Dynamics of the Tropical Middle Atmosphere: A Tutorial Review*

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ABSTRACT *The general circulation of the tropical stratosphere, mesosphere and lowermost thermosphere is discussed at a tutorial level. Observations of the quasi-biennial and semiannual oscillations by both in situ and satellite techniques are first reviewed. The basic dynamics controlling the zonal-mean component of the circulation are then discussed. The role of radiative diabatic cooling in constraining the zonal-mean circulation in the middle atmosphere is emphasized. It is shown that the effectiveness of this radiative constraint is reduced at low latitudes, allowing for the sustained mean flow accelerations over long periods of time characteristic of the quasi-biennial and semiannual oscillations in the tropics.*

The current view is that the dominant driving for the equatorial mean flow accelerations seen in the middle atmosphere derives from vertically-propagating waves. This process is illustrated here in its simplest context, i.e. the Plumb (1977) model of the interaction of monochromatic internal gravity waves with the mean flow (based on earlier work of Lindzen and Holton, 1968; Holton and Lindzen, 1972). It is shown that the dynamics illustrated by this simple model can serve as the basis for an explanation of the quasi-biennial oscillation.

The paper then describes some of recent developments in the theory of the quasi-biennial and semiannual oscillations, including aspects related to the interaction between tropics and midlatitudes in the middle atmosphere. The paper concludes with a discussion of the effects of the long period dynamical variations in the tropical circulation on the chemical composition of the stratosphere.

RÉSUMÉ *D'un point de vue pédagogique, on discute, dans cet article, de la circulation générale de la stratosphère, de la mésophère et de la plus basse couche de la thermosphère dans la zone tropicale. D'abord, on passe en revue les observations des oscillations semestrielles et quasi biennales de deux façons, soient par des techniques satellitaires et in situ. On discute ensuite de la dynamique de base qui contrôle la composante zonale moyenne de la circulation. On met l'accent sur les contraintes que le refroidissement radiatif diabatif impose à la circulation zonale moyenne de la moyenne atmosphère. On montre que l'efficacité de cette contrainte radiative est réduite aux basses latitudes. Ceci tient compte des accélérations*

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soutenues de la circulation moyenne sur de longues périodes de temps, propres aux oscillations semestrielles et quasi biennales dans les tropiques.

Il est de notoriété que l'élément dominant entraînant des accélérations de la ciculation moyenne équatoriale dans l'atmosphère moyenne provient de la propagation verticale des ondes. Ce processus est illustré ici dans son plus simple contexte, i.e., le modèle Plumb (1977) de l'interaction des ondes gravitationnelles internes monochromatiques avec la circulation moyenne (basé sur les travaux précédents de Lindzen et Holton, 1968; Holton et Lindzen, 1972). On montre que la dynamique illustrée par ce simple modèle peut servir de base pour l'explication de l'oscillation quasi biennale.

Cet article décrit enfin quelques développements récents de la théorie des oscillations semestrielles et quasi biennales, comprenant les aspects reliés à l'interaction entre les tropiques et les latitudes moyennes dans l'atmosphère moyenne. En conclusion, on discute des effets des variations dynamiques à longue pèriode dans la circulation tropicale sur la composition chimique de la stratosphère.

1 Introduction

There are a number of features that distinguish the dynamics of the tropical stratosphere and mesosphere from the dynamics elsewhere in the atmosphere and hence justify the separate discussion of the tropical circulation that is presented here. The small Coriolis parameter at low latitudes leads to a breakdown of the validity of the geostrophic approximation for the wind and invalidates the quasi-geostrophic theory that is so useful in explaining the large-scale circulation of the extratropical atmosphere. Another consequence of the small Coriolis parameter is that temperature observations (e.g., from satellite radiometers) are of limited use in inferring winds at low latitudes. Thus, it is much harder in the tropics to diagnose the detailed dynamics from available observations, leading to more reliance on indirect theoretical and modelling approaches to study the general circulation.

Perhaps the most distinctive features of the circulation in the tropical middle atmosphere are the large-amplitude, long-period oscillations seen in the zonally-averaged flow. In particular, the winds and temperatures of the equatorial stratosphere undergo a very strong quasi-biennial oscillation (QBO) while the region from the near stratopause to lowermost thermosphere displays a prominent semiannual oscillation (SAO). These are such spectacular phenomena that their study has dominated the field of tropical middle atmospheric dynamics. The present tutorial will begin by reviewing some of the detailed observations of the QBO and SAO (Section 2). This is then followed in Section 3 by a consideration of the role of eddy forcing and diabatic effects in controlling the zonal-mean circulation. This will provide a basic explanation for the existence of long-period fluctuations in the tropical mean flow. Section 4 will introduce a very simple version of a model of wave-mean flow interaction in the tropical middle atmosphere and will show that this can be used as the basis for an understanding of the QBO. Section 5 then briefly reviews extensions of the simple model of the QBO and application of similar ideas to the explanation of the SAO. Section 5 concludes with a consideration of the effects of the QBO and SAO on long-lived trace constituents in the stratosphere. A brief summary is given in Section 6.

This paper does not aim to be a comprehensive review, and many significant papers dealing with the tropical middle atmosphere will not be referenced. The discussion assumes that the reader has a familiarity with the governing equations of meteorology, the notion of mean flow/eddy decomposition, and other basics that are covered, for example, in the first few chapters of Holton (1992). Also assumed is a knowledge of the "transformed-Eulerian" formalism for the mean circulation (e.g., Chapter 12 of Holton, 1992). The intent of this paper is to present both observations and theoretical considerations at an introductory tutorial level, and also to indicate briefly the scope of more recent developments in the subject.

2 Observed features of the tropical circulation

a *Historical Introduction*

The first scientific knowledge of the winds in the tropical stratosphere was obtained from observations of the motion of the aerosol cloud produced by the eruption of Mt. Krakatoa (modern-day Indonesia) in August 1883. The optical phenomena caused by the aerosol were remarkable enough that their first appearance was widely noted. Russell (1888) collected observations from over 30 locations in the tropics and plotted the motion of the edge of the aerosol cloud (see Fig. 1). The regular westward motion is evident, and Russell computed a mean easterly wind velocity between about 31 and 34 m s^{-1}.

The wind in the tropical lower stratosphere was first measured with pilot balloons in 1908 by von Berson at two locations in equatorial East Africa. Over the next three decades these observations were followed by sporadic measurements at a number of tropical locations (see Hamilton (1998a) for a review of these early observations). The results sometimes indicated easterly winds and sometimes westerly winds, a state of affairs reconciled at the time by assuming that there was a narrow ribbon of westerlies (the "Berson westerlies") embedded in the prevailing easterly current revealed by the Krakatoa observations (e.g., Palmer, 1954).

Regular balloon observations of the lower stratospheric winds in the tropics began at a number of stations in the early 1950s. By the end of the decade it was obvious that both the easterly and westerly regimes at any height covered the entire equatorial region, but that easterlies and westerlies alternated with a roughly biennial period (Veryard and Ebdon, 1961; Reed et al., 1961). Initially it was thought that the period of the oscillation might be exactly two years, but as measurements accumulated it soon became clear that the period of oscillation was somewhat irregular and averaged over 2 years. By the mid-1960s the term "quasi-biennial oscillation" (QBO) had been coined to denote this puzzling aspect of the stratospheric circulation.

b *Modern Observations of the QBO*

Figure 2 shows the raw time series of monthly-mean 30-mb zonal wind computed simply by averaging daily balloon observations at Singapore (1.3°N) over a period of 8 years. This illustrates many of the key features of the QBO. Note that the time series is clearly dominated by an alternation between easterly and westerly

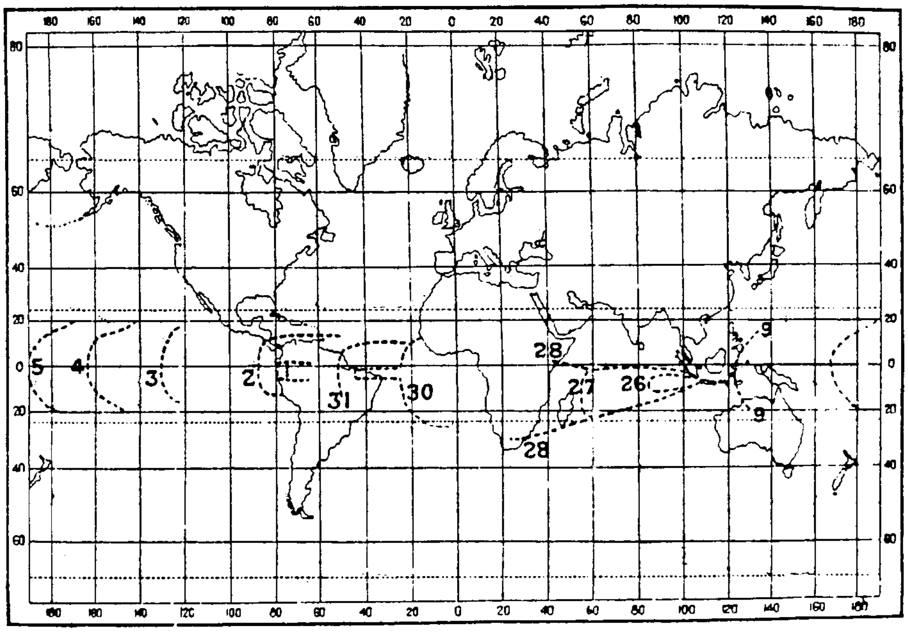

Fig. 1 The spread of the optical phenomena observed after the eruption of Mt. Krakatoa on 26 August 1883. The dotted lines give the western boundary of the region where the phenomena had been observed on successive days, 26 August, 27 August ... 9 September. Reproduced from Russell (1888).

wind regimes roughly every other year. The extremes in the prevailing winds vary from cycle to cycle, but the peak easterly (in the monthly mean) usually exceeds 30 m s^{-1}, while the westerly extreme is usually between 10 and 20 m s^{-1}. The time series has a rough square-wave character with rapid transitions (~2–4 months) between periods of fairly constant prevailing easterlies or westerlies.

The height-time evolution of the monthly-mean wind near the equator for over four decades is shown in Fig. 3 (an updated version of the figures in Naujokat, 1986; Marquardt and Naujokat, 1997). The wind reversals invariably appear first at high levels and then descend. At any level the transition between easterly and westerly regimes is rapid so that the transitions are also associated with strong vertical shear. These near-equatorial data suggest that the mean period of the oscillation is about 28 months and varies between about 20 months and 36 months (Marquardt and Naujokat, 1997). Much of the variability in period is associated with the changes in the length of the easterly phase, particularly above about 50 mb. The maximum amplitude occurs near 30 mb and the amplitude drops off to small, but apparently still detectable (Hitchman et al., 1997) values near the tropopause (~17 km). The dropoff in QBO amplitude above 30 mb is very gradual and the oscillation is still

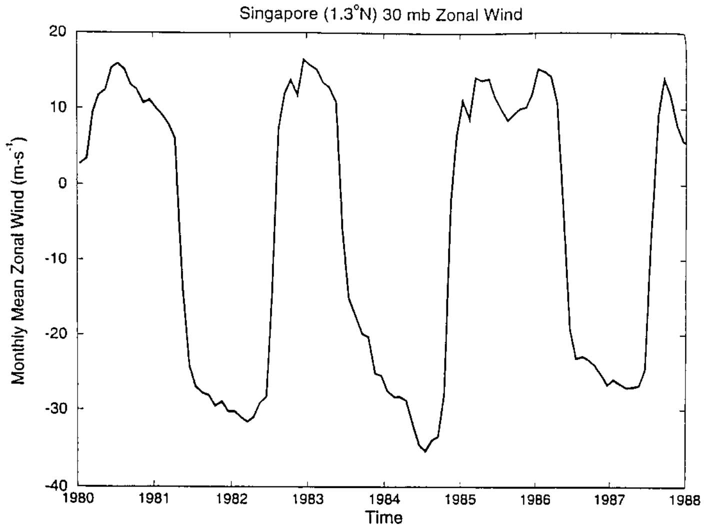

Fig. 2 Time series of the monthly-mean zonal wind measured by balloons at Singapore during the period 1980–88.

very strong at the 10-mb level. At almost all levels and all times the easterly-to-westerly transitions are more rapid than vice versa, and the associated westerly shear zones are considerably more intense than the easterly shear zones.

There is only a very limited network of radiosonde stations near the equator regularly reporting stratospheric winds. An assumption implicit in most observational studies is that the prevailing winds near the equator are essentially zonally-symmetric. Belmont and Dartt (1968) tried to check this with available radiosonde data and concluded that the QBO was indeed very nearly zonally-symmetric up to 50 mb (above this level they felt they had inadequate data for verification). Among the complications in determining the details of the QBO signal is the fact that the annual cycle becomes strong off the equator and itself has considerable geographical variability (Hitchman et al., 1997). Recently the advent of Doppler-radiometer observations of horizontal winds by the High Resolution Doppler Imager (HRDI) instrument on the Upper Atmosphere Research Satellite (e.g., Ortland et al., 1996) has provided another opportunity to examine this issue. Ortland (1997) finds that there may be some modest ($\sim$10 m s^{-1}) zonal asymmetries in the monthly-mean wind at the equator in the westerly phase of the QBO, particularly at and above 10 mb. This issue will be discussed further in Section 5b, but, to first order, the assumption of a zonally-symmetric QBO is appropriate.

Figure 4 shows a determination of the amplitude (solid contours) and phase (dashed contours) of the QBO in zonal wind as a function of height and latitude in the tropical stratosphere. It was derived by Reed (1965b) who fit a simple 26-month harmonic to about 8 years of zonal wind data at a number of low latitude stations.

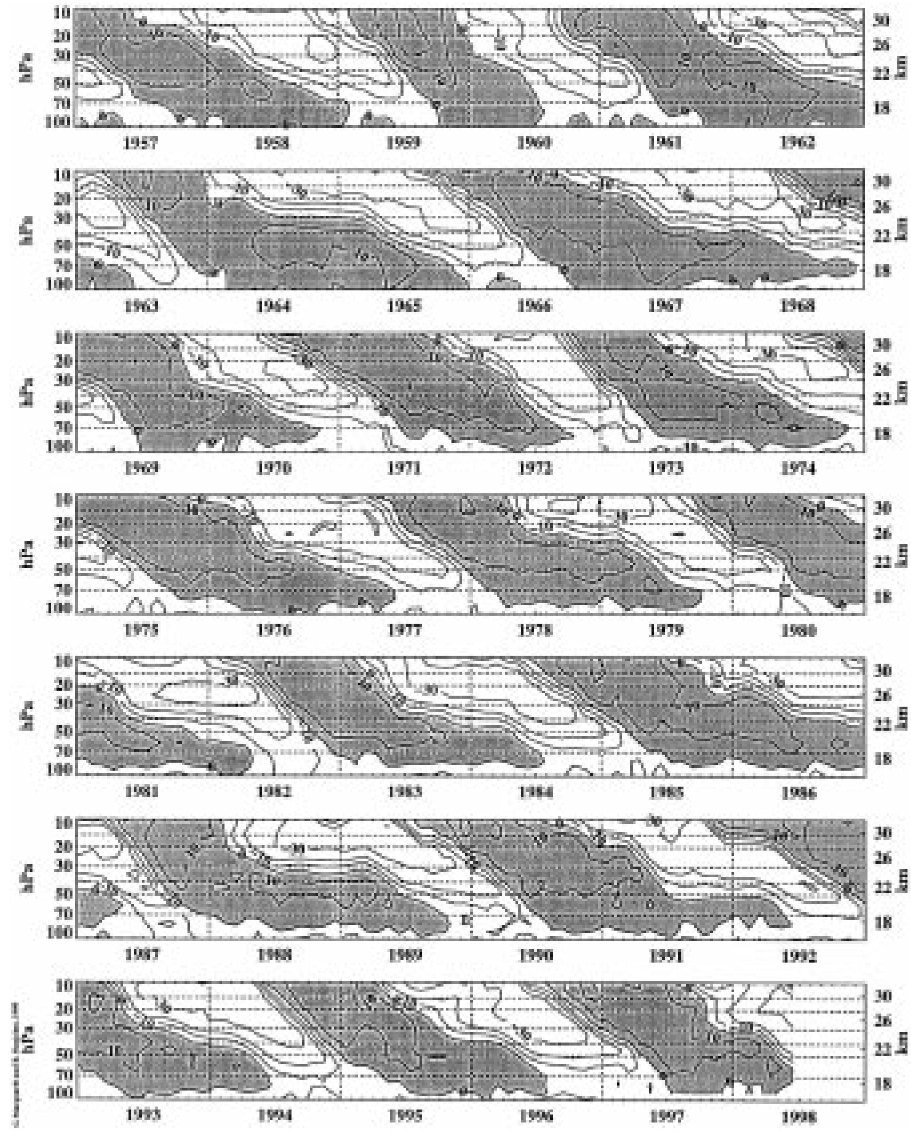

Fig. 3 Time-height section of the monthly-mean wind at stations near the equator. Results represent observations from Canton Island (2.8°S, 171.7°W) during 1957–1967, Gan (0.7°S, 73.1°E) during 1967–1975, and Singapore (1.4°N, 103.9°E) during 1976–1998. Westerly winds are shaded and the contour interval is 10 m s^{-1}. Figure provided by B. Naujokat.

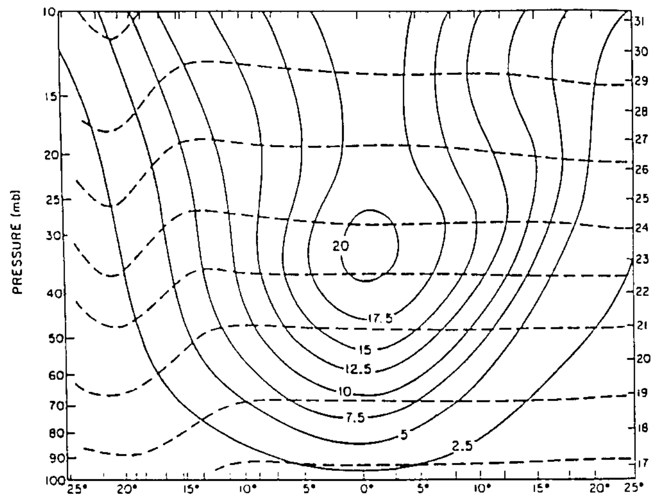

Fig. 4 Latitude-height section of the amplitude and phase of the QBO in zonal wind determined from radiosonde observations. Amplitude contours are solid and the contour interval is 2.5 m s^{-1}. The Northern Hemisphere is shown on the left. Phase contours are dashed and the contour interval is 1 month. The thin tick marks on the axis show the latitude of each of the stations used in the analysis. The scale on the right is a standard height (in km). Adapted from Reed (1965b).

The result for the amplitude shows a peak centred squarely on the equator and a roughly Gaussian dropoff in latitude with an *e*-folding width of between 13 and 15 degrees of latitude. The phase lines are remarkably regular indicating a steady downward propagation of about 2 km month^{-1} and very little phase variation in latitude. Similar results for the observed height-latitude structure of the QBO are given by Belmont et al. (1974).

A QBO in temperature has also been clearly observed. Reed (1962) used balloon observations to show that the QBO in temperature has a peak amplitude of $\sim$2–3°C. In general the usefulness of the geostrophic approximation breaks down at low latitudes, of course, but in fact the *zonal-mean* component of the circulation should be close to geostrophic balance. Reed (1962) showed that indeed, within observational error, the measured zonal-mean QBO temperature variations are in thermal wind balance with the zonal-mean zonal wind.

In recent years observational studies of the general circulation of the tropical lower and middle stratosphere have focused on characterizing some of the details of the QBO. Examples include studies of the evolution of the meridional structure of the zonal wind field through the QBO cycle (Hamilton, 1984, 1985; Dunkerton and Delisi, 1985), and studies of the variability of the QBO period from cycle to cycle (Quiroz, 1981; Dunkerton and Delisi, 1985; Naujokat, 1986; Maruyama and Tsuneoka, 1988). Studies of the influence of the QBO in the tropical upper stratosphere, mesosphere and lower thermosphere have been made using rocketsonde (e.g., Hamilton, 1981) and HRDI satellite observations (Burrage et al., 1996).

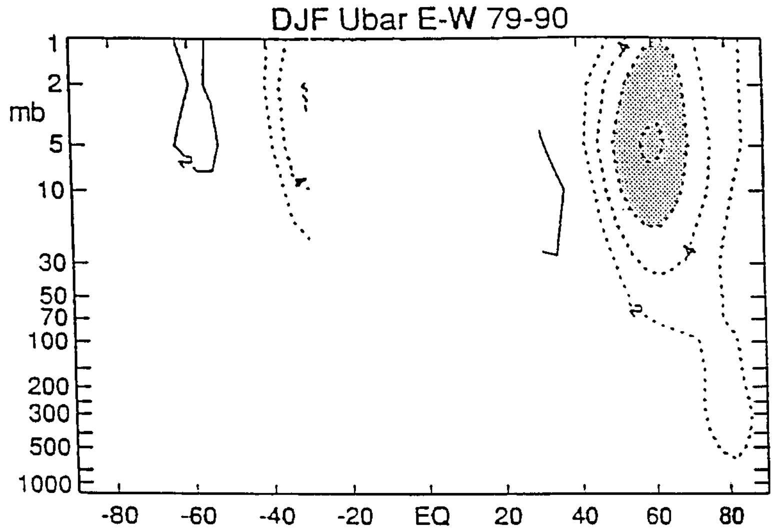

Fig. 5 The December–February zonal-mean zonal wind averaged over winters with easterly equatorial winds at 40 mb minus an average over winters with westerly equatorial winds at 40 mb. The data employed are for 1979–90. The dashed contours denote negative values and the contour interval is 2 m s^{-1}. The zonal winds here are determined geostrophically from global analyses of geopotential height. Results are not plotted at low latitudes where the observed geopotential analyses are of too poor quality for an accurate determination of the zonal wind. Adapted from Baldwin and Dunkerton (1991).

The issue of QBO effects beyond the tropical middle atmosphere has been addressed in a number of studies and is still an issue of current investigation and controversy. The clearest remote effects appear to be in the extratropical northern hemisphere winter stratosphere (e.g., Holton and Tan, 1980, 1982; Dunkerton and Baldwin, 1991; Baldwin and Dunkerton, 1991). Figure 5 shows the December–February zonal-mean zonal wind averaged over easterly phase QBO periods minus that averaged over periods of westerly QBO phase. Here the phase of the QBO used in the compositing is based on the sign of the 40-mb zonal wind measured at Singapore (1.3°N). The tendency for the polar vortex to be somewhat weaker in the easterly QBO phase is evident. Other studies have shown that midwinter sudden warmings of the northern hemisphere (NH) polar stratosphere are significantly more frequent during the easterly QBO phase than in the westerly phase (Dunkerton et al., 1988). The issue of extratropical QBO effects is discussed in more detail in Section 5b.

c *Observations of the Semiannual Oscillation*

The 10-mb level (approximately 30 km) is the usual ceiling for operational balloon soundings, and so knowledge of the wind field at higher levels was first obtained with rocket soundings. Figure 6 (from Reed, 1965a) shows all the zonal wind ob-

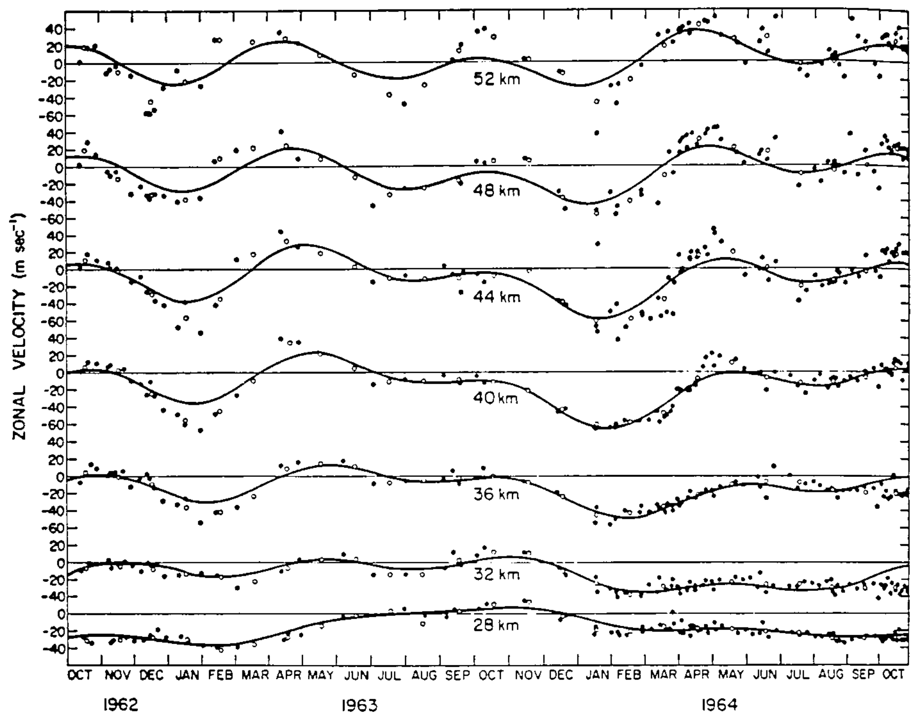

Fig. 6 Zonal wind measurements taken at Ascension island (7.9°S) during the period October 1962 through October 1964. The solid circles show individual measurements and the open circles are monthly means for months when there was more than one measurement available. Reproduced from Reed (1965a).

servations from rocket soundings during a two-year period at Ascension Is. (7.9°S). In the middle stratosphere the QBO appears quite clearly (at least to ∼40 km; see also Hamilton, 1981). However, at upper stratospheric levels the QBO is dominated by a shorter period oscillation. We now know that this is a semiannual oscillation (SAO). Unlike the QBO, the SAO is very clearly phase-locked to the calendar. Near the stratopause the easterly extremes are reached in January and July and the westerly extremes around April and October. Figure 7 shows a climatological annual march of the zonal wind deduced from rocket observations at a number of sites. At the equator the semiannual variation clearly dominates in a thick layer around the stratopause. The vertical structure of the westerly accelerations displays a downward propagation that is similar to the QBO wind reversals. The easterly accelerations are more uniform in height. The stratopause SAO has been seen in rocket observations of zonal winds, rocket observations of temperatures (e.g., Garcia et al., 1997; Dunkerton and Delisi, 1997), satellite radiometer measurements

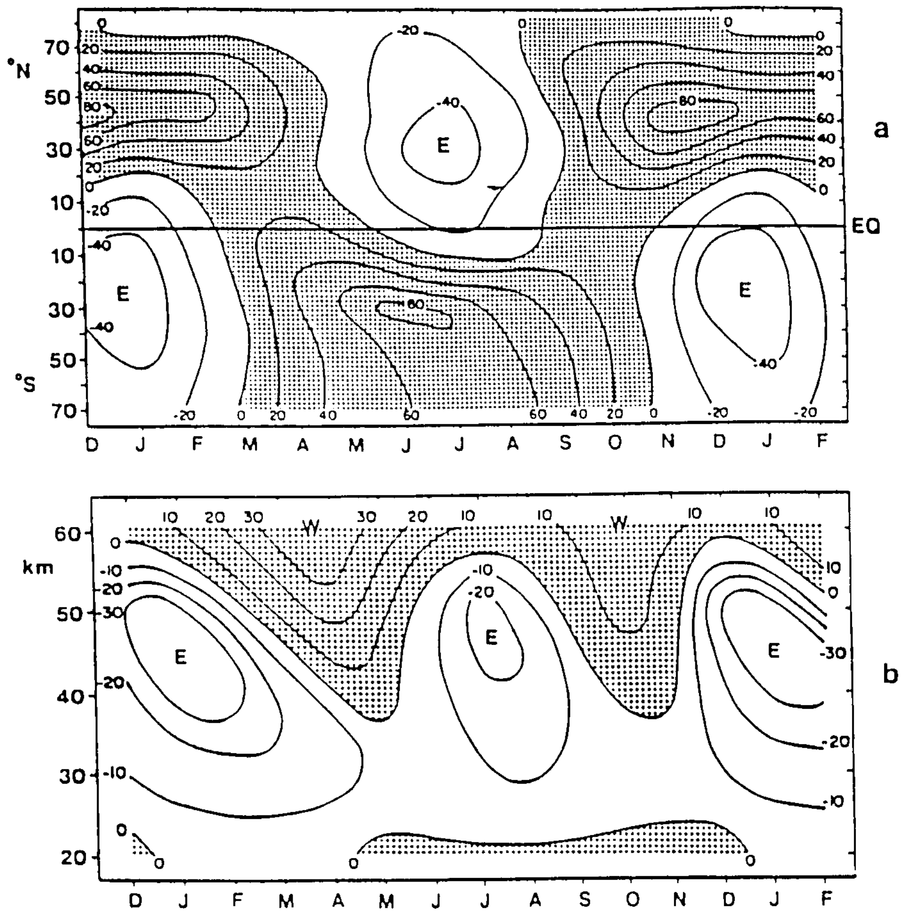

Fig. 7 (top) Latitude-time section of the climatological annual march of zonal-mean zonal winds at 50-km height determined from rocketsonde observations at several stations. Contour interval is 20 m s^{-1} and regions of westerlies are shaded. (bottom) Altitude-time section of the annual march of equatorial zonal wind determined from interpolation of observations at Kwajalein (8.7°N) and Ascension Island (7.9°S). Contour interval is 10 m s^{-1} and regions of westerlies are shaded. Reproduced from Delisi and Dunkerton (1988b) and based on an earlier figure from Belmont et al. (1975).

of temperature (Hitchman and Leovy, 1986; Delisi and Dunkerton, 1988a) and the recent HRDI Doppler radiometer measurements of winds (Burrage et al., 1996).

Using the limited number of rocket soundings available at very high altitudes, Hirota (1978) and Hamilton (1982a) were able to show that the amplitude of the SAO at low latitudes drops in the lower mesosphere, but rises again to a maximum near the mesopause. They also found that the SAO at the mesopause is nearly 180°

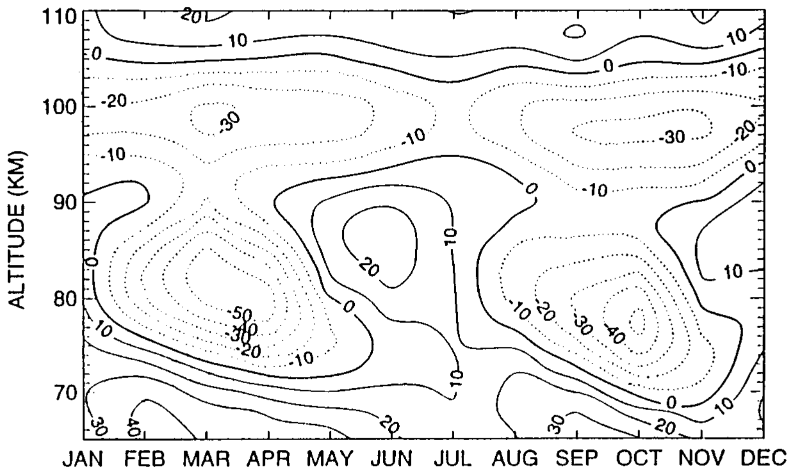

Fig. 8 Annual march of the zonally-averaged equatorial zonal wind measured by the HRDI Doppler radiometer. The contour intervals is 10 m s^{-1} and dashed contours denote easterly winds. Reproduced from Garcia et al. (1997).

out of phase with the stratopause SAO. These observations have been confirmed recently with the wind measurements from the HRDI instrument. Figure 8 (from Garcia et al., 1997) shows the equatorial zonally-averaged zonal wind as a function of height and time of year from the HRDI measurements in the height range 65–110 km. The prominent SAO at the mesopause seen in these data has amplitude comparable to that observed near the stratopause.

3 Basic dynamical considerations – The "Radiative Spring"

While not perfectly monochromatic, the equatorial QBO is the most spectacular example of a quasi-periodic oscillation in the atmosphere that is not astronomically forced (such as the annual and diurnal cycles). The QBO has a timescale of deterministic predictability of months or even years – in strong contrast to meteorologists' usual picture of predictability in the atmosphere. An attempt will be made to explain this surprising feature within the simplest possible contexts. We will first consider the mean flow response to eddy forcing and show why the circulation in the equatorial region is likely to behave quite differently from that in higher latitudes. Then in the next section we will treat the very simple case of two internal gravity waves interacting with the mean flow, to illustrate how long-period oscillatory equatorial mean flow variations may be produced.

We will discuss here what is perhaps the simplest useful model for the mean circulation in the middle atmosphere. In particular, we consider the steady-state response of the mean flow to a specified eddy forcing. The model has the realistic feature that diabatic effects act to restore radiative equilibrium. For simplicity we will model this radiative effect with a Newtonian cooling towards a state with no meridional temperature gradient (it is easy to generalize to the case where the "radiative equilibrium" state has a meridional gradient). Also, for mathematical tractability, we will consider only the Boussinesq equations on an f-plane (periodic boundary conditions in the x-direction will represent the zonal periodicity of the real atmosphere). The base state stability is assumed constant and some non-linear advection terms are ignored (along with the eddy terms often neglected in the primitive equation transformed Eulerian mean formalism). With all these assumptions the governing equations for the zonally-averaged circulation reduce to:

$$-f\bar{v}^* = \bar{F}_{\text{eddy}}$$

$$f\bar{u} = -\frac{1}{\rho_0}\frac{\partial \bar{P}'}{\partial y}$$

$$\frac{\partial \bar{P}'}{\partial z} = -g\bar{\rho}'$$

$$\bar{w}^*\bar{\rho}_z = -\alpha\bar{\rho}'$$

$$\frac{\partial \bar{v}^*}{\partial y} + \frac{\partial \bar{w}^*}{\partial z} = 0$$

where

$$\bar{\rho} = \rho_0 + z\bar{\rho}_z + \bar{\rho}'$$

and

$$\bar{P} = \bar{P}_0 + \bar{P}'.$$

The thermodynamic equation can be rewritten as

$$N^2\bar{w}^* = \frac{\alpha g}{\rho_0}\bar{\rho}'$$

where

$$N^2 = -\frac{g\bar{\rho}_z}{\rho_0}.$$

Here, f is the Coriolis parameter (a constant), $\bar{u}$ is the zonally-averaged zonal velocity, $\bar{v}^*$ and $\bar{w}^*$ are the meridional and vertical components of the resident mean meridional circulation, N is the buoyancy frequency, $\bar{P}$ is the zonal-mean pressure and $\bar{\rho}$ is the zonal-mean density. The density is the sum of a background state and the perturbation $\bar{\rho}'$. The background state density is the sum of the constant

Boussinesq mean density, $\bar{\rho}_0$, and a linear decrease with height. The background pressure $\bar{P}_0$ is assumed to be in hydrostatic balance with the background density. The body force per unit mass associated with the Eliassen-Palm flux divergence of the eddies is $\bar{F}_{\text{eddy}}$, and the Newtonian cooling has a relaxation coefficient α.

These linear, constant coefficient equations are easily solved if we write the solution in terms of Fourier harmonics. Consider one harmonic:

$$\bar{F}_{\text{eddy}}, \bar{P}', \bar{\rho}', \bar{u}, \bar{v}^*, \bar{w}^* \propto e^{i(ky+mz)}.$$

Substitution into the equations of motion gives

$$\bar{v}^* = -\frac{\bar{F}_{\text{eddy}}}{f} \tag{1}$$

$$f\bar{u} = -\frac{ik\bar{P}}{\rho_0} \tag{2}$$

$$im\bar{P} = -g\bar{\rho}' \tag{3}$$

$$\frac{\rho_0 N^2 \bar{w}^*}{\alpha} = g\bar{\rho}' \tag{4}$$

$$ik\bar{v}^* + im\bar{w}^* = 0. \tag{5}$$

Equations (3) and (4) can be combined to give

$$\frac{\rho_0 N^2 \bar{w}^*}{\alpha} = -im\bar{P}$$

or

$$im\bar{w}^* = \frac{\alpha}{\rho_0 N^2} m^2 \bar{P}.$$

Substituting this into (5) and using (1) to eliminate $\bar{w}^*$ and $\bar{v}^*$ gives

$$-ik\frac{\bar{F}_{\text{eddy}}}{f} + \frac{\alpha m^2}{\rho_0 N^2}\bar{P} = 0$$

or

$$\bar{P} = \frac{ik\rho_0 N^2}{f\alpha m^2}\bar{F}_{\text{eddy}}.$$

From equation (2) then

$$\bar{u} = -\frac{ik}{f\rho_0}\bar{P} = \frac{k^2 N^2}{f^2 m^2 \alpha}\bar{F}_{\text{eddy}}. \tag{6}$$

The response of $\bar{u}$ to the forcing is proportional to $1/f^2$, which implies that, for given spatial scales of the forcing, the constraint of the "radiative spring" on the mean flow is very much less effective near the equator than at higher latitudes. One factor of $1/f$ arises because the temperature perturbation associated with a given wind perturbation scales as f, and there is an additional factor of $1/f$ because the induced meridional circulation affects the zonal momentum equation through the $f\bar{v}^*$ term.

Since m^2 in equation (6) is equivalent to a second derivative in z on the zonal velocity, this acts as a vertical diffusion, i.e.

$$\frac{\partial^2 \bar{u}}{\partial z^2} = -\frac{k^2 N^2}{f^2 \alpha} \bar{F}_{\text{eddy}}$$

or in general, with a non-resting background state,

$$\frac{\partial^2 (\bar{u} - \bar{u}_{\text{rad}})}{\partial z^2} = -\frac{k^2 N^2}{f^2 \alpha} \bar{F}_{\text{eddy}}$$

where $\bar{u}_{\text{rad}}$ is the zonal mean wind that is in thermal wind balance with the radiative equilibrium temperature structure. The radiative spring acts as an "effective vertical diffusion" diffusing away any deviations from the radiative equilibrium state.

This process has a characteristic timescale $\bar{u}/F_{\text{eddy}}$ or $(k^2N^2/f^2m^2\alpha)$. At sufficiently small f this will become very long, and it is unlikely that the steady-state solution has much relevance. Near the equator the dominant balance is presumably between $\partial\bar{u}/\partial t$ and $\bar{F}_{\text{eddy}}$. Thus at low latitudes the mean flow can respond to eddy driving with sustained accelerations. The momentum in the zonal-mean flow at any time has responded to the accumulated effect of the eddy driving and serves as a "memory" for the atmospheric circulation.

This treatment of the mean flow effects in the tropical middle atmosphere has been highly idealized, of course. Some aspects involved in more realistic models are discussed by Dunkerton (1991). For example, Dunkerton notes that in a compressible atmosphere the effective diffusion does not act symmetrically in the upward and downward directions as it does in the simple Boussinesq case considered here. The work of Haynes et al. (1991) is also relevant. This is a more general discussion of the response of the zonal-mean circulation to imposed forces, including such aspects as the nonlinear advection of the relative mean momentum by the mean meridional circulation, and the spherical geometry. Their results also demonstrate the difficulty in finding a steady-state mean flow response at low latitudes that would be relevant to the real atmosphere.

4 Simple model of wave-driving of the mean flow

a *Introduction*

The considerations outlined above explain the propensity of the zonal-mean tropical middle atmospheric circulation to undergo slow variations. The basic explanation for why the variations can take the form of the quasi-regular QBO wind reversals was provided by Lindzen and Holton (1968) and Holton and Lindzen (1972). They showed that the eddy driving of the mean flow from a combination of eastward and westward travelling, vertically propagating, internal gravity waves or equatorial planetary waves could produce a long-period mean flow oscillation with many of the observed characteristics of the QBO. This section will discuss simple models that demonstrate how waves interacting with the mean flow may give rise to the long-period QBO phenomenon. Neither the details of the wave dynamics in the equatorial region (in particular the role of planetary rotation), nor the tropospheric forcing mechanisms for these waves will be considered here. Reviews of the standard theory and observations of equatorial waves and gravity waves may be found in Wallace (1973) or Andrews et al. (1987).

Plumb (1977) presented a slightly simplified version of the Holton-Lindzen model which simply considers the mean flow accelerations induced by two internal gravity waves. We will describe the Plumb model in some detail here. The first step will be a brief discussion of the properties of linear internal gravity waves.

b *Two-Dimensional Internal Gravity Waves*

We consider the case of purely 2-D, non-rotating flow in a domain with periodic boundary conditions in the x-direction. For simplicity we will treat the problem within the Boussinesq and hydrostatic approximations. Consider then the equations linearized about a time-independent, zonal-mean state characterized by mean flow $\bar{u}(z)$, and mean static stability $N(z)$.

$$\frac{\partial u'}{\partial t} + \bar{u}\frac{\partial u'}{\partial x} + w'\frac{\partial \bar{u}}{\partial z} = -\frac{1}{\rho_0}\frac{\partial P'}{\partial x} - \alpha u' \tag{7}$$

$$\frac{\partial \rho'}{\partial t} + \bar{u}\frac{\partial \rho'}{\partial x} - \frac{\rho_0 N^2}{g} w' = -\alpha \rho' \tag{8}$$

$$\frac{\partial P'}{\partial z} = -g\rho' \tag{9}$$

$$\frac{\partial u'}{\partial x} + \frac{\partial w'}{\partial z} = 0. \tag{10}$$

Where we have allowed for the possibility of wave dissipation by inclusion of the linear relaxations in the momentum and thermodynamic equations. Now assume travelling wave solutions

$$u' = U(z)\mathrm{e}^{i(kx-\omega t)}$$

$$w' = W(z)\mathrm{e}^{i(kx-\omega t)}$$

etc.

Substitution into equations (7)–(10), and assuming that $\alpha = 0$, gives

$$\frac{\partial^2 W}{\partial z^2} + \left(\frac{N^2 k^2}{(\omega - k\bar{u})^2}\right) W + \left(\frac{k}{(\omega - k\bar{u})}\right) \frac{\partial^2 \bar{u}}{\partial z^2} W = 0.$$

If we now assume that the background state does not vary with height so that $N^2(z)$ and $\bar{u}(z)$ are constant, then

$$\frac{\partial^2 W}{\partial z^2} + \left(\frac{k}{\omega - k\bar{u}}\right)^2 N^2 W = 0. \tag{11}$$

The solution to this equation is of the form $W \propto e^{imz}$ and the resulting dispersion relation is

$$m^2 = \frac{N^2 k^2}{(\omega - \bar{u}k)^2} = \frac{N^2}{\hat{c}^2}$$

where $\hat{c}$ is the intrinsic (Doppler-shifted) horizontal phase speed for the wave. Waves with high intrinsic horizontal phase speed have small m, or long vertical wavelength. Also, near critical levels, where the horizontal phase speed matches the background flow, the vertical wavelength should become very small.

The vertical group velocity is given by

$$c_{gz} = \frac{\partial \omega}{\partial m} = \frac{\partial(\omega - \bar{u}k)}{\partial m},$$

Using as our convention that the intrinsic frequency of the waves is positive, the appropriate root must be taken:

$$\omega - \bar{u}k = \pm \frac{Nk}{m}$$

where the signs of k and m determine the sense of phase propagation. This gives a vertical group velocity

$$c_{gz} = \mp \frac{Nk}{m^2} = \mp \frac{(\omega - \bar{u}k)^2}{Nk}.$$

This implies that for $m < 0$ (i.e., for waves with downward phase propagation), $c_{gz} > 0$ and so there is upward group propagation. Also, higher intrinsic horizontal phase speed waves have larger vertical group velocity. As $\omega \longrightarrow \bar{u}k$ the vertical group velocity $c_{gz} \longrightarrow 0$.

One can also easily show for $c_{gz} > 0$, that $\overline{u'w'} > 0$, if $k > 0$, and that $\overline{u'w'} < 0$, if $k < 0$. That is, for upward-propagating wave energy with eastward (westward) intrinsic phase speed the eddy momentum flux is positive (negative).

For the general case when the mean state varies with height, the full equation for $W(z)$ is needed. However, if we assume that the mean state varies sufficiently slowly with height relative to the wave phase, i.e.,

$$\frac{k}{|\omega - \bar{u}k|}\frac{\partial^2 \bar{u}}{\partial z^2} \ll m^2$$

then the dominant balance is just that described by equation (11). Again if $\bar{u}$ and N^2 vary only slowly with z over a vertical scale m^{-1}, then we can use the WKB solution

$$W(z) = W(z_0)\left(\frac{kN}{\omega - \bar{u}k}\right)^{1/2} \exp\left(\int_{z_0}^{z} -\frac{ikN}{\omega - \bar{u}k}dz\right) \tag{12}$$

where we have considered a wave with upward group velocity. In this case, one can go back to the governing equations and – assuming a slowly-varying mean state – show that $\overline{u'w'}(z) = \text{constant}$, which is consistent with the Eliassen-Palm theorem (i.e., that for steady, unforced and undissipated linear gravity waves the divergence of the Reynolds stress should be zero; see Eliassen and Palm, 1961).

Including the dissipation terms $-\alpha u'$ and $-\alpha\rho'$, in the above derivation simply replaces ω by $\omega + i\alpha$. In this case, the integral in equation (12) becomes

$$\int_{z_0}^{z} -\frac{kN}{\omega + i\alpha - \bar{u}k}dz$$

$$= \int_{z_0}^{z} -\frac{ikN}{(\omega - \bar{u}k)(1 + \frac{i\alpha}{\omega - \bar{u}k})}dz.$$

Assuming weak dissipation (i.e., that $\alpha \ll (\omega - \bar{u}k)$) and using a binomial expansion, the integral becomes

$$\approx \int_{z_0}^{z}\left(-\frac{ikN}{(\omega - \bar{u}k)} + \frac{i^2\alpha kN}{(\omega - \bar{u}k)^2}\right)dz$$

$$= \int_{z_0}^{z} -\frac{ikN}{(\omega - \bar{u}k)}dz - \int_{z_0}^{z}\frac{\alpha kN}{(\omega - \bar{u}k)^2}dz$$

so that the solution has both an oscillating character and an overall amplitude modulation with an exponential decay with height. In this case, the Reynolds stress is no longer constant but

$$\overline{u'w'}(z) \sim \exp\left(\int_{z_0}^{z} -\frac{2\alpha kN}{(\omega - \bar{u}k)^2}dz\right) = \exp\left(\int_{z_0}^{z} -\frac{2\alpha}{c_{gz}}dz\right).$$

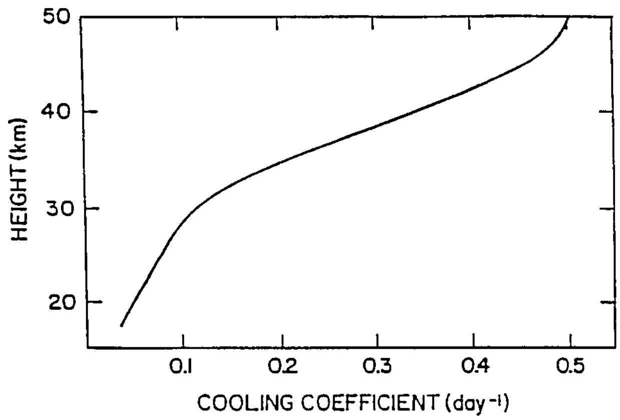

Fig. 9 A reasonable representation of the damping rate of temperature perturbations due to radiative effects in the stratosphere, appropriate for temperature perturbations with long vertical scales. Reproduced from Hamilton (1982b) and based on the radiative-photochemical results of Blake and Lindzen (1973).

Therefore, the slower the vertical group velocity, the longer the damping can act on the waves, the more rapid the decay of the Reynolds stress, and the stronger the local mean flow driving. Note that in general α can be a function of z. If the damping term is omitted from the momentum equation but retained in the thermodynamic equation, it is easy to show that

$$\overline{u'w'}(z) = \overline{u'w'}(z_0)\exp\left(\int_{z_0}^{z} -\frac{\alpha k N}{(\omega - \bar{u}k)^2} dz\right). \tag{13}$$

This is the case considered by Plumb (1977) in his original paper, and it is possibly a reasonable treatment of the stratosphere in the sense that large-scale motions may be regarded as rather inviscid, while the effects of radiative transfer will definitely act to damp the temperature perturbations associated with wave motions. The effectiveness of radiative transfer in damping waves is known to depend on the vertical wavelength, with short wavelengths more strongly damped (e.g., Fels, 1982). In the limit of very long vertical wavelength perturbations the timescale for the radiative damping can be estimated to range from $\sim$100 days in the lowermost stratosphere to $\sim$5 days near the stratopause. Figure 9 shows a typical profile used for the thermal damping rate in the stratosphere. The dissipation rates given in this figure are for long vertical wavelengths and thus are something of an underestimate of those appropriate for the waves (with wavelengths roughly 3–15 km) thought to be important in driving the mean flow accelerations in the tropical stratosphere (e.g., Wallace, 1973).

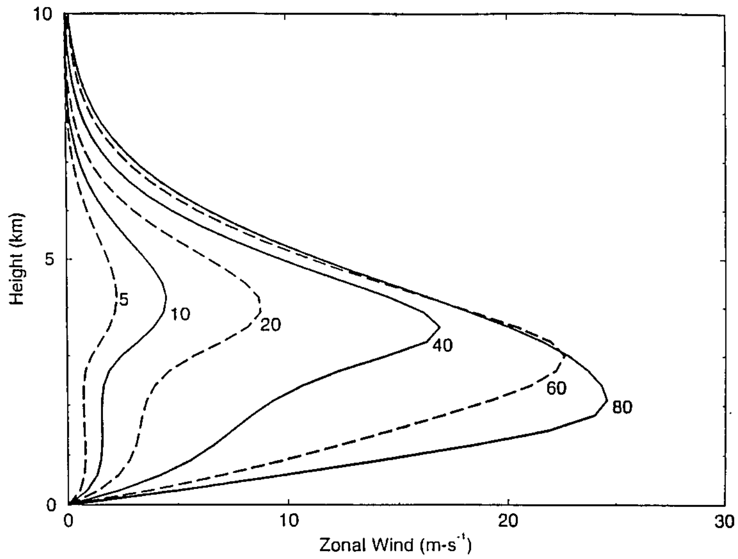

Fig. 10 The mean zonal wind in a simple Boussinesq one-wave Plumb model plotted at 5, 10, 20, 40, 60 and 80 days of integration starting from zero mean wind. The parameters employed are $F(z=0) = 0.02$ m^2 s^{-2}, $k = 2\pi/4 \times 10^7$m^{-1}, $k/\omega = 30$ m s^{-1}, mean flow diffusivity 0.3 m^2 s^{-1}, $N = 0.02$ s^{-1}, $\alpha = 0.01$ day^{-1} for $z < 3$ km and $\alpha = 0.01\ (1 + 3.3z)$ day^{-1} for $z > 3$ km.

c *One-Wave Plumb Model*

Consider a system with a zonal-mean flow $\bar{u}(z, t)$ and one internal gravity wave, with wavenumber k and frequency ω, forced at the lower boundary. Assume that there is a constant effective vertical diffusion K acting on the mean flow. Then the zonal-mean zonal momentum equation is

$$\frac{\partial \bar{u}}{\partial t} = -\frac{\partial}{\partial z}(\overline{u'w'}) + K\frac{\partial^2 \bar{u}}{\partial z^2}.$$

We follow the standard treatment of wave-mean flow interaction problems in assuming that the eddy fluxes can be adequately computed from linear wave theory. In addition we assume that the vertical wavelength of the wave is sufficiently short that the WKB scaling discussed earlier applies, and that the mean flow changes so slowly that the steady wave solution can be used. In this case we saw that the wave momentum flux is given by equation (13). In that equation z_0 is the height of the lower boundary, which can be regarded as corresponding to the tropopause. All the wave parameters, including the momentum flux at z_0, must be specified, along with $N(z)$ and $\alpha(z)$. Using boundary conditions on the mean flow such that $\bar{u}(z_0) = 0$ and $\frac{\partial \bar{u}}{\partial z} = 0$ at the top boundary, one can integrate numerically very easily, calculating the wave flux, then using the flux divergence as forcing in the mean flow equation for some brief time step, then recalculating the wave fluxes with the updated mean flow, etc.

Figure 10 shows the mean flow evolution in an integration of this model. The imposed gravity wave has a phase speed of +30 m s^{-1} and a period of about 15 days. With an initial zonal wind $\bar{u}(t_0) = 0$, and with radiative damping that increases the height, the initial acceleration of the zonal wind is strongest higher up. As the wind is accelerated, however, the intrinsic phase speed of the wave decreases, leading to increased dissipation at lower levels. Eventually the wave is almost choked off from reaching the upper levels and the accelerations there drop almost to zero. The net effect is to produce a mean flow jet with maximum that descends with time, very much like the wind regimes in the QBO. After 80 days the model is approaching a steady-state with the maximum mean flow comparable to the wave phase speed. At this point the wave is very strongly absorbed in the lowest few kilometres of the domain and the eddy forcing is counteracted by the mean flow diffusion of momentum down into the lower boundary.

The model described here is easily generalized to include the effects of compressibility (e.g., Plumb, 1977). The WKB solution for the Reynolds stress associated with the vertically propagating gravity wave becomes

$$\rho_0(z)\overline{u'w'}(z) = \rho_0(z_0)\overline{u'w'}(z_0)\exp\left(\int_{z_0}^{z} -\frac{\alpha k N}{(\omega - \bar{u}k)^2}dz\right)$$

where $\rho_0(z)$ is the mean density. The mean flow equation is generalized to

$$\frac{\partial \bar{u}}{\partial t} = -\frac{1}{\rho_0}\frac{\partial}{\partial z}(\rho_0\overline{u'w'}) + \frac{K}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0\frac{\partial \bar{u}}{\partial z}\right).$$

The reduction of mean density with height leads to more effective wave driving of the mean flow accelerations at higher levels. In the compressible model one can obtain a very pronounced downward progression of the jet maximum even with a wave dissipation rate, α, that is constant with height.

d *Two-Wave Plumb Model*

The example above showed that the interaction of a single dissipating internal wave with the mean flow can produce monotonic mean flow accelerations that are self-limiting. We now show that adding a second wave with phase speed in the opposite direction allows the mean flow to oscillate.

Consider a system with two waves such that $k_1 = -k_2$ and $\omega_1 = \omega_2$. The governing equation for the mean flow is now

$$\frac{\partial \bar{u}}{\partial t} = -\frac{\partial}{\partial z}(\overline{u'w'}) + K\frac{\partial^2 \bar{u}}{\partial z^2}$$

where the total Reynolds stress is

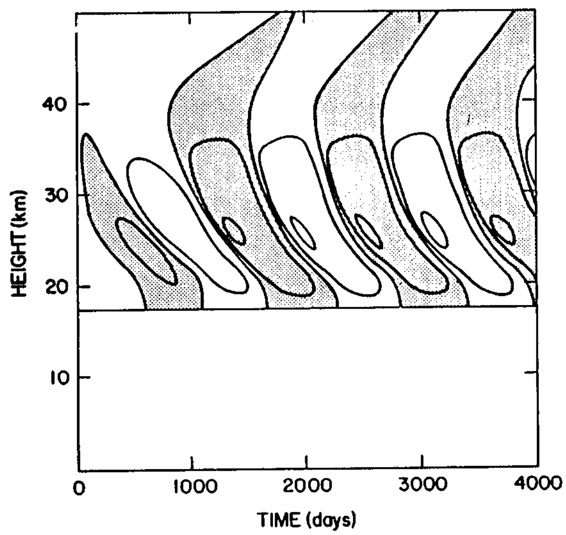

Fig. 11 Height-time evolution of the mean flow in a two-wave Plumb model. The model employed is a version that takes into account compressibility. The model is forced with two waves with phase +25 m s^{-1} and -25 m s^{-1} and the wave fluxes are specified at the lower boundary located at 17-km height. The contour interval is 10 m s^{-1} and regions of negative wind are shaded. Reproduced from Hamilton (1982b), where further details can be found.

$$\overline{u'w'}(z) =$$

$$\overline{u'w'}_1(z_0)\exp\left(\int_{z_0}^{z} -\frac{\alpha k_1 N}{(\omega_1 - \bar{u}k_1)^2}dz\right) + \overline{u'w'}_2(z_0)\exp\left(\int_{z_0}^{z} -\frac{\alpha k_2 N}{(\omega_2 - \bar{u}k_2)^2}dz\right).$$

For simplicity we suppose the waves have equal amplitudes at the lower boundary, so that

$$\overline{u'w'}_1(z_0) = -\overline{u'w'}_2(z_0).$$

Figure 11 shows the results from an integration of a 2-wave Plumb model using waves with phase speeds of +25 m s^{-1} and -25 m s^{-1}. The result shown is for a compressible version and other details can be found in Hamilton (1982b). The mean flow evolution displays an oscillation between positive and negative winds with a period of slightly over 1000 days, downward propagation of the wind reversals, and the development of intense shear zones. The basic resemblance to the observed QBO is striking and represents a nice demonstration that the downward propagation of the jets and shear zones can result purely from interactions with upward propagating waves, i.e. the basic forcing for the stratospheric QBO may be regarded as residing in the troposphere where the waves are excited.

The basic reason for the oscillatory behaviour in Fig. 11 is easy to understand. Once the flow develops the strong winds in, say, the $+x$ direction within the lower part of the domain, the propagation to upper levels of the wave with negative phase speed is enhanced. When this wave is dissipated at these upper levels a negative jet is formed. This is a kind of "shadowing effect" in which the effects of a spectrum of vertically-propagating waves tends to produce a mean flow anomaly aloft of opposite sign to that at lower levels. The negative jet produced aloft can descend just as in the 1-wave case. Once the negative jet has descended low enough the effects of the mean flow viscosity destroy the positive jet below, allowing the negative mean flow region to descend right to the lower boundary. The process can then continue with a positive jet produced aloft, and so on.

The behaviour of the simple 2-wave Plumb model described here has been reproduced in a laboratory experiment by Plumb and McEwan (1978). They put salt-stratified water into an annulus with a flexible membrane at the bottom. The membrane was then forced up and down in order to excite a standing oscillation at the bottom. Since the standing oscillation can be regarded as the sum of two waves with equal but opposite phase velocities, the laboratory experiment is almost a direct analogue of the 2-wave Plumb model (with the annulus geometry supplying the periodic boundary conditions). The results were quite dramatic, as the waves produced a much longer period oscillation in the annulus-averaged circulation with downward-propagating flow reversals.

5 Further developments

a *Issues in the Theory of the QBO*

The Plumb model, while very elegant, is obviously an extreme simplification of reality. In this section more recent work that has aimed at understanding some of the real world complications in the QBO will be briefly introduced.

Perhaps the most fundamental problem impeding our understanding of the dynamics of the QBO is the inability to diagnose accurately from observations the actual contributions of different parts of the wave spectrum to the mean flow accelerations in the tropics. A direct calculation of the vertical Reynolds stress requires knowledge of the vertical wind, which cannot be directly measured on a regular basis. Even the planetary-scale variations of the horizontal wind in the tropical stratosphere are hard to determine, given the sparse radiosonde network and the difficulty in obtaining reliable estimates of the wind from satellite temperature observations. This lack of a firm observational quantitative understanding affected the development of QBO theory from the beginning. In fact, the original model of Lindzen and Holton (1968) supposed that the forcing of the mean flow was due to gravity waves with a continuous spectrum of horizontal phase speeds, while the same authors four years later (Holton and Lindzen, 1972) showed that similar results could be obtained by including only one eastward-propagating and one westward-propagating large-scale equatorial planetary wave. Even today we cannot be certain of the relative contributions of planetary-scale waves and smaller-scale gravity waves.

The 1972 Holton and Lindzen paper presents a theory very similar to that of Plumb (1977), except that the expressions for vertical group velocity of the equatorial Kelvin and Rossby-gravity waves are used for the eastward and the westward-propagating waves, respectively (see Lindzen, 1971, 1972). The choice of wave parameters in the model was roughly based on available observations of planetary-scale equatorial waves (see Wallace, 1973, for a review). The results were quite encouraging, since with "reasonable" parameters the model was found to simulate an equatorial QBO with roughly the correct period, amplitude and vertical structure.

Much of the theoretical work on the QBO over the last 25 years has involved generalizations of the Holton-Lindzen (1972) model. Dunkerton (1981) considered the effects of wave transience in the model (i.e., he relaxed the assumption of steadiness in the calculation of the wave fluxes). Saravanan (1990) generalized the Holton-Lindzen model to include a large number of waves. Dunkerton (1983) added an additional easterly forcing to the Holton-Lindzen model which was meant to account for the effects of mean flow forcing from quasi-stationary planetary waves forced by topography in the extratropics.

A major limitation of the Holton-Lindzen and Plumb QBO models is their restriction to one spatial dimension, i.e., they solve for the height and time dependence of a mean flow meant to represent the flow averaged over some latitude band around the equator. Plumb and Bell (1982) and Takahashi and Boville (1992) constructed numerical models of the QBO forced by two waves, but included a treatment of the waves and mean flow in both height and latitude. Also included in the Plumb and Bell work was the effect of the vertical advection by the mean meridional circulation produced by mean flow radiative effects. They noted that the presence of westerly shear on the equator is associated with a warm anomaly at the equator and hence anomalous diabatic cooling, leading to mean sinking (or at least anomalously weak rising motion). This aspect of the QBO had been qualitatively discussed earlier by Reed (1965b), and Reed's schematic diagram of this effect is reproduced here as Fig. 12. Plumb and Bell noted that the vertical advection associated with this component of the mean meridional circulation should act to intensify the accelerations as westerly jets descend, since the downward advection of the westerly momentum from above adds to the local westerly wave-driving. Conversely, the mean advection effect should act to weaken the easterly accelerations. Plumb and Bell proposed this as the mechanism to explain the observed asymmetry between the strengths of the easterly and westerly shear zones in the QBO.

Recently the possibility that a broad spectrum of gravity waves might be responsible for driving the QBO has been revived (e.g., Dunkerton, 1997). In one interesting development, Alexander and Holton (1997) conducted high-resolution limited-area explicit numerical simulations of a tropical squall line and found that

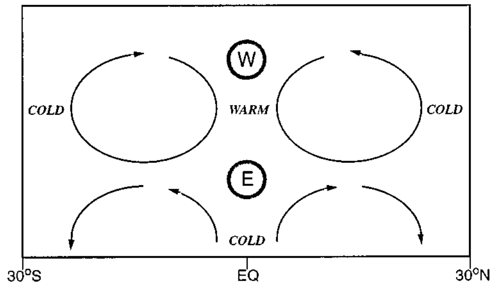

Fig. 12 Schematic view of how the QBO may affect the mean meridional circulation in the tropical stratosphere. The "E" and "W" refer to the maximum easterly and westerly zonal winds, and the peak warm and cold QBO anomalies at the equator are also marked. Adapted from Reed (1965b).

the momentum fluxes associated with the gravity waves forced by such storms could well be significant for the dynamics of the QBO.

Another concern has been the simulation of the QBO within comprehensive general circulation models (GCMs; see Hamilton, 1996, for a review of the application of GCMs to the middle atmosphere). Historically GCMs have usually produced a simulation that had weak easterlies in the tropical lower stratosphere, with very little year-to-year variability (e.g., Hamilton and Yuan, 1992; Hamilton, 1996). Only recently have the first GCM simulations of spontaneous QBO-like mean flow oscillations been reported (Takahashi, 1996; Horinouchi and Yoden, 1998; Hamilton et al., 1999). All three of these studies showed that a key to producing a QBO-like oscillation is employment of rather fine numerical grid resolution in the vertical (less than 1 km level spacing in all cases). However, there seem to be other aspects of model formulation involved as well. For example, Takahashi (1996) showed that his model simulated a QBO-like oscillation when one particular parametrization of moist convection was used, but not when another equally plausible parametrization was employed. It is also noteworthy that, while these models have produced equatorial mean wind oscillations with some features similar to the observed QBO (amplitudes, formation of strong descending shear zones), all the model oscillations have unrealistically short periods (between about 1 and 1.4 years). The ability of some comprehensive GCMs to simulate spontaneous equatorial wind oscillations opens up the possibility of detailed diagnosis of the dynamics involved. Undoubtedly much work in this area will be performed in the near future.

b *Interaction Between the Tropical and Extratropical Stratosphere*

As noted in Section 2b the phase of the QBO in the tropical lower stratosphere is observed to be related, at least in a statistical sense, to the state of the NH winter polar vortex. It is straightforward to show that the mid- and high-latitude effects of the QBO cannot be ascribed simply to the response of the zonally-symmetric circulation to eddy momentum fluxes in the tropics (e.g., Plumb, 1982). Thus it is clear that the QBO in the tropics must be modulating in some manner eddy fluxes that affect the circulation of the extratropics.

Holton and Tan (1980) in their original paper suggested that the tropical mean wind variations act to modulate the propagation of large-scale quasi-stationary waves forced by topography in the extratropics. It is known that the wave activity from such waves generally propagates upward and somewhat equatorward (e.g., Edmon et al., 1980; Hamilton, 1982c). The situation is illustrated schematically in Fig. 13. This shows the tropospheric subtropical jet, the polar night jet, and the equatorial jet associated with the QBO. In the situation depicted in the top panel, the QBO is in its westerly phase and stationary waves will be able to propagate across the equator and into the summer hemisphere without encountering a critical surface (where the zonal-mean zonal wind is zero). By contrast when there are easterlies on the equator (bottom panel of Fig. 13), the stationary waves will encounter a critical surface on the northern flank of the equatorial jet. This will presumably stop any further equatorward penetration of the waves. In addition there is reason to believe that there will be at least partial reflection of wave activity flux back into the poleward direction (e.g., Haynes and McIntyre, 1987). This may lead to more concentration of quasi-stationary wave activity in the NH extratropics, a situation depicted schematically by the thick arrows in Fig. 13. This in turn will lead to increased wave drag on the polar vortex, explaining the correlation between equatorial mean winds and the strength of the polar vortex. This process has been simulated successfully in a variety of nonlinear numerical models, ranging from purely barotropic one-layer idealizations of the global stratosphere (e.g., O'Sullivan and Salby, 1990), to mechanistic three-dimensional models of the middle atmosphere forced with prescribed stationary waves at the tropopause (e.g., Holton and Austin, 1991; O'Sullivan and Dunkerton, 1994), to comprehensive GCMs (e.g., Hamilton, 1998b).

The modulation of the equatorward propagation of the stationary planetary waves by the tropical mean flow has implications for the zonal asymmetries in the QBO, an issue discussed earlier in Section 2b. The stationary waves are topographically forced and so have a fixed geographical distribution. To the extent that these waves are modulated by the QBO in zonal-mean wind, they will appear as zonal asymmetries in the QBO, even as determined from monthly-mean observations. In fact the expected QBO modulation of the waves at low latitudes is observed (Ortland, 1997) and found in model simulations (Hamilton, 1998b). At levels where the mean easterlies dominate, the monthly-mean wind near the equator appears quite zonally-symmetric, but at levels with mean westerlies there is a significant ($\sim$5 m s^{-1}) amplitude, zonal wavenumber one, perturbation evident even at the equator. A similar effect is apparent at stratopause levels, where large-scale planetary waves are observed to penetrate deeply into the tropics at times of equatorial mean westerlies (e.g., Barnett, 1975; Hirota, 1976).

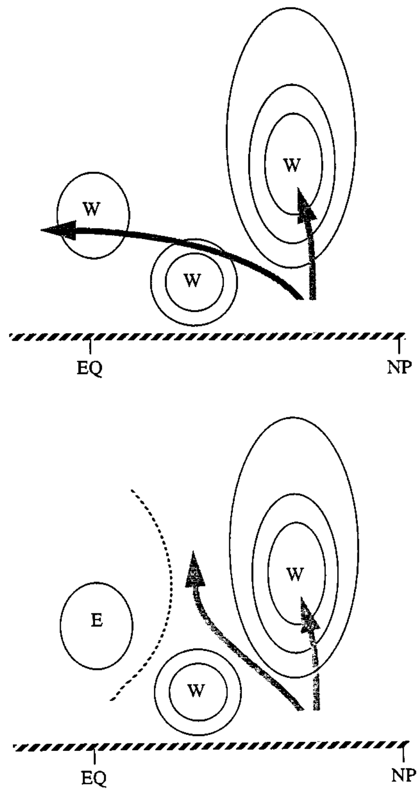

Fig. 13 A schematic diagram showing the zonal wind structure in the winter hemisphere in the westerly phase of the QBO in the lower stratosphere (top) and in the easterly phase (bottom). The dashed line in the bottom panel shows the location of the zero wind line. The thick arrows denote the dominant paths of wave activity propagation for quasi-stationary planetary waves forced near the surface in the extratropics.

c *Aspects of the Theory of the SAO*

The SAO would appear at first glance to be much less of a mystery than the QBO. The fact that the semiannual component of the seasonal cycle of a meteorological quantity might become prominent at low latitudes is not surprising, given the fact that the sun crosses the equator twice each year. Determining the detailed mechanisms responsible for the SAO has been challenging, however. The first significant theoretical paper was that of Dunkerton (1979), who noted the observed downward progression of the westerly acceleration phase of the SAO suggested that it might result from the interaction of a vertically-propagating wave with the mean flow. He constructed a one-wave Plumb model, and picked the wave parameters to be similar to those for a prominent Kelvin wave that had been identified in observations by Hirota (1978). In particular this “Hirota” wave has a faster phase speed

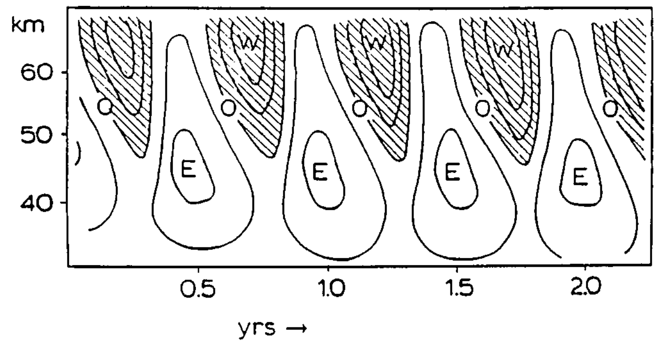

Fig. 14 The height-time evolution of the equatorial zonal-mean zonal wind in the simple model of Dunkerton (1979). The contour interval is 10 m s^{-1} and westerlies are shaded. Reproduced from Dunkerton (1979).

($\sim$50–70 m s^{-1}) than the waves normally employed in models of the QBO. The higher phase speed of this wave allowed it to propagate through the stratosphere with only modest attenuation and then to be strongly absorbed in the region of very fast radiative damping near the stratopause. Dunkerton showed that the interaction of such a wave with the mean flow could plausibly produce the observed westerly SAO accelerations. To make his model oscillate, Dunkerton simply assumed the presence of an easterly mean flow driving with a semiannual cycle, and no vertical phase variation. The result of one of Dunkerton's experiments is reproduced here as Fig. 14. The time-height section has considerable resemblance to the observed stratopause SAO (Fig. 7) at the equator. Hitchman and Leovy (1988) used satellite temperature observations to estimate the properties of the large-scale Kelvin waves in the equatorial stratopause region, and went on to show that indeed such waves represent an important driving mechanism for the westerly mean flow accelerations in the SAO.

A plausible mechanism for the easterly mean flow driving in the SAO was provided by Holton and Wehrbein (1980) and Mahlman and Sinclair (1980). They noted that the residual mean meridional circulation at stratopause heights is directed from the summer into the winter hemisphere. One result is to advect the summertime easterlies onto and across the equator. This has the effect of producing easterly maxima on the equator twice a year, shortly after the solstices. Holton and Wehrbein included this effect in a simple zonally-symmetric model of the middle atmosphere (with eddy effects simply included as a linear drag on the zonal wind). They were able to produce a simulation with an oscillation at the equatorial stratopause between strong easterlies (near the solstices) and weak easterlies (near the equinoxes). Takahashi (1984) generalized the Holton-Wehrbein model to include a westerly forcing due to dissipation of an equatorial Kelvin wave with properties similar to that considered by Dunkerton (1979) in his one-dimensional model. Takahashi was able to obtain a reasonably realistic simulation of the stratopause SAO with both easterly and westerly phases. The main deficiency in this simulation was a tendency for the easterly phase to be too narrowly confined near the equator. More recent work has considered the possibility that the easterly phase might be broadened and intensified by the wave-driving associated with quasi-stationary planetary waves propagating in from mid-latitudes (e.g., Hamilton, 1986). In contrast to the situation for the QBO, the stratopause SAO has been reasonably well simulated in a number of GCMs (Hamilton and Mahlman, 1988; Sassi et al., 1993; Jackson and Gray, 1994; Muller et al., 1997). This has allowed the mechanisms involved in driving the SAO to be examined in detail (at least within the context of the GCMs). It appears in the model simulations that both the cross-equatorial advection by the mean meridional circulation and the effects of quasi-stationary planetary waves contribute significantly to the easterly accelerations in the stratopause SAO.

Another issue that is still being investigated is the role of gravity waves with less than global scale in the SAO. Hamilton and Mahlman (1988) found that in their GCM simulation the westerly accelerations in the stratopause SAO were forced significantly by explicitly-resolved gravity waves of horizontal wavelength $\sim$1000s of kilometres, as well as by global-scale Kelvin waves. Sassi and Garcia (1997) constructed a simple mechanistic model in which the bulk of the westerly wave driving in the SAO came from convectively-excited gravity and Kelvin waves with scales of a few thousand kilometres. Recently Ray et al. (1998) included a parametrization of the effects of small-scale gravity waves in a simple diagnostic calculation of the zonal-mean momentum balance of the tropical middle atmosphere. They also determined that the gravity waves are very important for forcing the westerly phase of the stratopause SAO.

We have yet to attain a detailed understanding of the SAO near the mesopause. Dunkerton (1982) noted that, in the presence of a spectrum of vertically-propagating waves, the "shadowing" effect of the mean wind changes at the stratopause could lead to an SAO of exactly the opposite phase at higher levels. With an appropriate choice for the wave parameters, he was able to construct a Plumb-type model that produced a fairly realistic mesopause SAO when a realistic SAO was imposed near the stratopause.

d *Long-period Variations in Trace Constituents*

The quasi-biennial oscillation in wind and temperature is known to correlate closely with a QBO in ozone concentrations (e.g., Oltmans and London, 1982; Hamilton, 1989; Randel and Wu, 1996) and in the concentrations of other stratospheric trace constituents (e.g., Zawodny and McCormick, 1991; Hitchman et al., 1994; O'Sullivan and Dunkerton, 1997; Cordero et al., 1997; Randel et al., 1998). The bottom panel in Fig. 15 shows 14-year times series of the 40-mb Singapore zonal winds and the equatorial zonal-mean total column ozone determined from the Total Ozone Mapping Spectrometer (TOMS) instrument on the Nimbus-7 satellite. Both of the plotted time series had the annual cycle and all harmonics of the annual cycle

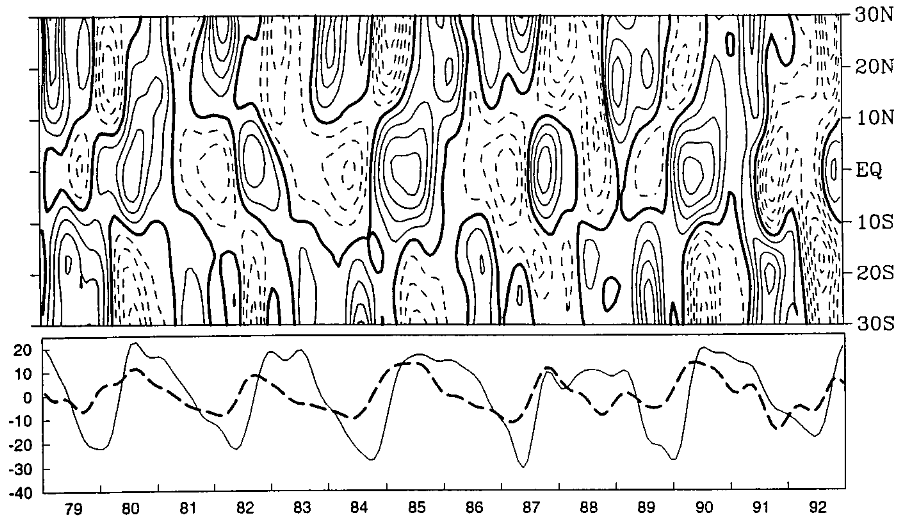

Fig. 15 The bottom panel shows the 1979–92 time series of 40-mb Singapore zonal wind (solid) and equatorial zonal-mean total column ozone from the Nimbus-7 TOMS instrument (dashed). Both the winds and ozone values have been deseasonalized and then band-passed filtered. See text for details. The scale is marked in m s^{-1} for the winds and Dobson Units for the total ozone column. The top panel shows a latitude-time section for the filtered 1979–92 total column ozone measurements. The contour interval is 3 Dobson Units and dashed contours denote negative values.

removed, and then were band-passed to select periods between 0.5 and 4.5 years. The interannual variations in the two quantities are clearly correlated, with the maximum ozone columns occurring around the time the 40-mb winds first reach the westerly extreme. The peak-to-peak amplitude of the QBO in ozone total column near the equator is $\sim$15–20 Dobson Units or $\sim$5–7% of the time-mean value. The basic explanation of the ozone QBO was provided by Reed (1965b), who noted that the changes in the temperature in the QBO should be connected with a QBO in the diabatically-induced meridional circulation. Again the situation is illustrated in Fig. 12. The anomalous rising and sinking motion associated with the QBO temperature changes cause variations in any trace constituent with a mean vertical stratification. If there is also a strong vertical gradient in the chemical lifetime of the constituent, then it is possible to produce a QBO in the integrated total column concentration. For the case of ozone, the mixing ratio increases rapidly with height in the lower and middle stratosphere and the chemical timescales decrease rapidly with height. As the QBO westerlies descend, the anomalous mean meridional sinking depresses the ozone isopleths near the equator. Since the chemical timescale is short at high levels, the ozone that is pulled down from the middle stratosphere is replaced rapidly, while at lower altitudes the ozone acts more nearly as a passive tracer. Thus the descent is associated with an increase in the total ozone column, and the equatorial ozone column reaches its maximum value after the westerly shear zone has descended through most of the stratosphere.

The QBO-related mean vertical motion at the equator must be balanced by flow of the opposite sign at higher latitudes (again as in Fig. 12). The temperature QBO has a node near 15° latitude, and so the diabatically-induced QBO vertical motion is presumably of opposite sign equatorward and poleward of about 15° latitude. The top panel in Fig. 14 shows the filtered ozone time series plotted as a function of latitude. A general tendency for the QBO-related ozone fluctuations to be out of phase across a latitude of about 15° is evident, as expected on the basis of the simple model in Fig. 12. However, the off-equatorial ozone anomalies have a more complicated appearance than those near the equator. The higher latitude variations appear to involve a significant QBO modulation of the annual cycle of atmospheric transport (e.g., Hamilton, 1989; Tung and Yang, 1994; O'Sullivan and Dunkerton, 1997).

Recent work has shown that Reed's original explanation for the ozone QBO needs to be modified in certain respects. One interesting issue is that the constituents that interact chemically with ozone also are transported by the QBO-related circulation. This turns out to be quite significant in the middle and upper stratosphere where the ozone chemical lifetime is short compared to the QBO transport timescale, but the lifetime of nitrogen dioxide is significantly longer than that of ozone. Since nitrogen dioxide acts to catalytically destroy ozone, the ozone concentration depends on the local nitrogen dioxide concentration. Chipperfield et al. (1994) showed that this effect should lead to a significant ozone QBO even in the middle and upper stratosphere.

Another significant modification to Reed's simple picture concerns the nature of the QBO-induced meridonal circulation itself. Jones et al. (1998a,b) and Kinnersely and Tung (1998) show that in simple models that include an annual cycle, the branch of the QBO-induced meridional circulation is considerably intensified in the winter hemisphere relative to that in the summer hemisphere. The analysis of these authors indicates that this can largely be attributed to the effects of meridonal advection of zonal momentum by the annually-varying cross-equatorial mean meridional circulation. Any jets that are formed near the equator by the action of vertically-propagating waves are advected somewhat into the winter hemisphere by this large-scale summer-to-winter flow. The thermal wind balance implies that the temperature perturbations associated with a given shear should increase as f, so even a small displacement of a jet off the equator can greatly modify the associated temperature structure (and hence the diabatic circulation). It may be that this effect can account for much of the annual synchronization of the ozone QBO noted above (Hamilton, 1989; Tung and Yang, 1994).

The basic theory for the QBO effects on trace constituent concentrations is also applicable to the SAO. Both observations and models of trace constituent concentrations in the tropical upper stratosphere support the notion that the dynamical SAO drives an SAO in atmospheric composition at low latitudes (e.g., Gray and Pyle, 1986). Once again, near the equator the tendency is for the descending westerly shear zone of the SAO to depress mixing ratio isopleths. This is nicely illustrated by the satellite radiometer determinations of nitrous oxide (N_2O) mixing ratio for the months of January, April and July shown in Fig. 16. N_2O has a tropospheric source and a chemical destruction mechanism that becomes much more effective with height, leading to the overall vertical stratification seen in these sections. Near the tropical stratopause the results have a strong dependence on season, with depressed contours at the equator in April (near the westerly extreme of the SAO at these heights) and a single peak structure in January and July. This is consistent with the notion that near the equator downward mean motion accompanies the descending westerly shear regime of the SAO.

6 Conclusion

In the roughly four decades since the discovery of the QBO there has been a remarkable growth in our knowledge of the circulation of the tropical middle atmosphere. Today there is a detailed observational record of tropical stratospheric zonal winds extending almost half a century. It is clear that the QBO is a rather stable feature of the circulation, although the possibility of very long-term modulation of QBO behaviour cannot be ruled out (e.g., Hamilton and Garcia, 1984; Teitelbaum et al., 1995). The operational in situ balloon observations of winds and temperatures have been supplemented in recent years by satellite-based radiometer observations of temperature, horizontal wind and trace constituent concentrations. These satellite observations show the very strong effects of the QBO in the dynamics and transport of the stratosphere. The satellite measurements have also been useful in delineating the structure of the SAO which is prominent in the regions of the tropical middle atmosphere above the ceiling of operational balloon observations.

These impressive observational advances have been matched by developments in theory and modelling. The explanation of the QBO in terms of the interaction of the mean flow with vertically-propagating waves (Lindzen and Holton, 1968; Holton and Lindzen, 1972; Plumb, 1977) represented a very important step not only for tropical stratospheric dynamics, but for the overall development of the theory of wave-mean flow interaction (e.g., Andrews and McIntyre, 1976).

Of course, there is still great scope for further investigation into tropical stratospheric dynamical processes. For example, the understanding of the interactions of the QBO in the tropics with the circulation of extratropical latitudes is far from complete. Another important issue that remains clouded by uncertainty is the simulation of the QBO by comprehensive general circulation models. In particular, it is not clear exactly what features distinguish the vast majority of models, which have no QBO at all, from the few models which have produced a QBO-like large-amplitude interannual oscillation of the tropical circulation. While GCMs typically differ somewhat in various aspects of their simulated climate (e.g., Boer et al., 1992), there is no other region of the atmosphere where the simulation seems to depend so spectacularly on the model employed. Even less clear at present is how

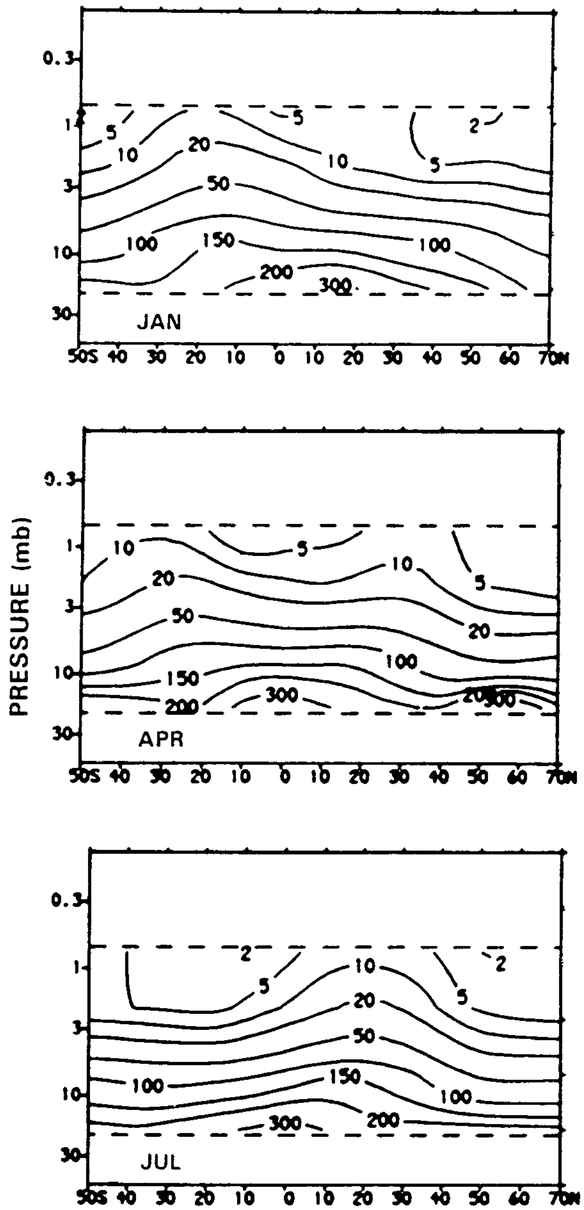

Fig. 16 Zonal-mean mixing ratios for N_2O determined for individual months using data from the Stratosphere and Mesosphere Sounder instrument on the Nimbus-7 satellite. Contours are labelled in parts per billion by volume and results are plotted only between 50°S and 70°N. Adapted from Gray and Pyle (1986).

one might modify the formulation of a GCM to lead to a truly realistic simulation of the QBO.

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