THE DISTRIBUTION OF ODD NITROGEN IN THE LOWER STRATOSPHERE AND POSSIBLE PERTURBATIONS CAUSED BY STRATOSPHERIC AIR TRANSPORT

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In the lower stratosphere a significant production of odd nitrogen results from the reaction \( \text{N}_2 \text{O} + \text{O}(^1\text{D}) \rightarrow 2 \text{NO} \). Since the transport is relatively slow, odd nitrogen builds up with a maximum mixing ratio of \( 2 \times 10^{-8} \) at 30 km. Profiles of odd nitrogen, for different latitudes, winter and summer, are computed from one-dimensional transport models. Variations with latitude are small. Horizontal transport is therefore not believed to alter our results significantly.

In order to evaluate the effect of odd nitrogen upon the ozone layer, \( \text{NO}_x \) profiles are calculated. OH is here a key component, since it converts \( \text{NO}_2 \) to \( \text{HNO}_3 \). In the region where ozone is determined by chemistry rather than by transport (above 25 km), \( \text{NO}_2 \) is found to be relatively abundant.

The effect of stratospheric transport on the \( \text{NO}_x \) distribution is shown to depend critically upon the height of emission. The effect increases by a factor of 5 or more for a change of flight level from 18 km to 23 km. This strong dependence should be duly considered when future stratospheric transport is discussed.
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THE DISTRIBUTION OF ODD NITROGEN IN THE LOWER STRATOSPHERE AND POSSIBLE PERTURBATIONS CAUSED BY STRATOSPHERIC AIR TRANSPORT

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Abstract

In the lower stratosphere a significant production of odd nitrogen results from the reaction \( \text{N}_2\text{O} + \text{O} (^{1}\text{D}) \rightarrow 2\text{ NO} \). Since the transport is relatively slow, odd nitrogen builds up with a maximum mixing ratio of \( 2 \times 10^{-8} \) at 30km. Profiles of odd nitrogen, for different latitudes, winter and summer, are computed from one-dimensional transport models. Variations with latitude are small. Horizontal transport is therefore not believed to alter our results significantly.

In order to evaluate the effect of odd nitrogen upon the ozone layer, \( \text{NO}_x \) profiles are calculated. \( \text{OH} \) is here a key component, since it converts \( \text{NO}_2 \) to \( \text{HNO}_3 \). In the region where ozone is determined by chemistry rather than by transport (above 25km), \( \text{NO}_2 \) is found to be relatively abundant.

The effect of stratospheric transport on the \( \text{NO}_x \) distribution is shown to depend critically upon the height of emission. The effect increases by a factor of 5 or more for a change of flight level from 18km to 23km. This strong dependence should be duly considered when future stratospheric transport is discussed.

1. Introduction

It is well established (Crutzen, 1971, Johnston, 1971, Hesstvedt, 1973), that nitrogen oxides act upon the ozone layer through the reaction

\[
\text{NO}_2 + \text{O} (^{3}\text{P}) \rightarrow \text{NO} + \text{O}_2 \quad k_1 = 9.2 \times 10^{-12}
\]

whereby odd oxygen is destroyed. Accordingly, reliable information about the distribution of nitrogen oxides in the lower stratosphere is essential in quantitative assessments of this effect. Production of odd nitrogen takes place in two regions (Crutzen, 1971, Isaksen, 1973). In the upper region, around 100km,
ionic processes result in an important production of NO and N. In the lower region, between 20 and 30 km, NO is produced when N₂O is destroyed by O (¹D). In this region the chemistry appears to be rather complicated. Modeling of the atmosphere with respect to odd nitrogen species, must also comprise a series of oxygen, hydrogen, and carbon species. The model has earlier been presented by Isaksen (1973) and will therefore not be discussed in detail here. We will, however, discuss production and loss reactions with O (¹D) and OH, which are day components for the nitrogen oxides in the lower stratosphere.

2. The model

Important parameters in the NOₓ calculations, such as eddy diffusion coefficient, temperature and the air density, are known to vary markedly with season and latitude. Calculations are therefore made for 60° and 30°, winter, for the Equator, and for 30°, 60°, and 90°, summer. Temperatures and densities are taken from CIRA (1965). The models are one-dimensional, and vertical eddy transport is included. The eddy diffusion coefficients used are given in Figure 1. Up to 27 km the coefficients are taken from Gudiksen et al., (1968). Above 27 km an exponential increase is assumed, as suggested by Lindzen (1971). His indication of a seasonal variation is also adapted.

Hesstedt (1973) has shown that a two-dimensional model should be considered for computations of the ozone distribution in the lower stratosphere. Since our model is one-dimensional we therefore decided to use observed values of ozone (Hering and Borden, 1964).

Our model is a steady state model, where the photochemistry is averaged over the day. Long lived species which are influenced by eddy diffusion are determined from equations of the type

\[ \frac{d[X]}{dt} = - \frac{d}{dz} \left( K_z [M] \frac{d\delta_x}{dz} \right) + P_X - Q_X [X] = 0 \]  

(2)

where \( K_z \) is the vertical eddy diffusion coefficient, \( M \) the total number density, \( \delta_x = [X] / [M] \) the mixing ratio of component \( X \), and \( P_X \) and \( Q_X [X] \) are averaged photochemical production and loss terms. This equation is solved by the direct implicit method given by Isaksen (1973), using finite differences.

Components with short chemical lifetimes, hours or less, are not influenced by transport processes. The transport term in equation (2) may then be disregarded, and average daytime values are determined from expressions of the type

\[ P_X - Q_X [X] = 0 \]  

(3)

where \( P_X \) and \( Q_X [X] \) are average daytime production and loss terms.
3. **Atomic oxygen in the excited state and hydroxyl**

Two components play an important role in the formation of odd nitrogen species; these components are O (\(^1\)D) and OH. Odd nitrogen is produced through the reaction (1) which appears to be the only effective source in the stratosphere. Nitrogen oxides will therefore depend strongly on the O (\(^1\)D) profiles, which are determined from

\[
O_3 + h\nu \rightarrow O_2 + O (^1D) \quad \lambda < 311\text{nm} \tag{4}
\]

\[
O (^1D) + M \rightarrow O (^3P) + M \quad k_5 = 6 \times 10^{-11} \tag{5}
\]

which gives

\[
[O (^1D)] = J_3 \cdot [O_3]/k_5 \cdot [M] \tag{6}
\]

In our model observed values are used for O\(_3\) and M. \(J_3\) is the daytime average dissociation rate of O\(_3\) for \(\lambda < 311\) nm. Its strong variation with height in the lower stratosphere (Isaksen, 1973) is a result of ozone absorption, and can therefore be calculated from the observed ozone profile.

O (\(^1\)D) profiles, calculated from equation (6) are given in Figure 2 for different latitudes, summer and winter. The rapid decrease below 50km is a result of the decrease in \(J_3\). Variations with latitude are small except at high latitudes, winter, where the production is very slow because of the sun.

Hydroxyl is a second component in the lower stratosphere. It is produced by the reaction

\[
NO + HO_2 \rightarrow NO_2 + OH \quad k_7 = 5 \times 10^{-11} \cdot e^{-\frac{1600}{T}} \tag{7}
\]

In addition to this reaction there is some production through the reaction

\[
O (^3P) + HO_2 \rightarrow OH + O_2 \quad k_8 = 7 \times 10^{-11} \tag{8}
\]

Below 30km, however, this reaction becomes negligible, since O (\(^3\)P) decreases rapidly below this height. OH is lost through several reactions. Important reactions in the lower stratosphere are found to be

\[
OH + CH_4 \rightarrow H_2O + CH_3 \quad k_9 = 5.5 \times 10^{-13} \cdot e^{-\frac{1886}{T}} \tag{9}
\]

3
\[
\text{OH} + \text{CO} \rightarrow \text{H} + \text{CO}_2 \quad k_{10} = 2.1 \times 10^{-13} \cdot e^{-\frac{115}{T}} 
\]

\[
\text{OH} + \text{HNO}_3 \rightarrow \text{H}_2\text{O} + \text{NO}_3 \quad k_{11} = 1.7 \times 10^{-11} \cdot e^{-\frac{1650}{T}} 
\]

\[
\text{OH} + \text{NO}_2 + \text{M} \rightarrow \text{HNO}_3 + \text{M} \quad k_{12} = 1.05 \times 10^{-11} \cdot e^{-\frac{170}{T}} (1 + f)^{-1} \quad (12)
\]

\[
f = 1.6 \times 10^{-2} \cdot T^4 / (4 \times 10^{-11} [\text{M}])
\]

Reactions (9) and (10) are real loss reactions, since \( \text{CH}_3 \) and \( \text{H} \) are ultimately converted to \( \text{HO}_2 \). Reaction (12), on the other hand, is not a real loss since most of \( \text{OH} \) lost by this reaction is reformed through the reaction

\[
\text{HNO}_3 + h\nu \rightarrow \text{OH} + \text{NO}_2 \quad \lambda < 546\text{nm}, \quad J_{a\text{HNO}_3} 
\]

The real \( \text{OH} \) loss is only the fractional part going through reaction (11) and reaction

\[
\text{HNO}_3 + h\nu \rightarrow \text{HO}_2 + \text{NO} \quad \lambda < 370\text{nm}, \quad J_{b\text{HNO}_3} 
\]

on the short wavelength side of \( \text{HNO}_3 \) dissociation. It appears that reaction (14) is much less effective than reaction (13). \( \text{OH} \) may then be determined from

\[
[\text{OH}] = \frac{(k_7 [\text{NO}] + k_8 [\text{O} (3\text{P})]) [\text{HO}_2]}{k_9 [\text{CH}_4] + k_{10} [\text{CO}] + 2k_{11} [\text{HNO}_3] + J_{b\text{HNO}_3}} \quad (15)
\]

\( \text{OH} \) can therefore only be determined in a model where all the components occurring in equation (15) are computed simultaneously. This is done in our model. Calculations are repeated until convergence is obtained. The resulting \( \text{OH} \) profiles are shown in Figure 3. \( \text{OH} \) has, for all latitudes, a maximum near 45km with a peak value of \( 2 - 3 \times 10^{17}\text{cm}^{-3} \). Down to the peak level \( \text{HO}_2 \) is effectively converted to \( \text{OH} \) through reaction (8), while the destruction through reactions (9), (10), and (12) is slow. Below 45km this is no longer true and we get a rapid decrease in \( \text{OH} \) down to the tropopause, where we find a minimum value near \( 10^5 \text{cm}^{-3} \). The strong variations with latitude below about 17km come from the variations in the tropopause level. At the Equator heights below 17km are in the troposphere, at high latitudes all heights above 8km are in the stratosphere.
4. Distribution of nitrous oxide

Destruction of \(\text{N}_2\text{O}\) through the reaction

\[
\text{N}_2\text{O} + \text{O} \left(^1\text{D}\right) \rightarrow \text{NO} + \text{NO} \quad k_{16} = 9 \times 10^{-11}
\] (16)

is the only important source of \(\text{NO}_x\) in the lower stratosphere. Realistic information on the spatial distribution of this component is therefore essential. \(\text{N}_2\text{O}\) has been found to be fairly uniformly distributed in the troposphere with negligible seasonal variations. A mixing ratio of \(2.5 \times 10^{-7}\) is regarded as representative for the troposphere (Schütz et al., 1970). Production of \(\text{N}_2\text{O}\) is expected to be through soil bacteria. So far no photochemical sources are known. It is therefore difficult to estimate production rates. Since the loss rate is small, vertical transport maintain a constant mixing ratio of \(\text{N}_2\text{O}\) in the troposphere, and the above value is used as boundary value at the tropopause. In the lower stratosphere the conditions are quite different. Diffusion is much lower and loss becomes more effective through the reaction

\[
\text{N}_2\text{O} + h\nu \rightarrow \text{N}_2 + \text{O} \quad \lambda < 337.5\text{nm}
\] (17)

and through the slower reaction

\[
\text{N}_2\text{O} + \text{O} \left(^1\text{D}\right) \rightarrow \text{N}_2 + \text{O}_2 \quad k_{18} = 9 \times 10^{-11}
\] (18)

in addition to reaction (16). \(\text{N}_2\text{O}\) profiles are therefore obtained from

\[
- \frac{d}{dz} \left( K_z [M] \frac{d}{dz} \left( \delta_{\text{N}_2\text{O}} \right) \right) - \left( J_{\text{N}_2\text{O}} + (k_{16} + k_{18}) [\text{O} \left(^1\text{D}\right)] \right) [\text{N}_2\text{O}] = 0
\] (19)

The results are shown in Figure 4. The slope of the profiles depends on the dissociation rates and on the eddy diffusion coefficient. Weak diffusion leads to a rapid decrease in the \(\text{N}_2\text{O}\) mixing ratio \((\delta_{\text{N}_2\text{O}})\). This occurs at 25-30km, where \(K_z\) has a minimum. Variations with latitude and season are relatively small, except for high latitude, winter. Apart from this case, production of \(\text{NO}_x\) through destruction of \(\text{N}_2\text{O}\) is confined to levels below 35-40km. At 60°, winter, the low sun makes photodissociation less important, and when in addition the eddy mixing is stronger, \(\text{N}_2\text{O}\) is transported to greater heights before it is dissociated.
5. Production and distribution of odd nitrogen

From the profiles of \( N_2O \) and \( O(1^D) \), shown in Figures 2 and 4, the production rates of odd nitrogen are easily calculated from

\[
P_{NO_y} = 2k_{16} \cdot [O(1^D)] \cdot [N_2O]
\] (20)

The efficiency of this production depends on how efficient reaction (16) is in breaking down \( N_2O \), compared to reactions (17) and (18). Figure 5 shows the loss rate of \( N_2O \) from reactions (16) and (17). Loss by reaction (18) is equal to the loss by reaction (16). The curves are for 30° summer. It is clearly demonstrated that reaction (17) is the main loss reaction of \( N_2O \). Only a few percent of the loss will produce nitrogen oxides through reaction (16).

Nitrogen oxide production from eq. (20) is given in Figure 6. In the summer production varies relatively little with latitude. Differences below 20km is a result of the variation in tropopause level, which influences the \( O(1^D) \) densities. At all latitudes, odd nitrogen production increases rapidly up to a marked maximum between 22-26km. In the winter the production is important over a much wider region. This is clearly demonstrated for 60°, winter, where a substantial production takes place above 40km, with a maximum around 35km. The winter profiles are a result of the combined effects of the absorption of radiation producing \( O(1^D) \) and the strong eddy diffusion, which effectively transports \( N_2O \) to greater heights. At 30°, winter, these effects are less pronounced, with the result that we get a profile more like the summer profiles. Maximum production rates vary from \( 1.5 \times 10^2 \text{ cm}^{-3}\text{s}^{-1} \) in the summer hemisphere, to \( 1.5 \times 10^1 \text{ cm}^{-3}\text{s}^{-1} \) at 60°, winter. If we add up the total production in a column, the average production for the whole globe becomes \( 10^8 \text{ cm}^{-2}\text{s}^{-1} \).

The distribution of total odd nitrogen in the lower stratosphere is easily calculated when its production rate is known. The only photochemical loss of nitrogen oxides in the atmosphere is through the reaction

\[
NO + N \rightarrow N_2 + O \quad k = 2.2 \times 10^{-11}
\] (21)

However, atomic nitrogen decreases rapidly below 60km (Isaksen, 1973). This reaction is therefore of no interest for our problem. Odd nitrogen is therefore removed from the lower stratosphere by vertical eddy diffusion alone. Since \( K_z \) increases with height, the particle flux of odd nitrogen, \( NO_y \)

\[
F = -K_z [M] \frac{d}{dz} (\delta_{NO_y})
\] (22)

is directed downwards.
The flux is proportional to the eddy diffusion coefficient and to \([M]\). It is, therefore clear that the increase in \(\delta_{NO}\), sufficient to transport away the odd nitrogen produced at a certain height, depends strongly on the heights where the production takes place. Assuming steady state conditions,

\[
- \frac{d}{dz} \left( K_z [M] \frac{d}{dz} (\delta_{NO}) \right) + P_{NOy} = 0 \tag{23}
\]

gives the height distribution of nitrogen in all odd forms, i.e., in the forms of NO, NO\(_2\), and HNO\(_3\); all other odd components are present in negligible quantities. Results of this calculation are given in Figure 7. As might be expected from the considerations above, there are latitudinal variations in odd nitrogen, due to variations in \(K_z\) and in production rates. Above 20km, where the effect on ozone is of importance, variations with latitude are very small in the summer. The mixing ratio of odd nitrogen increases strongly up to 25km, the region of the strongest production (Fig. 6). But even above this height there is substantial increase to above 30km, partly due to the low \(K_z\) values. Above 35km \(\delta_{NO}\) drops as a result of the low mesosphere values used for NO (Isaksen, 1973a). Maximum mixing ratios of \(1.2 \times 10^{-8}\) are found around 32km.

In the winter the low solar elevation leads to production at higher altitudes. Therefore the increase in mixing ratio takes place at higher altitudes. This is especially true for 60°, winter, where the strong atmospheric absorption gives high mixing ratios in the upper stratosphere. It is, therefore, an increase in odd nitrogen mixing ratios in the whole stratosphere.

6. The distribution of odd nitrogen on NO, NO\(_2\), and HNO\(_3\)

In order to estimate the effect on ozone, we shall next see how the total odd oxygen, as computed above, is distributed on the different odd nitrogen species. The initial product is NO through reaction (16). This component will rapidly be converted to NO\(_2\) through reaction

\[
NO + O_3 \rightarrow NO_2 + O_2 \quad k_{24} = 9.3 \times 10^{-13} \exp \left( -\frac{1235}{T} \right) \tag{24}
\]

The balance between NO and NO\(_2\) is maintained through solar dissociation of NO\(_2\)

\[
NO_2 + h\nu \rightarrow NO + O (^{3}P) \quad 397.5\text{nm} \tag{25}
\]

with some contribution from reaction (1).
The conversion to HNO$_3$ is much slower (reaction (12)). To obtain HNO$_3$ densities, the loss reactions must be compared with reaction (12). The main loss of HNO$_3$ is through reactions (13) and (14) with some contribution from reaction (11). NO$_3$ is rapidly converted to NO and NO$_2$ through the reactions

$$\text{NO}_3 + h\nu \rightarrow \text{NO} + \text{O}_2 \quad \lambda < 11\mu$$

$$\text{NO}_3 + h\nu \rightarrow \text{NO}_2 + \text{O} \quad \lambda < 700\text{nm}$$

(26)

NO$_3$ is unlikely to be of importance and will be disregarded here.

From the above reactions the distribution of odd nitrogen on NO, NO$_2$, and HNO$_3$ can be calculated. Equilibrium between NO and NO$_2$ is given by reactions (24), (25), and (1). Loss rates of NO and NO$_2$ are shown in Figure 8 for 30° summer. It is seen that equilibrium is established after a few minutes. Similarly, equilibrium between NO - NO$_2$ and HNO$_3$ is given by equations (11) - (14). Equilibrium in this cycle is much slower, as shown in the same figure. Lifetimes decrease from 1 day at 30km to 10 days at 15km. The short lifetimes in the NO to NO$_2$ cycle means that NO is converted to NO$_2$ a few minutes after sunset, when there is no photodissociation. Likewise, NO is also reproduced rapidly after sunrise from photodissociation of NO$_2$.

The long lifetimes in the NO - NO$_2$ - HNO$_3$ cycle result in negligible HNO$_3$ variations over the day below 30km. By confining the equations for equilibrium in the NO - NO$_2$ and in the NO - NO$_2$ - HNO$_3$ cycle, average daytime number densities for NO, NO$_2$, and HNO$_3$ are given by

$$[\text{NO}] = \left[ \text{NO}_y \right] \frac{J_{\text{HNO}_3} + k_{11} [\text{OH}]}{J_{\text{HNO}_3} + k_{11} [\text{OH}] + k_{12}' [\text{OH}] [\text{M}]} \cdot \frac{k_1 [O (^3\text{P})] + J_{\text{NO}_2}}{k_1 [O] + J_{\text{NO}_2} + k_{24} [O_3]}$$

(27)

$$[\text{NO}_2] = \left[ \text{NO}_y \right] \frac{J_{\text{HNO}_3} + k_{11} [\text{OH}]}{J_{\text{HNO}_3} + k_{11} [\text{OH}] + k_{12}' [\text{OH}] [\text{M}]} \cdot \frac{k_{24} [O_3]}{k_1 [O (^3\text{P})] + J_{\text{NO}_2} + k_{24} [O_3]}$$

(28)
\[
[HNO_3] = \frac{k_{12} \ [OH] \ [M]}{J_{HNO_3} + k_{11} \ [OH] + k_{12}' \ [OH] \ [M]} \tag{29}
\]

\(k_{12}'\) gives the efficiency of the coefficient \(k_{12}\) in converting \(NO + NO_2\) to \(HNO_3\):

\[
k_{12}' = k_{12} \frac{k_{24} \ [O_3]}{k_{24} \ [O_3] + k_1 \cdot [O (^3P)] + J_{NO_2}} \tag{30}
\]

The relative abundances of the three nitrogen species, given by equations (27), (28), and (29), are presented in Figure 9 for heights between 10 km and 40 km for 30° summer. \(HNO_3\) is seen to be the dominating component below about 27 km. At 25 km 70% of odd nitrogen present is in the form of \(HNO_3\), and it exceeds 40% at all heights below. The maximum \(HNO_3\) concentrations around 25 km is a result of the high ozone concentration. Below about 15 km \(HNO_3\) increases due to the increase in \(OH\). Above 35 km \(HNO_3\) is present in small quantities. \(NO_2\), which is of special interest in the ozone chemistry, is most abundant around 30 km, where 50% of the total odd nitrogen is in the form of \(NO_2\). Relative densities vary little below 25 km, between 20% and 30%. NO is the most abundant component above 35 km, compared with the other components. Below 30 km it is a minor component. The NO - \(NO_2\) cycle is of interest for ozone destruction through reaction (1). As we have already seen, lifetimes are very short. During the day, when ozone is lost through reaction (1), NO and \(NO_2\) are in photochemical equilibrium. When the sum of NO and \(NO_2\) is known, equilibrium values are easily calculated from equations (1), (24), and (25).

\(NO_X\) mixing ratios are given in Figure 10 for summer and winter hemisphere. Above 30 km the mixing ratios differ little from odd nitrogen mixing ratios, due to low \(HNO_3\) densities (Fig. 8). Below 25 km, there is a sharp drop down to 10 km, where mixing ratios of about \(10^{-9}\) are obtained for all latitudes. Variations with latitude are very small, except in the winter hemisphere where \(\delta_{NO_X}\) falls off at higher latitude.

7. Air transport in the stratosphere and its effect on the \(NO_X\) distribution

It is evident that nitrogen oxides produced from aircraft flying in the stratosphere will alter the stratospheric \(NO_X\) distribution. Since vertical eddy diffusion varies strongly with height, it is of interest to estimate the magnitude of this additional source when the aircraft are assumed to fly at given heights. We shall here consider the effect of a "reference fleet" of 200 four-engine aircraft, each performing 1000 flights of 2-1/2 hours per year. The fuel flow is assumed
to be 5000kg per hour per engine. The emission of NO (converted to NO₂) is assumed to be 21g per kg fuel. These 200 aircraft are assumed to fly at the same flight level uniformly routed around the globe. This gives an average production rate of $1.88 \times 10^7$ NOₓ molecules cm⁻²s⁻¹. Since we use finite differences with steps of 1km, we have, in our model, a production rate of $1.88 \times 10^2$ NOₓ molecules cm⁻²s⁻¹. This number will be used as an additional source in equation (23) at given flight levels. The variation in $\delta_{NOX}$ for flight levels between 15km and 32km, is given in Figure 11. The steady state profiles in Figure 10 for 60°, summer, and 30°, winter, are used as a basis, and the increase is given in percent. Since the ozone distribution below 25km is largely dominated by transport processes (Hesstvedt, 1973 a), the increase in NOₓ between 25km and 30km is of primary interest for the ozone problem.

As might be expected, the increase in NOₓ depends strongly on the flight level. NOₓ injection at high altitudes leads to a considerable increase in NOₓ. This can be explained by the very fast drop in eddy diffusion coefficient above the tropopause (Fig. 1). The added NOₓ has to be transported downwards into the troposphere. Production at heights where $K_z$ is small, therefore means that the gradient in $\delta_{NOX}$ must be much greater than the gradient is at lower heights with higher $K_z$ values, in order to maintain the flux of NOₓ given by equation (22).

The increase in $\delta_{NOX}$ is somewhat different in the two diagrams of Figure 11, a result of difference in $K_z$. In both cases, however, the perturbation in NOₓ increases rapidly as the source level, given along the abcissa, is moved upwards. With the source at 18km, the increase in NOₓ is approximately 2% for 60°, summer, and 0.5% for 30°, winter. With the source at 23km, NOₓ increases with 5-10% in the 25-30km region in both models, and with source at 28km, NOₓ increases by around 20% in both models. The calculations give similar results for the other latitudes. These results have been used in a two-dimensional model to show the effect on the ozone layer (Hesstvedt, 1973 a).

These numbers clearly demonstrate that the effect of stratospheric transport on the NOₓ densities depends strongly on the flight levels. Flight levels above 20km lead to a considerable increase in NOₓ. Having in mind that the perturbation (in %) of the ozone layer is roughly one third of the perturbation (in %) of NOₓ, the importance of the flight level should be duly considered when the effect of stratospheric transport upon the ozone layer is discussed.

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References


Figure Captions

Figure 1. Vertical eddy diffusion coefficient.

Figure 2. Vertical profiles of O (\textsuperscript{1}D) for 90° summer, 60° summer, 30° summer, 0°, 30° winter, and 60° winter.

Figure 3. Vertical profiles of OH, for the same latitudes as in Figure 2.

Figure 4. Vertical profiles of N\textsubscript{2}O for 90° summer, 0°, 30° winter, and 60° winter.

Figure 5. Vertical profiles of N\textsubscript{2}O loss rates from equations (16) and (18) at 30° summer.

Figure 6. Vertical profiles of production rates of odd nitrogen from equation (16) for the same latitudes as in Figure 4.

Figure 7. Profiles of odd nitrogen mixing ratios for the same latitudes as in Figure 2.

Figure 8. Vertical profiles of odd nitrogen loss rates in the NO - NO\textsubscript{2} cycle (curves $q_{\text{NO}_2} - \text{NO}$, and $q_{\text{NO}}$) and in the NO - NO\textsubscript{2} - HNO\textsubscript{3} cycle (curves $q_{\text{NO}_2} - \text{HNO}_3$ and $q_{\text{HNO}_3}$) for 30° summer.

Figure 9. Distribution of odd nitrogen on NO, NO\textsubscript{2}, and HNO\textsubscript{3}.

Figure 10. Vertical profiles of NO\textsubscript{x} mixing ratio.

Figure 11. Increase (in %) of NO\textsubscript{x} mixing ratio caused by a fleet of 200 SST aircraft as a function of flight level (horizontal axis).
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