Stratospheric ozone destruction by man-made chlorofluoromethanes

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Cicerone, RJ
Stolarski, RS
Walters, S

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Stratospheric Ozone Destruction by Man-Made
Chlorofluoromethanes

Abstract. Calculations indicate that chlorofluoromethanes produced by man can greatly affect the concentrations of stratospheric ozone in future decades. This effect follows the release of chlorine from these compounds in the stratosphere. Present usage levels of chlorofluoromethanes can lead to chlorine-catalyzed ozone destruction rates that will exceed natural sinks of ozone by 1985 or 1990.

Chlorofluoromethanes (CF₃Cl, CF₂Cl₂, and CFC₁₂), being produced as aerosol propellants and refrigerants in large and growing amounts, and their atmospheric concentrations are increasing (1–3). These compounds have been considered valuable as tracers of atmospheric motions because they are relatively inert chemically with atmospheric lifetimes exceeding 10 years (4). Unlike CCl₄, they seem to have no natural sources or sinks in the troposphere; their lifetimes are controlled by diffusion into the stratosphere where they can be photodissociated by ultraviolet light (5). Molina and Rowland (5) have noted that this stratospheric sink for CF₃Cl is also represents a potential sink for stratospheric O₃. This is so because the photodissociation of CF₃Cl releases chlorine atoms which can catalytically destroy O₃ through reactions like those of the nitrogen oxides (NOₓ) with O₃ (5–8). Because of the great importance of stratospheric O₃ (9), we have reexamined this potential effect and its likely time evolution. We find that current CF₃Cl usage levels and trends can lead to chlorine-catalyzed O₃ destruction rates exceeding all natural sinks of stratospheric O₃ by the early 1980's. Stratospheric changes will continue long after ground-level emissions cease. For example, if emissions were curtailed now, the resultant O₃ destruction could maximize around 1990 and would remain significant for several decades. Our calculations also indicate that the CIX concentrations (the sum of the concentrations of Cl, Cl₂, and HCl) will increase significantly in the stratosphere but not in the troposphere.

Hence, critical monitoring of this problem will require measurements in the stratosphere; tropospheric observations alone will not suffice.

After release, CF₃Cl molecules diffuse upward to be photolysed by solar radiation (chiefly 175- to 220-nm ultraviolet wavelengths) in the stratosphere. On the basis of several atmospheric measurements (1–3) and known chemical properties of the chlorofluoromethanes, stratospheric photolysis appears to be the major sink for CF₃Cl (5). Several other possible sinks have been suggested but appear to be negligible (10). After release of the first chlorine atom by a solar photon, chemical reactions will probably remove the remaining chlorine and fluorine atoms from the CF₃Cl radical, temporarily forming phosgene-type molecules (5).

Once free in the stratosphere, chlorine atoms can catalyze the recombination of O₃ and atomic oxygen. The key reaction is (5–8)

$$\text{Cl} + \text{O}_3 \rightarrow \text{ClO} + \text{O}_2$$

$$\text{ClO} + \text{O} \rightarrow \text{Cl} + \text{O}_2$$

$$\text{O}_3 + \text{O} \rightarrow \text{O}_2 + \text{O}_3 \quad \text{(net)}$$

Eventually, the chlorine released from CF₃Cl should reach the ground through downward diffusion and tropospheric rainout (6–8).

We have quantitatively estimated globally averaged rates of O₃ destruction due to the chlorine atoms released in the initial photodissociation into man-made CF₃Cl. Figure 1 displays the altitude integral of these rates for three model time histories of CF₃Cl and CFC₁₂ (11): model 1, exponentially increasing with a doubling time of 3.5 years, the current pattern; model 2, exponential increase from 1960 to 1975, then constant at the 1975 rate; and model 3, exponential increase from 1960 to 1975, then immediate cessation. Figure 1 also presents globally integrated O₃ destruction rates due to the chemical reactions of oxygen alone (labeled "Chapman") and for the NOₓ catalytic cycles. These two O₃ sinks are currently believed to control stratospheric O₃ concentrations (8, 12). Models 1 and 2 yield rapidly increasing O₃ destruction rates that will equal the natural sinks by about 1982 and 1986, respectively. Model 3 shows that, if CF₃Cl emissions were curtailed now, the ensuing O₃ destruction rates would maximize around 1990 at a rate comparable to major natural cycles and would persist for several decades. Even larger effects may be possible because in our calculations we consider only the first chlorine atom released from each CF₃Cl molecule; in reality, all four halogen atoms may be freed (13).

Important parameters in calculations leading to Fig. 1 appear in Table 1. To account for vertical transport in our time-dependent, one-dimensional (altitude) model we adopted and smoothed the relatively high eddy diffusion coefficient (K) profile of Wofsy and McElroy (14). To compute photodissociation coefficients, J, for CF₃Cl and CFC₁₂ we used available laboratory...
photoabsorption data (15). We evaluated the O₃-dedestroying effect of the chlorine oxides (ClOₓ) by using the photochemical reaction scheme of Stolarski and Cicerone (6) after adding the reactions ClO + NO → Cl + NO₂ and H₂ + Cl → HCl + H (16). In the calculations we assumed an atmospheric background of O, O₂, CH₄, OH, NO, and H₂ that was constant in time.

A constant O₃ profile representing present conditions is probably not consistent with the large O₃-deestroying rates of ClOₓ of Fig. 1, but through its use we can compare the effect of the ClF,Clₓ to the present natural O₃ sinks. Other procedures and input data are discussed in (17).

The accumulation (mixing ratio) of CFₓ,Clₓ molecules is shown as a function of altitude and time in Fig. 2, predicted from emission history model 2. Corresponding profiles are also shown for CIX. Because the relevant photochemical processes are fast relative to the transport processes, one may calculate the amounts of Cl, ClO, and HCl present at each altitude from the CIX concentration and the equations of photochemical equilibrium (6–8). The time evolution of the CIX profiles show large increases above 20 km and virtually no change below 15 km. The precise CIX concentration predicted for a given altitude and time depends on several computational parameters, especially K, but the profile shapes clearly indicate the effects of a mid-stratospheric source (see the Qᵥ values in Table 1), slow transport below 25 km, and eventual loss to the ground.

The 1985 chlorine production rate, Qᵥ for 1985 in Table 1, is 5 × 10⁸ cm⁻² sec⁻¹, integrated over altitude. Under steady-state conditions this figure will grow to 2 × 10¹⁰ cm⁻² sec⁻¹ if only one chlorine atom is taken from each ClFₓ,Clₓ and model 2 (constant) emissions are assumed. This man-made CIX source is comparable to the projected NOₓ input due to large supersonic transport (SST) fleets, and it appears that ClOₓ destroys O₃ more efficiently than NOₓ (6). McElroy et al. (12) find that O₃ depletion from SST NOₓ is sensitive to the injection altitude. The O₃ effect due to added CIX will be difficult to evaluate, but current models (6–8) indicate that the addition of 25 km is much more important than that added below.

We have not attempted to predict O₃ concentrations or their changes due to increasing amounts of CIX from ClFₓ,Clₓ usage. Predictions of O₃ concentrations and trends due to perturbations are still the subject of considerable debate (18). However, Fig. 1 indicates that, regardless of the precise magnitude of the effect of increasing destruction rates on O₃ content, chlorine atom production from present or exponentially increasing ClFₓ,Clₓ concentrations will eventually convert the O₃ layer from NOₓ to ClOₓ control. Uncertainties in the initial CIX profile, conversion rates to chlorine, or catalytic efficiency do not alter the basic conclusion, only the rate of passage to ClOₓ control. Studies should be pursued with more complex interactive photochemical models to assess the impact of these large predicted increases in O₃ destruction on the O₃ content. In addition to continued careful monitoring of O₃, other key stratospheric trace constituents must be measured. For example, the stratospheric concentrations of OH radicals and atomic oxygen, although central to questions of stratospheric chemistry, have not, to our knowledge, been measured. The present stratospheric abundance of gaseous chlorine, natural or otherwise, must also be determined.

RALPH J. CICERONE
RICHARD S. STOLARSKI
STACY WALTERS
Space Physics Research Laboratory,
University of Michigan, Ann Arbor 48105

References and Notes
4. L. Macht, paper presented as part of the Proceedings of the 2nd International Union of Theoretical and Applied Mechanics International Congress of Geodesy and Geophysics Symposium on Turbulent Diffusion in Environmental Pollution Control, Charlottesville, Virginia, 1973; see also (1, 2, 5).
8. P. Crutzen, ibid., p. 1569; see also (5).
10. Experiments by O. C. Taylor, N. E. Hester, and E. A. Cardis (personal communication) show that soils and plants act as sinks for ClFₓ,Clₓ and ClFₓ. Other studies (N. E. Hester, personal communication) show that these compounds are remarkably stable in simulated photochemical smog. Also, possible reaction with atmospheric O(1D) atoms, even at a rate of 10⁻¹⁰ cm³ sec⁻¹, would not be a significant sink because of small O(1D) concentrations. The search for natural sinks should continue.
11. Worldwide rates of production of ClFₓ,Clₓ and ClFₓ are currently about 0.45 × 10⁸ and 0.27 × 10⁷ km⁻², respectively (R. L. McCarthy, personal communication). Between 1960 and the present, this production grew exponentially with a doubling time of 3.5 years; see also (2, 4).
13. F. S. Rowland (personal communication) tells us that all the chlorine atoms are likely to be rapidly removed from the ClFₓ,Clₓ radical.

Table 1. Input data at several altitudes: K, the only diffusion coefficient; the NO concentration; J, the photodissociation coefficients for ClFₓ,Clₓ and ClFₓ, ClFₓ, respectively; Qᵥ, the predicted release rate of chlorine atoms assuming a model 2 emission time history. The J values were computed for a solar zenith angle of 60° and then divided by 2 to represent a diurnal average. Note: 1.5(4) means 1.5 ± 0.4.

<table>
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<tr>
<th>Altitude (km)</th>
<th>K (cm² sec⁻¹)</th>
<th>NO concentration (cm⁻³)</th>
<th>Jᵥ,ClFₓ (sec⁻¹)</th>
<th>Jᵥ,ClFₓ,ClFₓ (sec⁻¹)</th>
<th>Qᵥ,ClFₓ (sec⁻³ cm⁻²)</th>
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Fig. 2. Calculated fractional concentrations (mixing ratios (volume-volume)) of atmospheric ClFₓ,Clₓ (dashed curves) and those for CIX (solid curves). Each curve in this model 2 with a rapid temporal accumulation of ClFₓ,Clₓ resulting in a similar increase of stratospheric CIX. Ground-level ClFₓ,Clₓ concentrations computed for the present are in good agreement with measured values (1–3). Also, the computed initial CIX profile is not consistent with available measurements (17).
Methane Production in the Intertidal Waters of Sulfate-Depleted Marine Sediments

Abstract. Methane in the intertidal waters of anoxic Long Island Sound sediments does not reach appreciable concentrations until about 90 percent of seawater sulfate is removed by sulfatereducing bacteria. This is in agreement with laboratory studies of anoxic marine sediments sealed in jars, which indicate that methane production does not occur until dissolved sulfate is totally exhausted. Upward diffusion of methane or its production in sulfate-free microenvironments, or both, can explain the observed coexistence of measurable concentrations of methane and sulfate in the upper portions of anoxic sediments.

Amounts of methane in the marine environment in excess of that resulting from equilibration with the atmosphere most frequently occur in anoxic waters such as those found in fjords (1, 2) and in the intertidal water of anoxic sediments (3–6). In the absence of sources associated with shipping and industrial activity or natural seeps from oil and gas reservoirs (7), most of this methane results from anaerobic bacterial decomposition of organic matter. Methane bacteria are readily found in anoxic environments, where they are terminal organisms in the microbial food chain (8); moreover, there is some evidence that methane is not produced until dissolved sulfate has been previously removed by sulfatereducing bacteria (5, 6).

The intertidal waters of Recent organic-rich marine sediments are ideal for studying the relation between methane and dissolved sulfate distributions because of the large concentration changes over short depth intervals that result from high bacterial activity. In this report we present the results of a study of methane and dissolved sulfate in the intertidal waters of Long Island Sound sediments which suggest that significant production of methane does not begin until dissolved sulfate concentrations approach zero. Results of laboratory studies of time-dependent changes in the chemistry of anoxic marine sediments indicate that sulfate reduction and methane production are mutually exclusive metabolic processes.

Gravity cores were collected at three stations in Long Island Sound. Station TH is located approximately 2 km south of the coastal town of Guilford, Connecticut, at a water depth of approximately 7 m. Stations BS and SC are located in two shallow harbors near Guilford. The water depths at these two stations ranged from 1.5 to 4 m because of tidal fluctuations. Intertidal waters were sampled without coming into contact with air by transferring sediments from sealed core liners to a filter-press type squeeze through an interlock flushed with CO\2 or He (9). Dissolved methane was measured by liquid stripping techniques developed for measuring dissolved gases in seawater by Swinnerton et al. (10) and applied to intertidal water measurements by Reeburgh (11). Dissolved sulfate was measured by gravimetric analysis as BaSO\4. Blank corrections for precipitation of nonsulfate material led to uncertainties of ±0.5 mmole liter\(^{-1}\) in dissolved sulfate concentrations.

Concentrations of methane and dissolved sulfate in the intertidal waters are plotted as a function of depth in Fig. 1. Differences in the depth of complete sulfate reduction at the three stations are probably a result of variations in sedimentation rates and the content of organic matter in the sediments. The leveling off of methane concentrations at station SC at about 1.0 mmole liter\(^{-1}\) is consistent with reaching saturation with respect to methane. The solubility of methane calculated from the Satchenow relation using solubility data from Winkler (12) and Atkinson and Richards (2) ranges from 1.1 to 2.3 mmole liter\(^{-1}\) (25 to 51 ml liter\(^{-1}\)) for the temperature range (4° to 28°C) and salinity range (27 to 31 per ml) encountered in these sediments.

Reeburgh (4) has presented evidence of methane saturation in Chesapeake Bay sediments, which results in the formation of trapped methane bubbles that strip other dissolved gases such as N\(_2\) and Ar from intertidal waters in the sediments. We infer from the low concentrations of dissolved N\(_2\) and Ar at station SC that this process may be taking place there (13).

The data in Fig. 1 show that in the intertidal waters of Long Island Sound sediments high methane concentrations do not occur unless sulfate concentrations have been appreciably lowered. Only where dissolved sulfate concentrations approach zero do concentrations of methane attain saturation.

Four alternative hypotheses can be used to explain these results. The first is that methane is produced at roughly the same rate throughout the sediment column by methane bacteria but is consumed by sulfate-reducing bacteria through the reaction:

\[
\text{CH}_4 + \text{SO}_4^{2-} + 2\text{H}^+ \rightarrow \text{H}_2\text{S} + \text{CO}_2 + 2\text{H}_2\text{O}
\]