swenson (1976). The theoretically calculated cross-sections by Dalgarno and Lejeune (1971) predicted a branching ratio of $\sim 20\%$ to $O^+(^2P)$. Swenson (1976) concluded that a branching ratio of less than 5% was required to explain their measurements. Rees et al. (1982) analyzed a set of complementary auroral observations obtained by the VAE instrument and other instruments on board the AE-D satellite and derived the production efficiency of $O^+(^2P)$ to be 18% by electron impact ionization. The results are in harmony with the theoretical calculations of Dalgarno and Lejeune (1971), which implies that the early deduced quenching rate coefficients by Rusch et al. (1977) could be low by approximately 15–20% since the electron impact ionization as a source was neglected (Torr and Torr, 1982).

A theory for monitoring exospheric temperature from the 7320–30 Å emission was developed by Yee and Abreu (1982a) by measuring the rate of decreasing $O^+(^2P)$ vertical brightness after twilight. Their deduced temperatures from VAE data agree very well with temperatures simultaneously measured by the neutral atmospheric temperature experiment (NATE) on board AE.

Of significance, however, is that in the studies of Yee and Abreu (1982a) a large emission background near 400 km was observed which they attributed to the interaction of the spacecraft with the atmospheric environment. This had been observed earlier at low altitudes (Torr et al., 1977), and led to a series of studies using VAE data at several wavelengths to examine spacecraft–environment interactions (Yee and Abreu, 1982b, 1983; Yee et al., 1984, 1985a). These investigations set the foundation for understanding the optical glow observed on the shuttle (Banks et al., 1983).

4. NEBULAR TRANSITION OF OI, 6300 Å

The atomic oxygen emissions at 6300, 6364 and 6392 Å, arising from the ^3P–^1D transition, are still a topic of active study. The VAE photometers observed the brightest line of the three at 6300 Å and AE studies have been of crucial importance in determining the thermospheric chemistry of the $O(^1D)$ metastable. Early studies of the atmospheric excitation of $O(^1D)$ have been reviewed by Chamberlain (1961), Noxon (1968), Bates (1978, 1982) and Torr and Torr (1982). Bates and Dalgarno (1953, 1954) showed that the 6300 Å emission should be one of the strongest features of

the dayglow but also suggested that chemical deactivation was important at lower altitudes. Studies by Dalgarno and Walker (1964), Fournier and Nagy (1965), Hunten and McElroy (1966), Wallace and McElroy (1966) and Dalgarno and Lejeune (1971) examined the important mechanisms for production and loss of $O(^1D)$ in the upper atmosphere: photodissociation of O_2, impact of fast electrons on atomic oxygen, dissociative recombination of O_2^+, quenching by N_2 and O_2, and spontaneous emission.

The first use of VAE measurements to investigate the day and night 6300 Å emission was by Hays *et al.* (1978). Using nightglow observations to investigate the dissociative recombination and quenching reactions:

$$O_2^+ + e \rightarrow O + O(^1D)$$

$$O(^1D) + N_2 \rightarrow O(^3P) + N_2$$
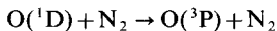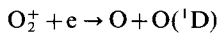
and employing the Garstang (1951, 1956) estimate for the total radiative transition probability $A_D^1 = 0.0091$ s^{-1}, a yield of 1.3 $O(^1D)$ atoms per recombination and a rate coefficient for quenching of 3×10^{-11} cm^3 s^{-1} was obtained. Applied to the dayglow, where the additional sources:

$$e^* + O \rightarrow e^* + O(^1D)$$

$$O_2 + h\nu \rightarrow O + O(^1D)$$
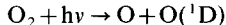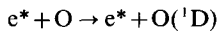
must be considered, these values were found to be compatible with theoretical estimates of the photo-electron flux and a low solar activity unattenuated rate for Schumann–Runge photodissociation $J_{O_2}(\infty) = 3 \times 10^{-6}$ s^{-1}. A characteristic dayglow altitude profile from this work is displayed in Fig. 3.

Following the suggestion (Rusch *et al.*, 1978; cf. Sharp *et al.*, 1979) that the reaction:

$$N(^2D) + O_2 \rightarrow NO + O(^1D)$$
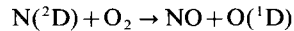
is an important source of $O(^1D)$ in the aurora, Torr *et al.* (1980) considered this reaction as a significant dayglow source as well. Here, the lower value for $J_{O_2}(\infty)$ of 1.5×10^{-6} s^{-1} established from *AE* solar e.u.v. measurements was used. However, Link *et al.* (1981) pointed out that Garstang's revised (1968) calculations found $A_D^1 = 0.0068$ s^{-1}, closer to the laboratory measurements of Kernahan and Pang (1975). Applying this value to the Hays *et al.* (1978) data, a rate coefficient for quenching by N_2 of 2.3×10^{-11} was calculated, in better agreement with laboratory measurements extrapolated to thermospheric temperatures. Using rocket observations of the 5200 Å/6300 Å ratio in the cleft aurora, Link (1983) also argued that the reaction of $N(^2D)$ with O_2 produces little $O(^1D)$. Torr *et al.* (1981) accepted the lower quenching rate, but, using VAE measurements of dayglow emission rate profiles, demonstrated that a $O(^1D)$ yield of ~ 0.8 from this reaction was indicated (see Fig. 4). A subsequent auroral study (Sharp *et al.*, 1983) using *AE-D* data also found that the "classical" sources of dissociative recombination and electron
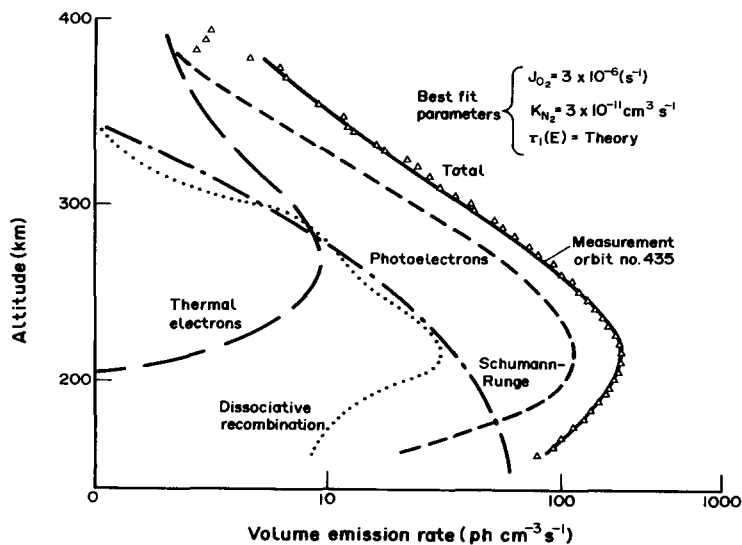

FIG. 3. VOLUME EMISSION RATE OF OI (6300 Å) DAYGLOW OBTAINED ON *AE-C* ORBIT 435 AND THEORETICAL BEST FIT COMPARISON.
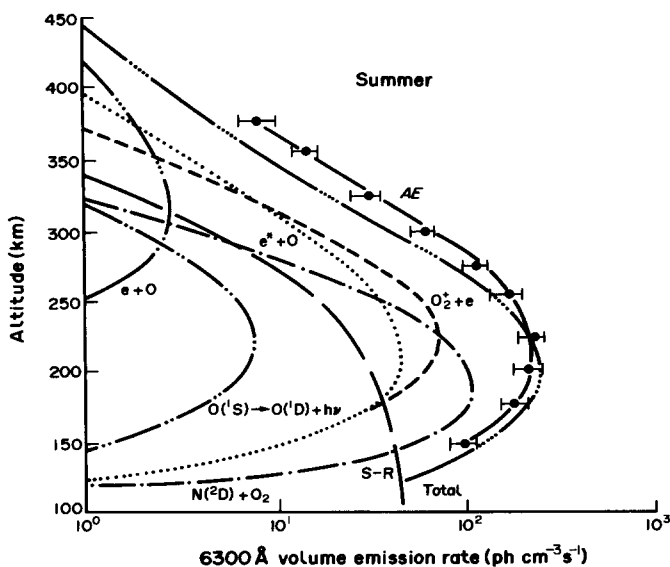(From Hays *et al.*, 1978.)

FIG. 4. DAYTIME OI 6300 Å VOLUME EMISSION RATES AS A FUNCTION OF ALTITUDE FOR JULY 1974.
(From Torr et al., 1981.)

impact were inadequate to explain the observed levels of 6300 Å emission, implying the existence of an additional production mechanism such as $N(^2D)+O_2$. However, laboratory measurements (Langford et al., 1986) indicate that the yield of $O(^1D)$ from this reaction is low.

New calculations by Froese-Fischer and Saha (1983) of a value for A_D^1 of 0.0093 s^{-1}, closer to the original Garstang estimate, and the suggestion (Yee et al., 1985b) that the quenching reaction

$$O(^1D)+O(^3P) \rightarrow O(^3P)+O(^3P)$$
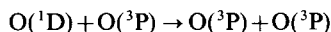
is of importance in the thermosphere, led to a study of the 6300 Å nightglow using VAE data (Abreu et al., 1986). This work found that with a quantum yield of $O(^1D)$ from dissociative recombination of 1.2 and the Froese-Fischer and Saha value for A_D^1, quenching rate coefficients for N_2 and O of 2.3×10^{-11} and 8×10^{-12} cm^3 s^{-1}, respectively, could be justified. These results were utilized in an examination of the 6300 Å auroral emission (Solomon, 1987; Solomon et al., 1987) using a tomographic inversion algorithm to compare VAE volume emission rate altitude profiles to those predicted by application of an electron transport code and chemical model to particle fluxes and ambient composition measured by AE instrumentation. In this study, values for the quantum yield of $O(^1D)$ from $N(^2D)+O_2$ of 0.1–0.2 were adopted, although higher values up to about 0.5 could not be ruled out. In addition, several minor aur-

oral sources were considered, as shown in Fig. 5. The "N^*+O_2" curve, representing a possible source of $O(^1D)$ from excited atomic nitrogen in the ^2D and ^2P states reacting with $O(^3P)$, is essentially speculative as the reaction rates are poorly established (see Section 6) and the products unknown, but the reaction of N^+ with O_2 has been measured in the laboratory (Langford et al., 1985, 1986) and found to result in substantial $O(^1D)$. Radiative cascade from $O(^1S)$ was a significant source of $O(^1D)$, but electron impact dissociation of O_2 and energy ransfer from $O^+(^2D)$ were not. Although electron temperatures during the period studied were not high enough to produce much thermal electron excitation of atomic oxygen, this mechanism could be important during high solar activity or in the dayside aurora, as it is in SAR-arcs.

The $N(^2D)+O_2$ reaction is likely to remain controversial (cf. Rees and Roble, 1986) until definitive laboratory confirmation of the $O(^1D)$ quantum yield. Current work (Solomon and Abreu, 1987) attempts to reconcile the new quenching rates, latest model atmospheres, and laboratory reaction measurements with VAE 6300 Å dayglow altitude profiles.

Nightglow morphology of the 6300 Å emission at low latitude has been examined by Abreu et al. (1982) and Fesen and Abreu (1987). Transport of ions and electrons by both electromagnetic and tidal effects was shown to be important. Another topic pertaining to transport is the question of $O(^1D)$ thermalization in the nightglow. 6300 Å line widths and Doppler shifts are commonly
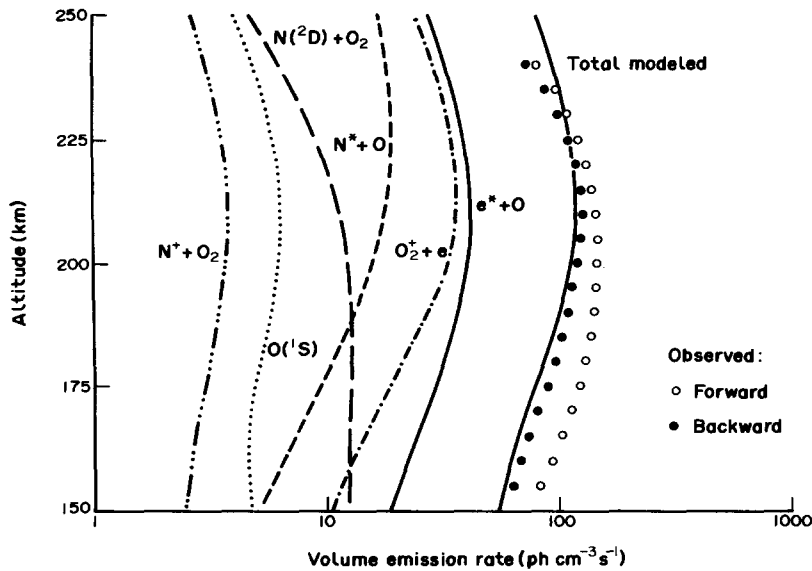
FIG. 5. AURORAL 6300 Å VOLUME EMISSION RATE ALTITUDE PROFILE FROM *AE-C* ORBIT 6136 COMPARING THE MODELED CONTRIBUTIONS FROM VARIOUS SOURCES TO VAE MEASUREMENTS MADE IN THE FORWARD-LOOKING AND BACKWARD-LOOKING DIRECTIONS.
(From Solomon *et al.*, 1987).

taken to be representative of the ambient gas, so quantification of the altitude range where this approximation is reasonable is necessary. Schmitt *et al.* (1981) found that at altitudes above 600 km the VAE emission altitude profile departs from the diffusive equilibrium profile, attributable to a non-thermal component. Recent work (Yee and Dalgarno, 1987) calculating the elastic and excitation exchange cross-sections implies that thermalization is incomplete at much lower altitudes as well (see accompanying paper, Yee, 1988).

5. AURORAL TRANSITION OF OI, 5577 Å

The metastable transition between excited ^{1}S and ^{1}D states of atomic oxygen results in the thermospheric green line emission at 5577 Å. Investigations of the *F*-region green line prior to the analysis of VAE data attempted to determine the sources of metastable O(^{1}S) atoms in the dayglow and at twilight (Wallace and McElroy, 1966; Feldman *et al.*, 1971; Hays and Sharp, 1973). These investigations established that photoelectron impact excitation of atomic oxygen and the dissociative recombination of O_2^+ were sources of O(^{1}S) above 140 km, although the relative magnitude of the two processes was somewhat uncertain.

Analysis of the *Atmosphere Explorer-C* composition and VAE data by Frederick *et al.* (1976) and

Kopp *et al.* (1977a) indicated that besides the sources mentioned above, a third source was required in order to account for the observed emission. The reaction of atomic nitrogen with O_2^+ was identified as the missing source. These sources were quantified in a self-consistent study, in which the complement of instruments on board the *AE-C* satellite for the first time allowed simultaneous measurements of most of the parameters required.

The VAE measurements at 5577 Å have also contributed to the question of the quantum yield of O(^{1}S) from the dissociative recombination of O_2^+. This problem has received recent attention following the laboratory work of Zipf (1980b) who concluded that his earlier (1970) measurement of a high yield of O(^{1}S) at about 0.1 was produced by a source population of vibrationally excited $O_2^+(^2\Pi_g)$ and that yields from the lowest vibrational levels of the ground state ion were considerably less, near 0.02. The variety of yields calculated by aeronomical studies (Hernandez, 1971; Hays and Sharp, 1973; Frederick *et al.*, 1976; Kopp *et al.*, 1977b; O'Neil *et al.*, 1979) could then be attributed to variations in the amount of vibrational development of the recombining O_2^+. Theoretical calculations of potential energy curves (Michels, 1979; Guberman, 1979) support this contention, but Bates and Zipf (1981) subsequently found that vibrational deactivation by atomic oxygen would proceed too rapidly to allow recombination to occur except in the

lowest vibrational states, and suggested an electron temperature dependence of the $O(^1S)$ yield instead.

Using *AE* data, Abreu *et al.* (1983) investigated this question by examining all the relevant parameters in the equatorial nightglow, where the dissociative recombination of O_2^+ should be the only significant source of $O(^1S)$. As shown in Fig. 6, the specific recombination rate was found to depend on the electron to atomic oxygen ratio, supporting the contention of a vibrational distribution effect, and suggesting that while deactivation of O_2^+ by O would proceed rapidly, at high altitudes it would not be complete before recombination occurred. No significant temperature variations were present during the orbits studied, so that question remains unresolved. Recent work by Guberman (1985, 1986) refines the earlier results, showing that recombination from even the $v = 1$, $v = 2$ states into the $O(^1D) + O(^1S)$ channel can be rapid, while that from the $v = 0$ level is not. This channel is favored over the $O(^3P) + O(^1S)$ channel as shown by analysis of the nonthermal $O(^1S)$ line shape measured by the Fabry–Perot interferometer on *DE*-2 (Killeen and Hays, 1983; Yee and Killeen, 1986).

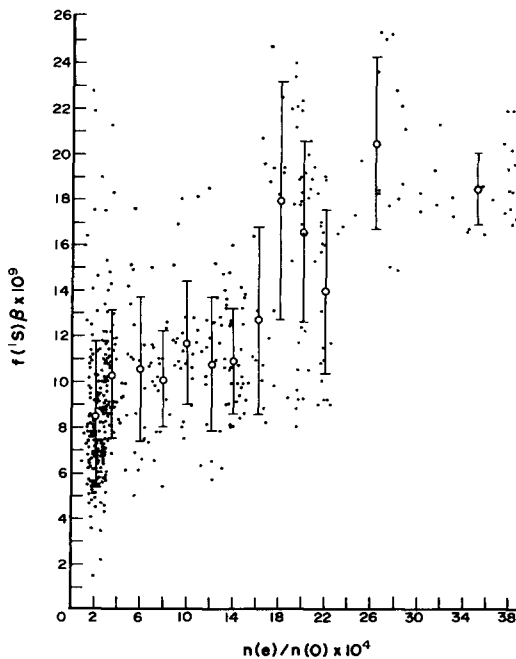The lower thermospheric and meospheric 5577 Å emission has been studied recently by Yee and Abreu

FIG. 6. $O(^1S)$ SPECIFIC RECOMBINATION RATE SHOWN AS A FUNCTION OF THE ELECTRON–OXYGEN RATIO.
Bin averages with statistical error bars are superimposed on the individual data points. (From Abreu *et al.*, 1983.)

(1987) using VAE data. They obtained the volume emission rate profile of the emission using a constrained linear inversion method. Volume emission rate profiles obtained while the satellite altitude was below 150 km were then used to map all available data, obtained at different satellite altitudes and zenith angles, into corresponding zenith intensities. In this fashion a morphological map of the mesospheric 5577 Å emission was generated, in which diurnal variations and the signature of dynamical processes can be observed.

6. NEBULAR TRANSITION OF NI, 5200 Å

The 5200 Å line emission originates from the metastable transition between excited ^2D and ^4S states of atomic nitrogen. The emission line is a direct monitor of the chemically active metastable $N(^2D)$ which plays an important role in the thermospheric nitric oxide cycle (Wallace and McElroy, 1966; Norton and Barth, 1970). Because of the small emission probability $(1.05 \times 10^{-5} \text{ s}^{-1})$ (Garstang, 1956), the emission brightness is very weak, approximately 100 Rayleighs (R) in the dayglow and less than 10 R in the nightglow.

Bates (1978) has summarized the early development of our understanding on the $N(^2D)$ photochemistry. The first theoretical attempts in analyzing the 5200 Å dayglow were based upon rocket measurements (Wallace and Nidey, 1964; Wallace and McElroy, 1966). They considered three possible source mechanisms:

$$N_2^+ + O \rightarrow NO^+ + N(^2D)$$

$$N_2^+ + e \rightarrow N + N(^2D)$$

$$NO^+ + e \rightarrow e + N(^2D)$$
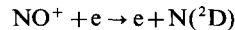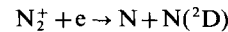
and loss via quenching by electron, O_2 and N_2. Since then, many theoretical and laboratory efforts have been directed at the determinations of the quenching rate coefficients and the yields of $N(^2D)$ atoms from dissociative recombination of N_2^+ and NO^+ ions. Lin and Kaufman (1971), for example, demonstrated that quenching of $N(^2D)$ by N_2 is slow and can be neglected and showed that quenching by O_2 follows not the path

$$N(^2D) + O_2 \rightarrow N(^4S) + O_2$$
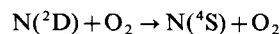
but the reactive path

$$N(^2D) + O_2 \rightarrow NO + O$$
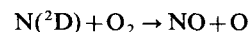
which is of great importance in the production of NO in the daytime thermosphere and aurora.

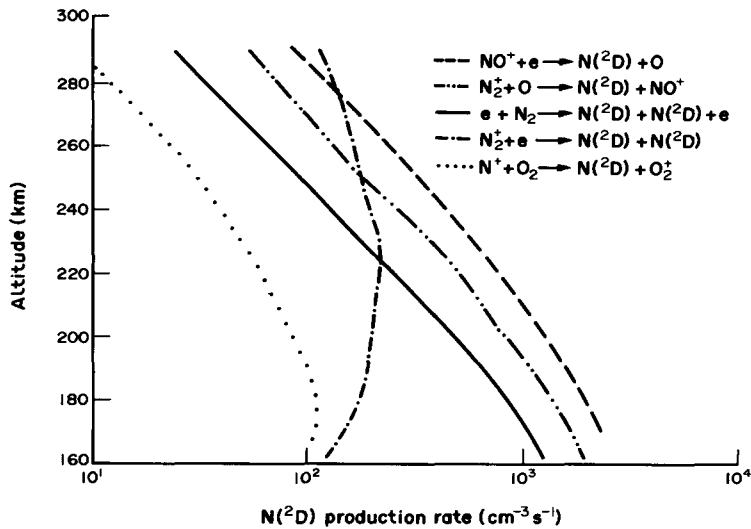The first quantitative investigations of $N(^2D)$ chem-

FIG. 7. PRODUCTION RATES OF $N(^2D)$ FOR AE-C ORBIT 1663 DOWNLEG.
The calculations assume that $N(^2D)$ is produced with the maximum possible efficiency. (From Frederick and Rusch, 1977.)

istry using the VAE database were described in four papers (Rusch *et al.*, 1957b, 1976; Torr *et al.*, 1976; Frederick and Rusch, 1977), which used simultaneous measurements of the 5200 Å emission and the atmospheric parameters to examine the production and loss mechanisms of $N(^2D)$ atoms. Figure 7 illustrates the major source mechanisms of $N(^2D)$, from which it can be seen that $NO^+ + e$ and $N_2^+ + e$ are comparable, with $N_2^+ + e$ becoming the major source at high altitudes. Figure 8 shows the loss rates by quenching with O, O_2 and electrons.

In the early study of Rusch *et al.* (1975b), they deduced yields of $N(^2D)$ atoms and the rate coefficients of quenching reactions, which are different from the values laboratory work had obtained. Subsequent investigations attempted to bring the two into harmony. Disagreement still remains on the quenching rate by atomic oxygen between the measured
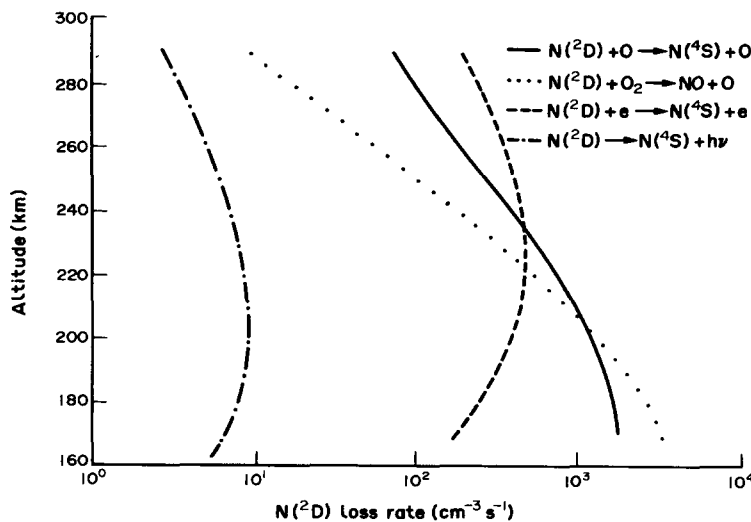

FIG. 8. LOSS RATES OF $N(^2D)$ FOR ORBIT 1663 DOWNLEG.
(From Frederick and Rusch, 1977).

rate coefficient of 1.8×10^{-12} cm^3 s^{-1} (Davenport et al., 1976) and aeronomically deduced values of only 4×10^{-13} cm^3 s^{-1} (Frederick and Rusch, 1977) and 7×10^{-13} cm^3 s^{-1} (Richards et al., 1981). Resolution of this issue is important to the thermospheric chemistry of O(^{1}D) (see Section 4) and NO, since the F-region N(^{2}D)/N(^{4}S) ratio is strongly influenced by this reaction.

7. FIRST NEGATIVE BAND OF N$_2^+$, 4278 Å

The first negative (1NG) band system of N$_2^+$ arises from vibrational–rotational transitions from the B$^2\Sigma_u^+$ configuration to the ground X$^2\Sigma_g^+$ state. The (0,0) band at 3914 Å is one of the prominent auroral features; the (0,1) band at 4278 Å, which is observed by the VAE photometers, only slightly less so. The 1NG bands are excited in the aurora by electron impact ionization of N$_2$. They are quite useful for auroral studies because the transition is allowed so emission occurs without quenching, and because the cross-section for electron impact ionization into the B$^2\Sigma_u^+$ state is proportional to the total ionization cross-section irrespective of the electron energy (cf. Rees, 1963; Dalgarno et al., 1965; Borst and Zipf, 1970). The volume emission rate of a 1NG band is therefore proportional to the rate of ionization of N$_2^+$, which, at low altitudes, is close to being proportional to the total ionization rate.

In the dayglow, the 1NG bands are excited primarily by resonant scattering, as was recognized by Bates (1949). A review and analysis by Broadfoot (1967, 1971) summarizes this effect. An experimental study by Sharp (1974) found resonant scattering to account for 0.90 of the observed daytime emission. Sunlit regions of the aurora such as high altitude cleft aurora may exhibit greatly enhanced 1NG emission due to this effect (Deehr et al., 1980; Sivjee and Deehr, 1981).

In the night aurora, this effect does not occur and the only other known source mechanism is impact ionization by energetic protons. In regions where protons may be neglected, either by direct measurement of their absence or by the absence of characteristic emissions which accompany proton fluxes (such as Hα, Hβ), electron impact ionization may be considered to be the only source. At any rate, the ratio of production of 1NG emission to N$_2$ ionization is probably similar for electrons and protons (Vallance Jones, 1974). Borst and Zipf (1970) determined that for electrons the ratio of ionization into the zeroth vibrational level of the B$^2\Sigma_u^+$ state to total ionization was 14.1. The ratio of radiative transition probabilities $A_{(0,0)}/A_{(0,1)}$ is 3.28 (Shemansky and Broad-

foot, 1971) so there are 46 N$_2$ ionizations for every 4278 Å photon emitted (Link, 1982). This proportionality has been used in auroral models (Rees and Jones, 1973; Rees and Luckey, 1974) to calculate the relationship between 4278 Å emission and total ionization rate; this will depend on atmospheric composition and the penetration depth of auroral electrons, and so varies with the energy of the electron flux. Rees and Luckey (1974) calculated the ratio of 4278 Å emission to electron energy flux to increase from 160 to 210 Rayleighs per erg cm^{-2} for electrons with characteristic energy in the 1–10 keV range; this was essentially verified by combination of ground-based and rocket observation (Rees et al., 1976), satellite measurement (Kasting and Hays, 1977), and ground–satellite coincidence study (Rees and Abreu, 1984). Kasting and Hays found somewhat higher emission than predicted for some AE orbits but attributed the deviation to precipitation of protons too energetic to be in the range detected by the LEE instrument. A rocket–satellite study (Rees et al., 1977) observed lower than expected 3914 Å levels below the coordination altitude but this was considered to be due to temporal variations of the electron flux.

The N$_2$1NG emissions are also commonly used in conjunction with the O(^{1}D–^{3}P) 6300 Å emission to indicate the energy of auroral electrons. Regardless of what processes excite the O(^{1}D) state, it is strongly quenched in the lower thermosphere, so the higher the energy of precipitating electrons the deeper they will penetrate and the lower the 6300 Å/4278 Å ratio. This ratio has therefore been used to obtain an estimate of the energy characteristics of the electron flux (e.g. Rees and Luckey, 1974). But because of the variety of source mechanisms for O(^{1}D) and the dependence on neutral atmosphere composition and density of the loss term, the 6300 Å/4278 Å ratio is not an entirely satisfactory tool. On AE-C, VAE had a mode with the 6300 Å filter on channel one and the 4278 Å filter on channel two and so was well equipped to study this question in conjunction with particle flux measurements. Solomon (1987) found reasonable correspondence between VAE observations of this ratio and model predictions based on measurement of the energetic electron flux by the LEE particle detectors on AE-C.

8. SECOND POSITIVE BAND OF N$_2$, 3371 Å

The second positive (2PG) band system of molecular nitrogen is produced in emission by allowed radiative transition from the C$^3\Pi_u$ to the B$^3\Pi_g$ state. About a fourth of the total system intensity is in the (0,0) emission band located at 3371 Å. The C$^3\Pi_u$ state

is excited in the dayglow and aurora by impact of low energy electrons; by photoelectrons in the former and by secondary electrons in the latter. Its cross-section for electron impact excitation, well established by laboratory measurements (Cartwright *et al.*, 1977), peaks at 14 eV and declines rapidly with increasing electron energy to a factor of 100 lower at 100 eV (Imami and Borst, 1974). Excitation from the ground state is electric-dipole forbidden and cascade from higher states is negligible (Vallance Jones, 1974) so the 2PG system is to good approximation an indicator of the low energy electron flux alone with photons and primary auroral electrons playing no significant role.

The 3371 Å dayglow emission was first identified by Barth and Pearce (1966); the volume emission rate profile was calculated by Nagy and Fournier (1965). Rocket flights have been used to analyze the photoelectron to emission relationship (Doering *et al.*, 1970; Van Tassel *et al.*, 1981; Conway and Christensen, 1985) and the *AE-C* VAE results have been presented (Kopp *et al.*, 1977b). Analysis of data from VAE and the photoelectron spectrometer (PES) on *AE-E* (Hernandez *et al.*, 1983) found a discrepancy between the 3371 Å emission calculated from the photoelectron observations, the emission measured by VAE, and that calculated from a theoretical model, with the PES calculation 20–30% higher than VAE and VAE in turn 30% higher than the model. Conway (1983) has suggested that blending of the Vegard Kaplan (0,9) band in the 3371 Å filter may account for 20–30% of the observed emission, depending somewhat on the rate of quenching of $N_2(A^3\Sigma_u^+)$, which would bring the VAE measurements into accord with the model. The two emissions have similar dayglow profiles, peaking around 150 km, so they are not easily distinguishable. In the aurora, however, the 3371 Å emission may peak in the 100–120 km region while the VK emission, being subject to quenching, will be somewhat higher. This effect was observed qualitatively (Solomon, 1987) in volume emission rate altitude profiles obtained by applying the tomographic inversion technique to VAE auroral measurements.

The auroral N_2 2PG emission has been observed from rocket (Sharp and Hays, 1974) and satellite (Rees and Abreu, 1984) with the former finding good agreement between measurements of the secondary electron flux and the 3371 Å emission and the latter reporting measured emission somewhat lower than modeled. This study used *AE-C* VAE to view the vertical column brightness and the Low Energy Electron (LEE) detector to measure electron flux. The 4278 Å emission was also observed and calibrated with a ground-based photometer as the satellite passed overhead. There was no "ground truth" for

3371 Å, however. The elevated 4278 Å/3371 Å ratio observed by VAE may be attributable to increased atmospheric absorption of the ground brightness at the shorter wavelength. The ground brightness component was subtracted from the signal using the method of Hays and Anger (1978) and Abreu and Hays (1981). The model values of 4278 Å/3371 Å, generally slightly less than unity, have been shown to be in good agreement with aircraft observation (Rees *et al.*, 1976), while the VAE ratios were in the range of 1.2–1.3. The possible contribution of the VK (0,9) band must also be considered. Further analysis of the 3371 Å auroral emission using VAE data to test the validity of models of the secondary electron flux and quantify the VK (0,9) contribution is in progress.

9. RESONANCE EMISSION OF Mg⁺, 2800 Å

Resonance scattering of sunlight by the Mg^+ produces the u.v. doublet at 2795.5–2802.7 Å in the day airglow. Metallic atoms, oxides and ions including those of magnesium are created by ablation of meteoric matter near the 100 km level in the Earth's atmosphere. The compounds of magnesium thus created may diffuse upward into the lower thermosphere where the oxides are reduced to neutral atoms by atomic oxygen and charge exchange with O_2^+ and NO^+ will ultimately ionize the Mg to form Mg^+. The lifetimes of all compounds of magnesium and its molecular ions are less than an hour above 110 km, while Mg^+ has a lifetime in excess of 1 day, rapidly increasing with altitude. Thus, Mg^+ is the dominant form of magnesium in the thermosphere and due to its long lifetime is an excellent tracer species for plasma motions in the *E*- and *F*-regions of the ionosphere, particularly since it is so easily observed photometrically.

The early observations of metallic ions (Vallance Jones, 1956; Duffay, 1958; Broadfoot, 1967) were carried out from the ground by observing the resonance emissions of Ca^+, but with the advent of rocket and satellite mass spectrometry (Hanson and Sanatani, 1970; Goldberg and Blumie, 1970; Grebowsky and Brinton, 1978) an understanding of the morphology began to evolve. The distribution of these ions showed great structure, with patches of Fe^+ observed up to 1000 km. Kumar and Hanson (1980) analyzed a very large database of satellite ion density data collected from *OGO*-6, *AE-C*, and *AE-D*, finding that below 250 km there was a high probability of finding metallic ions at all times. At low latitudes these ions were detected at extreme altitudes from 400 to 1000 km, with a detection probability that peaked strongly in the late afternoon. Optical observations of

the Mg$^+$ emission at 2800 Å at twilight (Borksenberg and Gerard, 1973; Gerard and Monfils, 1974, 1978; Gerard, 1976) looking upward from the 530 km orbit of the ESRO satellite *TD1A* indicated that the emission was often present at evening twilight, but never at dawn, anticipating the results of Kumar and Hanson (1980).

The VAE optical observations at 2800 Å have helped to clarify time history of the morphology of Mg$^+$. The first observations (Gerard *et al.*, 1979) of the time development of the equatorial ion densities indicated low ion densities in the morning, building from the *E*-region into the afternoon, with highly structured profiles in the late afternoon. This development was not regular and exhibited great variability from orbit to orbit. However, the concept of a developing "fountain of ionization" in the daytime, driven by the upward drift of plasma due to the Pederson current associated with the equatorial electrojet, had been proposed by Hanson *et al.* (1972). The complete VAE database for the 2800 Å emission was analyzed (Fesen, 1981; Fesen and Hays, 1982a) using the generalized two-dimensional inversion technique (Fesen and Hays, 1982b) providing Mg$^+$ density profiles throughout the day from 300 to 500 km over the magnetic latitude range from $-25°$ to $+25°$. These results showed clearly that throughout the equatorial region the Mg$^+$ density is very low (<10 cm^{-1}) until at least 10:00 L.T. The density increased sharply at or just North of the equator from 12:00 to 14:00 L.T.,

with continued evolution to higher density in the late afternoon depending upon the longitude sector of the observation. Figure 9 shows the Mg$^+$ ion density contours at 400 km derived from VAE observations.

The general behavior of Mg$^+$ and presumably the other metallic ions is in qualitative agreement with plasma motions (Fesen and Hays, 1983) that produce the equatorial anomaly in the ionosphere. This drift is generally upward at the equator through the day with a reversal occurring at evening twilight, causing plasma to rise over the equator during the day and diffuse downward along magnetic field lines to higher latitudes as the day progresses. This simple symmetric convection is complicated by cross equatorial winds and tides which cause asymmetries in the flow of the dominant O$^+$ plasma. These major ion motions entrap the minor metallic ions and lead to the evolving afternoon equatorial plume of Mg$^+$ observed by the VAE photometers. There is qualitative agreement between the theoretical models of this process and the observations, but details related to tides and the vertical electric fields remain to be quantified.

10. CONCLUSION

This review has demonstrated that a simple photometer in orbit can produce a great deal of information about the upper atmosphere. The vertical profile of many of the airglow features taken at various times of the day allow us to verify the rates of many
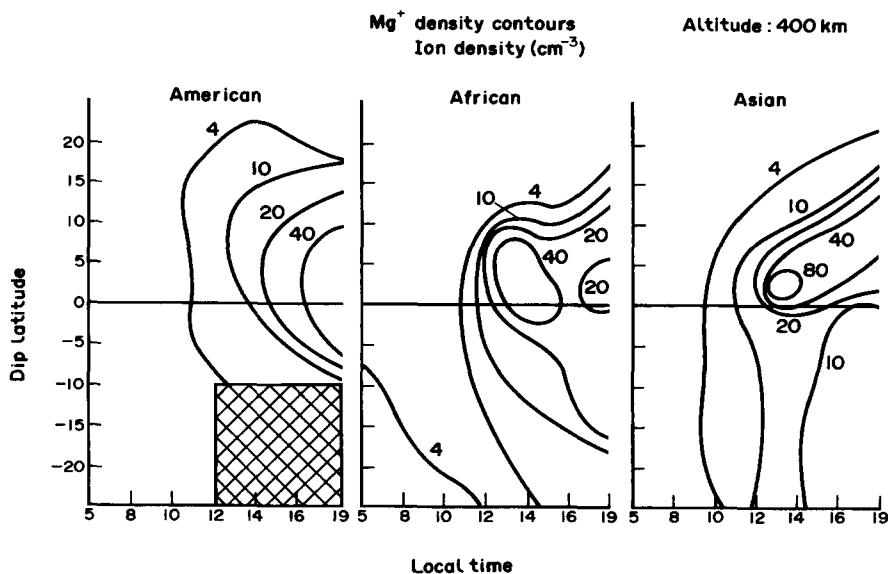

FIG. 9. CONTOUR MAPS OF THE Mg$^+$ DENSITY IN THE AMERICAN, AFRICAN AND ASIAN SECTORS AT 400 km, AVERAGE WINTER SOLSTICE CONDITIONS.
Hatched region indicates lack of data. (From Fesen and Hays, 1982b.)

key reactions that control the photochemistry of the thermosphere and ionosphere. Auroral observations provide additional insight into the physical processes that are occurring. In addition, long lived species such as Mg^+ are tracers of motions in the low latitude ionospheric plasma. Much of the information that was obtained through the VAE investigation has given us a body of knowledge that will allow the state of the thermosphere to be observed by remote sensing techniques in the future. There is a clear need to continue our observations of the thermosphere and indeed extend these observations into the lower thermosphere and mesosphere in the future. The usefulness of optical remote sensing of the upper atmosphere has been demonstrated by VAE and the techniques developed in these studies will play a significant role in future geophysical observations of the Earth from the surface as well as from space.

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