NOTES AND CORRESPONDENCE

On the Global Exchange of Mass between the Stratosphere and Troposphere

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ABSTRACT

Seasonal mean downward mass fluxes across the 100 mb level in the extratropics of both hemispheres are computed using the meridionally integrated residual vertical circulation as determined from the transformed Eulerian mean equations. The eddy momentum and heat flux data required for the calculation are taken from Oort's 15-year climatology. Upward mass flux from the troposphere to the stratosphere in the tropics is computed from mass continuity. The flux is a maximum during Northern Hemisphere winter and a minimum during Northern Hemisphere summer. The computed fluxes imply a 2.5 year turnover time for the global atmospheric layer above 100 mb.

1. Introduction

It is now widely recognized that the major portion of the observed departure from radiative equilibrium of the mean temperature distribution in the middle atmosphere is driven by the dynamical effects of dissipating eddies that propagate upward from the troposphere. These dissipating eddies tend to maintain the temperature in the extratropical middle atmosphere above radiative equilibrium, while the temperature in the tropics is maintained below radiative equilibrium. As a consequence, there is net diabatic heating and upward flux of mass across isentropic surfaces in the tropics, and net diabatic cooling and downward mass flux across isentropic surfaces poleward of about 30°. The implications of the eddy forcing of the diabatic circulation for transport of chemical tracers in the middle atmosphere is reviewed, for example, in Andrews et al. (1987).

The nature of the control exerted by the eddy dynamics on the mean meridional mass circulation of the middle atmosphere was discussed briefly by Haynes and McIntyre (1987) and McIntyre (1987) and in more depth by Haynes et al. (1990). In these works the authors enunciated a principle of "downward control." They showed that for steady, or slowly varying flow on a sufficiently large horizontal scale the zonally averaged downward motion across an isentropic surface at any extratropical latitude is determined by a vertical integral involving the eddy dissipation above the particular isentropic surface. Thus, as pointed out by McIntyre (1987), the time and zonal mean downward mass flow across the extratropical tropopause is controlled by dynamical processes in the first few scale heights above the tropopause, not by the details of tropopause folding and other processes more directly related to tropospheric dynamics.

Eddy dissipation in the stratosphere is apparently dominated by large-scale eddies (Rossby waves). The zonal mean zonal force due to dissipation of such eddies is measured approximately by the divergence of the quasi-geostrophic EP flux (Edmon et al. 1980). Thus, the distribution and seasonal variability of the eddy forcing can be determined from standard meteorological observations, and the downward control principle can then be used to estimate the mean rates of mass flow from the stratosphere to the troposphere in the extratropics. The purpose of this paper is to apply the downward control principle to estimate for each hemisphere and season the averaged extratropical downward mass flux across the 100 mb level implied by the climatological data of Oort (1983), and hence by continuity to obtain an estimate for the upward flux across the tropical tropopause.

2. Equations

The relationship between eddy dissipation and the mean meridional mass circulation can be exhibited in the clearest fashion if isentropic coordinates are used. For application to observations, however, it is simpler to express the equations in isobaric coordinates using the transformed Eulerian mean (TEM) form of zonal averaging. In the following we apply the notation of Edmon et al. (1980). For time-averaged conditions (a seasonal average, say) the residual circulation (\( \vec{b}^*, \vec{\omega}^* \)) approximates the mean meridional mass flux.
Thus, the total flux, \( F_m \), across an isobaric surface poleward of latitude \( \phi_0 \) is given by
\[
F_m = -2\pi a^2 g^{-1} \int_{\phi_0}^{\pi/2} \bar{\omega}^* \cos \phi d\phi.
\] (1)

But from the TEM form of the continuity equation (Andrews et al. 1987)
\[
\frac{1}{a} \frac{\partial}{\partial \phi} (\bar{v}^* \cos \phi) + \frac{\partial}{\partial \rho} (\bar{\omega}^* \cos \phi) = 0
\] (2)
so that \((\bar{v}^*, \bar{\omega}^*)\) can be expressed in terms of a mass flux streamfunction, \( \Psi \), defined by
\[
\frac{1}{a} \frac{\partial \Psi}{\partial \phi} = (\bar{\omega}^* \cos \phi); \quad \frac{\partial \Psi}{\partial \rho} = -\bar{v}^* \cos \phi.
\] (3)

Substituting from (3) into (1) and using the boundary condition that \( \Psi \) must vanish at the pole, we get
\[
F_m = \pm 2\pi a g^{-1} \Psi(\phi_0)
\] (4)
where the plus (minus) sign refers to the Northern (Southern) Hemisphere. The residual vertical motion is downward outside the tropics so that from (3) \( \Psi \) must be negative in the Northern Hemisphere extratropics and positive in the Southern Hemisphere extratropics. Thus, the total extratropical flux in each hemisphere is obtained by choosing \( \phi_0 \) in (4) so that the magnitude of \( \Psi \) is maximized. Thus, if we let \( \phi_N \) (\( \phi_S \)) designate the latitude at which \( \Psi \) reaches its minimum (maximum) value in the Northern (Southern) Hemisphere, the total extratropical downward mass flux for the hemisphere can be expressed from (4) as
\[
F_m(N) = +2\pi a g^{-1} \Psi(\phi_N) \quad \text{(North)}
\]
\[
F_m(S) = -2\pi a g^{-1} \Psi(\phi_S) \quad \text{(South)}.
\] (5)

By mass continuity the sum of the downward transport in the two hemispheres must equal the upward transport in the tropics
\[
F_m(T) = 2\pi a^{-1} g [\Psi(\phi_S) - \Psi(\phi_N)]
\] (6)
which implies that the upward tropical transport is driven indirectly by eddy dissipation in the extratropics, not by local radiative heating. The latter must be regarded as a consequence of the remote forcing by the eddies.

We next wish to show that \( \Psi(\phi_0) \) can be expressed in terms of eddy forcing. Now, for quasi-geostrophic scaling the TEM form of the zonal mean momentum equation is just
\[
\frac{\partial \bar{u}}{\partial t} - \bar{\eta} \bar{v}^* = (a \cos \phi)^{-1} \nabla \cdot \mathbf{F}
\] (7)
where \( \bar{\eta} = f - (a \cos \phi)^{-1}(\bar{u} \cos \phi) \) is the zonal mean absolute vorticity, and the zonal force per unit mass is given by the divergence of the EP flux:
\[
\nabla \cdot \mathbf{F} = (a \cos \phi)^{-1} \frac{\partial}{\partial \phi} (F_\phi \cos \phi) + \frac{\partial}{\partial \rho} (F_\rho)
\]
where
\[
F_\phi = -a \cos \phi \bar{v} u'; \quad F_\rho = \bar{\eta} a \cos \phi \bar{v} \bar{\theta}' / \bar{\theta}_p.
\]
For time-mean motions (3) and (7) can be combined to give
\[
\frac{\partial \Psi}{\partial \rho} \approx -\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\cos^2 \phi \bar{v} u')
\]
\[
+ \frac{\partial}{\partial \rho} \left( \frac{\bar{\eta} \cos \phi \bar{v} \bar{\theta}'}{\bar{\theta}_p} \right).
\] (8)

Integrating (8) from a reference level \( p_T \) to the top of the atmosphere yields an expression for the latitudinal dependence of the residual streamfunction at the level \( p_T \):
\[
\Psi(\phi, p_T) = -\frac{p_T}{a \cos \phi} \left( \frac{\partial}{\partial \phi} \cos^2 \phi \bar{v} u' \right)
\]
\[
+ \cos \phi \frac{\bar{v} \bar{\theta}'}{\bar{\theta}_p}(\phi, p_T)
\] (9)
where the angle brackets represent a pressure-weighted vertical average from level \( p_T \) to the top of the atmosphere. According to (5) and (9), the total mass flux across a given isobaric surface from the North (South) Pole to latitude \( \phi_N \) (\( \phi_S \)) is proportional to the sum of the vertical component of the EP flux at \( (\phi, p_T) \) and the pressure-weighted vertical integral of the divergence of the meridional component of the EP flux for latitude \( \phi \). Typically, both terms are negative in the Northern Hemisphere extratropics and positive in the Southern Hemisphere extratropics.

3. Global exchange rates

The data source used here was Oort's (1983) atlas of global atmospheric circulation statistics for 1958–1973. Oort's tables of zonal mean heat fluxes at 100 mb, momentum fluxes at 50 and 100 mb, and potential temperature at 50, 100, and 200 mb were used to evaluate numerically the right-hand side terms in (9). The flux data, which are tabulated at 5° latitude intervals, were fit to a twelfth order polynomial and the computations of latitudinal derivatives, and minima and maxima in \( \Psi \) were carried out using the Mathematica symbolic mathematics software package. Vertical finite differencing was used to estimate \( -\partial \theta / \partial \rho \) \( \approx 1.3 \times 10^{-2} \) K Pa \(^{-1}\) at the 100 mb level. The vertical average in (9) was estimated using the trapezoidal rule of numerical quadrature, and assuming a zero contribution from the \( \rho = 0 \) level. Thus, the data were weighted by taking \( \frac{1}{2} \) of the 100 mb flux plus \( \frac{1}{2} \) of the 50 mb flux. Since the observed flux at low latitudes tends to decrease rapidly with height above 100 mb
(Oort 1983) this weighting should be more realistic than an average that gives a larger weighting to the 100 mb values.

Table 1 shows for each season and hemisphere the computed downward mass flux across the 100 mb surface, and the latitude at which the streamfunction reaches its extremum (i.e., where the residual vertical motion changes sign). The annual mean flux in the Northern Hemisphere is 42.4 \times 10^8 \text{ kg s}^{-1} (1.3 \times 10^{17} \text{ kg yr}^{-1}), which is nearly double that in the Southern Hemisphere, consistent with the stronger eddy forcing in the Northern Hemisphere. This value is close to that estimated by Robinson (1980) from tracer mass balance considerations. The seasonal variation is also stronger in the Northern Hemisphere with a three times larger value in winter than in summer, and intermediate values in the equinoctial seasons.

Upward mass flow is confined to the latitude belt bounded by the latitudes given in Table 1. Typically this belt spans a latitude interval of about 12 to 20 degrees on either side of the equator. The exception is during the Southern Hemisphere summer season when according to the data used here weak upwelling extends to 37°S. The net flux is very small at that season, however, indicating only a small departure from radiative equilibrium.

For the 100 mb surface the net global downward flux is 64.8 \times 10^8 \text{ kg s}^{-1} (2.0 \times 10^{17} \text{ kg yr}^{-1}). Since the total mass of atmosphere above 100 mb is 5.2 \times 10^{17} \text{ kg}, the turnover time is about 2.5 years (i.e., about 40\% of the mass lying above the 100 mb layer is exchanged with the region below each year).

The net upward tropical mass fluxes at 100 mb for the four seasons as inferred from (9) are shown in Table 2. There is a strong maximum in the Northern Hemisphere winter (DJF) due to the maximum in the extratropical downward flux calculated for the Northern Hemisphere for that season. During the Southern Hemisphere winter, however, the flux is actually smaller than during the equinoctial seasons.

### 4. Discussion

The previous estimates, based on the TEM, are in approximate agreement with estimates based on computed diabatic heating rates. If it is assumed that the upward flux is uniformly distributed between about ±15° latitude, then the annual mean mass flux implies an area-averaged tropical upwelling velocity at 100 mb of 2.9 \times 10^{-2} \text{ m s}^{-1}. This in turn implies a diabatic heating rate at 100 mb of about 0.25 K d^{-1}, which is consistent with net radiative heating rates commonly quoted for this region; however, since the diabatic heating rates are computed as small differences between large contributions of opposite sign, the method used here, which depends on large scale eddy statistics, may provide a more accurate estimate of the global mass exchange rate.

Atmospheric chemists are generally interested in the exchange rate for various trace species rather than for air itself. The net trace species exchange depends on correlations between tracer concentrations and the mass flow, so that in the case of a tracer whose stratospheric concentration has strong temporal and spatial variability the seasonal variability in net cross tropopause tracer flux may be larger than the seasonal variability in total air mass flux. Furthermore, if such correlations are important, the mean mass flow quantities computed here may have only limited value in the study of tracer budgets. To examine this possibility we consider the budget of the tracer N₂O.

N₂O is a long-lived gas that is well mixed in the troposphere. It has sources at the surface and is destroyed by photolysis and oxidation in the stratosphere. According to WMO (1986) the source strength (measured in terms of the nitrogen flux) is 14.0 ± 3.5 Mt yr^{-1} (1 Mt = 10^9 kg). This is balanced by a stratospheric sink of 10.5 ± 3 Mt yr^{-1} and a net accumulation in the atmosphere of 3.5 ± 0.5 Mt yr^{-1}. Assuming that at the tropical tropopause the mixing ratio of N₂O is a uniform 300 ppbv, our deduced tropical mass flux gives a flux of nitrogen into the stratosphere in the form of N₂O of 61.7 Mt yr^{-1}. Comparing this to the estimated destruction rate in the stratosphere, we conclude that 51.2 ± 3.0 Mt yr^{-1} must return to the troposphere in the extratropics. If we assume that this downward flux is due to the mean residual circulation in the extratropical region, then the downward moving air must have a mean mixing ratio of 250 ± 15 ppbv.

Published observations of N₂O in the lower stratosphere are primarily based on balloon ascents near 45°N in

<table>
<thead>
<tr>
<th>Season</th>
<th>Flux (10^8 \text{ kg s}^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>DJF</td>
<td>93.3</td>
</tr>
<tr>
<td>MAM</td>
<td>56.1</td>
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<tr>
<td>JJA</td>
<td>47.2</td>
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<tr>
<td>SON</td>
<td>62.4</td>
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<tr>
<td>Mean</td>
<td>64.8</td>
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</tbody>
</table>

### Table 1. Seasonal mean downward mass flux at 100 mb for latitudes poleward of φn and φs.

<table>
<thead>
<tr>
<th>Season</th>
<th>Northern Hemisphere</th>
<th>Southern Hemisphere</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>φN</td>
<td>Fm (10^8 \text{ kg s}^{-1})</td>
</tr>
<tr>
<td>DJF</td>
<td>13°</td>
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<td>MAM</td>
<td>14°</td>
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<td>42.4</td>
<td>22.4</td>
</tr>
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</table>
the summer (Schmeltekopf et al. 1977; WMO 1982). These suggest a mixing ratio near the 100 mb level of the order 260–280 ppbv, although Robinson (1980) estimates a value of 250 ppbv.

Thus, the dynamically derived mass exchange rate appears to be consistent with the observed budget of N₂O. Of course, this assumes that the summer N₂O observations represent the entire year, and that the downward tracer flux can be estimated by multiplying the latitudinally average mixing ratio for the extratropics times the mean mass flux. In reality the tracer mixing ratio and residual vertical velocity both have strong latitudinal dependence so that latitudinal correlations should be considered. Such correlations appear to be important in the tracer balance for two-dimensional models (Ko et al. 1985). These factors might be sufficient to account for the apparent discrepancy. Application of the method used here to more recent datasets may help resolve this issue.

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REFERENCES


