Lagrangian estimate of global stratosphere-troposphere mass exchange

Kyong-Hwan Seo[1] and Kenneth P. Bowman

Department of Atmospheric Sciences, Texas A&M University, College Station, Texas, USA

Received 11 April 2002; revised 2 July 2002; accepted 10 July 2002; published 6 November 2002.

[1] Seasonal variations of the mass flux between the stratosphere and troposphere are investigated using a Lagrangian framework. Monthly three-dimensional trajectories are calculated for 1996 and 1997 from the United Kingdom Meteorological Office assimilated data. Mass exchange shows nearly the same annual variations as previous Eulerian studies: (1) the northern net downward flux across the extratropical tropopause shows a primary maximum in late spring and early summer and secondary in winter, which is consistent with observational measurements, (2) the southern net downward flux has a peak only in winter, and (3) the extratropical and tropical air exchange across the upper boundary of the middleworld exhibits a pronounced annual variation with a wintertime maximum and summertime minimum. The occurrence of the largest downward flux over the western edges of the Pacific and Atlantic and near the Mediterranean suggests that tropopause folding and baroclinic disturbances in midlatitude storm tracks are the primary agent for the intrusion of stratospheric air into the troposphere. Tropical upward transport occurs in association with the monsoon anticyclones, and the greatest diabatic injection into the tropical stratosphere takes place over the summer hemisphere subtropics according to the seasonal variation of the Hadley circulation. Comparison of the Lagrangian flux estimate across the 100-hPa surface with the estimate by the downward control mechanism using the residual mean meridional circulation shows reasonable correspondence. *INDEX TERMS:* 0368 Atmospheric Composition and Structure: Troposphere—constituent transport and chemistry; 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; *KEYWORDS:* stratosphere-troposphere exchange, middleworld-overworld exchange, mass flux, Lagrangian, storm track

Citation: Seo, K.-H., and K. P. Bowman, Lagrangian estimate of global stratosphere-troposphere mass exchange, *J. Geophys. Res.*, *107*(D21), 4555, doi:10.1029/2002JD002441, 2002.

1. Introduction

[2] The exchange of air and trace constituents between the stratosphere and troposphere is important to the variation and distribution of the chemical species in both regions. By carrying man-made ozone depleting chemicals from the lower troposphere into the stratosphere, cross-tropopause transport has a tremendous impact on stratospheric chemistry. The downward flux of ozone-rich stratospheric air into the troposphere contributes to the oxidizing capacity of the troposphere. The exchange also affects synoptic meteorology; for example, the intrusion of high potential vorticity air of stratospheric origin into the low-level baroclinic area can trigger cyclone development [*Throncroft and Hoskins*, 1990]. The downward influx of dry stratospheric air can cause an increase in the local outgoing radiance [*Appenzeller and Davis*, 1992] by diminishing the amount of tropospheric precipitable water. Moreover, explanation of the recent decreasing trend in the total column ozone in midlatitudes requires a dynamical contribution from poleward transport of ozone-poor subtropical air into the stratosphere [*Hood et al.*, 1999, and references therein].

[3] Stratosphere-troposphere exchange (STE) can be modeled by a single cell zonal-mean diabatic circulation over each hemisphere, comprising rising motion from the tropical troposphere to the stratosphere and sinking flow from the extratropical stratosphere to the mid and upper troposphere. Another component of the exchange, isentropic transport, is significant on the timescales shorter than that of the cross-isentropic diabatic transport. Various observational studies, for instance, of air parcels, water vapor and chemical constituents, verify that faster transport occurs along isentropic surfaces [*Appenzeller et al.* 1996a; *Lelieveld et al.* 1997; *Vaughan and Timmis*, 1998; *Ray et al.* 1999]. Accordingly, accurate calculation of STE on a timescale of a month requires consideration of the combined effect of diabatic and adiabatic transport near the tropopause. So a three-dimensional Lagrangian approach may provide a more realistic picture of mass exchange across the

[1]Now at Department of Meteorology, Florida State University, Tallahassee, Florida, USA.

0148-0227/02/2002JD002441$09.00

tropopause or other boundary surface. In this study, we use kinematic trajectories driven by three-dimensional analyzed winds to estimate the rate of air mass exchange between the stratosphere and troposphere.

[4] As a part of the stratosphere-troposphere exchange problem, the exchange between the middleworld and the overworld (MOE) [*Hoskins*, 1991] has also been examined. By applying the principle of "downward control" put forward by *Haynes and McIntyre* [1987], *McIntyre* [1987], *Haynes et al.* [1991] and *Holton et al.* [1995], *Holton* [1990], *Rosenlof and Holton* [1993], and *Rosenlof* [1995] estimated the global vertical exchange. According to downward control theory, the mean extratropical mass circulation across a given control surface (either isentropic or isobaric) is remotely controlled by the dynamical eddy forcing, including unresolved subgrid-scale gravity wave drag and turbulent motion, above that surface, not by the details of the physics below. It has been shown that the seasonal cycle of upward mass flux into the tropical stratosphere is determined by the annual cycle of dynamical eddy forcing, which is strongest in the Northern Hemisphere (NH) extratropics. The zonal momentum forcing in the stratosphere and mesosphere above 100 hPa or the 380 K isentropic surface induces a maximum net upward mass flux over the tropics in the NH winter and minimum in the NH summer. Another estimate of the upward and downward mass flux based on calculation of the zonal mean diabatic heating shows a qualitatively similar result [*Yang and Tung*, 1996]. An estimate of the amount of middleworld-overworld mass exchange will be carried out with three-dimensional Lagrangian methods as an independent comparison with the results above.

[5] Cyclogenesis or tropopause folding, and more broadly, transport associated with storm track activities are believed to be the primary processes responsible for upward (troposphere-to-stratosphere) and downward (stratosphere-to-troposphere) mass fluxes across the extratropical tropopause [*Danielson*, 1968; *Andrews et al.* 1987; *Lamarque and Hess*, 1994]. Cut-off cyclones and small-scale turbulence in the upper troposphere may also be important factors [*Hoskins et al.*, 1985; *Price and Vaughan*, 1993; *Lamarque and Hess*, 1994; *Wirth and Egger*, 1999].

[6] Although the mechanisms of the remote forcing and tropopause folding seem to act disparately, these may be put in the same frame as discussed by *Lamarque and Hess* [1994]. They suggested that mass exchange by tropopause-folding events is connected with the downward control principle through mass continuity near the tropopause. In this study the detailed mechanism for this exchange will not be explored rigorously, but instead by comparing the previous estimates from the above processes with the current Lagrangian estimates, the most plausible physical mechanism will be inferred.

[7] Recently, a variety of studies on the cross-tropopause exchange using Lagrangian methods have been devoted in relation to synoptic or mesoscale weather systems, revealing fairly detailed behavior of the different type of airstreams and relevant physical processes [*Wernli and Davies*, 1997; *Wirth and Egger*, 1999; *Stohl*, 2001; *Wernli and Bourqui*, 2002]. Especially, *Wernli and Bourqui* [2002] provides Lagrangian 1-year statistics of deep extratropical cross-tropopause exchange in the NH by employing a residence time criterion that can distinguish between the short- (less than 1–2 days) and long-lasting exchange events (less than 10 days). They identify the importance of tropopause folding or storm track for the cross-tropopause exchange (see also *Seo and Bowman* [2001]) and derive the annual cycle of net flux and two-way flux. In a few respects, our method differs from theirs. As *Stohl* [2001] indicated, filamentation by synoptic scale-eddies corresponds to stirring process and occurs on the relatively short timescale, whereas mixing process, which destroys parcels' original chemical coherency, is of practical importance, for example, in chemical impacts. In this sense, it is worth investigating the larger spatial and longer timescale air mass mixing; and the applied methodology in the following can effectively average out the small-scale and short-timescale (within a week) "reversible" features and stirring process. Furthermore, in this study global cross-tropopause mass exchange will be calculated including in the tropics and in the Southern Hemisphere. Mass exchange calculations easily extend between the middleworld and overworld.

[8] The objectives of this study are (1) to estimate global exchange of air between the stratosphere and troposphere and between the middleworld and overworld using a Lagrangian framework, (2) to compare the seasonal variation to previous observational and diagnostic studies, and (3) to investigate the physical processes for the exchange by analyzing the temporal and spatial distribution of the mass flux.

2. Trajectories

[9] In the Lagrangian approach, trajectories of individual fluid parcels are calculated by solving the kinematic equation of motion,

$$\frac{d\boldsymbol{x}}{dt} = \boldsymbol{v}(\boldsymbol{x}, t), \tag{1}$$

where, $\boldsymbol{v}$ is the velocity at the position of a parcel, $\boldsymbol{x}$. Trajectories of air parcels are calculated in the spherical coordinate with pressure as a vertical coordinate. Three-dimensional trajectories of individual parcels are computed using a standard fourth-order Runge-Kutta scheme [*Bowman*, 1993] with 32 time steps per day (45 min) and assimilated daily winds from the United Kingdom Meteorological Office (UKMO) [*Lorenc et al.*, 1991; *Swinbank and O'Neill*, 1994]. One-month trajectories are computed for 1996 and 1997 to investigate the seasonality of cross-tropopause exchange. Parcels are initialized on the first day of each month on 128 by 192 uniform longitude-latitude grids extending from 88°S to 88°N (2.8125° lon × 0.9215° lat) at 33 pressure levels with intervals of 25 hPa. Horizontal velocity and pressure vertical velocity $\left(\omega = \frac{dp}{dt}\right)$ are interpolated linearly in longitude-latitude-pressure coordinates and time from the gridded data to the positions of air parcels.

[10] Note that the pressure vertical velocity is subjective to considerable uncertainty, as in other vertical velocity data, because the variable is a diagnostic quantity produced from the numerical model used in the data assimilation system, which is a development of the scheme at the UK

Met Office for operational weather forecasting. Although it is difficult to quantify the effect of the vertical velocity on resulting mass flux, it is known that the vertical velocity appears to capture the broad-scale vertical circulation, but produces unrealistic features at smaller scales; so this serves to our purpose to some degree even though the small-scale deep convection in the tropics are not well resolved.

[11] Preliminary tests for several months of 1997 show that the mass flux from integration initialized 3 days earlier or later than the nominal initial day differs less than 10%. The change in the integral time step in the range between 30 and 90 min results in no more than about 10% difference in the mass flux, implying fairly robust results in the study.

3. Methods

[12] In this study, the tropopause is defined as the surface combining the portions of the 2 potential vorticity unit (1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1}) drawn from the poles into the tropical region and the potential temperature of 380 K from the equator into the subtropics, unless otherwise stated. This definition of the tropopause forms the continuous surface around the globe, which changes with time, and facilitates the comparison and validation of the results of other technique. The 380 K potential temperature surface also demarcates the upper boundary of the extratropical middleworld.

[13] Mass volumes are employed to facilitate the calculation of mass flux across the tropopause or boundary between the middleworld and overworld. A mass represented by each parcel in the three-dimensional space is calculated as follows:

$$M_{i,j,k} = -\frac{1}{g}\left(a_0^2 \cos\phi_j\right)\Delta\lambda\Delta\phi\Delta p, \qquad (2)$$

where a_0 and g are a radius of the Earth and a gravity acceleration, respectively, and λ, ϕ, and p denote respectively longitude, latitude and pressure. Parcels (indexed as i, j, and k) are located at the centers of individual mass boxes.

[14] By following each parcel it can be determined whether mass is transferred to other regions of the atmosphere. Since STE or MOE involves a change in potential vorticity or potential temperature following individual air parcels, mass flux across boundaries can be easily calculated by interpolating to find the initial and final values of potential vorticity or potential temperature for each trajectory.

[15] The spatial distribution of cross-tropopause exchange is examined to gain some insights into the three-dimensional exchange process. The locations at which parcels cross from one part of the atmosphere to another are used to find the spatial distribution of mass exchange. The mass fluxes are binned in 5° × 5° longitude-latitude boxes.

4. Results

4.1. Mass Flux Between the Lowermost Stratosphere and Troposphere

[16] Figure 1 compares our Lagrangian estimate of mass exchange with an Eulerian budget method developed by *Appenzeller et al.* [1996b]. We calculate the monthly net downward mass flux employing their mass flux balance equation. The net downward flux from the lowermost stratosphere into the troposphere is estimated by summing the downward flux from the overworld by the diabatic circulation and the rate of change of the mass of the lowermost stratosphere. The net downward flux is plotted as thin solid line in the NH and thin dotted line in the SH.

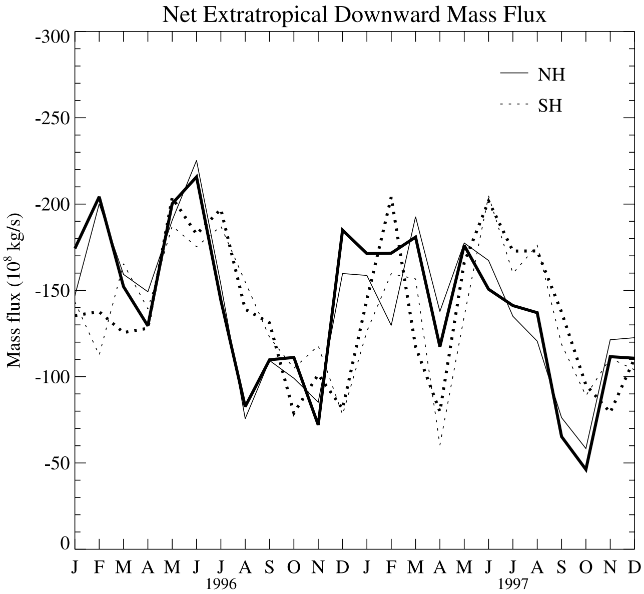

Figure 1. Monthly variations of the net downward mass flux from the lowermost stratosphere into the troposphere using an Eulerian mass budget equation in the northern hemisphere (solid lines) and southern hemisphere (dotted lines). The thick lines are the Lagrangian net downward mass fluxes in each hemisphere.

[17] While the budget equation yields only the net flux into the troposphere, the Lagrangian analysis explicitly provides the two-way mass exchange rate. The Lagrangian net downward mass flux is calculated by subtracting the upward mass flux of tropospheric air into the extratropical lowermost stratosphere from downward flux of the lowermost stratosphere, which is overdrawn as the thick lines for each hemisphere in Figure 1. Actually both components are 2 to 3 times larger than the net downward flux and the diabatic downward mass flux from the overworld.

[18] We see from this figure that the variation of the net cross-tropopause flux estimated by the balance equation exhibits a prominent seasonal cycle. As derived by *Appenzeller et al.* [1996b] and *Gettelman and Sobel* [2000], the primary maximum in the NH appears in spring or early summer and a secondary maximum occurs in winter. The former is related to the period of greatest baroclinic wave breaking during the phase of most rapid decrease of the lowermost stratospheric area, and the latter arises from the strongest winter downward flux from the overworld. The smallest net flux appears in the fall season, as was shown in the isentropic cross-tropopause mass exchange by *Seo and Bowman* [2001]. The magnitude of this flux compares relatively well to the results by *Appenzeller et al.* [1996b] and *Gettelman and Sobel* [2000]. The northern winter average mass flux into the troposphere by this Lagrangian method is 1.8×10^{10} kg s^{-1}, which is somewhat greater than their results of 1.4×10^{10} kg s^{-1} and 1.2×10^{10} kg s^{-1},

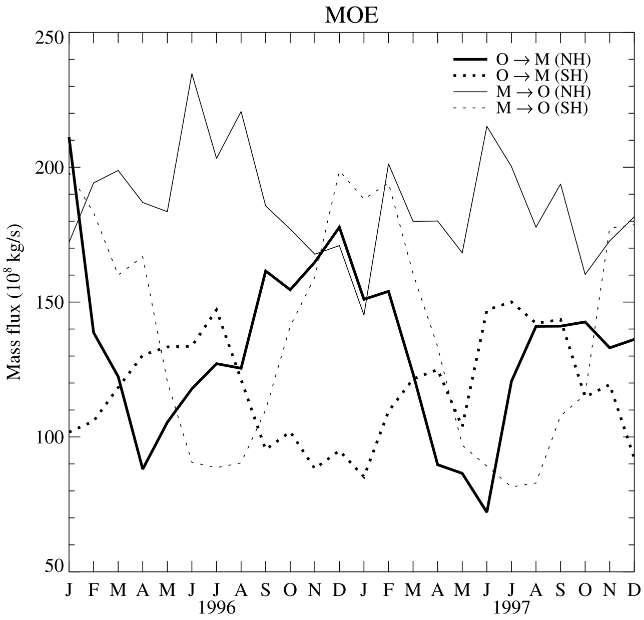

Figure 2. Monthly global mass fluxes across 380 K isentropic surface. The downward mass flux in the NH is denoted by the thick solid line and that in the SH by the thick dotted line. The thin solid line and dotted line represent the upward mass flux in the NH and in the SH, respectively.

respectively. The southern annual cycle peaks in winter but is less pronounced than the northern cycle.

[19] The match between the Lagrangian and the Eulerian net downward mass flux in the northern hemisphere phase and magnitude is quite good. The late spring maximum and fall minimum are in reasonable agreement with an observational evidence for ground-level concentration of the stratospheric radioisotope, Strontium-90 (Sr-90) from atmospheric nuclear bomb tests [*Fry et al.*, 1960].

[20] In order to test the dependence of the results on integration times, mass flux across the tropopause and between the middleworld and overworld (next section) is computed for different lengths of integration times with a 5-day increment (results not shown). Except for the rapid increase in mass exchange during the first few days, the exchanged mass increases approximately linearly as a function of integration time. Hence, the use of 1-month integration does not significantly change the net mass flux. Instead, this tends to represent the large-scale exchange that accompanies irreversible mixing between the two different air properties.

4.2. Mass Flux Between the Middleworld and Overworld

[21] A view of the monthly hemispheric mass fluxes across 380K isentropic surfaces is provided in Figure 2. The largest downward (overworld-to-middleworld) flux appears during the fall and winter in both hemispheres. This is mainly attributed to increased zonal momentum forcing in the cool seasons, which induces a large interannual variability. Consistent with the downward control principle, the NH peak in winter is large relative to the SH winter peak owing to the greater wave forcing in the extratropical stratosphere in the NH winter as shown by *Rosenlof and Holton* [1993] and *Rosenlof* [1995]. The smallest downward flux occurs in warm seasons.

[22] The upward (middleworld-to-overworld) mass flux has summertime maxima and wintertime minima in both hemispheres. Since major upward transport occurs in the tropics, the cycle of the upward mass flux follows the seasonal oscillation of the meridional mean circulation in the upper troposphere. Thus, tropical air parcels are injected across the boundary of the overworld and middleworld in different regions of the tropics at different times. Also, by this reason and partly the higher annual variation in the NH momentum forcing in the extratropical stratosphere, the seasonal cycle in the NH is less distinct than the SH. However, the NH flux is larger than the SH flux throughout the year except the SH summer and thus total tropical upward flux (not shown) exhibits the same annual cycle as the NH.

[23] To compare with the previous study using the transformed Eulerian-mean (TEM) residual mean circulation (see Table 1 of *Rosenlof and Holton* [1993]), we calculate the seasonal mean downward flux across the 100 hPa pressure as shown in Table 1. The Lagrangian result has an annual cycle in common with the residual mean circulation estimate and the northern total exchange per year is similar between two approaches. But the annual fluctuation by the Lagrangian method is greater than the remote forcing estimate by the residual mean transport. Also, the southern mass flux appears generally stronger than the TEM approach. Table 1 also shows the seasonal mean extratropical mass flux across the tropopause for each season.

4.3. Spatial Distributions of Mass Exchange

[24] The Lagrangian method makes it possible to identify where air parcels cross the tropopause and thus indirectly identify the mechanism of exchange. This is done by tracking all the parcels, and summing their masses at the point where the parcels cross the tropopause. Figure 3 shows December–January–February (DJF) and June–July–August (JJA) total mass fluxes and the locations at which parcels cross the tropopause in each direction. During DJF the major peaks of downward flux appear over East Asia and the western Pacific, the Atlantic, and the Mediterranean. During JJA the strongest downward mass flux appears over a longitudinal band from the southern Indian Ocean through Australia to the South Pacific. Another

Table 1. Seasonal Mean Extratropical Net Downward Mass Fluxes Across the Tropopause and 100 hPa Surface[a]

	Across Tropopause		Across 100 hPa			
			NH		SH	
	NH	SH	Lagr.	RH93	Lagr.	RH93
DJF	168	85	102	80.6	27	33.4
MAM	150	93	44	45.7	56	30.7
JJA	92	156	5	25.7	71	30.1
SON	70	77	56	42.7	40	27.6
Mean	120.0	102.8	51.8	48.7	48.5	22.4

[a]Units are in 10^8 kg s^{-1}. Compared with results from *Rosenlof and Holton* [1993] (here RH93).

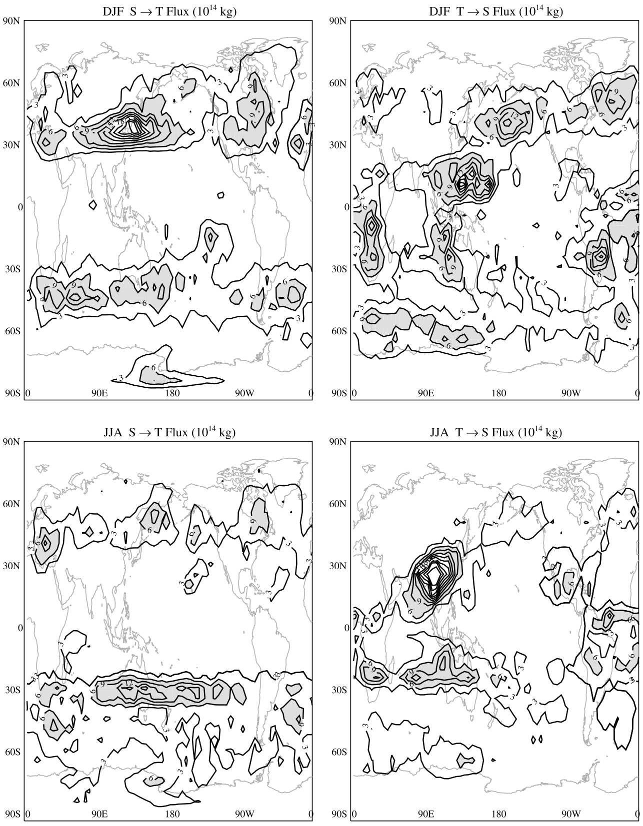

Figure 3. Season total downward and upward mass flux and location at which parcels cross the tropopause in DJF and JJA. The mass contour intervals are 3×10^{14} kg. The regions with the values greater than 6×10^{14} kg are shaded.

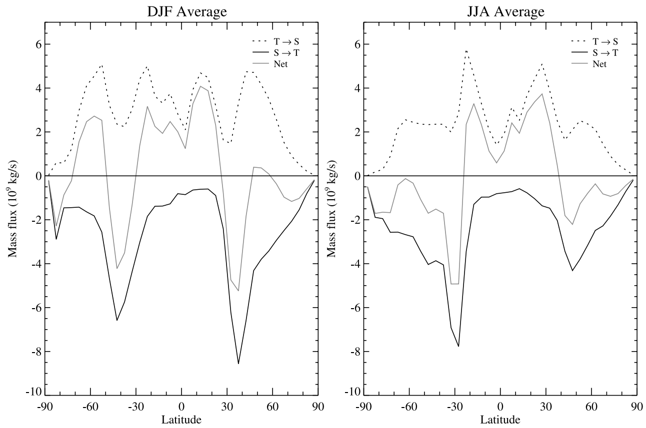

Figure 4. Zonal mean seasonal mean downward (solid lines), upward (dotted lines) and net (gray lines) mass fluxes across the tropopause in DJF and JJA.

relatively small peak appears over the South Atlantic Ocean. These regions for both winter hemispheres are consistent with the preferred routes for transport of stratospheric air into the troposphere. These peaks also coincide with the regions of the strongest cyclogenesis and the principal storm tracks [*Whittaker and Horn*, 1984; *Trenberth*, 1991; *Frederiksen and Frederiksen*, 1993] and the highest frequency of tropopause folding events [*Postel and Hitchman*, 1999]. This again indicates that transient baroclinic activities and associated Rossby wave breaking near the tropopause are the primary cause for exchange between the stratosphere and troposphere in the extratropics.

[25] Upward transport occurs in the subtropics as well as in the midlatitudes. DJF maxima in the NH are located in the western tropical Pacific, while JJA peaks appear over Southeast Asia and subtropical North America. The strong summertime peak in Asia is consistent with the southwest-northeastward tilted maximum region of the climatological specific humidity observed by *Dethof et al.* [1999]. The two peaks over Asia and Mexico are created by the monsoon convection in the Asian and Mexican anticyclones [*Chen*, 1995; *Dunkerton*, 1995; *Dethof et al.*, 1999]. The SH summertime peaks are located over the northern Australia, South Africa and South America. This indicates that tropical air parcels are transported upward in association with the monsoon in the SH [*Newell and Gould-Stewart*, 1981] as well.

[26] To more clearly characterize the latitudinal dependence of the net flux, the zonal-mean seasonal-mean upward and downward mass fluxes are calculated for both solstice seasons (Figure 4). The net mass flux (i.e., downward minus upward mass flux) is denoted by the gray lines. Two very distinct peaks appear in the downward flux in midlatitudes. The downward flux has a largest value in the winter hemisphere, leading to the largest net flux as well. Again this indicates that downward transport takes place mainly over the storm track latitudes.

[27] Other seasonal variation of the net flux can be seen in high latitudes. The DJF net flux is upward in 45°–60°N and 50°–70°S and the transport into the stratosphere occurs quasi-isentropically (the extratropical parcels exhibit a potential temperature change of less than 12 K, whereas the potential temperature change for the tropical parcels ranges from 22 K to 51 K). By contrast, the JJA net flux is downward at this latitude for both hemispheres. However, individual monthly plots of net flux show that net upward flux occurs throughout the year except for some months of the JJA season. Net flux in the tropics is perennially into the stratosphere via the cross-isentropic diabatic circulation.

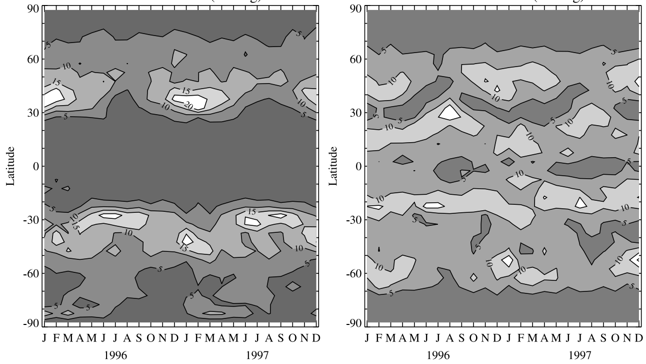

Figure 5. Time evolution of the zonal total mass at the time when parcels cross the tropopause in each direction.

[28] Figure 5 shows the time evolution of the zonal total mass flux across the tropopause in each direction. It is clearly seen that the downward mass transfer occurs principally in midlatitudes in both hemispheres. However, it also reveals a significant hemispheric difference. The northern hemisphere mass flux exhibits an annual march with a maximum in wintertime, while the southern hemisphere flux has a semi-annual cycle with maxima in both winter and summer. Furthermore, the magnitudes of the southern summer peaks are comparable to the winter counterparts.

[29] In contrast to the downward mass transport, the upward mass distribution has a more complex structure as also shown in Figure 5; there are four different bands for this transport; two in the tropics and two in the extratropics (see also Figure 4). The northern extratropical peak exhibits a similar seasonal cycle as the downward mass flux. On the other hand, the upward mass transport in the southern extratropics peaks in summer. The reason for this difference is not investigated in detail in this study. It is interesting to find that the greatest extratropical upward transport in both hemispheres takes place in more poleward regions than the corresponding downward mass transport, which also appears clear in Figure 3.

[30] The tropical bands are found to be formed by rising motion over the intertropical convergence zone (ITCZ) induced by radiative heating (not shown). Although the tropical distribution is more scattered, the major features can be identified by the horizontal distribution of the cross-tropopause mass flux; the northern winter peak is related to mass flux over the western tropical Pacific, and the summer maxima over the India, northern Australia, southern Africa and southern South America as shown in Figure 3.

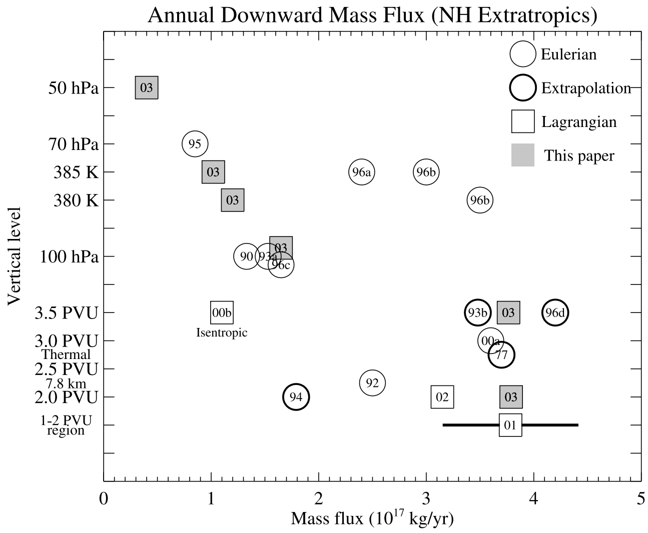

Figure 6. Annual total northern extratropical net downward mass flux in units of 10^{17} kg yr^{-1}. The circles represent an Eulerian method, the squares a Lagrangian method. The gray shading denotes an annual mass flux calculated by the extrapolation of the monthly statistics. The published year of each study is added to make distinctions: *Danielson and Mohnen* [1977], *Holton* [1990], *Follows* [1992], *Rosenlof and Holton* [1993], *Hoerling et al.* [1993], *Lamarque and Hess* [1994], *Rosenlof* [1995], *Yang and Tung* [1996], *Appenzeller et al.* [1996b], *Eluszkiewicz et al.* [1996], *Siegmund et al.* [1996], *Gettelman and Sobel* [2000], *Dethof et al.* [2000], *Stohl* [2001], *Wernli and Bourqui* [2002], and this study.

5. Discussion and Concluding Remarks

[31] An estimate of global-scale air mass exchange between the stratosphere and troposphere is made using the Lagrangian method. In higher latitudes, the net flux is into the stratosphere for DJF with the SH part much greater than the NH. This is consistent with the diagnosis from a global budget application by *Hoerling et al.* [1993] using the mathematical formulation of *Wei* [1987]. They showed that in subpolar latitude bands substantial mass is transported along isentropic surfaces from the troposphere into the lowermost stratosphere by stationary eddy and transient components of the adiabatic transport. This subpolar net upward transport verifies that a local abundance of water vapor at high latitudes is produced by the upward and poleward transport of water vapor-rich tropospheric air approximately following mean isentropic surfaces. In contrast to their results, however, it is shown from this Lagrangian calculation that the high-latitude upward net flux is a factor of more than 3 smaller than the tropical upward flux. For JJA the net flux is into the troposphere in both the tropics and high latitudes.

[32] The annual total extratropical net downward mass fluxes in the NH at various control surfaces are summarized in Figure 6 along with other published results. Our result shows that across the tropopause when defined as the 2 PVU surface, the greatest annual downward mass flux of $\sim 3.8 \times 10^{17}$ kg yr^{-1} appears. This is about 20% larger than the net mass flux of *Wernli and Bourqui* [2002], and is in the range of *Stohl*'s [2001] estimates where the tropopause is defined as the region between 1 and 2 PVUs. The use of a 3.5 PVU threshold in dividing the extratropical troposphere and stratosphere does not change significantly the downward flux relative to the 2 PVU tropopause, suggesting the annual mass exchange is not sensitive to the choice of a specific PV surface.

[33] The isentropic Lagrangian mass transport across the extratropical tropopause between the 322 and 370 K surfaces by small-scale filaments in the work of *Dethof et al.* [2000] underestimates significantly the annual mass flux if other estimates are compared. The extrapolated flux from a simulated tropopause fold by *Lamarque and Hess* [1994] is also smaller than the current estimate. Interestingly, however, the values estimated by extrapolation from individual tropopause folding events by *Danielson and Mohnen* [1977] or from January statistics using the Wei equation by *Hoerling et al.* [1993] and *Siegmund et al.* [1996], yield a similar range of values to our estimate. This implies that the dominant mechanism for the cross-tropopause transport is tropopause foldings and baroclinic disturbance activities along storm tracks in midlatitudes, and the larger-scale processes other than small-scale filament stripping or intrusion are more important in determining the annual exchange between the stratosphere and troposphere. If it is remembered that the strongest downward transport occurs in winter, a brute-force extrapolation made only from a winter estimate to an annual estimate produces an upper limit of

the mass flux. Accordingly, mass transport in our case seems vigorous relative to the other results.

[34] The annual downward mass exchange rate across the 100-hPa surface shows nearly the same magnitude between the Lagrangian estimate and the downward control principle estimate using the TEM residual circulation. Hence, it is seen that the nonlocal dynamical remote forcing mechanism effectively represents actual air parcel exchange between the middleworld and overworld on the meridional plane. Another feature to note is that using the Eulerian approach at 380 or 385 K yields an annual mass fluxes greater than ours by a factor of 2–3. The explanation for this discrepancy needs further investigation.

[35] This study contains several limitations. First, as *Wernli and Bourqui* [2002] indicated, deep convection responsible for rapid vertical exchange is not well resolved in the extratropics. The use of vertical velocity in the tropics produces inescapable errors with no knowledge of apparent and objective true values. These all might lead to an inaccurate estimate of mass fluxes. Second, the applied methodology provides at best an "estimate" of real mixing of air mass, because it is nearly impossible to quantify the actual timescale of mixing for individual air parcels; this mixing process is more important in assessing chemical impacts. Albeit imperfect, this work and other Lagrangian methods provide some useful insights on exchange or mixing process between the troposphere and the stratosphere.

[36] **Acknowledgments.** This study was funded by NASA grant NAG-1-1896.

References

Andrews, D. G., J. R. Holton, and C. B. Leovy, *Middle Atmosphere Dynamics*, Academic, San Diego, Calif., 1987.

Appenzeller, C., and H. C. Davis, Structure of stratospheric intrusions into the troposphere, *Nature*, *358*, 570–572, 1992.

Appenzeller, C., H. C. Davis, and W. A. Norton, Fragmentation of stratospheric intrusions, *J. Geophys. Res.*, *101*, 1435–1456, 1996a.

Appenzeller, C., J. R. Holton, and K. H. Rosenlof, Seasonal variation of mass transport across the tropopause, *J. Geophys. Res.*, *101*, 15,071–15,079, 1996b.

Bowman, K. P., Large-scale isentropic mixing properties of the Antarctic polar vortex from analyzed winds, *J. Geophys. Res.*, *98*, 23,013–23,027, 1993.

Chen, P., Isentropic cross-tropopause mass exchange in the extratropics, *J. Geophys. Res.*, *100*, 16,661–16,673, 1995.

Danielson, E. F., Stratospheric-tropospheric exchange based on radioactivity, ozone and potential vorticity, *J. Atmos. Sci.*, *25*, 502–518, 1968.

Danielson, E. F., and V. A. Mohnen, Project Dustorm report: Ozone transport, in situ measurements and meteorological analysis of tropopause folding, *J. Geophys. Res.*, *82*, 5867–5877, 1977.

Dethof, A., A. O'Neill, J. M. Slingo, and H. G. J. Smit, A mechanism for moistening the lower stratosphere involving the Asian summer monsoon, *Q. J. R. Meteorol. Soc.*, *125*, 1079–1106, 1999.

Dethof, A., A. O'Neill, and J. M. Slingo, Quantification of the isentropic mass transport across the dynamical tropopause, *J. Geophys. Res.*, *105*, 12,279–12,293, 2000.

Dunkerton, T. J., Evidence of meridional motion in the summer lower stratosphere adjacent to monsoon regions, *J. Geophys. Res.*, *100*, 16,675–16,688, 1995.

Eluszkiewicz, J., et al., Residual circulation in the stratosphere and lower mesosphere as diagnosed from Microwave Limb Sounder Data, *J. Atmos. Sci.*, *53*, 217–240, 1996.

Follows, M. J., On the cross-tropopause exchange of air, *J. Atmos. Sci.*, *49*, 879–882, 1992.

Frederiksen, J. S., and C. S. Frederiksen, Southern hemisphere storm tracks, blocking, and low-frequency anomalies in a primitive equation model, *J. Atmos. Sci.*, *50*, 3148–3163, 1993.

Fry, L. M., F. A. Jew, and P. K. Kuroda, On the stratospheric fallout of Strontium-90: The spring peak of 1959, *J. Geophys. Res.*, *65*, 2061–2066, 1960.

Gettelman, A., and A. H. Sobel, Direct diagnoses of stratosphere-troposphere exchange, *J. Atmos. Sci.*, *57*, 3–16, 2000.

Haynes, P. H., and M. E. McIntyre, On the evolution of vorticity and potential vorticity in the presence of diabatic heating and frictional or other forces, *J. Atmos. Sci.*, *44*, 828–841, 1987.

Haynes, P. H., C. J. Marks, M. E. McIntyre, T. G. Shepherd, and K. P. Shine, On the "downward control" of extratropical diabatic circulation by eddy-induced mean zonal forces, *J. Atmos. Sci.*, *48*, 651–678, 1991.

Hoerling, M. P., T. K. Schaack, and A. J. Lenzen, A global analysis of stratosphere-troposphere exchange during northern winter, *Mon. Weather Rev.*, *121*, 162–172, 1993.

Holton, J. R., On the global exchange of mass between the stratosphere and troposphere, *J. Atmos. Sci.*, *47*, 392–395, 1990.

Holton, J. R., P. H. Haynes, M. McIntyre, A. R. Douglas, R. B. Rood, and L. Pfister, Stratosphere-troposphere exchange, *Rev. Geophys*, *33*, 403–439, 1995.

Hood, L., S. Rossi, and M. Beulen, Trends in lower stratospheric zonal winds, Rossby wave breaking behavior, and column ozone at northern midlatitudes, *J. Geophys. Res.*, *104*, 24,321–24,339, 1999.

Hoskins, B. J., Towards a PV-theta view of the general circulation, *Tellus, Ser. AB*, *43*, 27–35, 1991.

Hoskins, B. J., M. E. McIntyre, and A. W. Robertson, On the use and significance of isentropic potential vorticity maps, *Q. J. R. Meteorol. Soc.*, *111*, 877–946, 1985.

Lamarque, J.-F., and P. G. Hess, Cross-tropopause mass exchange and potential vorticity budget in a simulated tropopause folding, *J. Atmos. Sci.*, *51*, 2246–2269, 1994.

Lelieveld, J., B. Bregman, F. Arnold, V. Brüger, P. J. Crutzen, H. Fischer, A. Waible, P. Siegmund, and P. F. J. van Velthoven, Chemical perturbation of the lowermost stratosphere through exchange with the troposphere, *Geophys. Res. Lett.*, *24*, 603–606, 1997.

Lorenc, A. C., R. S. Bell, and B. MacPherson, The Meteorological Office analysis correction data assimilation scheme, *Q. J. R. Meteorol. Soc.*, *117*, 59–89, 1991.

McIntyre, M. E., Dynamics and tracer transport in the middle atmosphere: An overview of some recent development, in *Transport Processes in the Middle Atmosphere*, edited by G. Visconti and R. R. Garcia, pp. 267–296, D. Reidel, Norwell, Mass., 1987.

Newell, R. E., and S. Gould-Stewart, A stratospheric fountain?, *J. Atmos. Sci.*, *38*, 2789–2796, 1981.

Postel, G. A., and M. H. Hitchman, A climatology of Rossby wave breaking along the subtropical tropopause, *J. Atmos. Sci.*, *56*, 359–373, 1999.

Price, J. D., and G. Vaughan, On the potential for stratosphere-troposphere exchange in cut-off low systems, *Q. J. R. Meteorol. Soc.*, *119*, 343–365, 1993.

Ray, E. A., F. L. Moore, J. W. Elkins, G. S. Dutton, D. W. Fahey, H. Vömel, S. J. Oltmans, and K. H. Rosenlof, Transport into the Northern Hemisphere lowermost stratosphere revealed by in situ tracer measurements, *J. Geophys. Res.*, *104*, 26,565–26,580, 1999.

Rosenlof, K. H., Seasonal cycle of the residual mean meridional circulation in the stratosphere, *J. Geophys. Res.*, *100*, 5173–5191, 1995.

Rosenlof, K. H., and J. R. Holton, Estimates of the stratospheric residual circulation using the downward control principle, *J. Geophys. Res.*, *98*, 10,465–10,479, 1993.

Seo, K.-H., and K. P. Bowman, A climatology of isentropic cross-tropopause exchange, *J. Geophys. Res.*, *106*, 28,159–28,172, 2001.

Siegmund, P. C., P. F. J. van Velthoven, and H. Kelder, Cross-tropopause transport in the extratropical northern winter hemisphere, diagnosed from high-resolution ECMWF data, *Q. J. R. Meteorol. Soc.*, *122*, 1921–1941, 1996.

Stohl, A., A 1-year Lagrangian "climatology" of airstreams in the Northern Hemisphere troposphere and lowermost stratosphere, *J. Geophys. Res.*, *106*, 7263–7279, 2001.

Swinbank, R., and A. O'Neill, A stratosphere-troposphere data assimilation system, *Mon. Weather Rev.*, *122*, 686–702, 1994.

Throncroft, C. D., and B. J. Hoskins, Frontal cyclogenesis, *J. Atmos. Sci.*, *47*, 2317–2336, 1990.

Trenberth, K. E., Storm tracks in the southern hemisphere, *J. Atmos. Sci.*, *48*, 2159–2178, 1991.

Vaughan, G., and C. Timmis, Transport of near-tropopause air into the lower midlatitude stratosphere, *Q. J. R. Meteorol. Soc.*, *124*, 1559–1578, 1998.

Wei, M.-Y., A new formulation of the exchange of mass and trace constituents between the stratosphere and troposphere, *J. Atmos. Sci.*, *44*, 3079–3086, 1987.

Wernli, H., and M. Bourqui, A Lagrangian "1-year climatology" of (deep) cross-tropopause exchange in the extratropical Northern Hemisphere, *J. Geophys. Res.*, *107*(D2), 4021, doi:10.1029/2001JD000812, 2002.

Wernli, H., and H. C. Davies, A Lagrangian-based analysis of extratropical cyclones, 1, The method and some applications, *Q. J. R. Meteorol. Soc.*, *123*, 467–489, 1997.

Whittaker, L. M., and L. H. Horn, Northern hemisphere extratropical cyclone activity for four mid-season months, *J. Climatol.*, *4*, 297–310, 1984.

Wirth, V., and J. Egger, Diagnosing extratropical synoptic-scale stratosphere-troposphere exchange: A case study, *Q. J. R. Meteorol. Soc.*, *125*, 635–655, 1999.

Yang, H., and K. K. Tung, Cross-isentropic stratosphere-troposphere exchange of mass and water vapor, *J. Geophys. Res.*, *101*, 9413–9423, 1996.

K. P. Bowman, Department of Atmospheric Sciences, Texas A&M University, College Station, TX 77843-3150, USA.

K.-H. Seo, Department of Meteorology, Florida State University, 404 Love Building, Tallahassee, FL 32306-4520, USA. (khseo@pacific.met.fsu.edu)