Two-dimensional model studies of equatorial dynamics and tracer distributions

By L. J. GRAY

SERC, Rutherford Appleton Laboratory, Chilton, Oxon

and

J. A. PYLE

Department of Physical Chemistry, University of Cambridge

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SUMMARY

A parametrization of Kelvin wave absorption in the equatorial stratosphere has been included in a two-dimensional model of the atmosphere in order to improve the simulation of the semi-annual oscillation. Additionally, a more detailed treatment of the momentum deposition associated with gravity wave breaking in the mesosphere has replaced the previous Rayleigh friction scheme. The model exhibits a semi-annual reversal of the zonal winds at the equator in the stratosphere, with the westerly phase descending with time, as observed. The inclusion of the improved gravity wave parametrization has led to a number of significant improvements; in particular, the model also exhibits a semi-annual oscillation in the mesosphere which is out of phase with that in the stratosphere. Meridional circulations associated with the westerly phase of the semi-annual oscillation produce ‘double peak’ structures in a number of modelled tracer fields, including N₂O, CH₄ and H₂O which compare well with observations by the SAMS and LIMS satellite experiments. These structures consist of a distinct minimum at the equator and double maxima in the subtropics; they are present for several months around each equinox. Inclusion of the more realistic westerly forcing has pointed to the need for additional easterly forcing in the model around each solstice. Further supporting evidence for the importance of momentum deposition associated with planetary waves in the easterly phase of the semi-annual oscillation comes from the difference in double peaks between the northern hemisphere spring and autumn, as suggested in an earlier paper.

1. INTRODUCTION

It has recently been shown that the semi-annual oscillation can play an important role in the distribution of long-lived tracers in the stratosphere, particularly over the equator and during the equinoxes. In a preliminary study, Gray and Pyle (1986), hereafter referred to as GP, forced a semi-annual oscillation (SAO) in the zonal wind field of their two-dimensional model by the inclusion of an additional empirical driving term in the momentum equation. During the westerly phases of the oscillation, which occur around the equinoxes, a meridional circulation was induced in the model with downward motion over the equator and upward motion in mid-latitudes. This had a profound effect on the modelled distribution of long-lived tracers; for example, in April and in October the latitude-height cross-section of N₂O exhibited a distinct minimum over the equator and double maxima in the subtropics. This was similar to the structure, known as the double peak, which was unexpectedly observed in N₂O and CH₄ data from the SAMS experiment on the NIMBUS 7 satellite (Jones and Pyle 1984; Jones 1984). Associated structures are seen in the LIMS measurements of ozone (Russell 1984), H₂O (Remsberg et al. 1984) and nighttime NO₂ (Gray and Pyle 1986) and these were also reproduced by the model.

In the initial study of GP the westerly phase of the semi-annual oscillation was forced with an acceleration of the zonal wind whose time dependence was specified using the positive half of a sine wave. This was adjusted so that the westerlies reached their maximum amplitude at each equinox. Suitable latitude and height distributions were chosen for the forcing and these were kept constant with time. During the negative half of the sine wave the forcing was set to zero so that the easterly phase of the SAO (around the solstices) was produced by advection of easterlies from the summer hemisphere, that is, by the term \( v(f - \bar{u}_z) \), where the symbols have their usual meaning. In this way the
model was forced to produce an SAO whose gross features compared sufficiently well with observations to confirm its possible importance to tracer distributions in the model. The simplicity of the approach, however, prevented the modelling of several details of the evolution of the SAO. For example, the westerly phase of the SAO is thought to be due to the deposition of westerly momentum associated with Kelvin wave absorption, which is dependent upon the Doppler-shifted phase velocity. The main deficiency of the scheme employed by GP was that the SAO forcing was independent of the background wind field and as a result the modelled zonal wind field at the equator did not exhibit the observed gradual descent of the westerlies with time. Secondly, a Rayleigh friction parametrization of the deposition of momentum in the mesosphere opposed the development of winds above about 50 km in the model.

In this paper we describe substantial improvements to the modelled SAO. A more physically based parametrization of Kelvin wave momentum deposition in the stratosphere has been introduced which depends upon the background wind field and characteristics of the wave itself (Dunkerton 1979). Additionally, the Rayleigh friction has been replaced by an improved representation of gravity wave momentum deposition in the mesosphere (Lindzen 1981; Holton 1982). The details of these two schemes are described in the next section. Results from the new version of the model are presented in section 3 and a comparison is made with previous model results and with observations. Finally, a summary of the results is given in section 4.

2. The model

The two-dimensional model of the atmosphere described by Harwood and Pyle (1975, 1977, 1980) was employed in this study. It calculates zonal mean values of temperature, wind components and chemical constituent mixing ratios with a resolution of $\pi/19$ in latitude, 0.5 in $\ln(p_o/p)$ in the vertical (where $p$ is pressure and $p_o$ is the surface value) and a 4-hour timestep. The model extends from pole to pole and from the ground to approximately 100 km. The model configuration is essentially that employed in GP with the addition of the improved Kelvin and gravity wave schemes. The gas kinetic data in the model have been updated in line with the recommendations of DeMore et al. (1983). Water vapour is now calculated in the stratosphere rather than being specified at 4 p.p.m.v. as in the initial study of GP. The tropospheric water vapour is still fixed; the lower boundary values for the stratosphere are taken to be the LIMS hygropause mixing ratios.

(a) Gravity wave parametrization

It has been recognized for some time that a form of damping of the zonal flow is necessary above about 50 km in numerical models of the atmosphere, in order to prevent the growth of unrealistically strong winds. Leovy (1964) first used a Rayleigh friction parametrization, which relaxed the winds to zero. This was generally rationalized as a qualitative parametrization of the wave drag due to the breaking of vertically propagating internal gravity waves. More recently, Lindzen (1981) and Holton (1982) have suggested improved schemes in which the passage of gravity waves through the atmosphere is modelled, with the associated momentum deposition and diffusion dependent on the background flow and details of the wave itself. The main advantage of these schemes is that they force the flow at mesospheric levels towards the phase speed of the wave, instead of towards zero. Because of selective absorption at critical levels in the stratosphere the gravity waves reaching the mesosphere generally have phase velocities in a direction
opposite to that of the flow at lower levels and the net forcing will also be in this direction. For a general review of gravity waves and their properties, see Fritts (1984).

The gravity wave parametrization in our model was developed by J. W. Barker (personal communication, 1985) and closely follows the schemes of Lindzen (1981) and Holton (1982) that have already been employed in similar models (e.g. Holton 1982, 1983; Garcia and Solomon 1985). For simplicity, the momentum forcing due to gravity waves is modelled but not the extra diffusion (note that chemical constituents are calculated in our model only up to about 60 km, below the level of maximum diffusion). Similarly, other refinements such as the net transport of chemical constituents by the gravity waves themselves and the effects of other gravity waves breaking nearby (Garcia and Solomon 1985), have not been included in this study. The parameters employed in this description of gravity wave breaking were partly chosen in order to produce a reasonable description of the zonal wind fields. Despite the many approximations this treatment is a significant improvement on the Rayleigh friction scheme.

The extra forcing term in the model momentum equation due to gravity waves is specified by

\[ F_x = -N^2 D/(\bar{u} - c) \quad z \geq z_b \]  

(1)

where

\[ D = \frac{K(\bar{u} - c)^4}{2N^3} \left( \frac{1}{H} - \frac{3d\bar{u}/dz}{\bar{u} - c} \right) \]  

(2)

and \( z_b \), the breaking level of the wave, is calculated using the equation

\[ z_b = z_o + 3H \ln((\bar{u} - c)/u^*) \]  

(3)

\( z_o \) is the lower boundary at which the wave amplitude is specified, \( N^2 \) is the static stability parameter, \( \bar{u} \) is the zonally averaged zonal velocity, \( c \) is the phase speed of the wave and \( K \) its horizontal wavenumber. The amplitude of the wave at \( z_o \) is specified through the parameter \( u^* \), defined by

\[ (u^*)^3 = B^2 N^2/[K^2(\bar{u}(z_o) - c)] = (2N/K)\bar{u}w' \big|_o \]  

(4)

where \( B \) is the amplitude of the vertical velocity perturbation at \( z_o \). To avoid discontinuities at the breaking level, the forcing was specified to decrease as \( \exp((z - z_b)/H) \) below \( z_b \) (\( H \) is the scale height of the atmosphere). A spectrum of gravity waves was specified with phase speeds \( c = 0, \pm 10, \pm 20, \pm 30 \) and \( \pm 40 \) m s\(^{-1} \). The amplitude \( u^* \) for each wave was specified using the formulation of Garcia and Solomon (1985):

\[ \bar{u}w' \big|_o = 10^{-2} \{\exp(-c_i/30)\}^2 \]  

(5)

for the \( i \)th component of the spectrum, so that

\[ u_i^* = [(2N/K) \times 10^{-2} \times \{\exp(-c_i/30)\}]^{1/3}. \]  

Note that Eq. (4) has no latitudinal dependence. There is no need in this model to introduce an arbitrary latitudinal distribution of forcing as we employ an inertial instability scheme (see below). Forcing that is constant with latitude represents the simplest experiment that can be performed. We would, of course, prefer to introduce a realistic latitudinal forcing distribution if one were available. The values of \( u_i^* \) are such that the waves achieve breaking levels in the mesosphere. Gravity waves in the atmosphere are believed to have wavelengths in the range 50–1000 km. Earlier studies (Holton 1982, 1983; Holton and Zhu 1984; Garcia and Solomon 1985) have found that using such values
in their parametrizations yielded too large a value of $F_x$. It is usually the practice to assume that wave breaking is not continuous either in time or in longitude at a given height and latitude, and hence to apply a so-called efficiency factor of approximately 10%. In this study, we specify a single horizontal wavenumber $K$ corresponding to a wavelength of 3000 km which may be considered to represent wavelengths of around 300 km that are present for only 10% of the time.

Acceleration of the zonal flow close to the equator can produce inertial instability. If this occurs in the model then an inertial adjustment scheme is activated in which the mean wind is adjusted back to a neutral inertial stability flow whilst preserving the mean angular momentum (Harwood and Pyle 1975; Dunkerton 1981; Holton 1983). Notice that the model will not actually represent the meridional circulations associated with inertial instabilities that occur in the atmosphere. Finally, above approximately 90 km we have added a Rayleigh friction which peaks at approximately 0.3 d$^{-1}$ at the top level of the model but falls off rapidly below 90 km. This may be regarded as a simple representation of the forcing associated with the breaking of the solar diurnal tide in the tropics (Lindzen 1981) and ion drag (Matsuno 1982; Garcia and Solomon 1985).

(b) The semi-annual oscillation

The westerly phase of the SAO is modelled in detail in this study by including the momentum deposition due to a thermally dissipating vertically propagating Kelvin wave. Kelvin waves are forced from below at the height $z_o$ (approximately 16 km) and are present at all seasons of the modelled year.

The formulation of the Kelvin wave momentum deposition follows that of Dunkerton (1979), who used the WKB approximation to derive an expression for the mean flow acceleration:

$$\frac{du}{dt} = A \exp\{(z - z_o)/H\} R(z) \exp\{-P(z)\}$$

where

$$R(z) = \alpha(z)N/(k(u - c)^2)$$ and $$P(z) = \int_{z_o}^{z} R(z') \, dz'.$$

$A$ is the vertical momentum flux at $z_o$, $\alpha(z)$ is the thermal damping rate, $N$ is the Brunt-Väisälä frequency and $k$ is the zonal wavenumber. The distribution of forcing in the meridional direction is assumed to be of normal mode (Holton 1975; Takahashi 1984), i.e.

$$\frac{du}{dt} = \left(\frac{du_{eq}}{dt}\right) \exp\{-2\Omega a \theta^2/(c - \bar{u}_{eq})\}$$

where $\bar{u}_{eq}$ is the zonal velocity at the equator, $\Omega$ is the rotation rate of the earth and $\theta$ is the latitude. The momentum forcing by a given wave therefore depends strongly upon the Doppler-shifted zonal wind speed at the equator, $(u_{eq} - c)$, as the wave propagates upwards, and upon the magnitude of the prespecified damping profile.

Following Dunkerton, we specify $c$, the phase speed of the Kelvin wave, to be 50 m s$^{-1}$ and the zonal wavenumber $k = 1$. The height of the tropopause, $z_o$, was taken to be approximately 16 km and the vertical momentum flux at this boundary, $A$, was assigned the value $7.0 \times 10^{-3}$ m$^2$s$^{-2}$. This is larger than used by Dunkerton, as his value was for a latitudinally averaged flux; also, the easterly phase in this model has a larger amplitude, so that greater acceleration is required to achieve a realistic westerly phase at the appropriate times. The thermal damping rate was chosen to be the 'slow' damping rate of Dunkerton (1979) with a functional fit given by
which peaks at approximately $2 \times 10^{-6}$ s$^{-1}$ at 50 km. Only the thermal damping of the Kelvin waves was considered in this model. Takahashi (1984) included an extra (mechanical) term dependent on the vertical eddy diffusion. Dunkerton (1979) showed that values greater than a threshold damping rate gave a severely reduced westerly acceleration at upper levels in his model. We also attempted to use faster, more realistic, damping rates but, like Dunkerton, found that this produced a poor representation of the zonal wind field. Similarly, in Takahashi’s model, the Kelvin wave momentum deposition did not produce sufficient westerly forcing in the region of 50–60 km to form a realistic SAO, possibly because his combined damping rates were too large.

The momentum deposition associated with Kelvin waves was present in our model at all times of the year and the amplitude of the forcing from the troposphere was kept constant with time. Recall that in GP the westerly forcing was not present during the solstices. It was found in the present study that, unlike GP, additional easterly forcing was necessary to produce the easterly phase of the SAO. Equatorward propagation of planetary wave momentum has been suggested as being important in forcing the easterly phase of the SAO. Indeed, it was this speculation that led GP to suggest that the SAO might be responsible for the double peak in tracer distributions, since the peak was more prominent following the disturbed northern hemisphere winter than it was six months later (Jones and Pyle 1984). It is not surprising that an additional easterly forcing is required in the model in order to simulate the easterly phase of the SAO. The model momentum fluxes are derived from satellite data which were necessarily averaged between 20°S and 20°N to obtain equatorial values, due to the lack of data. In this model then, an additional easterly momentum forcing was imposed at the appropriate seasons. The forcing was chosen to vary in time like $\sin(4\pi d/365)$ where $d$, the day number, was specified so that the maximum easterly winds occurred in June and again in December. Early estimates of the amplitude of the SAO (Reed 1966) indicated a zonal wind oscillation of $\pm 30$ m s$^{-1}$ at the equator. More recent estimates of zonal winds derived from the LIMS satellite data (Hitchmann and Leovy 1986; see Fig. 1) have suggested larger peak to peak variations, from approximately $-70$ m s$^{-1}$ to $+35$ m s$^{-1}$ in the region of 50 km. For this reason we chose to specify an easterly phase in our model that relaxed

![Figure 1](image_url)
toward the rather larger value of $-50\, \text{m s}^{-1}$ than was achieved in the initial study of Gray and Pyle. The velocity profile was relaxed towards $-50\, \text{m s}^{-1}$ on a timescale of 12.5 days. During the positive half of the cycle the forcing was set to zero. Thus the period of the SAO in the model was determined by the nature of the easterly forcing. The latitudinal profile of the easterly forcing was the same as for the Kelvin waves and was applied uniformly throughout the region 20–45 km.

3. Results

(a) Zonal wind fields

As a control run the model was integrated for almost four years with the gravity wave parametrization present, but no SAO. Figure 2 shows the time series of zonal wind at the equator for the last year of this integration, which we shall call run A. Without suitable westerly momentum deposition due to gravity or Kelvin waves the equatorial zonal winds in the model are easterly at all heights, due to the angular momentum constraint (see Fig. 2(a) of Harwood and Pyle (1980)). In run A, due to selective absorption of the gravity waves at lower levels, only those waves with westerly phase speeds are able to propagate into the mesosphere at the equator. Westerlies are evident above 55 km, overlying easterlies which peak at approximately $40\, \text{m s}^{-1}$ during the solstices. In a second integration (run B), both gravity waves and SAO were included; the model was initialized with data from year three of run A and run for several years to ensure that there was no significant drift of the fields with time. Figure 3 shows the zonal wind time series at the equator for this run. An SAO is evident between 30 and 50 km with westerlies in excess of $30\, \text{m s}^{-1}$ at the equinoxes. The gradual descent of the
westerlies with time compares extremely well with observations and is a considerable improvement on the wind fields of the original study of GP. A region of westerly acceleration of the zonal wind at the equator occurs in model run B in February at about 40 km (and again in August). This region gradually descends with time so that in March (September) it is maximum at about 30 km. In GP, with a 'forced' SAO, this descent was instantaneous. An SAO is also present in the mesosphere, with the stronger westerlies above the stratospheric easterly phase. Again, this is in good agreement with observations of the mesospheric SAO (Hirota 1978), which is out of phase with the stratospheric SAO.

(b) Meridional circulations

As shown in GP, the westerly acceleration induces a meridional circulation with descent over the equator and rising motion at mid-latitudes. GP showed that this induced circulation was sufficient to deform the contours of N₂O in their model and could account for the double peak observed in the SAMS nitrous oxide and methane observations as well as other related features observed in LIMS measurements. In Figs. 4(a) and (b) the latitude–height cross-sections of pressure-weighted velocity vectors in March are shown for model A and model B, respectively. In model run A two direct 'Hadley' cells are present with rising motion at the equator and descent at mid-latitudes. A strong indirect cell is present between 40°N and 60°N and a similar, but much weaker, cell is present in the southern hemisphere. In run B the equatorial circulation has been disturbed by the presence of the SAO; descending motion is now present over the equator at approximately 30 km. Solomon et al. (1986) and Hitchmann and Leovy (1986) used measurements from LIMS to diagnose a circulation of the stratosphere; the derived vertical velocities indicated a similar circulation, with descent at the equator during February and March and ascent at mid-latitudes. Although the latitudinal forcing of the SAO in our model is confined to equatorial regions the effects of the circulation that it induces are evident at much
higher latitudes. For example, the indirect cell at 40°N–60°N has decreased substantially, while in the southern hemisphere the indirect cell has increased in amplitude. These circulations at high latitudes in Fig. 4(b) are similar to the circulations of run A approximately 20 days further on in the run. The presence of the SAO has altered the temperature and hence the ozone distribution, so that the heating in the model has also changed. This has resulted in the solstice circulation developing more rapidly at high latitudes in run B than in run A.

(c) Tracer distributions

The meridional circulations induced by the gravity wave breaking and SAO formulations strongly influence the distribution of long-lived tracers in the model. Figure 5(a) shows the latitudinal cross-section of N₂O for March from a model run which contained neither the gravity wave breaking nor the SAO formulation (a Rayleigh friction forcing was present in order to close the stratospheric jets—see GP). A single maximum is situated at the equator and the mixing ratio decreases rapidly with height and towards the poles. The 2 p.p.b.v. contour shows a small ‘double peak’ feature due to the Rayleigh friction acting on easterlies above 55 km at the equator, which forces a small westerly acceleration. In Fig. 5(b) the corresponding plot from run A is shown, in which the
Rayleigh friction was replaced by the gravity wave-breaking parametrization. There is now evidence of a double peak structure extending down to 50 km. Finally, the corresponding plot from run B (Fig. 5(c)) which also included the SAO formulation, exhibits a strong double peak structure extending right down through the atmosphere to 35 km. Notice that it is the combined effects of both gravity waves and SAO forcing which gives the depth of structure seen in Fig. 5(c). A comparison of the methane distribution in the three model runs shows similar results.

One of the major improvements of this study over that of GP is that the Kelvin wave forcing is dependent on the background wind field. This leads to the descent of the! zero. wind isopleth with time (Fig. 3) and, in effect, produces a westerly forcing which is of longer duration than in GP. The double peak structure in the model is now present for several months around each equinox. The development of the double peak is illustrated in Figs. 6 and 7, which show the monthly mean cross-sections of \( \text{N}_2\text{O} \) from model runs A and B. In the following discussion we shall use the distribution of \( \text{N}_2\text{O} \) to illustrate the improvements in the modelled tracer fields. Similar improvements were also evident in the modelled \( \text{CH}_4 \) distribution. In run A, with gravity wave forcing but no SAO, the local maximum in \( \text{N}_2\text{O} \) above 45 km moves from 60\(^\circ\)S at the December solstice across to approximately 20\(^\circ\)N at the June solstice. Below 45 km the distribution of \( \text{N}_2\text{O} \) does not vary substantially with the time of year. The morphology of run B (Fig. 7) is very similar to that of run A above 45 km, with an off-equator peak in the mixing ratio situated in the summer hemisphere during the solstices. Below 45 km, however, a double peak structure is present for several months around the equinox in the periods March–May and September–November. The feature is coherent throughout these periods. The structure and time evolution of the modelled double peak compares favourably with the SAMS observations of \( \text{N}_2\text{O} \) (Jones and Pyle 1984; Jones 1984).

Several of the features in Figs. 6 and 7 suggest that there are still some problems with the treatment of tracer transport in the model. Firstly, the absolute mixing ratios of \( \text{N}_2\text{O} \) are lower than the SAMS measurements, particularly at the equator. Secondly,
advection by the indirect circulation in the winter hemisphere polar regions has resulted in fairly high mixing ratios at the winter pole in the model (see for example, 60°-80°N in January of both runs A and B). A similar feature is not evident in the SAMS data (although the limited latitudinal extent of the satellite coverage may mask this to some extent). This points to a possible deficiency in the cancellation of mean and eddy motions in the model.

A further improvement in this work is that H₂O is modelled in the stratosphere, rather than being specified as in GP (see section 2). Observations of H₂O show rather interesting double peak phenomena. For example, Remsberg et al. (1984) show double minima in the stratosphere below 40 km with a local maximum at the equator; on the other hand, at the stratopause maxima are found at ±30° latitude with a minimum at the
equator. In GP we speculated that this pattern is due to the fact that H$_2$O increases with height in the stratosphere and that the level of maximum momentum forcing associated with the SAO is just below the stratopause. Induced downward motion over the equator in the lower stratosphere will lead to enhanced water vapour mixing ratios there and hence double minima in the subtropics. Above the peak of the forcing, the circulation is opposite to that below, with rising motion over the equator and sinking in mid-latitudes (see Fig. 3 in GP). Such a circulation should produce a double maximum in the water vapour fields at the stratopause. Figure 8 shows a latitude section of water vapour mixing ratio at about 40 km in March from runs A and B. As expected, by including the SAO forcing a maximum is found at the equator in run B, with local minima to either side, due to the reasons outlined above. No convincing evidence was found of a double maximum at greater altitudes, which we attribute in this model to the proximity of the upper boundary for the chemical constituent calculations at 60 km.

(d) Asymmetries in the modelled SAO

An interesting feature of the zonal winds shown in Figs. 2 and 3 is the marked asymmetry between June and December. In run A, the westerlies are of larger amplitude at 65 km in June than in December, and they extend over a greater depth of the mesosphere. In run B, the asymmetry is even more marked. For example, in the region 60–70 km at the June solstice an extensive region with velocities greater than 20 m s$^{-1}$ is evident, with peak velocities reaching 30 m s$^{-1}$; at the December solstice the velocities are much reduced and barely reach 20 m s$^{-1}$. This situation is reversed in the region 45–60 km, with velocities of 30 m s$^{-1}$ in December and about 20 m s$^{-1}$ in June. Between 35 and 45 km the amplitude of the easterly phase of the SAO is greater in December than in June; for example, at 42 km the equatorial velocity in December is $-20$ m s$^{-1}$ while in June it is zero. It is important to be aware that the velocities observed at the equator cannot be interpreted in isolation from the mid-latitudes. The planetary wave momentum forcing in the model is derived from satellite data (for a discussion of this, see Crane et al. (1980)), and we expect to see asymmetries between the mid-latitude jets in the two
hemispheres. This, in turn, will affect the magnitude of the advection of easterlies into the equatorial regions. In run A, a possible cause of the asymmetry above 60 km is an asymmetry in the strength of the meridional circulation induced by gravity wave drag at mid-latitudes (via the dependence on \( \bar{u} - c \)). This is stronger in December than in June and the enhanced easterly advection results in weaker westerlies at the equator.

An asymmetry in the equatorial zonal wind at lower levels may also be a direct source of asymmetry at higher levels, and this appears to be the case in run B. The
amplitude of the momentum deposition due to Kelvin waves depends on the vertically integrated ($\bar{u} - c$) term in Eq. (1). Any asymmetry present in the background wind field due to an asymmetry in eddy momentum forcing or in the advection of easterlies from the summer hemisphere therefore has the potential to be felt at higher altitudes. The sensitivity of the equatorial zonal winds above 50 km to the asymmetry of the background winds at lower levels was investigated in a series of model runs in which we attempted to reduce this asymmetry. Figure 9(a) shows the zonal wind field at the equator from a model run (run C) which was identical to run B except that the easterly SAO forcing was imposed up to 60 km and on a much shorter timescale of 2-5 days. By relaxing towards the desired easterly profile on a faster timescale we hoped to reduce the asymmetry in the easterly phase. However, despite the faster relaxation forcing, there was still a significant asymmetry in the low-level easterlies between 35 and 55 km and in the westerlies above 60 km at the solstices. (Because the easterly forcing was imposed up to 60 km in this run instead of only 45 km, the easterly phase in this run extended to 50-55 km and compares more favourably with observations—see Fig. 1; however, the model tracer fields did not alter substantially as a result of this improvement.) In a subsequent experiment, in which the easterly forcing was forced on an even shorter timescale (half a day), the asymmetry in the low-level easterlies was still present. As the imposed easterly forcing in this latest model run was much larger than the forcing due to the eddy momentum fluxes derived from satellite data, it seemed likely that the asymmetry was, in fact, due to the advection of easterlies from the summer hemisphere. Therefore, in a further model run, the latitudinal extent of the imposed easterly SAO forcing was doubled (see section 2). This ensured that advection in the region up to 55 km would merely advect easterlies of a similar magnitude to those already present at the equator and hence no asymmetry could be introduced. Figure 9(b) shows the resulting time series of zonal wind at the equator; the asymmetry in the low-level winds at 45–55 km at the solstices has been eliminated and so too has the strong asymmetry in both the stratospheric and mesospheric westerlies. This series of experiments illustrates the sensitivity of the modelled SAO to the wind field at lower levels. We expect the SAO, therefore, to be modulated by the phase of the QBO (Dunkerton 1979), which has its maximum amplitude at approximately 25 km and exhibits alternating easterlies and westerlies with a period of about 27 months. The inclusion of the QBO in our model and its effect on the SAO and on tracer distributions will be the topic of a future study.

The asymmetry in the equatorial zonal winds of run B (Fig. 3) did not produce a noticeable asymmetry in the modelled N₂O fields (Fig. 7). The SAMS satellite data, on the other hand, display a marked asymmetry, with a more prominent double peak in March–May than six months later. GP proposed a connection with the greater level of planetary wave activity during the northern hemisphere winter. They suggested that the resultant stronger easterly SAO phase around December would give rise to stronger westerly accelerations of the zonal winds during the March equinox and hence to a stronger double peak. We have tested this idea with our model, by forcing the equatorial velocities towards $-50 \text{ m s}^{-1}$ during the December solstice (as in all previous runs), but towards only $-20 \text{ m s}^{-1}$ during the June solstice, thus introducing a much more substantial asymmetry at approximately 50 km than was already present. Figure 10 shows the time series of equatorial zonal winds for this run. In the region 30–50 km, larger westerly acceleration is evident in February and March than in August and September. Figure 11 shows the resulting N₂O distribution in the months September, October and November. In comparison with the corresponding fields from run B (Fig. 7) the amplitude of the double peak during these months has been severely reduced, but the fields in March, April and May are unchanged. This model run therefore strongly supports our suggestion
that the larger amplitude of the double peak observed by SAMS during March–May is associated with the higher level of planetary wave activity during the northern hemisphere winter.

4. CONCLUSIONS

In GP it was shown that the semi-annual oscillation plays an important role in determining equatorial tracer distributions. That study was preliminary, in that the SAO was forced by specification of a westerly momentum sufficient to reproduce the observed wind fields. In this paper, the dynamical treatment of the SAO forcing has been improved, with a number of aims: firstly, to see whether improved forcing leads to improvements in the modelled wind fields; secondly, to demonstrate whether this in turn leads to a better representation of the model tracer fields.

The first major enhancement compared with GP is in the treatment of the SAO westerly forcing, for which we employ the WKB approximation to equatorial Kelvin wave propagation, following Dunkerton (1979). The westerly momentum forcing in the model is present at all times. Its strength at any height depends on the background zonal wind fields, which leads to a mechanism for self-regulation of the SAO.

The second major enhancement is a treatment of the momentum deposition by gravity waves, following Lindzen (1981). This replaced the Rayleigh friction as a means of closing the mid-latitude jets in the upper mesosphere. The treatment depends similarly on the background wind fields, which has led to a number of interesting features in low latitudes. For example, we find an SAO in the upper mesosphere which is out of phase with the stratospheric SAO, in agreement with observations (Hirota 1978).

The introduction of a better physical basis for the treatment of Kelvin and gravity waves has led to significant improvements in the modelled wind fields and this in turn
Figure 11. Monthly mean cross-sections of $N_2O$ (p.p.b.v.) for September, October and November from a model run in which the applied easterly forcing of the SAO at the June solstice was reduced compared with the December solstice.

has produced a more satisfactory model simulation of equatorial tracer distributions. The descent of the zero wind isopleth with time implies a forcing of westerly winds which is of longer duration than in the study of GP and led, for example, to double peak structures that last for a number of months. This is in good agreement with observations and contrasts with GP where the zero wind isopleth was found simultaneously at all heights in the middle stratosphere, when the forcing was of short duration and when the double peak lasted typically about a month.

The gravity wave treatment has led to the double peak structure being obtained throughout the stratosphere and mesosphere. The structure now shows greater depth and coherence. In contrast, GP found a double peak situated at about 40 km; the Rayleigh friction treatment tended to oppose the production of the double peak at higher altitudes.

Advection of easterlies from the summer hemisphere was not sufficient to produce a reasonable easterly SAO phase in this model. Additional easterly forcing was therefore applied at the appropriate times of the year. A model run in which the amplitude of the
easterly phase of the SAO was reduced in June compared with December, confirmed the speculation in GP that the stronger double peak in N\textsubscript{2}O and CH\textsubscript{4} observed by SAMS in March–May may be associated with enhanced planetary wave activity in the northern hemisphere winter. We believe that this is further evidence for the importance of planetary waves in the easterly acceleration phase of the SAO.

Another interesting feature, which warrants a close analysis of available data, is the presence of an annual component in the equatorial zonal winds at most levels. This appears to be primarily due to asymmetric advection of easterly momentum. In our model there is a hemispheric asymmetry in the strength of the mid-latitude jets arising from the use of eddy momentum fluxes from satellite data. In a series of experiments, we have highlighted the sensitivity of the modelled SAO to the magnitude of the zonal winds at lower levels.

Finally, the inclusion of H\textsubscript{2}O as a modelled species has led to the simulation of the observed double minima in the mid-stratosphere, confirming the speculation of GP. The double maxima at and above the stratopause were not, however, reproduced due to the proximity of the upper boundary. We aim to investigate this feature, amongst others, in a further study with the photochemical scheme extended to the mesopause level.

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