variations previously discussed (2) for GCR-produced NO_x.

Other possible long-lived terminal negative ions such as ClO_x^- (whether natural or artificially introduced) might play a role similar to that postulated above for NO_x^-. However, it is conceivable that such ions with high electron affinities could simply reduce the NO_x^- population and thereby reduce, rather than enhance, the catalytic destruction of O_3 by negative ions.

Terminal negative ions may have an alternative sink (other than their eventual mutual neutralization with positive ions) in attachment to natural aerosols, generally thought (8) to be sulfuric acid droplets. It is known that negative ions in the stratosphere have pronounced vertical stratifications that are associated with sharp temperature inversions (9). It seems likely that the aerosol content can suppress negative ions and indirectly modify the O_3 density (10).

M. A. RUDERMAN
H. M. FOLEY
Department of Physics, Columbia University, New York 10027
J. W. CHAMBERLAIN
Department of Space Physics and Astronomy, Rice University, Houston, Texas 77001

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$$O_3^- + N_2O \rightarrow NO + NO_3^-$$

as a source of catalytic NO_x. We are very grateful to E. E. Ferguson for communicating to us the result of a recent experiment (F. C. Fehsenfeld and E. E. Ferguson, *J. Chem. Phys.*, in press), in which the rate for this reaction was determined to be less than 10^{-14} cm³/sec. Our proposal for O_3 destruction by this mechanism must accordingly be rejected.
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11. This work was supported by the National Science Foundation by grants through the Physics Section (M.A.R.) and the Atmospheric Sciences Section (J.W.C.) and by the Energy Research and Development Administration (H.M.F.).

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Stratospheric Ozone Effects on Temperature

Abstract. *Calculated surface temperature changes,* ΔT_s, *due to stratospheric ozone depletion (at 35°N latitude in April) are less than previously estimated and range between −0.6 and +0.9°K. The sign of* ΔT_s *is determined by the surface albedo and the presence or absence of a low-lying particulate layer (heating with particles, cooling without particles). The calculations indicate that a 90 percent stratospheric ozone depletion does not cause the temperature inversion at the tropopause to vanish, although it is weakened substantially.*

Since O_3 is known to be one of the major trace constituents responsible for thermal heating of the stratosphere (1) [it absorbs solar ultraviolet (190 to 350 nm), solar visible (450 to 650 nm), and infrared (9 to 10 μm) radiation], it has been suggested by Dickinson (2) and others (3) that depletion of stratospheric O_3 could have catastrophic effects on climate. This suggestion is based on coupling of the positive feedback mechanisms of ice cover (factor of 3 or greater) and conservation of relative humidity (factor of 2) with the cooling noted by Manabe and Strickler (4). While I have not explicitly included the former feedback mechanism in the study reported here, I have included the latter one. The original theoretical research relating O_3 and climatic effects was conducted by Manabe and Strickler (4). For a cloudless sky at 35°N latitude in April (considering the effects of CO_2, H_2O, and O_3 and assuming fixed absolute humidity), they calculated that complete removal of O_3 from the atmosphere would cause a decrease of less than 1°K in the earth's steady-state surface temperature, but would cause the temperature reversal that defines the tropopause to vanish. The latter effect is important since it could lead to much-enhanced vertical mixing and a dramatic shift in the mean wind field (5). Later Manabe and Wetherald (6) calculated radiative-convective thermal profiles (assuming average cloudiness and constant relative humidity) for O_3 distributions corresponding to 0°N, 40°N, and 80°N in April and found that the tropospheric temperature would decrease as the abundance of O_3 decreased and the height of the maximum O_3 concentration increased.

In the work reported here I took the Manabe-Wetherald (6) thermal equilibrium model, with three layers of water clouds (corresponding to average cloudiness at 35°N latitude in April), and introduced a low-lying layer of Mie-scattering particles in order to calculate the radiative-convective steady-state temperature profile under various conditions. The calculations were performed for a mean global surface albedo of 0.1, for the present distribution of stratospheric O_3 as well as for 90, 60, 50. 30, 10, and 0 percent levels. Additional calculations were made to test the sensitivity of these results to changes in other model parameters, including the presence or absence of particulate layers and changes in surface albedo, ω_s.

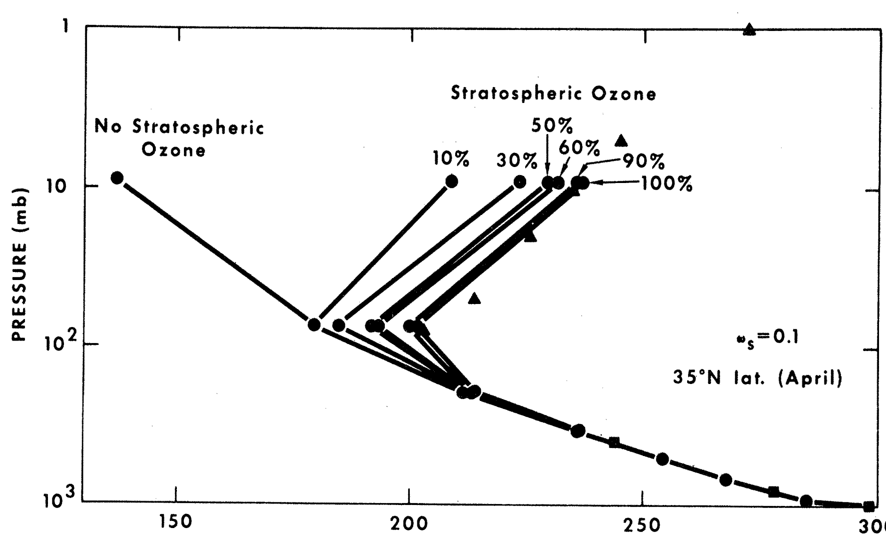

Fig. 1. Calculated steady-state temperature profiles as a function of pressure (millibars) for the present abundance of stratospheric O_3 and for 90, 60, 50, 30, 10, and ~0 percent of those values. Values from the subtropical (30°N) July model atmosphere (triangles) and experimental data interpolated for 35°N in April (squares) are included for comparison (10).

A description of the model and the parameters characteristic of 35°N latitude has been presented (*7*), and the reader is referred to the earlier work for details. The spectroscopic treatment remains unchanged, as well as the details of the method. A particulate layer having a mean visible optical density τ of 0.065 (refractive index $\eta = 1.5 - 0.1i$) was assumed (*8, 9*). It is unfortunate that a realistic global value of the imaginary component of the index of refraction is still uncertain.

The purpose of the calculations is to estimate the magnitude of the thermal effects for limited O_3 depletion, based on changes in the radiative-convective exchange. No thermal effects from reaction kinetics are included, since these require prohibitively small integration time steps. Parameterization of the atmospheric temperature profile in terms of the percentage of stratospheric O_3 depletion is considered, since only stratospheric O_3 appears to be involved. The model does not include circulation.

For reference, the assumed 100 percent and 0 percent stratospheric O_3 distributions as a function of pressure are given in Table 1. The 0 percent O_3 level is actually assumed to be 0.00002 g of O_3 per kilogram of air to avoid calculation problems which arise when zero values are used. In addition to the thermal effect of O_3 depletion, I consider the sensitivity of the thermal effect to the variation in surface albedo and cloud cover.

The results of the calculations are shown in Fig. 1 and Table 2. For comparison, Fig. 1 includes measured temperatures for April at the surface, at 800 mbar, and at 400 mbar, and six values of stratospheric temperature from the subtropical (30°N) model atmosphere for July (*10*). It is clear from Fig. 1 and Table 2 that the largest thermal effects occurred in the stratosphere.

With this model, calculated stratospheric temperatures are found to decrease at the 8.9-, 74-, and 189-mbar levels as the amount of O_3 decreases. Furthermore, the magnitude of the decrease is largest where the initial O_3 concentration is greatest (at 8.9 mbar).

As found by Manabe and Strickler (*4*), the temperature reversal is completely destroyed by complete removal of O_3 from the stratosphere. This means that the tropopause disappears independently of the water clouds and the particulate layer. However, my calculations show that the tropopause will vanish only as the last percentages of O_3 are removed. At the 8.9-mbar level, a 90 percent O_3 depletion leads to a 28°K drop in temperature from 237° to 209°K, but this is still 7° to 8°K above the initial temperature (at 100 percent O_3) calculated at the 74-mbar level.

With the present model the maximum calculated stratospheric temperature change for a 10 percent depletion is ~ 1.5°K at the 74-mbar level. The magnitude of this temperature difference is one-fourth that reported by Newell (*11*) (except for the sign difference) around the time of the major eruption of Mt. Agung, Bali (8°S, 115°E), on 17 March 1963. Hence, natural phenomena may produce stratospheric temperature fluctuations larger than that calculated for 10 percent O_3 reduction.

Turning to surface temperature differences, I find that removal of O_3, in the presence of the particulate layer, leads to heating of the surface (Table 2). The heating is the result of the combined effects of increased net solar radiation at the ground and increased water vapor in the atmosphere (Table 3). The magnitude of the surface temperature change (ΔT_s) increases as the surface albedo increases because of an increase in the amount of visible radiation being reflected and effectively "trapped" by the particulate layer. For total removal of stratospheric O_3, mean global surface albedo (0.1), and constant absolute humidity, the present model leads to a 0.1°K temperature increase (Table 2). If the calculation is repeated, assuming constant relative humidity (with no change in the particulate refractive index), the calculated surface temperature increase is 0.6°K. This positive temperature difference is in contrast with the results of Manabe and Strickler (*4*), who reported a temperature decrease of less than 1°K for total O_3 removal with a cloudless sky and no particulate layer. Hence, the results seem to indicate that the sign of ΔT_s due to O_3 depletion is determined by the presence or absence of absorbing particles in the atmosphere. I therefore repeated my calculations with the particulate layer removed from the model, and found a calculated surface temperature decrease in the presence of average cloudiness of

Table 1. Atmospheric O_3 distribution.

Pressure (mbar)	Mixing ratio of O_3 (g/kg): For 100% stratospheric O_3	For 0% stratospheric O_3
8.9	0.010167	0.000020
74.1	0.001560	0.000020
188.6	0.00040	0.000020
336.1	0.000197	0.000020
500.0	0.000119	0.000119
663.9	0.000069	0.000069
811.4	0.000050	0.000050
925.9	0.000039	0.000039
991.1	0.000038	0.000038

Table 2. Atmospheric temperature changes due to 100 percent stratospheric O_3 depletion (in the presence of three layers of water clouds and a low-lying particulate layer).

Pressure (mbar)	ΔT (°K), constant absolute humidity, ω_s = 0.1	ΔT (°K), constant relative humidity: ω_s = 0.07	ω_s = 0.1	ω_s = 0.6
8.9	−100.0	−100.0	−99.2	−85.8
74	−21.8	−20.8	−20.8	−17.5
188	−1.2	−3.2	−3.0	−2.3
336	+0.1	+0.5	−0.5	−1.8
500	+0.1	+0.5	+0.5	+0.6
663	0.0	+0.5	+0.6	+0.8
811	+0.0	+0.5	+0.6	+0.9
926	+0.1	+0.5	+0.6	+0.9
991	+0.1	+0.5	+0.6	+0.9

Table 3. Changes in net solar (downward) radiation, atmospheric water vapor content, and temperature near the earth's surface, due to 100 percent stratospheric O_3 removal, given for two extremes of surface albedo: $\omega_s = 0.07$ (dark soil) and $\omega_s = 0.6$ (snow). The change in net solar radiation is less over snow because more than 50 percent of the radiation striking the snow surface is reflected. The change in water vapor content is less over snow because there is less water vapor due to the lower temperature. The change in surface temperature is greater over snow because the low-lying particles absorb the back-reflected solar energy.

Surface albedo	Change: Net solar radiation (erg cm^{-2} sec^{-1})	Water vapor content (g/kg)	Surface temperature (°K)
0.07	4.58×10^4	0.28	0.50
0.6	2.01×10^4	0.03	0.90

Table 4. Changes in the temperature differences at constant relative humidity for 100 percent stratospheric O_3 depletion (column 4, $\omega_s = 0.1$, Table 2) due to changes in cloud parameters.

Pressure (mbar)	Change in ΔT (°K): No clouds	No high clouds	No low clouds
8.9	−10.4	−0.7	−8.0
74	−2.3	−4.3	−2.5
189	−1.0	−0.1	−1.4
336	−0.3	−0.2	−0.2
500	−0.3	−0.1	−0.2
664	−0.3	−0.1	−0.2
811	−0.3	−0.1	−0.1
926	−0.3	−0.2	−0.2
991	−0.3	−0.2	−0.2

$-0.2°K$ for fixed humidity and $-0.6°K$ for constant relative humidity. Hence the calculated bounds for average cloudiness are between $-0.6°$ and $0.6°K$ ($\omega_s \leq 0.1$). For the present model a 10 percent stratospheric O_3 depletion yields a ΔT_s of ~ 0.01 at 35°N latitude. A comparison of calculated temperature differences assuming constant absolute humidity and constant relative humidity is given in Table 2.

If in the model the height of the maximum in the O_3 profile is arbitrarily decreased from 8.9 mbar (31.6 km) to 74 mbar (18 km) (by removing 50 mass percent of the O_3 from the former level and placing it in the latter level), ΔT_s (for $\omega_s = 0.1$) is $+0.9°K$. It is interesting that this arbitrary decrease in the height of the O_3 maximum yields a larger ΔT_s than complete stratospheric O_3 removal. This is in agreement with the results of Callis *et al.* (*12*) and Coakley (*13*).

To test the sensitivity of the results to the model cloud parameters, separate calculations of the steady-state temperature profiles were made for the cases of (i) no water clouds, (ii) no low-lying water clouds, and (iii) no high-lying water clouds. These results are shown in Table 4. A set of calculations was also performed for an 18-level atmospheric model; no essential difference was found between these and the calculations for nine levels.

The important new conclusions to be drawn from these calculations are that (i) with low-lying absorbing particles surface heating occurs when stratospheric O_3 is depleted (in contrast to the cooling calculated for stratospheric O_3 depletion with no absorbing particles) and (ii) surface albedo affects the amount of surface cooling or heating. For a sky with or without water clouds and with no suspended particles, the O_3 depletion produces less cooling with high surface albedos than with low surface albedos. With absorbing particles the O_3 depletion produces more heating with high surface albedos than with low surface albedos. Since the O_3 abundance, the surface albedo, and the effective albedo change with latitude, all effects must be considered simultaneously when considering O_3 effects at different latitudes. The surface temperature change when the O_3 is depleted, for average cloudiness at 35°N latitude, April, and a surface albedo of 0.6 or less, is between $-0.6°$ and $0.9°K$, while the effect is negligible for 10 percent O_3 removal. The sign of the surface temperature change is determined by the degree to which airborne particles affect the radiative balance. The final effect of O_3 depletion on the temperature must include all the physical details of the atmosphere, including (i) the photochemical balance, (ii) feedback due to changes in surface albedo and ice cover, (iii) feedback due to changes in cloud cover, and (iv) general circulation.

RUTH A. RECK

Physics Department,
General Motors Research Laboratories,
Warren, Michigan 48090

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Uterotrophic Effect of Delta-9-Tetrahydrocannabinol in Ovariectomized Rats

Abstract. *Chronic administration of delta-9-tetrahydrocannabinol, an active component of marihuana, has significant uterotrophic effects in ovariectomized rats as measured by the uterine weight gain bioassay for estrogens.*

The possibility that tetrahydrocannabinol (THC), one of the major active components of marihuana, may have estrogen-like activity has been raised by recent reports on heavy, chronic marihuana use in men (*1*) and THC administration in male rodents (*2*). However, other studies of similar concern in men (*3*) and male rodents (*4*) fail to support these findings.

Since the effects of marihuana use in women and THC administration in female rats are not well documented, the possibility that THC may have estrogen-like activity led us to examine the effects of THC in female rats, using the uterine weight gain and vaginal bioassays for estrogens. The data presented here demonstrate that delta-9-tetrahydrocannabinol (Δ^9-THC) has significant estrogenic activity, as measured by uterine weight gain in spayed rats.

Estrogens often elicit similar morphological and physiological responses in reproductive tissues of male and female rodents and human beings (*5–8*). Heavy, chronic marihuana use, estrogen, or THC administration has the following effects in men and male rodents. In men, heavy marihuana use is reported to elicit gynecomastia, oligospermia, and depression in plasma testosterone levels (*1*). Estrogen administration in men results in depressed testosterone levels, gynecomastia, antiandrogenic effects on secondary sex glands, and azoospermia (*5*). In male mice, chronic administration of THC induces complete arrest of spermatogenesis and regression of Leydig cell tissue and accessory sex glands (*2*). Chronic THC is found to increase adrenal weight, stimulate male breast development, and depress seminal vesicle weights and somatic growth in male rats (*2*). Depressed somatic growth, increased adrenal weight, and depressed seminal vesicle weights are also seen in estrogen-treated male rats (*6*). These findings lend further support to the possibility that heavy, chronic THC administration may have a direct or indirect estrogen-like activity.

It is known that estrogenic compounds are implicated in endometrial carcinoma and breast cancer in women (*7, 9, 10*). In addition, administration of estrogen-like compounds during pregnancy induces vaginal cancer in human female offspring (*11*). In female rodents, estrogens also induce uterine, vaginal, and mammary cancer (*12*). Since oral administration of THC is now used as an antiemetic in patients receiving cancer chemotherapy (*13*), the similarities in human and rodent response to estrogen administration, as well as the possibility that marihuana (THC) may have estrogenic effects, were of interest.

Replacement therapy with estrogenic compounds is known to restore the reproductive tract of ovariectomized rats to the precastrate state (*8, 14*). Using the uterine weight response assay in a preliminary experiment, we found that the