RELATIONS BETWEEN THE AIRGLOW AND THE MESOSPHERIC CIRCULATION

by

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ABSTRACT

The airglow emissions of the vibration-rotation bands of the hydroxyl molecule, the atomic sodium D lines, and the green line of atomic oxygen are investigated. The observational and theoretical evidence regarding the nature and cause of the emission, the emission altitude, the variation of the emission features in both time and space, and the correlation ofthe emission with other airglow and geophysical phenomena is discussed for each of these three components of the airglow. Relations between the features and the variations of these three airglow components and the mesospheric circulation are proposed. It is suggested that the turbulence, tides, and circulation of the upper atmosphere are directly related with the variations of the airglow. The rotational temperature of the OH emission seems to be a good indication of the temperature and the temperature variations at the OH emission altitude.

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INTRODUCTION

In the early part of this century astronomers recognized a faint terrestrial component of light in the night sky. This light could not be accounted for by the scattering by the earth's atmosphere. The light originates in the earth's atmosphere at altitudes between 70 and 120 km and is known as airglow. Studies of the airglow have shown that it consists of a number of distinct lines and bands at various wavelengths and of differing intensities. Chamberlain (1) has made a thorough study of the information about the airglow published through 1959.

The airglow exists at all times in the upper atmosphere. It is termed dayglow and nightglow depending upon the local time of observation. In general, the dayglow and nightglow have different characteristics. At dawn and dusk certain components of the airglow are enhanced and their emission is known as the twilightglow.

In general, the energy for the airglow comes from the chemical recombination of substances dissociated by the action of sunlight. The recombination process is not necessarily direct. The dissociated components may undergo several different chemical transformations before returning to the original (undissociated) state. The chemical reactions give up the stored solar energy by exciting various electronic, vibrational, and rotational states of the reaction products. The airglow results from the return of the chemically excited reaction products down to their normal ground states. Some components of the dayglow and twilightglow emit as a result of resonance and flourescence of sunlight.

In this study the nightglow and twilightglow produced by the hydroxyl molecules (OH) and neutral sodium atoms (Na) are examined along with one component of the airglow produced by oxygen atoms (O). The variations of these components of the airglow with season, altitude, and latitude as well as their diurnal variations are presented. The mechanisms thought to be responsible for the production of these components are discussed. Correlations with various geophysical phenomena and other airglow components are examined. Each of the three airglow emissions is discussed separately and comprise the first three parts of this study. Information found in the literature through 1963 along with some published during the first part of 1964 is included in this study.

Several questions of meteorological interest arise from a study of the airglow. How are the dynamics of the atmosphere at and near the altitudes of the airglow regions connected with the observed variations in airglow intensity? Can the known facts concerning the upper atmospheric circulation and structure be fitted into a scheme which will predict the variations observed in the airglow? Can transport processes of the upper atmosphere supply the energy for airglow emissions during the arctic winter? Is it possible for substances emitting radiations in the airglow to be used as tracers of upper atmospheric motions? Do short term airglow variations indicate changes in the upper atmospheric circulation pattern? Answers to these questions would indeed be a valuable addition to the knowledge about the upper layers of the atmosphere.

The fourth section of this study is an attempt to find relations between the dynamics and structure of the upper atmosphere and the variations of the airglow. Eddy transport processes are discussed in relation to their ability to supply energy for the airglow and also to supply the substances which emit airglow radiation. Possible relations between vertical eddy transport and airglow altitudes and intensities are presented. A circulation mechanism is proposed which will explain some of the latitudinal and seasonal variations of the airglow as well as some of the observed features of the mesospheric temperature structure.

HYDROXYL NIGHTGLOW

A. Discovery

In the summer of 1948 Meinel (2) obtained spectra of the near infrared region (7700-8900Å) of the night sky. In this region he discovered a large number of intense emission features, apparently due to molecular radiations. Several other investigators attempted to associate these features with various systems of N_2 but were unsuccessful. In 1950 Meinel positively identified many of these features as belonging to the vibration-rotation system of the OH molecule (3,4). He also defined the vibrational and rotational structure of the OH molecule as well as the various constants of the system and predicted the positions of more bands farther in the infrared which he believed to be much more intense than the strong bands already observed (5). Later observations taken in Dec. 1949 and Jan. 1950 gave excellent agreement between theoretical and observed wavelengths of OH bands (5). Kron (6) investigated spectra farther into the infrared (9000-11000Å) and identified bands of OH predicted by Meinel and eliminated the possibility of their belonging to N_2. Roach, Pettit, and Williams (7) looked at spectra they had obtained during the summer of 1948 for the region from 6400Å and identified many of the Meinel OH bands.

In a later work, Meinel (8) predicted the relative intensities of the various OH bands and determined the temperature of the emitting layer from the rotational structure of some of the vibrational transitions.

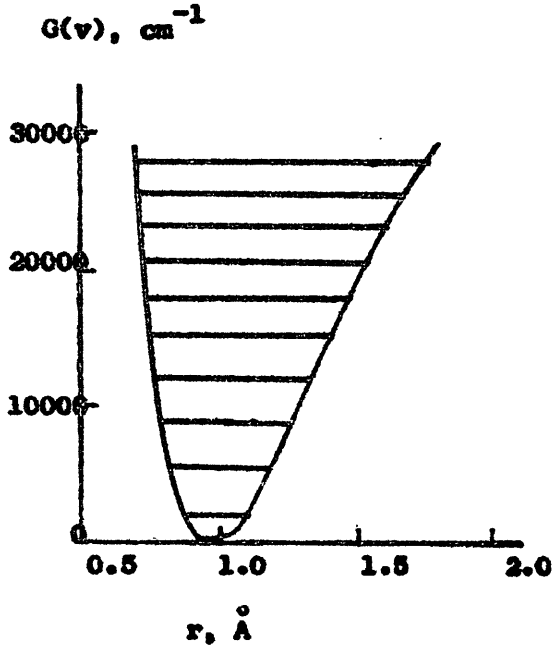
He was able to observe all vibrational levels up to v = 9 (3.248eV). A diagram of the ground state vibrational structure is shown in Figure 1. Meinel concluded that such a high level of energy could not be obtained by thermal excitation and that the excitation mechanism was due to an energy resonance mechanism which selectively populated the v = 9 level. The other lower vibrational levels were then populated by cascading downward from the v = 9 level. From the intensities observed in the near infrared and theoretical calculations Meinel predicted that the total emission by the OH bands was far stronger than all other airglow emissions combined but that most of the emission lay so far in the infrared (3-4μ) so as to be unobservable due to the lower atmospheric thermal emission. All in all, Meinel not only identified the OH emission in the nightglow but also laid a groundwork for further investigations and initiated several lines of investigation himself.

B. Nature of the Emission

The OH emission is composed of vibration-rotation bands formed in the ground state ($^2\Pi$) of the OH molecule. Each band is composed of P, Q, and R branches with the R branch forming the band head on the short wavelength side and the rotational structure evidenced in the R and P branches. The wavelengths of the bands range from 3817Å for the 9-0 band to 4.47μ for the 9-8 band. In general the transitions with highest Δv have the least probability of occurring and those transitions with the highest probability of occurrence ($\Delta v = 1$) have the lowest energy differences and hence emit at the longer wavelengths. The

FIGURE I

X $^2\Pi$ Ground State: Vibrational Structure*



v	G(v) cm^{-1}	$G_o(v)$ cm^{-1}	$G_o(v)$ eV
0	1847.78	0.00	0.000
1	5417.41	3569.63	0.443
2	8821.36	6973.58	0.865
3	12061.61	10213.83	1.266
4	15139.28	13291.50	1.648
5	18054.52	16206.74	2.009
6	20805.95	18958.17	2.350
7	23392.1	21544.3	2.671
8	25806.7	23958.9	2.970
9	28043.5	26195.7	3.248
10	30095.1	28247.3	3.502

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* After Krassovsky, V. I., N. N. Shefov and V. I. Yarin, Planet. Space Sci., 9, 895, 1962.

band origins and wavelengths have been computed by several authors (9,10) and a complete listing is given by Chamberlain (1).

The transition probabilities are given by several authors. The most recent work has been done by Ferguson and Parkinson (11) who used a fifth degree polynomial to represent the dependence of the dipole moment on inter-nuclear distance in order to calculate transitional probabilities but concluded that they cannot be uniquely determined since no observations of $\Delta v = 1$ transitions are available. Relative and absolute intensities of the various OH bands have been determined by several investigators. Meinel (8) was the first to report intensity ratios for bands in the near infrared. Roach, Pettit, and Williams (7) reported the first absolute intensities as well as relative intensities for many bands out to 1.16μ. Later work in 1955 for the 1 to 2μ region was hindered by water vapor absorption (12). Dufay (13) gave the observed relative intensities of the entire $\Delta v = 4$ sequence in 1957 and in that same year Harrison and Jones (14) reported absolute intensities for the $\Delta v = 3$ and $\Delta v = 2$ sequences. Additional observations of absolute and relative intensities from more accurate observations have been reported since 1958 (15,16,17,18,19,20). Beyond 2.5μ atmospheric thermal emission overcomes the OH emissions. In addition, in the 1.5 to 2.5μ region water vapor bands strongly attenuate the OH emission. Neither of these effects could be eliminated, even by choosing an exceptionally cold and clear night for observations (at 52° N) (19). Needless to say, there is considerable disagreement among the absolute intensities reported and even among the relative intensities. Part of this is due to the variability of the OH emissions but, as Chamberlain points out, the intensity measurements are quite subject to error (1).

Meinel (8) originally postulated that the excitation of OH was due to an energy resonance which selectively populated the ninth vibrational level and that the remaining levels were populated by cascading from the ninth level. At the time there was no detailed theory for the energy distribution produced by a chemical reaction and it seemed that resonance would be important. Later work has discredited this idea and Chamberlain and Smith (10) showed that there was no reason to expect the ninth or any other level to be preferentially excited. They proposed that all levels would be populated equally. This latter idea is in better agreement with the observations of relative band intensities and theoretical transition probabilities (19,21).

Another problem is the absence of excitation of vibrational levels greater than the ninth. The solution may lie in the energy available from the exciting chemical reaction. The exciting reaction(s) has yet to be irrevocably established. Shefov (22) has reported the observation of the tenth vibrational level at Zvenigorod in Dec. 1962 but this is the only report of a level greater than the ninth in the OH nightglow since the identification of the OH bands by Meinel in 1950 and hence its validity is rather doubtful. Krassovsky (23) has argued that if the tenth and higher levels are excited they will rapidly disappear through collisions with O_2 or N_2 before emission and hence not be observed in the nightglow. He states that OH($v \geq 10$) has sufficient energy to react chemically with O_2 and N_2 to form O_3 and N_2O respectively.

The energy released by the OH band far exceeds that of all other airglow components combined. The observed energy release in the 1 to 2μ region by OH is approximately 3.4×10^{-2} erg/cm^2 steradian sec (14). The total energy release by all OH bands is estimated to be several ergs by Ferguson and Parkinson (11) and 3.2 ergs/cm^2 sec by Chamberlain and Smith (10). Most of this cannot be observed due to atmospheric thermal emission. Wallace (24) estimates that 85% of the photons are emitted in the 2.8 to 4.5μ region. On the average each OH molecule emits four to five quanta before reaching the ground vibrational state (10,24). The maximum possible emission is nine quanta for a molecule excited into the ninth vibrational level and cascading down with a $\Delta v = 1$ at each transition. Using the assumption of exclusive excitation of the ninth level various investigators estimated the excitation rate of OH molecules to be between 10^{10} and 1.5×10^{12}/cm^2 sec (14,17,25,26,27). Using equal excitation of all levels, which is more likely, the rate of newly formed excited OH molecules is estimated to be 1 to 1.6×10^{12} OH/cm^2 sec (10,11,24,26,27). The actual rates are probably somewhat higher than the values given here due to collisional de-excitation of excited OH by other substances.

Determination of the rotational structure of the OH bands can furnish the temperature of the OH molecules. When the excited OH molecules cascade to lower vibrational levels, higher and higher rotational states are reached and the rotational temperature is thus increased. If the excited OH molecules undergo sufficient collisions with the ambient particles the rotational states will be redistributed and become in equilibrium with the ambient temperature. The lower vibrationally excited states have longer lifetimes (10^{-2} sec for v = 1) than the more highly excited states (2×10^{-3} sec for v = 9) and hence are more likely to have rotational temperatures in equilibrium with the ambient temperature (28). Several investigators have calculated that below 90 km the excited OH molecules will undergo sufficient collisions before excitation such that collisional redistribution of the rotational states will occur so that the rotational temperature will be equal to the ambient temperature (28,29,30). The further above 90 km that the emission takes place, the more likely it is that the higher vibrational levels will not have rotational temperatures in equilibrium with the ambient temperature. Also the rotational temperatures determined from the various OH bands will exhibit greater differences since the lower vibrational levels will be more nearly in equilibrium than the higher levels. A possible example of this effect can be seen in observations at Zvenigorod reported by Krassovsky and his co-workers (31). They encountered cases where the rotational temperature was greater than 270°K for vibrational levels six to nine while simultaneously observing lower rotational temperatures for the fourth and fifth vibrational levels. This result might also be due to different excitation mechanisms dominating at different altitudes with different temperatures (31,28).

A difficulty in the interpretation of the rotational temperatures is due to the fact that the OH emission occurs over a significant range of altitude in the atmosphere. The intensity of the emission varies with altitude and the ambient temperature varies with altitude also. The measured rotational temperatures will then be an average weighted according to the intensity and temperature variation with altitude in the emitting layer. McPherson and Jones (29) feel that OH rotational temperatures cannot be compared with other temperature measurements until the relation between intensity and temperature with height are known. Hunten (32) brings out the point that rotational temperature observations are affected by atmospheric absorption.

Recent observations in Russia of OH rotational temperatures, intensities, and relative population of the vibrational levels of the OH bands have shown some interesting features. Some observations during the winter 1957-58 at three locations showed that the intensity of emissions with different rotational temperature was the same (33). Observations by Fedorova (16) at 42°N showed no relation between rotational temperature and initial vibrational level. Observations at Yakutsk (62°N) during the winter 1959-60 show that for rotational temperatures greater than 250°K and high intensity there is an intensity increase accompanying rotational temperature increases (31). For rotational temperatures less than 250°K there was no relation between temperature and intensity. Another feature of the Yakutsk observations was the tendency for an increase in the population of the higher vibrational levels with increasing rotational temperature. Observations at Zvenigorod (56°N) during the same period disclosed no relation between the rotational temperature and intensity of the OH bands for any temperature range (31). From later observations at Zvenigorod, Berg and Shefov (34) report that the OH intensity depends directly upon the rotational temperature but that they were not at all well correlated.

C. Altitude of the OH Emission

Several methods are available for estimating the height of the layer from which the OH nightglow originates. The earliestis the Van Rhijn Method which assumes that the emission layer is a thin spherical shell around the earth and of uniform intensity. The height of the layer is deduced by observing the variation of the intensity of emission with zenith angle. Many of the earlier height determinations were made by this method. From observations taken in southern California during the summer of 1948, Roach, Pettit and Williams (7) obtained an altitude of 70±20 km for the altitude of the infrared emission of the nightglow which they later identified as being due to OH. Similar observations by Huruhata (35) during 1949 in Japan yielded heights of approximately 300 km. In 1952 Barbier (36) critically examined all height determinations made up to that time and concluded that the altitude of the emitting layer of OH was approximately 70 km. From observations during the winter of 1952-53, in France, Berthier (37) obtained heights of 140 to 150 km. More recent improved observations by the Van Rhijn Method from aircraft and the ground have yielded heights ranging from 63 to 69 km (20,38). All observations by this technique are subject to the validity of the assumptions involved, in particular to the uniformity of the intensity in the emitting layer. This assumption has been shown by many observers to be invalid since surveys of intensity over the entire sky reduced to the zenith show a very uneven pattern of intensity which fluctuates and moves with time.

Another method of determining the height of the OH nightglow makes use of the variation of the zenith intensity over the sky mentioned above.

This requires simultaneous observations by two stations located some distance apart. Each station makes surveys of the intensity variation over the sky and a comparison is made of the intensity patterns obtained in order to correlate features in the pattern in the portions of the sky observed by both stations. The height of the emitting layer can then be determined by triangulation on a feature seen by both stations. Nakamura (38) has made just such triangulation observations during Dec. 1959 and Jan. 1960 from stations in Japan located 35 km apart. His results showed the height of the OH layer to range between 71 and 76 km.

A third method of height determination is by use of the observed rotational temperatures. If the temperature distribution of the atmosphere with height is known and if the emitting OH molecules have their rotational structure in equilibrium with the ambient temperature, then the rotational temperatures can indicate the altitude of the emitting layer. As has been pointed out before, there are limitations on the altitude where rotational equilibrium is established before emission. Also the observed temperature is not for a single height but averaged over a finite layer over which the intensity of emission and temperature varies. From observations of rotational temperature in Norway, Kvifte (18) deduced an altitude of 70 km. Using observations over a year period, Wallace (24,30) found rotational temperatures which indicated that the altitude of emission was either 70 km or 100 km. Since the rotational temperatures were approximately the same for all vibrational levels observed, Wallace concluded that the emitting layer was below 80 km and hence at 70 km. From observations of the rotational temperature variation over the year and the variation of atmospheric temperature at 75 km over the year as deduced from rocket observations, Wallace concluded that the altitude variation of the emitting layer over the year was small. Krassovsky (39), in a series of observations, obtained a dispersion of rotational temperatures which he believes show a variation in the altitude of emission. He observed high rotational temperatures on some occasions which indicated to him that the emission takes place well away from the mesopause (which is located approximately 85 km). At Zvenigorod Berg and Shefov (34) obtained rotational temperatures which correspond to the temperature at 80 km as determined by rockets.

The best method for determining the height of the OH nightglow layer is by means of rocket observations. In this method photometers are oriented at various angles with respect to the rocket axis and equipped with filters so as to isolate the spectral region containing the emission to be observed. The readings of the photometers are taken as the rocket goes up through the atmosphere. The emitting layer is found at the altitude where the photometer readings drop off most rapidly (depending upon the photometer orientation). Most of the rocket observations have been made at White Sands, New Mexico. One observation has been reported by Russian workers.

The first observation at White Sands was reported by Heppner and Meredith (40) for a flight in July 1956. Their photometer was centered at 6300Å and they observed emission of the 9-3 OH band at 6252Å. The photometer gave readings from 56 km to 163 km (peak rocket altitude) and from these they deduced that almost all the OH emission originated below 100 km and that the lower boundary of the emitting layer was below 56 km. The 6300Å line of OI did not interfere with their measurements since it originates above 150 km. Tousey (41) later summarized the results of all rocket observations through 28 March 1957 taken at White Sands. The results are shown in Table I. In interpreting these altitudes, it must be remembered that in the 5200-5400Å region the OH bands are weak. From the 5410Å results it appears that a continuum emission exists in the region from 85 to 110 km and this was seen in the other observations instead of the OH emission. Krassovsky (42) later

TABLE I

Summary of Rocket Observations of the Height of the OH Layer at White Sands, New Mexico

Date	Max Rocket Alt	Filter	½Width	Nearest OH	Alt	Results
17 Nov 55 2 AM	105	5200Å	150Å	5203Å(9-2)	85-97km	Max at 95 km
12 Dec 55 10 PM	141	5320Å	30Å	5277Å(6-0)	80-110km	Max near 100 km
5 Jul 56 1 AM	163	5335Å	59Å	5277Å(6-0)	90-110Km	Max at 104 km
		6300Å	72Å	6258Å(9-3)	Below 100 km	Lower edge below 56km
28 Mar 57 10 PM	146	5230Å	22Å	5203Å(9-2)	85-110km	Max near 100 km
		5410Å	22Å	None	90-110km	Max near 100 km
		6300Å	25Å	6258Å(9-3)	80-110km	

used the 100 km heights deduced from these rocket observations to support a particular mode of excitation of the OH molecule but Bates and Moiseiwitsch (43), Packer (44) and also Dalgarno (45) disagreed with this interpretation claiming that a weak continuum was observed rather than the OH bands in that region. The latter interpretation appears to be the correct one.

A more recent rocket observation has been reported by Packer (44). This was also taken at White Sands on 6 Nov. 1959. One rocket photometer was equipped with a filter centered at 7280Å and a half width of 180Å and another photometer had a broad filter which was centered at 8800Å and observed the emission between 7400Å and 10,200Å. The first photometer observed the 8-3 OH band at 7275Å and the second included the entire $\Delta v = 4$ sequence plus 9-4 and 3-0 bands. The 8-3 band had 10% of the total emission between 52 and 72 km, 70% between 72 and 95 km, and 20% above 90 km. The peak was found to be at 83 km. The radiation above 95 km was apparently due to a continuum. The photometer with the broad filter gave no data above 91 km but it appeared that no more than 20% of the emission was above 95 km and most of this could be due to a continuum and the (0-1) atmospheric O_2 band at 8645Å. The maximum emission appeared to be near 90 km. No attempt was made to subtract out the continuum from these observations and doing so would only lower the OH altitudes slightly. Packer's final conclusion was that the OH layer was mainly between 70 and 95 km with a maximum between 85 and 90 km (44).

The most recently reported rocket observation was made by Tarasova (46) in Russia on 23 Sept. 1960. There were two photometers, one equipped with a filter centered at 8640Å and a 280Å half width, and the other had a filter admitting radiation in the 9100Å to 10,700Å range. The 8640Å photometer detected no emission below 74 km or above 130 km. The emitting layer had a sharp lower boundary at 74 km, and a very rapid rise of intensity to a maximum at 80 km, and an upper boundary between 100 and 120 km. The center of gravity of the layer was at 81 ±2 km. It was estimated that 50% of the emission was due to the (0-1) atmospheric band of O_2 and the rest due to the 7-3 OH band. The other photometer observed almost exclusively OH emission and showed a localized layer with a center of gravity at 78 ±2 km.

In summary, it seems that the rocket observations of Packer (44) and Tarasova (46) have most nearly isolated that which is being investigated, the intensity distribution with altitude of the OH nightglow. Packer's results for 7280Å of the 8-3 OH band seem to be particularly valid since the filter half-width is small and there are no other strong emissions near that wavelength. Packer's conclusion that the main OH emission originates between 85 and 90 km seems somewhat unjustified. The location of the rocket shot reported by Tarasova was not given. Also it is not known exactly if or how the altitude of emission varies with location, time of year, or time of day. Krassovsky's interpretation of rotational temperature variation from night to night as being due to variations in the altitude of emission (39) must also be taken into consideration. This temperature variation could also be due to changes of temperature at fixed altitudes in the emitting layer, unless the variations are so large and so rapid that heating or cooling of the layer by normal meteorological processes is excluded. Based upon the above data and considerations it is concluded that the peak of the OH emission originates between 80 and 85 km. Additional rocket observations with simultaneous readings from several narrow spectral intervals would be welcome. Rocket observations from further in the infrared of the $\Delta v = 2$ or 1 bands might be useful due to the high intensity of these bands but this must await instrument improvements.

D. Variations of the OH Emission

1. Intensity

The variation of the OH nightglow intensity over the sky and the attempts at triangulation on various distinctive features in order to determine the altitude of the emitting layer have already been mentioned. Nakamura (27) has described cases where the OH intensity pattern changed, grew, and spread throughout the night. His simultaneous observations at two stations show patterns which are not similar but which have a good correlation of diurnal variations of the mean intensity found by averaging over the entire sky. Generally all sky surveys show a patchy, irregular emission pattern.

The intensity variations have been reported by several authors. There appears to be a variation of intensity throughout the night, from night to night, and a variation of the diurnal pattern over the year. Observations on the variation during the night have been principally reported for stations located in Russia and France. Early observations of the intensity variation during the night have been made by Berthier (47) for observations at Haute Provence in France. From observations made in Jan. 1953 Berthier found a general decrease after sunset which lasted for three to four hours. This was followed by a steady rise to substantially higher intensities just before dawn. A more complete report was made by Berthier (48) in 1955 based upon almost two years of observations. He observed a general decrease from three to four and a half hours after sunset, then constant intensity till midnight, followed by an increase till just before dawn, and then a slight drop of intensity. The amplitude of the variation ranged from 1.6 to 2.9 averaging 2.05. The dawn intensities were higher than early evening intensities. In summer the minimum at midnight was much sharper with no period of constant intensity. Strangely enough Berthier (49) later reported that the variation during the night was random. Armstrong (50) also observed differing types of variation during the night which appeared to him to have no physical significance (however, his observations were contaminated with O_2 radiation). More recent results at Haute Provence by M. Dufay (51) disclose almost the same type of variation originally reported by Berthier (48,52).

In Russia, Fishkova and Markova (53) report a nightly intensity minimum near local midnight and a maximum two to four hours after midnight based upon observations over a year at Abastumani (42°N). For observations taken during the winter of 1956-57 at Bzurakan (40°N) Fedorova (16) finds the OH intensity remains fairly constant throughout the night varying by less than a factor of 1.5. In Japan Nakamura (27) found what seemed to be a monthly variation in the diurnal intensity variation.

The daily curves showed a better similarity in winter than in summer. The general winter pattern was an intensity decrease to midnight and then an increase to greater intensity at dawn. In summer an overall decrease or nearly constant intensity were most frequently observed. Nakamura thought that there seemed to be both a diurnal and a semi-diurnal component in the intensity variation. Nakamura's results were derived from observations taken over a period exceeding two years.

There is a fairly good agreement by most observers on the pattern of the annual variation of OH intensity. Berthier (52) finds an intensity minimum in midsummer (July and August), followed by an abrupt increase to a maximum near the beginning of November. From the maximum there is a slow decrease during the winter which becomes more rapid in March and finally reaches the summer minimum. Berthier reported that the minimum intensities were 20% less than the maximum. Berthier's observations were taken over a two-year period at Haute Provence in France. Later reports of the annual OH intensity variation at Haute Provence by M. Dufay (51) and Berthier (49) have confirmed Berthier's original observations. A somewhat discordant report was made by Barbier (54) based on observations made on 118 nights over a two-year period at Haute Provence. Barbier interpreted his data as showing a maximum at the autumnal equinox and a minimum at the spring equinox. A good portion of this seems to be a desire on Barbier's part for a symmetrical variation. Barbier (55) later reported observations at Haute Provence for the period of June through Sept. in 1957 which show a definite minimum intensity in late August. Observations over this same period at Tamanrasset (23°N) showed almost constant intensity (55). Observations in Russia at 41°N, 56°N and 62°N by Fishkova and Markova, Berg and Shefov, and Yarin respectively all agree on an early winter or late fall maximum and a minimum in late summer (August).

Most observers have remarked on a large variability of the OH intensity, not only the diurnal and annual patterns but also from night to night and month to month. Fishkova and Markova (53) found that their absolute intensities varied by a factor of four. Harrison and Jones (14) found a large variation of intensity at 1.68μ throughout the year. Yarin (56) reported that the nightly and night-to-night variations far exceeded his instrumental errors. Berg and Shefov (34) reported that the OH intensity varied most during the winter. From observations at Saskatoon, Shemansky and Jones (20) found much lower absolute intensities of certain infrared bands than reported by Harrison and Jones (14) from earlier observations at the same locations.

Latitudinal variations of OH intensity are difficult to determine due to the inaccuracies in absolute brightness calibrations of the instruments used. Nevertheless, there are some definite indications of a latitudinal variation. A comparison of observations made during the summer of 1957 at Tamanrasset (23°N) and Haute Provence (44°N) was made by Barbier (55) who found the intensities at Haute Provence much higher on the average. The intensity showed little variation at Tamanrasset. Shefov (22) found the mean intensity at Zvenigorod (56°N) was nearly the same as the minimum intensities observed at Yakutsk (62°N). There seems to be an overall increase of intensity with latitude. It also appears that the intensity variation is greater in midlatitudes than at low latitude. Nothing can yet be definitely said concerning the high latitude variation. There is also some evidence for longitudinal variations. Fedorova (16) found little variation during the night at 44°E (less than a factor of 1.5) while Berthier (52) reported an average nightly amplitude factor of 2.05 with some cases ranging up to 2.9 for observations at 0°E. Care must be used here since the two series of observations were made more than two years apart.

In summary, it can be said that the OH nightglow intensity exhibits great variability, not only in time but apparently in space as well. Over a single station the intensity is not uniform over the sky and also varies with time. The diurnal variation generally decreases after dusk to midnight and then rises to a maximum several hours past midnight. This variation is weak or non-existant in summer but well defined in winter. The annual intensity variation has a maximum in late fall or early winter, relatively high values throughout the winter and a minimum in the late summer. The intensity exhibits great night to night variability, particularly in middle and high latitudes. The intensity also seems to be generally stronger at higher latitudes.

2. Temperature

Rapid changes in the OH rotational temperature have been observed at most locations. Krassovsky (31) and Yarin (56) have noted large night to night variations and variations during a single night. It is almost impossible to observe a dirunal temperature variation with present-day instruments. The OH emissions in the wavelength region where photographic plates are sensitive is quite weak and hence long exposures are necessary in order to determine the band structure with the necessary resolution and accuracy. Kvifte (57) feels that the minimum instrument resolving power for rotational temperature work should be at most 50Å/mm. Observations of temperature from measurements in the infrared are not useful due to the poor resolving power of the instruments. During long winter nights Krassovsky (26) exposed two plates for half a night each and found temperature differences greater than 10°K. Since each of these exposures gives the average temperature for half the night, it is fairly certain that the true diurnal variation is much more than that observed by Krassovsky. Hunten (32) attributes the wide temperature variation to either a temperature change at fixed altitudes or variation in the altitude of the emitting layer in a region with a strong temperature gradient. Krassovsky (26) favors the latter mechanism.

Instrument limitations do not prevent the detection of an annual rotational temperature variation. In some cases, particularly during short summer nights, it is necessary to expose a plate over several nights in order to achieve a readable OH spectrum. Studies of rotational temperature over an entire year have been made by Wallace (30) at Yerkes

Observatory, 42.6°N, and by Nguyen-Huu-Doan (58) at the Observatory at Haute Provence, 43.9°N. Their results are shown along with others in Table II. From these two series of observations it is seen that the maximum temperature occurs in the winter (about February) and the minimum in the summer (July). The amplitude of the variation for these two stations is approximately 30°K. This general pattern of variation is supported by many other reports covering only part of the year. Berg and Shefov (34) found a maximum in December and January and a minimum in summer from observations at Zvenigorod (55.7°N) taken during a period of a year but lacking data for Sept. and Oct. Projudina's observations (61) at Zenigorod show high winter values with a sudden dropoff in April. At Yakutsk, 62.1°N, Yarin (31,56) found a midwinter maximum with a tendency for a rapid rise in Nov. and a rapid fall at the end of March (there were no summer observations since there is little or no true night at high latitudes). At Zsenigorod, 55.7°N there seemed to be no clear relation in the temperature variation (31,61). Observations at Saskatoon (12,20,29), 52.1°N, taken during several different years also show a definite tendency toward a winter maximum and a summer minimum.

It should be noted here that many of the earlier rotational temperatures were incorrect due to an error in the energy relation used. This was discovered in 1959 and a correction has been computed and applied to all incorrect temperatures (57,62,63). This correction has been applied to all rotational temperatures used in this paper.

A plot of rotational temperatures versus latitudes shows a definite latitudinal effect. A summary of all known average temperature

TABLE II

Summary of OH Rotational Temperature[*] Measurements Listed by
Month and Station

Author	Latitude	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Average
Blackwell et al (15)	16.3°S	-	294	-	-	-	-	-	-	-	-	-	-	294
Fedorova (16)	40.4°N	212	219	-	-	-	-	-	-	-	-	-	215	216
Wallace (30)	42.6°N	228	239	223	191	225	212	195	213	213	216	-	223	216
Doan (58)	43.9°N	228	230	226	226	201	205	197	-	-	235	228	222	220
McPherson and Jones (29) Gush and Jones (12) Shemansky and Jones (20)	52.1°N	248	239	239	210	199	-	-	-	-	200	248	-	226
Shefov (60) Prokudina (61)	55.7°N	227	230	226	219	-	-	-	-	-	-	-	228	226
McPherson and Jones (29)	58.8°N	-	-	227	-	-	-	-	-	-	-	-	-	227
Kvifte (57)	59.7°N	215	217	225	-	-	-	-	-	-	-	-	-	218
McPherson and Jones (29)	74.7°N	-	242	284	-	-	-	-	-	-	-	-	-	263
Chamberlain and Oliver (59)	76.6°N	-	-	-	-	-	-	-	-	-	-	300	-	300

[*] All temperatures are given in °K

determinations is shown in Table III. They are listed in order of increasing latitude without reference to longitude or period of observation. In Table IV the temperatures are broken down into the season of the observations and averages are given for all determinations at that season and latitude regardless of the year of observation. Several observers did not give the date or season of their observations and hence their temperatures could not be included in this table. From the winter data in Table IV it is readily apparent that in middle latitudes (40°- 55°) there is little, if any, variation in the rotational temperature. During the winter at about 60°, the rotational temperatures begin to increase rapidly with maximum values found at the poleward limit of observations (78.8°N). McPherson and Jones (29) attribute this latitude effect to a decrease in the height of emission with increasing latitude. This temperature correlation is much higher with geographic latitude than with geomagnetic latitude. Less complete tables similar to Table III have been published by Wallace (63) and Kvifte (57) but some of the Russian observations listed by them are in error as to latitude. Kvifte and Wallace have mistakenly used the geomagnetic latitude given in the original Russian articles for geographic latitude.

There also seems to be a latitudinal effect in the amplitude of the annual variation. Wallace (30) believes that the amplitude of the annual variation increases with latitude. A comparison of the observations at 42.6°N and 52.1°N seem to bear this out. It is unfortunate that the long days and twilight during the summer months at higher

TABLE III

Averate OH Rotational Temperature Determinations

Author	Latitude	1	2	3	4
Blackwell et al 1960 (15)	16.3°S	-	Aug 1958	294	294
Fedorova 1959 (16)	40.4°N	N45°	Dec56-Feb57	233±16	216±15
Meinel 1950 (8)	42.6°N	N	Dec49-Jan50	260±5	240±5
Wallace 1961 (30)	42.6°N	-	Dec58-Jan60	216	216
Cabannes et al 1950 (64)	43.9°N	NE75°	Aug49-Jan50	200±50	185±46
Dufay + Dufay 1951 (65)	43.9°N	NE	-	262±7	242±7
M. Dufay 1959 (51)	43.9°N	NE75°	Sep57-Nov57	250	230
Connes + Gush 1959 (66)	43.9°N	NE70°	Mar 1959	270	250
Connes + Gust 1960 (67)	43.9°N	NE70°	Mar 1959	240	220
Nguyen-Huu-Doan 1963 (58)	43.9°N	-	1961	220	220
M. Dufay + Bilquey 1963 (68)	43.9°N	-	1961	230±10	230±10
Gaynullina (62)	38.7°N	N60°	Dec57-Jan58	257	237
Gush + Jones 1955 (12)	52.1°N	N75°	May-Jun 54	200±20	200±20
McPherson+Jones 1960(29)	52.1°N	-	Feb-Mar 57 Feb-May 58	234±29	216±27
Shemansky+Jones 1961(20)	52.1°N	-	Oct-Nov 59	213	213
Prokudina 1959 (61)	55.7°N	N60°	Jan-Apr 57	240±20	222±19
Shefov 1962 (60)	55.7°N	-	Jan 58 Dec59-Apr60	240	222
McPherson+Jones 1960(29)	58.8°N	-	Mar 58	246	227
Kvifte 1959 (18)	59.7°N	SW30°	Jan58-Mar58	215±2	215±2
Kvifte 1961 (57,69)	59.7°N	SW30°	Jan58-Mar58 Feb59-Mar59	222	222
Shuyskaya 1959 (16)	60.2°N	S20° N70°	Winter	265±14 275	244±14 254
Yarin (62)	62.1°N	S60° N60°	Nov-Dec 59 Oct59-Apr60	250 265	250 265
M. Dufay+Bilquey 1963(68)	66.7°S	-	-	370±25	370±25 *
Prokudina 1959 (61)	68.3°N	Zenith N75°	Dec 58 Feb 57	280±20 330±20	258±20 304±19
McPherson+Jones 1960(29)	74.7°	-	Feb-Mar 57	296±38	273±35
Chamberlain+Oliver 1953(59)	76.6	-	Nov 52	325±25	300±33

* Defective photometer

1. Direction and Zenith Angle of Observation
2. Date of Observations
3. Reported Temperature °K
4. Corrected Temperature °K

TABLE IV

Average OH Rotational Temperatures

Latitude	Winter	Spring	Summer	Fall
39.7°N	237	-	-	-
40.4°N	216±15	-	-	-
42.5°N	226	214	208	215
43.9°N	226	218	201	231
52.1°N	247	211	-	213
55.7°N	222	226	-	-
58.8°N	-	227	-	-
59.7°N	222	-	-	-
60.2°N (Obs to south)	244	-	-	-
60.2°N (Obs to north)	254	-	-	-
62.1°N (Obs to south)	250	-	-	-
62.1°N (Obs to north)	265	-	-	-
63.3°N	300	-	-	-
74.7°N	273±35	-	-	-
76.6°N	-	-	-	300

latitudes prevent the accurate observations which are needed to confirm this variation.

In summary the OH rotational temperature exhibits great variability from night to night and also some, as yet, undetermined variation during the night. An annual maximum is found in midwinter and a minimum in midsummer. The amplitude of this annual variation seems to increase with latitude. In general, the rotational temperature in the winter is essentially constant in middle latitude, but poleward of about 55° it increases rapidly with latitude. The reason for this variation is not clear. As rocket observations have shown, the temperature at 75 km increases by about 40°K between 33°N and 59°N and also varies with time (70). But, whether it is the temperature variation at fixed altitudes, or altitude variations in a region of strong vertical temperature gradient which causes the observed temperature variations cannot be determined. Both are probably responsible in part for the observed behavior. It is also possible that longitudinal variations exist but verification of this would require observations over the same time period by several stations with widely differing longitude and approximately the same latitude. Such observations have not yet been made.

E. Correlations

The correlations of OH nightglow with other night sky emissions or geophysical phenomena are very few. Probably the best correlation is with the sodium emission. Berthier (71) made observations of OH and Na nightglow intensity during six nights in Sept. and Nov. of 1952. He found that there was a very definite correlation between the two emissions. The variations of the two emissions were nearly parallel throughout the night on all six nights. Barbier (54) and Berthier (52) later noted that the seasonal variation of the two emissions were very similar and that they also seemed to be correlated. Several other workers at various locations have remarked on the strong diurnal and seasonal covariance of these two emissions (1,24,30,72). Packer (44) reported a good correlation between the H and Na variations of intensity over a five-year period.

Ballif and Venkateswaran (73) have examined the correlation of OH and Na emissions from the viewpoint of the reaction mechanisms. They show that the Na intensity variation has a temporal component of variation dependent upon ozone concentration in the emitting layer. They believe that the reaction mechanism for the OH emission also depends upon the ozone concentration. Since the two emissions originate from nearly the same altitude region, they should then be correlated since they both depend upon the ozone concentration at their common emission altitudes.

Krassovsky and some of his coworkers (31) find a strong correlation of OH intensity with that of the atmospheric H_α line at 6563Å. This is to be expected since they both depend upon the hydrogen concentration. As yet no one else has observed this correlation.

Apparently there is no correlation of OH intensity with the atomic oxygen green line at 5577Å. Fedorova (16) found no diurnal correlation from his study of data taken over a year. However, Fishkova and Markova (53) find that the near midnight minimum of intensity for OH coincides with the 5577Å maximum.

There seems to be no correlation between the OH intensity and geophysical phenomena. Kvifte (18) reported that the OH does not seem to be enhanced during auroral displays. Berg and Shefov (34) could find no dependence of intensity or rotational temperature upon magnetic activity for a year's observations at Zvenigorod. Finally, Krassovsky (17) found the OH emission to be independent of solar activity.

F. Reaction Mechanism

Since the identification of the OH nightglow a great amount of work has been done in trying to establish the excitation mechanism for the OH molecule. This work has been accompanied by the publication of an equivalent amount of literature arguing the relative merits of the reaction mechanisms proposed. To date, there are two modes of excitation which appear to be feasible. One mechanism, suggested by Bates and Nicolet (25), involves a reaction between ozone and atomic hydrogen in order to produce vibrationally excited OH molecules. The other mechanism depends upon a reaction between vibrationally excited oxygen molecules and atomic hydrogen. Krassovsky is the proponent of this latter mechanism and has published numerous articles expounding his viewpoint. The main features of each excitation mechanism will first be presented separately and then both will be discussed as to their ability to explain the observed emission features.

1. Ozone Mechanism

The ozone-hydrogen mechanism was first proposed by Bates and Nicolet (25,28) in 1950 shortly after Meinel announced the identification of the OH bands in the nightglow. The source of atomic hydrogen is mainly from water vapor which diffuses upward through the tropopause into the mesosphere where it is broken down by reactions 14 and 29 (all numbers refer to the reactions listed in Table V). Upward diffusion of methane and molecular hydrogen also contribute to the atomic hydrogen concentration since they are not affected by the low temperatures in the stratosphere which greatly lowers the dew point of the water vapor. The molecular hydrogen is broken down directly by the action of sunlight and also by reaction 20 while the methane undergoes a series of dissociation reactions which begins with 15. Reaction 20 is almost complete well

TABLE V

Principal Reactions Associated with the OH Nightglow

1. $O_2 + h\nu_{HERZ} \longrightarrow 2O^*$
2. $O_2 + h\nu_{SHUM-RUN} \longrightarrow O^* + O^{*'}$ (at 100 km)
3. $O^* \longrightarrow O + h\nu$
4. $O_2 + O + X \longrightarrow O_3^* + X$
5. $O + O_3^* \longrightarrow O_2^* + O_2^*$
6. $O + O + X \longrightarrow O_2^* + X^*$
7. $O_2^{**} + O_2 \longrightarrow O_3^* + O$
8. $O_2^* + X \longrightarrow O_2 + X^*$
9. $O_3^* + X \longrightarrow O_3 + X^*$
10. $O_3^* + h\nu(\lambda > 1025\text{Å}) \longrightarrow O_2 + O^*$
11. $O_3^* + O_3^* \longrightarrow O_2^* + O_2^* \; O_2^*$
12. $O + e \longrightarrow O^- + h\nu$
13. $O^- + O \longrightarrow O_2^* + e$
14. $H_2O + h\nu \longrightarrow OH + H \quad \begin{cases} \lambda < 1300\text{Å} \text{ at 70 km} \\ \lambda < 1800\text{Å} \text{ at 100 km} \end{cases}$
15. $CH_4 + h\nu \longrightarrow CH_3 + H$ (followed by other breakdown reactions)
16. $H + H + X \longrightarrow H_2 + X$
17. $H + O_3 \longrightarrow OH^*(v \leq 9) + O_2^*$
18. $H + O_2^*(v \leq 27) \longrightarrow OH^*(v \leq 9) + O$
19. $O + H + X \longrightarrow OH^* + X$
20. $H_2 + O \longrightarrow OH + H$

(TABLE V continued)

21. $HO_2 + O \longrightarrow OH + O_2$

22. $HO_2 + h\nu \longrightarrow OH + O$

23. $O + H \longrightarrow OH^* + h\nu$

24. $OH^* \longrightarrow OH + h\nu$

25. $OH^* + X \longrightarrow OH + X^*$

26. $OH^*(v \geqq 10) + O_2 \longrightarrow O_3 + H$

27. $OH^{**} + O_2 \longrightarrow O + HO_2$

28. $OH^* + O_2 \longrightarrow OH^{*\prime} + O_2^*$

29. $OH + O \longrightarrow H + O_2^*$

30. $OH + H \longrightarrow O + H_2^*$

31. $H + O_3 \longrightarrow HO_2 + O$

32. $H + O_2 + X \longrightarrow HO_2 + X$

33. $H + HO_2 \longrightarrow H_2O + O$

34. $O_2^{**}(v > 27) + N_2 \longrightarrow N_2O^* + O$

35. $N_2O^* + O \longrightarrow N_2^* + O_2^*$

36. $OH^* + N_2 \longrightarrow N_2O + H$

37. $O_2^* + OH \longrightarrow OH^* + O_2$

38. $O_2^* + HO_2 \longrightarrow OH^* + O_3$

39. $O_3 + HO_2 \longrightarrow OH^* + O_2 + O_2$

below 100 km and 16 proceeds very slowly so that all hydrogen exists either in the atomic state or in compounds with oxygen (but not as water). The compound HO_2 produced by 27, 31 and 32 may have appreciable concentration in the lower part of the dissociation region. The atomic hydrogen concentration at 100 km is estimated to be about 10^8 to 10^9 atoms/cm^3.

Bates and Nicolet ruled out OH^* production by 19 and 23 as being much too slow to be effective in producing even part of the observed excitation rate. They finally settled upon 17 as being the most probable excitation mechanism followed by 24 and then 29 which serves to recycle the hydrogen. Reaction 17 is most effective at lower altitudes (about 65 km) and is fast enough to produce the observed excitation rate. Further support for this mechanism comes from the fact that 17 produces energy which can only excite OH vibrational levels of nine or less and is energetically incapable of exciting the tenth vibrational level which is not observed in the nightglow. Chamberlain (1) points out that this mechanism can be valid only if the OH emission layer is at about 65 km since the maximum emission rate, which occurs at 65 km, is required in order to produce the observed high excitation rate.

2. Excited Molecular Oxygen Mechanism.

In 1951 Krassovsky suggested that reactions 18, 24, 25, 29 and 30 were responsible for producing the OH nightglow. The first step in the production of O_2^* used in 18 is through the dissociation of O_2 by 1 and 2 into O^* which decays by 3. The resulting atomic oxygen then forms O_2^* by means of 4, 5 and 6. Krassovsky believes 4 and 5 to be more effective than 6 in regions where the concentration of O_2 is high. Reaction 5 is very fast and the O_3^* produced by 4 is shortlived and in low concentration at 100 km. Below 100 km the small concentration of O_3, O and H causes O_3^* to deactivate by 9 (23). A limit is placed on the excitation level of O_2^* by 7, 34 and 35 which removes O_2^* ($v > 27$). This explains, in part, why $OH^*(v > 9)$ is not observed. Deactivation of O_2^* by 8 is not believed to be highly probable by Krassovsky. Krassovsky (23) also includes 10 and 11 as effective means of keeping the O_3 concentration low at 100 km.

To further explain the absence of $OH^*(v \geq 10)$ Krassovsky (23) proposes that 26 and 36 are very effective in the de-excitation of OH^* ($v \geq 10$). These two reactions coupled with 7, 34 and 35 then serve to completely quench and eliminate any radiation from OH vibrational levels higher than the ninth. At 100 km the concentration of O is sufficient to maintain constant concentration of O_2^* and hence OH intensity throughout the night.

3. Discussion.

An argument as to which of the two above mechanisms is correct has been going on in the literature since 1955. Krassovsky (23) argued that 17 could excite the tenth vibrational level but this was completely disproved by Bates and Moiseiwitsch (74).

Krassovsky (23) further believed that 17 did not have a sufficient rate of reaction to explain the OH intensity. He attributed the low rate of reaction to the rapid disappearance of O through 4, 5 and 6. According to his calculations, 4 could cause a decrease to one-half the original concentration in 0.5 sec, 5 sec, and 10 min after sunset at 60, 70 and 80 km respectively. Reactions 5 and 6 serve to decrease the O concentration even more rapidly.

There would then be no means of recycling H since 29 and 30 would then be too slow. At 100 km the O concentration will remain sufficiently high throughout the night but 17 would be entirely too slow due to low O_3 concentration at this altitude. Bates and Moiseiwitsch (74) countered this argument by pointing out that Krassovsky had used too high a rate coefficient ($10^{-33}T^{1/2}$ cm^6/sec) for 4. Their rate coefficient ($5 \times 10^{-36}T^{1/2}cm^6/sec$), makes the atomic oxygen decay very slowly at 75 km (where the concentration of $x \approx 10^{15}/cm^3$ and concentration of $O_2 = 2x\ 10^{14}/cm^3$). At 75 km with the rate coefficient of Bates and Moiseiwitsch the time for the atomic oxygen concentration to decrease to one half of the initial concentration is many hours. They further stated that the photodissociation rate of O_2 at 75 km was sufficient to produce the necessary atomic oxygen needed in 29. The argument as to the correct rate coefficient continued (42,43). The slower rate coefficient proposed by Bates and Moiseiwitsch appears to be more reasonable and its use would completely nullify Krassovsky's argument.

The rocket measurements of the altitude of the OH layer made by Tousey (41) were used by Krassovsky (42) in support of his O_2^* mechanism since they showed that the OH emitting layer was apparently near 100 km. This is the altitude where Krassovsky's mechanism is expected to be most efficient and at this altitude the O_3 mechanism is unable to support the OH emission. Dalgarno (45) and Bates and Moiseiwitsch (43) argued against this claim on the grounds that although the 6-0 and 9-2 bands of OH were present in the spectral region observed by the rocket measurements, they are extremely weak and their emission was overcome by that of a nightglow continuum at 100 km which is actually what was measured by the rocket observations. The later, more accurate rocket observations of Packer (44) and Tarasova (46) seem to support, if anything, the O_3 mechanism due to the lower altitudes found for the OH nightglow layer.

Further points in the argument were made by Krassovsky (42) who stated that OH^* was rapidly deactivated by collisions at 75 km and hence couldn't emit at such a low altitude. Rocket observations seem to prove this statement as being incorrect. Bates and Moiseiwitsch (74) felt that O_2^* was rapidly removed by collision and hence wouldn't be available in sufficient concentrations to support the OH emission through 18. Very recently Dalgarno (75) has pointed out that the high efficiency of atom-atom interchange of O_2^* with O drastically reduces any effect that O_2^* might have upon the chemistry of the nocturnal atmosphere.

Kvifte (21) argues that if the O_2^* hypothesis is correct then O_2^* ($v = 26,27$) has two possibilities of exciting OH to the ninth vibrational level whereas three possibilities exist for O_2^* ($v = 23,24,25$) to excite the eighth vibrational level of OH. It follows that more $OH^*(v = 8)$ should be observed than $OH^*(v = 9)$. But, if anything, the reverse seems to be true, thus favoring the O_3 mechanism (21). Along this same line, it has been pointed out before that there is no reason for either 17 or 18 to preferentially excite $OH^*(v = 9)$. It seems that if O_2^* ($v \leq 27$) is the excitation mechanism then there would be a much greater tendency to excite the lower vibrational levels of OH since almost every level of O_2^* can excite $OH^*(v = 1)$ but only two levels of O_2^* ($v = 26,27$) can excite $OH^*(v = 9)$ and these two levels of O_2^* do not necessarily excite just that level. Observations seem to indicate that all levels of OH^* are excited approximately equally and this would support the O_3 mechanism.

Additional arguments in favor of the O_2^* mechanism were presented by Krassovsky (17) in 1958. He felt that lower altitudes (70 to 80 km) would favor the O_3 process and higher altitudes (90 to 100 km) would favor the O_2^* process. The OH intensity produced by the O_3 process should vary with H concentration which Krassovsky felt would vary with solar activity at 75 km. Since there is no observed correlation of OH intensity with solar activity, the higher altitude O_2^* process would be favored because at 100 km there should be no variation in the atomic hydrogen concentration with solar activity as H_2O and H_2 are completely dissociated at that altitude.

Observations by Krassovsky (17) showed the fourth and fifth vibrational levels of OH were somewhat preferentially excited. Krassovsky believed this supported the O_2^* reaction. He reasoned that 4 and 5 had enough energy to produce two equally excited O_2^* molecules each of which could excite OH^* (v = 4) thus giving the observed preferential excitation. As pointed out before, the O_2^* molecules do not necessarily transfer all their energy into the resulting OH^* of 18 and hence OH^* (v = 4) is not necessarily preferentially excited by the O_2^* resulting from 5 as Krassovsky predicts. Also 9, 10 and 11 must be considered in that they can remove or alter the O_3^* entering into 5 and thus change the resulting excitation level of the O_2^* produced by that reaction.

In a later work, Krassovsky (76) discussed the O_3 and O_2^* process for low H concentrations ($5 \times 10^6/cm^3$) at 100 km and below. He finds that 17 and 18 would then require unacceptably high concentrations of O_3 and O_2^* and proposes several new processes for the formation of OH^* which do not involve atomic hydrogen. They are listed in Table V as reactions 37, 38 and 39. Reaction 37 would be favored at lower altitudes where the O concentration is low so that 29 is ineffective and OH can remain free. Krassovsky predicts that 32 is 10^3 more effective than 4 and hence 38 and 39 may make significant contributions to the OH nightglow. Each reaction would excite its own relative population of OH^*, proceed at different rates, and have different altitudes of maximum excitation.

Observations of the OH nightglow made by Krassovsky and his co-workers during the IGY have shown some peculiar features (22, 31,56). One was that, at Yakutsk, for high intensity and rotational temperatures greater than 250°K there was an increase of intensity with rotational temperature. Such was not the case at Zvenigorod or at Yakutsk when the rotational temperature was below 250°K. Krassovsky (31,39) and Yarin (56) attribute this effect to 17 which is temperature dependent and which will cause an increase of intensity at higher ambient temperatures. At temperatures above 250°K at Yakutsk, 17 was dominant while, at Zvenigorod and at low rotational temperatures at Yakutsk, 18 (which is not temperature dependent) was dominant. Shefov (22) at Zvenigorod and also Yarin (56) at Yakutsk report a tendency for the increase of the relative population of the higher vibrational levels of OH^* with increasing rotational temperature. Krassovsky (39) says this shows that both 17 and 18 are taking place simultaneously and that this effect is being produced by 17, becoming more rapid at higher temperatures. Shefov (22) agrees and further states that this indicates that 18 is predominant at Zvenigorod.

Krassovsky's latest argument against the O_3 process (26) is rather complicated. He invokes reactions 25, 26, 27 and 36 which he believes will allow only non-equilibrium emission (giving high rotational temperatures) and only emission above 80 km, thus eliminating the O_3 process. Krassovsky states that the rotational temperature does not reflect the ambient temperature and that this can only be possible if OH^* is deactivated by collisions or disappears through chemical reactions after a small number of collisions in which equilibrium is not established between the rotational and ambient temperatures. Thus, the only emitters won't be in equilibrium and the others will react chemically before equilibrium emission can be attained. Above 80 km, Krassovsky says that the particle concentration is too small to attain sufficient collisions for equilibrium to occur, especially for $OH^*(v \geq 6)$. Reaction 26 will only operate with $OH^*(v \geq 10)$, but 27 will proceed for $OH^*(v \geq 6)$ and 36 will proceed for $OH^*(v \geq 8)$ and thus cause considerable extinction for $OH^*(v \geq 6)$. Reaction 25 will proceed for all OH^* and thus give non-equilibrium rotational temperatures, increase the relative population of lower vibrational levels, and quench all OH^* radiation at lower altitudes. Therefore, the OH emission will arise from altitudes above 80 km. At still higher altitudes, the chemical de-excitation effect of 26 and 27 will become less, thus giving increased rotational temperatures (due to non-equilibrium) and increased intensity of emission (due to less de-excitation). Reactions 25, 26 and 27 explain the temperature intensity dependence observed at Yakutsk and also quench all OH emission from lower attitudes where the O_3 mechanism would be expected to operate. There are several criticisms to this entire argument. It does not explain the observations of many other workers of nearly equal rotational temperatures obtained from several OH^* bands with both high and low initial vibrational levels. It cannot explain the OH emission found below 80 km by rocket observations. Finally, if the rates of reactions 25, 26 and 27 are as large as Krassovsky believes them to be, then, the amount of OH^* de-excited by these reactions will be so large that the O_2^* mechanism will not be able to produce enough excited OH^* so that, after de-excitation sufficient OH^* will be left over to give the observed emission rate. This would be true even if the emission altitude is at 100 km, which is contrary to rocket observations.

Probably the most complete investigation into the problem of which mechanism produces the OH nightglow was made by Wallace (24) in 1962. He examined a model of the hydrogen-oxygen atmosphere in local photochemical equilibrium using the most recent values for the reaction rates of the many processes involved. He then tested the ozone-hydrogen reaction, 17, and found that between 65 and 95 km the O_3 concentration is determined by atomic hydrogen. He calculated OH emission profiles based on 17 and found that it could produce the observed emission rate. Furthermore, since the O_3 concentration is determined by 17, then the emission rate of OH^* produced by 17 would be independent of temperature. However, the emission peak would be at 70 to 75 km (based upon a 12 hour day and a 12 hour night) which is below the altitude deduced from rocket observations. If deactivation of OH^* by 25 or 28 is assumed to supress the emission below 75 km so as to bring the emission peak into line with the rocket observations, then 17 is entirely incapable of producing the observed emission rate. Also, the seasonal and latitudinal variations of the OH emission are not explained by this reaction. During the artic winter, there would be an insufficient amount of atomic oxygen below 85 km to recycle H by 29. Wallace (24) concluded that 17, when considered within the framework of local photochemical equilibrium, fails to represent many of the observed features of the nightglow.

Recently Ballif and Venkateswaran (77) investigated the diurnal variation of the OH nightglow to be expected from reactions 17 and 18 separately. They included the principal chemical reactions and abundance variations in their calculations. They concluded that if the OH nightglow was produced by 17 three possible variations could exist: an increase of intensity throughout the night, a post-twilight increase followed by a decrease, and a decrease throughout the period of darkness. They argued that the post-twilight increase followed by a decrease for the rest of the night seemed most likely. If 18 controlled the OH emission, then the OH intensity variation through the night would be small and the intensity would be nearly steady. Neither of these processes gives the type of diurnal intensity variation most frequently observed and it must be concluded that some other process is operative or that local photochemical equilibrium is inadequate for the description of the diurnal OH intensity variation.

In the final analysis, neither process completely explains the observed features of OH nightglow. It is difficult to make completely accurate calculations due to the large uncertainties in rate coefficients for many of the reactions involved and in the actual concentrations of many of the components of the upper atmosphere between 60 and 110 km. Many of the arguments used by Krassovsky in support of 18 are either weak or proven to be erroneous. But 17, although it seems to be much more adequate, has several drawbacks when considered within the framework of local photochemical equilibrium. It seems very likely that dynamical processes must be considered in the production of the OH nightglow, particularly during the winter at high latitudes. These will be discussed in a later section.

SODIUM

A. Discovery

Prior to its identification in the nightglow, sodium was not expected to be found in any significant quantity in the upper atmosphere. The seeming improbability of sodium being a major component of the night sky radiations probably prevented the complete identification of the sodium nightglow for some time. In 1929 Slipher (78) reported a strong emission in the night sky at 5892Å that appeared to be a permanent part of the nightglow. He remarked that this emission was closely in the position of the sodium D doublet but apparently he did not consider that it might actually be due to sodium.

Final, conclusive identification of sodium in the nightglow was made in 1938 by Bernard (79,80) and by Cabannes, Dufay and Gauzit (81,82,83) independently. It had been shown earlier that the 5893Å radiation was of terrestrial origin due to the intensity increase from zenith to horizon. Bernard also identified the twilight glow at 5893Å as being due to sodium (78,79). These identifications provided conclusive proof for the existence of free sodium in the upper atmosphere (84)

B. Nature of the Sodium Emission

The sodium nightglow and twilightglow consists of two principal lines resulting from electronic transitions of the sodium atom into the ground state. One line, D_1, is located at 5896Å and results from a

$^2P_{\frac{1}{2}} \rightarrow {}^2S_{\frac{1}{2}}$ transition. The other line, D_2, is at 5890Å and results from a $^2P_{3/2} \rightarrow {}^2S_{1/2}$ transition. These two lines make up what is known as the sodium D doublet. The statistical ratio of these two lines (D_2/D_1) from transition probabilities is two to one. The mean life in the upper state ($^2P_{3/2}, {}_{1/2}$) is approximately 10^{-8} sec (85). The sodium D lines appear to be the only ones occuring in the atmospheric emissions. The 3303Å emission due to atomic sodium has been observed in sodium trails ejected from rockets in the twilight (86) but not in the natural night and twilight emissions.

The processes which produce the nightglow and the twilightglow are different. It is fairly well established that the nightglow is caused by chemical excitation through some sort of two-body exchange reaction (1,87,88). While the nightglow requires sodium atoms in the excited state, the twilightglow requires atoms in the ground state. Kooman and his co-workers (87) believe that the reaction which produces sodium atoms in the excited state is probably the same one that maintains free neutral atoms in the ground state. The twilightglow is produced by resonance scattering of sunlight (1,84,88,89,90,91). Resonance scattering was first proposed by Bernard along with his identification of sodium in the twilightglow (80). Vegard and his co-workers originally disagreed with this hypothesis and proposed that the twilightglow was excited through photodissociation by ultraviolet light which was screened by the ozone layer (92,93). In 1950 Bricard and Kastler found that the twilight emission was slightly polarized (90). This is predicted by scattering theory, hence this result favors resonance scattering. From studies of sodium vapor trails ejected from rockets in the twilight, Cooper and his co-workers have firmly established that the twilight emission is due to resonance scattering(86).

The sodium nightglow exhibits a wide variation of intensity ranging from the order of a few Rayleighs* to 400R (94). The average in mid-latitudes appears to be about 150R (85,95) based on over five years of data. The twilightglow is much brighter and has maximum values exceeding 6000R at the zenith (15,96). Nightglow intensity measurements are complicated by the emission of the 8-2 band of OH which overlies the entire region around 5893Å. Fortunately, the intensity of the 8-2 OH band is only about one-sixth of that of the sodium doublet (85).

The neutral, atomic sodium abundance deduced from nightglow and twilight observations also varies widely. Many measurements and estimates of the sodium abundance have been made. The values range from 10^9 to 2.5×10^{10} atoms/cm^2 column (58,91,96,97).

C. Reaction Mechanism

In 1939, shortly after the identification of sodium in the nightglow, Chapman proposed a mechanism for the production of electronically excited sodium atoms which would emit the yellow doublet radiation observed in the nightglow (98). Chapman reasoned that the energy for the sodium D lines came from an atmospheric store of energy that was always present and that that energy could only be dissociated energy. He proposed that

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* 1R = Rayleigh = an apparent emission rate of 10^6 photons/cm^2(column) sec.

the sodium was oxidized through 13 at lower levels and through 14 at upper levels (numbers refer to reactions listed in Table VI). Chapman gave 4 as the reaction which produced sodium in the excited state ($^2P_{3/2, 1/2}$).

In 1950 Bates and Nicolet (99) proposed some additional reactions to excite the sodium nightglow. Their reactions involved atomic hydrogen and are listed as reactions 2,3 and 6 in Table VI. Bates, in 1954, proposed a further reaction involving atomic hydrogen (100). His reaction is listed as 7 in Table VI.

The possibility of producing the sodium nightglow by sodium ion reactions has been discussed by Bates and Dalgarno (101). They considered the occurrence of reactions 8 and 9 in the ionosphere where the concentration of sodium ions is much higher than that of neutral sodium atoms. They found that during the day 19 could possibly recycle the sodium. At night only charge transfer reactions such as 18 are available to recycle the neutral sodium. Bates and Dalgarno found that the reactions of ionized sodium were able to give only a minute fraction of the observed sodium nightglow intensity due to the low sodium concentrations in the ionsphere and the low rate coefficient of 18. It was concluded that ionized sodium reactions were not responsible for the sodium nightglow. Observations of sodium vapor trails ejected from rockets at twilight between 75 and 200 km showed no evidence of ionic recombination (86) thus supporting the above conclusion. Huâten (102) states that the ionization reaction rates are too slow for equilibrium to be reached, even during the day. Chamberlain also finds that

TABLE VI

Principal Chemical Reactions Associated with the Sodium Airglow

✓1. $NaO_2 + O \longrightarrow NaO + O_2$

✓2. $NaO_2 + H \longrightarrow NaO + OH$

✓3. $NaO_2 + H \longrightarrow NaH + O_2$

4. $NaO + O \longrightarrow Na(^2P) + O_2$

5. $NaO + O^* \longrightarrow Na(^2P) + O_2$

✓6. $NaH + O \longrightarrow Na(^2P) + OH$

7. $NaH + H \longrightarrow Na(^2P) + H_2$

8. $Na^+ + O^- \longrightarrow Na(^2P) + O$

9. $Na^+ + e \longrightarrow Na(^2P) + h\nu$

10. $Na + M^* \longrightarrow Na(^2P) + M$

?11. $Na + O_2^*(v > 12) \longrightarrow Na(^2P) + O_2$

12. $Na(^2P) \longrightarrow Na(^2S) + h\nu\,(5893\,\text{Å})$

✓13. $Na + O_3 \longrightarrow NaO + O_2$

14. $Na + O + X \longrightarrow NaO + X$

✓15. $Na + O_2 + X \longrightarrow NaO_2 + X$

16. $Na + O_2^* \longrightarrow NaO + O$

17. $Na + OH^* \longrightarrow NaO + H$

18. $Na + M^+ \longrightarrow Na^+ + M^*\ (M = O \text{ or } N)$

19. $Na + h\nu \longrightarrow Na^+ + e$

20. $Na_2 + NaO \longrightarrow Na_2O^* + Na$

21. $Na_2O^* + Na \longrightarrow Na_2O + Na(^2P)$

22. $Na_2 + NaO \longrightarrow Na_2O + Na(^2P)$

23. $O + O + X \longrightarrow O_2^*(v = 50) + X$

24. $O + O_2 + X \longrightarrow O_3^* + X$

25. $O + O_3^* \longrightarrow O_2^*\ (v = 12 \text{ to } 16) + O_2$

26. $M^* \longrightarrow M + 48.5\ \frac{KCAL}{MOLE}$

recombination of sodium ions at night is not important (1).

Ogawa has reported laboratory studies of reactions between sodium and ozone (103). The ozone-sodium reaction was accompanied by the emission of the D lines only and was found to be much faster than reactions of sodium with molecular oxygen or a mixture of atomic and molecular oxygen. To explain his results, Ogawa, on the basis of the heat of reaction, proposed 12,13,20,21 and 22 to be the exciting mechanism. In the atmosphere the concentration of Na_2 used in 20 and 22 is much too small to furnish the observed sodium nightglow intensity. Hence Ogawa's laboratory mechanism cannot be responsible for the sodium nightglow.

In 1956 Hunten (104) postulated that 13 and 4 was the probable mechanism. He calculated that the ozone concentration increased by a factor of a hundred at sunset in the nightglow layer thus supporting 13 and 4. Koomen (105) believed that both 4 and 6 were responsible for producing the excited sodium but found that both were insufficient above 70 km (87). Omholt (88) ascribed the sodium nightglow to 4 with 6 becoming significant above 75 km. Ogawa (103) and Dalgarno (45) argue against 4 on the basis that it appears to be endothermic. Reactions 6 and 7 are exothermic but the reaction rates are largely unknown (45). Laboratory studies have shown that 7 does not occur (106). Chamberlain (1) suggested that 3, 6 and 15 might explain the nightglow. Dalgarno (45) states that the rate coefficient of 15 is high. This would tend to support the cycle of 3, 6 and 15.

Krassovsky (17) has proposed that below 100 km atomic sodium could be excited by collisions of NaO with oxygen atoms in the metastable excited states $^1S_0$ and $^1D_2$ as shown in 5. This process is exothermic and thus overcomes the principal objections to 4. Krassovsky uses 16 and 17 to recycle the neutral sodium. In this mechanism (5, 16 and 17) sodium would exist in the free state only where the O^* concentration is high. The sodium concentration would vary with the concentrations of O_2^*, OH^*, and O^*. Unfortunately the rate coefficients of these reactions are unknown. Krassovsky did not state the excitation levels needed by OH^* or O_2^* in order to make 16 and 17 exothermic and therefore practical. The excitation levels required of O_2^* and OH^* in 16 and 17 may be so high that the concentrations of the O_2^* and OH^* possessing the necessary energy are too low to produce the recycle rate needed to give the observed intensity.

Potter and DelDuca (106) and later Okuda (107) have postulated an excitation mechanism of the type 10 where the M^* must meet the requirement given by 26. Potter and DelDuca considered O^*, $O_2(^3\Sigma_u^+)$, N_2^*, and O_2^* ($v > 12$) as possible excited reactants for 10. All but O_2^* ($v > 12$) existed in insufficient concentrations at the emission altitudes to explain the sodium nightglow. Vibrationally excited oxygen molecules satisfying 26 are formed by 23 and by 24 and 25. Vibrationally excited oxygen is stable to collisional deactivation except with atomic oxygen. Even so, not all collisions with O deactivate O_2^* and those that do react de-excite only one vibrational level at a time. Thus O_2^* formed by 23 could remain sufficiently excited for some time. Using an estimated rate coefficient for 11 and an estimated de-excitation rate of O_2^*, Potter and DelDuca find that 11 can maintain the sodium nightglow intensity and also give the approximate layer distribution and altitude of maximum emission that have been observed by rockets. After sunset O_2^* production will occur principally through 23 above 95 km and through 24 and 25 below 90 km. Both processes will contribute since the free sodium is distributed over the range of both the O_2^* production processes.

Ballif and Venkateswaran (108) have considered sodium excitation by 4,6 and 11. They find that these processes can produce intensities and altitude distributions similar to those observed but that none is sufficiently close to those observed. They feel that the correlation of sodium nightglow with that of the hydroxyl molecule favors 6. Potter and DelDuca (106) claim that the Na^* and OH^* correlation favors the 11,24 and 25 process since both emissions would then depend upon the ozone concentration. If 11 and 23 were the excitation process, it would be expected that the sodium D lines and the atomic oxygen green line in the nightglow would have similar variations (106). This is not true and thus the process of 11 and 23 is not favored.

Observations of sodium trails injected between 60 and 140 km by rockets at night have been reported by Bedinger and his co-workers (109). They visually observed a dim glow at 60 km which decreased to 90 km. Above 90 km the glow increased to a very bright maximum at 100 km and then decreased again till at 120 km no glow was observed. Above 135 km a glow was again detected which became bright and persistent at 140 km. Bedinger et al (109) attribute the glow up to 90 km to 1,3,4,6,13,14 and 15. The maximum at 100 km and the glow above 135 km are attributed to O_2^* and N_2^* respectively (106).

The final choice of reaction mechanism is difficult to make. The process involving 11 involves several approximations and estimations. To explain the observed intensities Potter and DelDuca invoke a most favorable reaction rate and a somewhat low de-excitation rate of excited oxygen molecules (to give higher O_2^* concentrations). At the altitudes of interest the observations of minimum intensity reported by Bedinger et al (109) seem to support the chemical reaction of sodium compounds rather than the energy exchange process of 10. The process involving 1,2,3,6, 13,14 and 15 is on firm ground as far as the production of the observed intensity is concerned but fails to predict the observed altitude of emission. Ionic recombination and charge exchange processes can definitely be eliminated. Krassovsky's mechanism using reactions 5, 16 and 17 can be eliminated since it definitely predicts a covariance between the sodium D lines and the atomic oxygen green line in the nightglow. This covariance is not observed. The uncertainty as to the ability of 11 to furnish the necessary intensity in comparison with the chemical processes (1,2,3,6,13,14 and 15) seems to be the deciding factor in the choice of the excitation process. The chemical processes are favored but 11 cannot be definitely eliminated from consideration.

D. Altitude

The sodium airglow at night and at twilight has attracted the attention of many workers. Many attempts have been made to determine the altitude of the emitting layers (the nightglow and twilight glow do not necessarily originate at the same altitude). The methods used vary widely and involve complicated corrections in order to determine the true altitude of the emitting layer. The altitudes of the sodium twilightglow and nightglow will be discussed separately below.

1. Twilightglow

The high intensity of the sodium twilightglow makes it particularly suitable for height determinations by several methods. There is a large variation of intensity from zenith to horizon and also with time during sunrise and sunset. Interpretation of twilight measurements is made quite complex by several factors. The twilightglow altitudes predicted by chemical processes will be discussed first. This will be followed by a discussion of the various optical and photoelectric measurements.

The altitude of the twilightglow is the altitude of maximum atomic sodium concentration. The free sodium atom concentration is controlled by the various chemical processes already discussed. These chemical processes will now be examined to find the altitude predicted for the maximum neutral sodium atom concentration.

Hunten (102) considered 4, 13 and 14 to be the important reactions but felt that in the twilightglow layer 14 could be neglected as being too slow. On the basis of 4 and 13 Hunten found that the free sodium concentration depended upon the ratio of the atomic oxygen and ozone concentrations, $\left[n(O)/n(O_3)\right]$. In the mesosphere and lower ionosphere the ozone concentration decreases with altitude. This will cause the sodium to exist as free atoms at high altitudes and in the combined form (NaO) at low altitudes. At 80 km the Na and NaO concentrations will be nearly equal. The maximum sodium concentration altitude will depend upon and be controlled by the ozone to atomic oxygen transition. The variation of O and O_3 concentrations with altitude predict a maximum sodium concentration at about 85 km with a rapid decrease to small concentrations about 6 km below the maximum altitude. Above the maximum concentration altitude Hunten's scheme predicts that the free sodium atom decrease will approximately follow the barometric law. This will not be true if ionization is taken into account.

Potter and DelDuca (106) have made similar calculations but have included the effects of reactions 1, 3, 6 and 15. Their results give equal Na and NaO concentrations at 86 km and equal Na and NaO_2 concentrations at 81 km. Free sodium becomes predominant above these altitudes while combined sodium becomes rapidly predominant below these altitudes. The free sodium concentration will be a maximum at 88 km.

The free sodium atom distribution has also been computed by Dalgarno (45) using reactions 1,2,3,4,6,7,13 and 15. Dalgarno points out the importance of 15 which is known to have a very fast rate coefficient which decreases with increasing temperature. He finds that due to this reaction, 15, the free sodium will disappear rapidly below the mesopause and will probably have a maximum concentration at about 85±5 km.

Chamberlain (01) predicts a maximum free sodium concentration at 85 km. Oxidation into NaO and NaO_2 will occur below this altitude and the free sodium density above the maximum will be decreased principally by ionization. Chamberlain also predicts that the ratio of Na to NaO will depend upon both temperature and density while the ratio of NaO to NaO_2 will vary only with temperature (1) Above 80 km the ratio of Na to NaO will increase rapidly due to atomic oxygen.

The photoelectric and other optical measurements of sodium are difficult to interpret. The geometry of the measurements is relatively simple. In computing the solar flux incident upon the emitting layer the transmission of the atmosphere must be taken into account. It must be remembered that the solar radiation has already passed through the atomic sodium layer once before reaching the twilightglow layer. Resonance absorption and multiple scattering must be considered. In order to obtain accurate results the full use of radiative transfer theory is required. Since 1956 several papers have appeared in the literature in which resonance scattering by neutral atomic sodium in the upper atmosphere is discussed and developed for sodium airglow measurements. References 30, 33, 34, 56 and 84 are among the papers dealing with this problem. A condensation and overall review of this problem is given by Chamberlain (1). Twilightglow measurements made prior to 1956 have not incorporated all the corrections necessary for accurate altitude determinations and should be regarded with suspicion.

The earliest measurements of the altitude of the sodium twilightglow were made by Bernard in late 1937. From his measurements at Tromsø Norway, and at two locations in France he determined an altitude of approximately 60 km (79,84). From observations at Haute Provence, France, in early 1948 Barbier found the base of the emitting layer to be between 60 and 70 km and a concentration decrease above 70 km following the barometric law (110). In 1950 Barbier and Roach reported twilight observations which give a scale height for the sodium density of 250 km between 250 and 600 km (111). This is in strong contrast to the earlier results of Barbier (110). Blamont and Kastler found twilightglow altitudes ranging between 120 and 200 km from measurements made in the fall of 1950 at Haute Provence (112).

Vegard and his co-workers have assumed that the twilightglow is a result of photodissociation of a sodium compound caused by ultraviolet radiation. They believe ozone to be the screening layer for the exciting radiation. On this basis Vegard and his co-workers have reported a screening height of 43 km, an upper limit of the sodium layer of 109 km, and a layer thickness of 28 km for 35 observations made at Oslo between 1943 and 1949 (92,93,113). Their results for 27 observations at Tromsø show an average screening height of 34 km and an upper limit of the sodium layer of 104 km (93,114).

Hunten and Shepherd find the altitude of the maximum intensity of the sodium twilightglow layer to be 85±3 km. Their observations show a sharp concentration decrease below the maximum altitude and a sodium concentration scale height of 7 km above the maximum. Hunten and Shepperd's results are based upon 17 observations made at Saskatoon during July 1952, and the spring and summer of 1953 (115). Cronin and Noelcke report an average height of 84.2 km for observations made at Washington Grove, Maryland between Aug. 1953 and Jan. 1955 (116). The results for twilight observations made over a two year period at Saskatoon by Hunten show an average height of approximately 84.5 km with all heights falling within a 6 km range above and below the average altitude (117). Blamont reports an average height of 90.5 km for 91 observations at Haute Provence made between Nov. 1953 and July 1955 (118).

Observations made since 1956 to which all or most of the necessary corrections for resonance scattering have been applied are listed in Table VII. From Table VII it is readily seen that at all stations with observations extending over an entire year the altitude of the maximum sodium concentration is approximately 90 km. Blamont et al (121) find the thickness of the emitting layer to be 14 km thick on the average. They also found a difference between the altitude in the morning and evening twilight with the evening layer being 3.5 km higher on the average. Contrary to this report, Rees and Deehr (126) find no difference between morning and evening altitudes, but their findings are based upon only 11 observations while those of Blamont et al (121) are based upon 201 observations. The observations of Rundle et al (123) were tests of a new instrument and a new method for determining the sodium distribution in the twilightglow layer. They feel that their observations are of poor quality since not all of the errors of their instrument are known or understood. Bullock and Hunten found the sodium distribution to be symmetrical about the peak concentration. The thickness of the layer, defined as the height difference between the two altitudes where the intensity (concentration) is 20% of the peak intensity, averaged 29 km (124). The observations at Christchurch are evenly distributed throughout the year with two observations reported for each month. Of the 24 Christchurch observations, 18 show a maximum between 87 and 94 km and all 24 observations reveal a smoothly distributed layer which is symmetrical about the maximum (122).

The results of Donahue and Blamont (119) for Tamanrasset and Haute

TABLE VII

Altitude of the Sodium Twilightglow

Author	Location	Dates of Observations	No. Obs.	Ht. of Max.	Lower Hts (km)	Upper Hts (km)
Donahue and Blamont (119)	Tamanrasset 22.8°N	Jul 1958 - Dec 1959	-	89.8	53	170
Obrien (120)	Melbourne, Australia 37.7°S	Jun - Jul 1959	10	90±5	85	95
Blamont, Donahue and Weber (121)	Haute Provence, France 43.9°N	Nov 54 - Jul 56 Sep-Nov 57	201	87.6±0.9	-	-
Donahue and Blamont (119)	"	1955-1959	-	84.9	57	100
Hunten et al (122)	Christchurch, N.Z., 44°S	Jan-Dec 61	24	91±3	84	101
Rundle, Hunten and Chamberlain (123)	Saskatoon, Canada 52.1°N	May 1958	4	85	65	110
Bullock and Hunten (124)	"	Oct 1958 - Dec 1959	120	90.1*	83*	95*
Sullivan and Hunten (125)	"	5 Apr 62	1	95	80	115
Obrien (120)	Antarctic 60° to 66°S	Feb 1959	3	95±10	90	105
Rees and Deehr (126)	College, Alaska 64.9°N	Nov 1959 - Mar 1960	11	91.8	84.6	99
Delannoy (96)	Terre Adelie 66.7°S	Aug + Sep 1957+1958	-	90±5	-	-

* Correction of +3.5 km added to original values (127)

Provence seem to be of somewhat doubtful value. The large range of altitudes (53 to 170 km) found for the height of the sodium layer at Tamanrasset seems to be particularly open to question. The heights found exhibit great variability from day to day and from morning to evening. Much of this may be due to the height of the screening layer used. Donahue and Blamont applied screening heights ranging from 18 to 44 km. This range seems to be much too large. An examination of the data reveals that almost all of the questionable height values at both stations (extremely low or high heights) were obtained from observations at 180° with respect to the azimuth of the sun. Observations taken at both stations at 0° and 90° with respect to the sun's azimuth show a much smaller variability in the height values obtained and are closely grouped around the mean altitude. It seems probable that for observations at 180° azimuth either resonance absorption and scattering were not taken into account properly or that at that azimuth the method is much more sensitive to errors. All observations were made at a fixed 75° zenith angle.

Another interesting feature of the observations of Donahue and Blamont (119) is the difference between morning and evening heights. At Haute Provence the morning heights averaged 2.7 km higher than those for the evening. This is contrary to the earlier findings of Blamont et al (121) in which the evening heights were 3.5 km higher than the morning heights for 201 observations made at Haute Provence. At Tamanrasset the averaged results show the evening heights to be higher than the morning heights by 4.5 km. At Haute Provence the layer thickness averaged

8.1 km and at Tamanrasset 23 km. At both locations the morning layer was thicker than the evening layer. This was particularly true at Haute Provence where morning layer was almost twice as thick as the evening layer (119).

From the results presented here it is concluded that the sodium twilightglow layer is centered at 91±5 km. Due to the contradiction in the results of Blamont et al (121) and Donahue and Blamont (119) regarding the relative heights of the morning and evening twilight layers, it appears that there is little or no real difference between the average morning and evening heights. The layer thickness appears to be greater in the morning than in the evening. The average thickness seems to vary from one location to another. Observations at several locations using the same techniques in the reduction of the data are necessary before any conclusions can be made regarding the spatial variability of the thickness. There seems to be little variation of the layer height with latitude.

2. Nightglow

The first altitudes of the sodium nightglow were determined by the Van Rhijn Method. Cabannes et al (82,83) reported an altitude of 130 km from their measurements using this method in 1938. Dufay and Tcheng (128) obtained an altitude of 80±20 to 30 km for 189 observations taken between Oct 1940 and Jan 1944 at Haute Provence. Observations at College, Alaska, in Feb. 1949 using the Van Rhijn technique gave the altitude of the sodium nightglow as approximately 220 km (129). In a series of observations taken between 1948 and 1951 at Cactus Peak, California, Roach and Barbier (130,131) and Roach and Pettit (132) found heights of 275±50 km. In these observations at Cactus Peak a two color Van Rhijn method was used in which a second photometer is used to determine the background intensity at a spectral interval near the sodium doublet so that the background can be subtracted from the readings of the photometer centered on the sodium D lines. In 1952 Barbier critically examined all the sodium nightglow altitude determinations made up to that time. He concluded that the sodium nightglow layer was at approximately 300km (36). From measurements by the two color method taken during the winter of 1952-3 Roach and Meinel (133) derived altitudes ranging between 117 and 145 km.

Estimations of the altitude of the sodium nightglow layer have been made using the chemical reactions involved and estimated concentrations of the reactants. In his article proposing the excitation mechanism consisting of reactions 4, 13 and 14, Chapman estimated that the level of emission was between 60 and 80 km (98). Bates and Nicolet proposed that reaction 6 must also be considered. Using 6 in combination with 4 as the excitation reactions they concluded that the emitting layer was at approximately 70 km or lower (99). In 1953 Bates and Dalgarno (101) investigated reactions of ionized sodium as possible excitation mechanisms. They concluded that such reactions could not produce the sodium emission at the altitude of 300 km where it was concluded to originate by Barbier (36). They stated that the emission was probably due to excitation through chemical processes and hence could not originate much higher than 70 km (101).

Bedinger et al (109) state that reactions 1,2,4,6,13,14 and 15 can support the sodium emission only up to 90 km. Potter and DelDuca (106) estimate that with 4 and 6 as excitation mechanisms the nightglow maximum will be below 70 km. They propose two mechanisms, 23 and 11 and 24, 25 and 11. These two processes will produce a maximum nightglow layer at 88 and 83 km, respectively, according to their calculations (106).

The theoretical predictions of the sodium nightglow altitude and those determined by observations using the two color Van Rhijn technique seem irreconcilable. This difficulty was cleared up by Donahue and Foderaro (134) who found that self absorption within the sodium nightglow layer must be considered. This factor will cause a layer at 70 km to appear to be at 300 km when measured by the Van Rhijn Method. They concluded that sodium nightglow altitude determinations must be made through the use of radiative transfer theory so that the effects of self-absorption and of second and higher order scattering are taken into account.

Another method of estimating the altitude of the sodium nightglow layer was used by Bricard and Kastler. In 1944 they made measurements of the Doppler temperature of the sodium nightglow. They found a temperature of 240±50°K which they believed implied an altitude of approximately 74 km from their knowledge of the atmospheric temperature structure (89). Hunten, using a more recent model of the standard atmosphere, states that this temperature corresponds to a height of about 85 km (32).

The most complete and accurate observations of the altitude and distribution of the sodium nightglow layer have been made through the use of rockets. All sodium nightglow rocket observations reported to date have been made at White Sands, New Mexico. Dalgarno (45) has pointed out the necessity of subtracting out the emission due to the overlying 8-2 OH band, the continuum, and starlight, etc. from the intensity by photometers centered upon the sodium D lines. For all the rocket observations reported an effort has been made to do just this so as to isolate the emission due only to the yellow sodium D doublet.

The first rocket observation was made on Dec. 12, 1955 and the results have been reported by Koomen, Scolnik and Tousey (87,105) and by Tousey (41). They find the maximum sodium intensity at 85 km with the entire nightglow layer ranging from 70 to 100 km. Koomen et al (105) noticed a scattering of points which they attribute to layer patchiness and variation of the star background. In a reanalysis of the data from this rocket flight Packer finds the maximum intensity at 84 km and the layer boundaries at 70 and 107 km (44).

A second rocket flight was made on 3 July 1956. The results of this flight have been published by Heppner and Meredith (40) and by Tousey (41). The peak of the layer was found at 93±3 km. The upper and lower boundaries were rather indefinite. The lower boundary was reported to be between 73 and 85 km. Heppner and Meredith reported a slow decrease above 100 to 140 km where the emission intensity dropped to zero (40). Tousey disagrees with this somewhat and places the upper boundary at 110 km (41). The measurements for this flight appear to be significantly affected by OH emission (44).

Tousey has reported the results of a third rocket flight made on 23 March 1957 (41). He finds the maximum intensity at 95 km and the boundaries of the layer at 80 and 115 km. Packer's reanalysis of the data from this rocket observation locates the maximum at 89 km with the boundaries of the layer at 66 and 118 km (44). In this case Packer's results seem more reliable.

The results of all three rocket flights are summarized in Table VIII. The peak of the nightglow layer is concluded to lie at about 89±5 km and the layer appears to be 30 to 40 km thick. The lack of observations makes it impossible to tell whether the variation in the results is random, due to seasonal or diurnal effects, or due to errors in the measurements and data reduction. Many further rocket flights are needed in any case so as to confirm and elaborate the results already found.

TABLE VIII

Rocket Measurements of the Sodium Nightglow Altitude

Date	Maximum Rocket Altitude	Filter 1/2 width	Intensity Maximum	Lower Boundary	Upper Boundary
12 Dec 55, 10 PM	141	20Å	85	70	100
5 Jul 56, 1 AM	163	43Å	93	78-85	110-140
28 Mar 57, 10 PM	146	25Å	89	66	118

E. Variations

Large, short-term and seasonal variations exist in the height and intensity of the sodium airglow. Also, differences in sodium airglow features have been reported for different observing stations. This general variability is one of the most interesting features of the sodium airglow. The variations of the sodium nightglow and twilightglow will be discussed separately.

1. Twilightglow

Variations in the twilightglow were first noted by Bricard and Kastler in 1944 (89). Several workers have noted irregular dusk and dawn variations in altitude and intensity from one day to the next and even during the same day (93,118,119,123,124,135,136). Observations in different directions during the same twilight period have also revealed that the free sodium distribution is irregular and patchy (1, 137,138). The variations of twilight intensity are caused by abundance variations of sodium in the twilight layer.

Several workers have commented upon cases of abundances doubling over a very short period. Rundle et al (123) have a well observed case where the sodium abundance doubled in three days. Similar occasions have been observed by Bullock and Hunten (124) during June and July. In these cases both groups of workers have noted that the increased abundance is due entiraly to a sodium density increase at altitudes below the proviously observed density peak and result in a lowering of the density peak (123,124). The sodium density above the peak seems to be unchanged. Donahue and Blamont have found that the twilight sodium layer becomes higher and thicker with increases of intensity at Haute Provence (119). At Tamanrasset they find that cases of higher altitude layers are associated with slightly less intensity (119). The results for observations at Haute Provence seem to be in conflict with the results of Rundle et al (123) and Bullock and Hunten (124) at Saskatoon. The measurements of the latter two groups (123,124) appear to be far more reliable due to better instruments, a more refined theory, and a more thorough study of the events.

In 1954 Hunten (102) predicted that if ionization were important, then the morning intensity (abundance) would be higher than that found for the evening due to recombination of the sodium ions during the night. He also predicted that this morning-evening effect would be largest at the equinoxes and symmetrical about them. Since the average height of the twilight sodium layer is near 90 km, it appears that any effects of ionic recombination will occur principally above this altitude and thus cause an apparent increase in the height of the layer in the morning as compared to the evening height. Several observers have looked for these morning-evening height and intensity changes.

Cronin and Noelcke (116) reported a generally greater abundance of sodium at sunrise for observations taken over an 18 month period in Maryland. Observations by Donahue and Blamont (119) over a two year period revealed no constant morning-evening abundance variation. Nguyen-Huu-Doan (139) was unable to detect any difference between morning and evening abundances for his observations between Oct. 1959 and Dec. 1961. Blamont et al (135) state that a morning-evening abundance difference would result from a Doppler effect which would shift the intensity of the radiation which causes the sodium atoms to resonate.

A morning-evening height difference was found by Donahue and Blamont (119). For cases with the same sodium abundance, they found the evening altitudes to be lower than the morning altitudes at both Haute Provence and Tamanrasset. Blamont et al have found some indications that the sodium layer expands at night and contracts during the day (121).

Dayglow measurements by Blamont and Donahue reveal large sodium concentrations during the day which decrease within twenty minutes of sunset (140). From all these results it appears that if there is any ionization effect, it is either very rapid so as to be undetectable by the measurement techniques used, or it is so slow that it has very little effect over the period from dusk to dawn. Such effects as are observed may well be due to changes in the photochemical equilibrium from day to night.

Seasonal variations in the sodium twilightglow layer are very marked. A general pattern of strong twilight intensity in winter and weak intensity in summer was reported by Bricard and Kastler (89) in 1944. Several workers have reported the seasonal twilight abundance variation at Haute Provence for observations beginning in Nov. 1953 and continuing to Feb. 1962 (58,112,118,119,136,137,139,140). The results show a primary maximum twilight abundance in Nov. and Dec. (about 2.5×10^{10} atoms/cm^2). High abundances are found throughout the winter. After Dec. the abundance decreases slightly to a weak minimum in Feb. which is followed by a definite secondary maximum in March (about 2.5×10^{10} atoms/cm^2). The abundance decreases during the spring to a minimum in June and July (about 2 to 3×10^{9} atoms/cm^2) and then increases again to the Nov.-Dec. maximum. The months of maximum and minimum values have varied a little from year to year (58,119).

Nguyen-Huu-Doan has determined abundances from the ratio of the intensities of the two D lines (58,139,141) rather than from the absolute intensity. This method will give good results if the abundance is greater than 8×10^9 atoms/cm^2, but may give spurious results due to the patchy distribution of the sodium layer (138).

Observations of the twilight sodium abundance at Saskatoon have been reported by several groups of workers (1,45,91,117,124,136,138,142 and 143). Hunten (117) found a maximum in March and a minimum in June with a nearly linear intensity rise from June to the end of Feb. and a steep decline from the end of Mar. to the end of May. Hunten's observations were made over a period of two years and had excellent agreement from year to year. The monthly average abundance varied by a factor of five. Other observations at Saskatoon have essentially confirmed the results of Hunten (45,124,136,142). The total sodium abundance ranges from an average minimum value of 1×10^9 atoms/cm^2 to about 6×10^9 atoms/cm^2 at the maximum. Bullock and Hunten (124) observed minimum abundance values in Sept. rather than June and described the seasonal variation as being sinusoidal with approximately a three week phase lag behind the solstices for 120 twilight observations taken over a 15 month period.

The seasonal variation found by Cronin and Noelcke (116) at Washington Grove, Md., exhibits a maximum in Nov. and Dec. and a minimum in May with a secondary maximum in Mar. and a secondary minimum in Jan. At Oslo a definite maximum was detected at the winter solstice for observations taken between Sept. and May during several different years (93). A small annual variation was found at Tamanrasset which showed a weak maximum in Nov. and Dec. and a weak minimum in July. The average abundance at Tamanrasset was found to be about 4.5×10^9 atoms/cm^2 (119). Observations at Churchill showed an abundance variation similar to that at Saskatoon but with a smaller amplitude (124). Twenty observations of poor quality were made in England over a two year period and showed a principal maximum in Dec., a secondary maximum in July, and minimums in spring and autumn (144). Because of their sparsity and limitations these later observations will be omitted from consideration.

To confirm the seasonal variation of the sodium twilight abundance, several sets of observations have been made in the southern hemisphere. Observations at Terre Adelie (66.8°S) taken between April and Sept. reveal a maximum in July and August (145). Tinsley and Jones (146) made observations at Christchurch, New Zealand for a twelve month period. They found a maximum in the middle of winter, a sharp drop to a minimum in Sept., secondary maxima in Oct. and Mar., and a principal minimum in the early summer. More recent observations at Christchurch exhibit a maximum in the fall, a small secondary maximum in late winter, and a minimum in the latter part of spring. Abundances at Christchurch ranged from 3×10^9 atoms/cm^2 to 15×10^9 atoms/cm^2 (122).

The abundance variation of sodium in the twilight glow is definitely seasonal as shown by the observations made in the southern hemisphere. With the exception of the variation at Saskatoon and Churchill the general seasonal abundance variation exhibits a maximum in late fall and early winter, generally high values during the winter with a definite weak minimum in the middle or late winter and a secondary maximum near the spring equinox, and a primary minimum in the late spring and early summer.

A curious factor is the difference of the month of the maximum and minimum at various stations. At Haute Provence two maxima are observed in Nov. and March while at Saskatoon a single maximum is found in March. The strong seasonal variation may be due to changes in the photochemical equilibrium if sodium is combined in molecular form. The abundance variation may also depend upon the height of the mesopause temperature minimum which would change the value of some of the temperature sensitive rate coefficients controlling the sodium abundance in the emission layer.

Seasonal variations in the altitude of the twilight layer have been noted by some workers. At Washington Grove, Md., and altitude minimum (72 km) was detected in June and a maximum in Oct. (93 km). The height of the layer was generally lower during the summer and higher in the fall and winter (116). Blamont et al (121) report minimum altitudes in May and June for 201 observations at Haute Provence. They also find the emitting layer to be thicker during the autumn. Maximum heights in March (88 km) and minimum heights in Sept. (85 km) were found at Saskatoon (124). At Churchill at 10 km height variation has been reported with maximum values occuring in May and Dec. and minimum values in Feb. and Nov. (124). Donahue and Blamont (119) found the altitude of the twilight layer to average 88 km in Nov. and Dec. and 72 km in June with generally higher altitudes in the autumn and lower altitudes in the summer at Haute Provence. The altitude of the twilight layer varies more in the spring and summer at Haute Provence (119). At Tamanrasset the twilight layer was found at an average height of 105 km during May and Aug. and at 85 to 90 km for the rest of the year (119). The altitude varied most in the spring and least during the winter (119). Observations over a two year period at Saskatoon (117) and during a one year period at Christchurch (122) exhibited very little seasonal variation in the altitude of the twilight layer ($<$ 3 km).

A variation of sodium abundance in the twilightglow with latitude is evident from an examination of the observations reported at various stations. Observations taken at 16.4°S between June and Aug. (southern hemisphere seasonal abundance maximum) exhibit comparatively weak twilight intensities (15). At 22.8° N the average abundance was found to be only slightly larger than the summer minimum abundance at 43.9°N (119). The amplitude of the seasonal variation was also smaller at 22.8°N than at 43.9°N (119). At 52.1°N (Saskatoon) the abundance averaged one-half of that at 43.9°N (Haute Provence) and the amplitude of the seasonal variation was also smaller at 52.1°N (91,124,137). The amplitude of the seasonal variation at 58.8°N has been reported to be less than at 52.1°N (124). Observations in March at 55.2°N reveal an abundance of 8 × 10^9 atoms/cm^2 (147) which is less than one-half of the average March abundance found at 43.9°N. At 44°S the seasonal abundance was found to range from 3 × 10^9 atoms/cm^2 to 15 × 10^9 atoms/cm^2 (122). This is more than 2-1/2 times the abundances reported for 52.1°N. At 66.8°S the abundance during the winters of 1957 and 1958 averaged about 15 × 10^9 atoms/cm^2 (96).

The general latitude pattern appears to be one of large abundance and large amplitude of the seasonal abundance variation in middle latitudes (near 40°). At low and high latitudes there is strong evidence for a decrease of sodium abundance and a decrease in the amplitude of the seasonal abundance variation. The only exceptions to this are the winter observations at 66.8°S (Terre Adelie) which reveal greater abundances than observed at 52.1°N and 55.2°N. It may be that above 60°

the abundance increases again. Chamberlain points out that ionization effects predict the greatest seasonal abundance variation at the poles. Unfortunately series of high latitude observations have not been reported so that this effect could be confirmed.

Simultaneous twilight observations in different directions at Haute Provence have been reported by Blamont and various co-workers (135,137). These observations have shown a higher twilight intensity to the north of the station during the winter and to the south of the station in the summer. For the entire year of 1955 and the first half of 1956 a persistent, definite east-west difference was observed. There were 46 simultaneous east and west observations made during the first half of 1956 alone. Of these observations 33 showed a higher intensity to the west, two revealed a higher intensity to the east, and 11 were approximately equal. The ratio of the intensity to the west to that in the east averaged 1.43 and often reached values exceeding 2.0 (on one occasion the ratio was 3.5). This east-west effect exhibited great variability. There was no average difference in this effect between dawn and dust observations which helps to confirm the validity of the east-west difference. Observations taken during the fall of 1957 revealed no east-west difference. These observations strongly imply a standing eddy in the sodium concentration. No other observations have been reported which either confirm or deny the existence of an east-west effect. Additional thorough studies at other locations such as this one made by Blamont, Donahue and Stull (135) are needed to confirm the east-west effect. Such standing eddy effects might explain the difference between Saskatoon and Haute Provence in the time of the seasonal sodium abundance maximum.

2. Nightglow

It has been recognized for some time that variations exist in the sodium nightglow. In 1929 while announcing his discovery of the yellow lines in the night sky, Slipher remarked that the yellow emission was a permanent feature of the night sky but was not of constant intensity (78). Observations taken in France, Alaska, and Australia have all shown large night to night fluctuations approaching 50% of the mean intensity value (120,129,148). The night to night variations are largest in the winter (85). At Poona, India, (18.5°N) the night to night variation was found to be rather small (149).

Measurements have been made of the ratio of the twilight intensity to the night intensity after the end of twilight (120). This ratio was found to vary considerably (by a factor of three during a one week period). This variation in the intensity ratio indicates that a variation occurs in the nightglow excitation mechanism which would explain in part the night to night fluctuations of the nightglow intensity.

Observations have been taken of the nightglow intensity over the entire sky. Upon correcting the intensity to the zenith intensity, it was found that the nightglow intensity was highly non-uniform and that the distribution tended to be patchy (128,130). This non-uniformity appears to be the usual state of the sodium nightglow layer rather than the exception.

A diurnal variation of the sodium nightglow intensity has been found, but it seems to be affected by seasonal and latitudinal changes. Dufay and Tcheng, (150) on the basis of observations taken during 189 nights, reported the variation during the night to be small and irregular with no definite pattern. They did feel that there was a general post-twilight decrease for one hour and then a small increase during the remainder of the night. Barbier and Roach found the sodium intensity to be enhanced with the sun was 18 to 40° below the horizon (111). Roach and Pettit found that a diurnal maximum tended to occur during the end of the night (132). Roach later reported that there was little diurnal variation at Cactus Peak in the summer, but that in the winter there was a marked increase of intensity from midnight till dawn (95). Observations at Haute Provence (72) have yielded almost exactly the same results as these obtained by Roach at Cactus Peak. Observations by Pettit and his co-workers have shown minimum values during the first part of the night and maximum values at dawn (85,94). The amplitude of this diurnal variation ranges from 20 to 30% of the mean nightglow intensity (94,120).

At Tamanrasset and Abastumani a tendence for a maximum during the second half of the night has been found (63,119). At Poona the intensity was found to be high at the beginning of the night, but it decreased rapidly till 2300 local time and then decreased more slowly during the remainder of the night (149).

No observations concerning the variation in the altitude of the nightglow layer during the night have been reported. Ballif and Venkateswaran have examined the excitation mechanisms to see if changes in photochemical equilibrium would predict any changes in the height of the emitting layer. They found that if the sodium nightglow excitation is produced by reactions

4 or 6 (see Table VI) then the peak of the emitting layer will rise during the night. If the excitation is produced by reaction 11, the peak will remain steady during the night (108).

The seasonal variation of the sodium nightglow intensity is very definite. Cabannes et al were the first to comment on the seasonal variation. In 1938 they reported that the maximum intensity occurred during the winter and the minimum intensity occurred in the summer (83). Since 1938 many measurements of the seasonal variation of the intensity have been made by a number of workers at several locations.

The sodium nightglow intensity at Haute Provence exhibits a seasonal variation with a maximum in Nov., a weak minimum in Feb., a secondary maximum in Mar., a well defined minimum between June and Aug., and a sharp increase of intensity to the maximum in Nov. (72,137,148). This seasonal variation is the same as that observed for the twilightglow at Haute Provence.

At Cactus Peak, California, the intensity maximum occurred in Nov. and Dec. and the intensity minimum was found in June and July (85,95, 132,151). The winter intensity was found to be three to five times the summer intensity (78,82).

A seasonal variation similar to that observed at Haute Provence was found at Sacremento Peak, N. M. (94,152). The maximum intensity occurred in Nov. and the minimum intensity occured in June. The Nov. intensity was ten times the June intensity. There was a definite secondary minimum in late winter and a weak secondary maximum in April.

Observations of the nightglow at other locations in the northern hemisphere and at Melbourne, Australia, all show a general early winter maximum and an early summer minimum (45,53,119,120,126). The nightglow intensity and twilight abundance vary in the same manner.

Several observations have been made of the ratio of the D line intensities (D_2/D_1). Berthier found a ratio of 1.98±0.05 for five zenith observations at Haute Provence (153). Nguyen-Huu-Doan reports an average ratio of 1.9±10% for 14 zenith observations during the winter (141). Due to a self-absorption effect the ratio should vary with the season being low for high sodium abundance and high (near 2.0) for low sodium abundance (154). The D line ratio at a 75° zenith angle was found to vary with season as predicted and yielded sodium night concentrations ranging from 5×10^9 atoms/cm^2 in the summer to 2.5×10^{10} atoms/cm^3 in the winter (58).

Observations of the nightglow at Cactus Peak (85,95,151) Sacremento Peak (94), and Haute Provence (137) have all shown a difference between the intensity to the north and south of the station. At all three locations there is a higher average sodium nightglow intensity to the north of the station during the winter months and a higher average intensity to the south of the station during the summer months.

Almost no work has been done regarding the seasonal variation of the altitude of the nightglow layer. Observations at College, Alaska, between Nov. 1959 and March 1960 reveal an increase in height toward spring accompanied by an intensity decrease (126), but this result is based upon a small number of observations and hence cannot be given much weight.

As in the twilightglow, there is a definite latitude variation in the nightglow intensity. At 16.4°S the average nightglow intensity in Aug. was reported to be 24 R (15), a value well below that observed in middle latitudes. The average intensity at Tamanrasset (22.8°N) was found to be only half that observed at Haute Provence (43.9°N) (119). At Sacramento Peak (32.8°N) the average intensity was found to be about 160 R (94). At Abastumani (41.8°N) an average intensity of 88 R was reported for the sodium nightglow (53). Observations at Haute Provence during the summer of 1956 yielded an average intensity of 212 R (55). There is an inherent difficulty in comparing these results due to the inaccuracy of the absolute calibration of the photometers used. None the less, there seems to be a definite increase in the average intensity between the tropics and middle latitudes.

The amplitude of the annual intensity variation seems to exhibit a latitudinal effect. The annual variation at Tamanrasset was reported to be smaller than that at Haute Provence (119). At Sacramento Peak the intensity varied over the year by a factor of 10 (94). A factor between three and five was noted for the annual intensity variation at Cactus Peak (95,151). It is concluded that there is an increase in the seasonal variation of the sodium nightglow intensity between low and middle latitudes.

It is possible that the diurnal variation of the nightglow intensity may exhibit latitudinal effects. The night intensity variation at Poona (18.5°N) was observed to be a rapid intensity decrease from sunset to near midnight followed by a more gradual decrease of intensity till dawn. The observations from which this variation was derived were made between

Jan. and April (149). In middle latitudes during the winter months there is a general tendency to have a slight intensity decrease after sunset till local midnight, followed by a definite intensity increase between midnight and dawn. More observations at both high and low latitude stations are needed to confirm this latitude effect.

Both the sodium nightglow and twilightglow have very similar seasonal and latitudinal variations. There is also definite evidence of longitudinal variations in some of the sodium airglow features. Of particular interest in this later category are the east-west abundance difference found at Haute Provence and the difference between the time of the seasonal abundance maximum at Saskatoon and Haute Provence.

F. Correlations

There appear to be no correlations of the sodium airglow with solar or geophysical phenomena. Barbier has reported that the sodium airglow is not affected by sunspots (72). Pettit et al (85) have commented that there were apparently no correlations of the sodium airglow with sunspots. Dufay and Tcheng found the average sodium nightglow intensity to be constant over the three year period between the end of 1940 and the beginning of 1944 (148). During this period the sunspot number was in a decline from a maximum in 1937 to a minimum in 1944. This tends to show a lack of a correlation between the sodium nightglow intensity and sunspots. Dufay and Tcheng also reported that the sodium night intensity was not correlated with magnetic activity or magnetic perturbations (148). Hunten has confirmed this result from an examination of about 100 observations taken at Saskatoon over a two year period (117).

The sodium nightglow intensity is definitely correlated with the OH nightglow intensity. A comparison of the diurnal and seasonal variations of these two airglow components shows them to be very similar. The diurnal variation of both sodium and hydroxyl exhibit a tendency for a minimum around local midnight and a maximum in the second half of the night. Fishkova and Markova have found the diurnal variation of the sodium intensity to be parallel to that of OH (40). The seasonal intensity variation for OH shows a maximum in late fall or early winter, relatively high values throughout the winter, and a minimum in late summer while the variation for Na has a maximum in late fall, high values through the winter, and a minimum in early summer. The sodium nightglow intensity exhibits a secondary maximum in the late winter or early spring while at the same time the OH nightglow intensity is decreasing slightly.

Several workers have noted a definite, positive correlation of intensity between these two features (72,106). In a theoretical study of the various reactions controling these two emissions Ballif and Venkateswaran have found that the intensity of both emissions depends directly upon the ozone concentration and hence both will vary in the same way as the ozone concentration (73).

In a very thorough study Dufay and Tcheng have found a very high correlation between the sodium nightglow and the nightglow at 6300Å (148). They attributed this latter emission to the $^1D_2 - {}^3P_{2,1}$ transitions of atomic oxygen at 6300.3Å and 6363.8Å. Although some of the emission at 6300Å was undoubtedly due to the atomic oxygen red lines it is more likely that the emission of some of the OH bands in that region of the spectrum was the reason for the high correlation. Dufay and Tcheng cannot be blamed for this error since at the time of their report in 1947 the OH bands had not yet been discovered. The filters used to isolate the 6300Å region were undoubtedly of such band width to pass most of the emission of the 9-3 band of OH which ranges from 6234Å to 6417Å and also some of the 5-0 band of OH which emits between 6130Å and 6326Å. The intensity found for the red oxygen lines ranges between 50 and 100 R. The 9-3 OH band has an intensity of about 100 R and the 5-0 band an intensity of approximately 60 R. It is readily seen that the OH emission will probably dominate the atomic oxygen red lines. At a 75° zenith angle the sodium 6300Å correlation coefficient was found to be 0.832±0.027 for 127 observations and 0.737±0.048 for 91 zenith observations (148). Since the sodium emission occurs below 100 km and since the atomic oxygen red lines originate near 250 km it is rather unlikely that they will be correlated.

Several instances of simultaneous intensity variations of the sodium and atomic oxygen green line (5577Å, $^1S_0 - ^1D_2$) in the nightglow have been reported (107). Contrary to this Vegard et al could find no simple correlation between these two emissions (114). For 75° zenith angle Dufay and Tcheng found a correlation coefficient of 0.381±0.060 for 209 observations and a correlation coefficient of 0.103±0.104 for 91 zenith observations (148). Even here the 7-1 band of OH lying between 5542Å and 5696Å may be responsible for a large part of this corrélation since it is so small (the intensity of the 7-1 OH band is about 3 OR).

Several workers have attempted to find a correlation between meteor activity and sodium abundance or intensity. Krassovsky found intensity variations of the sodium airglow difficult to associate with meteorites (17). Chamberlain reported no correlation of sodium abundance with meteor activity (1). Vallance Jones studied the meteor influx rate for 24 hour periods at Christchurch, New Zealand. He found the sodium twilight brightness to be out of phase with the meteor influx rate and concluded that the sodium abundance in the twilight and night airglow was independent of the meteor influx rate (155). This lack of sodium airglow dependence upon meteor activity will be seen to be an important factor in determining the origin of the sodium in the upper atmosphere as well as the cause for the sodium airglow variations.

ATOMIC OXYGEN GREEN LINE

A. General Features

In the latter part of the 19th century evidence began to accumulate which suggested that the 5577Å green auroral line was present at all times in the night sky even in the absence of a visible auroral structure. Measurements of this "permanent aurora" or "non-polar aurora" by Rayleigh (156) in 1920 showed that this radiation was not polarized and hence could not be due to sunlight scattered by interplanetary gas. This fact has been confirmed by Bricard and Kastler in 1950 (90).

In 1921 Rayleigh reported that for a series of 145 observations at Terling, England, he was able to detect the green auroral line on photographic plates exposed over an entire night on about two out of every three nights (157). A continuing series of observations was conducted by Rayleigh. In 1922 he reported that he could regularly find the green auroral line in the night sky without any accompanying N_2 auroral bands (158). Simultaneous all night observations made to the north and south at Terling revealed no detectable difference in the intensity of the 5577Å line (159). From this Rayleigh concluded that this radiation which seemed to be present over the entire sky on most nights was not of auroral origin and termed it the non-polar aurora.

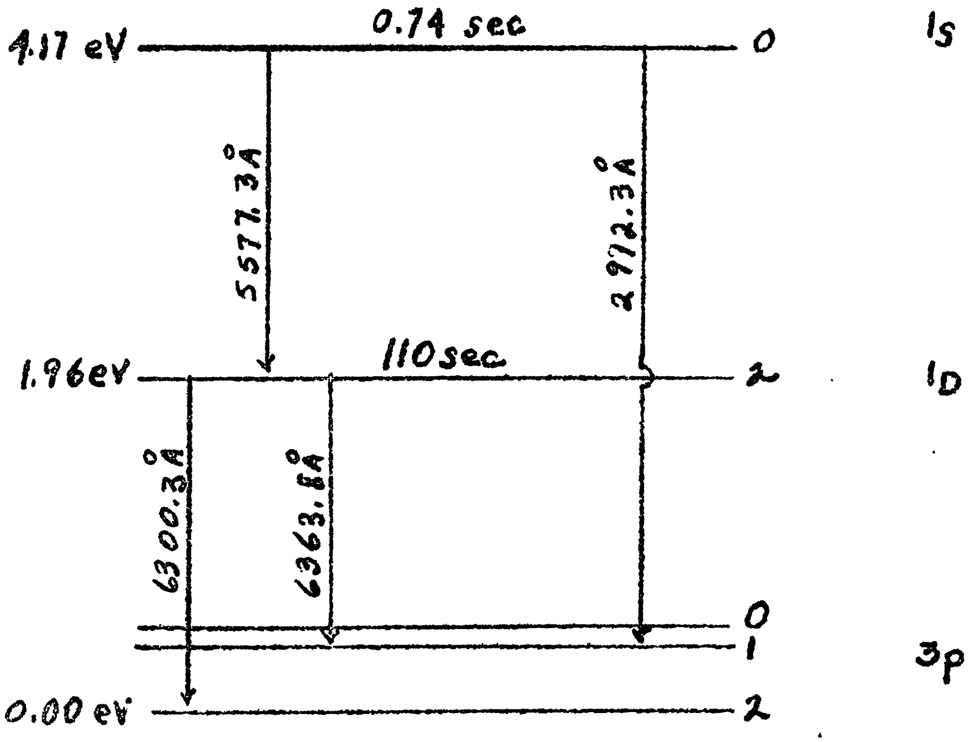
The basic differences between the 5577Å auroral and airglow (non-polar aurora) were defined by Rayleigh in 1924 (160). The basic characteristics of the 5577Å airglow are: it is not accompanied by the band spectrum of N_2.

it is present over the entire sky, it is essentially uniform over the entire sky (relative to the aurora), and it remains essentially constant for weeks at a time (in comparison to aurora).

Later attempts have been made to differentiate between auroral and airglow emission of the 5577Å line on the basis of intensity alone. St. Amand and Ashburn (161) state that the average 5577Å auroral intensity is about one and a half orders of magnitude greater than the 5577Å airglow intensity. Some workers feel that if the radiation is intense enough to be detected visually, then it is of auroral origin, but there is a question of whether or not aurorae exist below the visual threshold. Indeed, several workers report that the distribution of the intensity of the 5577Å radiation is smooth, unimodal, and positively skewed, indicating that the airglow intensities merge with those of the aurora (162,163). Roach, McCaulley and Marovich (162) have found this type of 5577Å intensity distribution to be common to all of the stations making observations during the IGY and propose that the 5577Å airglow and aurora emissions have a common origin. Barbier is in direct disagreement with this proposal. He believes the 5577Å aurora and airglow to be two distinct phenomena of entirely different character (164). In the light of the information accumulated over the past forty years since Rayleigh first differentiated between the 5577Å aurora and airglow, the major distinguishing feature between the two is that the aurora is accompanied by the first positive band system of the nitrogen molecule.

The substance responsible for the emission of the 5577Å radiation was not known for some time. At one time it was attributed to solid nitrogen. Some believed the emission to be due to krypton in the upper atmosphere. In 1925 McLennan and Shrum were able to reproduce the auroral green line in laboratory. They found the wavelength to be 5577.350±0.005Å and identified the emitter as atomic oxygen (165). Subsequent investigations have confirmed this identification and have shown that the emission arises from a $^1S_0 - ^1D_2$ transition of the oxygen atom. The $^1S_0$ state of atomic oxygen is a metastable state having a mean life of 0.74 sec and an energy of 4.17eV (166). The $^1D_2$ state is also a metastable state with a 110 sec mean life and produces a red doublet at 6300Å and 6364Å through the $^1D_2 - ^3P_{2,1}$ transitions (1). The ratio of the intensity of the 6300Å emission to that at 6364Å is three to one from the theoretical transition probabilities. The $^1S_0$ state may also decay to the $^3P_1$ state producing a line at 2972Å. The $^1S_0 - ^1D_2$ has a transition probability of 1.28/sec and the $^1S_0 - ^3P_1$ has a transition probability of 0.078/sec. Hence the 5577Å line is much more intense. Due to the relatively long mean lifetime of the $^1S_0$ and $^1D_2$ states collisional deactivation will play an important part in controling the emission of atomic oxygen in these states, particularly for the emissions having $^1D_2$ as the upper state. The energy levels and transition probabilities of the oxygen atom which appear in the airglow and aurora are shown in Figure II. Only the 5577Å airglow emission will be considered in this study.

Roach has estimated the mean intensity of the 5577Å airglow emission to be 250R (95). The median intensity of 21,088 observations of the 5577Å airglow taken at a number of widely separated stations during the IGY was found to be 254 R (167) in good agreement with Roach's estimate. The 7-1



LOWER ELECTRONIC STATES AND TRANSITIONS OF THE OXYGEN ATOM

FIGURE II

band of OH lies between 5542Å and 5692Å and hence contaminates observations of the 5577Å intensity. The 7-1 OH band is estimated to be responsible for about 6% of the intensity observed at 5577Å with the filters used at present (168). Chamberlain gives the total intensity of the 7-1 OH band as about 25 R (1).

Some observers have attributed a cellular structure to the 5577Å airglow. Roach et al estimate these airglow cells to be about 2500 km in diameter (169). Haug finds the cells to have a mean diameter of about 1000 km from his observations (170). The cells found by Roach et al appear to move at a speed of 92m/sec and have an energy emission rate of 1 to 15 × 10^{13} ergs/sec based upon a 2500 km cell diameter (171). The concept of large cells of 5577Å airglow emission drifting about in the upper atmosphere seems to have been generally accepted by the other workers studying the 5577Å nightglow.

B. Excitation Mechanism

Until recently it was thought that the problem of establishing the excitation mechanism for the 5577Å nightglow had been adequately solved. Until the identification of atomic oxygen as being responsible for the 5577Å emission the early workers were at a loss to explain the origin of this component of the nightglow. Rayleigh tentatively suggested that it might be due to a phosphorescence phenomena (160). In 1932 Chapman postulated a three body reaction mechanism between atoms of atomic oxygen which produces an oxygen molecule and an oxygen atom in the $^1S_0$ state (172). This mechanism is referred to as the Chapman process in the literature and is listed as reaction 1 in Table IX. The energy available from 1 is 5.2eV which is more than sufficient since the energy needed to excite the $^1S_0$ state of atomic oxygen is only 4.17eV.

The Chapman process was generally accepted by those studying the 5577Å airglow. There are several reasons for favoring this mechanism. Rocket measurements have shown that the 5577Å nightglow layer occurs at about the same altitude as the maximum atomic oxygen concentration (41,87), thus favoring reaction 1. Several workers at various locations have studied the intensity distribution of the 5577Å nightglow (107,164,173,174). They have found that the cube root of the intensity has a normal distribution. This would be expected if the Chapman process were operative, since the rate would depend upon the cube of the atomic oxygen concentration. Another reason for favoring the Chapman process is the high covariance of the intensities of the 5577Å emission and the molecular oxygen Herzberg bands (173). The three body atomic oxygen reaction can also proceed as shown by reaction 11 (see Table IX) and hence if the Chapman process were operative a high correlation between the 5577Å emission and the Herzberg bands would be expected. Roach et al state that any departures from exact covariance of these two emissions would be due to auroral enhancement of the 5577Å line (173). Tandberg-Hanssen and Roach have made estimates of the rate coefficient for the Chapman process. They found that based upon the atomic oxygen concentrations near 100 km the Chapman process could produce an emission intensity of the 5577Å line equivalent to that of a faint aurora which is far higher than the intensities normally observed in the nightglow (175). Other workers have also

TABLE IX

Reactions Associated with the 5577Å Nightglow

1. $O + O + O \longrightarrow O_2 + O(^1S)$ ✓
2. $O_2^* + CO \longrightarrow CO_2 + O(^1S)$
3. $O_2^* + N \longrightarrow NO + O(^1S)$
4. $OH^* + OH \longrightarrow H_2O + O(^1S)$
5. $O^+ + O^- \longrightarrow O + O(^1S)$
6. $O + M^* \longrightarrow M + O(^1S)$
7. $O + e(4.17eV) \longrightarrow O(^1S) + e$
8. $O_2^+ + e \longrightarrow O + O(^1S)$
9. $O_2(A\ ^3\Sigma_u^+) + O \longrightarrow O_2(X\ ^3\Sigma_g^-) + O(^1S)$ ✓
10. $X + YO \longrightarrow XY + O(^1S)$
11. $O + O + O \longrightarrow O_2(A\ ^3\Sigma_u^+) + O$ ✓
12. $O + O + M \longrightarrow O_2(A\ ^3\Sigma_u^+) + M$ ✓
13. $O^+ + O_2 \longrightarrow O_2^+ + O$
14. $O(^1S) \longrightarrow O(^1D) + h\nu$ (5577Å)
15. $O(^1S) + O_2(^3\Sigma_g^-) \longrightarrow O(^1D,\ ^3P) + O_2(^1\Sigma_g^+,\ ^1\Delta_g)$
16. $O(^1S) + N_2 \longrightarrow O(^1D, ^3P) + N_2^*$
17. $O(^1S) + O_2 \longrightarrow O_2^* + O$
18. $O(^1S) + N_2 \longrightarrow NO^* + N$
19. $O_2(A\ ^3\Sigma_u^+) \longrightarrow O_2(X\ ^3\Sigma_g^-) + h\nu$ (Herzberg)

found the rate of 1 to be sufficient to give the observed emission intensities (45,176).

Other excitation mechanisms have been proposed but have been found to be inadequate. Nicolet proposed a two step process involving reactions 13 and 8 (177). This process would be operative in the F region of the ionosphere, well above the presently accepted altitudes for the 5577Å nightglow, and will not produce the observed emission intensity.

Krassovsky suggested some reactions with vibrationally excited molecular oxygen which could produce atomic oxygen in the ^1S_o state (17). These are listed as reactions 2,3 and 4. Krassovsky himself believed 3 to be inadequate due to the low concentration of N in the emitting region but found that with concentrations of about 10^6 to 10^7 O_2^*/cm^3, 10^4 to 10^6 OH^*/cm^3, 10^6 to 10^7 CO/cm^3, and $10^8 OH/cm^3$, reactions 2 and 4 could produce the observed emission rate. Dalgarno argued that the concentrations of O_2^* and OH^* proposed by Krassovsky were much too high and that collisional deactivation was not sufficiently taken into account (45). In addition it seems that Krassovsky's proposed concentrations of CO and OH are too high if the emission takes place near the altitudes of maximum atomic oxygen concentration.

Reactions of the type shown by 10 were investigated by Gush and Sharma (176). For such a reaction to proceed the dissociation energy of XY must equal or exceed the dissociation energy of YO plus the excitation energy of O (^1S_o). They were not able to find any substances meeting this requirement in the upper atmosphere around 100 km.

Various processes involving ions were suggested but discarded for one reason or another (101,176,178). Martyn believes that charged particles are an essential constituent in the reaction mechanism and that they act as a catalyst (178). Bates and Dalgarno ruled out fast electrons and other excitation transfer processes as insufficient (101). Ion concentrations are too low to yield the observed emission intensities throughout the entire night (101,176).

Since 1959 several workers have supported an excitation process of a discharge nature (163,173,175,179). This process is shown as reaction 9. The electron would gain the necessary energy through acceleration by the electric field in the lower ionosphere. Roach has noted an increase of 5577Å intensity with current density in the lower ionosphere which, he believes, supports the discharge mechanism (163). Roach et al have found the intensity distribution curves to be unimodal and very similar for stations observing 5577Å airglow at low and high geomagnetic latitudes (173,174). The airglow observations at high geomagnetic latitudes are strongly contaminated with auroral 5577Å emission and this has led Roach et al to believe that a single process of a discharge nature is responsible for both the 5577Å auroral and airglow emissions.

Okuda has observed a simultaneous intensity increase over the entire sky which he believed implied a secondary excitation mechanism in addition to the Chapman process (107). He postulated that the additional excitation energy came from meteor showers and that the meteoric excitation (reaction 6) was two orders of magnitude less than that due to the Chapman process. In support of his proposal Okuda pointed out that if a factor of 2 to 3 R were subtracted from the 5577Å airglow intensities before the cube root was taken, the cube root of the remainder would exhibit a more nearly perfect normal distribution. Okuda found this to be true for the intensity distribution of all the stations he examined.

In 1961 Barth and Hildebrandt (181) reported the results of their laboratory studies of the Chapman process. In their work they investigated the rate coefficient of 1 and found it to be 8×10^{-38} cm^6/sec. Using this rate coefficient and the upper atmospheric densities of O_2 and O, they computed the rate of production of $O(^1S)$ by 1 and the amount of collisional deactivation of $O(^1S)$ by 15. The resulting emission rate of 5577Å by 14 was 0.2 R or roughly three orders of magnitude less than the observed emission rate (181). They concluded that the long favored Chapman process could not be responsible for the 5577Å nightglow.

Barth proposed a two step excitation mechanism to replace the Chapman process (182). The first step in Barth's mechanism involves the production of electronically and vibrationally excited oxygen molecules by reaction 12. The excited oxygen molecules then transfer the energy to oxygen atoms by 9. In 1962 laboratory studies were reported by Barth and Patapoff which confirmed the ability of 12 and 9 to produce the observed 5577Å emission rate (183). The use of O_2^* in the excitation of the OH and Na nightglow has already been discussed and rejected in both cases. One of the main reasons for doing so was that O_2^* is rapidly deactivated by collisions with O. It is this deactivation by collisions with O that produce the atomic oxygen nightglow.

Many of the reasons given in support of the Chapman process apply equally well to the mechanism proposed by Barth. The emission of the oxygen green line in Barth's process depends upon the production of $O_2(A^3\Sigma_u^+)$, but this is responsible for the emission of the Herzberg bands as shown by 19. Hence 12 and 9 predict the observed covariance between emission of the 5577Å atomic oxygen green line and the Herzberg bands of molecular oxygen in the nightglow.

Tohmatsu and Nagata have pointed out that both the Barth and Chapman processes will predict the same intensity distribution if collisional deactivation of $O(^1S)$ is sufficiently large (184). Gust and Sharma made a thorough study of the de-excitation of $O(^1S)$ by a number of possible reactions (176). They concluded that de-excitation reactions involving three body collisions or reactants with low concentrations such as N, Na, NO, H and OH were negligible in comparison with 15, 16, 17 and 18. These four reactions were found to be able to de-excite about one half the $O(^1S)$ produced before emission could take place (176). Thus, the condition for similar intensity distribution of the Chapman and Barth processes is satisfied and the emission produced by 12 and 9 will also be proportional to the cube of the atomic oxygen concentration. The Barth process is expected to operate best in the region of maximum atomic oxygen concentration which is about where the 5577Å airglow originates.

In summary, only the Chapman and Barth processes seem to fit the observed features of the atomic oxygen nightglow at 5577Å. The laboratory studies by Barth and Hildebrandt (181) and by Barth and Patapoff (183) seems to confirm that the process proposed by Barth (reactions 12 and 9) is responsible for the 5577Å nightglow emission. The proposal of a single excitation mechanism for both the auroral and airglow 5577Å emission involves some difficulty in that it fails to explain the absence of the auroral nitrogen bands in the nightglow. The ability of the Barth process to predict the observed emission intensity, altitude, and covariance with the O_2 Herzberg bands is strong evidence in support of the Barth process. The process of reactions 12 and 9 is concluded to be the 5577Å nightglow excitation mechanism.

C. Altitude

Several procedures have been employed to determine the altitude of the atomic oxygen 5577Å nightglow. The principal methods used to determine the altitude of 5577Å nightglow are the Van Rhijn technique, triangulation measurements, Doppler temperature measurements, and rocket observations. These methods will be discussed separately in the order mentioned above.

1. Van Rhijn Technique

All the earlier measurements of the altitude of the 5577Å airglow were made by the Van Rhijn technique. In order to obtain accurate measurements by this method corrections must be applied for absorption and scattering of the emitted radiation by the atmosphere. An allowance must also be made for the background emission in the same spectral interval.

A thorough study of the emission altitude was reported by Dufay and Tcheng in 1947 (128). For observations on 189 nights taken over a three year period they found the height to be 100±20 to 30 km. They later discovered an instrumental error that they had not taken into account (185). They applied a correction factor to their observations which increased the altitude to an average value of 215 km.

Abadie et al (186) deduced from their observations that there were two emission layers of 5577Å nightglow, one at 400 km and the other at 1000 km. They had difficulty in accounting for the background emission. It was later shown by Roach and Meinel (187) that the observations of Abadie et al could just as easily be interpreted as showing a layer at 130 km, the airglow layer, and a second, weaker layer at infinity, due to the starlight and zodiacal light background.

In the winter of 1948-49 Roach and Barbier (130,131) made measurements by a two color Van Rhijn method in which a second photometer is used to observe the emission near 5210Å. In the region near 5210Å there is no strong atmospheric emission, and it was assumed that the intensity observed in this region was the same as the background in the 5577Å region and hence would serve as a background correction. This is not entirely true since the 7-10H band near 5577Å must still be taken into account, but the second photometer observations do serve as a good correction for the continuum emission and the stellar background. Roach and Barbier made whole sky surveys and calculated the 5577Å emission altitude from the isophote maps which they had obtained from their surveys. They concluded that the altitude was 100±50 km from four observations by their two-color method (130,131).

The two-color Van Rhijn method was applied by several other workers to the determination of the 5577Å nightglow altitude. Altitudes were reported for measurements made at Haute Provence (188,189,190); Cactus Peak, Calif. (36,132,191,192,193,194); and Poona, India (168). All these different observations by different observers gave very consistent altitudes ranging between 200 and 250 km. Barbier critically examined all the observations reported up to 1952 and concluded the 5577Å nightglow altitude was 250±30 km (36).

In 1953 Bates and Dalgarno raised some objections to the observed 250 km emission altitudes (101). They argued that at 250 km no mechanism could be found to give the observed emission intensities and that if the Chapman process was responsible for the emission, then the altitude for the emitting region should be near 100 km (101). Barbier studied the corrections and assumptions used in determining the altitude by the two-color method (195). He concluded that the uncertainties in the extinction, the extraterrestrial light intensity, and the assumptions implied by the Van Rhijn method were not sufficient to change to 200 km to 250 km heights obtained if the results are based upon an extended series of observations (195). Manring and Pettit pointed out that the Van Rhijn method would give wide daily and diurnal variations in the height due to the patchiness of the emission (196).

Ashburn (197) found the error in the two-color observations in 1955. All the observations reported had used a correction for extinction by absorption in the ozone layer and by Rayleigh scattering in the lower atmosphere. Ashburn made measurements of the actual extinction coefficient at several locations. He found the measured coefficient to be much higher than those used in determining the 5577Å airglow heights. He found the difference to be due to scattering and absorption by haze and dust particles and water droplets in the atmosphere (197). Huruhata and Tohmatsu (198) concurred with Ashburn's conclusions.

Height measurements using the additional extinction correction proposed by Ashburn are much more in line with the altitudes predicted by the excitation mechanism. Two-color observations made at Jungfraujoch in the Swiss Alps by Elsasser and Siedentopf (199,200) have yielded a mean altitude of 90 km for the 5577Å nightglow. Observations at Cactus Peak on three nights gave altitudes for the 5577Å layer between 82 and 117 km (133). For observations made during 125 nights over a two year period at Sacremento Peak, N. M., Manring and Pettit found a height of 85±68 km (201,202). Other series of observations made in Japan (198), France (203), and the United States (at Fritz Peak, Colo.) (204) have all found the altitude of the oxygen green line nightglow to be at about 100 km in exact agreement with the altitude predicted by the chemical excitation mechanism.

2. Triangulation

The patchiness and irregularity of the 5577Å nightglow makes height measurements by triangulation quite feasible. Huruhata et al attempted triangulation height measurements from two stations in Japan located 1110 km apart (205,206). They were able to get only one simultaneous measurement on a small bright patch in the region of the sky seen by both stations. A height of 270 km was deduced from this single observation.

At Cactus Peak St. Amand et al (194) made triangulation measurements during two nights in conjunction with another station 317 km distant. They obtained altitudes of 100±10 km and 80±10 km for the two nights of observation. During the same nights observations by the two-color Van Rhijn method with the underestimated extinction coefficient resulted in an altitude of 200±20 km (194). Observations by Manring and Pettit (201, 202) from stations 220 km apart yielded altitudes between 80 to 100 km. Recent triangulation observations have been reported for stations 92 km apart (Churchill and Belcher, Manitoba (207). A statistical procedure was used to correlate the intensities found in the portions of the sky observed in common. The resulting altitude was found to be 95±5 km.

3. Doppler Temperatures

Several different measurements have been made of the Doppler widths of the 5577Å line. An interferometer was used in all the measurements. Measurements at the Lick Observatory yielded temperatures between 155 and 231°K (208,209). Observations in Ireland were made by Armstrong between Feb. and April 1956 (210,211). Measurements of the low order dispersion yielded temperatures between 150 and 235°K and higher order dispersion measurements gave temperatures ranging between 180 and 220°K. Taking the possible errors into account, Armstrong concluded that the mean temperature was about 190°K (210,211). Measurements by Wark at the Lick Observatory during May and June gave temperatures between 174 and 200°K with a mean value of 184±15°K (212). Winter measurements at Loparskaya, USSR (69°N) gave temperatures of 250±35°K and 260±35°K (213).

A model of the atmosphere above 50 km has been given by Nicolet. In his model the temperature is given as 150°K at 85 km and 200°K at 100 km (214). Observations of the temperature at 80 to 85 km in rocket grenade experiments at White Sands and Churchill show the actual temperature to be somewhat higher (20 to 30°K) than given in Nicolet's model (70). The measured Doppler temperatures are in general agreement and imply altitudes slightly under 100 km.

The measurements at 69°N are in disagreement with the others, but this is likely to be due to changes in the upper atmospheric structure with latitude and season. Measurements of the OH rotational temperature during the winter have shown the temperature to increase sharply with latitude above 60°N. This temperature increase undoubtedly occurs in the layer producing the 5577 Å emission also. Hence, it appears that the altitude indicated by the Doppler temperatures is about 100 km.

4. Rocket Observations

Five observations of the 5577Å nightglow layer have been made by use of photometers carried through the emitting layer on rockets. All of the rocket measurements reported in the literature have been made at White Sands, New Mexico.

The first successful rocket observation was made in Nov. 1955 (215). The 5577Å nightglow layer was found to range from 70 to 105 km. The peak emission was found to lie between 90 and 95 km. Some doubt exists in the height of the upper boundary since the maximum altitude attained by the rocket was only 105 km.

A second rocket shot was conducted in Dec. 1955. Koomen et al reported the results of this measurement (105). Data was obtained from the rocket for altitudes between 65 and 141 km. In a preliminary analysis the layer was found to lie between 80 and 115 km with a peak emission at 96 km. There was considerable scatter in the data due to background variation and layer patchiness. The final analysis showed the peak emission to be at 99 km, the lower boundary between 65 and 80 km, and the upper boundary between 110 and 120 km (87). The indefiniteness of the boundaries was attributed to layer patchiness.

A third rocket measurement was made in July 1956. From a preliminary analysis Heppner and Meredith found a sharp base at 90 km, a maximum emission at 94 km, a rapid intensity decrease between 100 and 104 km, and zero emission at 118 km (40). A full analysis of the data from this observation by Tousey (41) changed only the altitude of peak emission from 94 km to 97 km.

Tousey (41) and Packet (44) have reported the results of two additional rocket observations which give about the same values for the layer distribution as those reported above. The results of all five rocket observations are summarized in Table X.

Accurate measurements by four different methods have all yielded emission altitudes of the 5577Å nightglow layer of about 100 km. Once the early difficulties were overcome, the two-color Van Rhijn technique consistently gave emission altitudes near or below 100 km. With the exception of a single measurement of doubtful accuracy reported by Huruhata et al, the triangulation measurements all resulted in altitudes ranging from 80 to 100 km. The Doppler temperature measurements appear to give results consistent with altitudes slightly below 100 km. The peak emission altitude as deduced from five rocket flights through the emission layer is about 97 km. The heaviest weight is given to the rocket measurements due to the greatest accuracy of this method. It is concluded that the 5577Å atomic oxygen nightglow emission layer is between 30 and 40 km thick, has a peak emission between 95 and 100 km, and is distributed about the maximum with a smaller scale height below the peak then above the peak.

TABLE X

Summary of Rocket Observations of the 5577Å Nightglow

Date of Observation	1	2	3	4	5
17 Nov 1955 2 AM	105 km	-	90-95 km	70 km	105 km
12 Dec 1955 10 PM	141	20Å	99	80	115
5 Jul 1956 1 AM	163	29	97	90	120
28 Mar 1957 10 PM	146	18	98	95	107
6 Nov 1959 -	-	18	97	76	125

1 = Maximum Rocket Altitude
2 = Filter Halfwidth
3 = Altitude of Maximum Emission
4 = Lower Boundary
5 = Upper Boundary

D. Variations

The atomic oxygen nightglow emission at 5577Å exhibits considerable irregularity in both space and time. A large number of observational studies have been conducted at several locations in order to determine if there is any pattern to the intensity variations. The final portion of this section will deal with general latitudinal and longitudinal variations of intensity.

1. General Variability

In one of his early studies of the atomic oxygen green line Rayleigh stated that he found no sensible changes in the emission for weeks at a time (160). Rayleigh's instruments were rather crude and insensitive and Rayleigh was using the auroral emission as a basis for comparison. Observations with faster, more sensitive instruments have shown considerable variation of the intensity.

Large variations of intensity from night to night have been noted by observers in the United States (85,216), Russia (83), Australia (217,218) and India (219). Several studies have shown large changes in the diurnal pattern from night to night (83,217,218,220). Observations in Switzerland have shown that most intensity variations are less than 25% of the mean value but that in 5% of the cases the variation exceeds 50% of the mean intensity (170). Large intensity fluctuations during a single night have also been noted (221).

Observations by Dufay and Tcheng made at different points in the sky during the same night have revealed the 5577Å layer to be definitely not uniform (128). All sky surveys show the layer to be very patchy (130,222). Rocket observations have also detected considerable patchiness (87). Patchiness of the emission layer causes large apparent variations in the altitude of the layer during the night if the measurements are made by the Van Rhijn technique (131). The pattern of the intensity over the sky also seems to change with time (196,223,224).

Observations by Manring and Pettit revealed an intensity pattern which did not change shape from night to night but varied in intensity with time (196). Observations at Tamanrasset over a three year period generally revealed an arc of maximum intensity oriented E-W almost directly overhead. In general the intensity of the arc was relatively low at the beginning of the night but the relative intensity increased through the night until dawn such that it was the dominating feature of the intensity pattern (224).

There are rare occassions when the 5577Å nightglow intensity becomes bright enough to be detected visually. Several cases of this have been observed. Rayleigh noted one such occasion on 6 Nov. 1931. He made a thorough study of the emission and found the auroral nitrogen bands to be absent, indicating that the brightness was due to the nightglow (225).

2. Diurnal Variation

Early observations of the 5577Å nightglow by Rayleigh showed no definite diurnal variation (157,226). In 1928 McLennan (227) and Rayleigh (228) reported a distinct diurnal variation with a maximum near midnight for observations at Flagstaff, Arizona and Terling, England respectively. For observations over a period of one year Rayleigh found the ratio of the intensity at the beginning of the night to that at midnight to be 0.7 to 0.8 (228). In 1930 H. S. Jones reported a similar variation for observations at Canberra, Australia (229). Observations in France during the winter of 1935--36 revealed a wide variation in the time of the diurnal maximum with maxima occuring four to four and one half hours either side of midnight (230). Extended series of observations made in the early 1940's in France (150) and the United States (231) confirmed the existence of a general midnight maximum.

At Cactus Peak, California an extensive series of observations was made of the 5577Å nightglow variations (85,95,132,191,192,193,220,232). The diurnal variation generally exhibited a maximum within two and one half hours of local midnight. The maximum intensity was usually about twice the minimum intensity and the predawn intensities were higher than those just after twilight. In the summer the maximum tended to occur before midnight, and in the winter after midnight. The intensity pattern appeared to be fixed relative to the antisolar point. To an earth-bound observer the pattern appeared to move from east to west. There were local variations in the pattern and also changes in the pattern with time.

Observations at Sacramento Peak, N. M. exhibited a small average diurnal variation with a maximum at about one third of the way through the night (94). Three types of diurnal variation were noted: a midnight maximum (most common), a midnight minimum, and no variation at all (least common) (94).

Several studies have been reported for other locations. At Fritz Peak, Colo. a diurnal maximum was generally found to be near midnight (163). Observations at Haute Provence revealed a small diurnal intensity variation (less than 10% of the average intensity) and a diurnal maximum within two hours of midnight (188,190). Fishkova and Markova generally found a maximum near midnight for observations at Abastumani (53). The maximum occurred within three hours of midnight at Naini Tal, India (233). Observations on 179 nights at Maruyama, Japan (35°N) revealed a nightly maximum near midnight (234). At Camden, Australia, the maximum generally occurred three hours after local midnight (217,218). Barbier et al reported the results for observations made over a one year period at 49.3°S in the Kerguelen Islands (235). They found the average diurnal intensity variation to be only 8% of the average intensity, a diurnal maximum two hours past local midnight, and greater intensities during the second half of the night.

In contrast to the above reports Sandford found no diurnal variation at Scott Base, Antarctica during the winter (236). At Poona, India, observations made in 1934 showed a minimum beginning at midnight which lasted for two hours and a maximum at dusk and dawn (237). Recent results for observations made during the IGY at Poona showed a maximum three hours before midnight followed by a decrease of intensity till dawn (149).

Some observers have found a seasonal variation in the diurnal pattern. At Haute Provence Barbier reported that the diurnal intensity variation exhibited a midnight maximum between Nov. and Mar., a slight increase through the night between May and Aug., a slight decrease during the night during Apr. and Sept., and almost no variation in Oct. (72). These results were based upon 5-1/2 years of observation. Observations at Mt. Abu, India, detected a diurnal maximum one to two hours before midnight in Feb. and shortly after midnight in May (219). In Jan. and Feb. the intensity was generally greater before midnight and in Apr. and May the intensities before and after midnight were about equal. These observations were made between Jan. and June only due to the monsoon (219). The results of three years of observations at Sendai, Japan (38.1°N) show a seasonal variation in the diurnal pattern. The diurnal intensity variation showed a midnight maximum between Nov. and Feb., a general increase during the night between May and Aug., a general decrease through the night in Sept. and Oct., no variation in Mar., and a weak midnight minimum in April (223). The seasonal changes of the diurnal variation at Haute Provence and Sendai are seen to be very similar.

Recently there have been some reports of a twilight enhancement of the 5577Å emission. The early studies of Dufay and Tcheng showed no trace of a post-twilight decrease (150). In 1958 Dalgarno was unable to find any definite observational support for a twilight enhancement (45). Observations at Fritz Peak and Rapid City were reported by Megill in 1960 which showed a definite twilight enhancement (166). In a later study Megill et al reported that the twilight enhancement exhibited a seasonal dependence (238). For observations over a year at Rapid City they found that the twilight enhancement occurred most frequently in the fall and winter and almost never in the spring and summer.

3. Seasonal Variation

The earliest studies of the seasonal intensity variation were made by Rayleigh (160,226). For observations at Terling Rayleigh found an annual maximum in November and low values in the winter with a maximum to minimum ratio of 3.8.

Since 1930 a large number of studies have been made of the seasonal intensity variation. These are summarized in Table XI. A study of the data in Table XI reveals a general tendency at every station for an intensity maximum in the fall. There is also a general tendency for a minimum of intensity in the winter or spring. There seems to be a latitude effect in the seasonal variation. Stations located below 35° exhibit maxima in the early spring and early fall and minima in the winter and summer. Poleward of 35° there is a general tendency for a strong fall maximum and a prolonged minimum during late winter and spring. Many of the stations listed in Table XI reported smaller intensity variations which are not listed in the table.

TABLE XI

Summary of Seasonal Variation Studies

Location	Lat.	Ref.	Observation Record	J	F	M	A	M	J	J	A	S	O	N	D
				Northern Hemisphere											
				Southern Hemisphere											
				J	A	S	O	N	D	J	F	M	A	M	J
Tamanrasset Algeria	22.8°N	239	-	Min			Max			Min			Max		
		235	-	Min			Max				Weak Min		Max		
Mt. Abu India	24.6°N	219	Jan-May 57 Jan-Jun 58	Min				Max			No Observations				
Sacremento Peak, New Mexico	32.7°N	94	1-1/2 yrs	Min (125R)					Max (400R)		Min (125R)			Max (400R)	
		152	1-1/3 yrs			Max (680R)			Min (230R)				Max (470R)		Min (370R)
		234	-			Max (470R)				Min (250R)			Max (330R)		Min (250R)
Mt. Palomar Calif.	33.3°N	231	1 yr									Min		Max	
Cape of Good Hope	34°S	229	4 yrs		Min			Max			Weak Min		Max		
Canberra Australia	35°S	229	2 yrs			Min		WeakMax		Weak Min					Max
Maruyama Japan	35°N	234	4 yrs	Min (160R)				Max (310R)			Min (220R)		Max (380R)		
Cactus Peak Calif	36.1°N	232	4 yrs						Min (340R)				Max (420R)		

Table XI (continued)

Location	Lat.	Ref.	Observation Record	J J	F A	M S	A O	M N	J D	J J	A F	S M	O A	N M	D J
Cactus (cont.)	36.1°N	95	5 yrs					Min					Max		
		85	5-1/2 yrs	Weak Max	Min							Max			Min
Sendai Japan	38.1°N	223	3 yrs					Min					Max		
Abastumani USSR	41.7°N	53	1-1/2 yrs					Min					Max		
Haute Provence France	43.9°N	45	1-1/2 yrs			Min			Weak Max		Weak Min		Max		
		72	5 yrs		Min (260R)				Max (450R)			Min (300R)		Max (450R)	
		235	-		Min					Weak Max	Weak Min		Max		
Kerguelen Islands	49.3°S	235	14 mos				Min		Weak Max		Weak Min		Max		
Terling England	52°N	240	11 yrs		Min								Max		
Scott Base Ant.	77.9°S	236	May-Sep 59		Min					No Observations				High Intensity	

A seasonal variation in the average direction of maximum intensity has been found at several mid-latitude stations. At Sacramento Peak (32.7°N) the average direction of the maximum intensity is strongly north during winter, slightly to the south in the spring and fall, and slightly north or almost overhead in the summer (94,202). At Cactus Peak (36.1°N) between Dec. and Apr. and between June and Sept. the maximum intensity is almost always to the north of the station and generally to the south in April, May and Sept. through Nov. (85,95,232).

The ratio of the intensity at a 75° zenith angle toward the north to that toward the south was measured at Niigata, Japan (37.7°N). This ratio was highest in Jan. and greater than one in June and July indicating higher intensity to the north. In Apr. and Sept. the stronger intensity was found to the south (241).

A reversal takes place in the seasonal variation of the direction of maximum intensity at Haute Provence (43.9°N). Observations reported in 1953 gave the maximum to the south in Feb. and July (216). A later study reported in 1959 showed the maximum to be to the south between June and Aug. and in Nov. (239). Whereas the higher intensity was north in winter and summer and south in spring and fall for the three stations below 38°N, at Haute Provence the variation is rather irregular with a tendency toward stronger intensity to the south in the summer and winter.

4. Latitude and Longitude Variations

Rayleigh (242) was the first to conduct an extensive study of the variation of average intensity with latitude. With the cooperation of several other observers he obtained the average intensity measurements at

14 locations over the globe in both the northern and southern hemisphere. The data gathered by Rayleigh revealed a very irregular variation of intensity (242). About the only conclusion that could be made was that the intensity seemed to be lower at low latitudes.

Huruhata and Nakamura have reported the results of observations made aboard a ship traveling from Japan to the Antarctic (35°N - 65°S) and back. The first round trip was made between Nov. 1956 and Apr. 1957. The observational values exhibited considerable scatter but a general variation with geographic latitude was detected which showed a minimum intensity at the equator (243,244,245). A second voyage was made between Oct. 1957 and Apr. 1958. The intensity exhibited a minimum at the equator, a maximum at about 30°, a secondary minimum between 35° and 40°, and a rapid increase above 40° (246). The results obtained from the observations made during a third voyage were the same as the first two (247). In all three cases the variation was with geographic and not geomagnetic latitude.

Barbier made a comparison of the intensities at 43.9°N and 22.7°N. He found the intensity to be higher at higher latitude (55). Observations at 16.1°S (15) have yielded relatively low intensities (120 R compared to the accepted mean value of 250 R). At College, Alaska, the intensity between Jan. and April averaged 2400 R (248). This very large value was undoubtedly due to auroral contamination.

Some of the observations made during the IGY are summarized in Table XII. Only those stations reporting more than 500 observations which cover the entire calender year are included. As seen from Table XII the latitude

TABLE XII

Results of IGY Observations (167) of the 5577Å Nightglow Intensity* at Selected Stations

Station	1	2	3	4	5	6	7
Ondrejov	49.9°N	15.0°E	+49.8	696	239	335	450
Simferopol	44.8°N	34.1°E	+41.2	644	230	295	370
Rapid City	44.0°N	103.1°W	+53.1	967	195	270	375
Haute Provence	43.9°N	5.8°E	+45.8	2290	260	360	480
Memambetsu	43.9°N	144.2°E	+34.0	717	175	262	385
Fritz Peak	39.9°N	105.5°W	+48.7	1329	252	365	485
Sendai	38.1°N	140.6°E	+27.9	1170	185	245	325
Nugata	37.7°N	138.8°E	+27.4	696	215	280	365
Kakioka	36.2°N	140.2°E	+26.0	1505	165	215	280
Gifu	35.5°N	137.0°E	+25.0	612	135	200	280
Maruyama	35.0°N	140.0°E	+24.8	819	195	252	325
Sacramento Peak	32.7°N	105.8°W	+41.6	1232	200	279	411
Tamanrasset	22.8°N	5.5°E	+25.4	2750	128	170	220
Camden	34.1°S	150.6°E	−42.6	1984	124	173	241

* Intensity in Rayleighs (R)

1. Latitude
2. Longitude
3. Magnetic Latitude
4. Number of Observations
5. Lower Quartile
6. Median
7. Upper Quartile

variation of intensity is quite irregular. A good part of this irregularity is probably due to the difficulty in the absolute calibration of the photometers. No noticeable improvement is made if the intensity is taken as a function of geomagnetic latitude. In general the intensity is lowest in the tropics and increases with latitude. There is some indication of a secondary minimum between 35° and 40°. Some support for a secondary minimum is also given by observations made at four locations in Japan over a one year period of the direction of highest intensity. Three stations with latitudes ranging from 34.9°N to 36.2°N found a higher intensity to the south while the fourth station at 43.9°N observed a higher intensity to the north (249).

The amplitude of the annual intensity variation is affected by changes in location. Rayleigh and Jones found that the annual variation at Terling was much larger than that observed at either Cape or Canberra (240)). The variations shown in Table XI are quite irregular and show variations at a single station from one observing period to another. It can only be concluded that the amplitude of the seasonal variation varies from one station to another and also from year to year.

Several definite longitude effects have been observed. The 5577Å nightglow appears to be enhanced near the region of the South Atlantic geomagnetic anomaly (250,251). It can be questioned whether this is airglow or auroral enhancement. Observations over a two year period at Sacramento Peak revealed a consistently higher intensity to the east of the station which became more pronounced during the spring (202).

Some studies have been made of the diurnal variation to see if any longitudinal effects could be detected. Simultaneous observations have been made on several nights at Cactus Peak and Sacremento Peak. A difference was found in the local time of the diurnal maximum (94). It was concluded that the diurnal pattern exhibited small local variations and that there was some longitudinal variation (252,253). Simultaneous observations at Haute Provence and Cactus Peak confirmed the conclusion stated above (216).

Summarizing the different variations of the 5577Å nightglow the following general conclusions can be made. The emission is very patchy and exhibits considerable variability from night to night. The diurnal pattern changes with season in middle latitudes but on the average the intensity reaches a maximum value during the middle of the night. The seasonal variation pattern is a function of latitude. Below 35° a maximum intensity occurs near the equinoxes and a minimum intensity near the solstices. Above 35° there is a tendency for a fall maximum and a minimum during the winter and spring. The intensity is lowest near the equator and increases with latitude. There are definite indications that secondary intensity minimum exists between 35° and 40°. Definite intensity variations with longitude have also been found.

E. Correlations

The atomic oxygen green line in the nightglow is probably correlated with more phenomena than any of the other airglow emissions discussed in this study. There are several strong correlations with other nightglow emissions and also some relations with various geophysical phenomena.

A high correlation has been found between the 5577Å line and the O_2 Herzberg bands in the nightglow. Observations at Fritz Peak, Rapid City, College, and Thule showed departures from this correlation only during auroral enhancement (N_2 + bands observed) (173). For observations over a 2-1/2 year period at Fritz Peak, Roach found a correlation coefficient between the intensities of 5577Å and the 3670Å Herzberg band of 0.93 (163).

Several workers have reported a strong correlation with intensities in the continuum region. A series of observations at Haute Provence have revealed that the 5577Å nightglow line and the 5180Å region of the continuum vary similarly on both a diurnal and annual basis (188,190). Tohmatsu detected a good correlation of the atomic oxygen nightglow and the 5300Å continuum emission (254). Observations at Brisbane, Australia also revealed a high correlation with the intensity at 5300Å (174). Roach found a correlation coefficient of 0.86 between the 5577Å and 5260Å continuum emissions in the nightglow for data collected over a 2-1/2 year period at Fritz Peak (163). For the same observations Roach found a correlation of 0.84 between the 5577Å and 4400Å continuum emissions (163).

Several studies have revealed only a weak intensity correlation with the 6300Å atomic oxygen nightglow (101,148,236). Dufay and Tcheng made observations at the zenith and at 70° from the zenith and reported correlation coefficients of 0.159±0.100 (95 observations) and 0.356±0.077 (127 observations) respectively (148). It is not surprising that the correlation is so low since in the region of the 5577Å emission the long lifetime (110 sec) of the ^1D state which is the upper state for the 6300Å emission allows enough time for the ^1D oxygen atoms to be collisionally deactivated.

The intensity correlation with the sodium nightglow was found to be 0.103±0.104 for 91 zenith observations and 0.381±0.060 for 209 70° zenith observations at Haute Provence (148). No correlations have been found between the intensities of the atomic oxygen green line and the OH bands.

Some correlations with ionospheric phenomena have been found. St. Amand reported that the 5577Å intensity variation was similar to variations in the height of the F region (255). An increase of 5577Å nightglow intensity with current density in the lower ionosphere has been found by Roach (183). Observations at Fritz Peak of the 5577Å intensities which exceed 450 R show a good correlation with several ionospheric parameters (179).

There is some disagreement on the correlation between 5577Å nightglow intensity and magnetic phenomena. The early observations of Rayleigh showed no relation with the degree of magnetic disturbance (157,226). Later Rayleigh and Jones found indications of higher intensity on magnetically disturbed days (240). Observations over a 3-1/2 year period at Haute Provence revealed some correlation with strong magnetic activity and a strong correlation with magnetic perturbations (148). Observations during 66 nights at Cactus Peak revealed an apparent negative correlation of 5577Å intensity with K_p*(232).

Since 1959 a number of studies have been reported on the correlation of 5577Å intensity with magnetic phenomena. Observations at Camden, Australia (-42° Magnetic) reveal little dependence upon K except for K greater than 7 (217,218). After removal of autocorrelation no relation with K_p could be found in the data taken at Maruyama (234) and only a very weak inverse

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* K_p is the average planetary magnetic index, K is the local magnetic index.

relation with K_p (5% significant) could be detected in the observations at Sacremento Peak when K_p was seven or less (256). At Sacremento Peak there seems to be a direct relation between the intensity and K_p for K_p greater than seven, but this is based upon only ten cases (256). Okuda observed enhancements of intensity but no relation with geomagnetic activity (107). At Brisbane, Australia a possible correlation with high K was found (174). At Fritz Peak and Rapid City a strong correlation was found by Roach when K exceeded four (257). The range of observed intensity also increased with K. It was concluded by Roach that these relations were due to contamination by auroral emission (257). At College, Alaska both the intensity and the range of intensity variation increased with magnetic activity (248). Observations to the north and south at Invercargill, New Zealand (-51° Magnetic) showed a strong dependence of emission intensity on K_p in both directions whether aurora could or could not be detected visually (268). A strong intensity dependence upon K was found for observations at Scott Base (77.9°S) when K exceeded three (236). The relation was better with the local K than with K_p.

The 5577Å nightglow intensity was compared with the local components of the magnetic field. At Fritz Peak a good correlation was found with negative changes in H, particularly when the intensity exceeded 300 R (259). No periods of geomagnetic activity were included in this study. Observations at Brisbane have confirmed this result (174).

At Kakicka a semidurnal enhancement of intensity with lunar time was observed (260). Davidson examined the IGY observations at Camden, Haute Provence, and Rapid City for variations with lunar time. A variation was found at each station but the phase at each station was different and there was a large statistical uncertainty in the results (261). Davidson pointed out that most observations were made during periods without moonlight and when this was taken into consideration along with the uncertainties found above it must be concluded that little or no variation of intensity with lunar time exists (261).

Several attempts have been made to correlate the 5577Å nightglow activity with solar activity. The earlier studies by Rayleigh showed no relation with the transit of sunspots (157) or short term sunspot variation (160). A later report by Rayleigh and Jones for observations at Terling (11 years), Canberra (8 years) and Cape (8 years) gave correlation coefficients with the 11 year sunspot cycle of 0.60, 0.79 and 0.89 respectively (240). Dufay and Tcheng found a 0.5 correlation coefficent with the relative sunspot number for 3-1/2 years of observations (148). There was a definite correlation with sunspots for nine years of observations at Sacremento Peak (262). At Cactus Peak no apparent relation with sunspots could be detected for observations between Nov. 1948 and Mar. 1954 (85,232). Observations over periods less than a year showed no indication of correlations with sunspots (233,246). Observations over a five year period at Sendai also failed to show any relation with sunspots (263). There are observations of a 27 day recurrence of high 5577Å intensity which indicate the possibility of a direct solar influence (148,246).

The 5577Å nightglow emission is seen to be definitely correlated with the continuum and O_2 Herzberg emission. There are definite indications of a relation with magnetic activity and ionospheric variations. Whether these last two relations are due to auroral enhancement of the 5577Å line is open to question. Stations with a long record of observation show a definite correlation with the sunspot cycle while no relation is obvious at stations with a shorter observation period.

RELATIONS BETWEEN THE AIRGLOW AND DYNAMICAL PROCESSES

Photochemical equilibrium processes are entirely unable to explain certain features of the airglow emissions. At high latitudes the OH nightglow emission not only continues during the entire winter but also reaches maximum intensity during the middle of the winter when photochemical equilibrium predicts that the intensity should decrease. Several investigators have shown that neither ionization nor chemical combination can yield the large seasonal variation of sodium abundance that is observed in the middle and high latitudes (117,142,165). Tohmatsu has shown that the intensity of the atomic oxygen 5577Å nightglow depends upon the total amount of atomic oxygen in the column, the vertical distribution of the oxygen atoms, the reaction rate, and the transition probabilities. The reaction rate and transition probabilities are essentially constant so that the only variation would be the amount and distribution of atomic oxygen (atomic oxygen concentration). Tohmatsu found that the time constants for the change of atomic oxygen concentration by photodissociation or chemical recombination were quite large (greater than 10^6 sec) at 100 km such that photochemical equilibrium in a static atmosphere is totally unable to explain any of the diurnal or short term 5577Å intensity variations (264). The seasonal intensity variations of the atomic oxygen and sodium nightglow also appear to defy explanation on a basis of static photochemical equilibrium.

Under photochemical equilibrium conditions the ozone-hydrogen excitation process of the OH nightglow takes place at about 70 km. The results of rocket observations show that the maximum OH emission layer is 10 to 15 km higher than 70 km. If the OH excitation is due to $O_2^{\#}$ as proposed by Krassovsky, then the emission should originate at 95 to 100 km (about 10 to 15 km higher than the altitude deduced from rocket observations). The altitude of the sodium nightglow emission as predicted by the controlling chemical reactions is 15 to 20 km lower than that observed by the use of rockets (45,88).

Studies of the deformation of sodium vapor trails ejected from rockets, the deformation of ionization trails left by meteors, the spread of radioactive tracers injected by high altitude atomic tests, and studies by rocket grenade, chaff, and falling sphere experiments have all shown that the upper atmosphere is far from static. Strong winds have been found in the region of the airglow emissions. Several of these same experiments have also revealed the existence of strong turbulence and significant tidal effects at the altitudes of the airglow emissions. The observed turbulence, tides, diffusion and vertical and horizontal motions undoubtedly play a significant part in controlling the emission intensity of the airglow by changing the concentrations of the various reactants in the chemical processes producing the emission. Observations of the atomic oxygen nightglow have revealed movements of the nightglow pattern at speeds of 90 to 150 meters/second (171,174). Evidence has been found for a definite rotational motion of some of the features in the atomic oxygen nightglow pattern (171). The difference between the observed features of the airglow and those predicted by photochemical equilibrium in a static atmosphere must be due to dynamical processes.

A. The Effect of Turbulence

All three airglow emissions investigated in this study exhibit an irregular, patchy distribution of intensity. Tohmatsu (264) attributes the 5577Å nightglow patchiness to variations in the atomic oxygen concentration of the emitting layer. The patchy distribution of the OH and Na nightglow are also due to variations in the concentration of the reactants which produce the emission. The concentration variations are probably due to turbulence.

Several workers have investigated the wind structure of the layer between 80 and 110 km by observation of meteor trails (265,266,267) and the injection of sodium clouds (268,269,270,271,272). All these investigations have shown the airglow emission layer to be quite turbulent with a wide range in eddy size. The larger scale eddies could quite possibly be responsible for the patchy appearance of the nightglow. Silverman has observed an occasion of large short term intensity fluctuations throughout a single night. During this night there were no indications of auroral activity but strong radio star scintillations were observed indicating density variations in the upper atmosphere due to turbulence (221).

In order to explain the patchiness of the OH emission Nakamura (27) investigated the effects of localized turbulence upon the change of the reactant concentrations in the emitting layer. He based his calculations upon the ozone-hydrogen reaction and used the results of Bates and Nicolet

(28) who have determined the ozone and hydrogen concentrations based upon photochemical equilibrium. Nakamura found that if the OH layer is at 70 km the intensity will increase due to ascending currents into the emitting layer, which will locally increase the ozone concentration. Descending currents into the 70 km region will bring down excess atomic hydrogen from the region near 75 km where Bates and Nicolet (28) predicted the maximum atomic hydrogen to be located upon the basis of photochemical equilibrium and thus also increase the OH emission intensity. Nakamura also found that if the OH emission layer lies above 75 km and below 90 km, the intensity will be increased by ascending currents which will bring up higher ozone and hydrogen concentrations but will be decreased by descending currents which will bring down lower ozone and hydrogen concentrations. The vertical currents which occur as a result of turbulence in the mesosphere are easily capable of varying the OH emission intensity by an order of magnitude according to Nakamura's calculations (27). The rapid temperature decrease normally found in the upper mesosphere (70 to 85 km) will allow strong vertical currents due to turbulence to exist in this region.

The difference between the observed OH emission altitude and that predicted by photochemical equilibrium can also be explained by vertical mixing. Krassovsky, in one of his arguments against the ozone-hydrogen process, remarked that, at 70 km, there was insufficient atomic oxygen to support the OH emission by the ozone-hydrogen process. He argued that the ozone-hydrogen process could only be sufficient if strong vertical convection were occurring which would bring the needed atomic oxygen down to the emitting altitude (23). The atomic oxygen is formed above 100 km by dissociation of O_2. Krassovsky later found that, even for his O_2^* process, atomic oxygen must recombine below 100 km if the rate is to be sufficiently rapid and hence must be brought down into the 90 to 100 km region by turbulence (17). More recent calculations by Krassovsky have shown that, regardless of which process produces the OH nightglow, the hydrogen recycle process which involves atomic oxygen requires a large amount of atomic oxygen to enter the emitting region (26).

Turbulence could possibly alter the vertical distribution given by photochemical equilibrium of the concentration of the various chemical constituents in the airglow region. It could possibly increase and even maintain higher atomic oxygen concentrations below 90 km than those predicted by photochemical equilibrium. Vertical turbulent mixing could also increase the H, O_3, Na, NaO and NaO_2 concentrations above 75 km so as to give the observed altitude distribution of the OH and Na nightglow emissions. Even the atomic oxygen 5577Å emission seems to originate somewhat below (4 to 6 km) the altitude of maximum atomic oxygen concentration predicted by equilibrium considerations. Downward mixing of atomic oxygen by turbulence could very well be responsible for this maximum emission altitude difference. Three rocket measurements have been made of the O and O_2 concentrations between 62 and 175 km. Below 85 km and above 120 km the concentrations of O and O_2 agreed with theoretical predictions (273). Between 90 and 120 km, no meaningful analysis of the rocket data was possible. This result could be due to turbulent mixing in the 90 to 120 km layer. Nicolet has reported that rocket observations of the O_2 distribution above 100 km show that atmospheric mixing processes are operative at and above this level (274).

B. Tidal Effects

The dirunal variations of the OH and Na nightglow show a tendency for minimum intensities near or somewhat before local midnight. The atomic oxygen nightglow exhibits a tendency for an intensity maximum near midnight. In all three cases the diurnal variation seems to be dependent upon latitude and season. Tidal effects causing density (concentration) changes in the emitting region might explain the observed diurnal intensity variations of the nightglow.

Observations of meteor trails (266, 267, 275) have shown that above 80 km tidal effects are important. Both dirunal and semidiurnal tidal components are observed between 85 and 105 km. At 96 km a diurnal density variation with an amplitude of 13% of the mean value has been found from observations at 53°N during the winter (275). The amplitude of the 24 hour density variation increased by a factor of three between 85 and 105km and a 12 hour density variation with an amplitude of one half the 24 hour variation was also found. A seasonal variation in the phase of the diurnal tidal wind between 85 and 100 km was found for observations at Jodrell Bank (53°N) over a five year period. The phase changed by about 10° per month from Jan. to Aug. but between Sept. and Nov. the phase shifted rapidly and irregularly. The time of this sudden phase shift varied from year to year (267). A semidiurnal tidal wind component of 20 m/sec was also found at Jodrell Bank. The amplitude of both the 12 and 24 hr tidal wind components increased with altitude with the larger increases taking place during the winter (276).

The atomic oxygen emission intensity has been shown to vary with density if the emission layer expands or contracts adiabatically. A 10% change in density will cause a 30% change of intensity. Higher densities yield higher emission intensities. Vertical velocities of a few cm/sec are sufficient to cause the observed diurnal intensity variations (184,241). If tidal effects are responsible for the intensity variation the variation will be a function of local time. The 5577Å emission will be a function of location if turbulence is responsible for the intensity variation (176). Pressure oscillations due to tidal effects will also cause intensity variations since the atomic oxygen concentration in the emission region is essentially constant during the course of a single night (277). Simultaneous observations have been made of the 5577Å nightglow at Cactus Peak and Sacremento Peak (252,253) and also at Haute Provence and Cactus Peak (218). In both these cases the dirunal variation was found to be a function of local time with small local modifications thus showing that both tidal and turbulence effects are operative with the tidal effect being predominant.

The seasonal variation of the diurnal intensity pattern of the 5577Å emission may be associated with the seasonal phase change of the diurnal, tidal wind component of the type found at Jodrell Bank (267). The intensity variation and the diurnal wind components are both strongest and both exhibit irregularity during the autumn.

A study by Ballif and Venkateswaran (77) of the diurnal variation of the OH nightglow as produced by the O_2^* and O_3 processes has shown that neither of these processes can duplicate the observed diurnal variation when considered within a framework of local photochemical equilibrium. These two processes are the only ones known to be capable of producing the observed high emission intensity. Hence some dynamical process must be operating to change reactant concentrations in the emission layer.

Vertical motions were postulated by Nakamura in an attempt to explain the diurnal variations of intensity as well as the patchiness of the OH nightglow (27). Nakamura determined the variations of the absolute intensity by taking an average of the intensities over the entire sky after they had been reduced to equilivant zenith intensities. The nightglow layer was assumed to ascend and descend adiabatically. The changes in the emission intensity were assumed to be due only to changes in the reactant concentrations caused by the adiabatic expansion and compression of the emitting layer. Nakamura calculated the adiabatic change in the height of the emission layer necessary to reproduce the observed diurnal absolute intensity changes. He also calculated the vertical velocities necessary to accomplish the altitude change so that the rate of variation of the emission intensity would be equal to the observed rate. Nakamura's results showed that adiabatic descent would cause the intensity to decrease and ascent would cause the intensity to increase. The vertical velocities found in this manner were between 4 and 10 cm/sec, which could be produced by tidal effects.

Nakamura's observations of the OH nightglow revealed a seasonal variation of the diurnal intensity variation in which the daily curves showed a greater similarity in the winter than in the summer with a definite midnight minimum in the winter and constant or slightly decreasing intensity during the night in the summer. This variation might be related to the strong diurnal tidal wind component found during the winter and the weaker diurnal wind component observed during the summer. Nakamura also noted that his observations of the diurnal intensity variations seemed to show a diurnal and a semidiurnal component which would lend further support to a tidal excitation mechanism.

The diurnal pattern of the sodium nightglow intensity has been found to be very similar to that observed for OH with the Na diurnal pattern also exhibiting the same seasonal variation as that of OH. It seems very likely that the diurnal sodium intensity variation is controlled by the same mechanism as the OH diurnal intensity variation. Hence it seems to be very possible that all three components of the nightglow have diurnal variations due to tidal effects.

C. Short Term Intensity Variations

Observations of the sodium, hydroxyl, and atomic oxygen nightglows have shown that large, random, day to day variations are a characteristic feature of these emissions. These random intensity fluctuations could be associated with the large irregular, day to day variations about the seasonal mean wind that have been shown to exist between 85 and 100 km by meteor trail observations (266,267,276). Unfortunately there is no series of simultaneous observations of both the airglow intensity and meteor winds at the same location to show if these two phenomena are correlated in such a manner that the day to day differences in the wind will cause corresponding changes in the airglow intensity.

D. Origin of the Upper Atmospheric Sodium

One of the major problems arising from a study of the sodium nightglow is the explanation of the large annual variation of the atomic sodium abundance. Photochemical equilibrium processes and sodium ionization have been found to be completely unable to explain the observed annual variation. Shortly after the identification of the sodium nightglow it was suggested that the source of the sodium was extraterrestrial and that it was released into the upper atmosphere by the burnup of meteorites (82,83). Early twilight observations of the sodium abundance showed a sodium scale height of 250 km between 250 and 600 km, thus suggesting an extraterrestrial source of sodium (111). Some investigators have proposed that the sodium abundance is due to a Na solar wind similar to the solar proton flux (92,93).

Observations of the annual sodium abundance variation in the southern hemisphere have shown that the variation is seasonal and six months out of phase with the northern hemisphere abundance variation. This argues against an extraterrestrial source for sodium. Simultaneous observations of the meteor flux/24 hours and of the twilightglow sodium abundance have been made at Christchurch for a one year period. The results of these observations have shown that the sodium abundance does not depend upon the rate of meteor influx (155). Calculations of the sodium influx of the sodium solar wind have shown this source to be incapable of explaining the sodium abundance and variation (105,278).

Several investigators have suggested that the sodium is of marine origin. Observations of the lithium and potassium twilightglow have been made and the lithium and potassium abundances deduced from these observations have been compared with the observed sodium abundances. Donahue has found that the observed Li/Na ratio is between the Li/Na ratio found in meteorites and in the ocean (279). The observations of the Li abundance are quite possibly contaminated by lithium injected during nuclear tests which took place shortly before the Li twilight observations. The Na/K abundance radio has been found to be almost the same as the ratio of these two compounds in seawater thus supporting a marine origin for the sodium (125,280).

Junge has presented an argument against a marine origin for sodium (278). He has found that if sea salt particles are the sodium source, then the very size of the particles prevents them from being carried in sufficient quantity to altitudes above 70 km where the sodium atoms become detached from the particle and thus produce changes in the sodium abundance. Junge's calculations show that not even the smallest salt particles (0.1μ radius) will be carried above 50 km. He further found that if the Na "evaporation" from the salt particles is at or below 30 km then the salt particles can supply sufficient sodium by molecular dispersed NaO and NaO_2, but he considered "evaporation" of sodium from the salt particles to be extremely unlikely below 70 km.

Junge calculated the sodium flux due to extraterrestrial sources and found that only meteors could supply the observed sodium abundance. In fact he found that the annual sodium flux from meteors was 30 times higher than the observed winter sodium abundances (278). A factor favoring meteoric origin of the sodium is that the meteors burn up principally between 80 and 100 km, at just the altitudes where the maximum free sodium concentration is found.

Kalkstein (281) has found that the residence time above 20 km for radioactive debris injected into the atmosphere at 43 km with the major portion of the debris rising to about 100 km as a result of the nuclear blast is about 10 years. It would be expected that sodium entering the atmosphere from meteor burnup between 80 and 100 km would have a similar if not longer residence time since the meteor burnup would create finer, lighter particles than produced by atomic testing. This implies that the greater part of the sodium in the upper atmosphere exists in a combined form and that the total amount of free and combined sodium far exceeds the amount of free sodium found in the night or twilight airglow. Using the sodium influx rate due to meteors and a residence time of 10 years above 20 km it is seen that the total amount of sodium is more than two orders of magnitude greater than the amounts of free sodium observed in the airglow. This undoubtedly accounts for the observed independence of free sodium abundance upon meteor flux rate. In conclusion it appears that the sodium is probably of meteoric origin and that only a small fraction of the sodium exists in the free neutral state. The potassium abundance observed in the airglow strongly detracts from the conclusion as to the sodium origin.

E. General Circulation and Seasonal Airglow Variations

The last remaining topic to be covered is the relations between the seasonal airglow variations and the upper atmospheric circulation. The seasonal variations of observed upper atmospheric parameters will be presented first followed by a discussion of possible relations of these circulation features with each of the airglow components investigated in this study.

The temperature structure and variations in the upper atmosphere have been investigated by means of rocket soundings. The principal temperature measurements have been made at Guam (13.5°N)(282), White Sands, N. M. (32°N)(283,284), and Churchill (59°N)(70,285,286). At Guam nine firings were conducted during Nov. 1958. The temperature was found to decrease steadily between 50 and 80 km with a minimum temperature at the upper limit of the observations (80 km) of 190°K. At White Sands 12 rocket measurements were conducted which showed no seasonal changes and only a small diurnal variation in the temperature between 50 and 80 km. The lapse rate was fairly steady at about 2.5°K/km, and a minimum temperature of 200°K was found at 80 km (the upper limit of the measurements). At Churchill 10 rocket firings were made which showed a strong seasonal variation in the temperature above 50 km. In the summer the temperature decrease above the mesopeak (50 to 55 km) was quite steady and minimum temperatures of 160 to 170°K were found at the mesopause (80 to 85 km). The winter measurements at Churchill all showed strong secondary peaks in the temperature (250 to 290°K) between 70 and 80 km. The temperatures at the mesopause (85 km) were 220 to 225°K which is 40 to 60°K warmer than the summer temperatures at 85 km. In all cases where data were available a rapid increase of temperature with height was found above the mesopause.

The principal wind measurements above 80 km have been made through observations of meteor trails (266,267). Below 80 km a variety of techniques have been employed. Summaries of all the known wind observations have been made by Murgatroyd (276) and by MacDonald (287). Between 30 and 80 km the wind is from the east during the summer with average maximum velocities of 75 m/sec at 60 km in middle latitudes. During the winter in this same region the wind is from the west with maximum average speeds of 95m/sec in the core at 55 km between 40 and 45° latitude. A secondary maximum of somewhat doubtful existence of 64 m/sec is found at 50 km and 65° latitude (287). The winds between 80 and 100 km undergo large seasonal variations. In middle latitudes the wind is from the west during the summer and winter but lighter winds from the east prevail at times near the equinoxes. Above 80 km the meridional component of the circulation becomes fairly strong.

The OH intensity reaches maximum values in the late fall and early winter, maintains relatively high values during the winter, and has a minimum in the late summer. The average OH intensity seems to increase with latitude during the winter and stations in middle latitudes appear to have a larger annual intensity variation than locations at lower latitudes.

Wallace (24) has suggested that the high wintertime OH emission intensity could be maintained by downward mixing of atomic oxygen from the high atomic oxygen concentrations found above 100 km. Vertical atomic oxygen transport could result in atomic oxygen concentrations five times greater than that given by photochemical equilibrium used in previous calculations at 85 to 90 km (24). This increased O concentration will give higher emission rates at 85 to 90 km. Continuing downward O transport will maintain an emission peak at these altitudes and thus be more in line with the emission altitudes determined by rocket observations. The downward mixing of atomic oxygen would give an OH emission which would have intensities independent of temperature as is normally observed at middle and low latitudes (24).

The observed winter OH emission rate has been found to require 7×10^{18} atoms of O/cm^2 column for 100 days of emission during the arctic winter (24). The amount of photochemically dissociated atomic oxygen above 80 km is between 10^{18} and 10^{19} atoms/cm^2 column. This is inadequate to maintain the high emission rate throughout the entire winter. The depletion of atomic oxygen would gradually slow down the emission rate until it would be far below the observed rate and thus not duplicate the winter maximum. In order to have continued high OH nightglow intensity throughout the winter, atomic oxygen must be mixed poleward and downward from the tropical regions where solar radiation is still acting to dissociate O_2 into O.

The poleward and downward transport of O has been used by Kellogg to explain the winter heating of the arctic atmosphere (289). The recombination of atomic oxygen below 90 km would supply the energy to maintain the high winter temperatures. It has been found that the energy released by O recombination will primarily go into heating of the atmosphere (290). It follows that the ambient temperature will depend upon the atomic oxygen concentration. If the OH emission rate is dependent upon the atomic oxygen concentration also, then the observed high latitude intensity-temperature correlation is explained by the fact that both the intensity and the temperature depend upon the same factor (atomic oxygen concentration) instead of the intensity variation being a direct result of the temperature variation through the reaction rate coefficient. The winter temperature observations at Churchill show possible evidence of subsidence which supports the Kellogg proposal.

Radioactive tracer experiments using Rh-102 injected at high altitudes during atomic tests have been interpreted as showing that increased vertical mixing occurs at higher latitudes between 100 km and the lower stratosphere during the winter (281). The tracer experiments can also be interpreted as showing that in the region near 100 km fairly rapid poleward horizontal diffusion takes place by both molecular and eddy processes. At 100 km the characteristic time for the recombination of atomic oxygen is on the order of about two weeks and longer above 100 km. Thus poleward mixing of atomic oxygen is possible above 100 km.

The vertical mixing between the lower stratosphere and 100 km observed at higher latitudes during the winter could also transport O_3 and H up into the emission layer as well as bringing O down into the emission layer and thus furnish an additional mode of increasing the OH emission intensity. During the summer there are no observations of the OH nightglow at high latitudes. The sodium airglow seems to have a double intensity maximum in both the fall and spring with high intensity during the winter and a minimum of intensity in the early summer. Several investigators have suggested that the seasonal sodium airglow variations are due to horizontal and vertical transport of sodium (117,119,142,155). Transport of neutral

Na from over the winter pole into middle latitudes has been suggested as the cause for the high winter abundances, but then in this case the winter sodium would be highest over the poles. Observations have shown that the winter sodium abundance is higher in middle latitudes, thus opposing the argument for equatorward transport of sodium.

It has been pointed out above that most of the atmospheric sodium probably exists in a combined form such as NaO and NaO_2. Sodium will be almost entirely in the combined form below 80 km. The changeover of the mesospheric circulation from summer to winter and back again will undoubtedly result in a great deal of vertical mixing. The winds observed between 80 and 100 km show a direction reversal during the fall and the spring while the changeover at lower altitudes is taking place. Observations of meteor winds have shown that the maximum turbulence at 90 km (the altitude of the sodium airglow) occurs at the time of the equinoxes and minimum turbulence is observed during the summer and winter (291). The increased vertical mixing that occurs during the spring and fall could bring excess NaO and NaO_2 from lower altitudes and also bring down increased amounts of O from higher altitudes into the emission region thus producing the observed enhancements of intensity during these periods. During the winter continued downward mixing of O from above 100 km will help maintain increased emission intensities.

The atomic oxygen 5577Å nightglow exhibits a definite autumn maximum and a winter minimum of intensity at all stations. The seasonal variations during the remainder of the year seem to be dependent upon latitude. Below 35° latitude there are maxima in the spring and fall and minima during the summer and winter. Above 35° the maximum intensity occurs in the fall and a prolonged minimum is found in the winter and spring. The 5577Å emission also seems to vary with latitude, exhibiting an intensity minimum near the equator, a maximum at about 30° latitude, a secondary minimum between 35° and 40°, and increasing intensity poleward of 40°.

The seasonal and latitudinal variations of the 5577Å airglow are difficult to interpret on the basis of any simple circulation model. The fact that the emitting layer is principally above 95 km may in some way effect the intensity variations. The 5577Å intensity variation equatorward of 35° can be interpreted as due to increased vertical mixing in the fall and spring accompanying the changeover of the zonal circulation below 80 km. This fits the intensity variation found for the sodium nightglow in middle latitudes.

Above 35° the 5577Å nightglow exhibits a different annual variation pattern. The annual variation of the sodium airglow at Saskatoon (52°N) also differs from the sodium variation at lower latitudes. Here the maximum Na intensities are found in the spring and the minimum in the fall in opposite phase to the 5577Å variation above 35° latitude. This 180° phase difference in the intensity variation is not contradictory. The fact that two emissions show a different variation pattern at higher and lower latitudes further supports the possibility of a change in circulation pattern with latitude. Differences in the altitude of emission may be responsible for the difference in the latitude of the change of the emission pattern since the changeover occurs at about 35°N for the 5577Å airglow and between 45° and 50°

for the Na airglow. Thus the same differences in a circulation pattern can account for the variations of both the sodium and atomic oxygen airglow.

CONCLUSIONS

The emission of the hydroxyl bands, the sodium D doublet, and the atomic oxygen green line in the airglow is affected by dynamic processes in the upper atmosphere. The patchiness of these emissions appears to be due to turbulence in the upper atmosphere. Upper atmospheric tides seem to be directly related to the diurnal variations of intensity of the three airglow emissions studies.

The sodium in the upper atmosphere is most likely of meteoric origin. The meteor influx rate and the upper atmospheric storage times lead to the conclusion that the total amount of sodium in the upper atmosphere is much greater than the amount of free neutral sodium deduced from twilightglow and nightglow observations. A large percentage of the sodium probably exists in the form of sodium oxide (NaO) and sodium dioxide (NaO_2).

The temperature of the OH emission layer (80 to 85 km) as deduced from measurements of the OH vibration-rotation band structure is in equilibrium with and equivalent to the ambient temperature. The seasonal and latitudinal variations of the OH rotational temperature fit in remarkably well with the rocket temperature measurements at the altitude of the OH layer. Simultaneous rocket measurements of the upper atmospheric temperature and ground measurements of the OH rotational temperature and height should be made so as to determine if the OH nightglow observations can be used directly as a means of determining the upper atmospheric temperature.

It appears that the seasonal intensity variations of the sodium and atomic oxygen nightglow can be fitted into a single circulation scheme. Both emissions exhibit intensity maxima in the spring and fall and intensity minima in the summer and winter at lower latitudes. This is in good agreement with the observed variation of the turbulence intensity at the altitudes of these emissions. At higher latitudes the annual variation of the intensity of these two emissions is different, indicating a change in the circulation at middle latitudes. At 100 km this changeover appears to take place at about 35° latitude based upon the atomic oxygen 5577Å nightglow variations and at 90 km the changeover seems to occur between 45° and 50° latitude based upon the sodium airglow variations. Increased sodium emission intensity is probably due to upward mixing of part of the large atmospheric store of sodium oxides. The enhancement of the 5577Å emission results from increased downward mixing of atomic oxygen to altitudes where the recombination of the oxygen atoms is more rapid.

The intensity variation of the OH emission with latitude and season does not appear to readily fit into the same circulation pattern indicated by the 5577Å and 5893Å emissions. Enhancement of the OH emission by downward mixing of atomic oxygen so as to increase the rate of emission would require a simultaneous increase of 5577Å emission intensity. It has been shown that there is almost no correlation between the OH and 5577Å emissions which argues strongly against the enhancement of OH emission by increased downward mixing of atomic oxygen. The only recourse seems to be that the OH emission must be enhanced by upward mixing of some atmospheric constituent(s) from lower altitudes. Included in the list of those substances which could possibly increase the OH intensity by increased upward mixing are O_3, H and HO_2. The production of OH^* by reactions with HO_2 should be more thoroughly investigated.

The fact that the OH emission occurs principally below the mesopause may be the reason for the difference between the variations of intensity of OH and Na and O (5577Å). The observations of the OH intensity are relatively few in number as compared to the other two emissions. It may well be that the OH intensity does exhibit both fall and spring maxima but that the number, timing and quality of the observations have failed to show the actual variation. In any case frequent observations of the OH nightglow intensity and temperature for a period of not less than three years in length should be made within each 5° latitude belt between 25° and 60° latitude in order to firmly establish the true variations of the OH nightglow. A similar set of observations of the sodium and green atomic oxygen nightglow would also be useful.

Simultaneous observation of meteor winds and the intensities of the various airglow components would also be useful in the study of the upper atmospheric motions. These observations could be made at 25°, 35°, 45° and 55° latitude. The short and long term correlations between the variations in the winds and the airglow intensities may give considerable insight into the dynamical processes in the upper atmosphere in addition to furnishing a proportionately large amount of observational evidence.

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