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Reactions on the surfaces of polar stratospheric clouds (PSCs) (11), which are composed of condensed phases of H₂O and HNO₃ (12), catalyze the rapid conversion of HCl and CINO₃ to labile species of inorganic chlorine in the polar vortex during winter (13, 14). Three catalytic cycles involving halogens account for ~95% of photochemical loss of O₃ in the vortex.

1) The self-reaction of ClO forming its dimer, (ClO₂)₂, followed by photolysis

![Fig. 1. Observations and reconstructions for 20 January 1992. (A) Observed mixing ratios of ClO and HCl, inferred mixing ratios for Cl* and Cl, and (B) the 24-hour mean loss rates for ozone, computed using Eq. 4 and Cl* shown in (A). The total recombination rate and contributions from the ClO-ClO, BrO-ClO, and ClO-O catalytic cycles (reactions 1, 2, and 3) are shown. All measurements taken along the ER-2 flight track, plotted against universal time (GMT). Corresponding latitudes are indicated at the top of the figure.](image)

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Chemical Loss of Ozone in the Arctic Polar Vortex
in the Winter of 1991–1992

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Fig. 2. The 24-hour mean loss rates for \( \text{O}_3 \) plotted against potential temperature \( \Theta \), for observations taken on 20 January 1992 during the descent and ascent in sunlight (solar zenith angle <80°) at 52°N. The altitudes 20, 16, and 15 km correspond approximately to \( \Theta = 500, 420, \) and 400 K, respectively.

\[
\begin{align*}
\text{ClO} + \text{O}_3 & \rightarrow \text{Cl} + \text{O}_2 + \text{O} \\
\text{Cl} + \text{O}_3 & \rightarrow \text{ClO} + \text{O}_2
\end{align*}
\]

Eq. 3 represents a minor loss of \( \text{O}_3 \), with rates accelerating in spring when the concentration of \( \text{O} \) atoms increases.

Hydrolysis of \( \text{N}_2\text{O}_5 \) on sulfate aerosols efficiently converts \( \text{NO}_3 \) (NO + \( \text{NO}_2 \)) to \( \text{HNO}_3 \), which is very stable at low sun angles (6, 17). Catalytic cycles involving \( \text{NO}_2 \), normally important at mid-latitudes, are negligible in the winter polar stratosphere (18). The catalytic cycle involving reaction of ClO with \( \text{HNO}_3 \) represents less than 5% of the total loss of \( \text{O}_3 \) for the basis of calculated values for \( \text{HNO}_3 \), corroborated by column measurements of \( \text{HOCl} \) (19) and is therefore not considered here.

The photochemical loss rate of ozone is

\[
\frac{d[\text{O}_3]}{dt} = -2[k_{\text{ClO} + \text{O}_3}[\text{Cl}][\text{O}_3] + (k_{\text{ClO} + \text{BrO}} + k_{\text{ClO} + \text{BrO}_2})[\text{BrO}][\text{ClO}] + k_{\text{ClO} + \text{ClO}}[\text{ClO}]^2]
\]

where \( k \) is the rate coefficient for the specified reaction, \( M \) denotes an air molecule, and \( R \) is

\[
J_{\text{ClO},1}[\text{ClO}]^2
\]

\[
J_{\text{ClO},1}[\text{ClO}][\text{ClO}] + k_{\text{ClO} + \text{ClO}}[\text{ClO}][\text{ClO}]
\]

the fraction of \( \text{ClO}_2 \) that photolyzes, where \( J \) is the photolysis frequency for the specified reaction. Removal rates attributable to reaction 1 decline at temperatures above 220 K, where \( R \rightarrow 0 \) (20). Loss of \( \text{O}_3 \) requires sunlight to photolyze \( \text{ClO}_2 \), \( \text{BrCl} \) (the nighttime reservoir for \( \text{BrO} \)), and \( \text{O}_3 \) (to generate \( \text{O} \) atoms).

Equation 4 allows us to compute instantaneous rates for \( \text{O}_3 \) loss along a flight track directly from measured radicals (5, 21), apart from \( R \) and \( \Theta \), which are calculated with Eqs. 5 and 7 below. To determine global loss rates of \( \text{O}_3 \), we must infer radical concentrations throughout the polar vortex from observations along the flight track, and the loss rate must be integrated over 24 hours (3).

The diurnal variations of \( \text{ClO} \), \( \text{BrO} \), and \( \text{O}_3 \) are governed by

\[
\frac{d[\text{ClO}]}{dt} = 2[J_{\text{ClO},1}[\text{Cl}][\text{O}_3] + k_{\text{ClO} + \text{O}_3}[\text{Cl}][\text{O}_3] + k_{\text{ClO} + \text{ClO}}[\text{ClO}]^2 - 2C_{\text{ClO} + \text{O}_3} + M)[\text{ClO}]^2]
\]

\[
\frac{d[\text{BrO}]}{dt} = J_{\text{BrCl}}[\text{BrCl}] - k_{\text{BrCl} + \text{BrO}}[\text{BrO}][\text{ClO}]
\]

\[
\frac{d[\text{O}]}{dt} = J_{\text{O}_3}[\text{O}_3] - k_{\text{O}_3 + \text{O}_3}[\text{O}][\text{O}_2][\text{M}]
\]

Equation 5 balances production and loss of \( \text{ClO}_2 \), Eq. 6 balances production and photolysis of \( \text{BrCl} \), and Eq. 7 balances photolysis of \( \text{O}_3 \) with recombination of \( \text{O} \) and \( \text{O}_3 \); each process has a relaxation time shorter than 1000 s. We determined the concentration of reactive chlorine corresponding to each measurement of \( \text{ClO} \) by matching observations to diurnal calculations with Eq. 5. Concentrations of \( \text{BrO} \) were about twice as high inside the vortex than outside, with little change during the winter; hence, a fixed relation was adopted between \( \text{BrO} \) and \( \Theta \) (22).

Concentrations of \( \text{ClO} \) began to increase dramatically in late December after temperatures over Asia had cooled to the threshold for \( \text{PSCs} \) (5, 9). Concentrations of \( \text{ClO} \) peaked in late January and then declined slowly in February and March. On 20 January 1992, the ER-2 aircraft departed Bangor, Maine (44°44'N, 68°47'W), at 1300 GMT, flew north at ~18-kin altitude (\( \Theta = 460 \) K), and turned south to return at ~19 km (\( \Theta = 490 \) K), with brief descents to 15 km at 65° and 52°N. Figure 1A shows data for \( \text{ClO} \) and \( \text{HCl} \) and inferred values of \( \text{Cl}^* \) and total inorganic chlorine, \( \text{Cl}_2 \), computed from observations of \( \text{N}_2\text{O}_5 \) and organochlorine gases (23). A steep decline in \( \text{N}_2\text{O}_5 \) and an increase in \( \text{ClO} \) were observed just north of Bangor (Fig. 1A) at the edge of the polar vortex (\( PV = 2.1 \times 10^{-5} \) K m\(^2\) kg\(^{-1}\) s\(^{-1}\)). Mixing ratios for \( \text{ClO} \) rose to 0.8 parts per billion by volume (ppbv) (1 ppbv = 10\(^{-9}\) mole fraction in air) near the vortex edge, increasing to 1.2 to 1.4 ppbv further inside; less than 15% of \( \text{Cl}_2 \) was converted to \( \text{Cl}^* \) outside the vortex, 60% near the edge, and 75 to 85% in the interior (24). Figure 1B shows loss rates of ozone along the flight track, computed by integrating Eq. 4 over 24 hours. Removal rates exceeded 1.4% per day: 50% attributed to reaction 1, 35% to 2, and 15% to 3.

Ozone loss rates in the polar vortex (Fig. 2) were uniform for \( 40° < \Theta < 50° \) K (16 to 20 km), declining at lower \( \Theta \). The mixing ratios of \( \text{ClO} \), \( \text{Cl}^* \), \( \text{Cl}_2 \), and \( \text{N}_2\text{O}_5 \) versus \( PV \) for \( \Theta = 470 \pm 10 \) K are shown in Fig. 3. Monotonic increases of the \( \text{Cl}^* \) value on an isentropic (constant \( \Theta \)) surface were derived for 8 of the 10 flights in the vortex; similar values for \( \text{Cl}^* \) were observed for a given \( PV \) on flights a few days apart. These results imply a homogeneous distribution for \( \text{Cl}^* \) in the polar vortex, extending over more than 4 km in the vertical and as far north as 70°N. Values of \( \text{Cl}^* \) versus \( PV \) were similar in 1989 and 1992 (3).

We computed distributions of \( \text{Cl}^* \) and daily loss rates for \( \text{O}_3 \) throughout the vortex using global meteorological fields for \( PV \) and the observed relation between \( \text{Cl}^* \) and \( PV \) (Fig. 4). Regions with elevated \( \text{Cl}^* \) were exposed to sunlight in lobes of the vortex that extended to 45°N, one of which was sampled by the ER-2. The mean ozone loss rate for the vortex region (\( PV > 2.1 \times 10^{-5} \) K m\(^2\) kg\(^{-1}\) s\(^{-1}\)) was 0.4% per day, notably lower than rates along the flight track (Fig. 1B). This difference illustrates the importance of assimilating aircraft observations in a tracer model to account for distortion of the polar vortex from axial symmetry (2, 3). Ozone loss rates found here are similar to those inferred from
measurements of ClO in the Arctic lower stratosphere made between 1 and 13 January 1992 by the Microwave Limb Sounder on the Upper Atmosphere Research Satellite (25).

After 1 February, minimum temperatures in the vortex exceeded the threshold for PSCs (5, 9), ending PSC processing. Little or no denitrification was observed (26), and Cl* and rates for ozone destruction began to decline (Fig. 5). The influence of illumination is evident in Fig. 5B: Loss rates were larger for given Cl* at lower latitudes (lower PV). However, elevated Cl* persisted longer at high latitudes, and O3 losses integrated over the winter were similar for a wide range of latitudes. About 0.7 part per million by volume (ppmv) (1 ppmv = 10^-9 mole fraction in air) of O3 (~15 to 20%) was removed by reactions 1 through 3 at Θ = 470 ± 10 K in the Arctic polar vortex in 1991–1992, similar to losses for 1988–1989 (3, 27).

The ozone loss rates in Fig. 5B represent snapshots of an isentropic surface on which air parcels move adiabatically (28). Non-adiabatic processes, such as radiative cooling, can significantly modify PV and Θ over time scales of a month, with changes in Θ of about ~70 K over the winter (29). The seasonal integral at Θ = 470 K nevertheless may approximate net ozone removal at ~18 km because O3 loss rates are uniform in both vertical (420 < Θ < 500 K) (Fig. 2) and horizontal (Fig. 5B) dimensions. Transport of O3 to Θ = 470 K during the winter makes it difficult to detect net chemical removal of 0.7 ppmv (8). Analysis of temporal variations of N2O and O3 suggest reduction of 15 to 20% of O3 at 17 km (8) and 10 to 15% between 10 and 12 km (30), consistent with chemical loss rates computed here.

Season-long observations during AASE II provided a unique opportunity to test ideas about the photochemical processes regulating radical concentrations in the polar vortex. We calculated production and loss of O3 and the evolution of ClO, HCl, NO, and NO2 from 15 December 1991 to 30 March 1992 for a hypothetical air parcel at Θ = 470 K, 65°N (PV = 2.8 x 10^-8 K m^-2 kg^-1 s^-1), assumed to circulate around the vortex every 8 days, by solving time-dependent equations (31) with a comprehensive set of chemical reaction mechanisms, reaction rates, and photolysis cross sections (20). Data from the ER-2 in December were adopted to define initial concentrations of total nitrogen oxides (NOy) (6), HCl (7), O3 (8), H2O, CH4, temperature (9), and the surface area of sulfate aerosols (10) on 15 December. Values of Cl were inferred from measured N2O (23, 26). The reaction efficiency for hydrolysis of N2O3 on surfaces of sulfate aerosols was assumed to be 0.10 (32).

Mean temperatures within the vortex (~210 K in January, ~230 K in March) were adopted from meteorological analyses for 1991–1992 (9). Temperatures below the PSC threshold were assumed to occur 1 day out of 8, from 9 to 25 January (9). Production of Cl* by reaction of HCl and ClNO2 on PSC surfaces was assumed to be rapid and stoichiometric (5, 7). Complete return of HNO3 to the gas phase was assumed upon PSC evaporation, consistent with observations for 1991–92 (26).

Calculated changes of ClO, HCl and NO (Table 1, case A) correspond closely to observations (Fig. 6), indicating accurate simulation of rates for key processes regulating ClO after the cessation of PSCs: photochemical decomposition of HNO3, which regenerates ClNO2, and reaction of Cl with CH4, which regenerates HCl. A study that included detailed air parcel trajectories to examine recovery from PSC processing reached similar conclusions (33). The integrated removal of O3 was ~19%, similar to empirical values from the ER-2 data discussed above.

What conditions would allow high concentrations of Cl* to persist into spring, as observed in the Antarctic (34)? If PSC processing persisted for another month, O3 losses would increase modestly (Table 1, case B): Photolysis of HNO3 accelerates rapidly as the season progresses, limiting

![Fig. 4](image_url)

**Fig. 4.** Distribution of the 24-hour mean loss rate for O3 (percent per day) in the north polar vortex for the 470 ± 10 K isentrope, derived with global fields of PV and the relation between Cl* and PV shown in Fig. 3, as described in the text. Contours of PV (in units of 10^-8 K m^-2 kg^-1 s^-1) and the flight track of the ER-2 (black line, with circle denoting Bangor, Maine) are shown. Latitude circles 60° and 30°N and a map of the continents are denoted by green lines.

![Table 1](image_url)

**Table 1.** Photochemical model results for ozone loss at 18 km.

<table>
<thead>
<tr>
<th>Case</th>
<th>Model conditions</th>
<th>O3 removala (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>PSCs every 8 days until 25 January</td>
<td>19</td>
</tr>
<tr>
<td>B</td>
<td>PSCs every 8 days until 26 February</td>
<td>26</td>
</tr>
<tr>
<td>C</td>
<td>Case A plus 90% denitrification at first PSC event</td>
<td>44</td>
</tr>
<tr>
<td>D</td>
<td>Case A plus twice the present Cl*</td>
<td>48</td>
</tr>
</tbody>
</table>

*aStated relative to a model without PSCs (see Fig. 6).
loss of $O_3$, after each PSC encounter. Extremely cold temperatures throughout the vortex (<195 K) in March could produce larger losses of $O_3$ if HNO$_3$ were segregated in an aerosol phase. If HNO$_3$ were irreversibly removed by the sedimentation of PSCs (denitrification), CI* would remain elevated until equinox, and $O_3$ losses would increase significantly (31, 35) (Fig. 6 and Table 1, case C). Much of the Antarctic polar vortex is denitrified because of pervasively cold temperatures (36), but only sporadic denitrification has been observed in the Arctic (26, 36). Large-scale denitrification could occur in the Arctic if there were an anomalously cold winter or if ice cloud formation were enhanced by increased concentrations of stratospheric H$_2$O. Cooling of the polar vortex could occur as a result of increasing CO$_2$ (37) or declining concentrations of $O_3$ (38). Large ozone reductions would also be expected in the Arctic without denitrification if concentrations of inorganic chlorine were to double beyond present-day values (Table I, case D). This scenario appears unlikely if present revisions (39) to the Montreal Protocol of 1987 (40) are effective in reducing inputs of chlorobarbons and bromocarbons to the atmosphere.

REFERENCES AND NOTES

1. Potential temperature, $\Theta$, is the temperature that an air parcel would reach if it were compressed adiabatically to 1000 mbar. Potential vorticity, PV, is a measure of the ratio of the absolute vorticity of a fluid layer to its depth [B. J. Hoskins, M. E. McIntyre, A. W. Robertson, Q. J. R Meteorol. Soc. 111, 877 (1985)]. For the polar stratosphere, $\Theta$ and PV are conserved typically over time scales of several weeks.


4. A vortex circulation forms in the stratosphere over each pole in winter. Air cools radiatively and descends within the vortex, creating steep gradients for tracers across the vortex boundary (M. Loniewski, J. B. Porcilea, K. R. Chan, S. E. Strahan, Geophys. Res. Lett. 17, 477 (1990)). Temperatures below the threshold for the formation of polar stratospheric clouds are pervasive in the Antarctic vortex and more localized in the Arctic (11).


8. M. H. Proffitt et al., ibid., p. 1150.


10. J. C. Wilson et al., ibid., p. 1140.


23. The recently detached parcel with high CI* in early January (2.1 x 10$^{-5}$ K m$^{-2}$ kg$^{-1}$ s$^{-1}$) is an artifact: small-scale features with high CI* and low N$_2$O characterize air with high PV (31). Deep air parcels associated with low CI*, high N$_2$O air over 100-100 km scales. Meteorological fields of potential vorticity are typically unable to resolve features on this scale (21).

24. Supported by grants from the National Aeronautics and Space Administration's High-Speed Research Program and Upper Atmosphere Program and by the National Science Foundation's Atmospheric Chemistry Program. Helpful discussions with S. R. Kawa and A. Weaver are gratefully acknowledged. We thank L. M. Atalante and D. W. Toohey for helpful discussions and for making data on CI* and BrO available.

25. W. B. DeMore and E. Tschuikowa, J. Phys. Chem. 94, 667 (1990), which gave CI*Br$_3$ = 0.75 deep inside the vortex. The use of cross sections from (20) (including absorption at longer wavelengths reported by W. B. DeMore and E. Tschuikowa, ibid., p. 5585) gives CI*Br$_3$ = 1 inside the vortex. Uncertainties in CI* have little effect on computed loss rates of O$_3$ because recombination depends mainly on observed daytime concentrations of CI* (Eq. 4).


30. M. M. Proffitt et al., ibid., p. 1150.


32. J. C. Wilson et al., ibid., p. 1140.


41. The apparently detached parcel with high CI* in early January (2 x 10$^{-5}$ K m$^{-2}$ kg$^{-1}$ s$^{-1}$) is an artifact: small-scale features with high CI* and low N$_2$O characterize air with high PV (31). Deep air parcels associated with low CI*, high N$_2$O air over 100-100 km scales. Meteorological fields of potential vorticity are typically unable to resolve features on this scale (21).

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