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Origin of extreme ozone minima at middle to high northern latitudes

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Abstract. Extreme ozone minima represent localized and temporally brief (several days) reductions in column ozone amounts below some chosen absolute level. Although such minima at middle to high northern latitudes are known to be primarily dynamical in origin, a remaining issue is whether heterogeneous chemical loss processes may also contribute significantly to their formation. A case in point is the record low 165 Dobson units (DU) minimum occurring on November 30, 1999, when temperatures near 30 hPa at the location of the minimum were lower than the threshold for the formation of type I polar stratospheric clouds (PSC). An examination of Polar Ozone and Aerosol Measurement III data for times surrounding the event shows that PSCs were indeed present in the vicinity where the ozone minimum was observed. However, archived data show that a similar extreme minimum of 167 DU with characteristics comparable to those of the November 30, 1999, minimum occurred on October 30, 1985, when no PSCs were present. An ensemble of 71 extreme ozone minima with amplitudes under 215 DU exhibit a nearly linear relationship between ozone minimum deviations from the zonal mean and corresponding 30-hPa temperature deviations. Such a relationship is predicted by analytic transport models which assume that vertical motions (i.e., upwelling) are responsible for the ozone minima. Temperature deviations near 30-hPa were unusually large for both the November 30, 1999, and the October 30, 1985, events, implying unusually rapid upward transport for these events. All 71 minima occur in regions where deviations from the zonal mean of 330 K potential vorticity are negative, implying an additional contribution to their formation by quasi-horizontal transport. The timescale for column ozone reductions during extreme ozone minima events is also determined and found to be at least 20 times more rapid than expected from known chemical loss processes. The data are therefore most consistent with a purely dynamical origin for extreme ozone minima in general and the November 30, 1999, event in particular. As was shown by earlier work, the basic dynamical process involves a combination of isentropic transport of ozone-poor air from the tropical upper troposphere and rapid upwelling over upper tropospheric anticyclonic disturbances resulting from poleward Rossby wave breaking events.

1. Introduction

On November 30, 1999, the Earth Probe Total Ozone Mapping Spectrometer (TOMS) instrument recorded a column ozone amount of only 165 Dobson units (DU; 1 DU = 1 matm cm) between Scotland and southern Norway (55.5°N, 3.1°E). This column amount is the lowest yet measured at northern middle to high latitudes by any TOMS instrument since measurements began in 1979. Like other ozone minima of this type, the November 30, 1999, event was associated with a strong anticyclone in the upper troposphere that results in rapid upwelling in the lower stratosphere and a southward displacement of the polar vortex toward Europe (for a review, see *Petzoldt* [1999]). A dominantly dynamical cause is therefore indicated. Several preliminary stud-

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ies including chemical transport model simulations have also suggested the dominance of transport processes in producing the November 30, 1999, minimum [e.g., *Valks et al.,* 2000]. However, as will be discussed further in section 2 below, during this event, stratospheric temperatures near 30 hPa were lower than the threshold for the formation of polar stratospheric cloud (PSC) particles. An examination of Polar Ozone and Aerosol Measurement (POAM) III data shows that PSCs were indeed present in the vicinity of the November 30, 1999, ozone minimum location. The question is therefore raised of whether increased heterogeneous chemical losses resulting from anthropogenic halogen emissions may have also contributed to this low-ozone event. A similar issue has previously been raised by *Petzoldt et al.* [1994] and *Petzoldt* [1999], who note that the same dynamical processes that produce extreme ozone minima also produce low temperatures conducive to more rapid heterogeneous chemical losses.

The relative importance of dynamical and chemical processes for producing observed ozone variability is incompletely determined on interannual and decadal timescales as well as on short timescales. At northern midlatitudes in February, as much as 40% of the observed zonal mean ozone trend over the 1979-1998 period has been attributed to long-term trends in lower stratospheric zonal winds and Rossby wave breaking behavior [*Hood et al.*, 1999]. Long-term trends in the induced mean meridional (Brewer-Dobson) circulation have also been proposed as a significant dynamical component of midlatitude ozone trends by *Fusco and Salby* [1999]. At northern polar latitudes, chemical ozone losses of 30 to 60% by March 31 of each year have been estimated by using chemical transport model simulations for the last 5 years [e.g., *Goutail et al.*, 1999; *Guirlet et al.*, 2000]. However, a substantial component of interannual variability at these latitudes is also produced by dynamical transport processes. For example, although strong ozone losses were modeled at polar latitudes during March of 1997, circulation anomalies during the same month were consistent with dynamical contributions to the observed ozone reduction [*Petzoldt*, 1999]. Therefore further work is needed to determine more accurately the relative roles of dynamics and chemistry in producing ozone reductions on all timescales.

In this paper, the relative importance of dynamical and chemical processes in producing extreme ozone minima at northern middle to high latitudes is investigated observationally in more detail by using both Nimbus 7 and Earth Probe TOMS data. In section 2, the dynamical processes that initiate the development of extreme ozone minima are described, and the November 30, 1999, event is examined as a case study. Evidence for PSC formation in the vicinity of the ozone minimum during this event is presented followed by a discussion of possible false reductions in the TOMS retrievals caused by optical effects of the PSCs. In section 3, 71 extreme ozone minima with local values under 215 DU are first identified from the 1979 to 1999 period to allow comparisons with the November 30, 1999, event. In particular, comparisons are made with the characteristics of extreme minima occurring during the month of October when no PSCs are normally present. Also, the dependence of extreme ozone minima amplitudes on 30-hPa temperature is investigated and compared with the predictions of a simple transport model. The specific dynamical circumstances that led to an unusually small local value for the November 30, 1999, minimum are then investigated. In section 4, the average timescale for the formation of extreme ozone minima is estimated and compared with the timescale for heterogeneous chemical ozone losses under the observed conditions. Final discussion and conclusions are given in section 5.

2. Dynamical Transport Processes and the November 30, 1999, Minimum

2.1. Occurrence and Distribution of Extreme Ozone Minima

Localized and temporally brief (several days) reductions in column ozone amounts with a dominantly dynamical origin have been observed in both hemispheres and have often been referred to as ozone "miniholes" [*McKenna et al.*, 1989; *Rabbe and Larsen*, 1992; *Newman et al.*, 1994; *Petzoldt et al.*, 1994; *Peters et al.*, 1995; *Peters and Waugh,* 1996; *James et al.*, 1997; *James*, 1998a, 1998b]. Extreme ozone minima represent a subset of these minihole events whose absolute magnitudes are below a chosen threshold. Unlike miniholes (see the climatology of *James* [1998a]), extreme ozone minima are most numerous in the fall and early winter, when the background mean ozone column is smaller but when planetary-scale Rossby waves in the Northern Hemisphere are also able to penetrate into the stratosphere. This is illustrated in Figure 1, which compares the mean ozone daily minima observed at latitudes >45°N using Nimbus 7 TOMS data with the time-averaged zonal mean ozone column at 50°N and 70°N. It can be seen that the most extreme ozone minima tend to be observed during the October to January period of each year. Moreover, unlike the situation for miniholes in general, the distribution of extreme ozone minima is not zonally symmetric but is strongly biased toward the North Atlantic/European region. This bias is illustrated in Figure 2, which plots the locations of all minima occurring since 1979 at latitudes northward of 45°N with amplitudes of less than 215 DU.

In part, the strong tendency for ozone minima to occur in the North Atlantic/European sector is a consequence of the fact that the monthly mean ozone column in the fall and winter seasons is relatively small in this longitude sector (see, e.g., Figure 1 of *Hood and Zaff* [1995]). The latter property is a result of mean horizontal and vertical ozone transports associated with quasi-stationary Rossby waves forced from below by land-sea

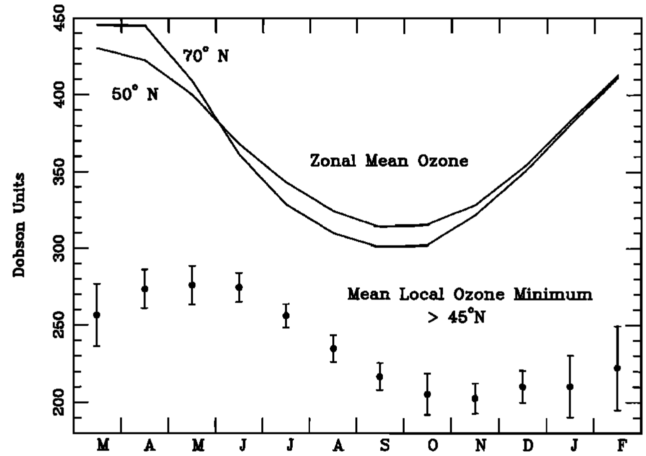

Figure 1. Zonal mean column ozone amount (solid curves) at 50°N and 60°N latitude as a function of season as calculated from 14 years of Nimbus 7 TOMS data. Solid circles and error bars are the mean and standard deviation, respectively, of the lowest column amount observed within a given month by TOMS at any location poleward of the 45°N latitude circle.

temperature contrasts and orography [*Kurzeja*, 1984; *Hood and Zaff*, 1995]. In addition, poleward Rossby wave breaking events produce transient column ozone decreases that are most numerous in this sector as well. In other sectors, equatorward breaking events are more numerous and are associated with intense tropospheric cyclones that locally increase the ozone column (see, e.g., *Olsen et al.* [2000]).

In general, Rossby waves are baroclinic disturbances of the zonal flow that amplify as a result of horizontal temperature advection. Occasionally, especially large poleward wave disturbances advect air from the tropical upper troposphere to the midlatitude lowermost stratosphere along isentropic surfaces. These poleward wave disturbances are associated with strong tropospheric anticyclones that can lead to rapid upwelling on a synoptic scale. Rossby waves "break" when a rapid and irreversible deformation of material contours occurs resulting in quasi-horizontal mixing and irreversible tracer transport [*McIntyre and Palmer*, 1983, 1984]. Poleward Rossby wave breaking events tend to occur most often when the zonal wind field in the lowermost stratosphere is either anticyclonic or only weakly cyclonic. At northern midlatitudes, this is most clearly the case in the longitude sector that includes the Atlantic and western European regions (see, e.g., Figure 21a of *Hood et al.* [1999]). In fact, the mean winter-spring wind shear is anticyclonic in the region 30°–45°N, 15°–45°W, where most poleward Rossby waves that lead to extreme ozone minima are initiated.

Poleward Rossby wave disturbances in the North Atlantic/European sector produce ozone column reductions at northern midlatitudes by two basic processes. First, as was noted above, ozone-poor air from the tropical upper troposphere is advected horizontally along isentropic surfaces into the midlatitude lowermost stratosphere (see, e.g., Figure 2 of *Hood et al.* [1999]). Second, extreme ozone minima (or miniholes) occur on a more synoptic scale in association with strong anticyclones generated by these same wave disturbances. As was pointed out originally by *Petzoldt et al.* [1994], the formation of extreme ozone minima above strong tropospheric anticyclones is primarily a consequence of uplift of advected air in the subtropical jet as it encounters a higher, colder tropopause. Adiabatic cooling resulting from the uplift results in lower temperatures at higher (~30 hPa) levels. Simultaneously, an ozone column reduction results by continuity since ozone-poor air converges below while ozone-rich air diverges above. In some cases, the amplification of Rossby waves on a planetary scale disturbs the vortex, produces a polar warming in the upper stratosphere, and displaces the vortex toward northern Europe. If the anticyclone and the displaced vortex overlap, the uplift above the anticyclone as the westerly jet encounters upward sloping isentropes can be particularly rapid. The most extreme minima are observed in such situations [*Petzoldt*, 1999]. If a cutoff anticyclone forms following a breaking event, a persistent blocking situation can develop, resulting in a significant regional anomaly in the monthly mean column ozone field.

To a first order, the asymmetry in the northern lower stratospheric zonal wind field can be understood as being due to the influences of land-sea temperature contrasts and orography. Specifically, the Icelandic and Aleutian low-pressure anomalies are especially pronounced in the fall to spring epoch because of the gen-

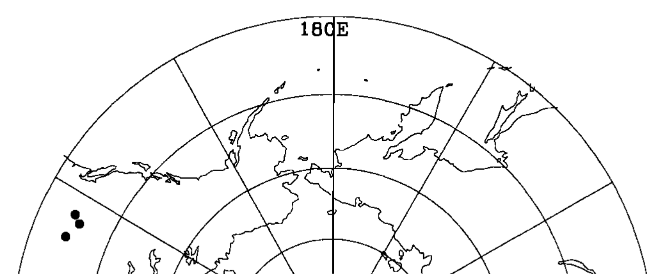

Figure 2. Distribution of extreme ozone minima with amplitudes of less than 215 DU observed by TOMS instruments poleward of 45°N latitude since 1979.

erally warmer surface temperatures in these dominantly oceanic high-latitude regions. As the associated quasi-stationary Rossby waves propagate upward, these low-pressure anomalies shift westward with height so that the low centered on Iceland at the surface is centered more nearly over northeastern Canada at the 250-hPa level. Lower stratospheric zonal winds are then accelerated across the northeastern United States and northern Atlantic, resulting in a less cyclonic wind shear regime at lower latitudes. A useful measure of the reduced cyclonic wind shear in this region is the North Atlantic Oscillation (NAO) index, defined as the difference in surface pressure across the North Atlantic (e.g., between Ponta Delgada in the Azores and Stykkisholmur in Iceland). Increases in the NAO index during the last 30 years have been shown to be related to observed ozone decreases over eastern Europe [*Appenzeller et al.*, 2000; see also *Hood et al.*, 1997]. Downstream of the North Atlantic region, higher-pressure anomalies associated with the colder land surface of Eurasia reduce the zonal winds at higher latitudes, producing a more cyclonic meridional shear. A similar, but less pronounced,

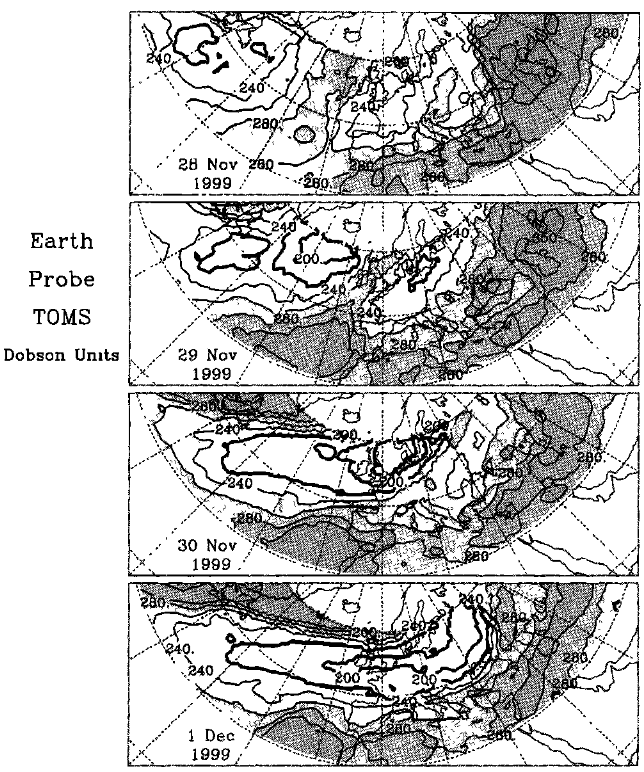

Figure 3. Contour maps of Earth Probe TOMS column ozone data for a series of days showing the development of an especially extreme ozone minimum (165 DU) that occurred on November 30, 1999. On this and subsequent maps, the Greenwich meridian is at center and the dashed lines indicate latitudes and longitudes at 15° intervals. The lowest latitude for which data are plotted is 30°N. The contour interval is 20 Dobson units (DU); contours at 220 DU and below are emphasized by a thicker line.

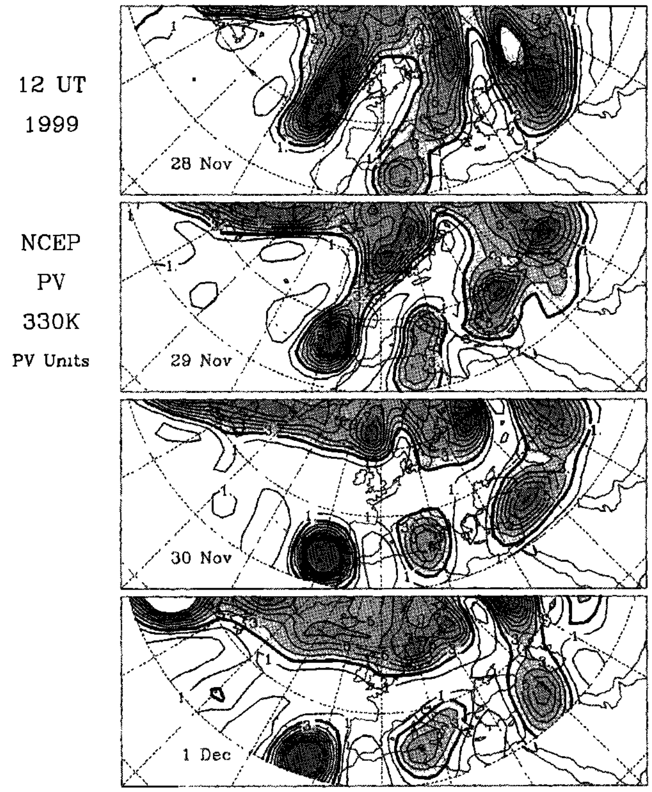

Figure 4. Contour maps of Ertel's potential vorticity (PV) at 1200 UT on the 330 K potential temperature surface in PV units (1 PV unit = 10^{-6} K m^2 kg^{-1} s^{-1}) for the same series of days shown in Figure 3. In this figure, the PV = 2 contour, which roughly follows the tropopause boundary, is emphasized by a thicker line. Values greater than 2, representing stratospheric air, are shaded, while values less than 2, representing tropospheric air, are unshaded.

reduced cyclonic shear regime occurs over the eastern Pacific associated with the Aleutian low and the North American high. Finally, as was argued by I.N. James [*James*, 1994], the strongest zonal wind maxima in the northern winter occur downstream of the two most important sources of orographic wave forcing, the Tibetan plateau and the Rocky Mountains. Thus the accelerated winds across the northern Atlantic that lead to a less cyclonic shear are also influenced by orography.

2.2. November 30, 1999, Minimum

As was noted in the introduction, the November 30, 1999, minimum of 165 DU at 55.5°N, 3.1°E was the smallest daily ozone column observed by any TOMS instrument in the northern extratropics during the 1979 to 1999 period. We therefore select this event to illustrate some of the above described characteristics. Figure 3 shows the Earth Probe TOMS data in the region of interest for a 4-day period including the observed minimum on November 30, 1999. For comparison, Figure 4 shows Ertel's potential vorticity (PV) on the 330 K surface calculated from National Centers for Environmental Prediction (NCEP) reanalysis data [*Kalnay et al.*, 1996]. The 330 K PV was computed on a 2.5° × 2.5° latitude/longitude grid by using methods described elsewhere (see, e.g., *Hood et al.* [1999]). For the purpose of a more direct comparison with TOMS data, which are acquired near local noon, we have used 6-hour NCEP data for 1200 Universal Time (UT) rather than daily averages.

Comparing Figures 3 and 4, it is seen that the November 30, 1999, event is initiated when a large poleward breaking Rossby wave injects a tongue of low-PV, upper tropospheric air from the subtropics in a northeastward direction toward Europe. On November 30 (third panel in each figure), a broad region of low-PV air extends up to 58°N at longitudes that include the location of the extreme ozone minimum. An examination of PV maps at 6-hour intervals on November 30 shows that an injection of low-PV air across the ozone minimum location (55°N, 3°E) occurs just near 1200 UT, when the backscattered radiance measurements were acquired by the Earth Probe TOMS instrument.

A close comparison of the third panels of Figures 3 and 4 shows that there is no clear minimum in 330 K PV at the location where the extreme ozone minimum occurs. There is only a broad zone of low PV (<1 PV unit) in the general region of low ozone. Other zones of similarly low PV on other dates (e.g., near 55°N, 45°E on December 1) are not characterized by such low column ozone amounts. Therefore direct horizontal transport of ozone-poor air from lower latitudes alone cannot easily explain the observed extreme ozone minimum of 165 DU.

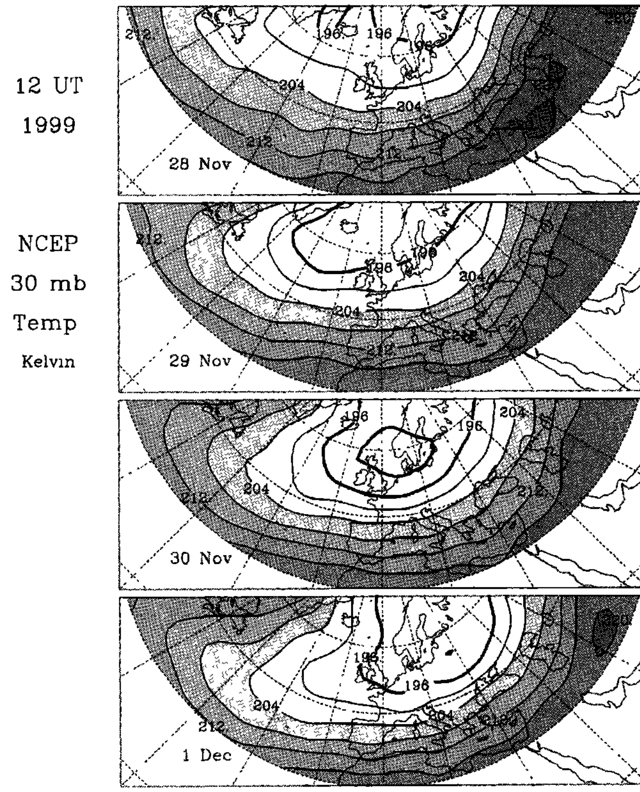

Figure 6. Contour maps of NCEP 30-hPa temperature at 1200 UT for the same series of days shown in Figures 3–5. The contour interval is 4 K, and contours of 196 K and less are thicker for emphasis.

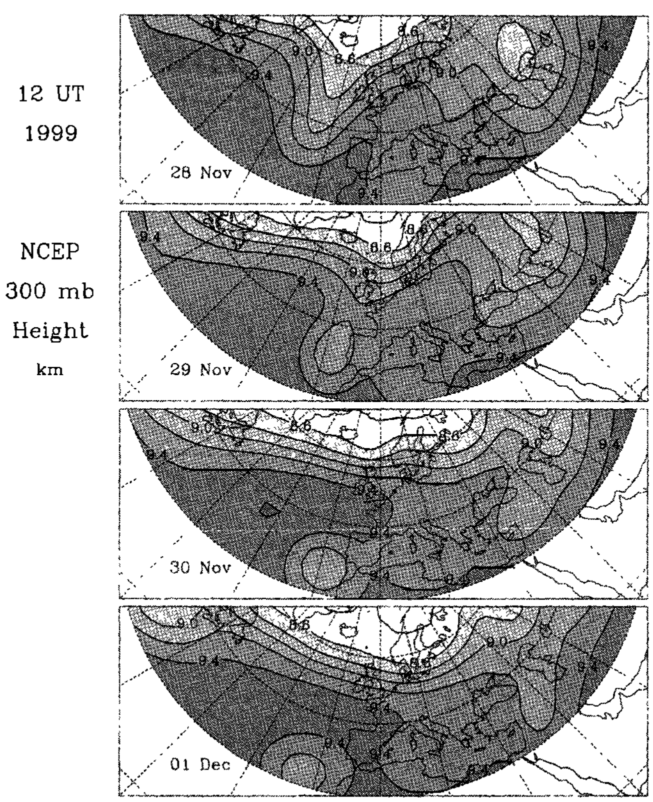

Figure 5. Contour maps of NCEP geopotential height at 1200 UT on the 300-hPa surface for the same series of days shown in Figures 3 and 4. The contour interval is 0.2 km.

As was noted above, previous work by *Petzoldt et al.* [1994] and others has shown that the most extreme ozone minima occur when a strong anticyclone develops in the upper troposphere followed by uplift of advected air and a southward displacement of the polar vortex. The signature of the upper tropospheric anticyclone is an increase in geopotential height near 300 hPa, while the most direct signature of vertical advection is a temperature reduction near the 30-hPa level. As is shown in Figure 5, the November 30, 1999, event is characterized by an increase in 300-hPa height as subtropical air of upper tropospheric origin is advected northeastward toward Europe. In addition, as is shown in Figure 6, a distinct temperature minimum of approximately 190 K at 30 hPa occurs on November 30, 1999. The temperature contours on successive days are consistent with a southward shift of the vortex with a maximum displacement reached on November 30, 1999. The near-spatial coincidence of the temperature minimum and the ozone minimum is shown in higher resolution in Plate 1. The most direct interpretation of the correlation shown in Plate 1 is that both the local ozone and temperature

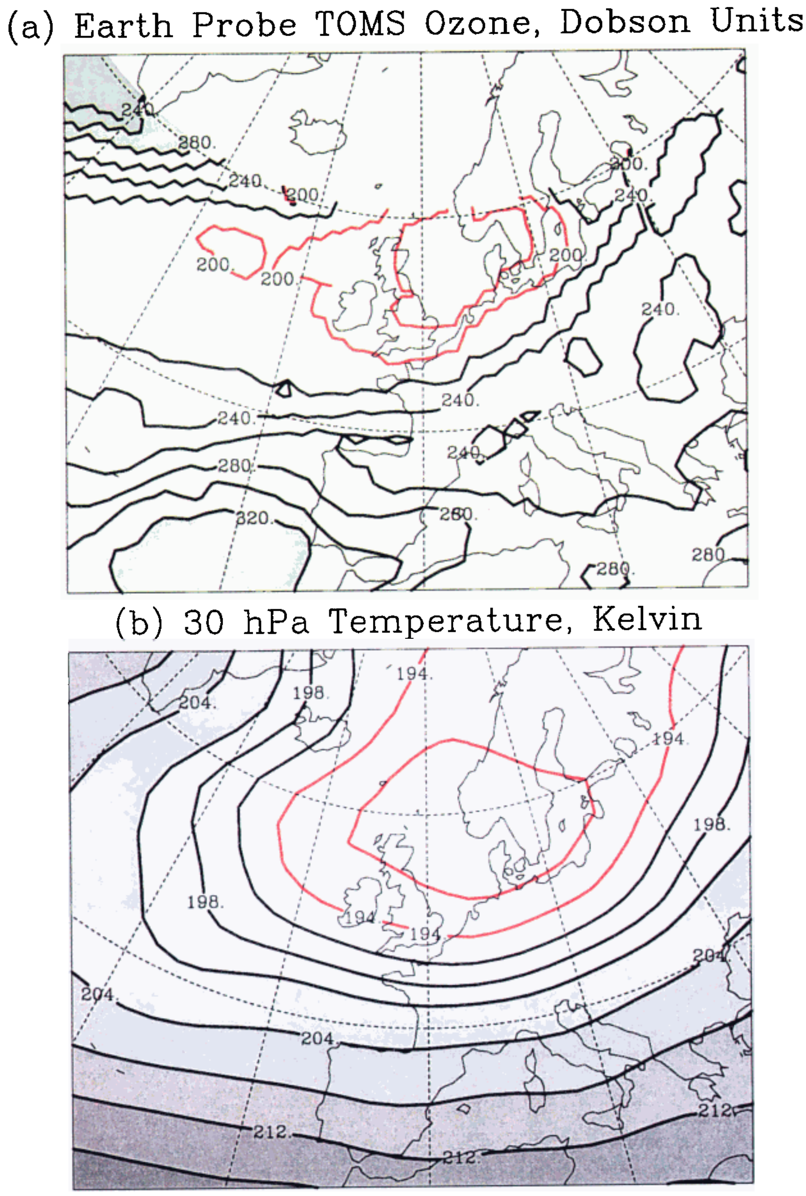

Plate 1. Comparison of (a) Earth Probe TOMS column ozone in Dobson units with (b) NCEP 30-hPa temperature at 1200 UT on November 30, 1999, for a region centered on the extreme ozone minimum at 55°N, 3°E. The contour intervals are 20 DU and 4 K (temperatures greater than 200 K) or 2 K (temperatures less than 200 K). In Plate 1a, contours of 200 DU and less are emphasized in red, while in Plate 1b, contours of 194 K and less are emphasized.

minima are a consequence of vertical transport of ozone and simultaneous adiabatic cooling.

2.3. PSC Aerosol Formation Near the November 30, 1999, Ozone Minimum

As is shown in Plate 1, the temperature minimum at the time of the ozone minimum is less than the threshold (192.6 K at 30 hPa; 195 K at 50 hPa) at which type I (nitric acid trihydrate) PSC aerosol may form [e.g., *Hanson and Mauersberger*, 1988]. Remote sensing observations demonstrate that type I PSCs do form in the 22- to 26-km altitude range at northern latitudes comparable to those at which extreme ozone minima are found [e.g., *Carslaw et al.*, 1998]. It is therefore possible that heterogeneous chemical ozone losses on type I PSCs are also contributing significantly to the formation of extreme ozone minima such as that on November 30, 1999.

For the purpose of investigating whether PSC aerosol was present within the region of collocated 30-hPa reduced temperatures and reduced column ozone during the November 30, 1999, event, we have examined Polar Ozone and Aerosol Measurement III data [*Lucke et al.*, 1999]. The POAM instrument is a visible/near-IR photometer operating in a solar occultation mode. The vertical resolution is 1–2 km; because of the viewing ge-

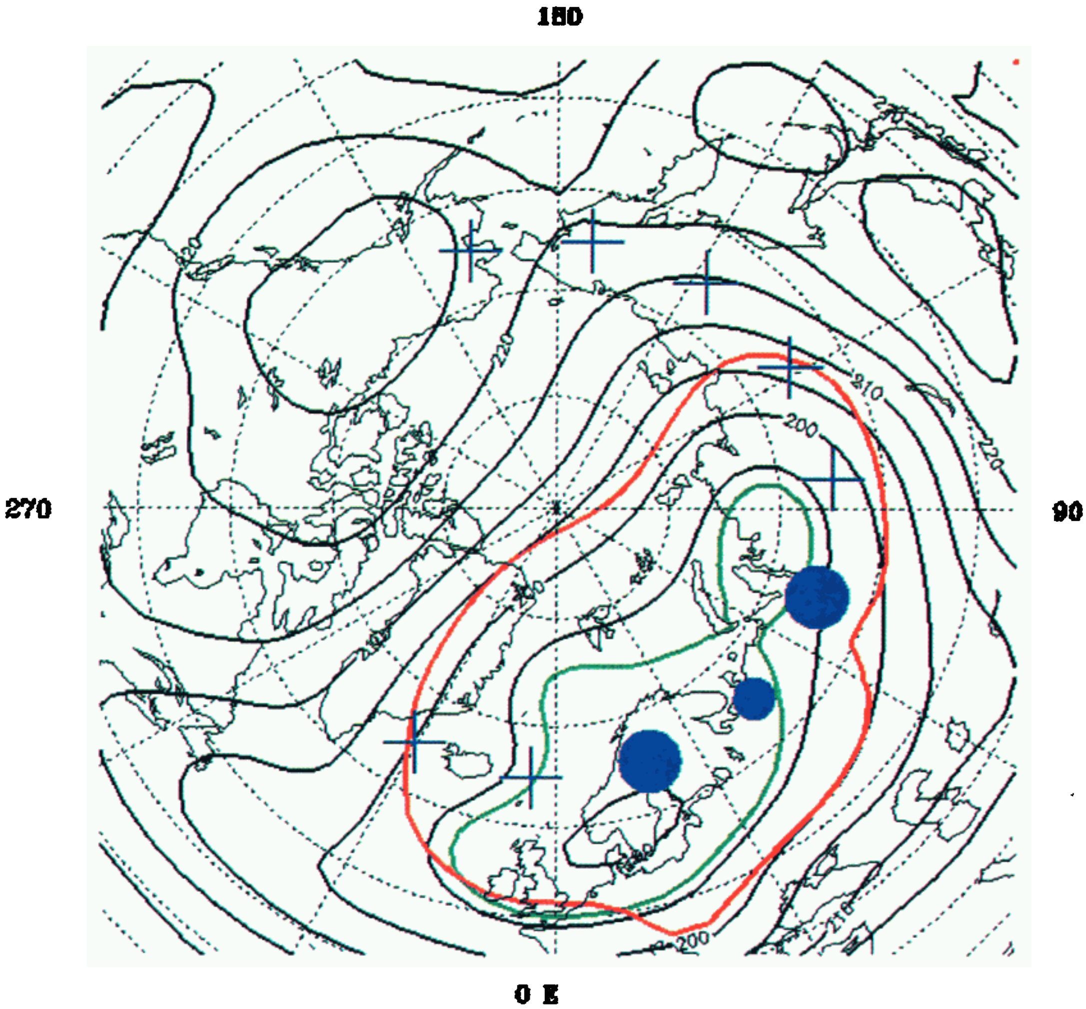

Plate 2. Comparison of POAM II measurements of PSC aerosol enhancements with 30-hPa NCEP temperatures on November 30, 1999. Locations of POAM measurements where no PSCs were detected are indicated by crosses. Locations of moderate aerosol enhancements are indicated by small blue solid circles, while locations of strong enhancements are indicated by large blue solid circles. The temperature contour interval is 5 K. The location of the polar vortex boundary on the 650 K surface, as determined by the method of *Nash et al.* [1996], is indicated by the red outline. The approximate threshold temperature for type I PSC formation (~193 K) is indicated by the green outline.

ometry, the horizontal resolution is about 200 km along the instrument line of sight, which is generally in the north-south direction, and 30 km perpendicular to the line of sight. POAM III is in a polar orbit and samples the high-latitude Northern Hemisphere 14 times per day around a nearly constant line of latitude. During the November 30, 1999, event, the measurement latitude was about 65°N. Although this is higher than the latitude of the ozone minimum (55°N), some inferences on the likely presence of PSCs should be possible from the combination of both the POAM data and the NCEP 30-hPa temperature data.

An examination of the POAM data shows that PSC aerosol enhancements were present in the longitude sector of the developing ozone minimum on November 28, 1999, and on November 30, 1999 [*Fromm et al.*, 1999]. POAM profiles in this longitude sector are not available for November 29, 1999. No aerosol enhancements were detected on December 1. The POAM PSC enhancements observed on November 28 and 30 are what *Fromm et al.* [1999] called "layer PSCs," which are distinct from so-called "HiZmin PSCs," interpreted by the authors as type II PSCs. We examined the layer PSC profiles of November 28 and 30 closely, in terms of extinction enhancement and the wavelength dependence of the aerosol extinction. We find marginal (defined as between 2.7 and 4 standard deviations above the normal background) enhancements as well as moderate enhancements (in excess of 4 standard deviations) and a considerable wavelength variation in these layer PSC observations. We interpret them to be type I (nitric acid trihydrate) PSCs, consistent with the fact that observed temperatures are higher than the formation temperature of type II PSC's (188 K at 30 hPa). On November 28, PSC enhancements were observed near 36°E and 62°E on the POAM measurement latitude circle at 65°N. Comparing these locations with the 30-hPa temperature map for this date (top panel of Figure 6), it is seen that PSCs are present roughly within the 196 K contour, which extends in latitude down to approximately 60°N. Therefore accepting 196 K as an upper limit on the threshold temperature for type I PSC formation in the 30- to 50-hPa pressure range, it is unlikely that the observed PSCs extended south of 60°N on November 28. On November 30, the date of the observed extreme ozone minimum, marginal to moderate PSC enhancements were observed on three consecutive POAM passes between 20°E and 70°E. This is illustrated in Plate 2, which compares POAM measurements with 30-hPa NCEP temperatures. The locations where PSCs were observed are roughly consistent with the region where 30-hPa temperatures are less than 193 K. This region extends southward to approximately 50°N along the Greenwich meridian and includes the location of the extreme ozone minimum observed on the same date. Therefore it is reasonable to suggest that type I PSC aerosol existed in the region of the extreme ozone minimum centered on 55°N, 3°E on this date.

Further confirmation of the presence of PSCs in the general vicinity of the November 30, 1999, minimum is obtained in the form of visual sightings near Kiruna, Sweden (68°N, 21°E), and over Oslo, Norway (59.9°N, 10.7°E) (S. Eckermann, M. Schoeberl, and G. Braathen, private communications, 2000). By December 1, however, the region of very low temperatures at 30 hPa was reduced in size, and no PSC aerosol was detected by POAM III. By this date, the main mass of transport-induced, ozone-poor air in the lowermost stratosphere had moved 30° eastward and was no longer beneath the coldest part of the vortex.

2.4. Possible PSC Effects on TOMS Retrievals

If a PSC with significant optical thickness ($\tau > 0.1$) is at an altitude even with or just above the ozone maximum, then incoming solar radiation could be reflected back to the TOMS instrument without absorption by ozone below the cloud level. This could result in an erroneously low column ozone retrieval that would be interpreted as an ozone minihole. This error source has been studied in detail by *Torres et al.* [1992], who find that the largest errors (~30 DU) are produced by type III PSCs (water ice; $\tau = 0.4$) at solar zenith angles greater than 85°. For type I PSCs ($\tau = 0.04$), retrieval underestimations are approximately 4 DU for solar zenith angles >84°. For the specific case of the November 30, 1999, event, no significant error is expected, since solar zenith angles were not greater than 84° and the PSCs in question were type I (P. K. Bhartia and O. Torres, private communication, 2000). However, for some extreme minima occurring later in the winter at very high latitudes where type III PSCs are more likely to be present, significant errors may occur.

3. Comparisons With Other Extreme Ozone Minima

Table 1 lists dates, locations, and values (or "amplitudes") of 71 extreme ozone minima observed by TOMS instruments on a series of spacecraft platforms during the months of October, November, December, January, and February over the period from 1979 to 1999. Specifically, Nimbus 7 TOMS data are used for the 1979 to 1992 period; Meteor TOMS data are used for the 1993 to 1994 period, and Earth Probe TOMS data are used for the 1996 to 1999 period. (No TOMS data are available for 1995.) To allow more direct comparisons with the November 30, 1999, minimum, which occurred at 55.5°N, 3.1°W, only those minima occurring at latitudes >55°N and at longitudes between 30°W and 30°E are listed in the table.

3.1. October 30, 1985, Event

A scan of Table 1 shows that, while the November 30, 1999, minimum has the smallest recorded value of 165 DU, 22 other minima have occurred with values of less than 200 DU. Of these, six minima had amplitudes of less than 190 DU, three had values under 180 DU, and one had a value under 170 DU. The latter minimum with an amplitude of 167 DU is only slightly higher than that of the November 30, 1999, minimum and occurred on October 30, 1985. As is shown in Plate 3, the October 30, 1985, minimum is also characterized by a temperature minimum near 30 hPa that correlates well geographically with the ozone minimum. This correlation is consistent with a predominantly dynamical origin involving rapid upward transport. Like the November 30, 1999, minimum (Plate 1), the temperature minimum is centered over Norway; this may reflect an additional upwelling effect due to air flow over Norway's mountain chain. Unlike the November 30, 1999, minimum, the lowest temperature of ~199 K seen in Plate 3 is still well above the threshold for PSC aerosol formation. In fact, no observations of PSC aerosol have yet been reported in the Arctic during the month of October because temperatures in the lower Arctic stratosphere virtually never decrease below 195 K. An examination of Stratospheric Aerosol Monitor II data for October 30, 1985, by one of us (M. Fromm) confirms that no PSCs were detected on this date. Consequently, it is reasonable to assume that the October 30, 1985, ozone minimum has no measurable chemical component and is entirely dynamical in origin.

3.2. Simple Ozone Transport Model

According to the physical model of extreme ozone minimum formation discussed in section 2.1, the ozone

Table 1. Observed Ozone Minima (>55°N, 30°W to 30°E) in October to February During the 1979 to 1999 Period

Year	Month	Day	Latitude	Longitude	O_3 Column, *DU*
1980	Oct.	15	66.5°N	29.4°W	212
1983	Oct.	10	77.5°N	4.4°W	215
1983	Oct.	25	56.5°N	1.9°W	196
1985	Oct.	7	78.5°N	30.7°E	214
1985	Oct.	30	60.5°N	13.1°E	167
1988	Oct.	17	66.5°N	30.7°E	211
1988	Oct.	28	65.5°N	21.9°W	209
1989	Oct.	6	68.5°N	15.6°E	213
1989	Oct.	31	69.5°N	28.1°E	202
1991	Oct.	11	67.5°N	25.6°E	197
1992	Oct.	16	74.5°N	9.4°W	199
1994	Oct.	6	67.5°N	1.9°W	208
1994	Oct.	10	72.5°N	20.6°E	208
1994	Oct.	25	71.5°N	25.6°E	215
1996	Oct.	18	68.5°N	20.6°E	201
1996	Oct.	24	62.5°N	13.1°E	209
1997	Oct.	1	58.5°N	24.4°W	215
1997	Oct.	16	63.5°N	30.6°W	215
1997	Oct.	22	68.5°N	26.9°W	212
1999	Oct.	28	66.5°N	23.1°E	193
1978	Nov.	28	57.5°N	9.4°W	210
1979	Nov.	3	63.5°N	3.1°E	192
1979	Nov.	16	59.5°N	29.4°W	210
1980	Nov.	8	62.5°N	19.4°W	215
1982	Nov.	11	63.5°N	30.6°E	208
1983	Nov.	11	59.5°N	8.1°E	205
1984	Nov.	15	59.5°N	30.6°E	205
1985	Nov.	12	65.5°N	1.9°W	215
1987	Nov.	9	62.5°N	1.9°W	193
1987	Nov.	24	62.5°N	24.4°W	184
1988	Nov.	6	57.5°N	0.6°E	198
1988	Nov.	20	59.5°N	21.9°W	198
1989	Nov.	15	65.5°N	8.1°E	198
1991	Nov.	5	62.5°N	11.9°W	208
1991	Nov.	16	56.5°N	24.4°W	214
1992	Nov.	8	65.5°N	6.9°W	211
1992	Nov.	23	57.5°N	3.1°E	195
1992	Nov.	30	57.5°N	15.6°E	195
1994	Nov.	13	57.5°N	3.1°E	198
1996	Nov.	13	61.5°N	16.8°W	190
1996	Nov.	30	56.5°N	14.4°W	178
1997	Nov.	8	65.5°N	18.1°E	208
1998	Nov.	3	63.5°N	30.6°W	210
1998	Nov.	9	63.5°N	5.6°W	206
1998	Nov.	21	57.5°N	3.1°E	207
1999	Nov.	12	63.5°N	15.6°E	210
1999	Nov.	20	64.5°N	13.1°E	196
1999	Nov.	30	55.5°N	3.1°E	165
1978	Dec.	9	55.5°N	0.6°E	213
1982	Dec.	30	55.5°N	21.9°W	208
1985	Dec.	9	56.5°N	16.9°W	215
1986	Dec.	22	56.5°N	29.4°W	203
1990	Dec.	13	60.5°N	3.1°E	205
1991	Dec.	7	57.5°N	4.4°W	194
1991	Dec.	14	58.5°N	5.6°E	203
1991	Dec.	28	59.5°N	10.6°E	201
1997	Dec.	16	60.5°N	23.1°E	206
1998	Dec.	8	55.5°N	3.1°E	202
1983	Jan.	23	63.5°N	8.1°E	209
1987	Jan.	7	56.5°N	14.4°W	204
1989	Jan.	31	62.5°N	15.6°E	176
1991	Jan.	23	59.5°N	20.6°E	208
1992	Jan.	9	56.5°N	11.9°W	205
1992	Jan.	16	59.5°N	8.1°E	190

Table 1. (continued)

Year	Month	Day	Latitude	Longitude	O_3 Column, DU
1992	Jan.	28	59.5°N	20.6°E	183
1984	Feb.	10	65.5°N	11.9°W	190
1986	Feb.	11	64.5°N	1.9°W	213
1989	Feb.	1	64.5°N	20.6°E	172
1990	Feb.	6	59.5°N	23.1°E	184
1993	Feb.	5	60.5°N	14.4°W	196
1993	Feb.	15	58.5°N	29.4°W	203
1993	Feb.	22	67.5°N	11.9°W	214

DU, Dobson units.

column reduction occurs through a combination of (1) large-scale, quasi-horizontal advection of ozone-poor air from the tropical upper troposphere into the midlatitude lowermost stratosphere during a poleward Rossby wave breaking event and (2) a more localized uplift of this ozone-poor air in the subtropical jet as it encounters an elevated tropopause over a cutoff anticyclone resulting from the breaking event. Poleward isentropic transport produces a reduction in PV that identifies the air as originating from the tropical upper troposphere (Figure 4). Adiabatic cooling of the uplifted air produces the pronounced temperature minimum near 30 hPa that correlates closely with the ozone minimum (compare Plates 1 and 3). A rough estimate for the contribution of horizontal transport to the ozone minimum can be obtained by using empirical regression relationships between 330 K PV deviations and column ozone deviations at northern midlatitudes in winter (see Figure 4 of *Hood et al.* [1999]).

In order to investigate the relationship between 30-hPa temperature deviations and column ozone deviations resulting from vertical transport, it is useful to consider a simple model in which only vertical motions are allowed (see, e.g., *Hood et al.* [1997]). Specifically, the steady state thermodynamic energy equation for an adiabatic process, when linearized about a zonal mean basic state, reduces to

$$\overline{u}T'_x + w'(\overline{T}_z + g/c_p) = 0, \tag{1}$$

where T is temperature; u and w are eastward and vertical velocities, respectively; overbars denote zonal means; primes denote deviations from zonal means; x and z subscripts denote partial derivatives with respect to the Cartesian coordinates; and g/c_p is the dry adiabatic lapse rate ($\simeq$9.76 K/km). Considering adiabatic flow and neglecting photochemical production and loss, the linearized ozone chemical continuity equation reduces to

$$\overline{u}r'_x + w'\overline{r}_z = 0, \tag{2}$$

where r is the ozone mixing ratio. Solving (1) and (2) simultaneously for r'_x gives

$$r'_x = \left[\frac{\overline{r}_z}{\overline{T}_z + g/c_p}\right]T'_x. \tag{3a}$$

Finally, integrating (3) gives

$$r' = \left[\frac{\overline{r}_z}{\overline{T}_z + g/c_p}\right]T', \tag{3b}$$

where r' is the zonally asymmetric component of the ozone mixing ratio at a given location and the constant of integration is zero, since all quantities have zero zonal means by definition. If vertical gradients in ozone mixing ratio and temperature are approximately constant over the lower stratospheric region where the column ozone amount is affected by vertical velocity changes, then, to first order, we have

$$\Omega' \simeq \mathcal{A}(y)T', \tag{4}$$

where Ω' is the column ozone deviation from the zonal mean and $\mathcal{A}$ is a density-weighted integral of $\overline{r}_z/(\overline{T}_z + g/c_p)$ over altitude. Specifically,

$$\mathcal{A} \simeq \int_{z_1}^{z_2} nA(z)dz, \tag{5}$$

where n is atmospheric number density, z_1 and z_2 are lower and upper altitude limits, and

$$A(z) = \frac{\overline{r}_z}{(\overline{T}_z + g/c_p)}. \tag{6}$$

For the purpose of allowing observational comparisons in the next section, it is of interest to estimate the value of $\mathcal{A}$ appropriate for vertical advection during the production of extreme ozone minima. Within the altitude range of interest (~17 to 27 km), $A(z)$ varies only slowly from 0.04 ppmV/K to 0.05 ppmV/K. For example, at 60°N in October at 20-km altitude, $\overline{r}_z \simeq 0.5$ ppmV/km, $T_z \leq 1$ K/km, and $A \simeq 0.05$ ppmV/K. Taking $A(z)$ outside of the integral, (5) can therefore be approximated as

$$\mathcal{A} \simeq \overline{A}n(z_1)H, \tag{7}$$

where H is the pressure scale height (~6 km in the subpolar lower stratosphere). Taking $\overline{A} = 0.045$ ppmV/K, $n(17\text{ km}) = 3.5\times10^{18}$ cm^{-3}, one obtains $\mathcal{A} \simeq 9.3\times10^{16}$ cm^{-2} K^{-1}, or $\mathcal{A} \simeq 3.5$ DU/K.

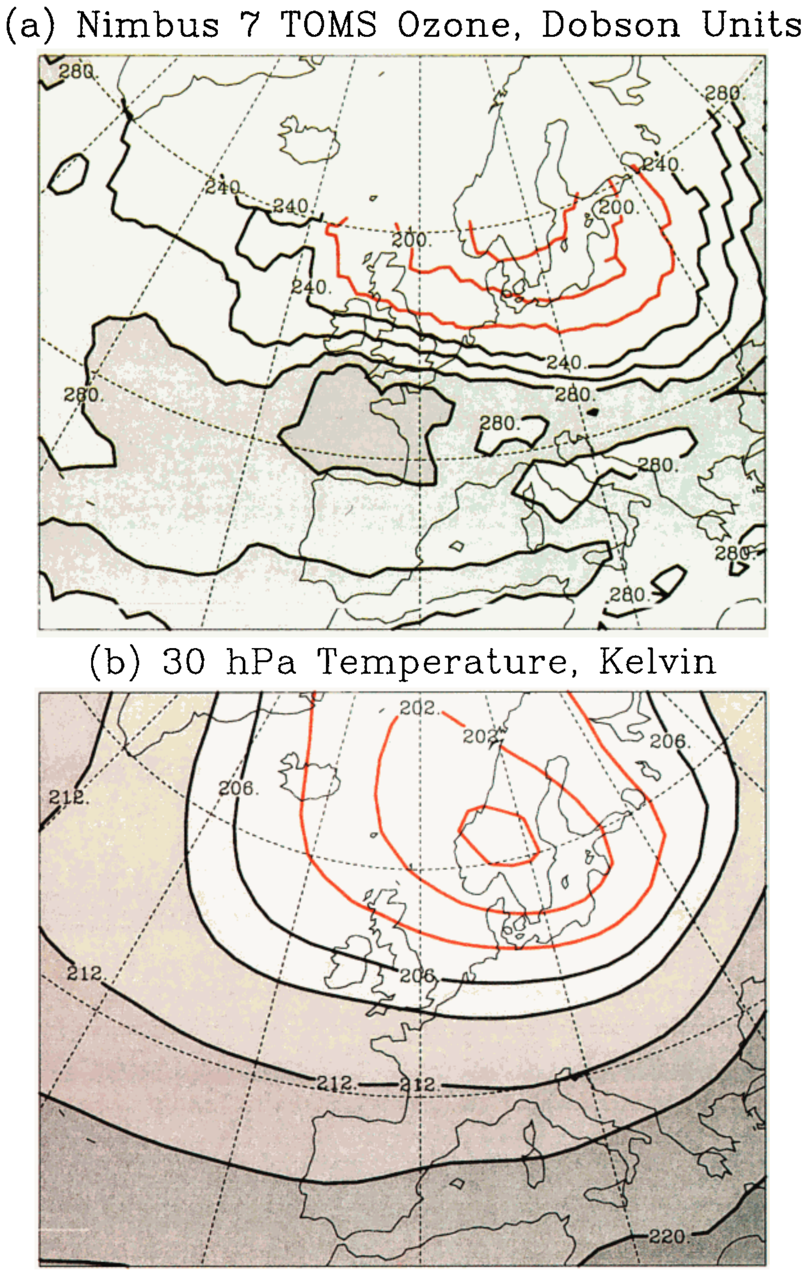

Plate 3. Same format as Plate 1 but for the extreme ozone minimum on October 30, 1985.

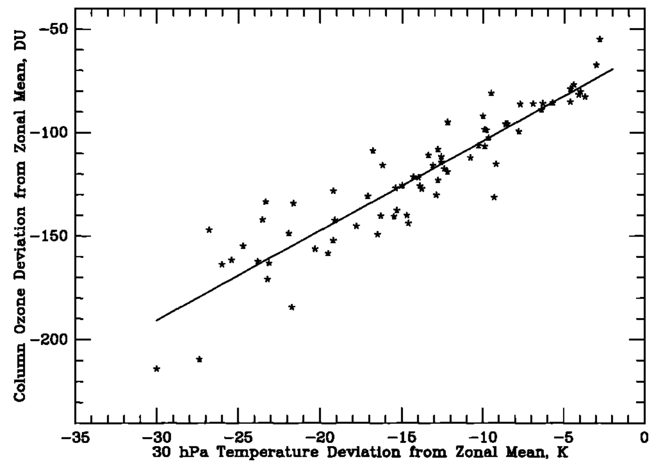

Figure 7. Plot of column ozone deviations from the zonal mean against corresponding 30-hPa temperature deviations for each of the 71 extreme ozone minima listed in Table 1. A regression line with a slope of 4.3 ± 0.3 DU/K is also shown.

3.3. Data Analysis and Model Comparisons

In order to investigate the relative importance of horizontal and vertical transports in the formation of extreme ozone minima, it is useful to plot the deviations from the daily zonal means of column ozone versus 330 K PV and 30-hPa temperature deviations on the dates and locations listed in Table 1. Figure 7 is a plot of the ozone deviations versus 30-hPa temperature deviations while Figure 8 is a plot of the ozone deviations versus 330 K PV deviations. In accordance with the predictions of equation (4), the distribution of data points in Figure 7 is approximately linear with a slope of 4.3 ± 0.3 DU/K. The proportionality between column ozone perturbations and lower stratospheric temperature deviations is supported on the broader global and climatological context by a number of studies [e.g., *Stanford and Ziemke,* 1996]. However, if only vertical advection were responsible for the extreme minima, then we would expect no correlation in Figure 8. As can be seen, although the strongest correlation is present in Figure 7, some correlation is also present in Figure 8. Specifically, Figure 8 shows that 330 K PV is usually 3–4.5 PV units lower in the vicinity of an extreme ozone minimum. Thus it must be concluded that both horizontal and vertical transports play a role in extreme minimum formation.

In order to estimate more quantitatively the contributions of horizontal and vertical advection to the formation of the ozone minima listed in Table 1, we may consider empirical regression relationships between 30-hPa temperature deviations, 330 K PV deviations, and ozone column deviations. In Figure 7, the fitted regression line has a slope of 4.3 ± 0.3 DU/K, only about 20% larger in amplitude than that (3.5 DU/K) estimated from the simplified transport model of the previous section. Considering the approximations adopted in the vertical transport model (adiabatic, steady state, photochemically inert, linearized velocities, climatological temperatures and ozone mixing ratios, etc.), the agreement is reasonable. Typical temperature deviations from the zonal mean of ∼−20 K at 30 hPa therefore imply contributions of ∼−80 DU to the column minima of Table 1. From Figure 4 of *Hood et al.* [1999], one finds that the sensitivity of column ozone deviations from the zonal mean to PV deviations at latitudes of 55°N to 70°N in winter is in the range of 20 to 25 DU/PV unit. Thus negative PV deviations of 3–4.5 units should imply transport-induced negative ozone deviations in the range of 60 to 100 DU. The ozone deviations expected from horizontal and vertical transport are therefore comparable in magnitude and together are sufficient to produce net deviations of at least ∼150 DU. Since the extreme ozone minima listed in Table 1 are

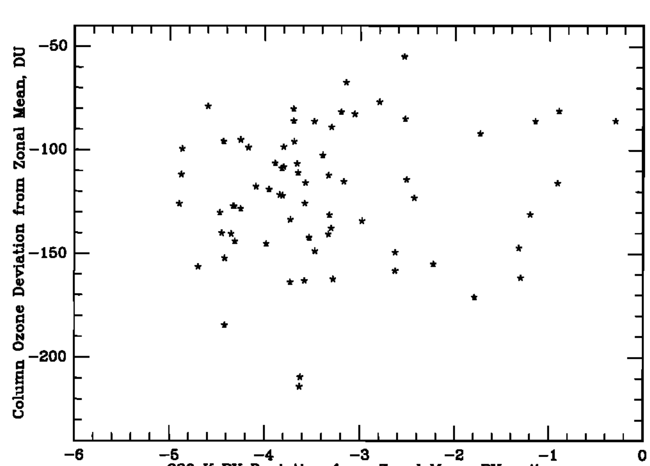

Figure 8. Plot of column ozone deviations from the zonal mean against corresponding 330 K PV deviations for each of the 71 extreme ozone minima listed in Table 1.

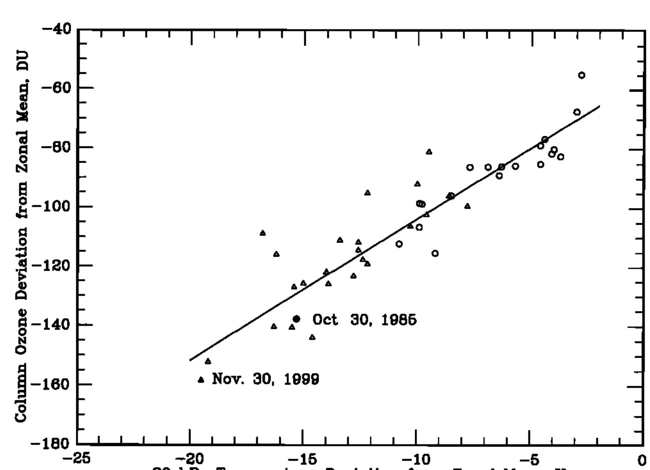

Figure 9. Same format as Figure 7 but for only those extreme ozone minima occurring during the months of October (circles) and November (triangles). A regression line with a slope of 4.4 ± 0.3 DU/K is shown.

100 to 200 DU less than the zonal mean near 60°N in the October to February period (Figure 1), the combination of horizontal and vertical ozone transports is more than sufficient to explain their low column amounts.

Figure 9 plots column ozone deviations versus 30-hPa temperature deviations only for the months of October (circles) and November (triangles). Data points for these months exhibit less variability than those for later winter months and are also less susceptible to possible retrieval errors of the type discussed in section 2.4. The data points corresponding to the November 30, 1999, and the October 30, 1985, ozone minimum events are indicated by solid symbols. A regression line derived from only the October and November data (slope 4.4 ± 0.3 DU/K) is also shown. Within the errors of the regression fit, both the November 30, 1999, and the October 30, 1985, events fall near the regression line, indicating that the combination of horizontal and vertical transport can explain the ozone minimum column amount in both cases. If significant heterogeneous chemical losses were contributing to the November 30, 1999, minimum, then a downward displacement of the corresponding data point well below the regression line would be expected. However, no significant displacement is detectable in Figure 9, and all points are consistent (within the errors of the fit) with ozone deviations driven by transport alone.

If chemical loss processes did not contribute measurably to the formation of either the November 30, 1999, minimum or the October 30, 1985, minimum, then it is necessary to ask what aspect of the dynamical transports led to unusually deep minima in both cases. We are especially interested in whether horizontal or vertical transport, or both, are responsible for these deeper minima. For this purpose, we have compiled a series of relevant quantities for all 71 cases listed in Table 1. Specifically, values at the time and location of a given minimum event, zonal averages, and deviations from the zonal mean were calculated for column ozone, 30-hPa temperature, 300-hPa geopotential height, and 330 K PV. In addition, the phase of the equatorial quasi-biennial wind oscillation (QBO) was determined for each event. An examination of these tabulated values (not listed here) shows that one unusual property characterized the November 30, 1999, and the October 30, 1985, events: The 30-hPa temperature deviation from the zonal mean was unusually large. As can also be seen from Figure 9, the 30-hPa temperature deviation for the November 30, 1999, event was the largest observed for a minimum event occurring in the month of November during the 20-year analysis period. Similarly, the deviation for the October 30, 1985, event was the largest observed for a minimum event during the month of October. In contrast, the 330 K PV deviations (representative of horizontal transport contributions) for these events were not larger than those for most other extreme minima. Since larger temperature (and ozone) deviations are expected for larger upwelling velocities, it follows that these two events were characterized by unusually strong uplift of advected air over their respective upper tropospheric anticyclones.

4. Timescale Comparisons

An additional approach toward distinguishing between a dynamically forced ozone reduction and a reduction with a significant chemical component is to compare the timescale of the observed ozone variation with that expected from heterogeneous chemical loss processes.

The expected loss rate of ozone resulting from heterogeneous chemical processes on polar stratospheric cloud particles is relatively slow. *Douglass and Stolarski* [1989] carried out a model simulation for PSCs at 65°N for January 15 conditions and estimated a total depletion of 1 to 2% over a 5-day period. *Evans* [1990] measured a trend in ozone mixing ratio near 20-km altitude of $-0.74 \pm 0.13\%$ per day at 82.5°N in January and February of 1989. This trend was attributed largely to chemical depletion, although contributions from dynamical transport could not be ruled out. *Schoeberl et al.* [1990] estimated an anomalous ozone loss rate above 20 km of $0.44 \pm 0.3\%$/d based on flight data from the 1989 Airborne Arctic Stratospheric Expedition. *Chipperfield et al.* [1996] have calculated area-weighted chemical O_3 losses within the Arctic polar vortex for a series of winters and obtained mean loss rates up to 0.3% per day. These calculations also show that cumulative ozone losses are negligible prior to January 1 and reach maximum amplitudes of more than 20%

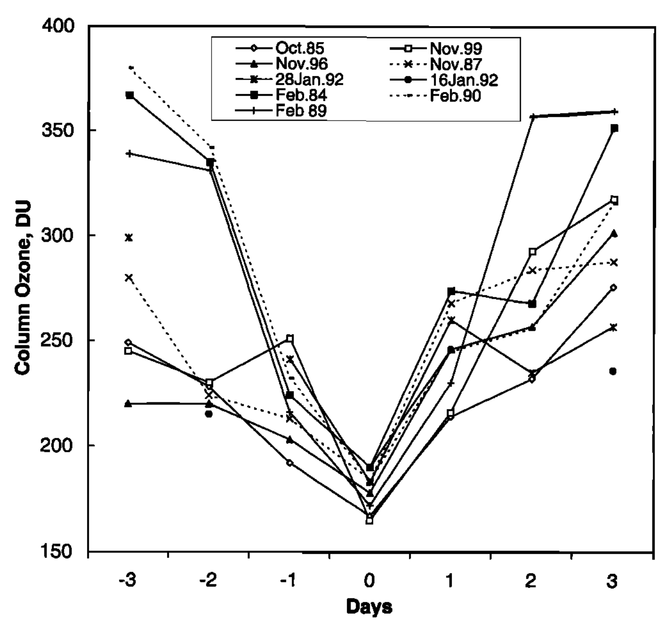

Figure 10. Time series of TOMS ozone values for dates preceding and following nine extreme ozone minima with amplitudes under 190 DU. Day 0 is the date of the minimum in each case. See Table 1 for further information.

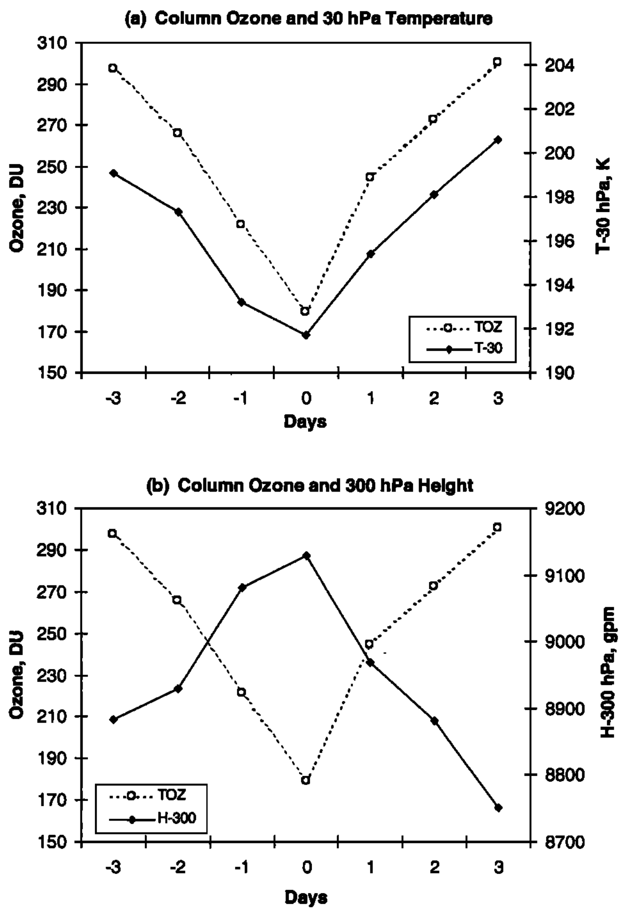

Figure 11. Mean column ozone time series for the nine extreme minima shown in Figure 10 compared with (a) mean 30-hPa temperature time series and (b) mean 300-hPa height time series.

in March. Finally, *Guirlet et al.* [2000] have recently estimated ozone losses of up to 60% poleward of 75°N in April of 2000. However, even at these high latitudes, the rate of ozone loss remains well under 0.5% per day.

Figure 10 shows time series of column ozone amounts at the locations of a number of extreme ozone minima whose values were under 190 DU. Each time series is ordered in relation to the date on which the extreme minimum was observed (day 0). The observed rates of ozone decrease range from as little as 30 DU/d to as much as 100 DU/d. Figure 11 shows comparisons of the mean ozone temporal variation during these nine selected events to the mean contemporaneous variations of 30-hPa temperature and 300-hPa geopotential height. As was discussed in section 2, the 30-hPa temperature is a close indicator of upwelling in the middle stratosphere, while the 300-hPa height is an indicator of the presence of an upper tropospheric anticyclone that leads to the upwelling. As is seen in Figure 10a, the mean ozone and temperature time series correlate closely, and the temperature minimum occurs on the same day as the extreme ozone minimum. As is seen in Figure 10b, the mean ozone and 300-hPa height time series correlate inversely with one another. The mean rate of ozone decrease prior to these extreme minima is ~40 DU/d, or about 14% per day.

From this comparison of the expected rates of heterogeneous chemically induced ozone loss (under 0.5% per day) with the observed rate of column ozone decline prior to extreme ozone minima (more than 10% per day), it is apparent that direct local chemical depletion plays a negligibly small role in the formation of these minima. Since significant (>10%) ozone losses within the polar vortex do accumulate after the beginning of January, it is possible that advection of ozone-depleted air during the southward displacement of the polar vortex contributes significantly to the formation of those minima occurring in January and February. However, this should not be the case for minima occurring in October and November (e.g., the October 30, 1985, and the November 30, 1999, minima), when cumulative ozone losses are negligible [e.g., *Chipperfield et al.*, 1996].

5. Summary and Conclusions

As was found in section 2.2, the November 30, 1999, ozone minimum has characteristics similar to those of other extreme ozone minima at middle to high northern latitudes in fall and winter. Specifically, dynamical processes associated with a poleward Rossby wave breaking event produced an anticyclonic disturbance in the upper troposphere and a southward displacement of the polar vortex toward northern Europe. Advected air in the subtropical jet was uplifted as it encountered the upper tropospheric anticyclone. The resulting region of strong upwelling simultaneously reduced temperatures at higher (~30-hPa) levels by adiabatic cooling and reduced the ozone column by upward advection of ozone-poor air. The location of this extreme minimum over northern Europe was due, in part, to the stationary wave induced mean ozone minimum in this longitude sector [e.g., *Hood and Zaff*, 1995] as well as to the tendency for poleward breaking Rossby waves to occur in this sector [*Peters and Waugh*, 1996]. However, as was found in section 2.3, POAM III data also provide clear evidence for the presence of type I PSC aerosol during the brief (several days) interval when 30-hPa temperatures were lower than the threshold for their formation. Therefore the possible significant contribution of heterogeneous chemical loss processes to the November 30, 1999, minimum may not be eliminated on the basis of the synoptic data for this event alone. This was the main motivation for the detailed analysis in sections 3 and 4. As was noted in section 2.4, it is unlikely that optical effects resulting from the presence of PSCs above the ozone maximum caused any significant false reductions in the TOMS retrievals during this event.

In section 3, a group of 71 extreme ozone minima with column amounts under 215 DU was identified to allow quantitative comparisons with the November 30, 1999, event. Of the six minima with column amounts under 190 DU, one had a value nearly as low (167 DU) as the November 30, 1999, event and occurred on October 30, 1985, when no PSCs could have been present. Like the November 30, 1999, event, the October 30, 1985, event was characterized by a deep temperature minimum of ~199 K at the location where the ozone minimum was observed. Because no PSCs were present, the collocated ozone and temperature minima are most directly explained if the local ozone minimum was caused mainly by rapid upwelling with no measurable chemical contribution.

In order to investigate whether the remainder of the 71 extreme minima may also be consistent with an origin involving horizontal and vertical transport, a simplified vertical ozone transport model was first formulated. The model predicts a linear dependence of ozone deviations from the zonal mean on temperature deviations from the zonal mean at a given level if only steady, adiabatic vertical motions are assumed to be responsible for both the ozone and temperature deviations. A plot of extreme column ozone minima deviations from the zonal mean versus 30-hPa temperature deviations for all 71 events (Figure 7) yielded a linear distribution with a slope of 4.3 ± 0.3 DU/K. This slope is consistent with the predictions of the vertical transport model when all uncertainties are considered. As is shown in Figure 8, however, all 71 ozone minima occurred in regions of negative deviations of potential vorticity on the 330 K potential temperature surface. As was discussed in section 3.3, these negative PV deviations imply an additional significant contribution to extreme minima formation from horizontal transport of ozone-poor air from the tropical upper troposphere to the midlatitude lowermost stratosphere. On the basis of empirical regression relationships between 30-hPa temperature deviations, 330 K PV deviations, and ozone column deviations, estimated contributions of horizontal and vertical transport were more than sufficient to explain the low column amounts of the extreme ozone minima of Table 1. As is shown in Figure 9, neither the November 30, 1999, event (for which PSCs were present) nor the October 30, 1985, event (for which PSCs were not present) lie significantly below the regression line as would be expected if chemically induced ozone losses were detectable. It is therefore concluded that the 71 extreme ozone minima studied here, including the November 30, 1999, event, are consistent with a purely dynamical origin involving a combination of quasi-horizontal isentropic transport and rapid upwelling over upper tropospheric anticyclones associated with poleward Rossby wave breaking events. It is noteworthy that the 30-hPa temperature deviation from the zonal mean for the November 30, 1999, event (-19.5 K) is the largest observed for any extreme ozone minimum in the month of November. Likewise, the 30-hPa temperature deviation for the October 30, 1985, event (-15.3 K) is the largest observed for any extreme minimum in the month of October. Unusually rapid upward transport was therefore most probably responsible for the unusually deep ozone minima observed during these two events. The fact that a similar deep minimum was observed on October 30, 1985, when no PSCs were present further supports this inference.

In section 4, the rate of column ozone reduction during extreme ozone minima was estimated as >10% per day. This rate is much larger than the rate of heterogeneous chemical ozone losses under present Arctic conditions when PSCs are present (<0.5% per day). Since temperatures lower than the threshold for type I PSC formation existed for only a few days around the November 30, 1999, event, it may be inferred on this basis that direct local chemical losses could have contributed no more than ~1% of the observed ~40% column ozone reduction during this event. Moreover, the November 30, 1999, event occurred too early in the year for significant ozone losses to have accumulated within the polar vortex. Therefore the southward displacement of the vortex also did not contribute measurably to the observed deep ozone minimum.

It is concluded that dynamical processes associated with poleward breaking Rossby waves in the North Atlantic/western European longitude sector are responsible for essentially all measurable ozone reductions occurring during extreme ozone minimum events. This includes the November 30, 1999, event which was characterized by unusually strong upwelling as evidenced by the observed 30-hPa temperature deviation. In a recent paper (brought to our attention by a reviewer), *James et al.* [2000] have undertaken a study with some parallels to the present one and have drawn a similar conclusion.

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