The Effect of Solar Corpuscular Radiation on the 1963 Final Spring Warming in the Antarctic

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ABSTRACT

The 1963 final spring warming in the antarctic stratosphere is discussed, with particular reference to the energy flux of electrons precipitated during the event. Temperature changes occurring throughout the atmosphere at this time are estimated. The change in thermospheric temperature (~110 km) due to particle influx is shown to be approximately 15K, when allowance is made for losses due to molecular conductivity, eddy transport and radiation. It is shown that this heating could lead to a greater deposition of gravity wave energy near 110 km in the auroral zone, and a further increase in local temperature. The resulting changes in the mid-latitude zonal wind above 80 km would lead to a modification in the large-scale wave activity at these heights. Although these mechanisms do not appear to constitute an initial triggering mechanism for the planetary wave which was associated with the stratospheric warming of 1963, they could lead to a correlation between the particle influx, and the mid-latitude stratospheric and thermospheric parameters.

Considerations of the photochemistry of ozone, as it is presently understood, suggest that the auroral emissions were too weak to introduce a mesospheric-stratospheric temperature perturbation capable of destabilizing the dynamic structure of that region. A more detailed knowledge of the photochemistry of ozone at heights up to 110 km, during particle influx events, should lead to a better understanding of the dynamics of this region. Finally, it is noted that there is insufficient atmospheric data to confirm the hypothesis that auroras or particle influx events are responsible for the onset of stratospheric warmings.

1. Introduction

A suggestion has been made by Willet (1968) that "solar corpuscular penetration" of the higher atmosphere, rather than an advective-dynamic process, may be responsible for the seasonal changes of temperature and ozone in the arctic and antarctic stratosphere. This possibility warrants discussion. The availability of satellite data on the energy flux of precipitated electrons permits a study of the 1963 final spring warming in the antarctic as a particular example of a rapid seasonal change having hemispheric dimensions. Meteorological data for this period are available from a descriptive study by Phillpot (1964). The major objective in this paper is to attempt to answer the question raised by Willet, as to how "the effective transmission of this (corpuscular) energy from the upper atmosphere where it is directly received downward to the lower atmosphere is accomplished."

Specific aspects of the study of particle influx will include: 1) the change in thermospheric temperature (~110 km) due to corpuscular heating and to cooling by molecular conductivity, eddy transport and radiation; 2) the effect of this change in temperature upon the propagation conditions for internal gravity waves, thermal tides and planetary waves; and 3) the effect of auroral emissions upon the temperature and dynamic stability of the mesosphere and stratosphere. A triggering mechanism for the stratospheric warmings is discussed.

2. The final spring warming of 1963

The period from September–December 1963 has been extensively studied by Phillpot (1964). From daily 100-mb temperatures for approximately 20 stations he showed that the major warmings, or departures from the average seasonal trend, occurred during the first part of November, and apparently progressed eastward around the Antarctic shoreline. Large regions where spring warmings (stratwarms) occurred are shown in Fig. 1, with dates of occurrence. These warmings mark the establishment of the summer easterly winds in the mid-latitude and polar stratosphere. The 1963 spring warming was comparable in magnitude to the warmings of 1957 and 1958 (Willet, 1968). At Wilkes, near W_2, a warming trend beginning on 26 October led to 100-mb temperatures which were 20C above the average seasonal curve by 1 November. A second spring warming which began 7 November led to a temperature of −35C on 9 November. December temperatures were 5–10C lower. The departure from the average seasonal curve began at Hallett and McMurdo on 1 November and temperatures of −25 and −30C, respectively, were reached by 17 November. These temperatures were 30C above the average curve and were

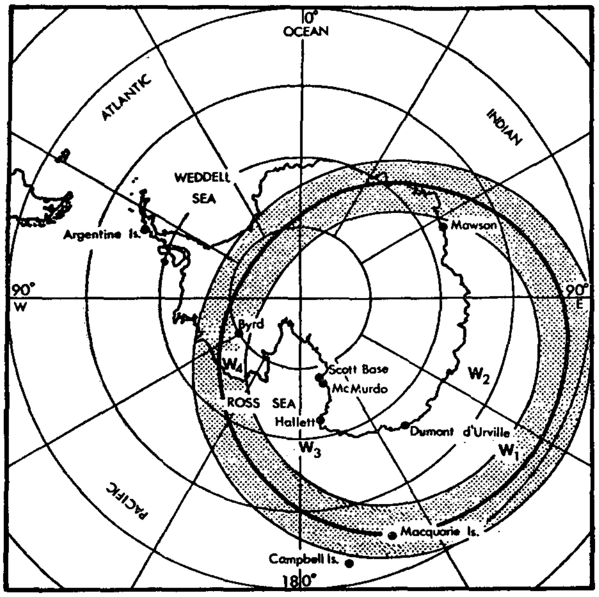

FIG. 1. The Antarctic continent, with the auroral zone defined by the invariant latitude of 67.5S. Regions where 100-mb warmings were first observed in 1963 are shown: W_1, 30 October; W_2, 10 November; W_3, 20 November; W_4, 30 November.

higher than the December values. Increases of 20C in four days during this period testify to the explosive nature of the warming.

From an analysis of tropospheric and stratospheric circulation during 1964, Zhdanov (1965) concluded that stratwarms first appeared between Mawson and Dumont d'Urville, where the tropospheric cyclones from lower latitudes reach the coast of Antarctica with the greatest frequency. Such a major feature of the Southern Hemisphere circulation is probably reproduced in broad detail from year to year and may be assumed to apply to 1963 also. Certainly the 1964 stratwarms tended to move eastward, as was noted the previous year.

Studies by Godson (1963) and Phillpot (1964) have shown that the antarctic circumpolar vortex has great strength and symmetry. Thus, although the vortex may break down in spring and form two centers during stratwarms (bipolarity), this does not appear to occur as often as in the Northern Hemisphere, where major disturbances of the vortex occur during midwinter warmings. The orographic properties of the antarctic land mass are probably responsible for the predominance of the asymmetric ($n=1$) planetary wave mode during the spring warmings. It is clear from the pressure and wind charts, however, that both the asymmetric ($n=1$) and bipolar ($n=2$) planetary waves are an important feature of the antarctic stratwarms (Phillpot, 1964). These stratwarms appear to be similar in terms of descriptive meteorology to the major stratwarms of the Northern Hemisphere, where significant alterations to the energy balance of the atmosphere occur simultaneously with the growth and decay of the planetary wave modes $n=1$, 2 (Julian and Labitzke, 1965; Perry, 1967). A visual inspection of the pressure charts shows that the asymmetric wave mode predominated during the antarctic final spring warming of 1963.

3. Corpuscular radiation during the 1963 event

At the time of the 1963 stratwarm, precipitated electron flux measurements were being made by means of a polar orbiting satellite (Sharp and Johnson, 1968). It is therefore possible to estimate the corpuscular energy input to the atmosphere during this period, and discuss its effect upon the energy balance or dynamic stability of the polar vortex.

The position of the auroral zone, at an average invariant latitude of 67.5°, is shown in Fig. 1. Calculations of invariant latitude by Campbell and Matsushita (1967) were used. The effective width of the zone has been defined using the average time during which particle precipitation was recorded by the satellite. The satellite crossed the auroral zones every 90 min at approximately 0100 and 1300 hours local time. Data were obtained for a 5-day period from 30 October to 3 November 1963. Several low energy particle detectors ($\geq$0.080 keV) were used by Sharp and Johnson to determine the total intensity of electron precipitation for each traverse of the auroral zone. The electron energy flux density was then available.

During the period of satellite observations moderate magnetic activity was recorded at ground stations, with Kp values varying from 0_+ to 8_0. Satellite data were obtained on one of the two days (29, 30 October) which were classified as geomagnetically "disturbed" in the CRPL Bulletin, CRPL-F 233 Part B. Apart from two periods of more intense magnetic activity (11–16 October, 6–12 November) these Kp values were typical of the magnetic conditions prevailing during the months of October and November and did not include any "quiet" day values. However, it is not possible to associate the measured electron energy flux density with that occurring during October and November, because the relationship between low energy particle precipitation (0.080–10 keV) and Kp is not known with sufficient accuracy.

The mean energy flux input for the observing period may be estimated as follows. During the southern nights flux densities of about 1 erg cm^{-2} sec^{-1} were most common, while during the days (where only Northern Hemisphere data are available) the flux densities were more nearly 0.25 erg cm^{-2} sec^{-1}. A mean flux density of 0.62 erg cm^{-2} sec^{-1} will be assumed. The mean energy flux within the previously defined auroral zone is then 1.6×10^{17} ergs sec^{-1} for the observing period. It is of interest to compare this value with previous estimates of precipitated electron energy flux. Akasofu (1964) suggested an energy deposition of 1.4×10^{21} ergs in 12 hr, during an auroral substorm, corresponding to a flux of 8×10^{16} ergs sec^{-1}. Considering the scarcity of available flux measurements, and the many assumptions which were made, Akasofu's estimate is in good agreement with the satellite measurements.

It is also necessary to estimate the average energy and spectral characteristics of the precipitated particles, so that the energy deposited at various depths in the atmosphere can be calculated. Taking the mean particle energy as 5 keV (Akasofu, 1964), a flux density of 1 erg cm^{-2} sec^{-1} corresponds to an electron flux of 10^8 cm^{-2} sec^{-1} for nighttime conditions. This flux and mean energy compares favorably with values obtained by rocket probes and satellites passing through particular auroras (Sharp *et al.*, 1964; McIlwain, 1960). The characteristic energy E_0 of an exponential flux shape $\{n(E) \propto \exp[-(E/E_0)]\}$ has been taken from Sharp and Johnson, for energies <30 keV. For energies >30 keV a value of $E_0=25$ keV has been used, since energy spectra are often characterized by two slopes (O'Brien, 1962). The use of a larger characteristic energy also places an upper limit on the fluxes at the lowest altitudes. Penetration depths for electrons of various energies have been taken from Whitten and Poppoff (1965). The fluxes, calculated for three heights in the mesosphere, are shown in Table 1.

TABLE 1. Auroral zone electron flux densities, October/November 1963.

Height (km)	Energy (keV)	Flux (cm^{-2} sec^{-1}) Night	Day
110	5	10^8	2.5×10^7
85	50	10^6	4×10^6
60	500	$<10^0$	$<10^0$

4. Heating effects

The first major indication of the 1963 antarctic stratwarm occurred on 30 October, in a region within the auroral zone (Fig. 1). An immediate question arises as to whether the electron flux densities measured during this period could be responsible for any direct heating of the atmosphere—and if so, which height ranges would be most affected.

There have been various estimates of the efficiency of conversion of kinetic energy of incident particles into heat. Chamberlain (1961) has suggested 15%, while Dalgarno (1964) estimates it to be 20%, with an upper limit of 50%. The value of 20% will be used here. Assuming that the energy is shared equally among the particles of the atmosphere above 105 km, they would gain energy at the rate of 3×10^{-8} eV sec^{-1}. Thus, if there were no energy losses, the temperature near 110 km would rise 30K from the unperturbed value of 240K, in one day. The unperturbed densities and temperatures have been taken from the 1961 COSPAR International Reference Atmosphere (CIRA). However, losses due to diffusion conductivity and radiation are large, and must be considered. The energy balance of the region under consideration—"the lower thermosphere," between the mesopause and 130 km—is not well understood and any calculation of the magnitudes of the different loss terms must of necessity be uncertain. It is again assumed that the kinetic energy of the incident particles is shared equally among the particles of the atmosphere above 105 km. This is a useful assumption when the temperature of the lower thermosphere is required relative to that of the mesosphere.

The vertical flux F_c of heat due to conduction is given by Nicolet (1960) as

$$F_c = -AT^{\frac{1}{2}}\frac{\partial T}{\partial z} \quad [\text{ergs cm}^{-2}\text{ sec}^{-1}], \qquad (1)$$

where A (atomic oxygen) $=3.6\times10^2$ ergs cm^{-1} sec^{-1} $(°K)^{-\frac{3}{2}}$, $A(O_2,N_2)=1.8\times10^2$ ergs cm^{-1} sec^{-1} $(°K)^{-\frac{3}{2}}$, and $\partial T/\partial z$ is the vertical gradient of temperature. Assuming equal concentration of atoms and molecules near 110 km, and using number densities and temperature gradients from the CIRA, equilibrium would be reached in two days. The perturbed temperature would then be 30K above the reference value of 240K.

At these low heights in the thermosphere, where the density is still quite large, eddy diffusion processes are also likely to be important. Energy loss due to eddy transport is given by

$$F_e = -\rho C_P K_T\left(\frac{\partial T}{\partial z}+\Gamma_d\right) \quad [\text{cal cm}^{-2}\text{ sec}^{-1}], \qquad (2)$$

where ρ is the atmospheric density, C_P the specific heat at constant pressure, K_T the thermal diffusivity, and Γ_d the adiabatic lapse rate. The values of K_T are discussed in a review by Curnow (1966), who presents values ranging from 10^6–10^7 cm^2 sec^{-1}. Difficulties in obtaining compatibility with other turbulence parameters and the thermospheric energy balance lead to this large variation in K_T. More recent calculations by Colegrove *et al.* (1965), and measurements by Justus (1967), suggest that a value of 3×10^6 cm^2 sec^{-1} near 110 km is likely. Since this parameter is known to be variable, a range of values from 1–5×10^6 cm^2 sec^{-1} will also be used in calculations. Combining the energy losses due to conductivity and eddy transport, equilibrium would be reached in 1 day at a temperature approximately 15K above that of the CIRA. The range of K_T values would produce temperature perturbations of 12–22K near 110 km.

The radiation losses are very difficult to estimate quantitatively, due to uncertainty of the concentration of various constituents from 90–130 km. Processes likely to be significant here are the radiative cooling by CO_2, transitions between the different levels of the ground term ^3P of atomic oxygen, and various air-glow emissions (Craig, 1965). Of these three processes, it is believed that the atomic oxygen transitions are most significant near 110 km. Bates (1951) developed an expression for this loss H_R, of the form

$$H_R = \frac{1.67\times10^{-18} n(\mathrm{O}) \exp(-228/T)}{1+0.6\exp(-228/T)+0.2\exp(-325/T)} \quad [\mathrm{ergs\ cm^{-3}\ sec^{-1}}], \quad (3)$$

the function not being strongly dependent on temperature in this height range. Energy losses for the comparatively small temperature perturbations involved here are therefore likely to be small. Using typical $n(\mathrm{O})$ values (Colegrove *et al.*, 1965), the energy loss is estimated to be less than 0.02 erg cm^{-2} sec^{-1}, which is much less than that due to molecular conductivity or eddy transport. The fractional change in $n(\mathrm{O})$ due to particle influx is presently thought to be small near 110 km (Maeda and Aikin, 1968), and hence would not modify this estimate significantly. For the purposes of this discussion it is therefore proposed to ignore radiation losses. This is due to necessity rather than to choice, but because of the uncertainty of concentrations in the 90–130 km region, there seems no alternative. It is believed, however, that the most important process, due to transitions in the ground term of atomic oxygen, has been considered adequately.

It is now necessary to consider the heating effects at lower heights in the atmosphere. An electron flux having the spectral characteristics discussed in Section 3 is most effective in increasing the temperature near 110 km. Below this height, decreasing flux density (Table 1) and increasing atmospheric density result in much smaller temperature perturbations. For example, at 95 km the temperature would be increased by less than 1K. There are two additional sources of energy at these lower heights, however. In the first case, heat from near 110 km would be transported downward by molecular conduction. Secondly, the downward transport of any additional atomic oxygen produced by electron precipitation would be effective in warming the lower region, due to the release of chemical energy (Kellogg, 1961). Unfortunately, uncertainties as to the atomic oxygen concentration and the magnitude of the vertical motions prevent a quantitative estimate from being made. This remains the major uncertainty in these calculations. Considering only the direct heating by the precipitated electrons, and the heating by molecular conductivity, the temperature perturbation near 95 km would be approximately 2K.

Summarizing, the overall heating effect of the particle influx measured from 30 October to 3 November would have been to increase the temperature gradient above the mesopause ($\sim$85 km). Temperature increases of 2K at 95 km and 15K at 110 km have been calculated. It has not been possible to include the heating due to the downward transport of any additional atomic oxygen formed during the particle influx.

5. The dynamics of the atmosphere: Effect of thermospheric temperature change

The direct heating previously discussed seems limited to heights above 85 km. In contrast, there is evidence to suggest that major stratospheric warmings, at least of the Northern Hemisphere's midwinter variety, are most intense near 45 km, and reverse or modify the stratospheric circumpolar vortex throughout its entire depth (Finger and Teweles, 1964). It is important to note, however, that modifications to the lower thermospheric temperature will affect the upward propagation of atmospheric waves of all scales, from internal gravity waves to planetary waves. Such waves are important to the energy balance of the atmosphere throughout its depth. A modification in wave propagation conditions is a possible means of coupling the atmosphere below 45 km with the perturbed lower thermosphere. For a comprehensive study the wind and temperature fields from the troposphere to near 110 km should be known, and the interactions of the various wave scales introduced. Such information is unavailable, and hence the propagation of each of the three scales of wave motions (gravity, tidal and planetary) will be discussed separately, with reference to existing analyses.

a. *Gravity waves*

The propagation of gravity waves to ionospheric heights, in the presence of various background winds and temperatures, has been analyzed by Hines and Reddy (1967). They showed, apart from directional filtering due to the wind profile, that modes with phase speeds $\geq$50 m sec^{-1} could be expected to reach 85 km (in average winter conditions) with little attenuation. Above this height the thermospheric temperature gradient introduces significant reflection of all modes whose phase speed is below 200 m sec^{-1}. Thus, an increase in temperature gradient (as was introduced in Sec. 4) could be expected to produce an increase in the absorption of gravity wave energy low in the thermosphere ($\sim$105–110 km). Hines (1965) has shown that gravity waves could comprise a significant source of thermospheric heating. Hence, a cascading process could develop which would lead to a further increase in auroral zone temperatures near 110 km, and an even sharper temperature gradient above the mesopause.

A second consideration is related to the source of gravity wave energy, low in the atmosphere. Gossard (1962) has estimated that the upward flux of gravity wave energy from the troposphere and lower stratosphere may reach 10^2 ergs cm^{-2} sec^{-1}, under favorable conditions (temperature inversions, frontal systems). Such a flux is very small with respect to a flux which would significantly reduce the energy density of the troposphere (10^3 ergs cm^{-3}), and hence it would not change the energy balance of the troposphere significantly. However, a feature of the development of midwinter stratwarms is the occurrence of strong tropospheric jet streams (Labitzke, 1965; Labitzke and van Loon, 1965). Such jet streams are likely to be an effective source of gravity waves. Hence, during a stratwarm an increased flux of gravity waves could be expected, which would constitute a larger source of thermospheric heating.

No causal relationship between particle influx events and stratwarms is evident from the above discussion. There is an indirect association, however, in that both types of event could lead to an effective coupling between the lower atmosphere below 45 km and the lower thermosphere. Correlations between the temperatures or winds of the troposphere and stratosphere, and temperatures, winds or ionization in the lower thermosphere would then be expected. Unfortunately, adequate measurements of these parameters do not exist for the 1963 antarctic event or any other major stratwarm.

b. Tides

The upward energy flux of the thermal diurnal and semidiurnal tides amount to several ergs cm^{-2} sec^{-1} at the equator (Lindzen, 1967a, 1968) but the trapping of the tidal modes at the higher latitudes of the auroral zone reduces the upward flux considerably. Using the latitudinal variation of the horizontal and vertical perturbation components shown by Lindzen (1967a), the 85-km diurnal tidal energy at Macquarie Island (55S) is 10^{-3} erg cm^{-2} sec^{-1}. Such a flux is several orders of magnitude less than the predicted gravity wave contribution (Hines, 1965). For this reason, tides do not appear to offer a strong coupling mechanism between the lower atmosphere and the perturbed thermosphere at high latitudes.

c. Planetary waves

Finally, the upward propagation of large-scale cyclonic and planetary waves must be considered. The effective refractive index for various seasonal wind and temperature profiles, and for wavelengths from 6000–14,000 km, have been calculated by Charney and Drazin (1961). Although the transmission coefficients are very small ($T \simeq 10^{-3}$) even for these large modes, the large tropospheric kinetic energy density ($\simeq 10^3$ ergs cm^{-3}) would allow an upper limit of 1 erg cm^{-2} sec^{-1} to reach heights near 90 km (Hines, 1963). Again, the study by Teweles (1965) shows that at least 10% of this tropospheric energy can be expected in the larger wave modes which have transmission coefficients of this order, so that fluxes of 0.1–0.25 erg cm^{-2} sec^{-1} can realistically be expected at 90 km. This is comparable to the gravity wave flux predicted by Hines (1965).

The most important factor governing the amplitude of planetary waves as a function of height is the zonal wind profile above 30 km (Charney and Drazin, Figs. 4, 7). Thus, a moderate westerly wind (10–30 m sec^{-1}) can result in a Doppler shift of the wave frequency to that of a freely propagating mode (Lindzen, 1967b). From the thermal wind equation, the thermospheric warming discussed in Section 4 would give rise to progressively weaker westerly winds above 80 km and eventually to strong easterly winds above 100 km in mid-latitudes. These conclusions are supported by the experimental results of Smith (1968). Thus, at latitudes (30–45S) equatorward of the auroral zone there would be a modification in the large-scale wave activity above 80 km during particle influx events; and if a planetary wave is capable of perturbing the electron densities of the lower ionosphere from 65–90 km by redistribution of ionizable constituents (Gregory, 1965; Geisler and Dickinson, 1968) or by temperature changes (Sechrist, 1967), the above mechanism could provide a coupling between particle influx events and/or magnetic storms in the auroral zone and perturbations in the mesosphere and lower ionosphere at mid-latitudes. This hypothesis is in sympathy with a recent study by Belrose and Thomas (1968) of changes in the middle latitude D-region associated with geomagnetic storms.

In conclusion, it will be noted that two coupling mechanisms between the troposphere and stratosphere and the perturbed thermosphere of 1963 have been suggested above. The first involved additional thermospheric heating by gravity waves at auroral latitudes, and the second, changes in the upward transmission of large-scale planetary waves at mid-latitudes. Measurements of the relevant parameters are required to clarify the processes involved. However, neither of these mechanisms appear to constitute a triggering mechanism for the initial development of the 1963 spring warming, or of the planetary waves which appear to be an inherent part of stratwarms in both hemispheres. The coupling between particle influx events and stratwarms could therefore be dependent on the chance occurrence of a suitable stratospheric planetary wave in the lower atmosphere. This is a reasonable hypothesis, since stratwarms have occurred during magnetically quiet periods, or when no major auroral activity was present.

6. Atmospheric emissions and the stability of the mesosphere

The remaining 80% of the mean precipitated electron energy flux ($\simeq 0.5$ erg cm^{-2} sec^{-1}) for the period from 30 October to 3 November, would be involved in the ionization and excitation of various molecular and atomic species. This will result in weak auroral emissions, and thermal energy in addition to that considered in Section 3. The effect of these emissions upon the temperature and dynamical stability of the mesosphere and stratosphere can be estimated by reference to analyses by Lindzen and Goody (1965) and Lindzen (1965). It transpires that variations in the solar ultraviolet much in excess of the total energy flux of emission from even an aurora of magnitude IBC III is required to give rise to significant temperature fluctuations near

30 km. Hence, the above analysis does not provide an immediate stratospheric triggering mechanism for a large aurora, or for the 1963 particle influx event considered here.

Lindzen (1966) has also shown that the stratosphere-mesosphere is unstable with respect to large baroclinic waves, which are similar to those which have been observed during Northern Hemisphere stratospheric warmings. A relatively simple model of the photochemistry of ozone was used in this analysis. In fact, the photochemistry of ozone and of oxygen at all heights up to 110 km is extremely complicated and is the center of considerable research (Willet, 1968; Hunten and McElroy, 1968; Evans *et al.*, 1968). Finally, during auroral activity the photochemistry of ozone is expected to be further modified (Maeda and Aikin, 1968; Isaksen *et al.*, 1968), and with it, the dynamics of the stratosphere and mesosphere.

7. Concluding remarks

The effect of particle influx upon the antarctic atmosphere during the 1963 final spring warming has been considered. No direct causality between these two events has been established, although several correlations between stratospheric and thermospheric parameters have been suggested by the analysis. At this time there appears to be insufficient data to confirm Willet's hypothesis that "solar corpuscular penetration" of the higher atmosphere may be responsible for the seasonal changes of temperature and ozone in the arctic and antarctic stratosphere.

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REFERENCES

Akasofu, S.-I., 1964: A source of the energy for geomagnetic storms and auroras. *Planetary Space Sci.*, **12**, 801–833.

Bates, D. R., 1951: The temperature of the upper atmosphere. *Proc. Phys. Soc.*, **B64**, 805–821.

Belrose, J. S., and L. Thomas, 1968: Ionization changes in the middle latitude D-region associated with geomagnetic storms. *J. Atmos. Terr. Phys.*, **30**, 1397–1413.

Campbell, W. H., and S. Matsushita, 1967: World maps of conjugate coordinates and L contours. *J. Geophys. Res.*, **72**, 3518–3521.

Chamberlain, J. W., 1961: *Physics of the Aurora and Airglow.* New York, Academic Press, 704 pp.

Charney, J. G., and P. G. Drazin, 1961: Propagation of planetary-scale disturbances from the lower into the upper atmosphere. *J. Geophys. Res.*, **66**, 83–109.

Colegrove, F. D., W. B. Hanson and F. S. Johnson, 1965: Eddy diffusion and oxygen transport in the lower thermosphere. *J. Geophys. Res.*, **70**, 4931–4941.

Craig, R. A., 1965: *The Upper Atmosphere.* New York, Academic Press, 509 pp.

Curnow, R. J., 1966: Upper atmosphere turbulence—a review. Tech. Note Pad 118, Weapons Res. Establishment, Salisbury, South Australia, 25 pp.

Dalgarno, A., 1964: Corpuscular radiation in the upper atmosphere. *Ann. Geophys.*, **20**, 65–74.

Evans, W. F. J., D. M. Hunten, E. J. Llewellyn and A. Vallance Jones, 1968: Altitude profile of the infrared atmospheric system of oxygen in the dayglow. *J. Geophys. Res.*, **73**, 2885–2896.

Finger, F. G., and S. Teweles, 1964: The mid-winter 1963 stratospheric warming and circulation change. *J. Appl. Meteor.*, **3**, 1–15.

Geisler, J. E., and R. E. Dickinson, 1968: Vertical motions and nitric oxide in the upper mesosphere. *J. Atmos. Terr. Phys.*, **30**, 1505–1521.

Godson, W. L., 1963: A comparison of middle stratospheric behaviour in the arctic and antarctic, with special reference to final warmings. *Meteor. Abhandl., Inst. Meteor. Geophys., Berlin*, **36**, 161–206.

Gossard, E. E., 1962: Vertical flux of energy into the lower ionosphere from internal gravity waves generated in the troposphere. *J. Geophys. Res.*, **67**, 745–757.

Gregory, J. B., 1965: The influence of atmospheric circulation on mesospheric electron densities in winter. *J. Atmos. Sci.*, **22**, 18–23.

Hines, C. O., 1963: The upper atmosphere in motion. *Quart. J. Roy. Meteor. Soc.*, **89**, 1–42.

——, 1965: Dynamical heating of the upper atmosphere. *J. Geophys. Res.*, **70**, 177–183.

——, and C. A. Reddy, 1967: On the propagation of atmospheric gravity waves through regions of wind shear. *J. Geophys. Res.*, **72**, 1015–1034.

Hunten, D. M., and M. B. McElroy, 1968: Metastable $O_2(^1\Delta)$ as a major source of ions in the D region. *J. Geophys. Res.*, **73**, 2421–2428.

Isaksen, I., E. J. Llewellyn, and A. Vallance Jones, 1968: Photochemical effects in the atmosphere resulting from auroral bombardment. CARDE Contract No. CD 601021, Rept. No. 2, Institute of Space and Atmospheric Studies, University of Saskatchewan, Saskatoon, Canada, 36 pp.

Julian, P. R., and K. B. Labitzke, 1965: A study of atmospheric energetics during the January–February 1963 stratospheric warming. *J. Atmos. Sci.*, **22**, 597–610.

Justus, C. G., 1967: The eddy diffusivities, energy balance parameters, and heating rate of upper atmospheric turbulence. *J. Geophys. Res.*, **72**, 1035–1039.

Kellogg, W. W., 1961: Chemical heating above the polar mesopause in winter. *J. Meteor.*, **18**, 373–381.

Labitzke, K., 1965: On the mutual relation between stratosphere and troposphere during periods of stratospheric warmings in winter. *J. Appl. Meteor.*, **4**, 91–99.

——, and H. van Loon, 1965: A note on stratospheric midwinter warmings in the Southern Hemisphere. *J. Appl. Meteor.*, **4**, 292–295.

Lindzen, R. S., 1965: The radiative-photochemical response of the mesosphere to fluctuations in radiation. *J. Atmos. Sci.*, **22**, 469–478.

——, 1966: Radiative and photochemical processes in mesospheric dynamics: Part IV. Stability of a zonal vortex at midlatitudes to baroclinic waves. *J. Atmos. Sci.*, **23**, 350–359.

——, 1967a: Thermally driven diurnal tide in the atmosphere. *Quart. J. Roy. Meteor. Soc.*, **93**, 18–42.

——, 1967b: Planetary waves on beta-planes. *Mon. Wea. Rev.*, **95**, 441–451.

——, 1968: Lower atmospheric energy sources for the upper atmosphere. *Meteor. Monogr.*, **9**, No. 31, 37–46.

——, and R. Goody, 1965: Radiative and photochemical processes in mesospheric dynamics: Part I. Models for radiative and photochemical processes. *J. Atmos. Sci.*, **22**, 341–348.

Maeda, K., and A. C. Aikin, 1968: Variations of polar mesospheric oxygen and ozone during auroral events. *Planetary Space Sci.*, **16**, 371–384.

McIlwain, C. E., 1960: Direct measurements of particles producing visible auroras. *J. Geophys. Res.*, **65**, 2727–2747.

Nicolet, M., 1960: The properties and constitution of the upper atmosphere. *Physics of the Upper Atmosphere*, New York, Academic Press, 17–71.

O'Brien, B. J., 1962: Lifetimes of outer-zone electrons and their precipitation into the atmosphere. *J. Geophys. Res.*, **67**, 3687–3706.

Perry, J. S., 1967: Long-wave energy processes in the 1963 sudden stratospheric warming. *J. Atmos. Sci.*, **24**, 539–550.

Phillpot, H. R., 1964: The springtime accelerated warming phenomenon in the antarctic stratosphere. Tech. Rept. No. 3, Intern. Antarctic Analysis Centre, Commonwealth Bureau of Meteorology, Melbourne, Australia, 35 pp.

Sechrist, C. F., 1967: A theory of the winter absorption anomaly at middle latitudes. *J. Atmos. Terr. Phys.*, **29**, 113–136.

Sharp, R. D., J. E. Evans, W. L. Imhor, R. G. Johnson, J. B. Reagan and R. V. Smith, 1964: Satellite measurements of low energy electrons in the northern auroral zone. *J. Geophys. Res.*, **69**, 2721–2730.

——, and R. G. Johnson, 1968: Some average properties of auroral electron precipitation as determined from satellite observations. *J. Geophys. Res.*, **73**, 969–990.

Smith, L. B., 1968: An observation of strong thermospheric winds during a geomagnetic storm. *J. Geophys. Res.*, **73**, 4959–4963.

Teweles, S., 1965: The energy balance of the stratosphere. Tech. Note No. 70, World Meteorological Organization, Geneva, Switzerland, 107–122.

Whitten, R. C., and I. G. Poppoff, 1965: *Physics of the Lower Ionosphere*. Englewood Cliffs, N. J., Prentice-Hall, 232 pp.

Willet, H. C., 1968: Remarks on the seasonal changes of temperature and of ozone in the arctic and the antarctic stratospheres. *J. Atmos. Sci.*, **25**, 341–360.

Zhdanov, L. A., 1965: The behavior of the antarctic stratospheric warming phenomenon in the spring of 1964. Tech. Rept. No. 4, Intern. Antarctic Analysis Centre, Commonwealth Bureau of Meteorology, Melbourne, Australia, 73 pp.