The semi-annual oscillation and equatorial tracer distributions

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SUMMARY

Some interesting low latitude features of tracer distributions revealed by recent satellite experiments have been studied using a two-dimensional atmospheric model. It is found that the equatorial semi-annual oscillation, forced in the model by simply prescribing a suitable momentum convergence, is important in determining the tracer behaviour. In particular, the meridional circulation associated with the semi-annual oscillation leads to a successful model simulation of the ‘double peak’ in N$_2$O and CH$_4$ mixing ratios, observations of which show, at certain times of the year, maxima on a fixed pressure surface in the subtropics with a local minimum at the equator.

1. INTRODUCTION

Perhaps the most surprising feature of the composition fields revealed by the satellite measurements on the Nimbus 7 satellite is the ‘double peak’ found for certain constituents. For example, for CH$_4$ and N$_2$O, maximum mixing ratios on a constant pressure surface can be found in subtropical latitudes with a local minimum at the equator. Jones and Pyle (1984) called this the double peak. It is most pronounced in the spring of the northern hemisphere with a much weaker feature six months later. At other times there is just one mixing ratio maximum in low latitudes.

Jones and Pyle (1984) presented CH$_4$ and N$_2$O data from just one year. Three years of data are now available (Jones 1984). Figure 1, taken from Jones (1984), shows the monthly averaged N$_2$O cross-sections (p.p.b.v.: parts per 10$^5$ by volume) from 1979 to 1981. The principal feature is the decrease of mixing ratio with height and latitude. At a fixed height maxima are found in southern subtropics in January and northern subtropics in July. A double peak is prominent in the observations in March, April and May and is clearly a feature of climatology. A much weaker double peak structure appears to be present in the November climatology. H$_2$O has a similar double peak feature (Russell et al. 1984) with a double minimum in the lower stratosphere but a double maximum around the stratopause. The satellite measurements of O$_3$ (Russell 1984) and nighttime NO$_2$ also show structure in equatorial latitudes with a local minimum over the equator at about 35 km.

Jones and Pyle studied the CH$_4$ and N$_2$O data from Nimbus 7 using a two-dimensional model. Although the gross features of the observations were modelled satisfactorily there were important differences of detail. In particular, the double peak was not reproduced. Secondly, at any given height, maximum model mixing ratios were always found at the equator, whereas the maximum observed values are found at latitudes ranging between about ±30°. Finally, although the vertical decrease of CH$_4$ was modelled well, the model N$_2$O decreased too little with altitude. All of these features could be associated with limitations in the modelled dynamics.

It is interesting to speculate in which manner the two-dimensional study is inadequate. One possibility is that the radiation schemes are incomplete. The lower

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stratosphere presents particular problems for modelling radiation transfer with a number of small compensating terms being important (Houghton 1978). Haigh (1984) has recently included a detailed radiation treatment in the model. She does not find double peaks (Haigh, private communication).

There is a rich spectrum of equatorial waves which were not included in the study by Jones and Pyle. A particularly interesting phenomenon in equatorial latitudes is the semi-annual oscillation of the zonal wind (Reed 1966). This oscillation has a period of six months with a peak amplitude at the stratopause level of about 30 m s⁻¹ (Belmont et al. 1974). Figure 2 shows the oscillation as presented by Reed. Hopkins (1975) suggests that the easterly phase is due to equatorially propagating planetary waves from the winter hemisphere. Other possibilities, e.g. advection of zonal easterly momentum, exist. The westerly acceleration phase is generally believed to be due to equatorial Kelvin waves (see Holton 1975).

If we take the hypothesis that the easterly phase of the semi-annual oscillation is due to planetary waves we should expect variations in the oscillation. In particular a stronger easterly phase should be expected to be associated with the northern hemisphere planetary waves which are generally larger than their southern hemisphere counterparts. This asymmetry provides a possible clue to the production of the double peaks. We speculate, first, that the semi-annual oscillation of the zonal wind and the associated oscillation in the meridional winds are responsible for the double peak. Figure 3, based loosely on Meyer (1970), shows the expected induced meridional circulation during the westerly acceleration phase of the oscillation. Taking a constituent with a mixing ratio which decreases with height, we expect the indicated perturbation to the mixing ratio field. Secondly, we speculate that the asymmetry in the double peak, which is pronounced in March, April and May but is quite weak six months later, is related to an asymmetry in the semi-annual oscillation due possibly to different levels of planetary wave activity in the two hemispheres.
We have tested some of these ideas in the two-dimensional model used by Jones and Pyle (1984). The basic model is described in the next section. The semi-annual oscillation has been forced by specifying a time-varying momentum source in equatorial latitudes. This is described in section 3. Finally, we present and discuss results for the wind fields and the distribution of various trace gases.
2. THE MODEL

The model used in this study is the two-dimensional mean circulation model described by Harwood and Pyle (1975). The model calculates the zonal-mean values of temperature, wind components, and chemical constituent mixing ratios with a resolution of $\pi/19$ in latitude, 0.5 in $\ln(p_o/p)$ (approximately 3-5 km) in the vertical (where $p$ is pressure and $p_o$ is the pressure at the surface) and with a 6-hour time step. A second-order partial differential equation is solved for the meridional streamfunction given the forcing by radiative and other diabatic heating and eddy heat and momentum fluxes. The dynamical and radiative formulation employed is that described by Haigh and Pyle (1982).

An important addition to the model is the inclusion of Rayleigh frictional damping of the zonal flow in the mesosphere. It has been found (Leovy 1964; Schoeberl and Strobel 1978; Holton and Wehrbein 1980; and Crane et al. 1980) that without such a damping there are insufficient sources and sinks of mean zonal momentum to close the polar night westerly jet and the summer easterly jet in the mesosphere. It is now generally accepted that Rayleigh friction may be rationalized as a qualitative parametrization of the wave drag due to the breaking of vertically propagating internal gravity waves (Lindzen 1981; Dunkerton 1982; Fritts 1984). The Rayleigh friction parametrization in this study is identical to that of Crane et al. (1980) who used the same model. The friction coefficient is zero below 50 km and increases to a maximum value of 1 d$^{-1}$ at $\sim$85 km (see their Fig. 15). Although this results in some changes in the tracer fields compared with Jones and Pyle, the $N_2O$ distribution of the model in the stratosphere is still dominated by a strong maximum in mixing ratio over the equator at both solstice and equinox.

A full photochemical scheme is included in the model. The scheme is essentially that used by Jones and Pyle (see also Haigh and Pyle 1982) with further updating of the kinetics data (DeMore et al. 1983; Baulch et al. 1982). $N_2O_5$ is now included in the photochemical scheme. For the purposes of the experiments described here, which are mainly of dynamical interest, the main limitation of the model is that stratospheric $H_2O$ is specified at 4 p.p.m.v. rather than calculated.

3. FORCING OF THE SEMI-ANNUAL OSCILLATION (SAO)

Holton and Wehrbein (1980) have discussed the semi-annual oscillation (SAO) found in their two-dimensional model of the atmosphere. Their model does not produce westerlies at the equator but the well-produced easterly phase of the oscillation is due to nonlinear advection of easterly momentum. Takahashi (1984) has included a simple treatment of Kelvin waves in his model which depends on constant, specified phase velocity and the zonal wind profile. In this way he has improved the simulation of the westerly phase of the oscillation. He concluded that the easterly forcing in his model, due to advection alone, was insufficiently strong and suggested that planetary waves may play a role.

Rather than follow Takahashi's approach, we have chosen simply to specify a westerly momentum forcing by inclusion of an additional term in the zonal component of the momentum equations. Following both Holton and Wehrbein (1980) and Takahashi (1984) we expect to produce an easterly phase of the oscillation by the nonlinear advection of easterly momentum from the summer hemisphere. Notice that the waves presumed responsible for the momentum forcing are assumed to play no further role in the model in producing diffusive mixing, in producing chemical eddy transports or in transporting heat. Unlike gravity waves which deposit momentum by the action of wave-breaking and hence cause a considerable amount of mixing, the main process by which Kelvin waves
transfer momentum to the zonal flow is believed to be by absorption and damping, processes which do not necessarily imply a large degree of mixing. It is possible that Kelvin waves may break in the lower stratosphere but we have not included extra mixing terms to account for this. Additionally we do not know, a priori, any reason why there should be a large eddy tracer transport associated with the Kelvin waves. Equally we cannot rule this out.

The additional westerly momentum inserted in the model in order to simulate the SAO is based on the zonal wind observations shown in Fig. 2. For example, at 44 km the equatorial zonal wind varies from about 30 m s\(^{-1}\) easterly to about 20 m s\(^{-1}\) westerly in three months. We have chosen a forcing which varied in time like \(\sin(2\pi D/365)\) where \(D\), the day number, was specified so that the maximum forcing of 35 m s\(^{-1}\)month\(^{-1}\) occurred at the end of February and at the end of August. The positive half of the cycle was used to force the westerlies; during the negative half of the cycle the forcing was set to zero. Since the largest amplitudes of the zonal wind oscillation are found in the tropics the forcing was confined to low latitudes, decreasing as \(\cos^2(\text{latitude})\) away from the equator. The precise details of the induced circulation depend critically on the vertical gradient of the momentum deposition. We have chosen to use the forcing distribution shown in Fig. 4 in which the strongest gradients (and hence induced circulation) occur at about 40 km. This distribution was arrived at, after some experiment, on the basis of the zonal wind structure it produced. It is, in fact, very similar to the vertical profile used by Dunkerton (1982) in his simulation of the SAO.

It should be emphasized that our approach is motivated by the desire to understand unexpected features of tracer distribution revealed by recent satellite experiments. We are not attempting to study the mechanism of the SAO but rather to look at the influence of the oscillation on stratospheric tracers. We have therefore striven for simplicity rather than strict accuracy in the modelling. Despite the limitations of our dynamical simulation we believe that the essential features of the relationship between the SAO and the distribution of stratospheric tracers should be revealed. There are several areas in which our modelling of the SAO could be improved. Perhaps the most serious deficiency is that the forcing is independent of the background zonal wind field. A simple method similar to that used by Takahashi (1984), with the momentum forcing dependent on the difference between mean wind speed and wave velocity \((\bar{U} - c)\), would be an improvement. Secondly, the fact that we have used a Rayleigh friction parametrization to model momentum deposition in the mesosphere has had an adverse effect on the modelled

![Figure 4. Height distribution of maximum acceleration (m s\(^{-1}\) month\(^{-1}\)) employed in run B in order to simulate the semi-annual oscillation.](image-url)
SAO, as it opposed the forcing of a westerly wind above 50 km. Dunkerton (1979) has shown that the Rayleigh friction approach is least satisfactory in the tropics and around the equinoxes. Thirdly, in order to keep the modelled SAO simple we have made no attempt to include the detailed time-dependence of the oscillation. The forcing peaks at all altitudes at the same time and the same vertical profile is used throughout. The observed SAO, on the other hand, has a structure in which the peak westerlies occur first at the highest altitudes and then descend during the subsequent months. Observations indicate, then, that the level of maximum vertical gradient of the forcing (and hence the level of maximum induced meridional circulation) slowly descends during the westerly phase.

In a further study, to be reported in a later paper, we have incorporated some of the improvements suggested above including forcing that is dependent on the background zonal wind field. This gives the correct time dependence of the SAO in the model. In accordance with our expectations, these improved dynamical treatments lead to only small changes to the behaviour of the tracer fields, which are discussed in the next section.

4. RESULTS

We have performed two basic model runs. Run A is the control run while run B includes the momentum forcing discussed above. Before describing the semi-annual oscillation experiment (run B) we point out those features of run A which differ from the previous experiments described by Jones and Pyle.

The inclusion of the Rayleigh frictional term has led to an improved zonal wind field and to a stronger, more realistic meridional circulation. It also led to an improvement in the modelled N₂O and CH₄ distribution above 50 km. Figures 5(a) and (b) show the latitude–height cross-sections of N₂O mixing ratio for April and July respectively. Note that the vertical scale extends from approximately 25 km to 60 km. This range includes the features in which we are interested, being the region for which data are available. In July, above 50 km, there is an asymmetry about the equator, with the maximum centred at approximately 18°N. In January (not shown) the maximum is situated at 18°S. This compares favourably with the SAMS data (see Fig. 1) and is a marked improvement on the corresponding plots shown in Jones and Pyle (1984) which had a maximum at the equator at all times of the year. Similarly, in April the distribution above 50 km is more realistic, showing signs of a weak double peak. However, not surprisingly, the inclusion of Rayleigh frictional damping has had little effect below 50 km and no double peak is evident there. The double maximum seen at about 55 km in Fig. 5(a) is related to the easterly winds at the equator associated with the re-establishment of the summer easterly jet; the Rayleigh frictional forcing is therefore in a westerly direction and induces a meridional circulation in the same sense as shown in Fig. 3 but at a somewhat higher level.

The observed changes in N₂O distribution as a result of including Rayleigh friction strongly support the need for a momentum forcing of a similar sign extending further down into the stratosphere, in order to reproduce the observed double peak. Note that it is difficult to justify the deposition of large quantities of momentum due to the breaking of gravity waves below about 50 km as it is doubtful that they could achieve the necessary amplitude below this height.

(a) Zonal winds

The evolution of the zonal winds at the equator in run B, which included the momentum forcing, is shown in Fig. 6. A semi-annual oscillation is evident. In the upper
Figure 5. Monthly-averaged latitude–height cross-section of N₂O volume mixing ratio (p.p.b.v.) from model run A for (a) April and (b) July.
stratosphere and lower mesosphere the wind changes from easterly to westerly with a peak to peak variation of about 50 m s$^{-1}$, similar to that observed. In accordance with our prescribed forcing the zero-wind line does not display a gradual descent with time; the flow changes from easterly to westerly almost simultaneously at all heights in late February and then again six months later. In the lower stratosphere the modelled winds are always easterly.

The zero-wind line does not extend further down than 40 km, a result of fixing the maximum vertical forcing gradient at the same level throughout the oscillation. Below 40 km during the equinoxes the easterlies are much stronger than observed, indicating insufficient westerly forcing at those levels. However, model runs in which the forcing was increased at lower levels in order to improve this feature displayed much reduced easterlies during the solstices. This strongly suggests the need for some additional easterly forcing (possibly planetary wave momentum deposition) if realistic easterlies are to be produced.

At approximately 50 km Fig. 6 shows good agreement with the observed zonal wind oscillation. Above this level, the modelled wind field decreases whereas the observations indicate a fairly constant maximum value of 30 m s$^{-1}$ at all heights above 55 km. This deficiency can be attributed to our use of Rayleigh friction, which strongly opposes the westerly forcing above ~50 km, as we have already mentioned. Thus, the zonal and meridional flows above about 55 km depend heavily on the frictional parametrization, which is known to be particularly poor in equatorial latitudes (Dunkerton 1979).

![Time-height cross-section of the evolution of the zonal winds at the equator in run B. Solid contours indicate westerlies.](image)

The equation for the model streamfunction is a linear equation forced by a sum of individual terms representing diabatic heating, eddy heat and momentum forcing, Rayleigh friction, etc. (see Harwood and Pyle 1977, 1980). Therefore, to find the circulation driven by the SAO forcing one needs only to set all the other forcing terms to zero.
Figure 7. Latitude–height cross-section of the induced vertical velocity (m ms⁻¹) due to the prescribed semi-annual oscillation forcing in run B. Solid contours indicate regions of rising motion, dashed contours indicate regions of descending motion. These values are representative of the westerly forcing phase.

Figure 7 shows the induced vertical velocity due to the prescribed SAO momentum forcing in March. For presentation purposes some small amount of smoothing has been performed. Smoothing occurs similarly in the model where fluxes are introduced to maintain static and inertial stability when required (see Harwood and Pyle 1975). The pattern in Fig. 7 is simple with a region of downward motion between ±25° latitude and rising motion, strongest at about ±55° latitude, elsewhere. The most vigorous circulation is found at the height of the sharpest vertical gradient in the forcing. The net vertical velocity field for March (not shown) has a small area of subsidence over the equator at 40 km, in excellent agreement with the diabatic circulation deduced by Solomon et al. (1986) from March LIMS temperature and ozone data.

The vertical forcing profile shown in Fig. 4 is constant above about 40 km, in an effort to maintain the zonal winds which are continually eroded by the Rayleigh friction term at upper levels. With a more realistic parametrization of gravity waves this would not be necessary. If the forcing had been made to decrease above 40 km, then a circulation similar to that in Fig. 3 would have been forced with rising motion in equatorial latitudes of the mesosphere and sinking in the extratropics.

The momentum forcing has been chosen to reproduce the observed semi-annual oscillation of the zonal winds so that the general agreement in Fig. 6 is to be expected. The important question is whether the induced circulations can subsequently explain the main features of the tracer observations and, if so, whether the limitations of the forcing, revealed by comparison of Figs. 2 and 6, might also be responsible for the detailed differences between model and observations.

(b) Tracer distribution

The discussion of the modelled tracer behaviour will concentrate first on N₂O which, because of its rapid decrease with altitude, is particularly sensitive to the representation
Figure 8. Monthly-averaged latitude–height cross-section of N₂O volume mixing ratio (p.p.b.v.) for April from run B.

Figure 9. Monthly-averaged latitude–height cross-section of N₂O volume mixing ratio (p.p.b.v.) for April 1979 from the SAMS satellite measurements.
of dynamics. Figure 8 shows the monthly average cross-section from model run B for April of the second year of integration. For comparison Fig. 9 shows the monthly mean SAMS data on the same scale. The run B N₂O distribution shows significant differences compared with the control run, run A (see Fig. 5(a)). In particular, the N₂O decreases with height in the tropics more rapidly in run B. Between 30 and 40 km the contours are flat due to the downward transport induced by the momentum forcing. In this region there is better agreement with the observations (Fig. 9) and at ~40 km, the level of maximum forcing, there is a feature suggestive of the double peak.

At higher latitudes, the additional upward contribution to the meridional circulation produces higher mixing ratios in run B than in run A. The observations generally indicate more latitudinal variation in higher latitudes than found in the model. In a recent comparison of classical Eulerian models and transformed circulation two-dimensional models (WMO 1986) it was found that the Eulerian models like this one predicted less latitudinal variation. This is possibly due to a rather high diffusive mixing in high latitudes which smoothes out strong gradients. Despite the differences in high latitudes, run B represents a significant improvement in modelling the observed distributions.

Figure 10 shows the model cross-section of N₂O from run B for July, to be compared with the observations (Fig. 11) and run A (Fig. 5(b)). The main improvement in run B is a more sharply defined maximum which is significantly off the equator in good agreement with the observations. This improvement occurs despite the fact that no forcing of the zonal flow occurs during the three months around the solstice. The changes are a reflection of the long response time of the tracer distribution to changes in the transport. Figure 10 shows the July cross-section for year 1. The corresponding cross-section for year 2 is rather more complicated as it still retains the presence of a double peak. This is perhaps an indication that the easterly phase of the oscillation in our model, which should remove the double peak, is not adequate in the long term. An examination of the zonal wind evolution in Fig. 6, which shows a gradual trend of increasing westerlies and decreasing easterlies, supports this observation.

The momentum forcing is the same for the vernal and autumnal equinoxes and run B shows no seasonal asymmetry; the double peak in October is similar to that in April.

Figure 12 shows the vertical distribution of N₂O at 9°N for April and July from run A, run B and the SAMS observations. 9°N was chosen because at this latitude the observations show a large temporal variation between April and July. For the control run (run A) little temporal variation is seen in mid-stratosphere (and this is true throughout the whole year) consistent with a maximum which remains very close to the equator. The values agree reasonably well with the SAMS measurements for July 1979 but, particularly at the higher altitudes, overestimate the spring values. For run B the agreement between model and observations is excellent. Run B shows a seasonal variation in agreement with the observations. For example, the April values at 40 km are a factor of two lower than the July values in both run B and in the SAMS data. The discrepancy between model and data in the lower stratosphere was noted by Jones and Pyle (1984). Compared with the in situ measurements, the SAMS data appear to be biased to high values below 10 mb.

Figure 13(a) and (b) show cross-sections of CH₄ from run A and run B of the second year of integration for April. The changes associated with the momentum forcing are similar to the case of N₂O with the additional induced downward flow causing a flattening of the contours in equatorial latitudes and the production of a weak double peak feature at around 40 km. Figure 14 shows profiles of CH₄ at 9°N from runs A and B compared with the SAMS observations. In this case run B underestimates the observed upper stratospheric CH₄. We do not believe that this is necessarily due to the model photo-
Figure 10. As Fig. 8 but for July.

Figure 11. As Fig. 9 but for July.
chemistry (OH and O(^1D)) are in accord with the observations and other models respectively while the chlorine content of the model is somewhat low) and reiterate the conclusion of Jones and Pyle (1984) that any discrepancy in the vertical profile of CH_4 or N_2O could well have a dynamical origin. In particular, the absence of easterly forcing could be responsible for the modelled values falling below the observations in the upper stratosphere.

In a third model integration, run C, the amplitude of the momentum forcing was approximately doubled so that the maximum forcing was 70 m s^{-1} month^{-1}; in all other respects the formulation of run C was identical to run B. The westerly phase of the SAO of this run was disproportionately stronger than the easterly phase reaching a maximum westerly wind of 90 m s^{-1}. The integration was continued for six months only; Fig. 15 shows the monthly averaged cross-section of N_2O for April. A prominent double peak is evident with maxima situated at ±20°. A comparison with the SAMS data for April in Fig. 9 reveals extremely good agreement in the spatial distribution and vertical gradients of N_2O. Due to the increased descent at the equator, however, the model underestimates the mixing ratio values. This experiment indicates that a double peak structure which shows good qualitative agreement with the SAMS data is possible provided the westerly momentum forcing is sufficiently strong.

An alternative possibility for our failure to reproduce the full vertical extent of the observed double peak is suggested by the work of Solomon et al. (1986). Having calculated a diabatic circulation from LIMS ozone and temperature data, they are able to reproduce the double peaks in CH_4 and N_2O using this as the transport circulation. Their best vertical structure is obtained using a very low value for the horizontal eddy coefficient, K_{yy}. It could be that at levels below the forcing level in our model, eddy transport is compensating too strongly for the mean transport induced by the SAO (as is often the case in Eulerian models like ours).

Generally speaking, the changes in the other species between runs A and B are less than for N_2O and CH_4, but qualitatively in the same sense. Thus, a low latitude seasonal
Figure 13. Monthly-averaged latitude–height cross-section of CH₄ volume mixing ratio (units are 10⁻² p.p.m.v.) for April from (a) run A and (b) run B.
Figure 14. Vertical profiles of CH₄ at 9°N for April and July from runs A and B compared with the measurements from SAMS.

Figure 15. Monthly-averaged latitude–height cross-section of N₂O volume mixing ratio (p.p.b.v.) for April from run C.
Figure 16. Monthly-averaged latitude–height cross-section of O₃ volume mixing ratio (p.p.m.v.) for April from run A.

Figure 17. As Fig. 16 but from the LIMS satellite measurements.
variation is modelled. Species whose mixing ratios decrease with height show a larger decrease at the equator in run B.

The results for ozone are particularly interesting. Figure 16 shows a model cross-section for April from run A, which may be compared with the LIMS observations (Fig. 17). The model overestimates the peak mixing ratios (possibly due to the low chlorine mixing ratios in the model) and fails to reproduce the observed low latitude structure seen between about 10 and 3 mb. A weak double peak structure is evident in run B (not shown). This feature is more pronounced in run C (Fig. 18). The qualitative behaviour of runs B and C is in excellent agreement with the observations.

Both dynamical and photochemical processes are involved in producing the ozone double peak. This is seen best by comparing the ozone fields with those for NO$_2$ (= NO + NO$_3$). Figure 19 shows LIMS nighttime NO$_2$ (equivalent for our purpose to NO$_3$) for April 1979. A double peak is clearly evident (Chi-rong Sun, personal communication). Notice also that the mixing ratios at the maximum are displaced into the southern hemisphere. The model results for runs A and C are shown in Figs. 20 and 21 respectively. The differences in magnitude between model and observations are not important. (For a discussion of the problem of modelling NO$_3$ see WMO (1986).) With the inclusion of the SAO (run C) the model maximum at about 40 km was displaced into the southern hemisphere and equatorial values were reduced. The model response is in fair agreement with the observations although the pronounced double peak extending up to the stratosphere is not reproduced. Comparing the ozone and NO$_2$ behaviour (Figs. 18 and 21), both fields show a double peak with equatorial minima at about 40 km. The gross behaviour shows that the fields are correlated and thus those features are of dynamical origin. In the mid-stratosphere, we notice that the maximum NO$_2$ mixing ratio is in the southern hemisphere but the ozone maximum occurs just north of the equator. This anticorrelation suggests a photochemical origin for this aspect of the modelled behaviour.
Figure 19. Monthly-averaged latitude–height cross-section of nighttime NO$_2$ volume mixing ratio (p.p.b.v.) for April 1979 from the LIMS satellite measurements.

Although we have not yet analysed individual terms in the continuity equations it appears that both equatorial dynamical processes and photochemistry are playing a role in producing the observed ozone behaviour. It will be interesting to calculate H$_2$O in a future modelling study, since the observed double peak in H$_2$O can be expected to lead to similar structures in OH and HO$_2$ which should further influence the ozone distribution.

5. CONCLUSIONS

The comparison of two model integrations, one with a representation of the semi-annual oscillation (run B) and one without (run A) has shown that the induced circulation due to the momentum forcing of the SAO is significant in the distribution of tracer species, especially in equatorial regions. In particular: (1) a double peak in N$_2$O and CH$_4$ has been reproduced which is qualitatively similar to that observed in the SAMS satellite dataset during March–May; (2) the single mixing ratio maximum which is present during the solstices moves away from the equator to 20°S in December to February and 20°N in June to September in a similar manner to that observed—in model runs without the SAO the maximum is always situated close to the equator; (3) the model displays significant seasonal variations in low latitudes, both qualitatively, with the change from a double peak to a single peak and back again, and quantitatively, with lower mixing ratios at the equator during the equinoxes than during the solstices—without the SAO the model displays little seasonal variation in N$_2$O distribution in the mid-stratosphere; (4) for ozone, the structure seen in the April LIMS observations is also modelled reasonably faithfully in runs including the SAO.

Although H$_2$O was not calculated in the photochemical model the observations are consistent with our ideas. If a circulation of the type shown in Fig. 3 is induced during
Figure 20. Monthly-averaged latitude–height cross-section of NO$_x$ volume mixing ratio (p.p.b.v.) for April from run A.

Figure 21. As Fig. 20 but from run C.
the westerly forcing phase then, since the H$_2$O mixing ratio increases with height in the stratosphere, a double minimum should be expected in low latitudes, as observed. Observations also show a double maximum at the stratopause. This perhaps confirms that the level of maximum momentum forcing is below the stratopause, so that the induced circulation above would be in the correct sense to give rising motion at the equator and sinking in the subtropics (see Fig. 3).

The model runs relied upon the nonlinear advection of easterly momentum from the summer hemisphere to produce the easterly phase of the SAO. Easterly forcing due to laterally propagating planetary waves from the winter hemisphere was not included in the model formulation. There is evidence from several sources that the easterly phase produced by the model was not adequate: firstly, the time-series of the zonal wind at the equator shown in Fig. 6 displays a slow trend towards stronger westerlies and weaker easterlies. The double peak is still evident in the N$_2$O cross-section for July of the second year by which time the easterly phase of the SAO should have established a single off-equatorial maximum. Secondly, when the westerly forcing was increased in the lower stratosphere in order to improve the zonal winds at that height the easterly phase became extremely weak; and thirdly, the model significantly underestimated the equatorial mixing ratios of CH$_4$. Easterly forcing would induce a rising motion in the equatorial stratosphere which could possibly improve the vertical gradients of CH$_4$, although a satisfactory simulation of both CH$_4$ and N$_2$O would probably not be achieved. The addition of an easterly momentum forcing in order to improve the easterly phase of the SAO would permit an increased level of westerly forcing while still retaining a realistic zonal wind field at the equator. The momentum forcing due to the absorption of Kelvin waves depends upon ($\bar{u} - c$) and is therefore sensitive to the background zonal wind field. This may possibly explain the asymmetry observed in the SAMS data which has a strong double peak in March–May but only a weak feature six months later. A strong easterly phase is expected in December to February as a result of the northern hemisphere planetary waves which are generally larger than those of the southern hemisphere. Hence a stronger westerly momentum forcing may follow as a result of the stronger easterly winds.

As mentioned above we have recently made a number of improvements to the dynamical model presented here and these will be discussed in a further paper.

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