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ON THE RELATIONSHIP BETWEEN THE THERMAL STRUCTURE OF THE STRATOSPHERE AND THE SEASONAL DISTRIBUTION OF OZONE

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Abstract. An attempt is made to put the phenomenon of Antarctic ozone hole in a global perspective by pointing out that there is a much larger and more persistent "hole" over the tropics. The thermal structure of the Antarctic lower stratosphere in winter is more similar to the tropical lower stratosphere than the Arctic lower stratosphere in winter. The dynamical transport that is responsible for creating the two column ozone minima has similar relationship to the thermal structure in the lower stratosphere in the two regions.

Introduction

Recent observational evidence [Stolarski et al., 1986; Schoeberl et al., 1986] shows that both the horizontal location and year-to-year variation of Antarctic column ozone appear to be correlated with the mean temperature in the Antarctic lower stratosphere (it should however be cautioned that the time series is too short to establish statistical significance, but these correlations may nevertheless serve to provide candidates for likely causes). It is unlikely that the decrease in the temperature is caused by the decrease in ozone [Rosenfield and Schoeberl, 1986]. An attempt is made in the present paper to provide some theoretical basis for suggesting that changes in lower-stratospheric temperature can lead to changes in column ozone concentration.

In the ongoing debate concerning chemical vs dynamical causes of the Antarctic ozone hole phenomenon, it should perhaps be pointed out that there is an even bigger and more persistent "hole" over the tropics, and that the accepted explanation of that "hole" is dynamical upwelling caused by the cold temperature in the tropical lower stratosphere. Photochemical equilibrium theory would have predicted an equatorial maximum concentration for ozone. Although photochemical processes are important in maintaining the vertical distribution of ozone mixing ratio, it is obvious that dynamical transports are needed to explain the observed latitudinal and seasonal distribution of its column density as depicted in Figure 1, which is almost opposite from that predicted for photochemical equilibrium.

The observed contrasts between the Northern and Southern Hemispheres in the distribution of ozone column density also have a dynamical/thermal origin. A main difference between the two hemispheres relates to the magnitude of the so-called Eliassen-Palm flux divergence, which measures large-scale eddy irreversible transports. The Southern Hemisphere generally has a lower level of large-scale wave activity due, presumable, to the lesser inhomogeneity of its surface topography. Such a dynamical contrast is expected to manifest itself in the zonal mean temperature distributions, and indeed it is known (cf Figures 6-1 and 6-2 in WMO (1985)) that the zonal mean temperature in the lower stratosphere in the Southern winter is about 20^0 to 30^0K colder than the corresponding quantity in the Northern winter. This thermal contrast leads to hemispheric differences in calculated radiative heating rates and diabatic circulations. Figure 2 is a schematic illustration of the different circulation patterns in the two hemispheres based on the radiative calculations of Rosenfield and Schoeberl (1986). For ozone, whose vertical distribution is determined by photochemistry to have its maximum mixing ratio located in the stratosphere, upward transport in the lower stratosphere generally tends to reduce its column abundance by bringing ozone-poor air from the troposphere into the stratosphere. This mechanism accounts for the equatorial minimum of column ozone, where there is prevailing upward motion due to the temperatures in the equatorial lower stratosphere which are colder than the local radiative equilibrium values. Conversely, when the same circulation cell transports ozone poleward and downward the column abundance of ozone is increased over the Artic. In this regard it is perhaps instructive not to compare the thermal structure of the Antarctic lower stratosphere with that over the Arctic region, but instead to compare it with the thermal structure of the tropical tropopause. In some sense, one can say that, thermally, the "tropopause" over Antarctica in winter and spring occurs in what we generically (in terms of height) refer to as the lower stratosphere, and that the thermal stratosphere does not exist until above 10 mb, where temperature starts to increase with height. Below that the vertical temperature gradient is generally negative, a feature we normally associate with a troposphere. The winter temperature in the 100 to 10 mb region is very cold, a feature also in common with the tropical tropopause region. So, if the thermal structure in the equatorial region is associated with an upward transport that produces an ozone column minimum, is it plausible to think that a similar thermal structure in the Antarctic can be associated also with an upwelling? Tung et al. (1986) suggested that upwelling may develop in early spring when the sun returns to Antarctica and rapidly increases the radiative equilibrium temperature, Te. For a brief period of about a month and a half, before the actual mean temperature, $\overline{T}$, catches up to the increasing Te, one may have $T < Te$. This situation then leads to an upward mean diabatic circulation, which lowers the column concentration of ozone in a similar manner as the situation in the tropics.

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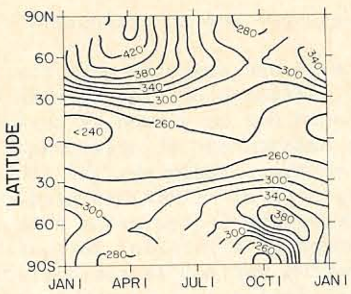

Fig. 1. Time- latitude section of the zonal-mean column ozone density in Dobson units, from Bowman and Krueger (1985).

Relationship between temperature and ozone.

Let Q(T) be the net radiative heating rate. Let Te be the radiative equilibrium temperature defined by Q(Te) = 0. The relationship between temperature and net heating is given approximately by [Dickinson, 1973]:

$$Q \cong -\alpha\Delta T, \tag{1}$$

where $\Delta T \equiv T - Te$ and $\alpha \equiv -(\partial Q/\partial T)_{Te}$. The relationship between the zonal mean "residual" meridional circulation ($\bar{v}^*$, $\bar{w}^*$) and Q (see Edmon et al., (1986) is, from the quasi-geostrophic form of the energy equation:

$$\Gamma\bar{w}^* = \bar{Q} - \frac{\partial}{\partial t}\bar{T}, \tag{2}$$

where overhead bar denotes zonal mean in isobaric surfaces, and $\Gamma \equiv \partial T/\partial z + g/c_p$ is the static stability parameter.

For a trace species such as ozone, whose mass mixing ratio is χ and whose rate of production by sources and sinks is S, the effect of dynamical transport on χ is described approximately by:

$$\frac{\partial}{\partial t}\bar{\chi} + \bar{v}_T\frac{\partial}{\partial y}\bar{\chi} + \bar{w}_T\frac{\partial}{\partial z}\bar{\chi} = \frac{\partial}{\partial y}(Kyy\cos^2\phi\frac{\partial}{\partial y})\bar{\chi} + \bar{S} \tag{3}$$

The transport circulation, ($\bar{v}_T$, $\bar{w}_T$), is in general different than the residual circulation ($\bar{v}^*$, $\bar{w}^*$) [WMO, 1985], but the latter can be taken to be a fair approximation for the former for our purpose. The diffusion term on the right-hand side of Eq. (3) represents irreversible mixing by transient eddies acting predominantly along isentropic surfaces. The source term $\bar{S}$ will be approximated her by $\bar{P} - \bar{L}\bar{\chi} = -\bar{L}\Delta\bar{\chi}$, where $\Delta\bar{\chi} \equiv \bar{\chi} - \chi_0$, and χ_0 is the photochemical equilibrium ozone mixing ratio defined by $\bar{S}(\chi_0) \equiv 0$. In (3) ϕ is latitude andy $y \equiv a\sin\phi$, and $z \equiv \ln(p_{00}/p)$.

Equatorial Lower Stratosphere. In regions where the horizontal gradient, $\partial\bar{\chi}/\partial y$, is weak, such as for ozone in the equatorial lower stratosphere. Eq. (3) becomes

$$\frac{\partial}{\partial t}\bar{\chi} = -\bar{w}^*\frac{\partial}{\partial z}\bar{\chi} + \bar{S} \tag{4}$$

In the tropical lower stratosphere, the observed zonal mean temperature is very cold. With $\bar{T} <$ Te, Eq. (1) gives $\bar{Q} > 0$, and Eq. (2) suggest $\bar{w}^* > 0$, if Te-$\bar{T}$ is large enough. According to Eq. (4), the resulting upwelling tends to reduce the local tracer concentration in regions where its vertical gradient is positive. For an inert tracer, this process will continue until the vertical gradient is eliminated. For a photochemical species such as ozone, the vertical gradient is maintained by photochemical processes, and so the decline in ozone concentration will continue until dynamical loss is balanced by photochemical production. The balance is expressed by

$$\frac{\partial}{\partial t}\Delta\bar{\chi} \cong -\bar{w}^*\frac{\partial}{\partial z}\chi_0 - \bar{L}\Delta\bar{\chi} \tag{5}$$

in regions of large vertical gradients of χ_0. Eq. (5) implies that upwelling tends to reduce local ozone concentration, leading towards a steady state value of

$$\Delta\bar{\chi} = -\frac{\partial}{\partial z}\chi_0 . \bar{w}^*/\bar{L} = \Delta\bar{T}\frac{\alpha\frac{\partial}{\partial z}\chi_0}{\bar{L}\Gamma} \tag{6}$$

Eq. (6) establishes the direct correspondence between temperature departure from radiaitve equilibrium and species concentration departure from photochemical equilibrium.

Antarctic Lower Stratosphere. Here during winter and early spring, large-scale irreversible mixing is usually low and perhaps getting lower [Mahlman and Fels, 1986]. For our present qualitative arguments, we shall assume $Kyy\cos^2\phi \sim 0$. The geostrophic form of the zonal momentum equation [see Edmon et al., 1980] is:

$$f\ \bar{v}^* = Kyy\ \cos^2\phi\ \frac{\partial}{\partial y}\bar{q} \tag{7}$$

where q is the quasi-geostrophic potential vorticity. Hence $\bar{v}^*$ is also weak inside the polar vortex. Eq. (3) now becomes, locally in the Antarctic lower stratosphere:

$$\frac{\partial}{\partial t}\bar{\chi} = -\bar{w}^*\frac{\partial}{\partial z}\bar{\chi} \tag{8}$$

when photochemical production $\bar{S}$ is ignored. Eq. (8) suggests that there is a correlation between the tendency of ozone decline and upwelling. As mentioned before, upwelling is correlated with cold temperature, i.e. $\Delta\bar{T} < 0$, only when the latter has sufficiently large amplitude to overcome the adiabatic cooling term in Eq. (2). Since the pattern of cold temperature in Antarctic lower stratosphere does not change appreciably before the onset of wave events, one then expects an approximate correlation between cold temperature and low ozone concentration for a period of time before the final warming. Such a correlation can be expressed by (6) approximately if one replaces the photochemical loss frequency L in (6) by $D \equiv -(\partial\bar{\chi}/\partial t)/\chi$, the dynamical loss frequency, and reinterpretes χ_0 as the pre-existing ozone distribution.

The observed correlation between the column density of ozone and lower stratospheric

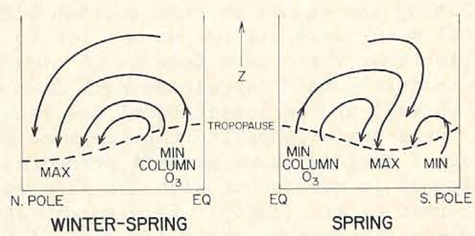

Fig. 2. Schematic diagram of the diabatic circulation patterns for the Northern (left) and Southern (right) Hemispheres.

temperature arises from (6) and the fact that the column density is significantly influenced by factors affecting $\Delta\chi$ near the altitude of its maximum partial pressure, which is the lower stratosphere.

Correlation for wave components. Let $\chi' = \chi' = \bar{\chi} - \chi$ be the perturbation from zonal average. Its distribution is governed by

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right)\chi' + v'\frac{\partial}{\partial y}\bar{\chi} + w'\frac{\partial}{\partial z}\bar{\chi} = 0 \qquad (9)$$

Similarly, temperature perturbation is governed by

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right)T' + \Gamma w' = 0. \qquad (10)$$

Combining (9) and (10) we find

$$\frac{\chi'}{\frac{\partial}{\partial z}\bar{\chi}} = \frac{T'}{\Gamma} - \eta'\frac{\partial}{\partial y}\bar{\chi} \Big/ \frac{\partial}{\partial z}\bar{\chi} \qquad (11)$$

where η' is the horizontal eddy displacement. Neglecting the last term in (11) in regions of small horizontal mean gradient, we find an approximate correlation between perturbation ozone amount and perturbation temperature. Taking typical values of $\Gamma \sim 7^0$K per kilometer, and a scale height for mean ozone, H_0, of 6 to 9 km, one finds for the Antarctic lower stratosphere:

$$\chi'/\bar{\chi} \cong T'/H_0\Gamma \sim 50\% \qquad (12)$$

for a thermal wave of amplitude $T' \sim 20^0$ to 30^0K, which is quite typical of the observed values in this region [Stolarski et al., 1986]. Eq. (12) implies that relatively large zonal asymmetry can be expected in column ozone distribution even in the Antarctic region as a result of wave perturbations.

This remark is relevant when one attempts to compare the ozone data at Halley Bay, say, with 1- or 2-D model results. In view of the fact that column ozone measurements at Halley Bay largely reflect the minimum value over all longitudes, and in view of the fact, consistent both with our result in this section and the observation of Stolarski et al. (1986), that the minimum value is partly a result of large-scale wave dynamics, one should in general expect a factor of two underprediction of the Halley Bay data value by 1- or 2-D models [Tung et al., 1986]. In other words, one should compare the predictions of a zonally averaged model with the zonally averaged data and not with the minimum values which are usually about one hundred Dobson lower than the zonal mean [Stolarski et al., 1986].

Combining the result of this section with that for zonal mean correlations, we are led to the conclusion that the column density of ozone should be fairly well correlated with the latitudinal as well as the longitudinal distribution of the large-scale temperature pattern at the altitude of maximum ozone partial pressure. This picture is consistent with the observation of Stolarski et al. (1986). It also appears that, since temperature and ozone are related, the causes for seasonal and interannual variations of ozone should be sought in factors that affect $\Delta\bar{T} \equiv \bar{T}$ - Te.

Seasonal and Interannual Variations

Recent evidence suggests that the Antarctic ozone hole, although confined inside the polar vortex, should not be viewed as an isolated phenomenon independent of developments outside the polar vortex. Stolarski and Schoeberl (1986) and Schoeberl et al. (1986) pointed out that the seasonal deepening of the ozone hole in spring is accompanied by a parallel intensification of the surrounding maximum in the midlatitudes in the Southern Hemisphere. Furthermore, this occurs in such a manner that the sum of ozone amount from 44^0S to 90^0S remains almost unchanged from August through November. Such a seasonal behavior is most suggestive of dynamical redistribution from the polar region to the midlatitudes [Stolarski and Schoeberl, 1986]. This redistribution may be accomplished by a reverse circulation in the lower stratosphere rising from the cold Antarctic vortex and subsiding over the warm midlatitude region outside the vortex. This mechanism was originally proposed by Tung et al. (1986), who suggested that $\Delta\bar{T}$ may become negative in early spring when Te increases after the sun returns. However, considerable uncertainty remains concerning the nature of radiative properties of the absorbers of solar insolation in early spring that are thought to be responsible for driving the seasonal reverse circulation in the Antarctic lower stratosphere. Tung et al. (1986) calculated an incremental increase (as the sun returns) of heating rate due to ozone alone of 0.5^0K per day in the Antarctic lower stratosphere in October. This is barely sufficient for explaining the seasonal decline of the zonal mean column ozone provided that the enhanced cooling due to the temperature increases, estimated to be about 0.3^0 per day, is compensated by additional heating due to other absorbers. The need for additional absorbers is pointed out by Rosenfield and Schoeberl (1986). Rough estimates by Tung et al. (1986) suggest that radiative heating by volcanic aerosols in the atmosphere may be comparable to ozone heating. A more detailed, but still preliminary calculation by Shi et al. (1986) suggest that the net heating due to stratospheric aerosols may amount to 0.25^0K per day in early October in 1982 over Syowa Station (69^0S).

The interannual variations of column ozone inside and outside the polar vortex also appear to be linked. Although on a percentage basis, the year-to-year decline of the October maximum at midlatitudes is less than the corresponding decline in the minimum inside the polar vortex, in terms of absolute magnitudes both appear to decline at roughly the same average rate (of the order of 10 to 20 Dobsons per year), as can be inferred from Table 2 of Stolarski et al. (1986). Combining this observation with the point mentioned earlier that the spring variations of the minimum and maximum tend to compensate each other, one may tentatively deduce that the interannual decline of October column ozone values is not simply related to the spring decline itself, but that the smaller, 20 Dobson a year decrease occurs mostly prior to August, before the development of the ozone hole in spring. While dynamical redistribution of ozone occurring South of 44^0S may explain most of the seasonal variation of column ozone in spring, namely the deepening of the Antarctic hole and the intensification of the midlatitude maximum, redistribution involving at least the entire Southern Hemisphere is needed for a plausible explanation of the interannual decline of both the Antarctic minimum and the midlatitude maximum. Interannual, changes in poleward transports from the equatorial ozone - production region over an annual time-scale can lead to variations in mid- and high-latitude ozone amounts. Mahlman and Fels (1986) speculated that reduced poleward transport may be caused by a reduction of wintertime planetary/cyclone scale disturbance activity in the troposphere, which also causes the temperature change.

The recent appearence of Antarctic ozone hole may be a result of a coincidental combination of two factors: the heavy loading of the stratosphere of volcanic aerosols, which affects Te, and a decrease of $\bar{T}$ of the Antarctic lower stratosphere, due to a decrease of wave transports. Without the additional heating by volcanic aerosols, the spring upwelling is found to be too weak even if the winter temperature is taken to be as cold as the radiative equilibrium values, as was assumed in the simple model calculation of Tung et al. (1986). Since the large increase after the El Chichon eruption, volcanic aerosols in the Antarctic stratosphere have been observed to decrease slowly but steadily since 1983 [Hofmann et al., 1986]. However, this decrease in aerosol concentration may not necessarily lead to a proportional decrease in the strength of the proposed Antarctic upwelling if it can be compensated by reduced cooling to space caused by the colder temperatures observed in recent years.

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