Variation in the Stratospheric Aerosol Associated with the North Cyclonic Polar Vortex as Measured by the SAM II Satellite Sensor

G. S. Kent
Institute for Atmospheric Optics and Remote Sensing, Hampton, VA 23666

C. R. Trepte
SASC Technologies, Inc., Hampton, VA 23666

U. O. Farrukh
Institute for Atmospheric Optics and Remote Sensing, Hampton, VA 23666

M. P. McCormick
NASA Langley Research Center, Hampton, VA 23665

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ABSTRACT

Aerosol extinction data obtained by the Stratospheric Aerosol Measurement II (SAM II) satellite instrument during the 1979/80 Northern Hemisphere winter season have been analyzed in relation to the cyclonic polar vortex. A synoptic approach has been employed to study the behavior of aerosol extinction ratio and optical depth between altitudes of 8 and 30 km as a tracer of mean atmospheric motions in and near the polar vortex. As the polar vortex intensifies, a gradient of extinction ratio is established across the polar-night jet stream, which is associated with subsidence within the vortex. Maximum subsidence occurs at the center of the vortex. Calculated descent rates relative to isentropic surfaces are of the order of $8 \times 10^{-4}$ m s$^{-1}$ near 20 km, at the center of the vortex between September and December. Below an altitude of 14 km, taken as the base of the vortex, and outside the vortex, horizontal movements occur freely, masking any systematic vertical motions. Extinction enhancements by polar stratospheric clouds and changes produced by sudden warmings in the second half of winter have prevented a similar study for this period. An estimate of the aerosol mass transferred downward through the base of the vortex for the entire season is 7000 tonnes. Comparison of the inferred stratospheric motions with earlier studies using radioactive tracers shows good agreement.

1. Introduction

The SAM II satellite system (Stratospheric Aerosol Measurement II) has been making systematic measurements of aerosol extinction at high latitudes since its launch in October 1978 (McCormick et al., 1979). These measurements have included sightings of polar stratospheric clouds (PSCs) (McCormick et al., 1981, 1982) which occur whenever the stratospheric temperatures fall to very low values below 200 K (Steele et al., 1983). More recently, it has been observed that, whether PSCs are present or not, the aerosol extinction shows strong horizontal gradients which appear related to the polar winter vortex. Measurements made in the arctic region in January 1983, approximately nine months after the eruption of El Chichon, in conjunction with airborne lidar observations, have been published by McCormick et al. (1983). These measurements showed that, at altitudes above 18 km, the inferred aerosol concentration was much less on the cyclonic side of the polar-night jet stream than on the anticyclonic side. In order to see if this phenomenon was a normal characteristic of the polar winter stratosphere and to obtain more information about its nature, one complete winter of SAM II observations has been selected for a detailed study.

The period selected is September 1979 to April 1980, covering the whole period during which the cyclonic polar vortex existed for this winter season. It is also the first complete winter of SAM II data available and, unlike subsequent winters, the onset of the polar vortex occurred while the stratosphere was relatively uncontaminated by effects of volcanic eruptions (Swissler et al., 1982; Kent and McCormick, 1984). The data have been analyzed on a synoptic basis; for each event during the period, the SAM II aerosol extinction has been classified according to the position of measurement and its relationship to the polar vortex. The emphasis of the analysis has been
on the use of the aerosol optical properties as a tracer for dynamic changes taking place inside the vortex. In this connection, we have been concerned with possible movements in the vertical, as well as the horizontal, directions. Our conclusions concerning stratospheric motions have been compared with observations made between 1958 and 1963 using radioactive tracers injected by nuclear weapon tests.

Our analysis has concentrated on the behavior of the background aerosol and we have attempted to eliminate PSCs from the data, as their inclusion strongly affects the mean value calculated for any aerosol optical parameter. We have also avoided studying in detail the changes taking place during stratospheric warming episodes in the second half of winter. The qualitative behavior of the aerosol extinction is discussed for the complete eight-month winter period, while the quantitative analysis is mainly confined to the period between September and December 1979.

The order of treatment of the data in this paper is as follows:

- Section 2. Discussion of the basic characteristics of the SAM II measurements and the application of microphysical temperature modeling and cloud filtering to the data. This is followed by a description of the way in which the SAM II data have been classified with respect to the vortex.
- Section 3. The results of the analysis of the SAM II data are presented. A qualitative description of the changes observed is given, together with the results of calculations on vertical velocity and mass flow.
- Section 4. Comparison is made of the results of the analysis of the SAM II data with earlier measurements using radioactive tracers.
- Section 5. Conclusions are summarized and discussed.

2. Background to analysis

a. SAM II measurement characteristics and data treatment

The SAM II satellite sensor consists of a sun photometer designed to measure atmospheric extinction, in a wavelength pass band centered on 1.0 μm, at each satellite local sunrise and sunset (McCormick et al., 1979). The satellite orbit is a high-noon sun-synchronous one with all sunrises occurring in the Antarctic region and all sunsets occurring in the Arctic region. The exact latitude of the sunset measurements varies quasi-sinusoidally with season between about 65 and 82°N. The measurement location reaches its most northerly latitude at the equinoxes and its most southerly at the solstices. This variation is of significance when considering the region of the polar vortex that is being sampled by SAM II. Consecutive measurement events are separated by approximately 26° in longitude.

The pass band at 1.0 μm is one with negligible atmospheric gaseous absorption and the extinction is that due to aerosol and molecular scattering alone. The attenuation is measured along a path approximately 300 km in length through the atmosphere with an instantaneous field of view of about 0.5 km.

The basic data product following the inversion procedure (Chu and McCormick, 1979) is an aerosol extinction profile for each satellite event at a wavelength of 1 μm. This has a vertical resolution of about 1 km and extends downward from 50 km to an altitude determined by the presence of cloud. A general discussion of high latitude satellite extinction and its variation with altitude and season has been given by McCormick et al., 1981. The SAM II maximum aerosol extinction in the stratosphere is reached below an altitude of 15 km with a value of $2 \pm 1 \times 10^{-4}$ km$^{-1}$. Above this level, the aerosol extinction decreases with altitude in a manner which depends upon both season and geographical location, becoming limited by noise between 20 and 35 km. A full discussion of errors in the inverted data has been given by Chu and McCormick (1979) and Kent and McCormick (1984).

In this paper, we shall present data in the alternative form of optical depth, which is obtained by integrating the extinction between two specified height levels. Optical depth is an approximate measure of the aerosol column mass concentration and, by employing a suitable conversion factor, may be used to calculate the aerosol mass loading. We shall also use extinction ratio $R$, which is defined as

$$R = (E_a + E_m)E_m^{-1}$$

where $E_a$ is the aerosol extinction and $E_m$ is the molecular extinction. In the absence of aerosol formation or loss, and subject to microphysical temperature and water vapor pressure corrections, the aerosol extinction ratio is conserved in a manner similar to a constituent mixing ratio. It is thus not dependent upon the atmospheric pressure and density changes occurring in either horizontal or vertical air mass movement.

Several problems may arise in the uncritical use of the aerosol extinction data. One of the most obvious of these is the inclusion of data taken in the presence of high altitude cloud. Cloud extinction values are very much greater than background aerosol extinction values, and their inclusion will seriously affect the calculation of any corresponding mean value. A study has been made of the frequency distribution of extinction values, both above and at the level of high cloud. Inspection of the complete frequency distribution of extinction values for all altitudes above 8 km shows two distributions, which may be separated by filtering the data at an extinction level of $4 \times 10^{-4}$
km$^{-1}$. Elimination of extinction values above this level reduces the amount of available data by about 20% at the altitude of the tropopause.

In addition, the presence of cloud produces perturbations in the inversion procedure in the immediate neighborhood of the cloud. Inverted extinction data from below a thin cloud layer are also subject to error, as they are obtained by assuming the cloud to be homogeneous at a fixed altitude, an assumption almost certain to be incorrect. Thus, we have also filtered out data below a cloud layer and data immediately above a very sharp gradient in extinction. This filtering procedure allows us to bring the analysis down to an altitude of 8 km (1 km below the mean tropopause altitude during the period of study).

Use is made of extinction ratio as a conservative quantity under transport. This is true only if the aerosol temperature remains constant. If a temperature change occurs, growth or evaporation of the aerosol takes place as the individual aerosols seek to maintain their equilibrium with the ambient water vapor. The effect of this change in aerosol particle size and composition on the extinction at 1.0 μm has been modeled for moderate changes in ambient conditions by Yue and Deepak (1981). To compare data taken in different months, we have chosen 223 K (−50°C) as a reference temperature. This value was selected as being close to the mean stratospheric temperature for altitudes below 30 km during August 1979, the last complete month preceding the formation of the polar winter vortex. Yue and Deepak (1981) have published graphs showing the temperature dependence of the 1.0 μm extinction between 190 and 250 K, for different values of the ambient water vapor concentration using a background nonvolcanic stratospheric aerosol model. We have used the results of this calculation and, assuming a constant stratospheric water vapor concentration of 4 ppmv (Murgatroyd, 1982), have determined a factor, $C(T)$, to be applied to the extinction to convert it to the equivalent extinction at 223 K. The factor $C(T)$ is defined as

$$C(T) = E_d(T)E_d(223)^{-1},$$

where $E_d(T)$ and $E_d(223)$ are the aerosol extinction at temperatures $T$ and 223 K, respectively. Table 1 shows a brief list of values for $C(T)$. It has been applied to the data using the temperatures supplied by the National Meteorological Center (NMC) (Gelman et al., 1981). At stratospheric altitudes above 10 km and for temperatures greater than 205 K, which represent the bulk of the data, it can be seen that the factor is at most about 1.5 and usually substantially less than this value. At lower altitudes, where we have assumed for convenience the same water vapor mixing ratio, and at lower temperatures, the factor can become large and very sensitive to small errors in the temperature assumed. In particular, at very low temperatures when the extinction may increase by several orders of magnitude, the calculation of an adjustment factor is unreliable.

At temperatures below about 200 K, the very large growth in particle size manifests itself in the SAM II data as PSCs, which are observed in the arctic region between altitudes of 14 and 23 km. Sightings of these have been reported by McCormick et al. (1981, 1982) and the mechanisms of formation has been discussed in detail by Steele et al. (1983). In theory, a threshold temperature might be chosen which would isolate all polar stratospheric clouds. In practice, neither the temperature nor the environmental water vapor pressure are known well enough for this to be possible. We have chosen 193 K as a compromise temperature which would eliminate most, but not all, PSCs without eliminating too much useful data. Weakly enhanced PSCs with a peak extinction less than $4 \times 10^{-4}$ km$^{-1}$ may not be removed from the data either by this temperature filter or by the previously discussed cloud filter.

It is useful to consider some of the approximations involved in the application of the factors in Table 1. Apart from those arising from approximations in the models assumed by Yue and Deepak (1981), we should consider the effect of having assumed possibly incorrect values for the water vapor mixing ratio and for the ambient atmospheric pressure. Fortunately, both these effects are relatively small. At an altitude of 20 km and a temperature of 200 K, a 50% error in the water vapor mixing ratio introduces an error of about 9% into the correction factor (which has a value of 1.47 for these conditions). The errors introduced by assuming a constant ambient pressure at a given altitude throughout the polar winter are smaller. The maximum change in pressure at the altitude of 20 km is about 25%, producing an error in the correction factor of less than 5%.

Errors of a similar magnitude are introduced when we consider the changes in ambient water vapor pressure produced by a vertical movement of an air parcel. A doubling in pressure at the reference tem-

<table>
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<td>0.84</td>
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</table>
perature of 223 K, as might be produced by an air parcel descending from 20 to 15 km, will produce a change in extinction ratio of about 5% (or less depending upon the initial value of the extinction ratio).

In general, we must regard the application of the factors given in Table 1 as a very useful first-order correction, subject to a certain amount of error arising from the assumptions made. Their use affects the quantitative, but not the qualitative, nature of our conclusions.

b. SAM II data classification

In order to quantitatively describe the variations in the aerosol optical properties across the vortex, SAM II measurements have been grouped together according to their position relative to the vortex. Since the stratosphere is in approximate geostrophic balance, air flows mostly parallel to the height contours on an isobaric surface. It follows that the region of highest wind speed, or the jet stream, is indicated by the maximum gradient of height contours. The center of the jet stream, when traced around the vortex by the maximum gradient of height contours, forms a quasi-continuous line, although the absolute wind velocity may vary along the jet stream axis and the flow lines of individual air parcels may depart from a pressure surface. This line of maximum wind speed may be viewed as a barrier in that there is little cross flow across the jet stream. The appearance of this barrier is illustrated by the distribution of potential vorticity. Studies by Krishnamurti (1959), Charney and Stier (1962), and McIntyre and Palmer (1983) have shown that above 20 km, potential vorticity for a given altitude is higher on the cyclonic flank of the jet stream than on the anticyclonic flank. A tight gradient in potential vorticity is located in a narrow band along the axis of the jet stream. Under adiabatic and frictionless conditions, air parcels must conserve their potential vorticity for time scales less than about a week (McIntyre and Palmer, 1983) and, consequently, are unable to cross the jet stream where the potential vorticity gradient is greatest. McCormick et al. (1983) observed a rapid change in the aerosol optical properties across the jet stream above 20 km and noted that the change corresponded with the expected variation in potential vorticity. On this basis, it is reasonable to view the center of the jet stream, or position of maximum wind, as the outer boundary of the vortex for a given altitude.

Because of the variation in optical properties across the jet stream above 20 km and because meteorological analysis maps at 30 mb contain more information retrieved from radiosondes than at higher altitudes, much of the analysis presented has been related on a 30 mb surface. The region of greatest wind velocity was found by determining the position of maximum density of height contours on the 30 mb meteorological map for each available day during the period of study. The value of the geopotential height at this point was noted and taken as the height of the axis of the jet stream around the vortex. It is recognized that the gradient wind provides a better estimate of the wind velocity, but for convenience and simplicity the geostrophic approximation was used to locate the axis of the jet stream. The height of the center of the vortex was also noted from the same map. These data were used to group the SAM II measurements according to three separate classification schemes, and were then used to examine different aspects of the variation of aerosol extinction.

1) Inside and outside the polar vortex: This method of grouping the SAM II data is illustrated in Fig. 1. A measurement has been classified as inside the vortex if the 30 mb height at the position of measurement is lower than the height at the center of the jet stream on the same isobaric surface.

2) Binning relative to the center of the polar jet stream: This binning scheme is a detailed expansion of the previous classification method and is illustrated in Fig. 2. The SAM II data have been placed into 16 sequential height bins on the 30 mb surface, each representing a change in geopotential altitude of 160 m. These bins have been numbered according to their position relative to the axis of the jet stream.

![Fig. 1. Location of SAM II measurements and the 30 mb meteorological analysis map for 27 December 1979. Geopotential altitudes are shown in decameters. Two selected SAM II measurements shown in Fig. 3 are indicated by triangles; remaining SAM II measurements are shown by circles. Events have been classified as inside the polar vortex if they lie on the cyclonic side of the polar jet stream (within the shaded region).](image-url)
Usefulness of these classification schemes depends on the uniformity of the SAM II measurement sampling. At certain times, portions of the vortex are poorly sampled. For example, at the end of winter, when SAM II measurements were being made at about 80°N and the polar vortex had been displaced from the pole, relatively few satellite observations were made inside the vortex. On the other hand, in October, when the SAM II latitude of measurements was also high and the vortex was centered about the pole, most sampling occurred within the vortex. For most of the data analyzed, the effect of such sampling problems has been minor.

3. Results of analysis

a. SAM II data

SAM II extinction profiles obtained inside and outside the polar vortex have been presented by McCormick et al., 1983. These profiles, obtained less than one year after the eruption of El Chichon volcano, show a marked difference in the aerosol concentrations above about 20 km, the aerosol extinction within the vortex being more than one order of magnitude lower than that outside.

A similar difference is observed for the period

![Image of aerosol extinction profiles](image_url)

FIG. 3. SAM II extinction profiles obtained on 27 December 1979. The locations of the measurements are as shown in Fig. 1. The solid line indicates the observation made at 65.0°N and 110.1°E (inside polar vortex). The dashed line indicates the observation at 65.0°N, 98.4°W (outside polar vortex). The dot–dash line represents the altitude of the tropopause.
under study, and two profiles showing original data obtained on 27 December 1979 are shown in Fig. 3. The error bars shown in the figure are typical of the random error occurring in the SAM II data. The location of these measurements are shown by triangles in Fig. 1. Once again, it may be noted that the aerosol extinction above 20 km is almost one order of magnitude lower within the vortex than outside. Application of the temperature adjustment described in Section 2a will increase this difference as temperatures inside the vortex are colder than those outside. Differences between the profiles at altitudes below 16 km are also apparent, where the aerosol extinction within the vortex is greater than that outside. These differences are typical of profiles inside and outside the vortex, although the detailed behavior depends upon the exact position of measurement. Measurements at altitudes above 25–30 km, where the extinction may be less than $1 \times 10^{-5}$ km$^{-1}$, show insufficient signal-to-noise ratio to permit a comparison to be made of individual profiles. Individual profiles have been binned according to classification Scheme 1), grouping together measurements made inside and outside the vortex. Weekly averages of these extinction profiles, without temperature adjustment, have then been calculated. Examples of these averages shown in Fig. 4 were chosen to be representative of different periods during the 1979/80 winter season. The early period of winter is represented in Fig. 4a. The difference in extinction between measurements made inside and outside the vortex is evident. This difference becomes much more pronounced in the middle of winter (Fig. 4b) and persists to the end of the winter (Fig. 4c).

Studies of these and other weekly averages show that the characteristics of the extinction profiles as discussed above are common to all the data and not associated with limited periods of measurements or selected profiles. The error bars shown in Fig. 4, which represent the standard deviation of the weekly mean values, are small and typical of most of the data. They indicate that the variability in extinction within each of the two groups is small. Only between 31 December and 25 February, when the data inside the vortex are contaminated by PSCs, is the variation larger.

b. Temporal variation of aerosol optical properties near the vortex center

An analysis of the aerosol data was performed using the binning classification scheme relative to the vortex center (Scheme 3). SAM II data obtained near the center of the vortex were averaged for consecutive individual weekly periods. Figure 5a shows the temporal variation of the isobaric (thick lines) and isentropic (thin lines) surfaces for data near the center of the vortex during the winter of 1979/80. Dashed lines in the figure indicate a period when the data were intermittent or occasionally ambiguous. (Note that data are present for a week in mid-March and the first week of April). As seen in Fig. 5a, the deepening of the depression is indicated by the descent of the pressure surfaces in autumn and the filling of the vortex is indicated by the ascent of these surfaces in March. The vortex reached its maximum intensity toward the end of December, when pressure surfaces were at their lowest altitudes. In contrast, the isentropic surfaces move slightly upward in altitude.

Figure 5b shows the plot of observed extinction

![Fig. 4. Weekly average extinction profiles obtained inside (open circles) and outside (solid circles) the polar vortex for selected weeks during the 1979/80 season (classification Scheme 1).](image-url)
ratio near the center of the vortex after it has been filtered for cirrus clouds and strong PSCs. In the figure, extinction ratio contours vary from a value of 1.2 to a value of 2.8 between the altitudes of 9 and 30 km. Figure 5c shows the same data adjusted to the temperature of 223 K. Comparing the two plots in Figs. 5b, c, it is evident that the general characteristics of the extinction ratio contours are not substantially altered by the application of the temperature adjustment, particularly before December. In the months of December, January and February, the average extinction ratio values are significantly lower in the temperature adjusted version. For example, the extinction ratio contour levels of 2.6 and 2.8 seen in Fig. 5b are removed from Fig. 5c. The adjusted version of the aerosol extinction ratio is a more conserved quantity since microphysical processes due to temperature are removed. This allows for a better assessment of dynamical processes taking place in the vortex and the following analysis uses this data set. Unfortunately, not all of the PSCs observed in the aerosol data set were removed. PSCs located near the center of the vortex were observed between 26 January and 23 February 1980.

We may divide Fig. 5c into two distinct periods. The first half of winter is characterized by the steady descent in altitude of the extinction ratio contours. During this period, the extinction ratio value of 1.2 moved downward approximately 11.5 km in altitude and the extinction ratio value of 2.2 (initially near an altitude of 18 km) descended about 5.5 km in altitude. It is interesting to note the almost linear nature of the individual contour lines. As the winter progressed, it is apparent that the contour lines converged together above about 15 km. Below 15 km, individual contour levels tend to vary more than at higher altitudes, but still show a downward trend. An important feature of this descent during the first three months of winter is that the temperature-adjusted peak extinction ratio stayed at a constant value between 2.2 and 2.4. The descent of the extinction ratio contours implies a downward movement of air, or subsidence. Since the isentropic surfaces shown in Fig. 5a for the same period move upward in altitude, the inferred motion of the air is not adiabatic. The isobaric surfaces shown in Fig. 5a do descend in altitude. However, inspection of the rate of descent of the surfaces in Fig. 5a shows that the height change is less for the same period than that in Fig. 5c. It is thus evident that hydrostatic settling of the pressure surface alone cannot explain the observed descent of the extinction ratio contours.

A possible explanation for the descent of extinction ratio contours is the gravitational sedimentation of the aerosol. An estimate of the aerosol sedimentation rates can be made by utilizing a commonly used log normal size distribution for nonvolcanic stratospheric aerosols (Pinnick et al., 1976; Russell et al., 1981; Steele et al., 1983; Yue and Deeppak, 1981). Some change in particle radius occurs with changing temperature. Steele and Hamill (1981) have shown that a particle at a pressure altitude of 50 mb and an ambient water vapor concentration of 6 ppmv (this value is greater than the value of 4 ppmv used elsewhere in this paper and would cause greater particle growth) would increase its radius by about 18% as the temperature was lowered from 220 to 200 K. This increase is not large and indicates that during
the first part of winter the mean particle radius would still be well under 0.1 μm. The sedimentation rate for spherical particles has been calculated by Kasten (1968), who shows that a 0.1 μm radius sulfuric acid particle, at an altitude of 20 km, would fall at a rate of about $4 \times 10^{-5}$ m s$^{-1}$ or about 0.3 km for the first half of winter. This is an order of magnitude lower than the descent of the extinction ratio contours near the vortex center, at the same altitude shown in Fig. 5c. Gravitational settling rates at other altitudes are similarly less than the observed subsidence rates during the October–December time period. The fact that gravitational settling may be ignored during the first half of winter should not be taken to imply that it cannot be a significant factor at any time. During January and February, lower temperatures occur in the stratosphere, leading to much greater particle growth and to the formation of polar stratospheric clouds. If the particle size were to increase tenfold to about 1 μm, the gravitational sedimentation rates would also increase by at least an order of magnitude (Kasten, 1968) and would be a significant factor in the vertical redistribution of aerosol. We believe that explanation of the steady descent of extinction ratio contours in the first half of winter must lie in the downward movement of the air mass across pressure surfaces.

It is of interest to use the contours in Fig. 5c to calculate the vertical mass flow of air for the first half of winter near the vortex center. The mass flow of air $F$ may be determined, based on continuity considerations, by the following expression:

$$F = \rho W = \rho_0 W T_0 P (T_0) \omega_1,$$

where $T$ is the temperature, $P$ is the pressure, $\rho$ is the density and $W$ is the vertical velocity of air (the subscript zero refers to the standard atmospheric sea level values). The rate of descent of each contour line from early September to mid-December was calculated by determining the slope of the line by linear regression. This was taken to be the absolute vertical velocity of air at the midpoint in the time and altitude of each regression line.

The mass flow was calculated using temperature and pressure for the midpoint of the regression line (first week of November 1979). In order to determine the diabatic component, a similar computation was carried out for vertical movement of the isentropic surfaces. Relative vertical velocity of air was determined at different heights. Table 2 lists the absolute and relative vertical velocities for the extinction ratio contours and the relative mass flow computed at different altitudes for the midpoint of this period of study. The uncertainty limits were evaluated from confidence intervals determined for the slopes of the regression lines and are also shown (as standard errors). The standard errors are minimum estimates, being based purely on the accuracy of the slopes of the regression lines and do not contain any estimate of additional inaccuracies arising from the uncertainty of assumptions in the preceding analysis. Data in this table do not extend to altitudes below 14 km. Later we will show that at lower altitudes it is not possible to distinguish between the inside and outside of the vortex, and consequently free mixing occurs.

The vertical mass flow can be seen to increase steadily with decreasing altitude between 24 and 14 km. If there were no flow across the polar-night jet, one would expect the mass flow to be constant from mass continuity consideration. The observed slow increase of the mass flow implies a small leakage into the vortex. An approximate value for the velocity of flow into the region near the vortex center [as defined by binning Scheme 3] may be calculated from the rate of change of vertical mass flow with altitude. In mid-November, this region had a mean radius of approximately 500 km. Use of the continuity equation leads to a mean radial flow velocity into the region, at an altitude of 20 km, of approximately $3 \times 10^{-2}$ m s$^{-1}$. This value may be compared to the mean subsidence rate within the region at the same altitude, of about $7 \times 10^{-4}$ m s$^{-1}$, and to the very much

### Table 2

<table>
<thead>
<tr>
<th>Extinction ratio contour</th>
<th>Altitude (km)</th>
<th>Downward vertical velocity ($\times 10^{-4}$ m s$^{-1}$)</th>
<th>Mass flow relative to isentropic surface ($\times 10^{-3}$ kg m$^{-2}$ s$^{-3}$)</th>
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<tr>
<td></td>
<td></td>
<td>Absolute (μ)</td>
<td>Relative to isentropic surface (μ)</td>
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<tr>
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<td>23.5</td>
<td>10.6 (0.7)</td>
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<td>1.4</td>
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greater wind velocity tangential to its boundary of about 10 m s⁻¹.

In the second half of winter, as seen in Fig. 5c between mid-December and the end of March, the aerosol data set contains weak, unfiltered PSCs inside the vortex. In addition, the vortex was disturbed by several warming events. These events caused uncertainties in the binning of data near the vortex center because of the rapid changes in size, shape and position of the vortex. An example of the influences of a warming event is seen in Fig. 5b in early January. Relocation of the vortex center has caused an apparent upward movement of the extinction ratio contours several weeks before the first sighting of a PSC near the vortex center. As a result, the interpretation of the extinction ratio contours seen in Fig. 5c during the second half of winter is difficult and the calculations of vertical velocity and the mass flow of air have not been attempted. Even if it were possible to calculate the vertical mass flow in the second half of winter, it would be necessary to consider both increased sedimentation at very low temperatures and increased vertical velocities during sudden warming episodes (Holton, 1980) in their interpretation.

The most important feature seen in the second half of the figure is the return to a vertical aerosol extinction ratio structure similar to that seen in late December. This indicates that the extinction ratio values stay approximately constant at a given altitude as the isobaric surfaces move upward in altitude, as seen in Fig. 5a. It also indicates that the vortex tends to preserve aerosol optical properties established earlier in the winter season.

c. Variation of extinction ratio across the vortex

In order to discuss the way in which extinction ratio varies with position relative to the polar-night jet stream, the SAM II data have been grouped according to month of measurement and then classified using binning Scheme 2. In any month, the data in the lowest number bin lie close to the vortex center. The extinction ratio values have been adjusted to the standard temperature of 223 K.

The results of binning the data in this manner for each month are shown in Fig. 6b–j. The center of the jet stream is marked by an arrow at bin position 8.5 and it can be seen that the SAM II measurements have been located fairly symmetrically about this position throughout the whole winter. Figure 6a shows a set of reference conditions indicating the state of the stratospheric aerosol before the formation of the polar vortex. These are based on the mean aerosol extinction profile for August 1979, and for the purposes of comparison with the other figure the extinction ratio has been shown in the form of a contour plot against bin number. This is intended as a reference only, the bin numbers have no significance as applied to data obtained before the formation of the vortex. In all these plots, the data have been analyzed (after filtering for cloud) down to an altitude of 8 km and the variation of the altitude of the tropopause with bin number shown on each figure.

The most significant feature of the contour plots shown in Fig. 6 is the variation in extinction ratio across the jet stream. This variation, which is visible even in the relatively shallow vortex of September 1979, becomes more pronounced as the vortex deepens. All months show a well-developed region of maximum extinction ratio which descends gradually within the vortex as winter progresses. In the first half of winter, the slope of the contours above the layer maximum remains fairly constant with time and with bin number. In the second half of winter, the transition across the jet stream becomes more abrupt. It is not clear if this change is real or the result of a sampling bias. Between September and December, SAM II sampled the vortex over a range of longitudes. In March and April, the vortex was displaced from the pole and only a limited section was sampled. In the former case, some longitudinal blurring of the transition region may have taken place which would not have occurred at the end of winter. In January and February, the contour bins within the vortex show strong irregularities due mainly to the presence of unfiltered PSCs which occurred only at the lower temperatures found inside the vortex. Despite these, the maximum extinction ratio values observed within the vortex remain fairly constant at a level just greater than 2.2 during the whole winter period. In contrast to the region near the vortex center, which shows a well-defined layer maximum, the region outside the center of the jet stream shows a flat, ill-defined layer maximum extending over a considerable vertical range. Below the region of maximum extinction ratio, the contour lines become approximately parallel to the line showing the mean variation of tropopause altitude.

Comparing the prevortex reference conditions with those in the middle of winter, we see that at the vortex center, the layer maximum has descended from approximately 19 km to about 14 km. In bins 8–10, near the center of the jet stream, the subsidence of the layer maximum is smaller and the separation of the contour lines is such as to make it difficult to obtain an exact value for this. Despite this, it is clear that subsidence has taken place in this region and that, at the layer maximum, its value is about one-half of that occurring at the vortex center. Outside the center of the jet stream, the behavior of the aerosol extinction is somewhat irregular. This is as expected since transport to these regions takes place much more freely than to the center of the vortex.

Values for the temperature adjusted extinction ratio of 2.6 or greater are observed between December and March at altitudes of 18–22 km outside the
Fig. 6. Variation of extinction ratio across the polar vortex. Data have been grouped by month and classified according to position relative to the center of the jet stream (classification Scheme 2). Panels (b)-(i) show individual monthly data from September 1979 to April 1980. Panel (a) shows a set of reference conditions based on data obtained prior to the formation of the vortex. [The bin numbers in (a) have no significance and are shown for comparison purposes only.] The vertical arrow at bin position 8.5 indicates the center of the jet stream and the dot-dash line indicates the altitude of the tropopause.
vortex. These values are larger than those observed, either within the vortex or in the polar region before the commencement of winter. Values of this magnitude are observed at similar altitudes in low latitudes (Kent et al., 1982). In the absence of any mechanism for the formation of new aerosol or for aerosol growth, and the consequent production of high extinction ratios, it seems likely that transport from low latitudes is responsible for these observations. Evidence for this comes from observations made by the SAGE (Stratospheric Aerosol and Gas Experiment) satellite system. SAGE measures aerosol extinction at a wavelength of 1 μm in a manner similar to SAM II but with a wide latitudinal coverage (McCormick et al., 1979). Figure 7a shows a contour plot of the extinction ratio measured by SAGE in December 1979. A region of high extinction ratio is centered over the equator at an altitude of about 25 km with a tongue of material extending northward at an altitude of 20–25 km up to the northern limit of SAGE measurements at this time of year (45°N). The peak value of extinction ratio observed at 45°N is similar to that observed by SAM II at similar altitudes just outside the vortex. Figure 7a also shows high extinction ratios at low latitudes just above the tropopause. These are absent from earlier SAGE data and are associated with the eruption of the Sierra Negra volcano in the Galapagos Islands (0.8°S, 91.2°W) on 13 and 17 November 1979. Figure 7b shows the same data plotted in terms of extinction rather than extinction ratio where the effect of the Sierra Negra eruption is emphasized. We believe that this fresh volcanic material has also been transported northwards modifying the optical properties of the aerosol just above the tropopause. Support for this interpretation comes from lidar observations made at Fukuoka, Japan (33.5°N, 130°E) (Fujisawa et al., 1982). These observations showed a layer of enhanced scattering at an altitude of 17 km on 11 December 1982, which persisted intermittently for several months and was attributed, by Fujisawa et al., to the injection of volcanic material into the stratosphere by the Sierra Negra eruptions.

The peak value of the extinction ratio within the vortex remains constant throughout the winter at a value between 2.2 and 2.4, while the altitude at which this is observed descends from about 18 km in September to 12 km in April. At the latter time, it has, in fact, descended to a level where there appears to be no differentiation between the inside and the outside of the vortex. We noted earlier that below the layer maximum the contour lines become approximately parallel to the tropopause. The altitude at which there is a transition from a steep gradient of contour lines to parallel contours may decrease slightly as the winter progresses. It is difficult to be precise about this, as this vertical transition does not become clearly visible until November, although it then persists through to April. In most months, there is clear evidence of a horizontal transition across the jet stream from high to low extinction ratios at all altitudes above 14 km and, in some months the transition may persist to slightly lower altitudes. The parallel nature of the contours below 14 km indicates the likelihood of horizontal transport and mixing at these heights.

**Fig. 7.** Contour plots of zonal mean 1 μm aerosol measurement data from the SAGE satellite, 22 November–31 December, 1979. (a) Extinction ratio; (b) extinction (10^-4 km^-1).
d. Optical depth variation

The aerosol optical depth between two altitudes may, using a suitable conversion factor (Kent and McCormick, 1984), be used to determine the approximate aerosol mass loading between those altitudes. It is of considerable interest to see how this mass loading has changed both inside and outside the vortex as the 1979/80 winter progressed. In order to take into consideration the apparent vertical transition near an altitude of 14 km, we have chosen this as a reference altitude and examined the optical depth behavior above and below this level. The aerosol optical depth has then been calculated, using temperature adjusted and filtered data, for two stratospheric altitude ranges: the first 9 km (mean tropopause altitude) to 14 km and the second 14 to 30 km (upper level of useful data).

Figure 8a shows the monthly running mean optical depth for the altitude range 9–14 km. A similar plot in Fig. 8b shows the optical depth for the height range 14–30 km. The optical depth has been plotted at weekly intervals from September 1979 to March 1980 for both the inside and the outside regions of the polar vortex as referred to the 30 mb pressure level. The most significant aspect of the data shown in Fig. 8a is a doubling of the optical depth between November and March and the very close correspondence between the variation inside and outside the vortex below 14 km. The increase in optical depth after November is believed to be associated with the transport of material from lower latitudes, particularly that injected by the eruption of the Sierra Negra volcano. Figure 7b shows that the majority of the material from the Sierra Negra eruptions lay quite close above the tropopause. We would therefore expect only small amounts to be located above our reference altitude of 14 km in the higher latitudes. The correspondence between the two curves in this figure agrees with the hypothesis that this altitude range lies beneath the polar vortex and that the division into inside and outside regions has no significance. Moreover, the close agreement during a period of significant optical depth variation and transport from lower latitudes provides evidence for good horizontal mixing up to the highest latitude sampled here (about 80°N).

Figure 8b shows the variation in optical depth inside and outside the vortex for the altitude range 14–30 km. The behavior here is in sharp contrast to that shown in Fig. 8a. Outside the vortex, significant changes are observed, which we believe to be associated with horizontal transport from lower latitudes. Nevertheless, the mean level at the end of the winter season is close to that at the beginning (1.0 × 10^-3) and there is no evidence of the strong increase in optical depth seen at lower altitudes. Inside the vortex, the optical depth was observed to decrease almost monotonically for the whole winter period. Except for a period during January and February when the data are contaminated by the presence of unfiltered PSCs, optical depth decreased linearly by about 6% of its original value per month. This decrease is interpreted as being due to subsidence, the aerosol being lost from the base of the vortex and becoming mixed into the lower stratospheric level (9–14 km).

![Fig. 8. Monthly running mean optical depths. Solid lines indicate data obtained outside the polar vortex and dashed lines indicate data obtained inside the polar vortex. (a) Altitude range 9–14 km; (b) altitude range 14–30 km.](image-url)
Horizontal motions beneath and outside the vortex are more rapid than descending motions inside the vortex; consequently, no accumulation of material is to be expected.

It is possible to use the change in optical depth inside the vortex to provide an approximate estimate of the mass of aerosol transferred downward by subsidence through the base of the vortex into the lower stratosphere. The polar vortex, inside the axis of the polar night jet, has an area between September and December of about $2 \times 10^7 \text{ km}^2$ at the 30 mb level. Over the same time period, the mean optical depth inside the vortex has fallen from an initial value of $9.5 \times 10^{-4}$ to about $7.2 \times 10^{-4}$. The mass loss may be computed from the equation:

$$\Delta M = A\eta \Delta \tau,$$

where $\Delta M$ is the mass of the aerosol lost by subsidence, $A$ is the area of the polar vortex, $\eta$ is the extinction to mass conversion factor, and $\Delta \tau$ is the mean decrease in optical depth. For our calculation, we have used a value for $\eta$ of $1.10 \times 10^9 \text{ m}^2 \text{ kg}^{-1}$ adopted by Kent and McCormick (1984), who chose this value as a mean between those values appropriate to prevolcanic and postvolcanic periods. Use of these figures gives a value for $\Delta M$, the mass loss by subsidence up to the end of December, of about 4,000 tonnes. This value does not take account of any mass loss or gain through horizontal exchange across the polar jet stream. It is simply an estimate of the apparent net change in mass loading above 14 km that has taken place in the aerosol that we see inside the vortex at the end of December. The vortex, after December, is characterized by a fluctuating area and by loss through detachment and horizontal mixing (McIntyre and Palmer, 1983). Nevertheless, it is clear from Fig. 8b that the mean optical depth continues to decrease at approximately the same rate as before and during December. It is likely, therefore, that the same process causes a similar rate of transfer of aerosol mass to the lower stratosphere between January and March, as occurred between September and December, making a total for the whole winter of about 7,000 tonnes. This figure represents about 1.4% of the total global aerosol mass loading in a nonvolcanic period (Kent and McCormick, 1984). Therefore, on a global scale during a nonvolcanic period, the significance of this subsidence within the north polar vortex is small. On an occasion in which a northern latitude volcanic eruption may have injected new and localized material that has entered the polar vortex (e.g., after the eruption of Mt. St. Helens in May 1980), it may well be a significant mechanism for vertically redistributing the aerosol.

The change in optical depth within the vortex for altitudes above 14 km may be used to calculate an approximate value for the mean subsidence rate within the vortex. During the 15-week period from 15 September to 30 December, the mean optical depth between 14 and 30 km was observed to decrease from $9.5 \times 10^{-4}$ to $7.2 \times 10^{-4}$. If we assume that this change is solely due to the transfer of aerosol downwards across the 14 km altitude level, we may determine the mean vertical velocity at 14 km using the formula

$$\Delta \tau = E_a W \Delta t,$$

where $\Delta \tau$ is the optical depth decrease ($2.3 \times 10^{-4}$), $E_a$ is the mean aerosol extinction at an altitude of 14 km ($1.6 \times 10^{-4} \text{ km}^{-1}$), $W$ is the vertical velocity and $\Delta t$ is the time period (15 weeks); $W$ is found to have the value $1.6 \times 10^{-4} \text{ m s}^{-1}$. After correcting for the movement of the isentropic surface at 14 km within the vortex over this time period, we obtain a value for the mean diabatic velocity of $2.1 \times 10^{-4} \text{ m s}^{-1}$. The corresponding vertical mass flow rate is $4.2 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$. This value is an average for the whole of the inside of the vortex. It is a minimum value as no correction has been made for any horizontal flow of material across the polar-night jet stream into the vortex. It is to be compared with the value obtained in Section 3b for the same altitude of $10.4 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ which referred to the subsidence at the center of the vortex, where it would be expected to reach a maximum.

4. Comparison with radioactive tracer observations

Atmospheric nuclear testing carried out during the period 1951–62 afforded an opportunity for studying stratospheric circulation patterns through monitoring of the distribution of radioactive debris. In particular, observations made in or near the north polar vortex showed several features consistent with those of the SAM II observations described. Three series of tests have provided results of interest: the United States “Hardtack” tests, carried out in the low latitude Pacific in the spring and early summer of 1958; a further United States series, also at low latitudes, in April–July, 1962; and the USSR series in 1961–62, near Novaya Zemlya, within the arctic circle.

Analyses of radioactive tracer observations following the 1958 “Hardtack” test have been published by Feely and Spar (1960) and Kalkstein (1962). These analyses are based on radioactive samples collected by Lockheed U-2 aircraft in the meridional-vertical plane in the Western Hemisphere, extending from about 75°N to 57°S. The May–July tests of 1958 resulted in the injection of Tungsten-185 into the lower stratosphere over the equator. This debris spread fairly rapidly in the stratosphere and was observed in north temperate latitude fallout in late 1958. Vertical profiles of Tungsten-185 for November–December 1959 (Feely and Spar, 1960) showed the existence of an equatorial reservoir at an altitude of about 20 km
with a layer of maximum activity sloping downwards to 75°N where the maximum occurred at an altitude of about 15 km. Feely and Spar (1960) interpret this observation in terms of a latitudinal mixing in the lower stratosphere. Other studies (Newell, 1961; Her- ring, 1966), show similar features using ozone as a tracer instead. Consistent with these observations is the interpretation of the SAM II and SAGE aerosol data presented in Sections 3c and 3d. Material from the equatorial stratosphere is mixed northward at all altitudes to the boundary of the polar vortex, and below about 14 km is mixed all the way to the pole.

In contrast to the Tungston-185 and ozone measurements, which are concerned with the dynamics of the lower stratosphere, those made on Rhodium-102 which are presented by Kalkstein (1962), are related to movements in the upper stratosphere. The Orange shot of the “Hardtack” series made on 11 August 1958 injected material containing the rhodium isotope at an altitude of 43 km above Johnston Island (15°S, 170°W). It is estimated that the debris rose rapidly to at least 100 km and probably higher. A sharp increase in Rhodium-102 was observed at high latitudes in the northern hemisphere in the late fall of 1959, the effects being strongest north of 60°N and not being seen in lower latitudes until early 1960. These observations, which were made at an altitude of 19.3 km, were interpreted in terms of high altitude (near 100 km) transport of Rhodium-102 to high latitudes, followed by a large-scale downward movement over the polar region. This was then followed by southward mixing of the newly subsided material during the early part of 1960, after the occurrence of warming pulses also believed to be responsible for much of the downward mixing. Telegados and List (1964) deduced a similar pattern of downward mass movement in the polar region following the same detonation over Johnston Island and also reported this motion to be evident a year earlier (winter of 1958–59) after a Soviet bomb was exploded at high latitudes. The reported observations of radioactive debris from the August 1958 explosion showed that between December 1959 and March 1960, material had descended from 20–14 km at an average rate of about 1.5 km month⁻¹ (6 × 10⁻⁴ m s⁻¹). The observations further showed that after March 1960 the downward progression of the debris ceased, coinciding with the annual breakup of the cyclonic polar vortex. Although the interpretation of the radioactive debris measurements, particularly that for the downward mixing, does not correspond exactly to that for the SAM II measurements given in Section 3, the actual measurements agree very well. In both cases we have subsidence within the vortex, shown to be a continuous phenomenon in relation to the SAM II data although it is almost certainly accompanied by an additional, more rapid subsidence during warming pulses (Holton, 1980). We have also, in both cases, observed the wintertime isolation of subsided material within the vortex, followed by its release in spring, during distortion and final breakup of the vortex.

The above model is supported by the observations reported by Telegados (1967). Measurements were made, at altitudes between 15 and 33 km, of Strontium-90 concentrations during January–May 1963. These observations, made by aircraft (15–21 km), at latitudes from the equator to 70°N and by balloon (15–33 km) at 31°N, followed a series of atmospheric tests carried out by the USSR over Novaya Zemlya (75°N, 55°E) in late 1962. The observations showed that the major fraction of the nuclear debris was injected into the polar vortex and remained there until February 1963. Its appearance at latitudes south of the vortex followed a split of the vortex into two cyclonic cells which migrated southward. A homogeneous stratospheric distribution of radioactive material was not, in fact, observed until May 1963, when these two vortices finally disappeared. This confinement of material within the vortex, at altitudes above 15 km, until the spring disappearance of the vortex, agrees well with the experimental picture described in Section 3, where the polar-night jet appears as a boundary, separating material inside and outside the vortex.

5. Summary and discussion

Aerosol extinction data obtained by the SAM II satellite sensor at altitudes between 8 and 30 km have been studied in relationship to the cyclonic polar vortex for one complete winter season (1979–80) in the Northern Hemisphere. The emphasis of this study has been on a synoptic approach and the use of the aerosol extinction ratio as a tracer for mean atmospheric motions.

The data have been filtered to remove the effects of high clouds and the more intense polar stratospheric clouds, and have been adjusted using a standard microphysical model to obtain the equivalent extinction at a constant reference temperature. This adjustment affects only the quantitative and not the qualitative aspects of the data and its interpretation. Data obtained during the first half of winter have been used to infer mean motions in and near the vortex. In the second half of winter the occurrence of PSCs and sudden warmings has made a similar quantitative analysis impossible.

The SAM II data showed that at altitudes above about 14 km, a systematic difference in aerosol extinction existed across the polar-night jet stream. Below this altitude, the aerosol optical properties were independent of measurement locations, indicating a region of free horizontal mixing. As winter progressed, a gradual lowering of the aerosol layer was observed.
above 14 km on the cyclonic side of the polar-night jet stream. This subsidence reached a maximum at the vortex center and persisted to the end of winter. On the anticyclonic side of the jet stream, transport of material from low to high latitude is apparent.

On the basis of these results, it is possible to visualize the polar vortex having a base at about 14 km and an outer boundary coincident with the jet stream axis. Calculations of mass flow rates inside the vortex show that material subsides downwards through the base into the well-mixed layer beneath the vortex. Mean vertical velocities inside the vortex range between $4-11 \times 10^{-4}$ m s$^{-1}$ and are about 4 orders of magnitude less than typical horizontal velocities. The variation with altitude of the vertical mass flow rates near the vortex center indicates a small cross-flow, or horizontal leakage into this region, with mean cross-flow velocities of a few centimeters per second. The integral mass flow due to this leakage between the altitude 14 and 24 km is approximately equal to the integral vertical mass flow entering the region from above 24 km. Although no calculation has been made of the cross-flow at other parts of the vortex, from continuity consideration a similar small cross-flow into the vortex would be expected. In addition, no attempt has been made to calculate any possible vertical flux outside the axis of the polar-night jet stream. This downward motion would be expected to be less than inside the vortex and strongly masked by changes produced by free horizontal movement.

From a study of the synoptic characteristics on the 100 and 200 mb surfaces, Hare (1959) concluded that these levels lay on either side of a transition zone separating the stratospheric and tropospheric circulation systems. The apparent existence of the base of the polar vortex at about 14 km is in agreement with these earlier findings. The abrupt change in aerosol optical properties near the axis of the polar-night jet stream is consistent with the calculated distribution of potential vorticity by Krishnamurti (1959), Charney and Stern (1962) and McIntyre and Palmer (1983). The small cross-flow of material into the vortex inferred from the long-term changes in the aerosol distribution is not in conflict with motions based on the conservation of potential vorticity. The latter applies only for time scales of less than one week (McIntyre and Palmer, 1983). Comparison of the inferred dynamical characteristics with those estimated by other tracer observations shows good agreement.

The dynamical characteristics are also consistent with the mean flow predicted by several meridional circulation models (Brewer, 1949; Dobson, 1956; Murgatroyd and Singleton, 1961). A more recent study using an approximate Lagrangian-mean formulation (Dunkerton, 1978), has features similar to these earlier models. All show circulation patterns with convergence and descent of air over the winter pole.

The present study has been performed for a Northern Hemisphere winter season in which the vortex contained a nonvolcanic background aerosol. Under the observed conditions, the inferred vertical mass movements did not act to significantly redistribute a major fraction of the global stratospheric aerosol burden. This may not always be true. For example, a volcanic eruption such as St. Helens or El Chichón may inject new aerosol into the stratosphere that becomes trapped within the polar vortex. Under these circumstances, when the global aerosol distribution is inhomogeneous and a large fraction of the aerosol mass may lie at high northern latitudes, vertical movements within the vortex could become an important factor. In the Southern Hemisphere, where the polar vortex is known to be more intense and to last for a longer period, larger mass movement could occur. This might account at least partially for the observed lowering of aerosol optical depth reported by McCormick et al. (1981).

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