Annu. Rev. Fluid Mech. 2005. 37:263–93
doi: 10.1146/annurev.fluid.37.061903.175710

STRATOSPHERIC DYNAMICS

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Key Words geophysical fluid dynamics, vortex dynamics, climate, weather, atmosphere

■ **Abstract** The large-scale dynamics of the extratropical stratosphere are reviewed. The role of Rossby waves and vortex dynamics in shaping the winter stratospheric circulation and the dynamics of the longitudinal mean flow are first discussed separately. The important effects of two-way interaction between waves and mean flow are then described, with emphasis on how mechanisms discovered in simple models can be followed through to models that are closer to the real stratosphere. A final topic is the possible effect of the stratosphere on the troposphere, with emphasis on dynamical mechanisms for such an effect.

1. INTRODUCTION

The atmosphere is conventionally subdivided in the vertical into regions according to the vertical temperature gradient, with temperature decreasing upward in the troposphere (0–10 km in global average), then increasing upwards in the stratosphere (10–50 km), then decreasing upward in the mesosphere (50–80 km), and then increasing upward again in the lower thermosphere (above 80 km). In the textbook *Middle Atmosphere Dynamics* (Andrews et al. 1987), the authors use the term "middle atmosphere" to describe the last three regions (up to about 100 km). The middle atmosphere can be distinguished from the troposphere below by the fact that it is remote from the Earth's surface and the dynamical effect of latent heating due to phase changes of water can be neglected (whereas in the troposphere it is extremely important). The upper limit corresponds roughly to the homopause, above which molecular processes rather than bulk fluid processes dominate transport, allowing differential gravitational settling of different chemical species according to molecular weight, and also to the level above which increased ionization implies a strong role for electromagnetic forces in the dynamics. The middle atmosphere can be approximated as well-mixed on small scales and electrically neutral.

This review focuses primarily on the lowest part of the middle atmosphere, the stratosphere, which has the potential significantly to affect conditions at the surface, as follows:

(*a*) Stratospheric ozone: Changes in stratospheric ozone imply changes in surface ultraviolet irradiance and also a changed supply of ozone to the troposphere. The latter may have a significant effect on surface ozone concentrations (Zeng & Pyle 2003).

(*b*) Radiative balance of the troposphere: The lower stratosphere plays an important role in the radiative balance of the troposphere, and changes in greenhouse gases such as ozone and water vapor in the lower stratosphere directly affect surface temperatures.

(*c*) Weather and climate: Whereas in weather forecasting and tropospheric climate modelling it has been customary to de-emphasize the role of the stratosphere, there is now evidence of downward dynamical links between the stratosphere and troposphere. These links may determine, for example, the effect on surface weather and climate of stratospheric aerosol changes due to volcanic eruptions and may imply a strong role for the stratosphere in determining future changes in the tropospheric climate due to increases in carbon dioxide and other greenhouse gases.

(*d*) Solar variability: Vertical coupling of dynamics between stratosphere and troposphere may provide an explanation for apparent signals of solar variability (whose direct physical and chemical effects on the mesosphere and upper stratosphere are significant and relatively well-understood) in tropospheric weather and climate.

Study of the stratosphere over the last two decades has been stimulated primarily by (*a*, above) the need to account for observed changes in stratospheric ozone (most famously the Antarctic ozone hole) and to determine the relation to anthropogenic chemical emissions, particularly emissions of long-lived halogen compounds such as halocarbons. The rate of emission of such compounds has now decreased, following the Montreal Protocol, and attention is now focused on the "recovery" of the ozone layer, as halogen concentrations in the stratosphere decrease. There is still concern that some of the shorter-lived replacements may cause significant ozone destruction (WMO 2003), but attention is shifting to the possible impact of future climate change on ozone concentrations. Meanwhile, the growing evidence that the stratosphere plays a wider role in the chemical-climate system (*b*, *c*, *d*, above) has stimulated new interest in this region.

Andrews et al. (1987) reviewed basic principles of the dynamics[1] of the middle atmosphere that remain relevant today. However, there have been many important developments since, including the period of intense research into stratospheric

[1]The convention in atmospheric science is to divide physical processes into dynamics on the one hand, meaning the part described by the momentum and thermodynamic equations (i.e., fluid dynamics), and 'physics' on the other, meaning processes such as radiative transfer, microphysics, and phase changes. In the middle atmosphere there is a three-way interaction between dynamical processes, physical processes (primarily radiative transfer), and chemical processes.

ozone depletion, prompted by the discovery of the Antarctic ozone hole, new satellite observations (e.g., from the Upper Atmosphere Research Satellite), the inclusion of stratospheric levels in numerical weather prediction models and routine meteorological data sets, and the extension of numerical general circulation models to include realistic representation of the middle atmosphere.

The emphasis on ozone and related chemical species has prompted much interest in the transport of chemical species and its representation in models (e.g., Plumb 2002). This review does not attempt to cover the topic of chemical transport (though clearly the transport characteristics of the stratosphere are intimately connected with its dynamics). Furthermore, there is no space in this review to give details of the combination of meteorological and chemical observations that now underpins our current knowledge of the dynamical state of the stratosphere.

The stratosphere, because of the increase in temperature with height, is strongly stably stratified. A convenient density variable, conserved in adiabatic motion, is the potential temperature θ, which increases monotonically upward when the stratification is stable. The dynamics on the large scale is described by theories of rotating stratified flows as developed over the last 60 or so years. Although the large-scale dynamics is now straightforwardly simulated by numerical models, the underlying theory is important in providing a framework in which behavior of models can be understood and their results can be interpreted. There is also significant small-scale, high-frequency motion in the stratosphere and mesosphere. Much of this is associated with inertio-gravity waves, which are forced in the troposphere by flow over topography, convection, shear instability, and sometimes by spontaneous emission from the large-scale flow. These inertio-gravity waves propagate up into the stratosphere and mesosphere where they break and dissipate. This implies a force on the large-scale flow where the breaking or dissipation takes place. There may be additional nondissipative forces due to horizontal refraction (Bühler & McIntyre 2003). Breaking also leads to localized patches of three-dimensional turbulence and hence to mixing of chemical species. The length scales of the inertio-gravity waves themselves, let alone the three-dimensional turbulence that arises from their breaking, are well below the spatial resolution of present-day global models, and the effects of inertio-gravity waves on the large scale therefore need to be represented in the numerical models by parametrizations. The whole subject of inertio-gravity waves and their effects is centrally important to understanding and modelling the middle atmosphere and was comprehensively reviewed recently by Fritts & Alexander (2003). It is therefore not covered in any detail here.

The circulation of the middle atmosphere varies strongly in height, latitude, and longitude. However, the most systematic variations are in latitude and height. This motivates a description where at the leading order the longitudinal variation is neglected (by taking the longitudinal average of the observed, three-dimensional state, for example). Figure 1 shows height-latitude cross-sections of longitudinal average of longitudinal wind (Randel 1992). There is an important difference between the winter hemisphere, where the flow is dominated

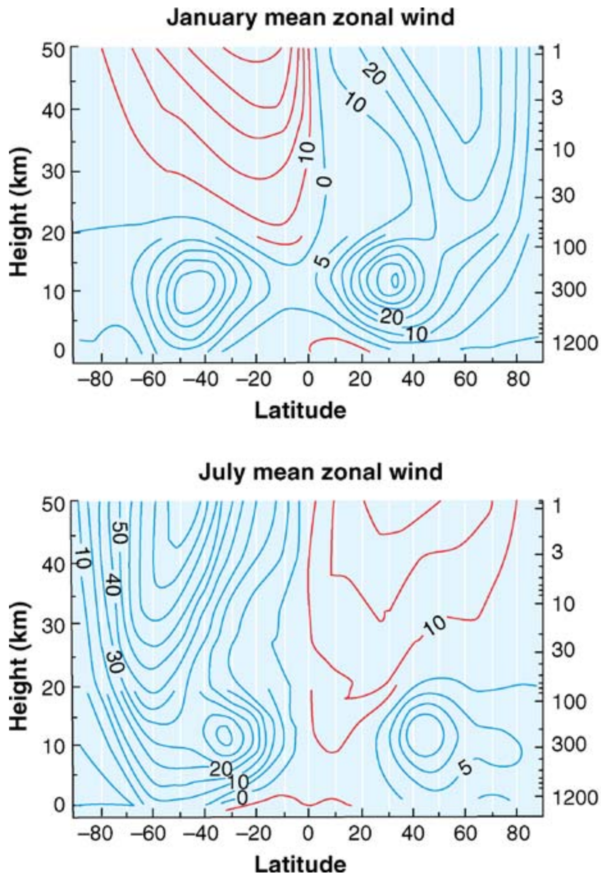

Figure 1 Longitudinally averaged longitudinal component of wind in troposphere and stratosphere for January (Northern Hemisphere winter) and July (Northern Hemisphere summer) (from Randel 1992). Negative regions (westward winds) are shaded. The winter hemisphere has strong eastward jets in the stratosphere (the "polar vortex"). The Southern Hemisphere jet (July) is stronger than the Northern Hemisphere jet (January) because of the weaker wave force in the Southern Hemisphere. (Copyright 1992, W.J. Randel. Reproduced with permission.)

by an eastward "polar vortex," and the summer hemisphere, where the flow is westward.

It has long been known that to explain the longitudinally averaged state it is necessary to take account of the systematic effects of the deviations, usually termed waves or eddies, of the actual circulation from the average or mean. The development of the theory of wave mean-flow interaction in the 1960s and 1970s was significantly stimulated by some of the questions posed by the observed state of the middle atmosphere. This theory is discussed in detail by Andrews et al. (1987) and references therein. One starting point for this review is that the wave mean-flow description continues to provide a useful quantitative framework for understanding the circulation of the middle atmosphere, although some of the limitations of this description are noted.

The molecular diffusivity in the stratosphere increases from 10^{-4} m^2 s^{-1} in the lower stratosphere to 10^{-2} m^2 s^{-1} in the upper stratosphere and is too small to play any direct role in the dynamics. The patches of three-dimensional turbulence noted above might in principle play a dynamical role, e.g., as a kind of vertical eddy diffusivity. But recent estimates of the mixing effect of these patches from chemical tracer observations (e.g., Legras et al. 2003) suggest a vertical eddy-diffusivity effect of at most 10^{-1} m^2 s^{-1} and probably at least an order of magnitude less, implying that their dynamical role is almost always negligible. A much more important "non-conservative" physical effect is radiative transfer, both short-wave heating primarily associated with ozone, and long-wave transfer, primarily associated in the stratosphere with carbon dioxide, ozone, and, in the very lowest part of the stratosphere, water vapor. The radiative transfer plays both a forcing role, in setting up large-scale pole-equator-pole temperature gradients and associated longitudinal velocity distributions (implied by geostrophic balance), and a dissipative role in that it acts to reduce the amplitude of fluctuations (e.g., wave amplitudes). An estimated thermal damping or relaxation timescale may be used as a measure of this dissipative role is estimated by a thermal damping timescale. This is determined, among other things, by the ambient temperature (smaller temperatures imply increased damping timescales) and by the vertical length scale (smaller scales imply decreased damping timescales). Estimates for vertical length scales of several kilometers based on radiative code calculations are 20–40 days in the lower stratosphere decreasing to 5 days in the upper stratosphere, although it is almost impossible to make such estimates precise (see Zhu 1993, 1997, including details of scale dependence; also Newman & Rosenfield 1997). As noted below, some representation of radiative transfer is an essential ingredient for any model of stratospheric dynamics on longer timescales than a few days.

The emphasis of this review is on the dynamics in the extratropics and its global implications. The topic of mean-flow interaction in the tropics is not treated in detail because one of the most important aspects, the quasibiennial oscillation, was recently reviewed by Baldwin et al. (2001). The structure of the remainder of the review is as follows. Section 2 discusses Rossby waves and stratospheric vortex dynamics. Section 3 considers the dynamics of longitudinally symmetric circulations. Section 4 discusses the extratropical stratospheric circulation, emphasizing the two-way interaction between waves and mean flows. Section 5 discusses dynamical coupling between troposphere and stratosphere (i.e., *c*, above).

2. ROSSBY WAVES AND STRATOSPHERIC VORTEX DYNAMICS

On the large scale and on timescales greater than a day or so the extratropical stratosphere is well described as a "balanced" system in which there is a single time-evolving scalar field, the potential vorticity (PV), which is materially conserved in adiabatic, frictionless motion, and from which all other dynamical fields may be instantaneously determined through a PV "inversion" (McIntyre 2003a,b and references therein). In the classical approach of analyzing small-amplitude disturbances to a self-consistent steady flow it is well known that such balanced systems allow, for given spatial structure, a single wave, usually described as a Rossby wave corresponding to the single time derivative in the PV conservation equation. The dynamics of Rossby waves involves horizontal advection (or more precisely, advection along θ-surfaces) of PV (which has a strong pole-to-pole gradient) with resulting changes in temperature and pressure fields and vertical displacement of fluid parcels. The balance assumption excludes other waves, such as inertio-gravity waves and acoustic waves.

An important part of the dynamics of the stratosphere is that planetary-scale Rossby waves (sometimes known as planetary waves) are excited in the troposphere, e.g., by flow over topography, by latent heat release, or through the nonlinear evolution of troposheric eddies (Scinocca & Haynes 1998), and then propagate up from the troposphere into the stratosphere and mesosphere. One important basic result, which may be derived under the rather strong assumptions of small-amplitude waves and background flow varying slowly only in the vertical, is the Charney & Drazin's classic result that waves propagate upward only through flow that is weakly eastward relative to phase speed (with maximum relative flow speed from propagation decreasing as length scale decreases) and only if the scale of the waves is sufficiently large. Given that the dominant forcing of stratospheric Rossby waves is geographically stationary, this provides a basic explanation of why the winter stratosphere (with eastward flow around the pole; see Figure 1) is much more disturbed than the summer stratosphere (with westward flow around the pole) and why the disturbances in the winter stratosphere tend to have much larger scales than is typical of the troposphere below.

The time evolution of the PV field in stratospheric flows gives a clear guide to the dynamical mechanisms operating. McIntyre & Palmer (1983, 1984), using PV maps calculated from observations, distinguished between reversible displacements and distortions of the polar vortex, which they associated with upward propagating Rossby waves, and the nonlinear stirring of the PV field outside the vortex, which they identified with the breaking of those Rossby waves. They called the region outside the vortex the stratospheric "surf zone." Many subsequent modelling and observational studies of the stratosphere further examined this vortex/surf-zone structure. Figure 2 shows a recent observational view of the PV field in the stratosphere (Simmons et al. 2004), from an "analysis" data set generated by the European Centre for Medium Range Weather Forecasts (ECMWF). Such data sets are produced by blending new observations with information from old observations that has been carried forward in time by a weather forecast model (in this case the ECMWF model).

The close mathematical relation between three-dimensional balanced systems and the two-dimensional vorticity equation led to the latter being studied as a simple and computationally inexpensive proxy for the three-dimensional system. Starting

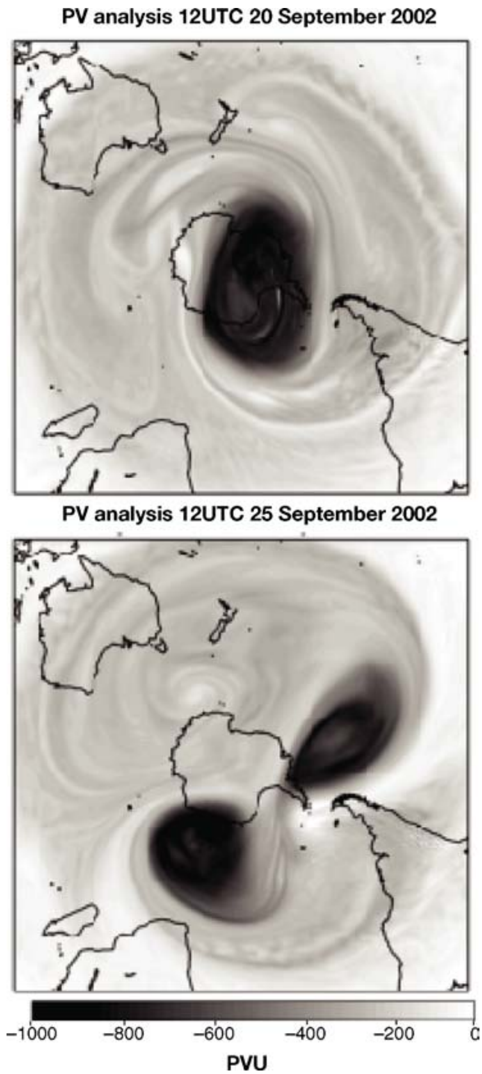

Figure 2 (From Simmons et al. 2004). Analyses of PV in the Southern Hemisphere on the 850-K potential temperature surface (corresponding roughly to 30 km), for 12 UTC September 20 and 25 September, 2002. Shading is −1000 PVU (*black*) to 0 (*white*) for potential vorticity (1 PVU $= 10^{-6}$ m^2 s^{-1} Kkg^{-1}). The PV distribution on September 20, 2002 shows the polar vortex (large negative values of PV, *dark shading*), surrounded by a "surf zone" in which streamers of air drawn out of the vortex and streamers of air from low latitudes are stirred together. The PV distribution of September 25, 2002 shows the vortex split into two in a sudden warming. This event is unique, in the Southern Hemisphere, over the last 50 years. (Copyright 2004, Am. Meteor. Soc. Reproduced with permission.)

with the work of Juckes & McIntyre (1987), a number of numerical studies of two-dimensional stratosphere-like flows gave important insights into the dynamics of the stratospheric polar vortex and surf zone, which have provided a useful framework on which to base interpretation of observations and of more realistic modelling studies as three-dimensional simulations have become more feasible. Juckes & McIntyre (1987) noted several important features including material coherence of the vortex (strictly speaking, the high PV core of the vortex), which, even for quite large-amplitude forcing, experienced reversible deformation but with almost no transport of fluid between interior and exterior, and the strong stirring effect of the disturbed flow outside the vortex, which tended to pull filaments of material out of the edge of the vortex and mix them into the exterior flow. They identified the region outside the vortex with the real stratospheric surf zone and noted that the confinement of filamentation and mixing to the exterior of the vortex naturally led to sharpening of the vorticity gradients surrounding the vortex through an erosion process. Another interesting feature of the Juckes & McIntyre simulations was the roll up into coherent small-scale vortices of filaments of high-vorticity air that were drawn out of the main vortex. The roll up may, in part, be explained by shear instability, since a PV profile taken across a filament exhibits a local maximum.

Later work with two-dimensional models confirmed this picture and explored the relevant processes in more detail. Some of this work was based on shallow-water models in which a relaxation of the thickness field to a suitable equilibrium distribution was used as a natural analogue of thermal relaxation due to radiative transfer. Juckes (1989) showed how the combined effects of such relaxation and wave forcing could lead to a persistent, sharp-edged vortex (see also Salby et al. 1990) and analyzed the transport characteristics of the flow. Polvani et al. (1995) focused on the subtropical part of the surf zone and showed how its structure depended on the presence of the relaxation term. Chen (1996) further explored the dependence of the structure of the surf zone, particularly its subtropical part, on wind profiles in the tropics and subtropics. The sharp PV gradients that form at the vortex edge through the erosion process motivated a piecewise-constant-vorticity model, with one value of vorticity inside the vortex and another value outside, which could be solved using the contour-dynamics method. This is attractive for reasons of both simplicity and computational economy. Polvani & Plumb (1992) used such a model to examine the process of filamentation and breaking and Polvani & Dritschel (1993) made a more general contour-dynamics-based study of vortex dynamics on a sphere.

Study of these stratosphere-like flows within two-dimensional models naturally focused attention on certain generic problems of two-dimensional vortex dynamics. One is the idealized erosion or "vortex stripping" problem (Legras et al. 2001 and references therein), where the action of a large-scale external flow applied to a pre-existing vortex is to remove the outer layers and leave a sharp jump in vorticity between the vortex itself and the exterior. If the external flow is not too strong then the jump eventually becomes large enough to give a dynamical stabilizing effect and hence inhibit the removal of further layers. This is a useful analogue of the action of Rossby wave forcing on the polar vortex, but is relevant to a much more general class of problems, e.g., to the evolution of vortices in two-dimensional turbulence.

A second generic problem is that of the effect of external flow on the evolution of filaments of vorticity. Studies of the stability of one-dimensional filaments show how the action of a strain flow on a filament is stabilizing (Dritschel et al. 1991). The effect of external shear may also be stabilizing (Dritschel 1989). In axisymmetric geometry the effect of the latter is such that a circular filament of vorticity surrounding an isolated vortex of the same sign may be stabilized by the shear of the irrotational external flow. The stabilizing effect is stronger on the sphere (Polvani & Dritschel 1993). These mechanisms imply that the roll up of filaments drawn out of the polar vortex into the surf zone may be relatively rare and are also relevant to the evolution of the filamentary vorticity structure that typically arises outside vortices in two-dimensional turbulent flows.

A basic paradigm for a three-dimensional balanced system is the well-known quasigeostrophic PV equation. This is closely analogous to the two-dimensional vorticity equation with the difference that the flow is layer-wise two-dimensional (in the horizontal plane) and the inversion operator that gives the stream function in terms of the quasigeostrophic PV is an inverse three-dimensional Laplacian-type operator, which is approximately isotropic if vertical distances are divided by Prandtl's ratio of scales f/N, where N is the buoyancy frequency and $f = 2\Omega \sin \phi$, where Ω is the rotation rate of the Earth and ϕ is the latitude, is the Coriolis parameter. The quasigeostrophic system has insufficient accuracy for stratospheric modelling and an effective way of investigating the dynamics of the extratropical stratosphere has been to use so-called primitive-equation models, which in principle allow inertio-gravity waves, but which in practice evolve almost as if balanced when used to simulate large-scale stratosphere-like flows. Concepts such as PV inversion may then be used diagnostically, rather than to predict the evolution of the flow.

Simulations in idealized models, some using the primitive equations (Haynes 1990, O'Neill & Pope 1988, Polvani & Saravanan 2000), some based on three-dimensional quasigeostrophic versions of contour dynamics (Dritschel & Saravanan 1994, Waugh & Dritschel 1999), show that in three-dimensional flows, where wave forcing is applied at low levels, there is upward propagation of Rossby waves and the nonlinear evolution of the flow at levels remote from the forcing generally exhibits the pattern identified by Juckes & McIntyre (1987), i.e., with persistence of the vortex as a material entity, strong stirring outside the vortex, which draws filaments of high PV air into the surf-zone, and sharpening of the PV gradients at the vortex edge through the process of vortex erosion. (The interpretation of the nonlinear evolution as Rossby wave breaking is natural in these flows, where there is a clear propagation phase, whereas in the two-dimensional flows described earlier the separation between propagation and breaking is not so clear and hence use of the term "wave breaking" at all is not so clearly justifiable.) Important features that occur in the three-dimensional case include the possibility of breaking at low levels, which destroys the vortex at those levels and prevents propagation of wave activity to upper levels (Dritschel & Saravanan 1994), sensitivity of the behavior to variation in the size and strength of the vortex in the vertical because of the implications for Rossby wave propagation (Polvani & Saravanan 2000, Waugh & Dritschel 1999), and three-dimensionality of the small-scale vortices that form as a result of the roll up of filaments (or, more correctly in the three-dimensional case, sloping sheets) of high PV that are drawn out of the main vortex during wave breaking. Polvani & Saravanan (2000) note that sheets drawn out of the vortex tend to be deep and speculate on the relevance of the vortex alignment process previously identified in quasigeostrophic turbulence.

Although the existence of small-scale vortices in the stratosphere has been argued on fluid-dynamical grounds, this has been very difficult to verify directly from observations. The main source of dynamical information has been from satellite radiometers, which generally have rather coarse resolution in either vertical or horizontal. The ECMWF analysis data set mentioned above is now at very high spatial resolution (40 km in horizontal, 1.5 km in vertical). Figure 3, taken from Simmons et al. (2004), shows pictures of small-scale vortices revealed by this data set. These features in the data set are strongly model-influenced, but Simmons et al. note that they fit well with actual observations. For example, the temperature and wind signatures of one of the small vortices shown on October 4, 2002 is in good agreement with measurements from Australian and neighboring radiosondes.

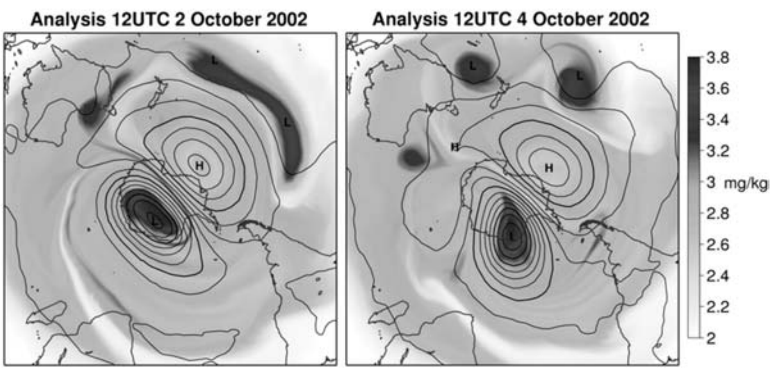

Figure 3 (From Simmons et al. 2004). Analyses of water vapor concentration (by mass) on the 850-K potential temperature surface (corresponding roughly to 30 km), for 12 UTC October 2 and October 4, 2002. Shading is from 2 mg/kg (*white*) to 3.8 mg/kg (*black*). The water vapor may be regarded as a tracer and high values correspond to high negative values of PV (recall Figure 2) and hence mark air that originated in the vortex. Over the two days shown, the band of high water vapor concentration/high PV air over the Pacific evolves into two separate vortices. (Copyright 2004, Am. Meteor. Soc. Reproduced with permission.)

The lifetime of small-scale vortices such as those seen in Figure 3 is likely limited, by decay due to radiative transfer, to a few days, with deeper vortices lasting longer than shallow vortices. Haynes & Ward (1993) discuss relevant mechanisms and estimates of timescales. Such vortices are also associated with anomalies in chemical species such as water vapor and ozone, and once the vortex itself dissipates, the chemical anomalies are vulnerable to deformation by large-scale shear, stretched into filaments/sheets and ultimately mixed into the background.

If the wave forcing is strong enough, the main vortex may be significantly displaced from the pole, strongly deformed in shape, or even split into two. These events are known to meteorologists as "sudden stratospheric warmings" because they are manifested as very rapid increases in temperature, due to adiabatic warming through descent (as implied by inversion of the evolving PV field). In the Northern Hemisphere sudden stratospheric warmings occur in mid-winter in about half of winters, on average, and there is also often a sudden-warming-like event at the end of winter. Disturbances to the vortex are generally weaker in the Southern Hemisphere than in the Northern Hemisphere, as is expected from the relative lack of topography and greater proportion of ocean versus land, and strong events tend to be confined to the end-of-winter transition—the "final warming"—in October or later. Therefore, there was considerable surprise and interest when, as seen in Figure 2, the Southern Hemisphere polar vortex split into two in September 2002, with a corresponding split of the "ozone hole"—the region of very low ozone air that currently forms annually over the Antarctic in the Southern Hemisphere winter. This event restimulated the interest of the broader atmospheric science community in the dynamics of the stratosphere, particularly in sudden warmings. An issue of the *Journal of Atmospheric Sciences* dedicated to the September 2002 warming will appear shortly.

The important ingredients of the dynamics of the extratropical winter stratosphere described in this section are upward propagation of Rossby waves on the PV gradients near the edge of the polar vortex and the breaking of these waves at levels where their amplitudes become large, giving rise to filamentation, deformation, and even splitting of the main vortex. In highly disturbed states the role of upward Rossby wave propagation may be small and a vortex-interaction description may then be more relevant.

In the remainder of this review the emphasis is less on vortex dynamics and more on the wave mean-flow description where the flow is divided into a longitudinal average or mean part and a disturbance or wave part. The essence of the theory of wave mean-flow interaction is that there is long-range momentum transport between the location where the waves are generated and the location where the waves break or dissipate (Andrews et al. 1987). The associated force is here called a "wave force." The theory may also be formulated in terms of transport of PV, which is local, within the breaking or dissipation region (e.g., McIntyre & Norton 1990). (The latter alternative may seem the more natural in view of the vortex phenomena described previously.) The relevant theories are most highly developed when the flow is weakly disturbed from a longitudinally symmetric state. Formal expressions for the wave force and wave momentum transport are valid in general, but what is lost when the flow is highly disturbed is any clear association with wave propagation. There is ongoing discussion and investigation of alternative formulations (e.g., see discussion in McIntyre 1982, also Thuburn & Lagneau 1999).

3. THE DYNAMICS OF THE LONGITUDINALLY AVERAGED CIRCULATION

In this section I discuss the dynamics of the longitudinally averaged circulation. As noted above, the systematic effects of waves or eddies are an essential part of this dynamics and are represented here by a longitudinal wave force $\mathcal{G}$ per unit mass. This exploits the simplification allowed either by the so-called "transformed Eulerian mean formalism" or by formulating the dynamical equations using θ as a vertical coordinate (e.g., Andrews et al. 1987). $\mathcal{G}$ represents the effect of waves, such as Rossby waves and gravity waves, on the mean flow. In the context of Rossby waves, $\mathcal{G}$ has to represent the propagation, breaking, and vortex interaction behavior described in the previous section and therefore has to be a complicated nonlinear, (and, as yet, undetermined) function of the mean flow and of wave sources. Gravity wave propagation and breaking also depends strongly on the background flow (Fritts & Alexander 2003) and has to be incorporated into $\mathcal{G}$. In this section we consider the much more straightforward problem of how the circulation responds to a given force.

3.1. Basic Principles

A suitable simplified set of model equations for this problem is as follows:

$$\frac{\partial \bar{u}}{\partial t} - 2\Omega \sin\phi \bar{v}^* = \mathcal{G} - \gamma \bar{u} \tag{3.1}$$

$$2\Omega \sin\phi \frac{\partial \bar{u}}{\partial z} + \frac{R}{aH}\frac{\partial \bar{T}}{\partial \phi} = 0 \tag{3.2}$$

$$\frac{\partial \bar{T}}{\partial t} + \bar{w}^* \left(\frac{HN^2}{R}\right) = \bar{Q}_s + \bar{Q}_l(\bar{T}) = \bar{Q}_s - \alpha \bar{T} \tag{3.3}$$

$$\frac{1}{a\cos\phi}\frac{\partial}{\partial \phi}(\bar{v}^* \cos\phi) + \frac{1}{\rho_0}\frac{\partial}{\partial z}(\rho_0 \bar{w}^*) = 0 \tag{3.4}$$

Here ϕ is latitude and z is a log-pressure coordinate. a is the radius of the Earth, R is the gas constant for dry air, H is the constant nominal density scale height used to define the log-pressure coordinates and ρ_0 is the background density in these coordinates, equal to $e^{-z/H}$. H is usually taken to be about 7 km, implying rough correspondence between the log-pressure coordinate z and geometric altitude. The independent variables are $\bar{u}$ (longitudinal velocity), $\bar{T}$ (temperature), and $(\bar{v}^*, \bar{w}^*)$ (latitudinal and vertical components of the circulation in the latitude-height plane, usually called the mean meridional circulation). Equations 3.1–3.4 are strictly valid only in the limit of small disturbances to a resting basic state in which the temperature varies only in height and for slow time variations (i.e., longer than a day or so). Equations 3.1–3.4 define a balanced system in the sense discussed in Section 2. (There are apparently two time derivatives, but one may be eliminated.)

The notation $\overline{(.)}$ emphasizes that the variables may be interpreted as longitudinal averages, with the meridional circulation taken as the "transformed Eulerian mean" circulation, in which case the effect of the eddies enters only through the wave force $\mathcal{G}$ in the longitudinal momentum equation (Equation 3.1). Following tradition, the term $\gamma\bar{u}$ is included in this equation to represent linear friction, but it is very difficult to argue that such a friction is at all relevant in the stratosphere. The second term on the left-hand side represents the Coriolis torque associated with the mean meridional circulation. Equation 3.2 expresses the so-called thermal wind relation between longitudinal velocity and temperature which follows from geostrophic and hydrostatic balance.

In the temperature equation (Equation 3.3), N^2 is the square of the buoyancy frequency, defined as

$$N^2 = \frac{R}{H}\left(\frac{dT_0}{dz} + \frac{\kappa T_0}{H}\right),$$

where T_0 is a reference temperature dependent only on z and $\kappa = R/c_p \simeq \frac{2}{7}$, where c_p is the specific heat of dry air at constant pressure. $\bar{Q}_s$ and $\bar{Q}_l$ are the short-wave and long-wave radiative heating rates, respectively. The latter is approximated as $-\alpha\bar{T}$, which is an extreme simplification but, as discussed above, captures the fact that the real radiative transfer is essentially relaxational. The final equation (Equation 3.4) expresses mass continuity for the meridional circulation.

Equations 3.1–3.4 describe the coupled response of the wind and temperature fields on the one hand and the meridional circulation in the latitude-height plane on the other, to applied wave force $\mathcal{G}$ and short-wave heating $\bar{Q}_s$. Any three independent variables can be eliminated to give a single equation for the fourth. The discussion below focuses on the meridional circulation, in particular the vertical velocity $\bar{w}^*$, but once this is known other fields can be deduced as required. To highlight the dependence of the response on the spatial structure and the time dependence of $\mathcal{G}$ and $\bar{Q}_s$, it is instructive to consider the case where they are assumed to be sinusoidal functions of time with given frequency σ, i.e., $\mathcal{G} = \mathrm{Re}(\hat{\mathcal{G}}e^{i\sigma t})$ and $\bar{Q}_s = \mathrm{Re}(\hat{Q}_s e^{i\sigma t})$, following Garcia (1987). It follows, by combining Equations 3.1–3.4, that the vertical velocity $\bar{w}^* = \mathrm{Re}(\hat{w}e^{i\sigma t})$ is the solution of the equation

$$\frac{\partial}{\partial z}\left[\frac{1}{\rho_0}\frac{\partial(\rho_0\hat{w})}{\partial z}\right] + \left(\frac{i\sigma+\gamma}{i\sigma+\alpha}\right)\frac{N^2}{4\Omega^2 a^2\cos\phi}\frac{\partial}{\partial\phi}\left[\frac{\cos\phi}{\sin^2\phi}\frac{\partial\hat{w}}{\partial\phi}\right]$$
$$= \frac{1}{2\Omega a\cos\phi}\frac{\partial}{\partial\phi}\left[\frac{\cos\phi}{\sin\phi}\frac{\partial\hat{\mathcal{G}}}{\partial z}\right] + \left(\frac{i\sigma+\gamma}{i\sigma+\alpha}\right)\frac{R}{4H\Omega^2 a^2\cos\phi}\frac{\partial}{\partial\phi}\left[\frac{\cos\phi}{\sin^2\phi}\frac{\partial\hat{Q}_s}{\partial\phi}\right] \tag{3.5}$$

Note first that the operator on the left-hand side is elliptic, so that the vertical velocity response to a forcing localized to a region spreads away from that region. This is as expected from PV inversion—the flow is assumed balanced and the non-local response to the applied forcing may be deduced by applying the appropriate inversion operator to the PV field, which is changing in time because of the applied forcing. Also note the fact that the different inverse timescales in the problem, the frequency σ and the frictional and thermal relaxation rates γ and α, respectively, appear only in the factor $(i\sigma+\gamma)/(i\sigma+\alpha)$ that appears both on the left-hand side and right-hand sides of Equation 3.5. Having displayed this dependence, the value of γ is set to zero, for reasons given above.

In the high-frequency limit, when thermal relaxation is negligible, the factor $i\sigma/(i\sigma+\alpha)$, which appears in two places in Equation 3.5, is approximately 1. The form of the operator on the left-hand side then implies that the response tends to have aspect ratio of the order of Prandtl's ratio of scales, f/N or $2\Omega\sin\phi/N$, mentioned above, with the structure of the response shallowing near the equator. The short-wave heating term on the right-hand side can be significant, implying that both wave force and short-wave heating may be effective in driving a mean circulation. As frequency decreases toward zero relative to the thermal damping rate, the modulus of the factor $i\sigma/(i\sigma+\alpha)$ also decreases towards zero. This has two important effects. First, the heating term on the right-hand side diminishes in importance. Second, the aspect ratio of the response changes, so that when $\sigma \ll \alpha$ the characteristic ratio of vertical to horizontal scales is $(2\Omega\sin\phi/N)(\alpha/\sigma)^{1/2}$, i.e., the response to given $\mathcal{G}$ deepens in the vertical (or narrows in latitude).

In the steady-state limit $\sigma/\alpha\to 0$ the short-wave heating term vanishes, implying that such heating drives no meridional circulation and is precisely balanced by the thermal relaxation, so that the right-hand side of Equation 3.3 vanishes. Furthermore, that second term on the left-hand side of Equation 3.5 also vanishes, so that the vertical velocity at a given latitude is driven only by the applied force (or, more strictly, the latitudinal derivative of the force) at that latitude. Vertical integration of Equation 3.4 in this limit, substitution from Equation 3.1, and application of the requirement that the vertical velocity w stays finite as $z\to\infty$, gives the result (Haynes et al. 1991)

$$\bar{w}^* = -\frac{1}{2\Omega\rho_0(z)\cos\phi}\frac{\partial}{\partial\phi}\left[\frac{\cos\phi}{\sin\phi}\int_z^\infty \rho_0(z')\mathcal{G}(\phi,z')dz'\right], \tag{3.6}$$

implying that, in the steady-state limit, the vertical velocity at a given level is determined only by the wave force directly above that level. Figure 4, taken from Holton et al. (1995), schematically shows the variation between the high-frequency response and the steady-state response.

The transition between the time-dependent and steady limits depends on the spatial structure and location of the force. Assuming that the height scale is Δz and the latitudinal scale is $\Delta\phi$, the estimate of the two terms on the left-hand side of Equation 3.5 implies that the transition between the two limits takes place when

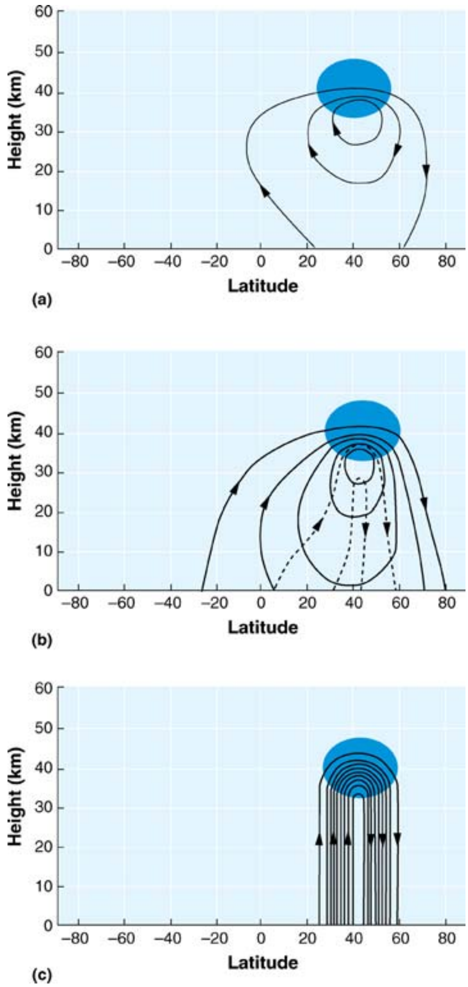

Figure 4 (From Holton et al. 1995). Idealized numerical experiments on the meridional circulation reponse of the stratosphere to a longitudinally symmetric westward force applied in the shaded region. Contours are streamlines, with the same contour interval used in each panel (and corresponding to a given mass flux). (*a*) Adiabatic response for $\sigma/\alpha \gg 1$. A smaller contour interval in this case would show an opposite-signed circulation cell above the shaded region and extension of the lower circulation cell well into the opposite hemisphere. (*b*) Response for $\sigma/\alpha \simeq 0.34$, corresponding to annual frequency and 20-day radiative damping timescale; the solid and dashed contours show the response that is respectively in phase and 90° out of phase with the forcing. (*c*) Steady-state response ($\sigma/\alpha \ll 1$). Note that the magnitude of the response in the mass stream function increases as σ/α decreases and that it is given qualitatively by Equation 3.6 for (*c*). The model is unbounded below so the ordinate has arbitrary origin. (Copyright 1995, Am. Geophys. Union. Reproduced with permission.)

$$\sigma \sim \frac{4\Omega^2 a^2 \sin^2\phi(\Delta\phi)^2}{N^2 \min\{\Delta z^2, \Delta z H\}}\alpha. \tag{3.7}$$

It follows that the timescale required for the steady-state limit to be achieved is longer at low latitudes. Near the equator, the $\sin^2\phi$ should be replaced by $(\Delta\phi)^2$ and rearranging, it follows that for given frequency σ, the steady-state limit is not achieved in a range of latitudes $\Delta\phi_\sigma$ given by

$$\Delta\phi_\sigma \sim \left(\frac{N^2 \min\{\Delta z^2, \Delta z H\}}{4\Omega^2 a^2}\right)^{1/4} \left(\frac{\sigma}{\alpha}\right)^{1/4} \tag{3.8}$$

(Holton et al. 1995).

3.2. Implications for the Mean Meridional Circulation

The sense of the circulations shown in Figure 4 corresponds to a westward force, as is expected to be present in the winter hemisphere of the real stratosphere, as a result of breaking and dissipating Rossby waves. In the high-frequency limit the response to a localized force extends significantly in latitude. This is manifested in the observed response to large wave force, e.g., associated with sudden warming events. It has been known for some time that at times of significant westward wave force, which drives downward motion and hence warming at high latitudes, there is an opposite response at low latitudes and in the opposite hemisphere, with observed cooling that may be explained by the compensating upward motion (e.g., Randel 1993).

The dependence of the response on σ/α shown in Figure 4 has important implications for the seasonal variation of the mean circulation and for the long-time mean circulation. The annual variation is of particular interest. Yulaeva et al. (1994) argued that the well-known annual cycle in tropical lower stratospheric temperatures, with cold temperatures in Northern Hemisphere winter and warm temperatures in Northern Hemisphere summer, was likely due to the annual cycle in the extratropical wave force, with relatively large force in Northern Hemisphere winter compared to Southern Hemisphere winter (for reasons discussed in section 2). For the annual frequency the ratio $\sigma/\alpha \simeq 0.3$ (corresponding roughly to panel b of Figure 4), so it is important to consider the reduction in latitudinal penetration of the meridional circulation as σ/α decreases. On this basis Holton et al. (1995) argue that the wave force relevant for the annual cycle is likely subtropical rather than extratropical. (Equation 3.8 gives an estimate of the relevant range of latitudes.) Randel et al. (2002) made a detailed study of the time variation of tropical upwelling and the relation to extratropical wave force and suggest that the annual cycle response is particularly large because the radiative relaxation timescale α^{-1} in the tropical lower stratosphere is very long (perhaps 100 days). This is consistent with some aspects of their results, but implies a significant difference between the phase of the maximum upward velocity and the phase of the maximum temperature and seems at odds with other observational studies, e.g., Rosenlof (1995). The overall plausibility of the Yulaeva et al. (1994) hypothesis (with the "subtropical" caveat) was demonstrated by Scott (2002) in simulations with an idealized "mechanistic" model of the stratosphere including seasonal variation. (Such models are discussed in more detail in section 4). With no forcing of Rossby waves in the extratropics, and hence no wave force, there is no annual and only a weak semiannual cycle in tropical lower stratospheric temperatures. With forcing of Rossby waves only in one hemisphere (a simple representation of the real interhemispheric difference), there is a significant annual cycle of temperature with the expected phase. However, Plumb & Eluszkiewicz (1999) point out that the observed vertical velocities, which tend to maximize on the summer side of the equator, are not wholly explained by the response to wave forces, and argue that heating, i.e., the annual variation of $\bar{Q}_s$ in Equation 3.3, also needs to be considered, supporting their arguments with simple numerical simulations.

The mean meridional circulation is particularly important on long timescales because of its role in transporting chemical species in the vertical across θ surfaces. In the tropical lower stratosphere there is systematic year-round upwelling, which plays an important role in carrying species with sources in the troposphere, many crucial for ozone chemistry, up into the stratosphere (e.g., WMO 1999, 2003). Considering long timescales naturally focusses attention on the steady-state limit, $\sigma/\alpha \to 0$, in which the result Equation 3.6 holds, as illustrated by Figure 4*c*. This result is discussed in detail by Haynes et al. (1991), who describe it as a "downward control principle" for the effect of wave force $\mathcal{G}$ on the mean meridional circulation. It is important to realize the limitations of this result—the downward control puts no constraint on what caused the wave forces and, in the case of most such forces in the stratosphere, the waves responsible have propagated up from the troposphere—but, nonetheless, it provides useful insights into the workings of the mean meridional circulation. For example, one implication is that forces exerted high in the middle atmosphere, e.g., in the upper stratosphere and mesosphere, may be important in determining the vertical velocity and, hence, temperatures in the lower stratosphere. In practice, the density factor within the integral works against this, but such forces may be important if wave forces in the lower and middle stratosphere are very small, as they may be, for example, in Southern Hemisphere mid-winter. (See further discussion in section 4.) Garcia & Boville (1994) demonstrate an important role for mesospheric gravity wave forces in this case.

The result, as expressed by Equation 3.6, that in the steady-state limit the meridional circulation is driven only by wave forces, not by heating (i.e., $\bar{Q}_s$), has important implications for understanding the time-averaged mean meridional circulation in the stratosphere. The term "extratropical pump" is sometimes used to describe the action of wave forces on the meridional circulation (Holton et al. 1995), with the emphasis on the extratropics due, in part, to the fact that the wave force is apparently strongest there. But, as Plumb & Eluszkiewicz (1999) note, great care is needed in distinguishing tropics from extratropics for this purpose. As implied by the $\sin\phi$ factor in the denominator of Equation 3.6, small wave forcing close to the equator could be highly significant.

However, a more fundamental point here is that the assumptions leading to Equation 3.6 cannot be justified close the equator. First, the linearization about a state of rest neglects (in Equation 3.1) the latitudinal (and vertical) gradients of the mean longitudinal flow u, relative to the Coriolis parameter. Equation 3.6 may be generalized appropriately (Haynes et al. 1991) and implies control that is downward along angular momentum surfaces rather than exactly downward. This allows the possibility of latitudinal, as well as vertical, penetration of the effects of wave forces, particularly near the equator (Dunkerton 1991, Tung & Kinnersley 2001), although this penetration is not likely very important in the lower stratosphere (Plumb & Eluszkiewicz 1999). Nonetheless, Semeniuk & Shepherd (2001b) demonstrate some latitudinal penetration of the effects of a westward force applied in the subtropics. They also explore the possibility of a middle atmosphere Hadley circulation in the sense of Held & Hou (1980), meaning a region where angular momentum is constant and a steady thermally driven circulation is therefore possible without any wave forcing, as suggested by Dunkerton (1989). Semeniuk & Shepherd (2001a,b) find that in the upper stratosphere, but not in the lower stratosphere, such a circulation leads to significant time-averaged upwelling.

Second, if there is weak frictional relaxation then this must inevitably become important near the equator. For example, with the frictional relaxation term $\gamma\bar{u}$ retained, the Equations 3.1–3.4 imply, in the steady state if $\gamma \ll \alpha$, a thin equatorial layer of thickness

$$\Delta\phi_\gamma \sim \left(\frac{N^2 \min\{\Delta z^2, \Delta z H\}}{4\Omega^2 a^2}\right)^{1/4} \left(\frac{\gamma}{\alpha}\right)^{1/4} \tag{3.9}$$

(analogous to the $\Delta\phi_\sigma$ in Equation 3.8) in which frictional relaxation enters the dynamical balance and, in particular, allows a meridional circulation without any local wave force (Plumb & Eluszkiewicz 1999). The 1/4-power dependence on γ makes it essentially impossible to exclude frictional effects from low latitudes in numerical models.

More generally, again as discussed by Plumb & Eluszkiewicz (1999), one key problem of this approach in analyzing the forcing of the mean circulation near the equator is that the wave force $\mathcal{G}$ is treated as given. This is justifiable in the extratropics where the change in velocity due to the applied forcing is relatively small. (It may be calculated via change in temperature implied by the steady state form of Equation 3.3 and thermal wind balance Equation 3.2.) However, the change in velocity increases as the equator is approached and, in the relevant time-dependent problem, a significant part of the wave force must directly drive a change in the mean flow (Haynes 1998). Given that the spatial distribution of the wave force likely depends strongly on any changes to the mean flow, it follows that a nonlinear problem must be solved in which wave force and change in mean flow are deduced together. Similar considerations apply in the subtropical upper troposphere (Held & Phillips 1990). In the mechanistic model simulations reported by Scott (2002) there is a self-consistent calculation of both the waves (and hence $\mathcal{G}$) and the meridional circulation response. The small amount of wave force within 10° or so of the equator appears to play an important role in the momentum balance that allows a broad tropical upwelling, but arises naturally from the breaking and dissipation of the Rossby waves represented in the model. It may well be the systematic westward sign of a Rossby-wave-induced $\mathcal{G}$ and the inevitability of some small penetration of such $\mathcal{G}$ to very low latitudes that accounts for the apparent robustness of the real tropical upwelling.

In summary, the theoretical approach of taking the wave force as specified suggests that, in principle, it would be possible for the force to stop abruptly at say, 20° with, by Equation 3.6, no upwelling equatorward of that. But this situation is not likely relevant to the real atmosphere. The realistic behavior of breaking and dissipating gravity waves and, above all, Rossby waves likely leads to small wave forces at low latitudes.

4. WAVE MEAN FLOW INTERACTION IN THE EXTRATROPICAL STRATOSPHERE

We now consider the implications of combining the longitudinally symmetric dynamics reviewed in section 3 with Rossby wave propagation, as discussed in section 2. The Equations 3.1–3.4 are still relevant, but with the important extra ingredient over the discussion in section 2 that the wave force $\mathcal{G}$ now depends on the $\bar{u}$ and $\bar{T}$ fields because the wave propagation, breaking, and dissipation depend on the background state. Important guiding principles are that Rossby wave propagation is generally inhibited by westward or strong eastward winds $\bar{u}$ and enhanced by strong PV gradients. A relevant model must have some kind of radiatively determined temperature field $T_r(\phi, z, t)$ toward which temperature field is relaxed by long-wave radiative transfer. In Equation 3.3 this would be achieved by choosing $Q_s(\phi, z, t) = \alpha T_r(\phi, z, t)$. The time dependence of T_r allows representation of seasonal variation. The question to address is what dynamical ingredients are needed to account for the observed seasonal and interannual variation of the stratospheric circulation, and the interhemispheric differences in that variation.

A natural simplification is to consider models that are "stratosphere-only" (though in practice to avoid problems with artificial upper boundaries such models usually extend into the mesosphere or higher). However, the effect of tropospheric wave sources must be included. Therefore, stratosphere-only models are usually formulated with an artificial lower boundary, e.g., at about 10 km, and with some representation of tropospheric sources by forcing at that boundary, e.g., through an artificial topography. Such models are often described as mechanistic models. Interhemispheric differences might be captured in such models by comparing large wave forcing (Northern Hemisphere) with small or moderate wave forcing (Southern Hemisphere). A next stage of sophistication is to use a general circulation model (GCM) with a dynamically active troposphere.

In studying the stratospheric circulation it is natural to emphasize two extreme possibilities (Yoden et al. 2002). One is where the stratospheric circulation is governed entirely by the troposphere. In this view dynamical events such as sudden warmings arise as a result of increased wave forcing from the troposphere. In a stratosphere-only model such an event would naturally be initiated by increasing the amplitude of the wave forcing at the artificial lower boundary. The opposite extreme is that of "independent stratospheric variation" in which, in the presence of nonzero but constant tropospheric wave forcing, the nonlinear internal dynamics of the stratosphere give rise to dynamical events such as sudden warmings. Reality appears to be somewhere in between.

Both extremes involve a two-way interplay between waves and mean flow in the extratropical stratosphere, as was first made clear using models with a restricted spatial representation of the different dynamical fields. In particular, Holton & Mass (1976) formulated a quasigeostrophic model in a mid-latitude channel in which (*a*) only a single Fourier mode was used to represent the spatial structure across the channel and (*b*) only the longitudinal mean and a single longitudinal wave mode were retained. The evolution equation reduced to two coupled partial differential equations in z and t, one describing the effect of the wave force $\mathcal{G}$ on the mean flow and the other describing the effect of the mean flow on the wave propagation and the force $\mathcal{G}$. A model incorporating the approximations (*a*) and (*b*) is referred to as a Holton-Mass model.

Holton & Mass (1976) demonstrated for steady radiative conditions (i.e., T_r independent of t) that for small wave forcing at the lower boundary a steady state was possible, whereas for larger forcing the interaction between waves and mean flow led to a vacillating or oscillating state, with the oscillation manifested both in the mean-flow strength and the wave propagation. The Holton-Mass vacillations are a simple example of oscillations arising from two-way interaction between waves and mean flow for Rossby waves in the extratropics (with the relaxation to the radiatively determined state playing an important role) and in that sense are the extratropical analogue of Plumb's simple model of the tropical quasibiennial oscillation (Plumb 1977). Holton & Mass argued that sudden warmings might be interpreted as the manifestation of vacillations arising from the internal dynamics of the stratosphere (for fixed but sufficiently large wave forcing). Yoden's (1987) detailed mathematical analysis of the Holton-Mass model revealed the bifurcation structure of the model and clearly revealed the possibility of multiple steady states and bifurcations from a steady state to an oscillating state, and later analysis by Christiansen (2000) revealed further bifurcations leading to chaotic solutions in some parameter regimes.

Many recent studies suggest that the phenomena exhibited by the Holton-Mass model can be identified in more realistic models. Scaife & James (2000) and Scott & Haynes (2000) demonstrated the existence of multiple steady states and vacillations in stratosphere-only models, with fixed lower-boundary wave forcing, and noted the role of latitudinal, as well as vertical, wave propagation. Boville (1986) gave some indication of multiple steady states, one a strong-flow, weak-wave state and the other a weak-flow, strong-wave state, in a perpetual-January GCM simulation including a realistic troposphere. Christiansen (1999) also showed stratospheric vacillations in a GCM and furthermore demonstrated that they were independent of tropospheric variability and disappeared when the Northern Hemisphere topography was removed, consistent with the Holton & Mass finding that the vacillations were present only for sufficiently strong tropospheric wave forcing.

A natural next stage is to consider stratosphere-only models in which the lower-boundary wave forcing remains fixed, but a seasonal cycle is imposed in the relaxation temperature field. The Holton-Mass model was investigated in this context by Yoden (1990), who argued that a quasisteady analysis provided a useful qualitative guide, but not a complete guide, on which to base interpretation of the time-dependent evolution, and showed that for moderate-amplitude wave forcing the flow began the winter in a weak-flow, strong-wave regime and then made a rapid transition to a strong-flow, weak-wave regime. Near the end of winter there was corresponding reverse transition. This captured some aspects of the observed Southern Hemisphere behavior. For large-amplitude forcing the flow throughout the winter remained in the weak-flow, strong-wave regime, with associated vacillations, a simple representation of Northern Hemisphere behavior. (Recall the interhemispheric differences shown in Figure 1). Scott & Haynes (2002), following on earlier studies by Holton and Wehrbein (1980, 1981), conducted a similar study in a model with latitudinal structure. This model also exhibited strong differences between moderate wave forcing, where the mean flow remained close to the radiative equilibrium flow throughout the winter, and strong wave forcing, where sudden warmings take place in mid or even early winter (earlier with larger wave forcing). Scott & Haynes also showed that two different late winter states, one with strong flow, one with weak flow, were possible for the same value of late winter wave forcing, and were achieved by different early winter forcing.

A different angle on the idea of multiple states and sensitivity of evolution is provided by Gray et al. (2003), who carried out mechanistic model integrations with fixed lower boundary wave forcing, applying radiative conditions for perpetual mid-winter. For each value of wave forcing the model was integrated from many different initial conditions, all corresponding to late summer, but differing in small details. For small forcing amplitudes all cases tended toward a strong flow state. For large forcing amplitudes all cases gave repeated warmings and hence exhibited a weak flow state. A range of intermediate forcing amplitudes showed considerable variation from one initial condition to the other, some giving sudden warmings and others (over a period of 200 days or so) remaining in a strong flow state.

The mechanistic models also suggest the possibility of interannual variability of the stratospheric circulation arising through the internal dynamics of the stratosphere (i.e., with lower-boundary wave forcing and radiation that repeat identically each year). Even in the Holton-Mass model, the chaotic behavior found by Christiansen (2000) is almost certain to imply interannual variability if an annual seasonal cycle is included. A clear mechanism for interannual variability in a model with unconstrained latitudinal structure was found by Scott & Haynes (1998), who demonstrated, for an intermediate range of values of wave forcing, a spontaneous biennial oscillation in which weak wave, strong flow and strong wave, weak flow winters alternated. An important part of the mechanism was the long damping timescale for $\bar{u}$ anomalies at low latitudes, arising from the weakness of the associated $\bar{T}$ anomalies (consistent with the reasoning in section 3 leading to Equation 3.8) endowing the low-latitude $\bar{u}$ field with an interannual memory, like a "flywheel."

In the mechanistic models just described the wave forcing is applied as a specified perturbation at the artificial lower boundary, but the flux of wave activity through the lower boundary varies strongly (e.g., Dunkerton et al. 1981, Gray et al. 2003, Scott & Haynes 2002), almost certainly as a result of partial backreflection of Rossby waves within the stratosphere. Some aspects of the behavior may be a consequence of the artificial lower boundary condition, but the presence of vacillation behavior in models that include a troposphere (Christiansen 1999, Scott & Polvani 2004) suggests that, in reality, it is possible for the stratosphere to play a role in determining the net wave flux it receives from the troposphere.

The interplay between dynamical variability imposed by the troposphere and the two-way interaction between waves and mean flow in the stratosphere was explored by Taguchi et al. (2001) and Taguchi & Yoden (2002a,b) in a series of simulations with a simplified GCM. Large-scale Rossby waves were generated both directly by a simple topography and also through nonlinearity by baroclinic eddies in the model troposphere. In perpetual January conditions large-amplitude topographic wave forcing gave a vacillation regime in the stratosphere, though, because of the highly variable tropospheric circulation, the vacillations were more irregular than typically seen in mechanistic models (Taguchi et al. 2001). With an imposed seasonal cycle (Taguchi & Yoden 2002a) weak/intermediate topographic wave forcing gave significant interannual variability only in late winter (typical of Southern Hemisphere), whereas stronger forcing gave strong interannual variability throughout the winter (typical of Northern Hemisphere). Taguchi & Yoden (2002b) examined the year-to-year interannual variability more carefully and showed that there was no indication of any year-to-year memory, suggesting that the mechanism for interannual variability described by Scott & Haynes (1998) is overwhelmed by the effect of tropospheric variability.

The overall picture from mechanistic-model and GCM studies is that the interplay between Rossby waves and mean flow in the winter stratosphere implies both sensitivity and variability. The sensitivity is not of the precise form implied by the simplest models of the Holton-Mass type, but is important nonetheless and strongly modulates the variability imposed by the troposphere. The sensitivity of the dynamical behavior plays a role in the "cold pole" problem of numerical models of the stratosphere, where errors in radiation schemes or the absence of gravity wave drag can lead to very large errors in the circulation (e.g., Boville 1995). The variability, particularly in view of evidence for natural variability of the coupled troposphere-stratosphere system on timescales out to at least a decade (e.g., Butchart et al. 2000), makes it very difficult to identify systematic change (e.g., due to radiative effects of changes in stratospheric ozone purely from observations) and implies that any numerical modelling study requires long integrations or large ensembles to ensure statistical significance (WMO 2003).

5. STRATOSPHERE-TROPOSPHERE COUPLING

Traditionally, the dynamical coupling between the troposphere and the stratosphere has been considered a one-way effect of the troposphere on the stratosphere, with the troposphere acting as a source of Rossby waves and gravity waves, which propagate upward into the stratosphere. However, observational and modelling evidence is now so strong that the stratosphere does not always play a passive role in the coupled dynamics. Indeed, it may sometimes be appropriate to regard changes in the stratosphere as causing changes in the troposphere.

Much of the evidence focuses on the Northern Hemisphere and Southern Hemisphere annular modes (NAMs and SAMs, respectively) (e.g., Thompson & Wallace 2000), which are dominant signals in variability in the troposphere and believed to arise primarily from two-way interaction between baroclinic eddies and the tropospheric mid-latitude jet (e.g., Feldstein & Lee 1998, Lorenz & Hartmann 2003, Robinson 1991). In Northern Hemisphere winter there is significant correlation between the annular mode index defined on the basis of the surface pressure field and the stratospheric circulation, so that, for example, when there is a strong pole-to-equator pressure gradient at the surface, indicating strong eastward surface flow, there are also strong eastward winds throughout the mid-latitude troposphere and in the mid-to-high latitude stratosphere (Thompson & Wallace 2000). There is corresponding organization in the wave fluxes, indicating propagation and momentum transport, both in the troposphere (Limpasuvan & Hartmann 2000) (as is expected from the accepted mechanism for the variability) and also in the stratosphere (Hartmann et al. 2000) (see Figure 5), implying an associated correlation in the wave force $\mathcal{G}$ in the stratosphere.

Baldwin & Dunkerton (1999, 2001) showed that the vertical structure of NAM variation in Northern Hemisphere winter typically shows a downward progression from middle stratosphere to troposphere (see Figure 6). Does Figure 6 imply a direct effect, with some delay, of anomalies in the circulation in mid-stratosphere on the troposphere?

There are several possible mechanisms by which the stratosphere might affect the troposphere. One possible mechanism is via the nonlocal PV inversion operator. Any change in the PV in the lower stratosphere will instantaneously give rise to changes in wind and temperature in the troposphere. Hartley et al. (1998), Ambaum & Hoskins (2002), and Black (2002) show explicit calculations to illustrate this point. This might account for the lower part of the pattern in Figure 6. A precisely equivalent statement, applied to longitudinal mean fields, is that a wave force $\mathcal{G}$ localized to the lower stratosphere will, through the instantaneous induced meridional circulation, give rise to an acceleration in the troposphere below.

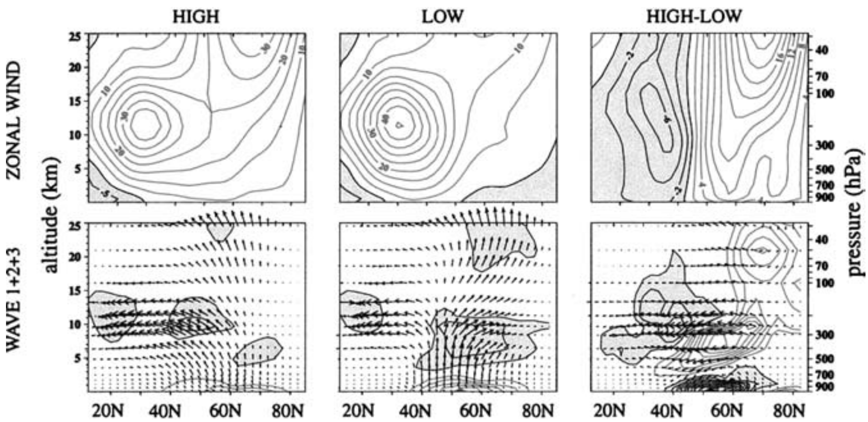

Figure 5 (From Hartmann et al. 2000.) Composites for periods of high and low NAM index and their difference (*left*, *center*, and *right*, respectively) in longitudinal wind (*top*) and Eliassen Palm flux, which indicates wave propagation and transport of westward momentum, and its divergence, which indicates eastward wave force. Positive contours are gray, negative contours are black, and negative regions are shaded. The Eliassen Palm flux is calculated only for longitudinal wavenumbers 1, 2, and 3. In the "high" phase wave fluxes tend to be directed equatorward within the troposphere and to converge in the subtropical troposphere, whereas in the "low" phase wave fluxes tend to be directed upward from troposphere to stratosphere and to converge, implying an anomalous westward wave force, in the mid- and high-latitude stratosphere. (Copyright 2000, Natl. Acad. Sci, U.S.A. Reproduced with permission.)

Furthermore, as noted in section 3, on longer timescales in the presence of radiative damping, the meridional circulation tends to be narrower and deeper below, potentially allowing an enhanced tropospheric response to a stratospheric wave force. Song & Robinson (2004) suggest that the downward penetrating response in the mean circulation communicates the effect of stratospheric wave forcing to the troposphere, where the response is amplified by the eddy (i.e., wave) feedbacks associated with annular variability.

A different mechanism for communication in the vertical is via Rossby wave propagation. Thus, one interpretation, probably the most obvious interpretation, of the patterns shown in Figure 5 is that upward propagating Rossby waves communicate the effect of annular mode changes in the troposphere to the stratosphere. But the changes in the wave flux might be caused by variation in the refractive properties of the lower stratospheric flow (Hartmann et al. 2000, see also Limpavusan & Hartman 2000), or by downward reflection from higher in the stratosphere (e.g., Perlwitz & Harnik 2003). Song & Robinson (2004) found that the effect of stratosphere on troposphere in their model simulations is significantly reduced when planetary waves are damped in the stratosphere and on this evidence rejected their original hypothesis of downward penetration of the mean circulation and argued

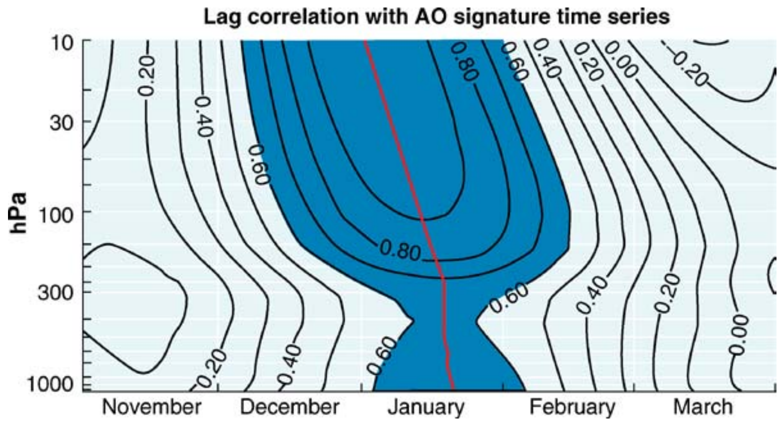

Figure 6 (From Baldwin & Dunkerton 1999.) Correlations between the timeseries at each level of a measure of the NAM with that at 10 hPa. (10 hPa corresponds roughly to 30 km, 100 hPa to 15 km.) There is clear downward progression of the correlation in the stratosphere, but this may be "phase propagation" rather than propagation of information. (See text for further details.) (Copyright 1999, Am. Geophys. Union. Reproduced with permission.)

that Rossby waves likely play a significant role in downward communication of information.

A further possible mechanism for downward communication of information is two-way interaction between waves and mean flow. However, Plumb & Semeniuk (2003) demonstrated that downward patterns similar to the upper part of the Baldwin & Dunkerton pattern shown in Figure 6 (which are also similar to those seen in the vacillation behavior described in section 4) arise naturally, in a Holton-Mass type model, as a response to forcing in the lower stratosphere. Indeed, in simple one-dimensional models of the equatorial quasi-biennial oscillation, which exhibit similar downward progressing features, one may argue with certainty that no downward propagation of information is possible (Plumb 1977). This argument is possible because of the neglect of rotation (the Coriolis parameter is zero at the equator), implying a local relation in the vertical between wave forcing and acceleration and the assumption that the waves may be modelled using the slowly varying approximation. In the extratropical stratosphere neither of these assumptions can be justified—in particular the relevant waves are Rossby waves, which have large vertical wavelengths and for which the slowly varying approximation is not justified. The Plumb & Semeniuk (2003) evidence notwithstanding, this leaves the possibility that there is some real downward causal influence. Indeed, numerical experiments that investigate the effect of perturbations in the upper stratosphere, representing solar variability effects, on the extratropical stratospheric circulation (e.g., Gray 2003, Kodera et al. 1990) in many cases show that the effects of those changes propagate downward and are significant in the lower stratosphere.

Other evidence that the stratosphere plays an active, rather than a passive, role in tropospheric variations associated with the NAM comes from observed and modelled changes in the timescale of these variations. Baldwin et al. (2003) showed that this timescale is significantly longer at times of the year (Northern Hemisphere winter, Southern Hemisphere spring) when there is strong Rossby wave propagation into the stratosphere. Correspondingly, artificial suppression of stratospheric variability in model simulations reduces the timescale of the tropospheric NAM (Norton 2003). The dynamical mechanism here is likely that, when there is significant flux of Rossby waves into the stratosphere, the flow in the stratosphere acts as an integrator (and hence low-pass-filter) of this flux (or rather the variability in this flux) because stratospheric damping times are relatively long. Any stratospheric effect on the troposphere therefore tends to increase the timescales of the variability. When there is little flux of Rossby waves into the stratosphere (in summer, or in Southern Hemisphere mid-winter) the effect is absent.

6. CONCLUSION

Efforts to predict future changes in the stratosphere are inevitably moving toward coupled chemistry-climate models (Austin et al. 2002). But dynamics remains a central issue. The patterns of variability and likely dynamical links between stratosphere and troposphere discussed in section 5 are part of increasing evidence that the stratosphere may play an active role in the tropospheric circulation, perhaps leading to changes in the troposphere as a result of stratospheric ozone changes (Hartmann et al. 2000) or stratospheric injections of volcanic aerosol (Robock 2000). These links may also offer scope for improving the skill of extended-range tropospheric forecasts (Baldwin et al. 2003).

A key outstanding dynamical problem in climate modelling for the stratosphere is reliable prediction of changes in the wave force $\mathcal{G}$. There is evidence from some numerical simulations with general circulation models that an increase in carbon dioxide implies an increase in planetary wave propagation into the stratosphere and hence an increase in the strength of the mean meridional circulation (e.g., Butchart & Scaife 2001, Rind et al. 2002 and references therein). However, this predicted increase is not certain—for example, Gillett et al. (2003) show an increase in wave propagation into the stratosphere only in the Northern Hemisphere, whereas Butchart & Scaife (2001) show changes in both hemispheres. Further (and possibly greater) potential uncertainty enters through the extreme difficulty in simulating possible changes in gravity wave sources in the troposphere.

ACKNOWLEDGMENTS

The author is particularly grateful to Michael McIntyre for detailed comments on a draft of this review. Other useful comments were received from Pavel Berloff, Steven Hardiman, and David Thompson.

The *Annual Review of Fluid Mechanics* is online at http://fluid.annualreviews.org

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