Radiative–dynamic effects of the Antarctic ozone hole and chemical feedback


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SUMMARY

A southern hemisphere winter/spring is simulated with the Coupled Stratosphere–Mesosphere Interactive Chemistry model, a mechanistic model of the middle atmosphere with an interactive chemistry scheme. Simulations are initialized and forced at the lower boundary with dynamic fields taken from UK Meteorological Office assimilations for 1992. Radiative and dynamic effects of ozone depletion are investigated by comparing runs that include heterogeneous chemistry and hence develop an ozone hole, with otherwise identical control runs with no heterogeneous chemistry and no ozone hole. Temperatures in the ozone hole and control runs start to diverge from around mid-September, by late October the lower stratosphere is up to $\sim$12 K cooler when an ozone hole is present. In the mid and upper stratosphere, a strengthened general circulation and an enhanced radiative heating rate in the ozone-depleted runs lead to slightly higher temperatures at the high southern latitudes than in the control runs. The initially smooth evolution of the temperature difference at the upper levels is interrupted in mid-October by more transient events associated with planetary-wave activity.

The feedback on the chemistry of the physical changes wrought by using the depleted ozone in the radiative calculations is investigated by comparing runs of a chemical transport model forced by winds and temperatures calculated with and without an ozone hole. The ozone-hole induced cooling does not change the spatial extent of the chlorine and bromine activation, and only slightly extends its duration. Consequently, the depth and size of the ozone hole are unchanged by chemical feedback. Local deficits in the ozone column of up to $\sim$30 Dobson units which arise in the cooler run in October are due to differences in horizontal transport. The increased diabatic descent resulting from using the depleted ozone for the radiation calculations does not significantly alter the vertical transport of polar ozone, the rate of which is low compared with the rate of chemical ozone destruction.

KEYWORDS: Chemistry Dynamics Feedback Ozone Radiation

1. INTRODUCTION

The unprecedentedly low column ozone amounts that have occurred over Antarctica during the springs of the last two decades are now known to be due to rapid removal of ozone from the lower stratosphere by catalytic chemical cycles (WMO 1995). Heterogeneous reactions on polar stratospheric clouds (PSCs) release chlorine and bromine from relatively inert reservoir species into more reactive forms which, in the presence of sunlight, catalyse the destruction of ozone (Solomon et al. 1986; Anderson et al. 1989). Anthropogenic emissions have increased the chlorine and bromine loading of the atmosphere and so led to the current severe springtime depletions of Antarctic ozone. The key chemical and dynamic processes controlling the ozone change were elucidated by modelling studies (e.g. Prather and Jaffe 1990; Granier and Brasseur 1991). Ozone destruction continues through winter and early spring until, after the temperature becomes too high to sustain PSCs, chlorine and bromine are re-sequestered in their reservoir species. The rate of sequestration bears an inverse relationship to the extent to which the stratosphere has been denitrified by sedimentation of PSCs, which in turn depends on the antecedent temperatures. The total ozone loss on the seasonal time-scale is thus strongly dependent on the evolution of the temperature of the polar stratosphere; if everything else is equal, the longer sub-PSC temperatures persist, the more ozone loss there will be, although the ozone in the lowermost stratosphere may be entirely removed before the PSCs evaporate.

The temperature of the stratosphere is controlled by the radiative and dynamic heating rates, and because ozone is a radiatively active gas, it, itself, has an influence on the temperature. Ozone heats the stratosphere by absorbing ultraviolet (UV) and visible
light coming from the sun, and infrared (IR) light, mainly in the 9.6 μm band, upwelling from below. As the ozone in the stratosphere is reduced, these sources of heating are also reduced. The impact of Antarctic ozone depletion on stratospheric temperature has been investigated in models of varying complexity. Shine (1986) examines the purely radiative effect of imposing an ozone hole on a one-dimensional model and finds a cooling of ~6 K in the lower stratosphere, accompanied by a smaller warming higher up. A two-dimensional modelling study by Chipperfield and Pyle (1988) ascribes the radiative warming at the upper levels to the increased flux of 9.6 μm radiation passing through the ozone-depleted lower layers to reach the atmosphere above. However, Kiehl et al. (1988) find that, in a general-circulation model (GCM), dynamic changes resulting from the imposition of an ozone hole contribute more to the warming of the upper stratosphere, which maximizes at ~6 K, than do the radiation changes. The dynamic warming is due mainly to increased adiabatic heating resulting from stronger descent over the pole in the presence of an ozone hole. In the lower stratosphere the temperature decrease of ~8 K found by Kiehl et al. is consistent with the radiatively determined response.

Observations also have suggested a link between ozone depletion and lower-stratospheric cooling. Butchart and Austin (1996) note, for the Octobers of 1979 through 1993, a high correlation between the downward trend in the Antarctic ozone column recorded by the Total Ozone Mapping Spectrometer (TOMS) and the brightness temperature at 86 hPa recorded by the Microwave Sounding Unit. (The expected radiative response to the ozone trend is calculated by Forster et al. 1997.) Moreover, time records of stratospheric temperatures from both the Antarctic and Arctic reveal cooling during the 1980s and 1990s coincident with observed ozone depletions (Randel and Wu 1999). On the seasonal time-scale, if the reduced radiative heating resulting from ozone depletion extends the duration of sub-PSC temperatures, then the chemical–radiative interaction provides a route for positive feedback both of ozone destruction and of radiative cooling in the lower stratosphere of the polar vortex. Knowledge of the extent of any such feedback is necessary for forecasting probable and possible responses of polar ozone to future atmospheric conditions, and complements studies of the sensitivity to changes in other radiatively active trace gases (e.g. Austin et al. 1992; Shindell et al. 1998a, b). Of course, the chemical–radiative interaction occurs against the background of the many other factors controlling the meteorological conditions of the polar vortex, such as the tropospheric forcing and the phase of the quasi-biennial oscillation (QBO).

The close coupling of the radiative, dynamic and chemical processes in the polar regions can make it difficult to separate cause from effect; characterizing the important interactions requires the application of radiative–dynamic–chemical models which incorporate the major feedback mechanisms. A number of simulations have been performed with GCMs in which the heating rates and the ozone field are linked via a temperature-dependent parametrization of the ozone chemistry, and the results so obtained compared with those obtained when the ozone hole is not accounted for in the radiative calculations. Studies of this type by Cariolle et al. (1990) and Mahlman et al. (1994) confirm an ozone-hole induced strengthening of the residual circulation, and find a cooling in the lower stratosphere of the order of 6 to 8 K, similar to that seen in the experiments using prescribed ozone fields. The results also suggest that the ozone hole delays the final warming of the Antarctic vortex by around 10–14 days. While such studies illuminate the radiative and dynamic consequences of the ozone hole, they do not address the question of whether there exists a closed chemical feedback loop. That is, whether there is significant self-perpetuation of the ozone-destruction processes via the induced stratospheric cooling. Tackling this problem requires a more detailed treatment
of the chemistry to capture the subtleties of the chemical system, and a mechanism for allowing/disallowing chemical feedback. (Dynamic models with fairly detailed interactive chemistry schemes are in existence and have been used to develop relationships between the QBO, total chlorine and the Antarctic ozone hole (Butchart and Austin 1996), and to examine the sensitivity of ozone and temperature to vertical resolution (Austin et al. 1997).)

This paper describes an experiment performed with a new model created by coupling two existing 'stand alone' models: a mechanistic dynamic model, and a chemical transport model (CTM) with a comprehensive chemistry scheme. A feature of this formulation is that the CTM with one treatment of chemistry can be driven by winds and temperatures previously calculated with a different treatment of chemistry. This affords an insight into how radiative/dynamic changes induced by chemical changes can feed back onto the chemistry. Additionally, the use of observed geopotentials to force the mechanistic model at the lower boundary helps to ensure a realistic simulation of planetary-wave propagation from the troposphere. This is important because both the polar ozone distribution (Austin and Butchart 1992; Butchart and Austin 1996) and the dynamic response to the Antarctic ozone hole (Shindell et al. 1997) show considerable sensitivity to the tropospheric planetary-wave forcing. The work presented below comprises: (i) an examination of the radiative and dynamic impact of the Antarctic ozone hole within the new model, with particular emphasis on the consequences for wave propagation in the stratosphere, and (ii) investigation of how these induced changes feed back onto the ozone distribution in the model via both chemistry and transport.

2. THE COSMIC MODEL

The UK Universities' Global Atmospheric Modelling Programme (UGAMP) Coupled Stratosphere–Mesosphere Interactive Chemistry (COSMIC) model is a recently developed, three-dimensional, mechanistic model of the middle atmosphere that uses a prescribed tropospheric forcing at the lower boundary. Radiative calculations are performed using a time-varying ozone field calculated by a photochemical scheme. The COSMIC model was created by coupling two existing models: the UGAMP stratosphere–mesosphere model (USMM) and an offline CTM. These dynamic and chemical models are run in tandem, synchronized with the OASIS code, a suite of programs for coupling numerical models (Terray and Thual 1995). Every 6 hours, zonal and meridional winds and temperatures are passed from the USMM to the CTM, and ozone values are passed from the CTM to the USMM for use in the radiation calculations. If the resolutions of the two models differ, the fields being passed are interpolated to the three-dimensional grid of the receiving model. Some details of the USMM and the CTM are given below.

(a) The USMM

The USMM is a development of the Extended UGAMP General Circulation Model (EUGCM) and, as such, traces its origins from cycle 27 of the European Centre for Medium-Range Weather Forecasts (ECMWF) forecast model (Tibaldi et al. 1990). Some details of the history of the USMM are given by Mote et al. (1998). Unlike the EUGCM, the USMM does not include the troposphere, but uses a prescribed tropospheric forcing at the lower boundary. For this study, the model was configured to have 34 isobaric levels extending from 89.5 to 0.01 hPa, giving a vertical resolution in the stratosphere and mesosphere of around 1.6 km. The levels are shown in Fig. 1.
The lower boundary, at 100 hPa, was forced with observed geopotentials. Positioning the boundary at 100 hPa ensures that it lies above the tropical tropopause so convection can be neglected. Also, at this pressure the smaller-scale upwardly propagating convection waves will have been largely filtered from the observed geopotential field, leaving only the longer wavelengths which are more reliably captured in the assimilated data. The USMM runs discussed below were at a spectral resolution of T21, i.e. triangular truncation of the spherical harmonics retaining the first 21 spectral coefficients. The time step was 15 minutes. Gravity-wave drag was parametrized by Rayleigh friction, with an altitude-dependent damping coefficient ranging from $1 \times 10^{-7} \text{ s}^{-1}$ at 50 km and below to $8 \times 10^{-6} \text{ s}^{-1}$ at 80 km.

The radiation code, based on the middle-atmosphere scheme (Shine 1989), includes short-wave heating due to ozone and diatomic oxygen, and long-wave heating due to CO$_2$, ozone and H$_2$O. Below 10 hPa, the time-varying ozone passed from the chemical model was used in the heating calculations, and between 10 and 5 hPa this was merged with a monthly ozone climatology. The contribution of H$_2$O to the heating is calculated as a function of local temperature with a globally invariant, fixed mixing ratio profile assumed implicitly; the model H$_2$O is not used. Seasonally and meridionally varying, upwelling fluxes of IR radiation in the 9.6 $\mu$m and 15 $\mu$m wavelengths were calculated using climatological temperatures assuming that emission at 9.6 $\mu$m originates at 700 hPa and emission at 15 $\mu$m originates at 130 hPa (W. A. Lahoz (1996), personal communication).

(b) The offline chemical transport model

The CTM used here is an updated version of the model described by Chipperfield et al. (1993). This isobaric model was run at the same 34 pressure levels as the USMM, with chemistry switched off above 1.5 hPa. The concentrations of chemical species in
TABLE 1. SPECIES IN THE PHOTOCHEMICAL SCHEME

<table>
<thead>
<tr>
<th>Transported species</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \text{O}_3 + \text{O}^3\text{P} + \text{O}^{(1}\text{D}) ), \text{H}_2\text{O}_2</td>
</tr>
<tr>
<td>( \text{N} + \text{NO} + \text{NO}_2 ), \text{NO}_1, \text{N}_2\text{O}_3, \text{HNO}_3, \text{HO}_2\text{NO}_2</td>
</tr>
<tr>
<td>( \text{Cl} + \text{ClO} + 2\text{Cl}_2\text{O}_2 ), \text{ClONO}_2, \text{HCl}, \text{HOCl}, \text{OCIO}</td>
</tr>
<tr>
<td>( \text{Br} + \text{BrO}, \text{BrONO}_2, \text{BrCl}, \text{HBr}, \text{HOBr} )</td>
</tr>
<tr>
<td>\text{CH}_4, \text{N}_2\text{O}, \text{CO}, \text{H}_2\text{O}</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Non-transported species</th>
</tr>
</thead>
<tbody>
<tr>
<td>\text{CH}_3, \text{CH}_3\text{O}_2, \text{CH}_3\text{O}, \text{CH}_2\text{O}, \text{HCO}, \text{CH}_3\text{OOH}</td>
</tr>
<tr>
<td>\text{H}, \text{OH}, \text{HO}_2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Fixed species</th>
</tr>
</thead>
<tbody>
<tr>
<td>\text{O}_2, \text{N}_2, \text{H}_2</td>
</tr>
</tbody>
</table>

Family groups are enclosed in brackets.

The bottom model level, centred at 89.5 hPa, were held constant at their initial values to provide an unchanging lower-boundary condition. A horizontal resolution of 5° latitude by 7.5° longitude was used, similar to the T21 resolution of the dynamic model. The chemical tracers were advected with the winds passed from the USMM, using the second-order moments advection scheme of Prather (1986). This is an accurate, non-diffusive scheme well suited to preserving sharp horizontal or vertical gradients. Vertical velocities in the CTM are calculated by integrating the divergence of the supplied horizontal winds.

The comprehensive photochemical scheme includes 40 chemical species, 102 gas-phase thermal reactions, and 26 photolytic reactions. The modelled species and their grouping into families are shown in Table 1. The chemical continuity equations are integrated by the semi-implicit symmetric method of Ramaroson et al. (1992) using a time step of 15 minutes. Rate coefficients were calculated using the temperatures passed from the USMM. A treatment of heterogeneous chemistry on hydrated \text{HNO}_3 (type 1) and water-ice (type 2) PSCs is included (Chipperfield et al. 1993, 1994). The occurrence of type 1 PSCs is calculated from the model fields of \text{HNO}_3, \text{H}_2\text{O} (both prognostic variables) and temperature using the algorithm of Hanson and Mauersberger (1988). The occurrence of type 2 PSCs is calculated from Teten’s equation in the form given by Murray (1967). PSCs are assumed present wherever they are thermodynamically possible. Type 1 and type 2 particles are sedimented from the model with fall velocities appropriate for particles of radius 1 \( \mu \text{m} \) and 10 \( \mu \text{m} \) respectively. The heterogeneous reactions included in the chemical scheme are shown in Table 2. Chemical processing on liquid sulphate aerosol was not included in the runs discussed here.

Photolysis rates are calculated with a scheme based on Lary and Pyle (1991) and Lary (1991). The scheme uses a four-dimensional look-up table to interpolate pre-computed photodissociation rate coefficients to the appropriate pressure, solar zenith angle, temperature and ozone column. Multiple scattering and spherical geometry are accounted for, and photolysis rates can be calculated for solar zenith angles up to 96°. Photochemical kinetic data were, in the main, taken from DeMore et al. (1994).

3. Runs performed

The runs performed for this study are summarized in Table 3. The basis of the study is a comparison of the meteorology and chemistry of separate runs of the COSMIC model that included and excluded the radiative effect of the ozone hole on the model.
TABLE 2. Heterogeneous reactions in the photochemical scheme

\[
\begin{align*}
N_2O_5(g) + H_2O_2(a) & \rightarrow 2HNO_3(g) \\
ClONO_2(g) + H_2O_2(a) & \rightarrow HOCI(g) + HNO_3(g) \\
ClONO_2(g) + HCl(g) & \rightarrow ClO_2(g) + HNO_3(g) \\
N_2O_5(g) + HCl(g) & \rightarrow ClO_2(g) + NO_3(g) + HNO_3(g) \\
HOCI(g) + HCl(g) & \rightarrow 2ClO_2(g) + H_2O_2(g) \\
HOBr(g) + HBr(g) & \rightarrow 2BrO_2(g) + H_2O_2(g) \\
HOCl(g) + HBr(g) & \rightarrow BrCl(g) + H_2O_2(g) \\
HOCI(g) + HBr(g) & \rightarrow BrCl(g) + H_2O_2(g) \\
BrONO_2(g) + HBr(g) & \rightarrow 2BrO_2(g) + HNO_3(g) \\
BrONO_2(g) + HCl(g) & \rightarrow BrCl(g) + NO_3(g) \\
ClONO_2(g) + HBr(g) & \rightarrow BrCl(g) + HNO_3(g) \\
BrONO_2(g) + H_2O_2(a) & \rightarrow HOBr(g) + HNO_3(g) \\
N_2O_5(g) + HBr(g) & \rightarrow BrO_3(g) + NO_3(g) + HNO_3(g)
\end{align*}
\]

TABLE 3. Summary of the runs performed

<table>
<thead>
<tr>
<th>Run</th>
<th>Model</th>
<th>Start date (1992)</th>
<th>Initial dynamics</th>
<th>Initial ozone</th>
<th>Lower-boundary forcing</th>
<th>CTM forcing</th>
<th>Heterogeneous chemical reactions</th>
</tr>
</thead>
<tbody>
<tr>
<td>OH-92A</td>
<td>COSMIC</td>
<td>14 Aug.</td>
<td>UKMO</td>
<td>MLS</td>
<td>UKMO</td>
<td>Coupled</td>
<td>Yes</td>
</tr>
<tr>
<td>CO-92A</td>
<td>COSMIC</td>
<td>14 Aug.</td>
<td>UKMO</td>
<td>MLS</td>
<td>UKMO</td>
<td>Coupled</td>
<td>No</td>
</tr>
<tr>
<td>OH-92B</td>
<td>COSMIC</td>
<td>06 Aug.</td>
<td>UKMO</td>
<td>MLS (zm)</td>
<td>UKMO</td>
<td>Coupled</td>
<td>Yes</td>
</tr>
<tr>
<td>CO-92B</td>
<td>COSMIC</td>
<td>06 Aug.</td>
<td>UKMO</td>
<td>MLS (zm)</td>
<td>UKMO</td>
<td>Coupled</td>
<td>No</td>
</tr>
<tr>
<td>OH-92C</td>
<td>COSMIC</td>
<td>06 Aug.</td>
<td>UKMO</td>
<td>MLS (zm)</td>
<td>ECMWF</td>
<td>Coupled</td>
<td>Yes</td>
</tr>
<tr>
<td>CO-92C</td>
<td>COSMIC</td>
<td>06 Aug.</td>
<td>UKMO</td>
<td>MLS (zm)</td>
<td>ECMWF</td>
<td>Coupled</td>
<td>No</td>
</tr>
<tr>
<td>CHEM-92A</td>
<td>CTM</td>
<td>13 Sep.</td>
<td>OH-92A</td>
<td>OH-92A</td>
<td>n/a</td>
<td>CO-92A</td>
<td>Yes</td>
</tr>
</tbody>
</table>

zm indicates a zonal-mean ozone field. See text for further explanation.

physics and dynamics. Several pairs of fully-coupled COSMIC runs, identical except that one run of the pair included heterogeneous chemical reactions and the other did not, were carried out. The runs with heterogeneous chemistry simulate the true conditions of the Antarctic winter/spring whereas the control runs without heterogeneous chemistry simulate the conditions in the absence of an ozone hole caused by heterogeneous processing. The simulations cover the period from early August, when rapid loss of ozone begins as the Antarctic vortex emerges from polar night, through to the time of the vortex break-down following the final warming. Differences in these pairs of runs are compared with differences between runs with the same chemistry but different initial or lower-boundary conditions.

In addition to the runs of the fully-coupled model, other runs were carried out using winds and temperatures from previous control runs to drive the CTM with heterogeneous reactions included. These runs are equivalent to running the model with full chemistry, but no radiative–chemical coupling. Comparing the chemistry in these runs with the chemistry in the fully-coupled runs reveals the significance for the chemistry of allowing it to interact with the radiative calculations and hence the dynamics.

(a) **Dynamic model**

The USMM runs were initialized with wind and temperature fields from the appropriate day of the UK Meteorological Office (UKMO) assimilated data (Swinbank and O’Neill 1994); an extrapolation using thermal-wind balance in the zonal mean was applied in the mesosphere between the top UKMO level at 0.3 hPa and the top model level at 0.01 hPa. At the lower boundary, the runs were forced with fields of geopotential taken either from the UKMO assimilated data, at 24-hourly intervals, or from ECMWF
assimilated data, at 6-hourly intervals. The geopotentials at intermediate times were found by linear interpolation.

(b) Chemical transport model

For the chemical initialization of the CTM, the southern-hemisphere ozone field was taken from Microwave Limb Sounder (MLS) observations in the 205 GHz band. To reduce random errors, ozone retrievals from the day and night sides of the orbit were gridded separately then averaged. A single day’s measurement was considered sufficiently representative for this relatively long-lived species, and initializing with ozone and dynamic fields retrieved on the same day provides consistency in the vortex regions where both fields may be highly azonal. The northern-hemisphere ozone and the other chemical species were initialized with zonal-mean fields taken from a two-dimensional model, with the concentration of the chlorine and bromine compounds scaled to give total inorganic chlorine and inorganic bromine loadings of 3.4 parts per billion (10^9) by volume (p.p.b.v.) and 18 parts per trillion (10^12) by volume respectively. The initial H_2O field was calculated from the model methane (CH_4) field assuming 2CH_4 + H_2O = 6.5 p.p.m.v. (parts per million by volume).

4. RESULTS AND DISCUSSION

In the discussions that follow, individual runs are referred to by the names shown in the first column of Table 3. COSMIC runs which included heterogeneous chemistry, and hence produced rapid destruction of Antarctic ozone, have names beginning OH, for ‘ozone hole’; the corresponding runs without heterogeneous chemistry have the OH replaced by CO, for ‘control’. Non-interactive runs of the chemistry model on its own are labelled CHEM.

(a) Ozone evolution

Runs OH-92A and CO-92A were started from 14 August 1992 with identical ozone fields taken, in the southern hemisphere, from MLS observations on that day. As discussed earlier, the model runs with and without heterogeneous reactions were intended to simulate current and pre-ozone hole conditions respectively. To compare the modelled ozone columns with the observed columns from the different eras, the modelled ozone has been plotted alongside Nimbus 7 TOMS data from the years 1979 and 1992. The October mean of the 1979 TOMS data is similar to that of the 1970 to 1973 data from Nimbus 4 BUV (Backscatter Ultra Violet) (Stolarski et al. 1997) so the 1979 data can be taken as representative of pre-ozone hole behaviour. The evolutions of the modelled and observed ozone are shown in Fig. 2(a) as area-weighted averages of the column polewards of 65°S. The initial difference between the column amount in the model and in the 1992 TOMS data is largely due to the modelled column not accounting for ozone below 100 hPa, but additional differences arise because TOMS data are not available from the high latitudes that are in polar night at this time. On 14 August, when the model runs begin, the TOMS column from 1992 is already significantly smaller than its counterpart from 1979, and the difference continues to grow. The two model runs were initialized with the same ozone field, and although the column in OH-92A immediately starts falling below the column in CO-92A, the peak difference in the modelled columns, in early October, is only about 40% of the synchronous difference between the TOMS columns. However, Fig. 2(b), which emphasizes the change in the ozone columns over the analysis period, reveals that the divergence of the two modelled
Figure 2. (a) The trend in the ozone columns from Nimbus 7 TOMS 1979 and 1992, and from the model runs OH-92A and CO-92A (see text). All fields are averaged, with area weighting, between 65°S and the pole. (b) The departure of the quantities in (a) from their initial values on 14 August 1992.

columns is similar to that of the two observed columns. The initial downward trend in the two TOMS columns is partly due to the measurements extending polewards as the polar-night terminator retreats, and some differences in the TOMS trends will be due to the dynamics responsible for the day-to-day variability, but the underlying differences can probably be largely ascribed to excess chemical destruction occurring in 1992. By early October, when TOMS measurements cover the whole vortex, the difference in the amount by which the two TOMS columns have changed is similar to the difference in the amount by which the two modelled columns have changed. The latter difference being confidently attributable to chemical differences. That the separation of the modelled trends tends to be slightly smaller than the separation of the observed trends is probably a consequence of the modelled ozone being chemically active only above 79 hPa. Nevertheless, the majority of the ozone destruction takes place above this height, and, overall, the results suggest that the chemical destruction is being modelled reasonably accurately in OH-92A; the difference between this run and CO-92A can be taken as representative of the anthropogenic effect on Antarctic ozone.
(b) Temperature evolution

(i) Comparison of simulations and observations. Figure 3 compares the evolution of the zonal-mean temperature at 56 hPa from OH-92A and from the UKMO assimilated data for 1992 (which were used to initialize the model). At the southern polar latitudes, the model cools relative to the assimilated data, with differences of up to \( \sim 14 \) K developing by day 60 (13 October). The cooling trend is established from near the beginning of the run, before the chemical feedback has a significant impact on the model temperatures, and so is a feature of the basic model configuration, independent of the interactive chemistry (bearing in mind that the model was initialized with the observed ozone). However, the maximum temperature difference is approximately doubled by the inclusion of radiative feedback from the ozone hole (see next subsection). Manney et al. (1996) show that UKMO temperatures in the lower vortex have a slight warm bias, relative to radiosonde data, when temperatures are below about 200 K, but this bias is small in relation to the discrepancy between the UKMO and modelled temperatures. Many factors, such as the model resolution, the treatment of upwelling long-wave radiation, and the use of Rayleigh friction to represent gravity-wave drag, may play a role in producing the excessively low temperatures in the model; a too-cold southern polar vortex is a common feature of global circulation models, and a full investigation of its origin in this experiment is beyond the scope of the current paper. Nevertheless, notwithstanding the cold bias, the modelled temperature evolution for the most part closely resembles the observed evolution. This confirms that the geopotential forcing applied at the model’s lower boundary is, to some degree, constraining the model to the observations. No attempt was made to ‘tune’ the model to improve the temperatures because such adjustment could impinge on the internal feedback processes. The implications of the overly-low temperatures for the model results are discussed in section 4(c).

(ii) Impact of ozone depletion on model temperatures. Figure 4 shows time versus pressure sections of the differences in the zonal-mean ozone and zonal-mean temperature from OH-92A and CO-92A at 75°S. This latitude was chosen as representative of conditions inside the polar vortex; plots at other high southern latitudes look similar. Because the other species used for the heating calculation are fixed, all physical and dynamic differences between the two runs must stem from the ozone difference. As already mentioned, the ozone in the lowest model level, centred on 89.5 hPa, was held constant in all the runs performed. Although this condition affects the local heating rates and thus the local temperature, tests show that reducing the ozone in the bottom level has little effect on the IR heating rates at the levels above. This implies that the fixed chemical boundary condition should not affect the temperature difference at the upper levels. In the lower stratosphere, the ozone difference grows steadily from the start of the run until about day 30, and then remains fairly constant for the next 40 days except that, above the region of maximum ozone difference, the disparity continues to grow slightly. The corresponding trend in temperature difference in the lower stratosphere is somewhat different: the temperature difference remains below 0.5 K until about day 25 (8 September), and then starts to grow at an increasing rate until it reaches a maximum of \( \sim 6 \) K at 55 hPa around day 70 (23 October). (A peak temperature difference of \( \sim 12 \) K is found close to the pole at this time.) The increasing rate of growth of the temperature difference while the ozone difference remains approximately constant is consistent with the increasing insolation intensifying the difference in the heating rates arising from the ozone disparity.
Figure 3. Time versus latitude sections of the zonal-mean temperature at 56 hPa from (a) OH-92A, (b) the UKMO assimilated data, and (c) the difference OH-92A minus UKMO. Contour interval is 2 K both for the temperatures and for the difference. The zero contour is dashed; negative contours are dotted. See text for further explanation.

The evolution of the temperature difference in the middle and upper stratosphere and mesosphere is more complex than in the lower stratosphere. During the early part of the run there is a slight temperature excess above the ozone-depleted region of OH-92A, but around day 57 (10 October), the vertical structure of the temperature difference is abruptly disrupted, and the ozone-depleted run changes from being warmer in the upper stratosphere and mesosphere to being colder throughout the whole depth of the atmosphere. Following this event, the temperature excess in the upper levels of OH-92A is re-established, with the relative warming starting in the mesosphere and propagating downwards as far as the mid stratosphere. This behaviour cannot easily be explained from purely radiative considerations, and is strongly suggestive of a dynamic origin. This raises a question concerning the extent to which dynamic differences between runs with and without heterogeneous chemistry are systematic and reproducible. It may be that apparently significant differences between the runs are, in fact, due to
Figure 4. (a) Time versus pressure section of the zonally-averaged ozone at 75°S from OH-92A minus the equivalent field from CO-92A (see text). (b) As (a) but for temperature instead of ozone. Note that the altitude range extends to 1 hPa in (a), and to 0.01 hPa in (b). The contour intervals are 0.25 p.p.m.v. for the ozone difference, and 1 K for the temperature difference. Negative contours are dashed.
the nonlinearity of the model’s dynamic processes chaotically amplifying much smaller differences stemming from the radiative feedback via the ozone in the chemical scheme. Figure 5 shows a time versus pressure section of the actual temperature from OH-92A. The temperature evolution in CO-92A (not shown) is, aside from the warmer lower stratosphere, very similar; there is no marked discontinuity in the temperature of either run corresponding to the discontinuity in the difference plot around day 57. This suggests that the temperature differences are caused by rather subtle effects, and are not due to any major dynamic events occurring in one run and not the other.

To investigate the reproducibility of the features seen in Fig. 4, a second pair of runs was performed. The new runs, OH-92B and CO-92B, were identical to OH-92A and CO-92A respectively except that (1) they were started from 6 August 1992, dynamically initialized with UKMO data for that day; and (2) they used a zonal initial ozone field obtained by interpolating between MLS ozone observations from 12 July and 14 August 1992. OH-92B and CO-92B thus comprise a repeat of OH-92A and CO-92A but starting from different initial conditions. A temperature comparison plot from OH-92B and CO-92B (equivalent to that from OH-92A and CO-92A shown in Fig. 4(b)) is presented in Fig. 6. It is striking that the evolution of the temperature difference between the new runs is very similar to that of the original runs. The similarity and the distinctive structure of the evolving temperature difference in the two experiments suggest that the difference is largely due to the different ozone amount in the two runs forming each experiment. This conclusion is supported by the fact that plots of the temperature difference between the runs with the same chemistry, but different initial conditions (not shown), exhibit no such distinctive structure. As a further test of the dynamic sensitivity, OH-92B and CO-92B were repeated using 6-hourly geopotentials from the ECMWF as the lower-boundary forcing instead of the 24-hourly fields from the UKMO. The temperature difference between these two new runs, OH-92C and CO-92C, is presented in Fig. 7, and again the same distinctive pattern is in evidence. This suggests that any role played by the lower-boundary forcing in controlling the development of the temperature difference between the ozone-depleted and control runs is consistent with the forcing in the real atmosphere and is not due to any characteristics peculiar to a particular dataset. Having established that the main features of the temperature differences between OH-92A and CO-92A are reproducible and largely independent of small changes in the boundary conditions, further diagnosis will focus on these two runs.
Figure 6. Time versus pressure section of the zonally-averaged temperature at 75°S from OH-92B minus the equivalent field from CO-92B (see text). Contour interval is 1 K. Positive contours are shown in unbroken black, the zero contour in unbroken white, and negative contours in dashed white.

(c) Radiative and dynamic heating

Stratospheric temperature is controlled by a combination of radiative and dynamic processes: the thermodynamic equation can be written in simple form as

\[
\frac{\partial T}{\partial t} = Q_r + Q_d
\]

(1)

where \(\partial T/\partial t\) is the net temperature tendency, \(Q_r\) is the net radiative heating and \(Q_d\) is the net dynamic heating, including horizontal advection. The dynamic term comprises mainly adiabatic heating associated with vertical air motions.

Figure 8 shows the net radiative heating rate, the dynamic heating rate, and the resultant temperature tendency from CO-92A at 56 hPa and 4.5 hPa as 9-day, centred, running means of their zonal averages at 75°S. (Dynamic heating rates were not calculated explicitly, but were inferred from the temperature tendencies and the net radiative heating rates using Eq. (1).) The, relatively small, differences between these heating rates and their equivalents from OH-92A are shown separately on a different scale, as are the temperatures from the two runs. At 56 hPa, the temperature tendency is dominated by the rapidly fluctuating, generally positive, dynamic heating term.
Figure 7. Time versus pressure section of the zonally-averaged temperature at 75°S from OH-92C minus the equivalent field from CO-92C (see text). Contour interval is 1 K. Positive contours are shown in unbroken black, the zero contour in unbroken white, and negative contours in dashed white.

countered by the more slowly varying, negative, radiative heating term. Over the first 40 days of the runs, the radiative heating in OH-92A gradually falls below that in CO-92A (see Fig. 8(b)) leading to a maximum difference of around 0.2 K day$^{-1}$. This excess radiative cooling in OH-92A is partially offset by a greater dynamic heating, but the net temperature tendency is generally smaller (less positive) in OH-92A than in CO-92A. After day 40, the difference in the radiative heating rates starts to fall, and around day 80 its sign reverses, indicating that the greater long-wave cooling due to the higher temperature of CO-92A now outweighs the greater solar heating resulting from the extra ozone in this run. At around the same time as the difference in the radiative heating rates starts to fall, the difference in the dynamic heating rates begins to oscillate and develop deep peaks and troughs, so that in the latter half of the period shown, the relative temperature trend between the two runs owes more to dynamic than to radiative differences.

Turning to 4.5 hPa, Fig. 8(e) shows that in the early stages of the runs, both the radiative and the dynamic heating rates are slightly greater in OH-92A than in CO-92A, explaining the initial slight warming of the upper levels of OH-92A relative to CO-92A. The disruption of the smooth evolution of the temperature difference at the upper levels
Figure 8. (a) Net (solar plus long wave), radiative and dynamic heating rates and temperature tendency from C0-92A zonally-averaged at 75°S and 56 hPa. All quantities are 9-day running means. (b) The difference between the quantities in (a) and their counterparts from OH-92A (i.e. OH-92A minus CO-92A). (c) Temperatures from OH-92A and CO-92A zonally-averaged at 75°S and 56 hPa. (d), (e) and (f) As (a), (b) and (c) but at 4.5 hPa. See text for further explanation.

seen in Fig. 4 around day 50 coincides with a sudden increase in the disparity of the dynamic heating rates. First the dynamic heating in OH-92A falls below that in CO-92A, causing a temporary reversal in the sign of the temperature difference, then it shows a rapid relative increase, leading to a peak positive difference; thereafter, the difference reduces, but the dynamic heating in OH-92A remains significantly greater than in CO-92A for the remainder of the run. The temperature differences caused by the differences in dynamic heating induce differences in the radiative heating which are of opposite
sign to and slightly lag the dynamic differences. Thus, unlike at 56 hPa, the most significant radiative differences at 4.5 hPa occur only after temperature differences have been induced by dynamics. The general pattern of temperature response to the ozone hole seen here with the radiative cooling of the ozone-depleted region accompanied by a dynamically induced warming above is similar to that found in other GCMs (Kiehl et al. 1988; Cariolle et al. 1990; Shindell et al. 1997). A comparison of the dynamic behaviours of the ozone hole and control runs is the subject of the next subsection.

(d) Vortex behaviour and planetary-wave diagnostics

In the COSMIC model the forcing of the stratosphere by the troposphere is represented by the geopotential field applied to the model’s lower boundary. Runs OH-92A and CO-92A were forced with identical, observed geopotentials, and thus were somewhat constrained to the same dynamic state. Nonetheless, radiative differences between the runs may affect the propagation of the applied forcing to the levels above, as well as impacting the dynamics at these levels more directly. One indicator of dynamic differences is the relative form and strength of the polar vortices. Figure 9 compares potential vorticity (PV) maps from OH-92A and CO-92A on day 78 (31 October), at 520 K potential temperature (θ), approximately one scale height above the lower boundary. At this time and height the temperature difference between the runs is near its maximum, and the vortex has a distinctive shape which should highlight any differences. Even so, the PV fields from OH-92A and CO-92A are similar in shape, and the −40 PVU (1 PVU = 1 × 10^{-6} K m^2 kg^{-1} s^{-1}) contours marking the approximate vortex edge enclose almost exactly the same areas. The PV magnitudes within the vortex are, however, markedly larger in OH-92A. The difference plot shows a maximum inequality of ~25% in the vortex core, while around the edge of the vortex the differences are small. This pattern of difference prevails throughout the model runs even in late November and December when the vortex is eroding and becoming increasingly distorted. The evolution of the minimum PV values at 520 K, shown in Fig. 10, indicates that during spring the weakening of the vortex in the ozone-depleted OH-92A lags that in CO-92A by about 12 days. This relative strengthening of the lower-stratospheric vortex is the expected dynamic response to the increased meridional temperature gradient arising from the excess radiative cooling in OH-92A.

In the middle stratosphere, above the ozone-depleted region, the induced zonal-mean temperature difference is smaller, and of changeable sign. Figure 11 presents PV maps from OH-92A and CO-92A at 1100 K (~6 hPa) on day 62 (15 October). This is during the period when OH-92A is temporarily cooler than the control run. Both maps have a form characteristic of a planetary-wave breaking event, with a filament of high PV air emanating from the vortex. As at 520 K, the vortex shapes, and the PV fields in general, are similar in the two runs, but the difference plot reveals a more complex relationship than was present at the lower height. At 1100 K the discrepancy between the runs is not correlated simply with the PV magnitude, and there are both positive and negative differences within the vortex region. While the similarity of the PV fields themselves evinces that similar planetary-wave events are taking place in the two runs, the filamentary appearance of the difference field suggests that the discrepancies may stem from minor differences in these wave processes. Examination of plots from other days reveals that the greater structure in the PV difference at 1100 K compared to 520 K is a robust feature present throughout the run, and it is found also in the differences in the temperature and in the vertical velocities. It appears, therefore, that whereas in the lower stratosphere the dynamic differences are largely explainable from the zonal-mean perspective, in the middle atmosphere the wave differences must be invoked also.
Figure 9. Isentropic polar stereographic maps at 520 K of the potential vorticity (PV) from (a) OH-92A, (b) CO-92A, and (c) the percentage difference between them, i.e. \((\text{OH-92A} - \text{CO-92A})/\text{OH-92A}\) \times 100. The date is 31 October, and the area shown is from the south pole to 30°S. The contour interval is 10 PVU for the PV (1 PVU = \(1 \times 10^{-6}\) K m²kg⁻¹s⁻¹), and 2% for the difference. See text for further explanation.

The planetary waves in the model were examined in more detail by performing Fourier decompositions of the geopotential fields at 75°S. Time versus pressure sections of the amplitudes of wave numbers 1 and 2 from OH-92A and CO-92A are shown in Fig. 12. (Wave number 3 behaves similarly to wave 2, but with smaller amplitude.) Wave 1 has a minimum just before day 40, then grows rapidly from around day 40 onwards, as does wave 2, the amplitudes peaking near day 45 and 50 respectively. For both waves the maximum amplitude is located near 1 hPa, although with wave 1 the altitude of the greatest amplitude decreases with time. At around the same time as the wave amplitudes are peaking near 1 hPa, the amplitudes at the levels above start to reduce quite rapidly. Over the next ~25 days this reduction travels downward until there is no significant wave activity above about 10 hPa. It is during this period of diminishing
wave activity that the largest differences between OH-92A and CO-92A arise; wave numbers 1 and 2 (and 3) in OH-92A retain a larger amplitude than their counterparts in CO-92A. This period coincides with the reversal of the temperature difference in the middle stratosphere (Fig. 4), and the initiation of relatively large differences in the dynamic heating rates (Fig. 8). Planetary-wave transience and dissipation can impact the dynamic heating rates through the momentum transferred to the mean flow. Waves breaking in the high latitudes usually decelerate the prevailing westerly wind, inducing a response in the meridional circulation which leads to an enhanced downward flow over the pole and a greater adiabatic heating. It seems that the waves breaking slightly earlier in the control run heat the middle stratosphere and reverse the zonal-mean temperature gradient. Several days later the breaking waves in OH-92A re-establish (and increase) the temperature excess in that run. Studies with other models have also attributed an increased dynamic heating above the ozone hole to enhanced wave activity (Butchart and Austin 1996; Shindell et al. 1997).

The influence of planetary waves on the zonal-mean flow is indicated by the divergence of the Eliassen–Palm (EP) flux (e.g. Andrews et al. 1987). The EP flux divergence is negative where waves are exerting an easterly zonal force on the atmosphere and positive where they are exerting a westerly force. Comparing the divergences in OH-92A and CO-92A reveals that they are similar up until about day 50, when larger differences start to arise, synchronous with the aforementioned differences in the wave amplitudes. Figures 13(a), (b) and (c) show, for the middle and high southern latitudes, the transformed Eulerian-mean divergences from OH-92A, CO-92A and their difference averaged between day 50 and day 70. While the fields themselves have the same general form, the difference plot shows inequalities of both signs, forming a complex pattern. This disturbed difference field is what might be expected from a period of strong wave dissipation in both runs. Figures 13(d), (e) and (f) show the same fields averaged between day 70 and day 90, by which time the waves have largely disappeared from the middle stratosphere. The EP flux divergence is now much less negative in both runs, but the difference between the runs is more systematic: the magnitude is greater in OH-92A virtually everywhere above ~30 hPa. This, by the downward control principle (Haynes et al. 1991), implies a greater mean descent, and hence a greater adiabatic heating in OH-92A. In the stratosphere, the wave forcing of the zonal flow as indicated by the EP flux divergence always dominates that from the imposed Rayleigh friction. Time–pressure
sections of the zonal-mean zonal wind at 75°S are presented in Fig. 14. These show that the top-down disappearance of the planetary waves from the model coincides with the reversal of the wind direction from westerly to easterly. Quasi-stationary planetary waves cannot propagate in easterly flows because their intrinsic phase propagation must be westward (Charney and Drazin 1961). The difference plot indicates that initially there is little between the zonal winds in the two runs; below about 1 hPa the westerlies in OH-92A are marginally (less than 1 m s⁻¹) stronger, and above that pressure they are marginally weaker. But as the wind at the upper levels starts to reverse direction, the remaining westerlies in the stratosphere of OH-92A become stronger than those in CO-92A, and stay so until just before day 120.
Overall, the dynamic diagnostics indicate a relatively small difference between the runs up until about the end of September when easterly winds start to appear at the upper model levels. This heralds a more chaotic period as the planetary waves dissipate in the mid and upper stratosphere. Thereafter there is a pronounced relative strengthening of the residual circulation in the ozone-depleted run, with enhanced descent and stronger westerlies in the lower stratosphere. This strengthening of the residual circulation owing to Antarctic ozone depletion is a finding common to several studies (e.g. Mahlman et al. 1994; Butchart and Austin 1996; Shindell et al. 1997).

(e) Chemical results

Having examined the radiative and physical effects of the ozone hole, we turn now to the impact of these effects on the ozone chemistry. In particular, the question of whether the lower temperatures resulting from ozone depletion can lead to a greater enhancement of reactive chlorine and bromine, and hence produce a positive feedback on the ozone loss. To address this question, the CTM was run on its own with the same chemical scheme as was used in OH-92A, i.e. including heterogeneous reactions, but forced with winds and temperatures from CO-92A, which had no heterogeneous reactions. This run, CHEM-92A, was integrated for 45 days starting from 13 September 1992 with initial dynamic and chemical fields taken from OH-92A on that date (day 30). September 13
Figure 13. Zonal-mean, latitude versus pressure sections of the Eliassen–Palm flux divergence from (a) and (d) OH-92A, (b) and (e) CO-92A, and (c) and (f) the differences OH-92A minus CO-92A. Results have been averaged for two separate periods: (a), (b) and (c) days 50 through 70, and (d), (e) and (f) days 70 through 90. The contour intervals are 2 m s\(^{-1}\) day\(^{-1}\) for the divergence and 0.2 m s\(^{-1}\) day\(^{-1}\) for the differences. The zero contour is dashed; negative contours are dotted. See text for further explanation.

was chosen as the start date for CHEM-92A because before then there was little physical difference between OH-92A and CO-92A.

Because CHEM-92A and OH-92A used the same chemical scheme, any differences in the evolutions of the chemical species in these two runs are attributable to differences in the forcing winds and temperatures: the forcings in CHEM-92A were calculated using undepleted ozone, and those in OH-92A with depleted ozone. Figure 15 shows the evolution of the zonal-mean temperature from OH-92A, and the zonal-mean reactive chlorine, ClO\(_x\) (= Cl + ClO + 2Cl\(_2\)O\(_2\)), and reactive bromine, BrO\(_x\) (= Br + BrO + BrCl) from OH-92A and CHEM-92A at 56 hPa. (BrCl is here included in the BrO\(_x\) family because it is the major night-time reservoir of BrO when ClO\(_x\) is enhanced.) The initial ClO\(_x\) and BrO\(_x\) abundances, taken from OH-92A, are highly enhanced
Figure 14. Time versus pressure sections of the zonal-mean zonal wind at 75°S from (a) OH-92A, (b) CO-92A, and (c) the difference OH-92A minus CO-92A. The contour intervals are 10 m s⁻¹ for the wind and 1 m s⁻¹ for the difference. Westerly winds and positive (westerly) differences are shown with unbroken contours; easterly winds and negative differences are shown with dotted contours. The zero contour is dashed. See text for further explanation.
Figure 15. Zonal-mean, time versus latitude sections of (a) and (b) the temperature, (c), (d) and (e) ClO$_x$, and (f), (g) and (h) BrO$_x$ fields at 56 hPa from (a), (c) and (f) OH-92A and (d) and (g) CHEM-92A, and (b), (e) and (h) the differences OH-92A minus CHEM-92A. The heavy unbroken lines on the temperature difference plot indicate the positions of the 195 K contours from OH-92A and CHEM-92A. The contour intervals are 5 K, 0.5 p.p.b.v., and 2 p.p.t.v. for the temperature, ClO$_x$, and BrO$_x$ respectively, and 1 K, 0.1 p.p.b.v., and 0.25 p.p.t.v. for the differences. Negative contours are dotted; the zero contour is dashed. The difference fields for the ClO$_x$ and BrO$_x$ have been smoothed slightly. See text for further explanation.
(and had been so continuously from the start of OH-92A). The ClOₓ abundance is strongly anti-correlated with the temperature: the highest ClOₓ values coincide with the lowest temperatures, and the region of enhanced ClOₓ retreats towards the pole as the atmosphere warms. The fall in ClOₓ is a consequence of the dissipation of PSCs and the cessation of the heterogeneous reactions required to sustain the enhanced reactive chlorine. The BrOₓ abundance also falls as the temperature rises, but the contrast between enhanced and non-enhanced concentrations is smaller for bromine than for chlorine because the bromine reservoir compounds are more reactive in the gas phase. Figure 15 also shows the difference in the temperature, ClOₓ and BrOₓ fields from the two runs. In the lower stratosphere, the forcing temperatures for OH-92A, calculated with depleted ozone, are lower than those for CHEM-92A, and consequently the ClOₓ is more enhanced in OH-92A than in CHEM-92A. Nevertheless, during the early part of the period shown, when the ClOₓ abundances are greatest, the temperature difference is small, around 1–2 K, and the difference in the ClOₓ amount is also small, with no extension of the enhanced ClOₓ to lower latitudes. The biggest difference in the temperatures occurs after they have risen above the PSC threshold (∼195 K), and after the enhanced ClOₓ has dissipated. The unbroken lines on the temperature-difference plot are the 195 K contours from OH-92A and CHEM-92A; these show that the longevity of the sub-PSC temperatures is significantly extended by the ozone hole only at the highest latitudes, where the maximum extension is ∼6 days. It is at this location and time that the largest differences in ClOₓ occur as the concentration is diminishing back toward non-enhanced levels. The differences in BrOₓ are unlike those in the ClOₓ: at latitudes equatorwards of ∼80°S the BrOₓ abundance is always higher in OH-92A, but near the pole the sign of the difference is changeable. Bromine chemistry is characterized by rapid cycling between different species, and it appears the BrOₓ inequalities between OH-92A and CHEM-92A are due more to differences in the chemical environment than to differences in the temperature. Be that as it may, the BrOₓ differences are relatively small, and less important for the ozone chemistry than the ClOₓ differences. Plots of the ClOₓ difference at other pressures show a pattern similar to that at 56 hPa, and there is no indication of the chlorine activation extending to greater heights due to cooling from the ozone hole.

A time-latitude section of the zonal-mean ozone at 56 hPa from CHEM-92A is given in Fig. 16. Even in this run, which does not account for the radiative effects of the ozone hole, the ozone at the high latitudes is almost removed by mid-October when the largest differences in ClOₓ between CHEM-92A and OH-92A arise. Thus, in the altitude range of greatest ozone destruction, there is little scope for an additional destruction due to the higher ClOₓ resulting when the ozone-hole cooling is accounted for. Nevertheless, at other heights where the ozone is less depleted, the temperature and ClOₓ differences between the runs might have more significance for the ozone destruction. A measure of the net effect is given by the ozone column. Figure 17 shows the evolutions of the zonal-mean ozone columns from OH-92A and CHEM-92A, and the differences between them. The columns were calculated between 80 hPa and 1.3 hPa only, i.e. for the model levels where the chemistry was active. The polar ozone column is already much reduced by 13 September, when CHEM-92A begins, and it continues to diminish for a few days, but by late September there is a positive trend at the high latitudes. This increase in total ozone while ClOₓ concentrations are enhanced, though declining, may be partly a consequence of the model not including the ozone destruction in the lowermost stratosphere, thus increasing the relative contribution of ozone enrichment by transport from middle latitudes. Up until about mid-October, the differences between OH-92A and CHEM-92A are small, but thereafter larger differences arise with the polar ozone
column in OH-92A falling below that in CHEM-92A. The largest deficit of ~25 Dobson units (DU) occurs in late October when the columns are increasing rapidly. The overall pattern of the difference is not easily explainable in purely chemical terms, and suggests that transport is playing a role. The effects of chemistry and transport can, to an extent, be separated using passive tracers. CHEM-92A and (the CTM portion of) OH-92A included inert tracers which on day 30 of OH-92A and the initial day of CHEM-92A were given the same fields as the chemical ozone. The evolving difference in these (initially identical) tracer fields gives an indication of the contribution of transport to the evolving difference in the chemical ozone field. The final panel of Fig. 17 shows the column difference remaining after subtraction of the difference in the passive tracers. Most of the structure is removed from the plot by this subtraction, and the aforementioned large difference in the columns in late October is much reduced, indicating that it was largely due to transport. The differences that do remain are small and suggest little excess ozone destruction in OH-92A. There does remain a deficit of 5–10 DU in OH-92A close to the pole in mid-October, coincident with the largest ClO$_\times$ excess in that run. Figure 18 presents polar stereographic maps of the four fields on a day from that period. The ozone columns from the OH-92A and CHEM-92A are much alike, with the ‘chemical vortices’ marked by the depleted columns being of similar size and shape, and both showing a filament of low-ozone air emanating outwards to lower latitudes. This similarity of the chemical vortices is consistent with the similarity of the dynamic vortices as indicated by the PV contours: regions of high PV gradient are expected to act as barriers to transport and contain the chemically perturbed air. The largest differences in the ozone columns (~30 DU) are associated with the filament of air streaming from the vortex, and subtracting the transport differences reveals that these are entirely dynamic in origin. The ‘chemistry-only’ difference reveals no systematic change either in the depth of the ozone hole, or in the horizontal extent of the depleted region. Thus, while including the radiative effects of the ozone hole in OH-92A has had a substantial, mostly negative, impact on the polar ozone column, the difference is due almost entirely to transport rather than to chemical feedback. The largest column deficit in OH-92A, in late October, is probably a result of the stronger vortex in that run, inhibiting the replenishment of the polar ozone by transport from middle latitudes.

As explained above, the radiative cooling from the ozone hole does not significantly extend the horizontal or vertical range of the region with enhanced ClO$_\times$ and BrO$_\times$, but it does extend the time for which the ClO$_\times$ remains enhanced. Thus, the key factor
Figure 17. Zonal-mean, time versus latitude sections of the ozone columns from (a) OH-92A, (b) CHEM-92A, and (c) the difference OH-92A minus CHEM-92A. (d) The column difference due to chemistry alone, with the contribution from transport differences removed. The contour intervals are 20 DU for the columns, and 5 DU for the differences. The zero contour is dashed; negative contours are dotted. See text for further explanation.
determining the extent of chemical feedback is the amount of ozone loss that takes place in this extended period. There was little ozone loss during this time in OH-92A, because the ozone concentration was already low. It was pointed out earlier that the lower-stratospheric temperatures from OH-92A and CHEM-92A have a cold bias relative to those from the UKMO assimilations. If this caused the ozone abundance at the time of the transition from sub- to super-PSC temperatures to be unrealistically low, then the model might miss a chemical feedback that would otherwise be present. The important issue in this regard is not the absolute temperature difference between the model and the assimilations but the difference in the longevity of the sub-PSC temperatures. Figure 3 showed that in the assimilations these temperatures persisted, in the zonal mean, until 29 September. Figures 16 and 17 show that by then the ozone depletion in the model.
was near its peak, and did not markedly increase in the interval up to mid-October when the sub-PSC temperatures dissipated from the model. Thus, had the model temperatures been more accurate, the same argument for a lack of chemical feedback would still apply. Moreover, because the ozone-hole induced temperature difference increases with time, the earlier the PSCs dissipate, the shorter the ozone-hole induced extension of the enhanced period is likely to be. The severe depletion of lower-stratospheric ozone seen in the model is not atypical of other modelling and observational studies (WMO 1995).

Heterogeneous chemical reactions on types 1 and 2 PSCs are sufficiently fast, and the rate of airflow through the clouds sufficiently high, that in relatively coarse-resolution modelling of chemical processing it is not vital to reproduce the true PSC distribution with great fidelity. The distribution of the different cloud types may, however, be important for the chemical recovery of the vortex after the PSCs have dissipated. PSCs crossing the tropopause irreversibly remove their constituent H2O and HNO3 from the stratosphere. Type 2 PSCs sediment faster than type 1, so the greater the formation of type 2, the greater the dehydration and denitrification of the vortex. The extent of denitrification may affect how active chlorine is converted back to the reservoir species. Modelling (e.g. Prather and Jaffe 1990; Austin and Butchart 1992) and observational (Santee et al. 1996) studies suggest that in the denitrified southern vortex of 1992 the primary recovery product was HCl rather than the ClONO2 which is formed in the less denitrified northern vortex. If lower temperatures produced by feedback lead to an increased formation of type 2 PSCs then the resulting excess denitrification could affect the chemical recovery. Figure 19 shows time–pressure sections of the areas occupied by type 1 and type 2 PSCs in OH-92A and CHEM-92A, and the differences between the runs. The evolutions of the type 1 PSCs in the two runs are similar up until about the end of September when the PSCs at the upper levels dissipate abruptly. The dissipation occurs a few days earlier in CHEM-92A than in OH-92A. In October the differences between the runs are greater: the clouds extend to higher altitudes in OH-92A and, at the lowest model levels, persist for several days longer. Type 2 PSCs occupy a smaller area than the type 1 and are more confined to the bottom model levels. The differences in their evolutions in the two runs are more marked than for type 1s; type 2 PSCs are virtually gone from CHEM-92A by the end of September, whereas they occupy a significant area in OH-92A throughout September and into early October.

As a result of the prevalence of type 2 PSCs and their high sedimentation rate ($\sim 1.5$ km day$^{-1}$ in the model), the vortex in OH-92A is almost completely denitrified by early September. This early and severe denitrification precludes the possibility of seeing a difference between the HNO3 evolution in OH-92A and that in CHEM-92A, which was initialized with the HNO3 field (and other chemical fields) from OH-92A. The low HNO3 abundances present when the PSCs dissipate mean that in both runs almost all the reactive chlorine is initially converted back into HCl. A small difference in HNO3 of the order of a few tenths of a p.p.b.v. does arise in mid-October when the PSCs have almost gone from CHEM-92A. The difference in the H2O evolutions in the two runs is small also, with a maximum excess dehydration in OH-92A of $\sim 0.5$ p.p.m.v. roughly coincident with the maximum HNO3 difference. The minimum mixing ratios of H2O in the lower stratosphere of the vortex are in the range 1.25–2.5 p.p.m.v. which is similar to MLS observations at 465 K potential temperature during the 1992 southern winter (Santee et al. 1995). The greater denitrification and dehydration in the colder run partially counteract the tendency of the lower temperatures to favour the formation and retention of PSCs. The lower the abundances of HNO3 and H2O the greater is the sensitivity to their concentrations; according to Teten's equation, reducing H2O from 2.0 to 1.5 p.p.m.v. lowers the threshold temperature for type 2 PSCs by $\sim 1.5$ K.
Nevertheless, it is the temperature differences between the runs that largely control the differences in PSC distribution. As noted earlier, the modelled H₂O is not used in the heating calculations, so any radiative effects of H₂O differences cannot be quantified.

Although the prevalence of type 2 PSCs in both OH-92A and CHEM-92A denitrifies the polar vortex nearly completely, and so precludes the possibility of seeing a difference in the mode of chemical recovery, the significantly greater occurrence of the type 2 clouds in OH-92A does suggest that in a less cold vortex with an earlier warming the effect of radiative chemical feedback on the temperature may lead to an excess denitrification sufficient to affect the routes by which the reactive chlorine is returned to the reservoir species.

(f) Vertical transport

As discussed above, the model runs with depleted ozone have a stronger residual circulation than the equivalent undepleted runs. The intensity of the residual circulation
Figure 20. Zonal-mean, time versus latitude section of the inert $\theta$-like tracer from OH-92A (see text) at 450 K minus the equivalent field from CO-92A (see text). The contour intervals is 2 units ($\equiv$K).

affects the rate at which the ozone-depleted polar lower stratosphere is replenished by descent of relatively ozone-rich air. The rate of diabatic transport in the runs was examined using an inert tracer initialized in the USMM with the same field as the potential temperature. The change in the distribution of this $\theta$-tracers with respect to the $\theta$ surfaces gives a measure of the diabatic motion in the model. In all runs, the tracer indicates the expected diabatic descent at the high latitudes during winter. Figure 20 shows the difference in the evolution of the $\theta$-tracers from OH-92A and CO-92A on the 450 K surface. The difference remains very small until about day 30, then grows at an increasing rate over the next 40 or so days with the maximum difference of $\sim$25 units (equivalent to a difference in diabatic descent of 25 K) occurring at the pole. This enhanced descent onto the 450 K surface in the run with the ozone hole is consistent with the increased intensity of the residual circulation already discussed. It is also consistent with the greater radiative cooling and dynamic heating at 56 hPa in OH-92A than in CO-92A shown in Fig. 8.

Imposing the descents implied by the $\theta$-tracers onto the initial ozone fields in OH-92A and CO-92A gives ozone increases at 450 K of around 0.2–0.4 p.p.m.v. in the polar regions by day 60, with a maximum difference between runs of $\sim$0.1 p.p.m.v. Over the same period the reactive ozone in the CTM of OH-92A falls by $\sim$2 p.p.m.v. This implies that the rate at which the ozone is being dynamically replenished is small in relation to the rate at which ozone is being chemically destroyed. The rate of dynamic replenishment seen here is also small relative to the rate of chemical ozone loss inferred directly from observations of ClO within the Antarctic vortices of recent years (Murphy et al. 1991; MacKenzie et al. 1996).

Because the chemical ozone loss in OH-92A is large, this experiment probably shows nearly the maximum effect that feedback can have on the vertical transport in the model. Nevertheless, the calculated effect of the altered transport on the ozone distribution is small in relation to the chemical ozone changes within the vortex. Similarly, the effect on the ozone chemistry of the altered vertical transport of chlorine and other species is likely also to be small.

5. SUMMARY AND CONCLUSIONS

A southern hemisphere winter and spring was simulated in a mechanistic model of the middle atmosphere with an interactive chemistry scheme. Runs which included the
Radiative effects of ozone depletion were up to ~12 K cooler in the lower stratosphere than runs which did not. Above the ozone-depleted region, temperatures were slightly warmer than they would otherwise be; this is due to increased dynamic heating from a strengthened residual circulation and to an enhanced radiative heating rate. The evolution of the vertical profile of the temperature difference has a distinctive structure which is insensitive to small changes in the initial conditions and in the lower-boundary forcing.

These results confirm that, within the COSMIC model, Antarctic ozone depletion, due, at root, to the currently unnaturally high loadings of chlorine and bromine, can have a significant effect on the atmospheric temperature and dynamics via the radiation calculations. The largest temperature differences arise after the temperatures have risen above the PSC threshold, but differences of several kelvin are present while ClO\(_X\) and BrO\(_X\) remain enhanced. However, the ozone-hole induced cooling is largest close to the pole, and it does not significantly extend the enhanced reactive ClO\(_X\) to lower latitudes. Nor does it increase the vertical extent of the enhancement. Moreover, there is little increase in the magnitude of the ClO\(_X\) and BrO\(_X\) concentrations within the enhanced region. Thus, while sub-PSC temperatures remain, the ozone destruction rate is largely unaffected by the temperature deficit due to the ozone hole. The largest differences in ClO\(_X\) between runs with and without an interactive ozone hole arise in the interval between temperatures rising above the PSC threshold in the non-interactive and in the interactive runs. In the experiments performed, this interval maximized at ~6 days in the highest latitudes during mid-October. By this time, the polar lower stratosphere was almost devoid of ozone so there was no additional loss due to the excess ClO\(_X\).

Dynamically, the radiative effects of the ozone hole have little impact on the morphology of the polar vortex but, in the lower stratosphere, the PV magnitudes inside the vortex are increased, while in the middle stratosphere the differences are more variable and transient. The decay of planetary waves from the stratosphere is delayed by the ozone hole and, following the period of strongest decay, the residual circulation is markedly stronger in the ozone-hole run. However, the increase in the replenishment of lower-stratospheric ozone through the enhanced diabatic descent is slight. The feedback from the radiative impact of the ozone hole onto the ozone column acts almost exclusively through changes in horizontal transport arising from the strengthening of the polar vortex.

The experiments described here for 1992 have subsequently been repeated using lower-boundary forcings from 1993 and 1994; the main findings discussed above were found to hold for these years also. Overall, the results suggest that, while the radiative impact of the ozone hole can affect polar ozone via transport changes, under current atmospheric conditions there is no feedback onto the ozone chemistry. This is because the induced cooling does not change the spatial extent of the chlorine and bromine activation, but only its duration. With present-day chlorine and bromine loadings, the chemically perturbed region of the Antarctic is effectively drained of ozone even without accounting for the radiatively induced cooling. Nevertheless, as chlorine and bromine levels fall during the next century, and ozone destruction becomes less complete, prolonged enhancement of ClO\(_X\) and BrO\(_X\) due to radiative cooling may become more important.

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REFERENCES


MacKenzie, I. A., Harwood, R. S., Froidevaux, L. W., Reid, W. G. and Waters, J. W.


Mahlman, J. D., Pinto, J. P. and Umscheid, L. J.


Mote, P. W., Stott, P. A. and Harwood, R. S.


Murphy, D. M.


Murray, F. W.


Prather, M. J.


Prather, M. and Jaffé, A. H.


Ramaroson, R. A., Pirre, M. and Cariolle, D.


Randel, W. J. and Wu, F.

1999 Cooling of the Arctic and Antarctic polar stratospheres due to ozone depletion. J. Climate, in press


1996 Chlorine deactivation in the lower stratospheric polar regions during late winter: Results from UARS. J. Geophys. Res., 101, 18835–18859


1998a Increased polar stratospheric ozone losses and delayed eventual recovery owing to increasing greenhouse gas concentrations. Nature, 392, 589–592

Shindell, D. T., Wong, S. and Rind, D.

1998b Climate changes and the middle atmosphere. Part IV: Ozone response to doubled CO2. J. Climate, 11, 893–918

Shindell, D. T., Rind, D. and Lonergan, P.


Shine, K. P.


Solomon, S., Garcia, R. R., Rowland, F. S. and Wuebbles, D. J.


Stolarski, R. S., Labow, G. J. and McPeters, R. D.


Swinbank, R. and O’Neill, A.


Terral, L. and Thual, O.

1995 OASIS: le couplage océan–atmosphère. La Météorologie, 10, 50–61

Tibaldi, S., Palmer, T. N., Brankovic, C. and Cubasch, U.


WMO