Thermospheric Temperature Dependence of the Atomic Oxygen 6300 Å Emission in the Twilight Airglow

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(Received November 19, 1975; Revised February 17, 1976)

The atomic oxygen 6300 Å emission in the airglow was observed in Logan, Utah during the period from September, 1970 to April, 1971. The emission rate of the morning twilight is found to be correlated with the thermospheric temperature which is derived by using the flux of the solar 10.7 cm radio wave and the geomagnetic disturbance index, K_p . There are considerable fluctuations in the emission rate which are not correlated with either the solar or geomagnetic disturbances.

1. Introduction

Enhancement of the atomic oxygen 6300 Å airglow occurs in the sunlit thermosphere due to the production of $O(^1D)$ by photodissociation of O_2 in the Schumann-Runge (SR) continuum at $\lambda=1300\sim1750$ Å (BRANDT, 1958; ROBLE and NORTON, 1972; LEJEUNE and DALGARNO, 1973). According to the result of the observation by rockets, the dayglow of the 6300 Å emission is excited predominantly in the SR photodissociation process (NAGATA, TOHMATSU, and OGAWA, 1965, 1968; WALLACE and McELROY, 1966). In the twilight the rate of SR photodissociation changes quite rapidly as the terminator of the sunshine migrates across the local atmosphere, causing at the same time a quick change of the emission rate of the 6300 Å airglow. The twilight observation is of particular interest as it provides the ground-sited observers with a useful means to remote-sense the vertical distribution of O_2 in the thermosphere.

It is theoretically expected that the diffusive equilibrium is quickly established for the O_2 in the thermosphere in spite of variations in the atmospheric temperature and the baseline concentration of O_2 in the lower thermosphere, that is the turbosphere (e.g. CIRA, 1972; JACCHIA, 1965). The total abundance of O_2 in the thermosphere is basically determined by a combining effect of its photodissociation by solar ultraviolet radiations and its transport in various

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modes of atmospheric motions and diffusions. Meanwhile, its relative population profile in the vertical direction depends on the thermospheric temperature which is in turn controlled by the solar activity, or more explicitly by the amount of the solar energy input in the forms of ultraviolet radiations and energetic charged particles.

NOXON and JOHANSON (1972) discovered a prominent seasonal variation in the emission rate of the 6300 Å twilightglow; the maximum emission rate appearing in early summer and the minimum in early winter. They speculated the cause of the seasonal variation to be due to the vertical transport of O₂ associated with either a general circulation-type wind or the turbulent diffusion in the lower thermosphere. The solar control of the 6300 Å twilightglow is expected too, as a consequence of a change in the atmospheric temperature, in response to changes in the solar ultraviolet irradiance and energetic particle flux as well. In a statistical analysis of the morning twilight, SUZUKI and OGAWA (1975) showed that the emission rate of the 6300 Å airglow had a positive correlation with the thermospheric temperature which had been deduced from the flux of the 10.7 cm solar electromagnetic radiation along with the planetary geomagnetic disturbance index K_p .

2. Observation

Observations of the airglow were carried out on the campus of Utah State University in Logan, Utah (41°44'N, 111°48'W) from September, 1970 to April, 1971. The airglow photometer is a tilting filter-type which was originally designed to measure the four atomic emissions in the polar aurora, H 4861 Å, O 5577 Å, Na 5893 Å and O 6300 Å. Details of the instrumental design and the procedures for the photometric calibration are described by FUKUNISHI and TOHMATSU (1973).

The photometer was directed to the zenith through the roof of the observatory to receive the radiation from the sky in a circular field of view of 3 degrees in diameter. The observations were usually continued under clear sky conditions after dark through midnight to the morning of the next day until the solar zenith angle (χ) reached a value of 95 degrees. For the values of χ smaller than 95°, and with the filter's transmission bandwidth of 5 Å, the background of the signal due to atmospheric Rayleigh scattering of the solar radiation became too high to be eliminated from the observed signal.

3. Data Analysis

The observations were carried out in 23 clear mornings from September,

Fig. 1a. Apparent emission rate of the atomic oxygen 6300 Å morning twilightglow observed in Logan, September-October, 1970.

Fig. 1b. Apparent emission rate of the atomic oxygen 6300 Å morning twilightglow observed in Logan, October-November, 1970.

Fig. 1c. Apparent emission rate of the atomic oxygen 6300 Å morning twilightglow observed in Logan, November-December, 1970.

Fig. 1d. Apparent emission rate of the atomic oxygen 6300 Å morning twilightglow observed in Logan, January–February, 1971.

Fig. 1e. Apparent emission rate of the atomic oxygen 6300 Å morning twilightglow observed in Logan, February–April, 1971.

1970 to April, 1971. The χ -dependence of the observed emission rate for each morning is displayed in Figs. 1a–e for five successive groups of observational periods. The experimental errors both from the photometry and data processing are estimated to be within 10 percent.

Besides the SR photodissociation, there are several excitation sources of the 6300 Å airglow. Therefore, it was necessary that the elimination of these extra components from the observed emission rate be made with caution.

The dissociative recombination of O_2^+ with electrons (DR) in the *F*-region ionosphere is a well-known source of the excitation of the 6300 Å emission. The nightglow of the 6300 Å emission in non-auroral conditions can fully be attributed to the DR excitation source. The emission rate can be related to the

F -region parameters, the virtual height ($h'F$) and the critical frequency (f_oF2) by using Barbier's empirical relation (BARBIER, 1959, 1962; DELSEMME and DELSEMME, 1960; CARMAN and KILFOYLE, 1963; PETERSON, VAN ZANDT, and NORTON, 1966) as

$$Q(\text{SR}) = A(f_oF2)^3 \exp [-(h'F - 200)/H] + B, \quad (1)$$

where H stands for the scale height of O_2 in the thermosphere, and A and B are constants which are determined by the least squares fit of the observed data.

To estimate the DR component of the emission we used the hourly values of the f_oF2 and $h'F$ from Boulder, Colorado ($40^\circ 00' \text{N}$, $105^\circ 14' \text{W}$). A time lag of 26 minutes was considered to compensate for the difference in longitude between Boulder and Logan. The constants A and B were determined by the least squares fit of the observed data for $\chi > 105^\circ$. As described in a later section the SR part of the emission, $Q(\text{SR})$ became prominent only for $\chi < 105^\circ$. However, the emission rate at $\chi = 105^\circ$ after subtracting $Q(\text{DR})$, obtained in the above procedure, still had a residual of 15 to 25 rayleighs. The origin of this residual is uncertain. A minor fraction of the residual may be ascribed to the excitation of O due to geomagnetically conjugate photoelectrons (CARLSON, 1966; COGGER and SHEPHERD, 1969; DEEHR, 1969; INGHAM, 1969; NICHOL, 1970; OKUDA and MISAWA, 1969; NOXON and JOHANSON, 1970), but we infer that the majority of this arose from the failure of Barbier's formula to approximate the DR source in the dawn ionosphere. In the present analysis we assume that the residual emission rate at $\chi = 105^\circ$ gives a constant baseline to that for $95^\circ < \chi < 105^\circ$, as expressed in the equation below.

$$Q(\text{SR}, \chi) = Q(\text{OBS}, \chi) - Q(\text{DR}, \chi) - [Q(\text{OBS}, 105^\circ) - Q(\text{DR}, 105^\circ)], \quad (95^\circ < \chi < 105^\circ). \quad (2)$$

4. Derivation of the Thermospheric Temperature

The thermospheric temperature referred to in this article is determined using the empirical formula proposed by JACCHIA (1965) and ROEMER (1971). Their formula is based on numerous data on the atmospheric drag on artificial satellites. According to Jacchia the coldest temperature (in Kelvin) of the thermopause in the world, for given solar and geomagnetic activity conditions, is given by

$$T_c(t) = 379 + 3.24\bar{F}_{10.7} + 1.3[F_{10.7}(t) - \bar{F}_{10.7}] + \Delta T_\infty, \quad (3)$$

with the correction term for the geomagnetic activity,

$$\Delta T_\infty(t) = 28.0\bar{K}_p(t - \Delta t) + 0.03 \exp [\bar{K}_p(t - \Delta t)], \quad (4)$$

where $F_{10.7}(t)$ and $\bar{F}_{10.7}$ denote respectively the flux of solar 10.7 cm radio emission at the time of observation (t), and its average for the past three solar rotations, in units of $10^{-22} \text{ W} \cdot \text{m}^{-2} \cdot \text{Hz}^{-1}$, and $\bar{K}_p$ the 0.4-day mean of the planetary geomagnetic disturbance index. A time-delay of $\Delta t = 5.5$ hours is introduced usually to obtain the best statistical fit to the observations. The temperature at a particular latitude and time of the day and year can be modeled by using a

Table 1. Solar 10.7 cm fluxes, geomagnetic K_p indices and the thermopause temperatures at the time of the 6300 Å airglow observations.

Date	(1) $F_{10.7}$	(2) $(F_{10.7})_{\text{mm}}$	(3) $K_p(t)$	(4) $K_p(t-7_{\text{hrs}})$	(5) ΣK_p	(6) $\Delta T_{\infty}(K)$	(7) $T_{\infty}(K)$	(8) Remark
1970, Sep. 30	138.7	144.6	2	3+	16	94	961	M
Oct. 7	128.0		1	1+	7-	37	890	Q
16	140.2		5+	1	28-	28	897	D
17	134.0		3	4-	33	104	965	D
18	131.5	147.3	3	5-	32+	140	991	D
23	149.0		4	4-	32	104	985	D
27	187.1		1-	1+	9	37	969	Q
28	191.8		3-	3-	22+	75	1013	M
29	190.3		3	2-	21-	47	983	M
Nov. 2	163.3		2	2-	8+	47	965	Q
14	179.1	158.5	1-	2	15-	56	989	M
15	186.1		1	1-	8-	19	961	Q
Dec. 12	172.6		1+	0+	7-	9	918	Q
26	124.0	148.0	0	2-	8+	47	892	Q
1971, Jan. 4	134.7		3+	4-	25-	104	982	D
21	169.2	157.4	3-	4-	21+	104	1030	M
28	166.7		4-	4	36	114	1037	D
Feb. 1	170.1		3	4	23	114	996	D
21	136.2	134.4	2+	3-	14	75	921	Q
22	143.6		0	2+	7-	66	922	Q
Mar. 20	109.1		1	4	19	66	832	M
30	103.4	110.8	2	0+	15	28	786	Q
Apr. 1	102.9	117.5	1	5-	18+	66	837	M

- (1) $F_{10.7}$ one day prior to the observation.
- (2) Monthly mean of $F_{10.7}$.
- (3) K_p at the time of observation.
- (4) K_p 7 hours prior to the observation.
- (5) Sum of K_p 's of the observed day.
- (6) The correction term of the thermopause temperature for the geomagnetic activity.
- (7) The computed thermopause temperature.
- (8) Q: quiet, M: medium disturbed, D: disturbed (see text).

formula containing T_c in such a manner that

$$T_\infty(\phi, t) = T_c(t)f(\phi, t, t_Y), \quad (5)$$

where f is a complicated function of latitude (ϕ), local time (t) and time of the year (t_Y) whose explicit mathematical expressions can be seen in the original text (JACCHIA, 1971; CIRA, 1972). Values of T_c and the derived local thermopause temperature (T_∞) for the 23 mornings of the observation are listed in Table 1 along with the data for $F_{10.7}$ and K_p .

5. Computation of the Theoretical Emission Rate

The 6300 Å airglow emission rate at the zenith arising from the SR photodissociation can be formulated, neglecting the atmospheric extinction, as (SUZUKI and OGAWA, 1975)

$$Q(\text{SR}) = \int_0^\infty \frac{A_{22} dz}{A_{21} + A_{22} + k_1 n(\text{N}_2, z) + k_2 n(\text{O}_2, z)} \int_{1300 \text{ Å}}^{1750 \text{ Å}} \sigma(\lambda) F(\lambda) n(\text{O}_2, z) \exp \left[-\sigma(\lambda) \int_{\text{path}}^{\text{opt.}} n(\text{O}_2, z') dz' \right] d\lambda. \quad (6)$$

The symbols used in (6) have the following meanings and values:

λ : Wavelength.

z : Altitude.

$A_{21} = 2.18 \times 10^{-3} \text{ sec}^{-1}$, $A_{22} = 6.92 \times 10^{-3} \text{ sec}^{-1}$: Probabilities of the spontaneous transitions for $^1\text{D}_2 - ^3\text{P}_1$ and $^1\text{D}_2 - ^3\text{P}_2$ (WIESE, SMITH, and GLENNON, 1966).

$k_1 = (5 \pm 2) \times 10^{-11} \text{ cm}^3 \cdot \text{sec}^{-1}$, $k_2 = (6 \pm 2) \times 10^{-11} \text{ cm}^3 \cdot \text{sec}^{-1}$: Coefficients of

Fig. 2. A comparison of the theoretical emission rate of the 6300 Å twilightglow in the Schumann-Runge photodissociation process, with the observed emission rate averaged for three groups of geomagnetic disturbance condition, Q (White circle): $\Sigma K_p \leq 15$, M (Filled square): $15 < \Sigma K_p < 23$, and D (Filled circle): $23 \leq \Sigma K_p$.

the quenching by N_2 and O_2 (NOXON, 1970).

$n(N_2, z)$ and $n(O_2, z)$: Number densities of N_2 and O_2 (JACCHIA, 1971; CIRA, 1972).

$\sigma(\lambda)$: The photodissociation cross section of O_2 in the Schumann-Runge continuum (METZGER and COOK, 1964).

$F(\lambda)$: Solar ultraviolet irradiance (WINDING, PURCELL, and SANDLIN, 1970).

The solid curves in Fig. 2 illustrate the χ -dependence of the SR emission rate which is calculated using formula (6) for the atmospheric densities corresponding to the thermopause temperature of 800, 900 and 1000 K respectively. To compare with the theoretical emission rate we divided the observed data in three groups according to their corresponding K_p values; Quiet ($\Sigma K_p \leq 15$), Medium ($15 < \Sigma K_p < 23$) and Disturbed ($23 \leq \Sigma K_p$). The average emission rate for each of Q, M and D groups is obtained and illustrated in Fig. 2. We observe that the calculated χ -dependence for $T = 900 \pm 100$ K represents fairly well the observed feature of the 6300 Å twilightglow variation.

6. Results and Discussion

In order to see more readily the correlation of the 6300 Å emission rate with the thermospheric temperature, we produced a scatter diagram of these quantities for various phases of the morning twilight with values of χ ranging between 95° and 105° . The temperature was basically determined by using formulae (3) and (5), but in this case an allowance had been made for the time lag Δt to vary from zero to 11.5 hours. Figs. 3a, b and 4a, b are examples of

Fig. 3. The scatter diagrams of the 6300 Å emission rate and the thermopause temperature. (a) $\chi = 100^\circ$, $\Delta t = 2.5$ hours, (b) $\chi = 99^\circ$, $\Delta t = 2.5$ hours.

Fig. 4. The scatter diagrams of the 6300 Å emission rate and the thermopause temperature. (a) $\chi=100^\circ$, $\Delta t=5.5$ hours, (b) $\chi=99^\circ$, $\Delta t=5.5$ hours.

such scatter diagram for the four cases of $\chi=99^\circ$ and 100° , combined with $\Delta t=2.5$ and 5.5 hours. The solid straight line in each figure is the least square fit to the observation, while the dashed straight line represents the theoretical temperature dependence of the emission rate which was calculated by using (6). The correlation coefficient was calculated too, for each combination of χ and Δt , to find the best correlation of 0.55 for $\chi=99^\circ \sim 100^\circ$ and $\Delta t=2.5$ hours. Also we notice that the calculated emission rate gives the best agreement with the observations for the same values of χ and Δt .

There are several reasons to infer why the best correlation is found for $\chi=99^\circ \sim 100^\circ$. For $\chi=99^\circ \sim 100^\circ$, the emission of the 6300 Å airglow is most efficiently produced at an altitude between 250 and 300 km (NOXON and JOHANSON, 1972), and it is the region of the atmosphere where the concentration of O_2 changes drastically in response to the solar and geomagnetic disturbances. In fact, if the temperature raised from 600 to 1500 K, according to CIRA, 1972 Model Atmosphere, the concentration of O_2 would increase by a factor of ten at a level of 200 km, 40 at 250 km and by more than 200 at 300 km. For larger values of χ , on the other hand, only the very high and rare part of the atmosphere with a lower concentration of O_2 is illuminated by solar ultraviolet radiations so that the production of $O(^1D)$ in the SR photodissociation process is quite reduced. Consequently if the DR part of the emission plays the major role in the observed emission rate, then the detection of the SR component involves a risk of having large errors. The SR emission rate increases considerably when $\chi < 98^\circ$. However, the atmospheric Rayleigh scattering of solar radiations increases more rapidly than the SR emission, thus introducing a possibly large

Fig. 5. Time variations of the 6300 Å emission rate in the morning twilight ($\chi=100^\circ$), the geomagnetic disturbance index, ΣK_p , and the 10.7 cm solar radio wave flux, $F_{10.7}$.

photometric error in the data. In conclusion, we find that the observation at $\chi=99^\circ \sim 100^\circ$ is most appropriate both from experimental and theoretical points of view, to study the correlation between $Q(\text{SR})$ and T_∞ .

The geomagnetic control of the emission rate is more complex and cannot be explained in a straight forward manner. It should be noted here that the satellite determination of the temperature is related to the integral and large-scale feature of the atmosphere due to solar and geomagnetic activities, while the emission rate of the 6300 Å airglow discussed here is related to the localized structure of the thermosphere at a particular latitude and altitude. The correction term for the temperature (4) has been deduced in a statistical analysis of the atmospheric drag of a satellite in its entire orbits, therefore it should be considered to represent merely a globally averaged status of the temperature and composition modifications due to geomagnetic disturbances. Heating of the upper atmosphere arises from various modes of magnetospheric-thermospheric interaction; the heating of the polar region atmosphere by auroral particles, the plasma heat conduction in the magnetosphere through the plasmasphere, Joule heating of the ionosphere by the electrostatic field resulting from geomagnetic disturbances, and so on. The time constant of the thermosphere in response to these heating processes is difficult to determine, but we expect whatever the mechanism involved may be that the disturbance effects propagate from a higher latitude to a lower one and also from a higher altitude to a lower one as we infer from the electromagnetic as well as plasma-dynamical structure of the magnetosphere. The time lag for K_p in (4) may then depend on the altitude as well as the latitude of the observation.

In the second point, the disturbances should be understood in a statistical sense in the present analysis. Fig. 5 illustrates the daily variations of $Q(\text{SR}, 100^\circ)$, ΣK_p and $F_{10.7}$ during the observation period. The correlation between

$Q(\text{SR}, 100^\circ)$ and ΣK_p is obscure in this simple comparison in the figure. $Q(\text{SR})$ decreased gradually as $F_{10.7}$ declined continuously from October through April. However, we notice in this figure a number of fluctuations of $Q(\text{SR}, 100^\circ)$ which are unassociated with the changes in $F_{10.7}$ or K_p . Especially prominent is a quasi-periodic variation observed in October, 1970 which occurred without any enhancement of $F_{10.7}$ and K_p . Other causes of such fluctuation may be sought more specifically in the lower thermosphere. As shown by a number of other experimental evidences, the abundance of O_2 in the thermosphere is critically dependent on its baseline concentration at the turbopause altitude. The baseline concentration of O_2 will be controlled by the strength of the mass transport provoked by the atmospheric motions in various time and spatial scales; the turbulence, the gravity wave-type oscillations, the planetary waves, the general circulation of the atmosphere, and so on. The diffusive equilibrium of O_2 will be quickly established in the upper region of the thermosphere corresponding to a change of the baseline concentration in the lower-lying turbosphere. These intrinsic changes in the atmospheric composition could cause the fluctuations of the emission rate of the 6300 Å twilight-airglow which are independent of the solar and geomagnetic activities.

The observatory and some experimental facilities for the present observation were provided by the courtesy of the Center for Research in Aeronomy, Utah State University when the senior author was a Residential Foreign Research Scientist of the National Science Foundation in 1970–71. The authors express their thanks to Dr. Clayton Clark, Director of CRA for his interest in and arrangements for conducting the experiment. We are grateful to Drs. Toshihiro Ogawa and Katsuhisa Suzuki of GRL for their critical reading of the manuscript.

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