The Stratopause Semiannual Oscillation in the NCAR Community Climate Model

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(Manuscript received 29 October 1992, in final form 16 March 1993)

ABSTRACT

The middle atmospheric version of the NCAR Community Climate Model (CCM2) has been used to study the development of the equatorial semiannual oscillation (SAO) in the stratosphere. The model domain extends from the ground to about 80 km, with a vertical resolution of 1 km. Transport of nitrous oxide (N₂O) with simplified photochemistry is included in the calculation to illustrate the influence of tropical circulations on the distribution of trace species. Diagnosis of model output reveals two distinct phases in the evolution of the zonal mean state on the equator. In early December, a strong and broad easterly jet appears near the stratopause in connection with a midlatitude wave event (sudden stratospheric warming) that reverses the winter westerlies of the Northern Hemisphere throughout the upper stratosphere. When the wave forcing dies out, the radiative drive allows the westerlies to recover at midlatitudes, while easterlies persist in the tropics. The resulting strong meridional gradient of the zonal mean wind provides favorable conditions for the development of inertial instability at lower latitudes. The meridional circulation associated with the instability shapes the "nose" of the easterly jet, reducing the extension of the unstable region.

In equinoctial conditions, a jet of westerlies appears in the lower equatorial mesosphere and descends to lower altitudes; positive accelerations associated with the descending westerlies are due primarily to Kelvin waves. The descent of the westerly jet does not reproduce well the observed behavior of the SAO westerly phase, either in amplitude or in the extent of downward propagation. As a consequence, the model does not simulate the "double peak" observed in the tropical distribution of N₂O. Comparison of wave amplitudes in the model with those derived from satellite observations shows that the calculated amplitudes are larger than observed in the upper stratosphere. It follows that inadequate Kelvin wave forcing is not the cause of the weak westerly phase in the model, and that some other mechanism must be responsible for the generation of the strong westerly phase observed.

1. Introduction

The existence of a semiannual oscillation (SAO) in the tropical middle atmosphere was discovered by Reed (1962) using temperature data obtained from radiosonde observations. Subsequent work by Reed (1965, 1966) with rocketsonde data demonstrated the existence of an associated oscillation in zonal wind extending to the stratopause, with an amplitude of some 30 m s⁻¹ near 50 km. Reed's findings have since been confirmed by numerous observational studies that have documented semiannual oscillations in zonal wind and temperature in both the stratosphere and mesosphere (Quiroz and Miller 1967; vanLoon et al. 1972; Belmont et al. 1974, 1975; Cole and Kantor 1975, 1978; Hopkins 1975; Delisi and Dunkerton 1988a,b; Garcia and Clancy 1990). Hirota (1978, 1980) and Hamilton (1982) have shown that the mesospheric zonal wind SAO extends to the top of the mesosphere, its amplitude peaking in the vicinity of the mesopause. The work of these authors also shows that the stratospheric and mesospheric oscillations are related and approximately out of phase between the stratopause and mesopause.

Initial attempts to understand the stratospheric SAO focused on the seasonal variation of ozone heating near the tropical stratopause (Reed 1962; Webb 1966). However, Reed himself recognized that the latitudinal extent of the oscillation appeared to be too great to be consistent with radiative heating, whose semiannual harmonic decays rapidly away from the equator. Later work by Mayer (1970) using a simple numerical model demonstrated that seasonal variations in tropical insolation are insufficient to produce the observed SAO and, in fact, that a realistic SAO could be generated only if it was assumed that a semiannually varying eddy momentum source was present. It is also possible to show by simple analytical arguments (Andrews et al. 1987) that time-varying radiative forcing in the tropics cannot account for the magnitude of the observed SAO.

It is now generally agreed that deposition of zonal
momentum by waves plays a central role in the SAO. The case is particularly compelling for the westerly wind phase, since no zonally symmetric mechanism can produce absolute westerlies at the equator. In spite of this consensus, the precise nature of the relevant waves remains uncertain 30 years after the discovery of the oscillation. Holton (1975) suggested that high-frequency Kelvin waves propagating from the lower atmosphere could account for the westerly phase of the stratospheric SAO. This idea received support from Hirota's (1978, 1980) analyses, which showed the presence of fast Kelvin waves ("Hirota waves," of period 4–10 days) in the upper stratosphere. Further support for the Kelvin wave hypothesis was added by the work of Salby et al. (1984), who used data from the Limb Infrared Monitor of the Stratosphere (LIMS) to confirm the existence of the Hirota waves and to document the existence of "superfast" Kelvin waves (period 3.5–4 days) at the highest levels (60–65 km) observed by LIMS. At the same time, analysis of model simulations suggests that the same kind of waves can play a role in driving the system (Hayashi et al. 1984; Boville and Cheng 1988; Boville and Randel 1992), although most of these numerical experiments do not yield the correct evolution of the mean state on the equator.

The westerly phase of the stratospheric SAO has been attributed to easterly momentum deposition by planetary waves propagating into the tropics (Hopkins 1975) and to nonlinear advection of mean easterly momentum by the summer-to-winter mean meridional circulation (Holton and Wehrbein 1980). The two processes are not mutually exclusive; on the contrary, insofar as a mean meridional circulation is driven by planetary wave Eliassen–Palm (EP) flux divergences, wave forcing and mean advection both arise from the same basic mechanism (dissipation of planetary waves). This explanation of the easterly phase of the stratospheric SAO is rather different from its counterpart for the quasi-biennial oscillation (QBO) in that it does not require the presence of equatorial waves capable of producing easterly forcing. The connection of the SAO easterly phase with the annual cycle (planetary wave forcing and mean momentum advection maximize during the winter season of each hemisphere) also distinguishes the SAO from the QBO, and evidently is responsible for the period of the former being a harmonic of the annual cycle while that of the latter can vary significantly from its mean value of ~27 months.

The picture of the stratospheric SAO outlined in the preceding paragraphs must be modified in view of the recent studies of Hitchman and Leovy (1986, 1988) and Hitchman et al. (1987), who showed that 1) the westerly momentum forcing by Kelvin waves deduced from LIMS data is insufficient to account for the observed westerly accelerations, and 2) inertial instabilities are commonplace in the tropical upper stratosphere during the solstices and may play an important role in the development of the easterly phase. It is conceivable that the additional acceleration that appears to be required to drive the SAO westerly phase is provided by smaller-scale gravity waves. The gravity wave hypothesis is attractive in that it could also help account for the occurrence of the mesospheric SAO, and for the phase relationship between the stratospheric and mesospheric oscillations. Numerical modeling by Dunkerton (1982) suggests that, given a spectrum of gravity waves that contains components with positive and negative phase velocities, the interplay of these waves with the easterly phase of the SAO in the stratosphere can give rise to a mesospheric SAO out of phase with the stratospheric SAO, more or less as observed. The modeling work of Dunkerton also indicates that the SAO may be "self-enhancing," in the sense that a strong easterly phase will be followed by a strong westerly phase, and vice versa. Delisi and Dunkerton (1988b) have used this idea to explain the seasonal asymmetry of the stratospheric SAO, which is stronger during the cycle encompassing northern winter and spring than during the remainder of the year. Seasonally asymmetric behavior, which also could be attributed to the self-enhancement mechanism, has been found for the mesospheric SAO by Garcia and Clancy (1990).

It should be clear from the foregoing discussion that, although the mechanism of the SAO may be understood in general terms, many of its components are poorly understood in detail. In particular, the roles and relative importance of planetary-scale Kelvin waves and small-scale gravity waves are uncertain. In the present study, we use a middle atmospheric version of the NCAR Community Climate Model (CCM2) whose domain has been extended to an altitude of 80 km to study the generation of the stratospheric SAO. The mesospheric oscillation, which affects primarily altitudes above 65 km, cannot be studied with the model. Although the top boundary of the model is located near the mesopause, altitudes above 65 km act as a "sponge layer" to prevent reflection of upward-propagating wave activity and do not provide a realistic representation of the dynamics of the upper mesosphere. For the easterly phase of the stratospheric SAO, we focus attention on the role of inertial instabilities and of hemispheric asymmetries in planetary wave driving during winter. For the westerly phase, we attempt to identify the eddy motions responsible for producing westerly accelerations.

The remainder of the paper is organized as follows. Section 2 provides an overview of the version of CCM2 used in the study; section 3 contains a description of one model annual cycle, while section 4 analyzes the processes responsible for generating the easterly and westerly phases of the SAO. Conclusions are presented in section 5.
2. Description of CCM2 and experiments

This study uses a recent version of the NCAR Community Climate Model (CCM2), configured for simulation of the troposphere and stratosphere. The formulation of CCM2 differs significantly from that of the previous version (CCM1), on which several previous stratospheric simulations were based. Detailed descriptions of CCM2 are presented elsewhere and a brief description will be given here, since only the horizontal truncation and vertical layering differ from the standard model.

CCM2 is a global general circulation model that solves the primitive equations using the spectral transform method in the horizontal and finite-difference approximations in the vertical. The transport of water and trace constituents (N2O in the present case) uses the three-dimensional semi-Lagrangian transport method described by Rasch and Williamson (1990). The hybrid vertical coordinate described by Simmons and Struifing (1981) is used with coefficients set so that the model surfaces reduce to constant pressure surfaces above 100 mb.

The focus of this study is on tropical processes and the horizontal and vertical resolution were chosen accordingly. These choices have been guided by the results of resolution experiments by Boville (1991) and Boville and Randel (1992). A T31 horizontal truncation was used in the present study, resulting in a shortest resolved zonal wavelength of ~1300 km, and corresponding approximately to a 4° grid resolution. This truncation appears to be adequate to resolve the important large-scale dynamics in the tropics while being inexpensive enough to allow very high vertical resolution. The vertical domain extends from the earth's surface to a rigid lid near 80 km (actually 2 Pa) using 75 levels. The spacing of the levels [ln(p)] increases slowly through the lower troposphere to a uniform Δ ln(p) = 0.15 (~1 km) above 650 hPa. The results of Boville and Randel (1992) indicate that 1-km vertical resolution should be adequate to resolve most of the equatorial waves believed to be important in the SAO.

The chemistry of N2O is extremely simple, since it is dominated by photolysis in the upper stratosphere. In the present study we parameterize this process as

\[ \frac{\partial q}{\partial t} = - \frac{q}{\tau(y, z, t)} \]

where y, z are latitude and height, respectively, and t is the time of the year; q is the N2O mixing ratio, and \( \tau(y, z, t) \) has been interpolated for CCM2 from the monthly mean photochemical lifetime (Solomon et al. 1986). The initial condition for N2O was specified from the simulation of Garcia and Solomon (1983). In order to reduce the effects of the initialization (1 September), the model has been run for two years; in this work we will present results from the second year of simulation.

Several physical parameterizations are of particular relevance to models extending through the stratosphere. Momentum flux divergence by stationary gravity waves is parameterized following McFarlane (1987), and also by Rayleigh friction following Boville and Baumhefner (1990). The shortwave radiation parameterization (Briegleb 1992) uses the 6-Eddington method and incorporates both diurnal and annual cycles of insolation. The longwave radiation parameterization has been updated to allow for Voigt line shapes (principally affecting the upper stratosphere). Biharmonic (7°) horizontal diffusion is included with coefficient 2 \( \times 10^{16} \) m² s⁻¹, as in Boville (1991). Vertical diffusion depends on the local stability (through the Richardson number) as described in Holtslag and Boville (1993), with a (negligible) minimum diffusivity of 0.01 m² s⁻¹.

The stratospheric simulation is also indirectly affected by parameterizations that are primarily active in the troposphere. These parameterizations alter the tropospheric circulation and, thus, the generation of both planetary and gravity waves and the upward radiative flux. The cloud fraction parameterization used for radiative purposes is a generalization of the method proposed by Slingo (1987). The diagnosed cloud fraction depends on relative humidity, vertical motion, static stability, and precipitation rate. Once clouds appear, their liquid water concentration is specified as a function of latitude and height for the radiation calculations. Boundary-layer transports are parameterized using a nonlocal diffusion approach (Holtslag and Boville 1993). Shallow and deep convection are parameterized by a simple mass flux scheme. Land temperatures are predicted using a diffusion equation for the surface and three subsurface layers with differing heat capacities. Sea surface temperatures are prescribed by linear interpolation between climatological monthly mean values.

3. Model annual cycle

In this section we illustrate some of the results (monthly means of zonal mean wind and N2O) obtained with the model described in section 2; we will address the principal features of the zonal mean state in solstitial and equinoctial conditions during the modeled second year, focusing our attention on equatorial events.

Figure 1 illustrates the daily evolution of the zonal mean jet on the equator from 1 September through 31 August of the following year. It is apparent that the two westerly phases are quite weak (with the spring phase stronger than the fall one) and do not descend below ~1 mb. Observations (Hitchman and Leovy 1986) indicate that the zero wind line should reach 8–9 mb; this model simulation falls ~10 km short. In-
Fig. 1. Time–height profile of equatorial mean winds, from 1 September through 31 August. Contour interval is 10 m s\(^{-1}\).

Spectroscopic analysis of the equatorial peak of westerlies (~0.4 mbar) in late March (not shown) reveals a latitudinal extent of 10°–13°, while LIMS data show broader westerlies (see Hitchman and Leovy 1986) extending up to 20°; this seems to happen in connection with weaker westerly accelerations in the modeled stratopause SAO. The vertical shear of \(\vec{u}\) is also weaker than observed, which implies weak sinking motions on the equator. Differences in the easterly phases are evident. During northern winter, the easterly phase develops quite rapidly with the formation of the jet in early December, while during the southern winter the evolution is slower and smoother. The appearance of easterlies early in December at the stratopause is observable in LIMS data (i.e., Hitchman and Leovy 1986; see their Fig. 3); while these data cover only the northern winter season, other model simulations have revealed different evolution of the zonal mean winds between the two solstices (see, for example, Hamilton and Mahlman 1988, their Fig. 6).

Figure 2 shows the monthly means for September; the zonal mean wind (Fig. 2a) illustrates the westerly phase during Northern Hemisphere fall equinox. The equatorial westerlies are quite weak (the maximum strength of the jet is 11 m s\(^{-1}\)) in the vicinity of the stratopause, and the nitrous oxide field (Fig. 2b) does not show the double-peak feature observed in data from the Stratosphere and Mesosphere Sounder (SAMS) by Jones and Pyle (1984). Although the observed oscillation in N\(_2\)O is much weaker in September–October than in March–April (e.g., Jones and Pyle 1984), there is always an indication in SAMS data of a distinct flat-

Fig. 2. Monthly means for September of the zonal mean wind (a) and zonal mean mixing ratio of nitrous oxide (b). Contour intervals are 10 m s\(^{-1}\) and 20 ppbv.

Fig. 3. Same as in Fig. 2 but for December.
of the jet intrudes into the winter hemisphere, and its vertical dimension is reduced, forming an easterly "nose," much narrower than the corresponding feature in December. In the equatorial upper stratosphere, the isopleths of N_2O have risen slightly in comparison with the previous month. The associated meridional circulation is examined in detail in the next section.

Approaching the March equinox (Fig. 5) a westerly jet appears at the upper levels in the tropics, together with a positive temperature anomaly (not shown) where the vertical gradient of \( \bar{u} \) reaches its maximum strength. The temperature anomaly can be easily explained considering the positive acceleration associated with the descending westerlies and the transport circulation in the meridional plane, which balances this acceleration in the zonal momentum equation. Thus, the presence of a positive temperature anomaly identifies the region of equatorial downwelling (see Hitchman and Leovy 1986). We note that, according to the weak shear of \( \bar{u} \), the temperature anomaly in this model run is about half of the one in LIMS data (~11 K). In the N_2O field, there is some flattening of the isopleths in the upper stratosphere compared to January but, contrary to SAMS observations, no indication of a double peak: in SAMS data (see Jones and Pyle 1984)

FIG. 4. Same as in Fig. 2 but for January.

FIG. 5. Same as in Fig. 2 but for March. Also plotted is the 5-ppbv isoline.
this feature is evident even in the 20 ppbv isoline, which descends to about the \( \sim 3 \) mb level. In Fig. 5b we have also plotted the 5 ppbv isoline, which carries some indication of a double peak at \( \sim 0.4 \) mb; the location is coincident with the maximum westerly shear in March (compare to Fig. 1) and is indicative of the downwelling associated with it, even though at a much higher altitude and much weaker than observed.

In April (not shown), the westerly jet descends to about 50 km, compared to 30–40 km in observations from LIMS reported by Hitchman and Leovy (1986). Furthermore, the jet is weak (barely 20 m s\(^{-1}\) in the monthly mean), while LIMS observations suggest values over 30 m s\(^{-1}\). By the end of April (see Fig. 1), the zero wind line has descended to \( \sim 0.8 \) mb. The weakness of the jet (and hence, of the associated downwelling) again fails to produce a double-peak feature at the stratopause level. Instead, there is some additional flattening of the isopleths of \( \text{N}_2\text{O} \) on the equator in the upper stratosphere (not shown).

During the remainder of the model seasonal cycle, we can observe the formation of the subsequent easterly phase. Figure 6 illustrates the mean fields in July. At the equatorial stratopause, the zonal mean wind is \( \sim -30 \) m s\(^{-1}\), compared to \( \sim -20 \) m s\(^{-1}\) in June (not shown). We note in passing that the depression (between \( \sim 10 \) and 20 mb) of \( \text{N}_2\text{O} \) mixing ratios over Antarctica (Fig. 6b) is due to descent associated with diabatic cooling following sunset at those latitudes (the temperature—not shown—drops from \( \sim 225 \) K in March to \( \sim 195 \) K at 10 mb). The strength of the equatorial easterlies is much less than in January (Fig. 3). We show in the next section that, contrary to the situation in December–January, a strong inertially unstable region does not form near the equator in June–July. Instead, the evolution of the mean state suggests that dissipation of incipient inertial instabilities takes place at this time.

4. Discussion

4a. Easterly phase: Formation and evolution

Observations (e.g., Hitchman and Leovy 1986) indicate that the formation of an easterly jet near the equator takes place sometime before the solstices, at about stratopause height. As mentioned in the Introduction, different authors have suggested that coupling between equatorial regions and midlatitudes might occur via quasi-stationary planetary waves. In the following discussion we show that the presence of wave activity from the winter hemisphere can be very important in the formation of the easterly phase. Figure 7 shows the evolution of the zonal mean wind from 30 November to 22 December, each panel about 7 days apart. Figure 7a (30 November) illustrates a typical wintertime condition, with predominant easterlies in the summer hemisphere and westerlies in winter. Near the equator the winds do not show any particular feature indicating the formation of the broad easterly "nose" seen in the December monthly mean (Fig. 3). In addition, although the maximum strength of \( \bar{u} \) is about \( -50 \) m s\(^{-1}\) at 45 km and 10°S, on the equator itself the zonal mean wind is quite weak at the stratopause. Just a week later (Fig. 7b, 7 December), the speed of the polar night jet is reduced sharply (from more than 100 m s\(^{-1}\) at 55 km and 60°N to 33 m s\(^{-1}\) at the same location) and easterlies appear in the subtropics. Between the equator and 30°N, the winter stratopause is dominated by easterlies. The upper stratosphere is evidently undergoing a modification of its mean state that assumes dramatic proportions by 13 December (Fig. 7c), when zonal winds become easterly throughout the Northern Hemisphere between 40 and 60 km. The core of the tropical easterly jet reaches \( -88 \) m s\(^{-1}\) at 10°S, while on the equator at 50 km typical values are \( -60 \) m s\(^{-1}\). The situation depicted in Fig. 7c can be classified as a major sudden warming. By the time the wave event dies out (Fig.

\[\text{1 It is worth noting that the official designation of "major warming" applies to the lower stratosphere (see Baldwin and Dunkerton 1989); however, the same dynamical mechanisms responsible for the lower stratospheric warming also account for the upper stratospheric event. The official definition is unduly restrictive.}\]
7d, 22 December), the radiative forcing is able to restore westerlies at midlatitudes, while easterlies persist near the equator. This produces very large meridional gradients of $\tilde{u}$ in the subtropics, which can exceed the local value of the Coriolis parameter. It can be easily shown that if

$$ \frac{\partial \tilde{u}}{\partial y} > f, \quad (4.1) $$

then the angular momentum of air parcels increases toward the axis of rotation, with consequent unstable motions in the meridional plane. Eventually, meridional transport will redistribute angular momentum on the northern flank of the equator, according to Dunkerton's (1981) picture of inertially unstable motions.

The evolution of the zonal mean wind gives us only a hint of what is going on, but a satisfactory illustration can be achieved using a more sophisticated tool. Ertel's potential vorticity (EPV) can be used to visualize the evolution of material elements under the influence of wave motions, especially on time scales of a few days, for which the motions can be considered adiabatic. Figure 8a shows time-averaged (5–7 December) EPV contours on the 1800-K surface (about 1.1 mb). In addition to the large planetary-scale wave at middle and high latitudes, a wide surf zone is easily identifiable throughout the subtropics of the winter hemisphere. McIntyre and Palmer (1983, 1984) have studied the breaking of planetary waves in the stratosphere and noted that, in the surf zone, linear propagation fails and wave overturning takes place with nonlinear mixing of parcels on all scales. The process depicted in Fig. 8a involves differential advection of air with high EPV from midlatitudes and air with low EPV from lower latitudes and even from the summer hemisphere. As a consequence, “tongues” of air with negative EPV appear in the Northern (winter) Hemisphere.
Comparing Fig. 8b with the EPV map of Fig. 8a, we observe regions of strong divergence at approximately the same location of negative EPV. This implies that parcels entering the region of negative EPV are accelerated and diverge in the horizontal plane (positive divergence patterns) while, on leaving the region of negative EPV, restoring forces decelerate the parcels and cause convergence (negative divergence patterns). Eventually, dissipative processes would allow for an irreversible mixing of momentum. The same conclusion has been reached by O'Sullivan and Hitchman (1992) using a mechanistic model of the middle atmosphere. They showed that the correlation between anomalous EPV and divergence/convergence patterns is not accidental: interaction between the tropics and midlatitudes via large-scale Rossby waves yields differential advection of EPV with consequent parcel acceleration. In the evolution that ensues, the winding up of EPV tongues produces an unstable region, converting energy from rotational into divergent motions. The behavior illustrated in Fig. 8 is an example of zonally asymmetric inertial instability (Dunkerton 1983) facilitated by the rearrangement of potential vorticity by extratropical Rossby waves (Hitchman et al. 1987). More recently, Dunkerton (1993) has studied with a simplified shallow-water primitive equation model the occurrence of inertial instability near the equator in the presence of a stationary, wavelike perturbation of the mean zonal flow. His results suggest that in the middle atmosphere local stationary inertial instability is more likely to occur and has the highest growth rate. The behavior illustrated in Fig. 8 appears to be of this kind; the pattern shown in Fig. 8 remains stationary throughout the duration of the wave event.

We conclude from the foregoing that, in the model, the initial stages of the evolution of the SAO easterly phase are strongly influenced by the presence of large-scale planetary waves from the winter hemisphere. Later on, as the wave event dies down, the same picture no longer holds. In fact, a similar computation of EPV in mid-January (Fig. 9a) shows only weak wave activity in midlatitudes. On the other hand, there is a large belt of negative EPV between the equator and 15°N, associated with a sizable horizontal divergence pattern (Fig. 9b). We emphasize that, contrary to the situation in December, the inertially unstable region is nearly zonally symmetric at this time. Model results for January do not reproduce any of the features of asymmetric inertial instability. It appears, then, that the evolution of the easterly phase in this particular model run may involve several processes: reversal of the zonal mean wind in the upper stratosphere, formation of a surf zone all across the winter hemisphere, asymmetric inertial instability, and, at other times, symmetric inertial instability.

An analysis of the differences in the evolution of the easterly phase between northern and southern winter sheds light on the role of wintertime wave activity. Fig-

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2 Isobaric and isentropic surfaces are not coincident at all latitudes; the isobaric levels shown in Figs. 8b, 9b, and 10b are, strictly speaking, coincident with the isentropic surfaces of Figs. 8a, 9a, and 10a only in some latitude range (~20°) near the equator where T varies less than 5 K. The noncoincidence of the two surfaces may introduce a slight phase difference between the anomalous EPV and the divergence field. The reader should keep these considerations in mind, even though they do not affect the physical interpretation we present here.
Fig. 9. As in Fig. 8 but EPV on 2300-K surface and horizontal divergence at 0.521 mb averaged over 14–16 January.

Fig. 10. As in Fig. 8 but on 1900-K surface and horizontal divergence at 0.949 mb, averaged over 14–16 June (Southern Hemisphere).

Fig. 11. Mean meridional residual velocity for December. Contour interval is 0.4 m s\(^{-1}\).

ure 10a shows EPV (1900 K) and horizontal divergence (0.95 mb) during mid-June; there is no evidence of extratropical wave activity, but near the equator the meridional gradient of \(\bar{u}\) is large enough to produce a narrow zonally symmetric region of positive EPV values. (Note that, for the Southern Hemisphere, positive EPV regions are potentially inertially unstable.) However, as shown in Fig. 10b, there is no evidence of zonally symmetric divergence/convergence patterns near the equator. This picture does not change appreciably in later months. In mid-July (not shown) the region of positive EPV has undergone some modification, but it remains narrow and unable to support any strong meridional overturning. We will show in the next subsection that the modification of the zonal mean state during the easterly phase of the northern winter is related to the strength of the inertially unstable circulation.

b. Redistribution of momentum in the easterly phase

The development of an inertially unstable region (zonally symmetric or otherwise) gives rise to motion that can modify the meridional distribution of angular momentum. In both cases the mean residual circulation plays an important role. Figure 11 shows the monthly average for December of the transformed Eulerian mean (TEM) meridional component, \(\bar{v}^*\) (see Andrews et al. 1987 for a definition of the TEM circulation). In agreement with theory (Dunkerton 1981), we identify several cells stacked over the equator and superimposed on the seasonal flow from the summer to the winter hemisphere. The mean meridional velocity attains a maximum strength of about 4 m s\(^{-1}\) near 55 km. Figure 12a shows the monthly mean of \(\bar{v}^*\) for January. It is evident that, by this time, we are left with a single, strong meridional jet (maximum \(\bar{v}^* \sim 7\) m s\(^{-1}\)) at about 52 km, between the equator
shown in Fig. 12 is to redistribute angular momentum in the meridional plane and so relieve the conditions of inertial instability. In fact, comparing Figs. 3a and 4a, it is quite evident that the easterly jet has undergone major modifications, reducing the range of altitudes where the meridional gradient of $\vec{u}$ is greater than the local Coriolis parameter. Further, as the easterly nose is advected into the winter hemisphere, the maximum gradient of $\vec{u}$ moves to higher latitudes, where the Coriolis parameter eventually becomes dominant and the flow is no longer inertially unstable.

This picture changes markedly when we look at the easterly phase during Southern Hemisphere winter. The mean meridional residual velocity in June (Fig. 13a) shows a weak jet ($\sim -2 \text{ m s}^{-1}$) just below 50 km and several other jets stacked above and below. The strength of the main jet is about one-half of that of its counterpart in December (see Fig. 11a). Figure 13b shows the mean meridional residual velocity for July. The meridional jet at $\sim 50$ km has been selected at the expense of the other cells, but its magnitude has not increased substantially. At the same time, the associated vertical velocity (not shown) is barely $1 \text{ mm s}^{-1}$ in

and 30°N. The residual mean vertical velocity, $\vec{w}^*$, is shown in Fig. 12b. At the level of the meridional jet, there is vertical convergence just south of the equator and divergence near 30°N. The largest values of $\vec{w}^*$ ($\pm 3 \text{ mm s}^{-1}$) are found in the convergence region. These patterns of the meridional circulation compare well with Hitchman and Leovy's (1986) residual circulation derived from LIMS data; note also that the vertical dimension of the meridional circulation cell implied by the ($\vec{u}^*$, $\vec{w}^*$) fields shown in Fig. 12 is only about 5 km. Thus, the process that gives rise to this meridional circulation could not be accurately simulated in models with vertical resolution much coarser than used here (1 km).\(^3\) The role of the circulation cell

\(^3\) We have verified this point by carrying out a simulation with a coarser resolution ($\Delta z \approx 2.5 \text{ km}$). In this case, the inertially unstable circulation, and hence the strength of the easterly jet, are weakened considerably. Presumably this occurs because, as $\Delta z$ is increased, the model is unable to resolve the faster growing inertially unstable modes that have small vertical scales. See also Boville and Randell (1992, their Fig. 2), where they show the influence of vertical resolution on the evolution of the easterly phase.

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[Figures 12 and 13 related to the text are not included in this text representation.]
both months. This weak meridional circulation (together with the relatively small meridional shear of \( \bar{u} \)) is unable to affect the mean state as strongly as during the Northern Hemisphere winter. As noted before, although there is some increase in the strength of the equatorial easterlies, there is little evidence of the development of a sharp easterly “nose” over the equator as was the case in January.

The residual mean meridional velocity (\( \bar{u}^* \), \( \bar{w}^* \)) has been used to evaluate the TEM zonal momentum budget:

\[
\frac{\partial \bar{u}}{\partial t} = \left( f - (a \cos \phi)^{-1} \frac{\partial}{\partial \phi} (\bar{u} \cos \phi) \right) \bar{v}^* - \frac{\partial \bar{u}}{\partial z} \bar{w}^* + \frac{1}{\rho} \nabla \cdot \mathbf{F} + X \quad (4.3)
\]

where \( a \) is the mean planetary radius, \( \phi \) is the latitude, \( z \) is the vertical coordinate in log-pressure altitude, \( X \) is the diffusion computed by the model, and all other symbols are the standard ones in the TEM formulation (Andrews et al. 1987).

Figure 14 shows the monthly mean of the first two terms on the rhs of (4.3) for January. These terms describe horizontal and vertical advection of zonal mean momentum, respectively. Between 50 and 55 km, within 10° north of the equator, the meridional advection term gives rise to negative zonal mean tendencies (\( \sim -5.0 \) m s\(^{-1}\) day\(^{-1}\)) and positive accelerations poleward. This pattern is related to the zonal mean meridional jet seen in Fig. 12a. The vertical advection term produces positive accelerations (\( \sim 3 \) m s\(^{-1}\) day\(^{-1}\)) above and below 52 km, which are clearly associated with the upwelling/downwelling regions near the stratopause shown in Fig. 12b. The pattern of accelerations shown in Fig. 14 is characteristic of the atmospheric response to inertial instability: the meridional motion advects the easterly jet across the equator, while vertical advection tilts the gradient of \( \bar{u} \) from the meridional to the vertical direction. The pattern also agrees with the Hitchman et al. (1987) suggested mechanism for the redistribution of momentum by inertially unstable cells (see their Fig. 8). The contribution of the Eliassen–Palm flux divergence (Fig. 14c) is relatively small (\( \sim 2.5 \) m s\(^{-1}\) day\(^{-1}\)) equatorward of 10°N, but becomes larger in the subtropics of the winter hemisphere. The pattern shown in Fig. 14c is very similar to that deduced by Hitchman et al. (1987) from LIMS observations and is also consistent with the findings of Hamilton and Mahlman (1988) for the GFDL SKYHI model.

We cannot arrive at the same conclusions when looking at Southern Hemisphere winter. Although the tropical atmosphere between 40 and 50 km supports several inertially unstable cells (not shown), their magnitudes are quite weak and barely reach \( \sim 2 \) m s\(^{-1}\) day\(^{-1}\). Looking at the daily evolution of EPV (not shown), it seems that dissipation of these cells takes place on smaller and smaller horizontal scales on isentropic surfaces: “blobs” of anomalous (positive) EPV are advected poleward and diluted with the higher neg-
ative EPVs at midlatitudes. It is not clear to what extent the suppression of these small-scale cells in the model is physically realistic, although we might expect that structures with very small horizontal scales would also be dissipated rapidly in the real world.

In conclusion, southern winter reveals some of the features observed in northern winter, but the evolution of the equatorial jet is much slower and does not show the remarkable intensity observed in northern winter. The striking difference between the two easterly phases appears to be due ultimately to the lack of strong wave activity in southern winter, which precludes the development of strong gradients of $\tilde{u}$ in the tropics of the winter hemisphere. Similar considerations apply in other model simulations (e.g., Hamilton and Mahlman 1988) where the relative weakness of wave driving during the southern winter compared to northern winter has been emphasized.

c. Westerly phase

We noted in section 3 that the model does not simulate the correct behavior of nitrous oxide in tropical regions near the equinoxes and that this is probably due to the poor representation of the strength and downward propagation of the westerly jet. Figure 15 shows the eddy acceleration (i.e., the divergence of the Eliassen–Palm flux) in the tropics ($30^\circ$S–$30^\circ$N), for March and April. In March above 60 km (Fig. 15a), positive accelerations ($\sim 2.5$ m s$^{-1}$ day$^{-1}$) are asymmetric with respect to the equator, the strongest accelerations appearing in the Southern Hemisphere. In April (Fig. 15b) an equatorially symmetric region of positive acceleration ($1.5$ m s$^{-1}$ day$^{-1}$) shows up at $\sim 50$ km; however, this region does not descend below 50 km and this results in weak downward propagation of the westerly jet. Further analysis of zonal eddy accelerations against the zonal wavenumber (not shown) suggests that the off-equator accelerations near 60 km and $20^\circ$S seen in Fig. 15 originate from resolved gravity waves, whose selective vertical propagation is controlled by the zonal mean state at lower altitudes. Hamilton and Mahlman (1988) have found similar behavior at the stratopause in the GFDL SKYHI model. In particular, they show that the vertical eddy momentum flux at 1 mb during February is shifted toward the Southern Hemisphere, peaking near $15^\circ$–$20^\circ$S (see their Fig. 28). Hamilton and Mahlman note that much of this momentum flux is contributed by gravity waves of zonal scale small enough not to be affected by the equatorial waveguide and that the shift of momentum flux toward the Southern Hemisphere is possibly associated with the location of the tropospheric convectively active region.

Although the eddy acceleration in April shows a prevalent equatorial symmetry, it is still not clear whether it is due to Kelvin waves. To elucidate this point, we analyzed the Eliassen–Palm flux divergence from 1 March to 30 April, using Hayashi’s (1971) technique. Figure 16 shows the result of this analysis for all westerly waves (i.e., the entire spectrum of positive frequencies) from wavenumber 1 to 4; the contribution from easterly waves is not significant at this time on the equator. Note that the eddy forcing attains its maximum values for wavenumber 3 and $\sim 60\%$ of the positive eddy acceleration (on the equator, at $\sim 0.5$ mb) shown in Fig. 15b is accounted for by the first four wavenumbers; contribution from higher zonal wavenumbers wanes rapidly after $k = 4$. The EP flux divergence is nearly symmetric about the equator for all wavenumbers, consistent with dissipation of Kelvin waves. Hitchman and Leovy (1988) suggested from LIMS data that the contribution of large-scale Kelvin waves is in the range 20%–70% of the required westerly accelerations.

To support our contention that the EP flux divergences shown in Fig. 15 are due to Kelvin waves, we have computed the power spectrum of eddy temperature. In Fig. 17 we show the dependence of the spectrum on the height for $k = 1$ and $k = 3$. There is a typical shift of the spectrum toward high frequencies...
as waves propagate upward, related to the absorption of the slower eddy components by the underlying mean flow (see, for example, Garcia and Salby 1987). At the same time, waves with higher zonal wavenumber attain larger amplitude at the upper levels. For wavenumber 3 (Fig. 17b) at \( \sim 0.6 \text{ mb} \) (roughly the height of maximum deposition of positive eddy momentum in Fig. 15b), the temperature harmonic attains a maximum at \( \sim 0.35 \text{ cycles day}^{-1} \) (period \( \sim 2.9 \text{ days} \)), which implies a vertical wavelength, \( \lambda_z \), of \( \sim 17 \text{ km} \) from the Kelvin dispersion relationship and a meridional \( e \)-folding width of \( \sim 14^\circ \) (\( \sim 1500 \text{ km} \)). From model results, we can also estimate the actual \( \lambda_z \) using a coherence analysis. Figure 18 shows the eddy temperature
not coherent at these locations and this time of the year; this is consistent with theoretical expectations for Kelvin waves, and agrees with Boville and Randel's (1992) findings on stratospheric equatorial waves in GCMs.

The importance of Kelvin waves in our calculations differs from the results of Hamilton and Mahlman (1988), who found that in the SKYHI model smaller-scale structures (zonal wavenumber \( k > 10 \)) contribute more to westerly acceleration even at the equator than large-scale waves. Our results, on the other hand, clearly indicate a rapid decrease of the eddy forcing near the equator with increasing horizontal wavenumber beyond \( k = 4 \). Moreover, spectral analysis of the eddy forcing and temperature field also indicates the predominance of Kelvin wave-like signal in the model results. It is not clear why smaller-scale gravity waves play a more important role in the GFDL model than they do in present simulation. The version of the GFDL used by Hamilton and Mahlman had a horizontal resolution of \( 3^\circ \times 3.6^\circ \), compared to the approximately \( 4^\circ \) resolution corresponding to the \( T31 \) truncation in our model. Insofar as moist convection in the tropical troposphere is an important forcing mechanism (Hess

**Fig. 17.** Altitude–frequency plot of eddy temperature power (K²) as a function of latitude and frequency (cycles day⁻¹) for zonal wavenumbers 1 (a) and 3 (b). Contours are logarithmic with three contours per decade. Dark shading denotes values from \( 10^{-1} \) through \( 9.9 \times 10^{-1} \); light shading, values between \( 10^{-2} \) and \( 9.9 \times 10^{-2} \); no shading, values lower than \( 10^{-2} \).

Field (wavenumber 3) in a spectral window encompassing periods between 2 and 3.7 days, with the base point at 1 mb. Contours are plotted only for coherence values \( \geq 0.9 \). At 0.6 mb the amplitude of the wave is \( \sim 1.75 \) K, and the horizontal phase lines yield a vertical wavelength of \( \sim 19 \) km, very close to the theoretical estimate. The meridional scale at \( \sim 0.6 \) mb is close to \( 13^\circ (\sim 1400 \) km), which is also in good agreement with theory. The consistency of the calculated wavelength and period with the dispersion relationship, together with the equatorial symmetry, suggests that we are looking at Kelvin waves. Further analysis, not shown here, indicates that the meridional component of the eddy motion field is several orders of magnitude smaller than the other eddy velocities, and definitely

**Fig. 18.** Eddy temperature field for \( k = 3 \) obtained from coherence analysis of model results over the period 1 March through 30 April, using a spectral window of 2 to 3.7 days. The base point is at 1 mb; horizontal lines indicate phase lag against the base point. The contour interval is 0.25 K. Contours and phases are shown only where the coherence is \( \geq 0.9 \).
et al. 1993), differences in convection parameterizations may account for at least some of the discrepancies between our results and those of Hamilton and Mahlman.

The predominance of equatorial forcing by Kelvin waves in our model results raises the question of how the calculated Kelvin waves compare with observations and, hence, whether the associated eddy forcing is reasonable. We have performed a Hayashi analysis on LIMS data for the period 1 March to 30 April 1979 using the latest version of these data (V5.6.0) with a base point at 10 mb, for wavenumber 1, encompassing all the westerly waves (Fig. 19a). The same analysis has been performed, for comparison purposes, on the model output (Fig. 19b). No significant coherence is achievable in LIMS data at stratosphere level; for this reason the comparison is limited to the middle stratosphere. The overall features are similar (vertical and meridional scales), although calculated temperatures are ~60% larger than observed by LIMS. One may infer from this that the failure to produce a realistic westerly phase is not due to inadequate forcing of large-scale Kelvin waves. This is consistent with Hitchman and Leovy's (1988) analysis of the momentum budget in LIMS data: the authors were not able to balance the observed zonal wind tendencies with estimates of Kelvin wave accelerations. At this point, to our knowledge, it is still an open question whether higher zonal wavenumbers or other processes related to tropospheric forcing may account for a deeper descent of the stratosphere westerly jets. Nevertheless, it is conceivable that Kelvin waves may contribute only in part to the downward propagation of the westerly phase and that the latter may be determined by processes not accurately simulated by the model.

5. Summary and conclusions

In the previous sections we have illustrated the results of a model run with the stratosphere/mesosphere version of the NCAR Community Climate Model (CCM2); the main objective has been to study the dynamical processes that take place during the stratospheric SAO. In order to illustrate transport effects, we calculated the seasonal evolution of nitrous oxide with simplified chemistry. Since the early 1980s, when satellite observations revealed a semiannual signal in the equatorial distribution of tracers (i.e., N₂O, CH₄), there has been growing interest on the physical aspects underlying the formation and evolution of both phases of the SAO. Although some simplified models can account quantitatively for certain of the observed transport features, we find that CCM2, in common with other general circulation models, fails to represent the observed behavior during the westerly phase.

The easterly phase is quite well reproduced in the model: both the magnitude and many details of the evolution compare well with observations. This allowed
us to document extensively the fluid dynamical processes taking place at this time. During Northern Hemisphere winter, the development of the easterly phase is influenced by a sudden stratospheric warming that takes place in December and reverses the zonal mean circulation at stratopause height. Large amplitude planetary Rossby waves propagating into the tropics draw tongues of air with negative Ertel’s potential vorticity (EPV) into the winter hemisphere, thus producing an inertially unstable region. Parcels entering this region of negative EPV are accelerated by inertial instability and redistribute momentum in the meridional plane. At the same time, nitrous oxide is advected in the meridional plane: upwelling and poleward transport are clearly related to the mean circulation that redistributes zonal mean momentum.

Comparison of the foregoing results with the easterly phase that develops during June and July reveals that weak wave activity in southern winter hinders the evolution of zonal mean easterly winds. This happens for two reasons. First, weak wave activity is unable to produce strong easterly accelerations, either directly or through wave-induced advection of zonal mean easterly momentum. Second, the resulting weak meridional gradient of $\mathbf{u}$ is unable to support vigorous inertial instability. We showed that, during June and July, the formation of a thin belt of anomalous EPV is clearly identifiable. However, this inertially unstable region supports only cells of small vertical scale, which are eroded by dissipative processes in the model.

The westerly phase is not well reproduced either in amplitude or in downward propagation. In the model run, the large-scale eddy field near the equator is dominated by Kelvin waves. The accelerations associated with these waves peak at wavenumber 3; for larger wavenumbers (smaller horizontal scales) the contribution to the modification of the mean state is small. Although these findings are consistent with conventional understanding of equatorial dynamics, the failure to simulate realistically the equinoctial dynamics results in a poor representation of the evolution of nitrous oxide in the tropics. A comparison of equatorial wave amplitudes in the present model against satellite observations reveals that calculated wave amplitudes are some 60% larger than implied by observations. Nevertheless, the westerly jet is unrealistically weak and its descent stalls at stratopause level.

These results suggest that the failure to simulate the SAO westerly phase might be due to poor representation of small-scale, vertically propagating gravity waves. Such waves, with wavelengths of 100 to 1000 kilometers, cannot be represented explicitly in the model and must instead be parameterized. The present version of CCM2 uses McFarlane’s (1987) parameterization for unresolved gravity waves; the physical idea used in this scheme is based on Lindzen’s (1981) formulation, but it differs from the latter in that generation of gravity waves is assumed to be due only to flow over orographic features. Although it is generally believed that this is an important mechanism for gravity wave forcing (Mahlman and Umscheid 1984; Boville 1984), other processes can also be effective. In the tropics, where flow over orography can be expected to be less effective than in middle latitudes, it is likely that deep convection plays the major role in the generation of small-scale gravity waves.

Acknowledgments. The authors would like to thank Peter Hess for his comments on the original version of the paper. One of the authors (FS) has been supported by Piano Spaziale Nazionale Italiano. One of the authors (BAB) has been supported by NASA Grants W-16215 and W-18181.

REFERENCES


Hamilton, K., 1982: Rocketsonde observations of the mesospheric


