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Thermal Response To Ozone Fluctuations
In The Stratosphere

by

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Thesis Director's Signature:

A handwritten signature in dark ink, appearing to read "J. H. ...", written over a horizontal line.

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ABSTRACT

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We have investigated the extent to which temperature fluctuations in the stratosphere, at low-latitude, are controlled by O_3 absorption of solar and ground radiation. From a one dimensional model, we calculated the expected local heating due to a local O_3 increase. The measured thermal response at different heights then allows us to estimate the local fluctuations in the O_3 abundance, which in turn yields a correlation between the total measured O_3 fluctuations and that expected from the local temperature fluctuations. Thus we compared the theoretical expectation of ΔT versus $\Delta[O_3]$ with a regression analysis of the total O_3 and local radiosonde temperature at 100, 50, 30, 20, and 10 mb for four low-latitude stations. The expected total O_3 fluctuations are comparable to the observed variations of total O_3 , suggesting therefore, that O_3 has important controlling influence on the local temperature fluctuations in the stratosphere.

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I. INTRODUCTION

Radiative processes play a complex role within the stratosphere in determining the thermal structure. Absorption of solar ultraviolet (UV) radiation by O_3 and absorption of solar radiation by CO_2 , and H_2O constitute the heat sources, with a relatively nonnegligible contribution by O_3 absorption in the $9.6\ \mu m$ in the lower stratosphere. The heating rate is dominated by O_3 absorption of solar UV in the middle and lower stratosphere, with a maximum heating rate of $10^\circ K/day$ at 50 km (Newell and Gray, 1972). That part due to the infrared bands of CO_2 , and H_2O , is only about $0.25^\circ K/day$ in this altitude range (Kuhn and London, 1969) and is therefore negligible. In the lower stratosphere, below 30 km, radiative heating due to O_3 absorption of solar UV is still an order of magnitude greater than the radiative heating due to solar absorption by CO_2 , and H_2O . Below approximately 25 km, the heating rate due to the $9.6\ \mu m$ band of O_3 is stronger than that due to the long wavelength bands of CO_2 , and H_2O and approaches within a factor of two the heating due to the absorption of solar UV by O_3 , just above the tropopause (Manabe and Moller, 1961). The heating rate is therefore proportional to the O_3 concentration, $[O_3]$.

Radiative cooling is due to infrared (IR) emission

primarily by the $15\ \mu\text{m}$ band of CO_2 , the complex rotational bands of H_2O , and the $9.6\ \mu\text{m}$ band of O_3 (above 30 km). Dickinson (1973) has reduced the enormous computer time required for determining the amount of IR cooling. He parameterized the cooling rates by a linear cooling approximation which depends upon the temperature deviations from equilibrium. The radiative cooling is then proportional to some function of space, times the temperature deviation. So, with the cooling rate determined by $T(z)$, and the heating rate determined by $[\text{O}_3(z)]$, the expected change in radiative temperature, $\Delta T(z)$, can be expressed in terms of the perturbation in local O_3 concentration, $\Delta[\text{O}_3(z)]$.

The existence of the interdependence of O_3 and T is unquestioned, but the mechanism responsible for this interdependence is. The analysis of recent Nimbus 4 observations has led Prabhakara et al. (1976) and Barnett et al. (1975) to different cause-effect relations. Prabhakara et al. (1976) interpreted the large standard deviation of zonal mean temperature near the polar regions at 100, 30, and 10 mb as being indicative of planetary wave activity. A similar pattern in the deviation of total O_3 , Ω , was also observed at this time. The resulting strong correlation between the temperature and total O_3 was then concluded as being due to associated transport. Therefore the fluctua-

tions of T and O_3 , are not due to the fluctuations of each other. Barnett et al. (1975) have also obtained a strong correlation between T and $[O_3]$ during a similar time period but at a much higher altitude of around 1 mb. From an O_x - NO_x - HO_x photochemical model the $[O_3]$ equilibrium value was calculated as an explicit function of T by expressing the various reaction rates in terms of their dependence on T . The theoretical variation of the dependence of $[O_3]$ to T was then calculated and shown to be in good agreement with the dependency from a regression analysis using the Nimbus 4 observations. Temperature was then concluded to be controlling $[O_3]$. This concluded interdependency of $[O_3]$ and T is different from that of Prabhakara et al. (1976), and both are different from the findings of this paper. The former argument concludes that an external agent is responsible for the observed fluctuations in T and O_3 where the latter argument concludes that T controls O_3 . Our argument contends a third possibility; O_3 controls T .

The aim of this paper is to ascertain the dominance of radiative processes in governing the long term fluctuations in $T(z)$. This was accomplished by studying the thermal response of the atmosphere due to perturbations of Ω . A regression analysis of mean monthly Ω to the radiosonde temperature observed at 100, 50, 30, 20, and 10 mb yielded

the empirical ratio, β , of ΔT to Ω fluctuations, $\Delta\Omega$. From radiative arguments alone, the amount of $\Delta[O_3]$ required to account for the observed ΔT can be expressed in terms of this regression coefficient at each height. The calculated $\Delta[O_3]$ were then compared to the variabilities of O_3 calculated from numerous ozoneosonde observations. A $\beta(z)$ distribution was then constructed based on the five regression coefficients. This allows us to qualify the relation of $\Delta[O_3]$ to control ΔT at levels unobserved. The constructed model was then qualified by comparing the calculated $\Delta[O_3]$ to the observed $\Delta\Omega$. An error analysis was also incorporated to account for the variations in these parameters affecting the response of ΔT to $\Delta[O_3]$, but which were not measured by us. We find that radiative processes play a significant role above the tropopause in governing the long term ΔT incurred by the observed $\Delta[O_3]$.

II. THEORY

Radiative processes play an important role in governing the mean thermal structure of the upper atmosphere. The equilibrium temperature is established by a balance between the radiative heating and cooling processes. The total heating rate depends on $[O_3]$, and a perturbation of $[O_3]$ changes the temperature which in turn alters the radiative cooling rate. Rodgers and Walshaw (1966) found that most of the IR cooling depends only on the temperature and composition of the level being calculated. This suggests that the perturbed IR cooling could be approximated by using a linear cooling expression. Dickinson (1973) utilized this result and generated a simple two term parameterization for IR cooling. One term, $b(T_0)$, represents the exact cooling for the temperature profile of the 1962 Standard Atmosphere Model. A second term, $a(T-T_0)$, gives the change in the radiative cooling for temperature that depart from this standard model. The equation for the radiative heat balance is,

$$\eta(z)[O_3(z)] = b(T_0(z)) + a(T(z) - T_0(z)) \quad (1)$$

where η is the heating rate per O_3 molecule per day, due to O_3 absorption of solar UV. The height dependence of $a(z)$ and $b(z)$ are tabulated in Dickinson (1973) and are reproduced in Table 1.

A sudden change in $[O_3]$, will not produce an instantaneous change in T . Rather, it will produce a sluggish increase in T at the rate,

$$\frac{\Delta T(z)}{\Delta t} = \eta(z) \Delta [O_3(z)] - a(z) \Delta T(z). \quad (2)$$

Therefore the local relation of ΔT and $\Delta [O_3]$ can be written,

$$\Delta T(z) = \frac{\Delta [O_3(z)] \eta(z)}{(a(z) + \tau^{-1}(z))} \quad (3)$$

where τ is the thermal response time. From the cooling coefficients of Table 1, it is seen that with a change in the heat input the atmosphere requires approximately two weeks at 30 km to adapt to the modified cooling rate. If the $[O_3]$ in a given air parcel fluctuates in a time of $\tau < a^{-1}$ then τ rather than a^{-1} is the dominant lifetime. A typical daily temperature fluctuation plot is shown in Figure 1 for the observed T at Nashville, Tennessee during the summer months. It is noted that this case exists for all the altitudes (16.64 to 31.66 km) observed. A portion of this paper will therefore be devoted to calculating the distribution of $\tau(z)$.

The discussion so far has been concerned only with local O_3 fluctuations. With $\Delta \Omega$ equal to the summation of $\Delta [O_3]$, the relation between ΔT and total O_3 fluctuations can be written,

$$\begin{aligned}
 T(z) &= \beta(z) \int_{-\infty}^0 \Delta[O_3(z)] dz \\
 &= \beta(z) \Delta\Omega
 \end{aligned}
 \tag{4}$$

where $\beta(z)$ is some proportionality constant. Substituting in (3) for $\Delta T(z)$ we get,

$$\Delta[O_3(z)] = \frac{\beta(z) \Delta\Omega [a(z) + \tau^{-1}(z)]}{\eta(z)} . \tag{5}$$

This tells us that for some distribution of $\beta(z)$, the amount of $\Delta[O_3(z)]$ required to give $\Delta T(z)$ can be related to $\Delta\Omega$.

To test the controlling influence of O_3 on T requires generating from observations $\beta(z)$ and calculating $\Delta[O_3(z)]$. This gives the magnitude of $\Delta[O_3(z)]$ required to account for the $\Delta T(z)$ observed. The test is affirmative if the calculated $\Delta[O_3(z)]$ is within the observed $[O_3]$ variability. We now will discuss the empirical analysis and data used to generate $\beta(z)$ which was used in this study.

III. EMPIRICAL ANALYSIS

Monthly radiosonde temperature and total O_3 were collected from four stations (Table 2) in the Northern Hemisphere with an average latitude of $32.0^\circ N$. The data were recorded for the summer months (May, June, and July) of the sixteen year period, 1958 through 1973. These data totaled 1,152 paired values of T and O_3 .

For each height observed, the temperature fluctuation about the mean monthly temperature was correlated to the total O_3 fluctuation about the monthly mean total O_3 . The raw temperature and raw total O_3 values were expressed as,

$$D(s, h, y, m) = D(s, h, y, m) - \sum_y D(s, h, y, m) / n \quad (6)$$

where the matrix element D represents the raw data, ΔT or ΔO_3 . The matrix element D symbolizes the collected data, n is the number of points in D , the index s represents the station, h is height, y is year and m the months in the summer season. The correlation coefficients were then calculated from,

$$C(h) = \frac{\sum_{s, y, m} [\Delta T(s, h, y, m) \Delta O_3(s, y, m)]}{\sqrt{\sum_{s, y, m} [\Delta T(s, h, y, m)]^2 \sum_{s, y, m} [\Delta O_3(s, y, m)]^2}} \quad (7)$$

Table 3 presents the five correlation coefficients calculated and a measure of their reliability at the first confi-

dence level.

The regression coefficient for temperature on total O_3 was also computed,

$$\beta(z) = C(z) \frac{\sigma_T(z)}{\sigma_{\Omega}(z)} \quad (8)$$

where $\sigma_T(z)$ and $\sigma_{\Omega}(z)$ are the standard error of estimates for T on Ω ; an error of assigning average values using a correlation type analysis. This $\beta(z)$ gives the slope of the best fit line for $\Delta\Omega$ versus ΔT . The values obtained are given also in Table 3. Deviations of the regression coefficients ($\beta_{\min}$ to $\beta_{\max}$) were calculated from the standard deviation of the correlation coefficients ($R_{\min}$ to $R_{\max}$). The resulting range in $\beta(z)$ at each of the five heights analyzed is shown in Table 3.

The $\beta(z)$, defined by (8), is interpreted to be a real number. That is, the standard deviations of temperature and total O_3 were not considered noise. There are seven sources of instrumental error contained in the radiosonde. The stability of the ground equipment, the readability of the recorder chart, and the temperature evaluation, account for a total incremental error of 0.2°C ; a large part of the total deviation, 0.36°C . The five σ_T 's calculated were on the average five times greater than this value and therefore, real.

The total O_3 measurements were all recorded by a Dobson spectrophotometer. It is generally agreed that instrumental errors can lead to standard deviations of 1.0 percent. The σ_Ω calculated is a factor of two and a half times this, and therefore real also.

IV. COMPARISON BETWEEN THEORY AND OBSERVATIONS

The values of the four remaining parameters in (5) have yet to be defined. However, a couple ($\eta(z)$ and $a(z)$) have been previously examined and are summarized below along with the derivations of the remaining two terms, $\Delta\Omega$ and $\tau(z)$. The variation of these parameters and their affect on $\Delta[O_3]$ are also discussed.

A. Parameters of Equation (5)

Numerous articles have appeared demonstrating the latitudinal and seasonal variations of η (Dopplick, 1972; Hering, 1967). Heating rates for the latitude and season of concern in this paper were calculated by Park and London (1974), and McElroy et al. (1974). The latter values have served as the bases for the heating rate calculations in this paper.

Their $\eta(z)$ were calculated at height intervals of 2 km for an O_x - NO_x - HO_x atmosphere at 30°N during 12° solar declination. The $\eta(z)$ were adjusted to compensate for our smaller observed O_3 abundance (0.321 atm-cm). At any season, the deviations of $\eta(z)$ due to different $[O_3(z)]$ are negligible relative to the variations due to the solar hour angle (Boykov, 1976). Being in the same season and latitude, the adjusted $\eta(z)$ varied little with those of McElroy et al. (1974). The values were obtained by calcula-

ting the dependence of $\eta(z)$ on $\Omega(z)$. The $[O_3(z)]$ distribution of McElroy et al. (1974) was then reduced by the ratio of the Ω observed and the Ω used by McElroy et al. (1974). The new $\Omega(z)$ was then calculated at each height and the corresponding value of $\eta(z)$ was obtained. A comparison of the two values is given in Figure 2.

The cooling coefficients for altitudes above 27 km were interpolated from the values of Table 1. For altitudes below this level, an upper and lower limit on $a(z)$ was assumed. The minimum cooling coefficient, a_1 , was determined by assuming the value at each level was equal to the last value in Table 1 (26.90 km). The maximum cooling coefficient, a_2 , was extrapolated from the last interval of Table 1. The values of $a_2(z)$ obtained were no more than a factor of two of that found at 27 km. Figure 1 presents the two cooling coefficient distributions calculated.

Daily temperature fluctuations were calculated at the tions of Table 1 during the period May, June, and July of 1963. Calculations show, generally τ is less than a^{-1} . An autocorrelation of the temperature fluctuations about the mean monthly values gave thermal response times in the range of 4 to 8 days in the lower stratosphere. The comparison of τ with a^{-1} is given in Figure 1.

The final unmentioned parameter in (5) is $\Delta\Omega$. Christie

(1973) analyzed Ω for 70 stations in the Northern Hemisphere over the extended period, 1958 through April, 1971. The monthly residual range, after removing the annual cycle, was found to be approximately $\pm 1.0(-2)$ atm-cm at the first confidence level. We have found a comparable value of $\pm 7.95(-3)$ atm-cm at the first confidence level. The latter value was used in this paper and yielded a relative $\Delta\Omega$ of 2.47 percent.

B. Test of Hypothesis at 100, 50, 30, 20, and 10 mb

Equation (5) was then solved at the five heights analyzed from the various parameters defined above. Twelve relative $\Delta[O_3]$ were generated at each altitude due to the twelve possible combinations of the regression coefficients ($\beta_{\min}$, β , and $\beta_{\max}$), cooling coefficients (a_1 and a_2) and the thermal relaxation times (τ_1 and τ_2). The relative $\Delta[O_3]$ values calculated at the five levels are given in Table 4. For $\Delta[O_3]$ to control ΔT , the values of $\Delta[O_3]$ calculated have to be comparable with the observed O_3 variability.

Boykov (1967) has done an extensive analysis on the $[O_3(z)]$ variation in the Northern Hemisphere. The relative fluctuations were shown to be about 15 percent in the lower stratosphere for the two month period (March and April). The values decreased to around 3 percent at the 40 km level. This variation with height agrees well with that calculated for a five month period (May through September) by Spankuch et al.

(1973) over Berlin, West Germany, although his values are larger by a factor of two. The larger variations of Spankuch et al. (1973) is to be expected due to the larger averaged period. The daily observed temperature over any three month period, demonstrates that the seasonal variation is a significant part. Variations about the mean monthly values are the defined fluctuations in this paper. With this in mind, the values of Table 4 are in good agreement above 100 mb. The magnitude of $\Delta[O_3]$ is large enough to explain the observed ΔT . Therefore, the $\Delta[O_3]$ is significant in magnitude to control ΔT .

The disagreement below 100 mb is understandable on the bases of the heating processes responsible for this region. This region is the troposphere. The U.S. Standard Atmosphere Supplement of 1962 gives the lower stratosphere-tropopause altitude at approximately 16.5 km. This is just below the 100 mb level. But frequency distributions (Downie 1972) of tropopause height show half widths on the order of 2 km with Reiter (1963) observing the boundary as high as 80 mb. Therefore the 100 mb level is not totally free of tropospheric influence. This is clearly indicated by the sharp drop of β from 100 to 50 mb. Tropospheric $T(z)$ is governed mainly by convection, with little influence on T due to radiative processes. The calculations for the 100 mb

region were left in for this explicit purpose; to demonstrate that β is a good indicator of ozone's controlling influence on temperature, for even regions outside radiative control. A similar analysis was carried out at the 150 mb level to demonstrate this disagreement and is summarized below. Calculations indicate that for $\Delta[O_3]$ to govern the observed ΔT at 150 mb, that a magnitude of $\Delta[O_3]$, a factor of 3 the $[O_3(z)]$ observed, is required. This is a factor of 20 times the $\Delta[O_3]$ calculated by ozone sondes and UMKHR observations (Boykov, 1967) and therefore impossible. The relatively strong correlation of ΔT to $\Delta\Omega$ at these lower altitudes (Table 3) can be explained by other reasons (Normand, 1952).

Dobson, Harrison, and Lawrence (1927) were first to examine the upper air quantities in search of the properties which possessed the strongest relation to O_3 . They discovered maximum positive deviations of daily temperature from the monthly mean are generally found to the rear of surface low-pressure areas, while maximum negative deviations are found to the rear of surface highs. This indicates that higher (lower) total O_3 values are due to advection from the richer (poorer) O_3 abundances of the northern (southern) latitudes of the Northern Hemisphere, in the former (latter) case.

Meetham (1937) expanded upon the work of Dobson, Harrison, and Lawrence (1927) to yield a more significant statistical study. He had shown that total O_3 and temperature were highly correlated with values as large as 0.59 in the upper troposphere and lower stratosphere. Measurements were taken at 12, 15, and 18 km at Oxford, England and Arosa, Switzerland. The explanation offered by Meetham (1937) for this relation was the first baroclinic stability argument linking O_3 and T. Suppose a cyclone forms in the troposphere, then the tropopause will fall since cyclones are low-pressure centers. The air from the stratosphere then subsides, and since $[O_3]$ increases with altitude, a larger O_3 abundance occurs. This, coupled with the increase in potential temperature, yields a strong correlation; as observed. This explanation is examined in detail by Reed (1950).

The intrusion of polar fronts has also a dynamic affect on the ozone-temperature relation. The moving polar front, richer in $[O_3]$, wedges under the warmer air ahead of the front. A temperature inversion forms, prohibiting convection. The increasing total O_3 is positively or negatively correlated to T depending on the height the data are retrieved from. Attmannspacher and Hartmannsgruber (1973), and Sreedharan and Tiwari (1973) have demonstrated this principle by analyzing several cases observed

in Poona, India and Hohenpeissenberg Observatory (65 km southwest of Munich, Germany), respectively.

The accompaniment of thunderstorms with fronts is another potential for an ozone-temperature correlation. Total O_3 amounts are observed to increase as much as 600 percent during storms accompanied by lightning. The actual O_3 production mechanism responsible is questionable, but several explanations have been offered (Orville, 1968, and Shlanta and Moore, 1972).

These ozone-temperature links summarized above, are examples of where strong correlations do not necessitate a cause-effect relation. The variations of O_3 and T are in phase, but one is not the cause of the other. The parameter $\beta(z)$ attests to this fact.

C. Test of Hypothesis for the Entire Stratosphere

We now wish to extend our analysis to regions that were not covered empirically. The observed temperature fluctuations for these regions are assumed to be reflected by the observed β . The hypothesis that $\Delta[O_3(z)]$ controls $\Delta T(z)$ for the entire stratosphere, based on extrapolating β from heights analyzed, was tested under the restriction that the sum of the local O_3 fluctuations must be equal to the observed total O_3 fluctuation. This last restriction was required, due to the lack of observa-

tions to calculate a reliable $\Delta[O_3(z)]$ in the upper atmosphere.

The $\beta(z)$ distribution is plotted in Figure 2. The values of β above 10 mb turns out not to be critical in determining $\Delta\Omega$. The concentrations of O_3 are very low at these altitudes. Over two thirds of Ω is found below 30 km with approximately two hundredths of Ω above 40 km. With an average relative $\Delta[O_3]$ of 15 percent for the heights analyzed, then only 0.3 percent can be due to this region.

The $\beta(z)$ distribution of the troposphere has been ignored based on the discussion in Section B. The argument is based on the random occurrence of O_3 produced by thunderstorms and other nonperiodic sources and sinks. Assuming long term averaging destroys these variations, the $\Delta\Omega$ observed is due entirely to the stratosphere; there is no real long term fluctuation of O_3 in the stratosphere.

Equation (5) was then integrated from the tropopause with the possible values of the previous distributions of $a(z)$, $\tau(z)$, and $\eta(z)$. Table 5 summarizes the agreement obtained between the calculated $\Delta[O_3(z)]$ distribution and the observed $\Delta\Omega$. Table 5 shows that within the various values of τ and a , stratospheric and mesospheric $\Delta T(z)$

is controlled by $\Delta[\text{O}_3(z)]$. This conclusion depends critically upon the thermal response time and the accuracy of the height of the tropopause. This latter point is clearly seen by examining Table 4.

V. CONCLUSION

Ozone fluctuations have been shown to be strong enough to control the observed temperature fluctuations for regions in the stratosphere (and possibly up to at least 80 km). The numbers in Table 4 and 5 were obtained due to the range in values of τ and a . Other parameters not yet mentioned, may also have an affect on the values in Table 4 and 5 and will now be briefly mentioned.

The $[O_3(z)]$ distribution used in this calculation is not a unique profile. Observations of regions near 30°N during the summer (Hilsenroth et al., Johnson et al., 1952; Krueger, 1973; and Hering and Borden, 1967) show that the difference in $[O_3(z)]$ distributions is small though. The choice of $[O_3(z)]$ distributions used, will have an affect only on the values of Table 4 and not Table 5. These effects are much smaller though, than the uncertainties previously considered.

The choice of $[O_3(z)]$ distributions does have an affect on $\eta(z)$, but again its affect on the values in Table 4 and Table 5 is negligible. Hering et al. (1967) have found that η is primarily dependent on the solar zenith angle and the duration of sunlight. These variables account for over 94 percent of the variance in the calculated mean monthly heating rates at the 10, 15, and 25 mb levels.

Two other independent variables of η are the absorption cross-sections for O_3 and the solar UV irradiance. The disagreement on the values of the absorption cross sections is negligible (Ackerman, 1971; Inn and Tanaka, 1953; Hearn, 1961; and Vigroux, 1953). The disagreement in the solar irradiance variation is not totally negligible and may be in fact the cause of the long term O_3 fluctuations. It has been shown experimentally that there exists a relation between increased sunspot activity and enhanced solar radiation in the UV (Heath, 1971). This is contrary to the Smith and Gottlieb (1975) review of the variation of solar UV flux to various stages of solar activity.

But whatever the mechanism responsible for O_3 fluctuations, Angell and Korshover (1973) have shown them to be real. Our indications are that they are strong enough to control the temperature fluctuations above the tropopause.

LEGEND

Table 1. Total infrared cooling rates and cooling coefficients at various pressure levels, derived from 1962 Standard Atmosphere Model.

Table.2. Stations used for observational analysis.

Table 3. Correlation, (R) , and regression, $\beta(z)$, coefficients of ΔT on $\Delta \Omega$ for the five heights analyzed and a measure of their reliabilities at the first confidence level.

Figure 1. Daily temperature fluctuations observed at Nashville, Tennessee during May, June, and July, 1963.

Figure 2. Comparison of the adjusted heating rates used in this paper with McElroy et al. (1975).

Figure 3. Comparison of the calculated thermal response times and the cooling coefficients of Dickinson (1973).

Table 4. Percentage of $\Delta[O_3]/[O_3]$ for the five heights analyzed $(\Delta[O_3(z)] = \beta(z) \int \Delta[O_3(z)](a(z) + \tau^{-1}(z)) \eta^{-1} dz)$.

Figure 4. Empirical β^{TO} vs. height and their $\Delta\beta$ at the first confidence level.

Table 5. Agreement between constructed local $\Delta[O_3(z)]$ and the observed $\Delta\Omega$.

HEIGHT		COOLING RATE	COOLING COEF.
Z (Mb)	Z (KM)	b (°K/DAY)	a (DAY ⁻¹)
0.01	80.70	-1.7	.016
0.04	72.69	0.7	.062
0.11	66.01	4.2	.125
0.17	62.91	7.6	.172
0.37	57.02	9.5	.200
0.67	52.32	12.1	.220
1.00	49.07	11.4	.212
2.23	42.69	6.7	.135
8.17	33.07	2.8	.08
20.09	26.90	2.0	.06

TABLE 1

TOTAL OZONE			RADIOSONDE TEMPERATURE		
STATION	LATITUDE	LONGITUDE	STATION	LATITUDE	LONGITUDE
MAUNA LOA, HAWAII	19° 32' N	155° 35' W	HILO, HAWAII	19° 42' N	155° 04' W
TALLAHASSEE, FLORIDA	30° 26' N	64° 20' W	MONTGOMERY, ALABAMA	32° 23' N	86° 19' W
NASHVILLE, TENNESSEE	36° 15' N	86° 34' W	SAME		
BOULDER, COLORADO	40° 01' N	105° 15' W	DENVER, COLORADO	39° 44' N	104° 59' W

TABLE 2

HEIGHT (Mb) (km)	R-ΔR	R	R + ΔR	$\beta - \Delta\beta$ (°K cm ²)	β (°K cm ²)	$\beta + \Delta\beta$ (°K cm ²)
10 31.66	0	.10	.20	2.6051(-20)	5.4596(-19)	1.0904(-18)
20 26.78	-.01	.08	.17	-4.0938(-20)	3.1485(-19)	7.1083(-19)
30 24.13	-.04	.05	.14	-1.5631(-19)	2.0469(-19)	5.5080(-19)
50 20.88	-.01	.10	.19	4.0938(-20)	3.9077(-19)	7.6293(-19)
100 16.64	.29	.37	.45	1.6450(-18)	2.1139(-18)	2.5530(-18)

TABLE 3

DATE TEMPERATURE RECORDED
FIGURE 1

FIGURE 2

FIGURE 3

HEIGHT (M_b) (KM)	THERMAL RESPONSE TIME	$(\Delta[O_3]/[O_3])_{MIN}$		$\Delta[O_3]/[O_3]$		$(\Delta[O_3]/[O_3])_{MAX}$	
		a_1	a_2	a_1	a_2	a_1	a_2
10	τ_1	.08	.08	1.63	1.63	3.37	3.37
31.66	τ_2	.04	.04	.80	.80	1.65	1.65
20	τ_1	.23	.23	1.75	1.75	3.95	3.95
26.78	τ_2	.11	.11	.83	.83	1.86	1.86
30	τ_1	1.24	1.27	1.70	1.67	4.53	4.47
24.13	τ_2	.60	.59	.80	.77	2.13	2.07
50	τ_1	.58	.56	5.53	5.32	10.81	10.39
20.88	τ_2	.46	.25	4.36	2.39	8.52	4.66
100	τ_1	57.83	53.70	74.26	68.96	89.75	83.34
16.64	τ_2	27.20	23.07	34.93	29.62	42.21	35.80

TABLE 4

FIGURE 4

TROPOPAUSE ALTITUDE (KM)	THERMAL RESPONSE TIME	$(\int \frac{\Delta[\text{O}_3]}{\Delta\eta} dz)_{\text{MIN}}$		$\int \frac{\Delta[\text{O}_3]}{\Delta\eta} dz$		$(\int \frac{\Delta[\text{O}_3]}{\Delta\eta} dz)_{\text{MAX}}$	
		a_1	a_2	a_1	a_2	a_1	a_2
18	τ_1	.538	.407	1.466	1.384	2.594	2.462
	τ_2	.266	.191	.881	.818	1.558	1.425
17	τ_1	1.334	1.126	2.183	2.032	3.868	3.612
	τ_2	.741	.629	1.308	1.177	2.318	1.815
16	τ_1	2.131	1.846	2.900	2.679	5.142	4.764
	τ_2	1.217	.987	1.736	1.535	3.079	2.205

TABLE 5

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