The Atmospheric General Circulation and its Variability

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Abstract

Progress in understanding the general circulation of the atmosphere during the past 25 years is reviewed. The relationships of eddy generation, propagation and dissipation to eddy momentum fluxes and mean zonal winds are now sufficiently understood that intuitive reasoning about momentum based on firm theoretical foundations is possible. Variability in the zonal-flow can now be understood as a process of eddy, zonal-flow interaction. The interaction of tropical overturning circulations driven by latent heating with extratropical wave-driven jets is becoming a fruitful and interesting area of study. Gravity waves have emerged as an important factor in the momentum budget of the general circulation and are now included in weather and climate models in parameterized form. Stationary planetary waves can largely be explained with linear theory.

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1. Introduction

I attempt here to give a summary of progress in understanding the general circulation of the atmosphere in the past 25 years, in celebration of the 125th anniversary of the Meteorological Society of Japan. This is a daunting task, especially within the confines of an article of reasonable length. I will therefore concentrate on a few areas that seem most interesting and important to me and with which I am relatively familiar. I will not attempt to give a full list of references, but will attempt to pick out a few important early references and some more recent references. Interested readers can use these data points to fill in the intervening developments.

The general circulation is defined to be the complete statistical description of large-scale atmospheric motions. A complete understanding of the general circulation requires an understanding of the role of small-scale motions, radiation, convection and interaction with the ocean and land surface. One is then nearly dealing with the problem of climate, which is too broad for this review. I will therefore focus primarily on large-scale dynamics, but will from time to time indicate where smaller scales or non-dynamical processes are critical to understanding the large-scale flow. I will focus primarily on the lower part of the atmosphere, including the troposphere and stratosphere.

2. The zonal mean and eddy viewpoint

One of the earliest questions about the general circulation of the atmosphere was why the surface winds tend to be westerly in midlatitudes and easterly in the tropics in all oceans. The first theories for this were posed in a longitude-independent framework (see reviews by Lorenz (1967) and Schneider (2006) for historical references). Although science has advanced to incorporate a fully three-dimensional view of the general circulation, division of the flow into a zonal mean and departures from the zonal mean, or eddies, is still a compact and useful viewpoint that has provided new insights in the past 25 years.

2.1 Zonal mean statistical climatology

Figure 1 shows the zonally averaged zonal wind for two seasons, averaged over the period from 1957-2002 from ERA-40 (Uppala *et al.* 2005). Near the surface the winds are easterly in the tropics, especially in the winter hemisphere. In the winter hemisphere at about 200 hPa the westerly winds form a jet at about 30˚ latitude, which is often called the subtropical jet. A tropospheric jet is also evident in the summer season, but it is displaced poleward to about 45N and 45S. The Southern Hemisphere exhibits strong surface westerlies at about 50S in all seasons. In the austral summer season (DJF) the surface westerly wind maximum is aligned with the upper level westerly jet. This alignment of surface and upper level westerly wind maxima is a signature of an eddy-driven jet, where the westerly winds are sustained by the meridional convergence of eddy momentum fluxes.

In the extratropical stratosphere the zonal wind is westerly in winter and easterly in summer. The winter stratospheric jet is much weaker in the Northern Hemisphere, because the larger stationary planetary wave forcing there forces strong wave, mean-flow interaction during winter and spring, which warms the polar stratosphere and weakens the polar stratospheric jet. The equatorial winds are easterly in the stratosphere in both seasons because the thermal forcing gradient reverses with the seasons and a pole-to-pole circulation exists in the solstitial seasons (Holton *et al.* 1995).

A useful understanding of the momentum budget of the upper troposphere can be gained by considering the zonal mean wind equation for a nondivergent fluid.

$$\frac{\partial \bar{u}}{\partial t} = f\,\bar{v} - \frac{\partial}{\partial y}\left(\overline{u'v'}\right) - \alpha\,\bar{u} \qquad (1)$$

Three terms are shown in the tendency equation for zonal wind. The Coriolis term indicates that poleward motion will cause a westerly acceleration. The eddy flux convergence term indicates that covariance between poleward and eastward eddy velocities can move momentum, and that where this eddy momentum flux converges a zonal acceleration can be induced. The frictional drag term postulates that frictional drag will always oppose the zonal flow, with a drag coefficient α. If the baroclinic version of (1) is averaged vertically over the mass of the atmosphere, the Coriolis term disappears and any tendency or frictional drag must be balanced by the eddy flux convergence term. Since the extratropical westerly wind maximum can only be sustained against drag by eddy momentum fluxes, the extratropical jets in Figure 1 must be eddy-driven.

The mean meridional wind is shown in Figure 2. In the upper troposphere strong flow from the summer to the winter hemisphere extends from the summer tropics to the winter subtropics. This meridional motion induces an easterly acceleration in the summer hemisphere and a westerly acceleration in the winter hemisphere (1). If angular momentum were conserved along a trajectory from the summer Tropics to 30 degrees in the winter hemisphere, we would expect easterlies at the equator and a strong westerly jet at 30 degrees in the winter hemisphere. This reasoning gives a subtropical jet in about the right position, but much too strong. Near the surface, meridional flow from the winter to the summer hemisphere provides a westward Coriolis acceleration necessary to produce the observed tropical easterly winds in the winter hemisphere. In the extratropics, the mean meridional winds are weak and in the opposite direction necessary to explain the zonal winds at upper levels, and it is again clear that the eddy fluxes of momentum play a central role there.

The total eddy meridional flux of zonal momentum is shown in Figure 3. Five centers of maximum eddy flux magnitude are present in the troposphere during both seasons. In the extratropics of each hemisphere a poleward eddy flux maximum is centered at 30-40 degrees of latitude, while an equatorward eddy flux maximum appears in high latitudes. The larger, subtropical flux center dominates, and its effect is to move momentum out of the tropics and supply it to the extratropics. The midlatitude westerlies would not exist without this momentum flux, so we think of the midlatitude westerlies as eddy-driven, while the divergence of eddy momentum flux in the tropics offsets the westerly Coriolis accelerations associated with the Hadley circulation, and eddies thereby weaken the subtropical jet.

An interesting feature is the maximum eddy flux over the equator, which takes momentum in the opposite direction to that of the mean meridional wind. This momentum flux is mostly contributed by stationary waves and indicates wave propagation in the same direction as the mean meridional wind. This wave propagation plays an important role in keeping the response to tropical stationary wave forcing symmetric across the equator, even though the heating that drives the tropical stationary waves is primarily in the summer hemisphere, as will be discussed in section 4.

The poleward transport of heat in middle latitudes is provided primarily by atmospheric eddies (Oort 1971; Trenberth and Stepaniak 2004). Figure 4 shows the covariance between eddy meridional wind and temperature, which is proportional to the meridional flux of sensible heat by eddies. Eddy sensible heat flux occurs mostly in the lower troposphere and has its largest magnitudes near 50N and 50S. Most of this heat flux is associated with baroclinic eddies, especially in the Southern Hemisphere. The flux has a large seasonal cycle in the Northern Hemisphere, but is nearly independent of season in the Southern Hemisphere, where the Southern Ocean maintains a strong meridional gradient of surface temperature in all seasons. A secondary maximum in poleward eddy heat flux occurs above the tropopause, which is associated with the baroclinic waves that cannot propagate into the stratosphere and whose amplitudes decay quickly in the lower stratosphere (Holton 1974). In the winter stratosphere the heat flux extends deeply into the stratosphere because planetary waves can propagate into the winter westerlies (Matsuno 1970). This heat flux is larger in the Northern Hemisphere, where the forcing of planetary waves by surface asymmetries is much larger than in the Southern Hemisphere.

2.2 Theoretical understanding of mid-latitude eddies and jets

When Lorenz wrote his classic monograph on the general circulation (Lorenz 1967), it was not understood why midlatitude eddies transport momentum preferentially poleward. As a result of work done in the late 1970's and subsequent refinements, we now have a better theoretical understanding of the relationship of meridional and vertical eddy fluxes of momentum to wave generation and propagation (Andrews and McIntyre 1976a; b). From the zonal mean momentum balance for a non-divergent barotropic fluid one can show that

$$\frac{\partial \bar{u}}{\partial t} = -\frac{\partial}{\partial y}\left(\overline{u'v'}\right) = \overline{v'\zeta'} \qquad (2)$$

The enstrophy budget for this system would be,

$$\frac{d}{dt}\left(\frac{1}{2}\overline{\zeta'^2}\right)+\beta_{eff}\,\overline{v'\zeta'}=\overline{F'\zeta'} \tag{3}$$

where F is some forcing of the vorticity field, such as the generation of eddies by baroclinic instability, and β_{eff} is the gradient of mean absolute vorticity. If this forcing adds enstropy in a stirring region, and the enstrophy remains steady, then an up-gradient flux of vorticity must occur in this stirring region. If eddies propagate outside the stirring region and increase the enstrophy there, then the vorticity flux must be down-gradient in those regions where the eddies arrive. Thus, in the presence of a positive gradient of mean vorticity, stirring of vorticity will result in a down-gradient flux of vorticity in regions removed from the stirring, and consequent easterly accelerations on either side of the stirring region, and a westerly acceleration within the stirring region. Thus we expect that a westerly acceleration should occur in a region of stirring, if waves are able to propagate away from the stirring zone (Held 1975).

This result can be generalized to quasi-geostrophic flow on a sphere (Edmon *et al.* 1980),

$$\frac{\partial \overline{u}}{\partial t}-f\,\overline{v}^{*}=\overline{v'q'}=\left(\rho_0\, a\cos\varphi\right)^{-1}\nabla\bullet\mathbf{F} \tag{4}$$

where q is the quasi-geostrophic potential vorticity and $\overline{v}^{*}$ is the residual mean meridional velocity (*c.f.* section 3). We obtain in (4) the additional benefit of expressing the potential vorticity transport as the divergence of a vector **F** that can be shown to be parallel to the group velocity for Rossby Waves. The structure of the Eliassen-Palm Flux vector, **F** (hereafter EP flux), shows the relationship between meridional wave propagation and eddy fluxes of momentum and heat.

$$F_\varphi = -\rho_0\, a \cos\varphi\, \overline{u'v'} \qquad (5)$$

$$F_z = f \rho_0\; a \cos\varphi\, \overline{v'\theta'} / \overline{\theta}_z \qquad (6)$$

The meridional component of the eddy wave activity flux is proportional to the negative of the meridional momentum flux (5) and the vertical component is proportional to the poleward eddy heat flux (6). The poleward heat flux represents a downward eddy momentum flux that is better understood as form drag in isentropic coordinates (Andrews 1983). As Rossby waves propagate meridionally away from source latitudes, zonal momentum is transported into the source region by departing eddies. This property of quasi-geostrophic eddies in the atmosphere is critical to understanding the momentum balance of the atmospheric general circulation and its variability. Eastward angular momentum is transferred from the surface to the atmosphere in the tropical easterlies. Angular momentum moves upward and poleward in the meridional overturning circulations in the tropics and this momentum is carried poleward and downward as a concomitant of the upward and equatorward propagation of eddies generated near the surface in midlatitudes.

The principle sources of eddy activity in midlatitudes are baroclinic instability and forcing of waves by zonal variations of surface topography and surface heat sources. Linear wave theory indicates that waves generated in the extratropics propagate preferentially toward the equator (Matsuno 1970), giving a predominantly poleward eddy momentum flux. Even waves that initially propagate toward the pole will eventually turn equatorward, if they are not damped first (Hoskins and Karoly 1981; Karoly and Hoskins 1982). The speed and direction of the eddy propagation depends on the spatial structure and frequency of the waves (Hoskins *et al.* 1983). Even in the absence of stationary wave forcing from surface features, generation of energy by baroclinic instability will result in a rich spectrum of waves from rapidly growing fast baroclinic waves to quasi-barotropic, quasi-stationary waves. Wave energy generated by baroclinic instability rapidly collects on barotropic waves (Simmons and Hoskins 1978b), and eddy energy tends to become equally partitioned between potential and kinetic forms (Schneider and Walker 2006).

A rich spectrum of wave activity results from the conservation of energy and enstrophy in a shallow layer of fluid. Fjortoft (1953) showed that in a two-dimensional fluid energy must cascade to both larger and smaller scales, and this result was extended to quasi-geostrophic flows by Charney (1971). The largest-scale and least damped structures in the atmosphere are zonal jets, and random fields of motion that are weakly damped will eventually organize into zonal jets (Williams 1979a; b; Yoden and Yamada 1993). Baroclinic regions that are broader than the Rossby radius of deformation also organize into zonal jets and associated storm tracks (Panetta 1993). The up-scale turbulent energy cascade is interrupted by a transition to wavelike behavior (Rhines 1975), and observations suggest that the inertial cascade stops around total wavenumber 8-10 (Boer and Shepherd 1983). Transition from two-dimensional turbulence to zonal jets is aided by the collection of energy onto Rossby waves and the tendency of Rossby waves to duct energy westward in zonally elongated structures (Rhines 1994).

The scale at which turbulence transitions to wavelike behavior marks a sharp reduction in the energy cascade rate. This scale can be estimated from the linear vorticity balance in the limit of small frequency,

$$\bar{u}\frac{\partial \zeta}{\partial x} + \beta v = 0 \tag{7}$$

or introducing a Fourier representation for streamfunction, $\psi = \Psi\, e^{i(kx+ly)}$,

$$k^2 + l^2 = \beta / \bar{u} \qquad (8)$$

we obtain the scaling for quasi-stationary Rossby waves on a beta-plane. For zonally invariant flow with k=0, (8) gives a meridional wavenumber of $l = \sqrt{\frac{\beta}{\bar{u}}}$, the Rhines scale, which suggests that Earth should have ample room for one eddy-driven jet between the equator and the pole, but not for two. For Earth, a meridional scale equivalent to the Rossby Radius of deformation gives a similar estimate for the eddy scale and jet spacing (Held and Larichev 1996; Schneider 2004). On larger planets with more rapid rotation such as Jupiter, many jets are expected. Barotropic decay experiments by Huang and Robinson (1998) show that the strongest interactions occur between jets and relatively small-scale eddies, and that the role of the eddies at the Rhines scale in energy exchange is modest, although they contain a large amount of energy. Schneider and Walker (2006) suggest that the baroclinic eddies interact with the thermal structure to make the atmosphere quasi-linear and inhibit the turbulent energy cascade as traditionally envisioned.

From these considerations, if no zonal variations in the forcing of the atmosphere existed, we would still expect a full spectrum of waves interacting with zonal jets. Although the scales associated with baroclinic instability are prevalent, quasi-stationary Rossby wave structures are present and dominate the low-frequency wave variability even in the absence of stationary eddy forcing (Yu and Hartmann 1993). In addition, the cascade of wave activity in the atmosphere, combined with the influence of stationary zonally asymmetric forcing, can lead to phenomena such as blocking (Tanaka and Terasaki 2006).

Since disturbances in the atmosphere tend to behave in a wavelike manner, one can use wave propagation theory as an aid understanding the flux of momentum by eddies. A test of this idea is illustrated in Figure 5, taken from Randel and Held (1991), which shows the transient eddy momentum flux as a function of latitude and phase speed. Most of the eddy momentum flux is provided by eastward traveling waves with zonal wave numbers 4-7. Linear wave theory suggests that waves should not be able to propagate past critical lines where their phase speed equals the zonal wind speed, and that the EP flux convergence should be confined to a narrow range around the critical line. Figure 5 indicates that the meridional momentum flux is concentrated in the region where the phase speed is less than the zonal wind speed, as expected, but the absorption of wave energy implied by the convergence of the momentum flux occurs 10˚-20˚ in latitude away from the critical line. To understand why the waves rarely reach their critical lines requires consideration of nonlinear processes.

It is apparent that equatorward propagating waves often break through nonlinear processes before they reach their critical lines. This is particularly clearly demonstrated in idealized simulations of baroclinic wave growth and equilibration (Simmons and Hoskins 1978a), and the nonlinear wave breaking aspects of these simulations show interesting sensitivities, particularly to the barotropic shear of the initial flow field (Thorncroft *et al.* 1993; Hartmann and Zuercher 1998). The use of potential vorticity as a framework for understanding the dynamics of the general circulation has contributed to progress (Hoskins *et al.* 1985; Hoskins 1991). Descriptive work using potential vorticity fields in the stratosphere clearly shows the effect of wave breaking on the absorption of wave energy on the equatorward margin of the westerlies (McIntyre and Palmer 1983). The stretching and deformation of vorticity by breaking Rossby waves is now believed to be the principle means of wave dissipation, but theory and observation suggest that wave reflection at critical lines is also possible (Magnusdottir and Haynes 1999; Abatzoglou and Magnusdottir 2004).

3. Variability of the zonal mean flow

In addition to including climatological features of interest, a zonally averaged view also includes the dominant mode of unforced low-frequency variability in the extratropics. The strong contribution of zonally averaged structures to variability arises both because zonal mean winds are weakly damped and because baroclinic eddies can provide a strong feedback to sustain jet anomalies. Because for Earth the midlatitude baroclinic zone is a bit wider than necessary for a single eddy-driven jet, the jet can move north and south within the baroclinic zone and this behavior arises even in simple models of the general circulation (Robinson 1991; James and James 1992; Yu and Hartmann 1993; Lee 2005). Since the baroclinic zone associated with the upper level jet is the source of eddies, we expect a source of baroclinic eddies in close association with the jet. If these eddies are able to propagate away from the jet, then they will provide a flux of momentum that will sustain the jet, via mechanisms described in section 2.2.

Figure 6 shows a sketch illustrating the first order momentum and heat budgets for an eddy-driven jet in the Eulerian frame of reference. At upper levels (1) holds approximately with $\alpha \approx 0$, and momentum flux convergence is balanced by an equatorward mean meridional wind. Near the surface the eastward acceleration associated with poleward mean meridional velocity is balanced by surface drag.

$$f\,\bar{v} \approx \text{surface drag} \approx \alpha\,\bar{u} \tag{9}$$

The heat budget at any level is given by,

$$\frac{\partial\bar{\theta}}{\partial t} = -\frac{\partial}{\partial y}\left(\overline{v'\theta'}\right) - \bar{w}\frac{\partial\bar{\theta}}{\partial z} + \bar{Q} \tag{10}$$

so that the eddy heat flux convergence approximately balances the sum of diabatic heating, Q, and adiabatic heating associated with mean vertical motion. The heat and momentum balances are closely linked through the mass balance and the mean meridional circulation. The eddy fluxes of heat and momentum support a reversed meridional circulation cell, suggesting that the heat transport by eddies is more than sufficient to balance the diabatic heating term.

An interesting question to pursue is whether eddy-driven jets are self-sustaining and can shape the climate, or whether they are a passive response to imposed diabatic heating gradients. Reasoning about the dynamical role of jet-eddy interactions is more fruitful in the transformed Eulerian mean framework, in which the zonal momentum equation and heat equations are (e.g. Andrews et al. 1987) ,

$$\frac{\partial\bar{u}}{\partial t} - f\,\bar{v}^{*} = \left(\rho_0 a\cos\varphi\right)^{-1}\nabla\bullet\mathbf{F} \tag{11}$$

$$\frac{\partial\bar{\theta}}{\partial t} + \bar{w}^{*}\frac{\partial\bar{\theta}}{\partial z} = \bar{Q} \tag{12}$$

where $\nabla\bullet\mathbf{F}$ is the divergence of the EP flux vector whose components are defined by (5-6), and

$$\bar{w}^* = \bar{w} + \frac{\partial}{\partial y}\left(\overline{v'\theta'}\left(\frac{\partial \bar{\theta}}{\partial z} \right)^{-1} \right) \tag{13}$$

is the residual vertical velocity, an approximation to the diabatic vertical velocity.

In addition, one can manipulate these equations to form the downward control principle, which in the quasi-geostrophic approximation and under an assumption of steady state conditions can be written (Haynes *et al.* 1991).

$$\bar{w}^* = -\frac{1}{a^2 \rho_0 \cos\phi}\frac{\partial}{\partial \phi}\left\{ \int_z^{\infty} \left\{ \frac{\nabla \bullet \mathbf{F}}{2\Omega \sin\phi} \right\}_{\phi = const.} dz' \right\} \tag{14}$$

The downward control principle (14) indicates that the residual vertical velocity is proportional to the meridional gradient of the integrated wave driving *above* that level.

Tanaka et al. (2004) have computed the residual mean circulation using more precise isentropic analysis, and their results for two seasons are shown in Figure 7. The mass stream function has a large center near the equator that represents the Hadley Cell, and smaller secondary centers in midlatitudes that are driven by eddy form drag, mostly by transient baroclinic eddies. The contribution of transient eddies to forcing the mass streamfunction is illustrated in the lower portion of Figure 7.

Most of the wave driving in midlatitudes is associated with the convergence of the upward component of the EP flux vector, which is associated with baroclinic wave growth and strong form drag in isentropic coordinates. Baroclinic eddies produce a strong downward flux of zonal momentum and a strong easterly acceleration in the middle and upper troposphere. Note that in this framework the surface winds are not maintained by the meridional circulation, but rather by the strong eddy form drag in the jet region that transports westerly momentum downward to the surface where it balances friction. The wave drag in turn by (14) drives the residual midlatitude overturning circulations shown in the lower panels of Figure 7.

Diagrams illustrating the EP flux vectors, their divergence, and the resulting residual vertical motion field associated with a midlatitude storm track are shown in Figure 8. Figure 8a shows the case in which the eddies propagate only vertically, while Figure 8b shows what happens when the EP flux vectors diverge meridionally away from the baroclinic zone as they propagate upward. The wave driving is spread over a broader range of latitudes when the waves propagate meridionally. Figure 8c shows the anomalies in EP flux divergence and residual vertical velocity that would occur as a result of meridional propagation away from the latitude of greatest wave generation. In the vicinity of the strongest baroclinic eddy generation, the transport of heat by the residual circulation is weakened. Thus, if the wave driving in the upper troposphere is spread over a wider latitude band and the externally applied heating gradient remains the same, then one expects a stronger equilibrium meridional temperature gradient at the wave source location when the eddies propagate away from the jet and break, than if they break without any meridional propagation (via 12 and 14). This in turn provides more baroclinicity to generate eddies and the jet can be self-sustaining rather than a simple diffusive process that responds only to the applied thermal forcing. Robinson (2006) has performed experiments with a two-level model in which eddies actually enhance the meridional temperature gradient and thus are self-sustaining by this measure.

In Figure 8c the meridional eddy flux is shown as roughly symmetric, whereas in nature the eddy flux is preferentially equatorward. If the wave propagation is preferentially equatorward, this displaces the wave breaking and consequent residual sinking motion to the warm side of the jet. A preference for equatorward wave propagation and breaking provides a mechanism for poleward propagation of jet anomalies (Lorenz and Hartmann 2001; Lee *et al.* 2007) that has been observed in some studies (Feldstein 1998).

Thus, in the presence of suitable broad-scale forcing of temperature gradients by radiation or surface fluxes, midlatitude eddy-driven jets are sustainable and their meridional position is not narrowly defined. Therefore, the dominant mode of low-frequency variability in the extratropics is north-south movement of the midlatitude jet (Nigam 1990; Thompson and Wallace 1998; 2000). Observational studies show persistent meridional jet displacements that are maintained by eddy momentum fluxes. The clearest example of this occurs in the Southern Hemisphere, which is more zonally symmetric and has a strong eddy-driven midlatitude jet in all seasons because of the large area of the southern ocean and the strong meridional gradient of SST in all seasons. North-south shifts in the position of the eddy-driven jet dominate the low frequency variability in all seasons and these shifts are supported by changes in the eddy fluxes of momentum (Hartmann and Lo 1998).

A rough understanding of midlatitude zonal flow, eddy interactions in a zonally-averaged context can be patched together from phenomenological studies and linear theory. The dominant process in middle latitudes is the development of baroclinically unstable eddies and their interaction with the zonal mean climate. Baroclinic instability efficiently modifies the mean state of the atmosphere to bring it near a state of neutral stability through the fluxes of heat, momentum and water produced by the disturbances generated by the instability. These disturbances have wavelike properties, so that linear wave propagation theories provide useful insights into the general circulation.

Diagnostic studies using observed and model data indicate that high frequency transient baroclinic eddies feed back positively on zonal mean wind anomalies and give them persistence (Robinson 1991; Yu and Hartmann 1993; Feldstein and Lee 1998; Robinson 2000; Lorenz and Hartmann 2001; 2003). From the analysis above it appears that the high frequency transient eddies would foster self-sustaining jets. They have a strong source in the baroclinic zone associated with the jet and so are likely to have net propagation away from the jet and provide a net momentum flux into it. Also, if they can propagate away from their source region, they are highly likely to approach a critical line as they propagate into regions of weaker zonal wind.

Baroclinic wave energy also cascades to equivalent barotropic waves, which typically have lower phase speeds and so can exist and propagate in the weaker winds on the edges of the jet. The dispersion relation for these barotropic waves is such that they are refracted back into the jet and in so doing pump momentum out of the jet (Lorenz and Hartmann 2001). One can imagine then that a competition exists between high frequency eddies whose propagation out of the jet acts to strongly reinforce the jet, and low-frequency external modes that are refracted into the jet and thereby weaken it. The intensity and structure of eddy-driven jets are thus partly determined by a balance between the effects of baroclinic and barotropic eddies.

It is attractive to seek theories in which the effect of eddies can be incorporated in a closed form theory of the general circulation (Schneider 2006), and many of the statistical properties of transient eddies can be obtained from the properties of the mean flow field (Farrell and Ioannou 1994; Branstator 1995; DelSole and Farrell 1996; Newman *et al.* 1997; Whitaker and Sardeshmukh 1998; Jin *et al.* 2006a; b). It is not yet clear whether dynamical complexity introduced by eddy, mean-flow interactions can be easily incorporated into simple theories of the general circulation. The interaction of eddies with the larger-scale and more slowing-varying components of the flow seems able to generate new forms of variability that may not be captured by treating eddies as parametrically related to the mean flow. In the stratosphere, the dominant mode of low frequency variability is also produced by wave, mean-flow interaction, but the primary waves of interest are quasi-stationary planetary-scale waves, which are forced in the troposphere and propagate upward to produce episodic wintertime stratospheric warmings (McIntyre 1982).

4. Stationary waves

Understanding of the time-averaged zonal asymmetries of the extratropical wintertime circulation has received intensive study (Held *et al.* 2002). The separation of the time-mean flow into its zonal average and the departures from zonal symmetry has been a fruitful area of investigation, and linear models seem to be quite effective in explaining the stationary wave component of the general circulation as a response to forcing by thermal contrasts, topography and transient eddies. It is somewhat surprising that linear theories can explain the time averaged flow so well, when the time-averaged flow seems such a minor part of the phenomenology observed in instantaneous weather maps. The relative importance of thermal versus orographic forcing of stationary waves remains a question of interest and the answer appears to depend sensitively on the wind distribution.

Another issue with stationary wave modeling is the role of transient eddies. Diagnostic studies suggest that transient eddy fluxes are a small term in the balance compared to the larger advection terms associated with the zonal flow and stationary eddies, yet transient eddies seem to play an important role in determining the extratropical response to tropical SST anomalies (Kushnir *et al.* 2002; Orlanski 2005) and transient eddies provide a positive feedback to meridional displacements of eddy-driven jets in a zonally averaged framework (Lorenz and Hartmann 2001; 2003). An outstanding problem is to understand how the strong positive baroclinic eddy, mean-flow feedback described above in a zonally-symmetric framework is applied to the more localized jets observed in nature. Modeling work shows that the structure of modes of low frequency variability follows that of the storm tracks as zonal asymmetry is increased, and one can transition from the nearly symmetric modes of the Southern Hemisphere to the more asymmetric modes of the Northern Hemisphere by adding asymmetry to the forcing (Yu and Hartmann 1995; Cash *et al.* 2005). When strong zonal asymmetries are present, the interaction of the stationary waves and localized jets with the zonal mean flow and transient eddies are both important (DeWeaver and Nigam 2000; Limpasuvan and Hartmann 2000)

Forcing of Rossby waves by tropical convection is important for understanding both the tropical and extratropical responses. Sardeshmukh and Hoskins (1988) pointed out the importance of advection by the divergent component of flow driven by the heating. Dividing the velocity into rotational $\mathbf{V}_\psi$ and divergent $\mathbf{V}_\chi$ components, the conservation equation for the vertical component of absolute vorticity in the absence of friction can be written.

$$\frac{\partial \zeta_a}{\partial t} + \mathbf{V}_\psi \bullet \nabla \zeta_a = -\mathbf{V}_\chi \bullet \nabla \zeta_a - \zeta_a \nabla \bullet \mathbf{V}_\chi = -\nabla \bullet \zeta_a \mathbf{V}_\chi \tag{15}$$

The conservation of absolute vorticity following the rotational flow sees a source associated with the usual stretching term of absolute vorticity, plus a term that is the advection of vorticity by the divergent flow. The combination of these two terms on the far right of (15) has been called the Rossby wave source. In the tropics the divergent component of wind is relatively large and can project onto vorticity gradients and contribute the majority of the Rossby wave source associated with tropical heating. Cognizance of the forcing by divergent wind advection is necessary to understand both the extratropical response to tropical heating (Sardeshmukh and Hoskins 1988), and the surprising equatorial symmetry of tropical stationary waves (Dima and Wallace 2007; Kraucunas and Hartmann 2007). This is especially true because absolute vorticity tends to be homogeneously small in regions of tropical convection, so that $\left|\mathbf{V}_\chi \bullet \nabla \zeta_a\right| \gg \left|\zeta_a \nabla \bullet \mathbf{V}_\chi\right|$.

Meridional propagation of stationary waves across the equator from the summer to the winter hemisphere in the presence of easterly winds is allowed by the presence of mean meridional winds in the same direction (Watterson and Schneider 1987; Esler *et al.* 2000; Kraucunas and Hartmann 2007). Kraucunas and Hartmann (2007) used a shallow water model of the upper troposphere to show why the tropical stationary wave pattern is symmetric across the equator in all seasons, even though the thermal forcing of the waves is in the summer hemisphere. Figure 9 shows the wave response of a shallow water model to localized heating at 10˚N. In the upper panel the zonal mean forcing is symmetric about the equator as for observed equinoctial conditions on Earth, but in the lower panel the zonal forcing is for solstitial conditions, with a stronger westerly subtropical jet in the winter hemisphere and mean meridional winds from the summer to the winter hemisphere. In the solstitial season the stationary wave response is nearly symmetric across the equator, even though the wave forcing is entirely in the summer hemisphere.

There are two fundamental mechanisms that make the solstitial response symmetric across the equator. First, the stationary wave can propagate across the equatorial easterlies in the presence of a mean meridional flow from the summer to the winter hemisphere. Second, the advection of vorticity by the divergent wind is larger in the winter hemisphere, where the zonal mean potential vorticity gradient is larger. For these reasons, although the thermal forcing is in the summer hemisphere, the response of the eddy wind and pressure fields tends to be symmetric across the equator in this simulation as they are in nature (Dima and Wallace 2007).

Significant progress has been made in understanding the interaction between climatological wavelike features in the tropics and the release of latent heat by convection. Concepts such as the gross moist stability (Neelin and Held 1987), vertical modal decompositions (Neelin and Zeng 2000) and the weak temperature gradient approximation (Sobel and Bretherton 2000; Sobel *et al.* 2001), have provided a simple framework for better understanding the tropical circulation and its variability (Bretherton and Sobel 2002; 2003).

5. Tropical Hadley circulation

Because of the weak vertical component of planetary rotation near the equator, relatively weak horizontal gradients of temperature or heating will generate overturning circulations there, and so equilibrium temperature gradients are weak. By using the constraints of balanced flow and conservation of angular momentum, it is possible to provide relatively simple explanations for the existence and structure of the Hadley Circulation (Schneider 1977; Schneider and Lindzen 1977; Held and Hou 1980). Held and Hou (1980) showed that solutions to the zonally-symmetric general circulation require a meridional overturning near the equator to fulfill the requirements of angular momentum conservation, but that this circulation need not extend to the pole as hypothesized by Hadley and other early workers, but can transition to solutions in radiative equilibrium at a latitude that is near the observed poleward extent of the Hadley circulation. If angular momentum is approximately conserved in these overturning circulations, then they produce very strong jets at their poleward extremities. Conservation of angular momentum implies jets of near 130 m s^{-1}, rather than the observed 40 ms^{-1}, and it is the large-scale eddies that are primarily responsible for taking the momentum surplus in the upper branch of the Hadley circulation and transporting it poleward and downward to sustain the surface westerlies of midlatitudes.

Heating gradients in the tropics generate meridional overturning circulations that efficiently flatten the temperature gradients and bring them into balance with the wind field. If the heating is displaced off the equator, these circulations are strongly asymmetric with respect to the equator (Lindzen and Hou 1988), much as is observed in the Earth's atmosphere during the solstitial seasons, when the mean meridional circulation is dominated by a single cell with rising in the summer hemisphere and sinking in the winter hemisphere and strong cross-equatorial flow in the lower and upper troposphere (Figs. 2 and 7). Although the solstitial circulation looks like a single overturning cell, Dima and Wallace (2003) have shown that the annual variation of the Hadley cell can be represented by a symmetric circulation with rising at the equator that varies little with season, plus a single cell with cross-equatorial flow that varies with the season with the upper branch always flowing from the summer to the winter hemisphere.

It is obvious from the earliest diagnostic studies that the export of momentum from the tropics by eddies is needed to explain both the weakness of the subtropical jets and the existence of surface westerlies in the extratropics. This transport of momentum from the tropics to midlatitudes is associated with eddies propagating from sources in midlatitudes to sink regions in the tropics. Waves that propagate into the tropics and break there will mix potential vorticity into the Hadley circulation. The injection of high potential vorticity air into the tropics alters the dynamical response to a hypothetically simple Hadley circulation. Recent work has begun to explore the interactions between tropical and extratropical flows that are associated with the coexistence of a tropical Hadley Circulation and extratropical eddies that propagate into the tropics and influence the zonal momentum and potential vorticity balances there.

The interaction between the tropical and extratropical circulations can be viewed from the perspective of the interaction between the Hadley circulation and extratropical eddies. Hou (1993) studied the response of a simple, dry GCM to moving the tropical heating toward the summer pole. This caused the Hadley circulation to increase and the winter extratropics to warm. Chang (1995) related the extratropical changes to an equatorward shift of the jet stream and associated baroclinic eddies. Seager et al. (2005) computed the GCM response to tropical warming associated with El Niño and showed a similar equatorward shift of the jet and storm tracks and associated zonally-oriented anomalies in precipitation. Global warming simulations indicate that the tropical upper troposphere will warm more than the extratropical troposphere, which introduces an enhanced gradient of temperature between tropics and midlatitudes. This can also increase baroclinic wave activity near the tropical-extratropical transition, and has been shown to cause an increase in the Brewer-Dobson circulation (Eichelberger and Hartmann 2005). Normally, the subtropical jet is less unstable to baroclinic eddies than the extratropical atmosphere, except when the subtropical jet is sufficiently strong, or the baroclinic zone is sufficiently equatorward (Lee and Kim 2003).

6. Gravity waves

In the past 25 years it has been recognized that the vertical fluxes of momentum by gravity waves play an important role in determining the general circulation of the atmosphere (Fritts and Alexander 2003). The most important role of gravity waves, other than adjusting the mass field toward geostrophic balance, seems to be vertical momentum transport, although breaking gravity waves may also provide significant mixing. The importance of gravity waves in the momentum budget increases upward in the atmosphere, since gravity waves can propagate freely and their amplitudes increase as the density decreases. The role of gravity waves is paramount in the mesosphere, where a qualitatively correct general circulation can only be obtained by incorporating the convergence of vertical momentum flux by gravity waves (Holton 1983; Garcia and

Solomon 1985). The body force associated with gravity wave momentum flux convergence is needed to explain the warmth of the winter polar mesosphere. Mesospheric gravity wave momentum flux convergence has a large effect as low in the stratosphere as 30km, particularly when the forcing by planetary scale waves is weak (Garcia and Boville 1994). Gravity wave momentum fluxes are used to correct cold biases in simulated winter polar stratospheric temperatures (via 12 and14). Variability such as the equatorial quasi-biennial oscillation is importantly influenced by the effects of gravity waves (Dunkerton 1997; Baldwin *et al.* 2001).

Gravity wave momentum fluxes are also important for tropospheric climate and weather prediction. Numerical forecast models have shown systematic errors in upper tropospheric winds whose spatial structure can be associated with surface topography. These biases can be alleviated by incorporating the gravity wave momentum flux convergence in a parameterized way (Palmer *et al.* 1986; McFarlane 1987).

The principle sources of gravity wave activity include topography, convection, unbalanced winds and shear instabilities. Gravity waves propagate upward from their source regions and if they do so conservatively their amplitude will grow as the air density decreases. Their upward propagation may cease because their amplitude gets so large that they break, or they may encounter a critical line where their intrinsic frequency is zero and beyond which they cannot propagate. In either case their vertical momentum flux will cease and the convergence of their momentum flux will cause an acceleration of the mean winds there. The nature and location of this gravity wave momentum flux convergence depends on the amplitude and phase speed spectra of the gravity waves present, and the wind and static stability that the gravity waves encounter as they propagate. In general, gravity waves with shorter horizontal wavelengths are reflected or absorbed lower down in the atmosphere, and only gravity waves with wavelengths of 10-100km can easily penetrate the stratosphere.

Linear theory for the case of waves propagating through an atmospheric mean state that depends only on height is well developed and provides a logic for developing gravity wave parameterizations (Lindzen 1981; Holton 1982; Matsuno 1982; Hines 1997; Fritts and Alexander 2003). This theory suggests that the acceleration associated with gravity wave absorption will have the same direction and sign as the horizontal phase propagation in the frame of reference moving with the mean flow (the intrinsic phase speed of the wave). In the mesosphere, gravity waves can break simply because of their amplitude, but in the stratosphere waves are more likely to break as a result of approaching a critical level. Because of the latter effect, gravity wave momentum forcing tends to be the same sign as the local vertical shear. For example, gravity waves in the lower stratosphere exert a westward force above the tropopause jet maximum. Reliable assessments of the role of gravity waves in the general circulation suffer from a lack of detailed observations of the spectrum and global distribution of gravity wave activity (Alexander *et al.* 2002; Pulido and Thuburn 2006).   Nonetheless, most realistic general circulation and climate models contain a gravity wave parameterization, which can have a significant effect on the circulation, especially if the model includes a stratosphere.

7. The tropical tropopause

The tropopause is a dynamically and climatically important feature of the atmospheric circulation. Its definition and explanation are a continuing theme of general circulation research (Highwood and Hoskins 1998; Pan *et al.* 2004; Randel *et al.* 2006). The tropical and extratropical tropopauses are disconnected at the subtropical jet stream and appear to be maintained differently. Thuburn and Craig (2000; 2002) showed that radiative effects of CO_2 and ozone are critical in defining the location and temperature of the tropical tropopause, but Kuang and Bretherton(2004) suggest that overshooting convection can also play an important role in the energy balance.

Recent work has shown that the tropical tropopause is not a sharp boundary between tropospheric and stratospheric air. Ozone, temperature and isotopic data suggest that a transition between tropospheric and stratospheric air begins several kilometers below the tropical tropopause (Folkins *et al.* 1999). This has given rise to the term upper troposphere-lower stratosphere (UTLS) to describe the region between about 13 and 17 kilometers where the air characteristics change from tropospheric to stratospheric values. The region of efficient vertical mixing by convection in the tropics extends to about 13 kilometers, but above that level the air is sufficiently cold, and therefore dry, that radiative cooling by longwave emission from water vapor and parcel heating by condensation both become inefficient (Folkins *et al.* 2000; Hartmann *et al.* 2001a; Folkins 2002; Hartmann and Larson 2002). Cloud tops are observed most frequently in the vicinity of 200 hPa (Hartmann *et al.* 2001b) and a very small fraction of convective parcels actually reach the tropical tropopause (Dessler *et al.* 2006). The vicinity of the tropical tropopause is thus a region where radiative relaxation rates are small, convection is infrequent and small changes in large-scale advection can have significant effects on the temperature balance.

It is important to understand why the tropical troposphere transitions from well-mixed tropospheric air to a mix of stratospheric and troposphere air at a level well below the tropopause. Explanations based on vertical mixing have been proposed (Folkins and Martin 2005), but lateral mixing between the extratropical stratosphere and the tropical upper troposphere across the subtropics through the gap between the tropical and extratropical tropopauses is probably important in this region. This would mix ozone-rich air from higher latitudes into the tropical upper troposphere below the tropopause. The subtropical jet forms a potential vorticity gradient barrier in this region, but the subtropical jet does not extend across all longitudes with equal strength. It is known that eddies propagate into the deep tropics, especially where and when weak westerlies extend to the equator (Webster and Holton 1982; Waugh and Polvani 2000). One can infer from Figure 3 and (2) that a substantial amount of wave breaking and potential vorticity mixing occurs in the subtropical upper troposphere.

The annual cycle of temperature near the tropical tropopause has a maximum in July-August and a minimum in December-January, the annual temperature anomaly is confined near the tropopause and is the largest variation in tropical zonal-mean temperature. Since tropical tropopause is coolest during the Northern Hemisphere winter (Yulaeva *et al.* 1994), it has long been held that this annual variation in tropical tropopause temperature is driven in part by annual variations in the wave driving of the stratosphere and the associated Brewer-Dobson circulation, in which air rises in the tropics and subsides in the winter extratropical stratosphere (Holton *et al.* 1995). Recent diagnostic and modeling work suggests that the annual variation of tropopause temperature may be controlled by processes largely within the tropical UTLS.

Kerr-Munslow and Norton (2006) performed diagnosis of the heat budget of the tropical tropopause and found that the temperature variations were driven by nearby stationary eddy fluxes with a surprisingly large role played by vertical momentum fluxes. Norton (2006) has performed simple numerical experiments to show that the temperature at the tropical tropopause is coldest when the thermal stationary wave forcing is nearer the equator. This would suggest that the tropical tropopause is warmer when the thermal forcing is displaced far off the equator as it is during the northern summer monsoon season. Cold tropical tropopause conditions appear above superrotating westerlies in Norton's simulations, since the cold zonal mean temperatures arise in part to balance equatorial westerlies below.

The superrotation in Norton's simulations results from Rossby waves propagating away from the equator and transporting westerly momentum to the equator, following the mechanisms described in section 2.2. Strong, zonal-mean superrotations are not observed in the troposphere. Kraucunas and Hartmann (2005) suggest that superrotation is suppressed because the tropical heating rarely stays on the equator for very long and when it is off the equator a Hadley circulation develops that generates easterly accelerations on the equator, which compensate for the westerly momentum source associated with thermal wave driving in the tropics (see also Lee 1999). The asymmetry of the Hadley circulation with respect to the equator thus plays a key role in determining zonal mean tropopause temperature through its effect on the zonal momentum budget. Nonetheless, it seems that seasonality in the forcing of tropical stationary waves plays an important role in determining the annual variation of tropical tropopause temperature.

8. The general circulation and climate change

An important test for our understanding of the general circulation is whether we can predict changes in the general circulation that might be associated with past or future climate changes. It would be especially useful to predict changes associated with global warming. Changes in the location, intensity or seasonality of major climatological features of the general circulation could be more important than average temperature changes, particularly where these changes might affect local hydrologic balances. Joseph et al. (2004) found that stationary waves weaken in a global warming simulation, producing a more zonally invariant climate.

One question is whether climate change will significantly affect the location or intensity of midlatitude storm tracks and associated jets. Because the wave, mean-flow interaction in midlatitudes naturally produces low-frequency variations in the latitude of the jets, it is reasonable to think that a modest climate change might significantly affect the position of jets and their associated storm tracks. Changes in jet position could be driven by tropospheric climate change (Yin 2005), or indirectly from the stratosphere (Hartmann *et al.* 2000), where change associated with ozone depletion and greenhouse gas increases is currently larger than surface climate change, and where connections to the surface climate have been demonstrated (Kodera *et al.* 1990; Baldwin and Dunkerton 1999; Thompson *et al.* 2000; Kuroda and Kodera 2004; Baldwin and Dunkerton 2005). Modeling of the response of midlatitude jets to greenhouse gas induced warming suggests that a drop of sea level pressure over the poles and a northward shift of the midlatitude jets is likely (Fyfe *et al.* 1999; Kushner *et al.* 2001; Bengtsson *et al.* 2006; Miller *et al.*

2006), but this signal is affected by the response of ENSO to global warming, which is less certain (Yamaguchi and Noda 2006).

Water will play a very important role in the response of circulation to climate change in the Tropics. As the climate warms up, the amount of water in the surface air and the moist adiabatic lapse rate shift significantly, and this increases both the efficiency of convection in heating the atmosphere and the dry static stability. Relatively simple models also suggest an important role for moist convection in setting the lapse rate of the extratropical atmosphere, although moisture does not seem to affect the dominant scales of motion in the extratropics (Frierson *et al.* 2006). Since more latent energy is released in upward motion because of higher absolute humidity, and more adiabatic heating is produced by subsidence because of increased dry stability, it is expected that the strength of the tropical overturning circulations will weaken in a warmed Earth (Knutson and Manabe 1995; Vecchi *et al.* 2006), but it is unclear yet how this interacts with the extratropical circulation.

Both tropical and extratropical atmospheric circulation changes will be tied closely to the interaction of the atmosphere with the ocean, which is beyond the scope of this review. A shift in the width or position of the Hadley Cell (Hu *et al.* 2005; Fu *et al.* 2006) or the extratropical wave-driven jet (Kushner *et al.* 2001) would have important implications for hydrology and many other things.

9. Conclusion

The past 25 years have seen significant advances in understanding the role of wave, mean-flow interaction in the general circulation of the atmosphere and its variability. In particular, the relationships among wave propagation, momentum transport and zonal flow accelerations are much better understood and have been applied to a wide range of problems. Many challenges remain, especially in understanding the interactions between scales of motion, particularly the interaction of the general circulation with gravity waves and convection. The role of moisture is increasingly being considered in recent studies, as Lorenz (1991) had expected. Our understanding of the general circulation will be tested and enhanced by efforts to predict the response to greenhouse gas increases.

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Figure Legends

Fig. 1 Latitude versus height cross-section of zonal mean wind for DJF and JJA derived from ERA-40 reanalysis (contour interval 5 ms^{-1}).

Fig. 2 Same as Fig. 1 except for zonal mean meridional wind (contour interval 0.5 ms^{-1}).

Fig. 3 Zonal average eddy meridional flux of zonal momentum for DJF and JJA seasons (contour interval 5 m^2 s^{-2}).

Fig. 4 Zonal average eddy meridional flux of temperature for DJF and JJA seasons (contour interval 5 ˚K ms^{-1}).

Fig. 5 Contours of 300mb transient eddy momentum flux as a function of latitude and phase speed for Dec.-Mar. season (left) and Jun.-Sept. season (right). Heavy lines indicate the zonal mean wind speed. Taken from Randel and Held (1991).

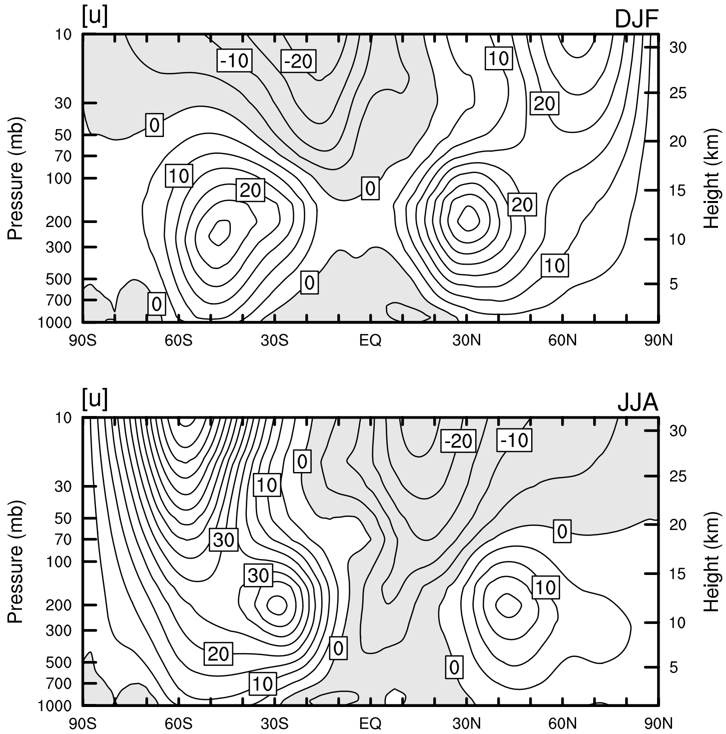
Fig. 6 A schematic showing the eddy heat (thin contours) and momentum (thick contours) fluxes and the associated Eulerian mean meridional circulation (heavy arrows).

Fig. 7 Zonal mean mass streamfunction diagnosed from isentropic analysis by Tanaka et al. (2004). The top two panels show the total mass streamfunction and the bottom two show the mass streamfunction that can be associated with transient eddy forcing (Units 10^{10} kg s^{-1}).

Fig. 8 Schematics showing the EP flux vectors and divergence in latitude–height cross sections for (a) wave fluxes that propagate upward and converge, (b) wave fluxes that propagate upward and then meridionally outward, and (c) the difference in

EP flux convergence and residual vertical velocity when the eddies diverge meridionally away from the source region. C and W indicate latitudes where the anomaly in residual vertical velocity would cause cooling and warming, respectively. The temperature gradient across the core of the jet is increased when waves can propagate meridionally out of the jet.

Fig. 9 Stationary wave response to heating centered at 10N for (a) Equinox mean wind conditions, and (b) Solstitial wind conditions for Southern Hemisphere winter. Shading indicates heat source, contours are height field and vectors represent winds. Note the strong symmetry with respect to the equator of the lower panel, despite the wave forcing being in the summer hemisphere. (From Kraucunas and Hartmann 2007) (Reference wind vector is 10 ms^{-1}).



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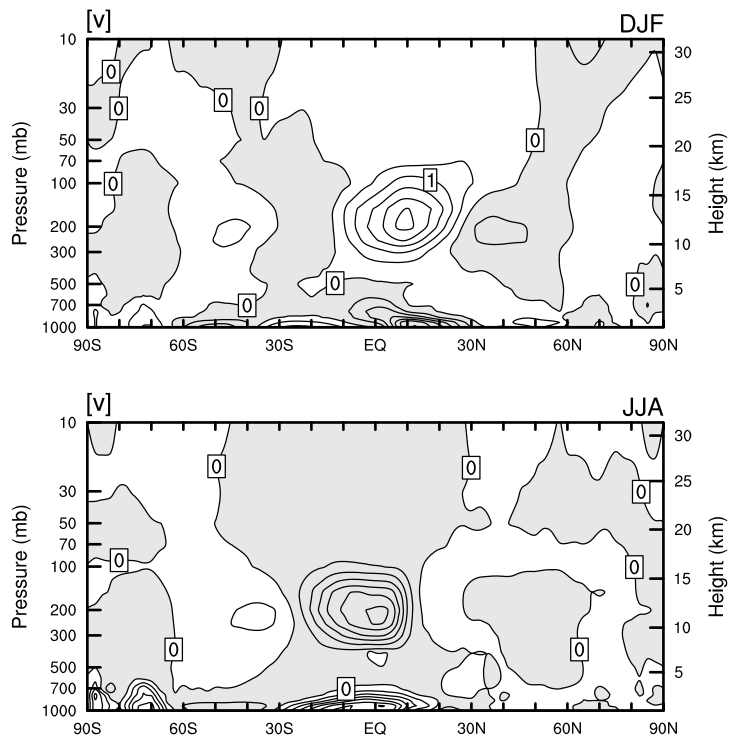

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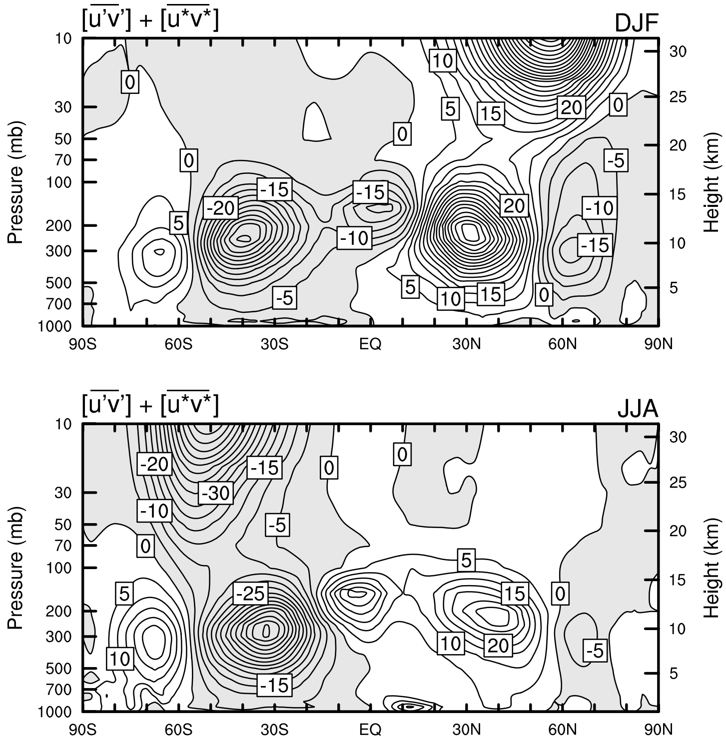

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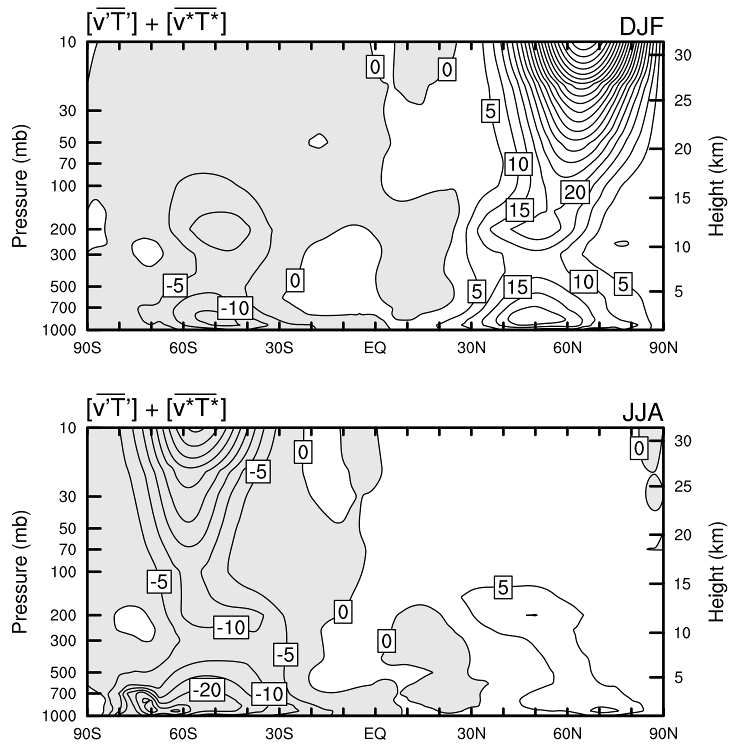

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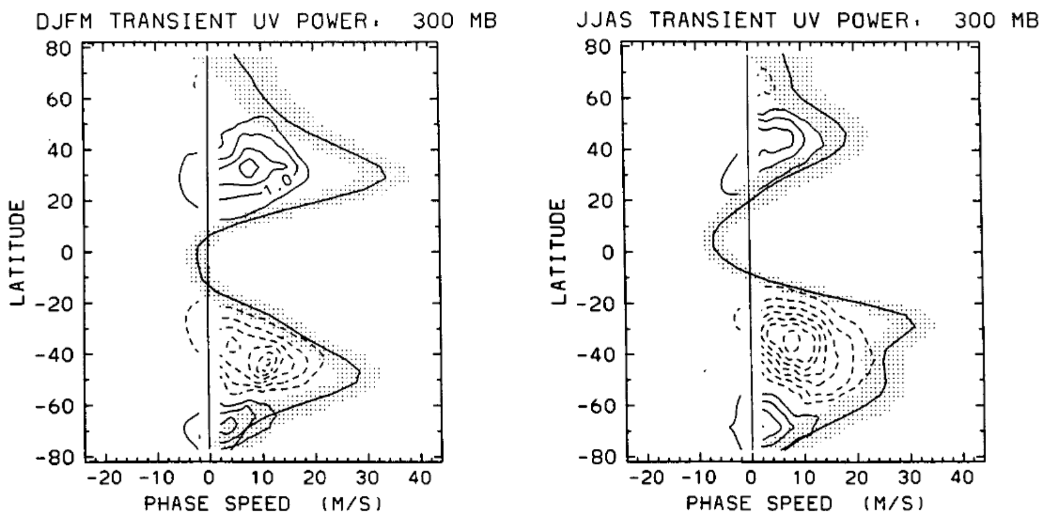

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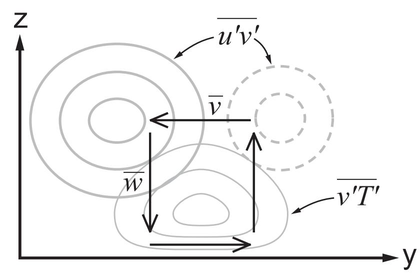

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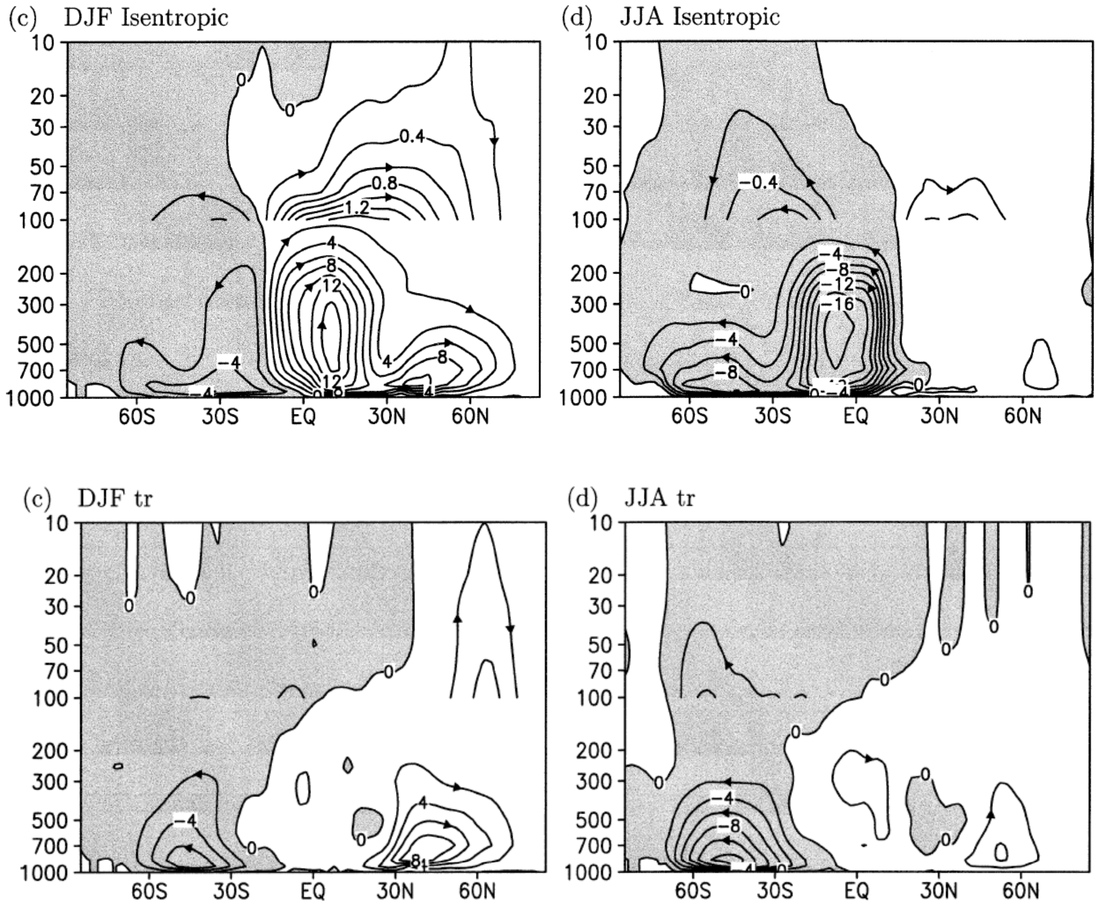

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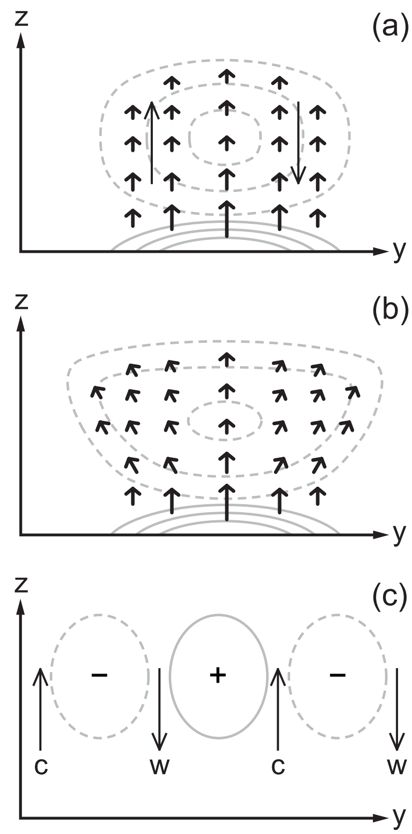

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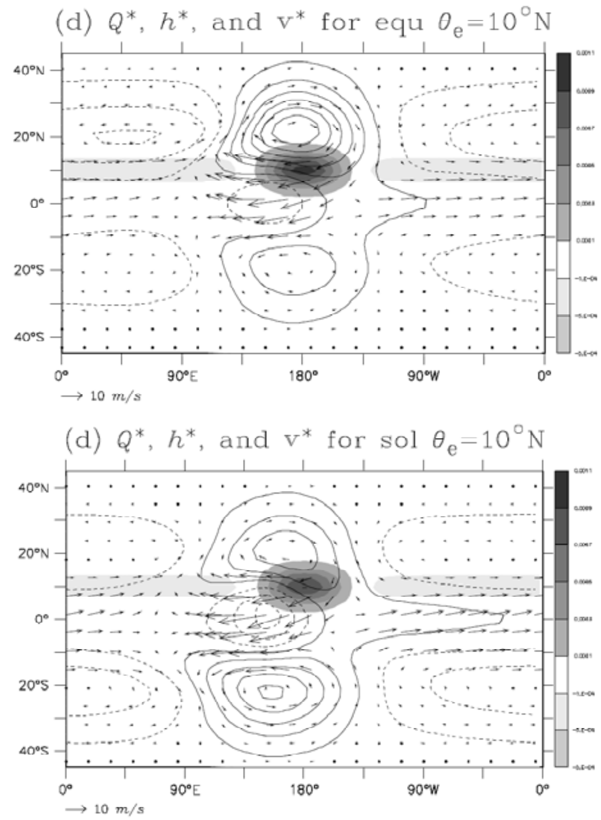

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