Comparison of Global-Scale Lagrangian Transport Properties of the 
NCEP Reanalysis and CCM3

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ABSTRACT

The global-scale transport properties of the NCEP reanalysis winds for the period from 1979 to 1997 are compared to a standard simulation with version 3 of the NCAR Community Climate Model (CCM3) forced by observed sea surface temperatures for the same period. The transport properties of each dataset are defined by the climatological Green’s function for the mass conservation equation for a conserved, passive tracer. Characterizing the atmospheric circulation in terms of material transport provides a very different view of the circulation than standard Eulerian-mean statistics. The Green’s functions are estimated from large numbers of Lagrangian (kinematic) particle trajectories computed by using the NCEP and CCM3 resolved winds. Generally the Green’s functions computed from the two datasets agree well. The transport circulation is dominated by two thermally direct cells, one in each hemisphere. There is a substantial seasonal cycle in the transport, particularly in the Tropics. From a transport point of view, the atmosphere can be divided into three regions: the Southern Hemisphere extratropics, the Tropics, and the Northern Hemisphere extratropics. Particle dispersion within each region is relatively rapid, while exchange between the regions is slower. There are partial barriers to transport between the Tropics and extratropics. Differences between the transport characteristics of NCEP and CCM3 are most noticeable in the Tropics, where CCM3 has stronger subsidence in the ITCZ compared to NCEP. The transport circulation is slightly faster in NCEP than in CCM3. Interhemispheric transport rates computed from the Green’s functions are compared with measurements of long-lived trace species from the Atmospheric Lifetime Experiment/Global Atmospheric Gases Experiment (ALE/GAGE) network. The ALE/GAGE station data for the northern and southern extratropics give an interhemispheric time lag of 1.8 yr for long-lived tracers such as CFCs. Fitting the transport data to a three-box model gives interhemispheric time lags of 1.8 and 2 yr, respectively, for NCEP and CCM3.

1. Introduction

Three-dimensional global climate models (GCMs) are increasingly being used to simulate the chemistry and aerosol composition of the atmosphere. Accurate simulation requires the correct representation of dynamical, thermodynamical, chemical, and radiative processes of the atmosphere. In current and foreseeable models, all of these processes are represented only approximately; and model simulations are imperfect. Uncertainties result both from our incomplete understanding of the relevant processes and our inability to completely and accurately represent those processes in models. In addition, while models allow controlled experiments to be carried out with simulated atmospheres, the complex interactions that occur among different components of the climate system in comprehensive models make it very difficult to unravel cause and effect relationships. Therefore, process models and diagnostic studies that simplify or ignore parts of the system are still essential for developing a detailed understanding.

Atmospheric dynamics (transport) remains an area that contributes substantially to uncertainties in understanding atmospheric chemical processes. Even in the ideal case of passive, long-lived tracers, the complexity of atmospheric motion makes understanding and predicting transport very difficult. At present, many aspects of the transport circulation in the atmosphere are poorly understood. Fortunately, most atmospheric substances with variable concentrations are present in small concentrations and have little direct effect on atmospheric motions. These substances can, therefore, be treated as though they are passively transported by the atmospheric circulation.

One approach to investigating transport is through Lagrangian methods. While it is possible to develop dynamical models in a Lagrangian framework, the label Lagrangian generally refers to kinematic studies. That is, the atmospheric velocity field is assumed to be known at some resolution from a model or from observations. That velocity field is used to compute the trajectories of hypothetical massless air parcels in the evolving flow.
field; and the trajectories are analyzed to study transport, mixing, etc. Lagrangian methods have proven to be very useful in understanding stratospheric transport problems (Hsu 1980; Matsuno 1980; Kida 1983; Austin and Tuck 1985; Schoeberl et al. 1992; Bowman 1993; Fisher et al. 1993; Pierce and Fairlie 1993; Chen 1994; Sutton et al. 1994; Bowman 1996). Schoeberl et al. (2000) recently reviewed some of the strengths and weaknesses of Lagrangian approaches to understanding trace gas distributions in the stratosphere. Although Lagrangian kinematic analysis methods have received widespread use in the stratosphere and for the interpretation of tropospheric field experiments, with the exception of a few studies trajectory methods have not been applied extensively to analyze the large-scale transport circulation of the troposphere (e.g., Kida 1983; Pierrehumbert and Yang 1993). As observations and simulations of the chemistry of the troposphere advance, there is a need for methods to characterize and evaluate the transport circulation on various scales. In this paper we compare the global-scale transport circulation of the atmosphere, taken from the National Centers for Environmental Prediction (NCEP) reanalysis data, with a simulation using a global climate model [the third version of the National Center for Atmospheric Research Community Climate Model (NCAR CCM3)]. Direct, quantitative comparisons are made by computing the climatological Green’s functions of the transport equation for both NCEP and CCM3 winds.

2. Methods

a. The Green’s function method

In order to characterize the transport circulation, we use particle trajectories to estimate the Green’s functions of the mass continuity equation for a conserved tracer,

\[
\frac{\partial s}{\partial t} + \mathbf{v} \cdot \nabla s = 0, \quad s(x, t_0) = s_0(x),
\]

where \(x\) is position, \(t\) is time, \(s\) is the mass mixing ratio of the tracer, and \(s_0(x)\) is the initial condition at \(t = t_0\). If \(v\) is known, then (1) is a linear differential equation for \(s\).

A formal solution to (1) can be found through a Green’s function approach (Hall and Plumb 1994; Holzer 1999; Holzer and Boer 2001; Bowman and Carrie 2002). The Green’s function \(G\) is the solution of (1) for all possible \(\delta\)-function initial conditions (all \(x_0\)); that is,

\[
\frac{\partial G}{\partial t} + \mathbf{v} \cdot \nabla G = 0, \quad G(x, x_0, t_0) = \delta(x - x_0).
\]

If \(G\) is known, the solution to (1) for an arbitrary initial condition \(s_0(x)\) can be found from

\[
s(x, t) = \int_{x_0} s_0(x_0) G(x, x_0, t) \, dx_0.
\]

This approach can be easily extended to find the climatological transport properties of the atmosphere. Given an ensemble of \(v\) fields, (1) can be solved repeatedly for the same initial condition. The ensemble-mean solution \(\langle s \rangle\) is found by taking the ensemble mean of (3), which yields

\[
\langle s(x, t) \rangle = \int_{x_0} s_0(x_0) \langle G(x, x_0, t) \rangle \, dx_0.
\]

Therefore, for any specified initial distribution \(s_0\) the ensemble-mean tracer distribution at future times can be found from the ensemble-mean Green’s function (Bowman and Carrie 2002). More importantly, perhaps, \(G\) provides a quantitative description of the climatological transport of a conserved passive tracer from an arbitrary initial location \(x_0\). Thus \(\langle G \rangle\) is one way to represent the climatological transport circulation of the atmosphere.

Here \(\langle G \rangle\) could be estimated by solving (2) repeatedly, using a GCM, for example; but the computational costs are high. It is possible, however, to estimate \(\langle G \rangle\) from air parcel (particle) trajectories because of the close connection between the solutions to the trajectory equation,

\[
\frac{dx'}{dt} = v(x', t) \quad x'(t_0) = x'_0,
\]

and the solutions to (2). In (5), \(x'\) is the position of the particle as a function of time \(t\), \(v\) is the velocity, and \(x'_0\) is the initial location of the particle at \(t = t_0\). (Primes are used to explicitly denote a particle trajectory.) The Green’s function for (1) is

\[
G(x, x_0, t) = \delta(x - x'_0),
\]

where \(x_0 = x'_0\) and \(x'(x'_0, t)\) is the solution to the trajectory equation (5). This occurs because the trajectories are simply the characteristics of (1). An alternative way of stating this result is that the transport operator simply advects the \(\delta\)-function initial condition without changing its shape (in the absence of diffusion). The path followed by the \(\delta\) function is the same as the path followed by a particle with the same initial location, \(x_0\). Particle trajectories can be used, therefore, to construct the Green’s function.

The following properties of the Green’s function are worth noting. If the value of \(x_0\) is fixed, \(\langle G(x, x_0, t) \rangle\) describes how air initially at \(x_0\) disperses throughout the atmosphere as a function of \(t\). Similarly, if the value of \(x\) is fixed, \(\langle G(x, x_0, t) \rangle\) describes where the air located at \(x\) came from at earlier times. Thus, \(\langle G(x, x_0, t) \rangle\) can be used to describe both to where air goes and from where it comes.

b. Input data

For the purposes of this paper, trajectories are computed using two different sets of global, three-dimen-
sional winds. The first is the NCEP–NCAR reanalysis project dataset (Kalnay et al. 1996), which is referred to here as NCEP winds. The second is a standard simulation of the NCAR CCM3, which is referred to here as CCM3 winds.

1) NCEP WINDS

The NCEP reanalysis winds were chosen because of their temporal and spatial coverage and availability. The reanalysis process uses a frozen, state-of-the-art, global data assimilation system. For this study, data from the 19-yr period 1979–97 are used. See Kalnay et al. (1996) for a detailed description of the assimilation system and input data.

Three-dimensional wind components \((u, v, \omega)\) are provided on a \(2.5^\circ \times 2.5^\circ\), global, latitude–longitude grid at 12 pressure levels (Table 1). Some variables are available at higher levels, but \(\omega\) is not. Results above 100 hPa should be disregarded. Instantaneous velocity values are available at 0.6-h intervals. A small adjustment is made to the vertical velocity field at each pressure level to ensure that the global-mean vertical velocity is zero.

2) CCM3 WINDS

The CCM3 winds are from a standard 18-level simulation at T42 resolution (\(\sim 2.8^\circ \times 2.8^\circ\) transform grid). Winds are archived at 6-h intervals. Three-dimensional winds \((u, v, \eta)\) are taken from the model’s hybrid \(\eta\)-coordinate system. Because CCM3 uses a staggered grid in the vertical, the horizontal velocity components are linearly interpolated from model “levels” onto model “interfaces” where the vertical velocity \(\eta\) is carried (Table 2). The interfaces include the bottom and top of the domain. The model has realistic geography and a seasonal cycle. A single integration is carried out from 1 September 1978 to 31 December 1997 using sea surface temperatures (SSTs) provided by the Global Model Intercomparison Project (Sheng and Zwiers 1998; Gates et al. 1999; Taylor et al. 2000).

c. Trajectory model

To compute the particle trajectories, (5) is solved numerically using a standard fourth-order Runge–Kutta scheme with 32 time steps per day and the resolved three-dimensional velocities from NCEP or CCM3 (Bowman 1993). Sensitivity experiments with the trajectory model have shown that this time step size is quite adequate to solve the trajectory equations accurately given the resolution of the wind data. Velocities at arbitrary \(x\) and \(t\) are computed by linear interpolation in space and time. Vertical motion is calculated in either \(p\) or \(\eta\) vertical coordinates, depending on the input dataset. At the top and bottom boundaries the vertical velocity, \(\omega\) or \(\eta\) is set to zero. In pressure coordinates (NCEP) this causes some problems, as the vertical velocity may not be exactly zero at 1000 hPa (i.e., the surface pressure may vary with time). To avoid possible problems, the global mean vertical velocity at each level is adjusted to be exactly zero. Additionally, areas of steep terrain have persistent downward vertical velocities at 925 hPa (the first NCEP data level above the surface) and tend to accumulate particles near the surface in some locales. Other areas with persistent upward vertical velocities (over the oceans) tend to lose particles from the surface. To keep particles from getting “stuck” near the ground, where horizontal velocities are small, all particles that move to pressures greater than 987.5 hPa (one-quarter of the depth of the lowest data layer) are arbitrarily moved vertically to 987.5 hPa. This effectively keeps particles out of the lowest \(\sim 125\) m of the atmosphere where horizontal winds are weak and, in practice, avoids particles accumulating near the surface. A similar scheme is used at the model top for the NCEP calculations. Because the CCM3 calculations are done in the model’s terrain-following coordinate, similar corrections are not necessary for the CCM3 winds. Particles move across the tropopause in both directions, but the stratosphere is poorly resolved in both the NCEP reanalysis and this version of CCM3, so detailed trajectories above 100 hPa cannot be considered reliable. The zonal-mean circulation in the stratosphere is slow compared to the troposphere.

Although CCM3 has a convective parameterization scheme, in this study we use only the large-scale vertical velocities computed from the continuity equation. Clearly, convective transport plays an important role in the

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**Table 1.** Vertical coordinates \((p, \text{hPa})\) of the NCEP grid at which the vertical velocity \(\omega\) is available. The vertical velocity is set to zero at 0 and 1000 hPa.

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**Table 2.** Vertical coordinates \((\eta,\text{ nondimensional})\) of the CCM3 grid at which the vertical velocity \(\eta\) is available. The vertical velocity is zero by definition at \(\eta = 0\) and 2.515 hPa.

<table>
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general circulation and trace species transport in both GCMs and the atmosphere. Here we follow the philosophy of beginning with simple models and moving to more complex models as our understanding grows. The dual problems of representing convective transport in Lagrangian calculations and understanding the effects of convection on large-scale transport are left for later work. In this study, transport due to the convective parameterization scheme has been neglected. This will likely result in an underestimate of the magnitude of the transport circulation in some areas.

In order that individual particles approximately represent equal masses of air, at 0000 UTC 1 January 1979 \( N = 4 \times 10^5 \) particles are randomly initialized in space [longitude \( \lambda \); sine of latitude (\( \sin \theta \)); and “altitude” \( p \) or \( h \)]. From their random initial conditions, particle positions are integrated forward continuously from 1 January 1979 to 31 December 1997.

d. Computing Green’s functions

As was shown in Bowman and Carrie (2002), a discrete approximation to the Green’s function can be estimated by computing the discrete probability density function of the particles

\[
\langle G_i(x, x_0, t) \rangle = \frac{\int \delta[x - x_i'(x', t)] \, dx}{N},
\]

where \( m_i \) is the number of \( \delta \) functions initially at \( x_i \) at time \( t_0 \) that are in the volume \( x_i \) at time \( t \). That is, the distribution is computed by dividing the domain into a regular, three-dimensional array of grid boxes and counting the number of particles in each grid box, giving a distribution \( p_i(x_i, t) = n_i(x_i, t)/N \), where \( n_i \) is the number of particles initially at \( x_i \) at time \( t_0 \) that are in grid box \( i \) at time \( t \). The ensemble mean is approximated by averaging together the \( G_i \) for all initial conditions that have their initial time \( t_0 \) within a given window, such as a 1-month period. These are then averaged together to compute seasonal or climatological means.

Bowman and Carrie (2002) discuss the sampling errors that arise in trying to estimate \( \langle G \rangle \) by counting particles. Examples below will demonstrate that, except near the edges of the particle distributions, where relative errors are large, \( \langle G \rangle \) is smooth. This indicates that the sampling errors are generally small.

3. Results

In this paper we concentrate on the climatological-mean circulation of the two datasets in the latitude–height plane. Future work will describe other aspects of the circulations. For this problem we compute a discrete Green’s function on a 50 \( \times \) 20 latitude–height grid. The grid is equally spaced in the sine of the latitude. Grid boxes are 50 hPa deep. By integrating over longitude and ensemble averaging over 1710 initial conditions (3 months \( \times \) 30 initial conditions per month \( \times \) 19 years) to produce climatological seasonal means, the climatological Green’s function for a single initial grid box is estimated from \( \sim 684 \, 000 \) particle trajectories.

a. NCEP transport circulation

Figure 1 contains representative slices through the NCEP Green’s function to illustrate the climatological transport circulation for the two solstitial seasons, December–February (DJF) and June–August (JJA). The top panels (a) and (b) show the seasonal cycle of the Hadley circulation in the Tropics. During DJF particle tends to ascend to the south of the equator and subside to the north. The pattern reverses during JJA. This is consistent with the seasonal motion of the intertropical convergence zone (ITCZ) and the large differences between the NH and SH Hadley cells during the winter and summer seasons. Particles ascending in the rising branch of the Hadley circulation in the summer hemisphere tend to cross the equator in the upper troposphere. Note that while the most probable motion (the peak of each distribution) tends to follow the natural sense of the Hadley circulation, there is substantial dispersion both horizontally and vertically. Also, in both hemispheres, but more strongly in the winter hemisphere, there is a strong gradient in particle density and an apparent barrier to transport that slopes upward and poleward in the subtropical troposphere from about 15° to 30° (more on this follows).

The middle two panels (c) and (d) show the dispersion of particles released in the extratropical lower troposphere (41.3°, 775 hPa). In the extratropics, particles tend to disperse rapidly along isentropic surfaces, which slope upward toward the poles in the lower troposphere, and somewhat more slowly across those surfaces (Bowman and Carrie 2002). Particles that descend into the boundary layer tend to warm and move upward across the isentropes, which is equatorward into the sub tropics and Tropics. In each hemisphere the dispersion is slightly greater during the winter than summer. The dispersion in the Northern Hemisphere is slightly larger than the corresponding season in the Southern Hemisphere. This is presumably due to the difference in large-scale eddy activity in the two hemispheres during equivalent seasons. The bulge in the NH in (d) indicates that during JJA some particles do move from the midlatitude lower troposphere more or less directly into the tropical Hadley circulation, without entering the Tropics in the shallow trade wind circulation.

The bottom two panels show the transport away from a low-altitude extratropical release site just outside the Tropics. A substantial number of particles are pulled at low levels from the winter hemisphere across the equator into the ITCZ and then lofted into the middle and
upper tropical troposphere. Particles moving from the summer hemisphere into the Tropics tend not to cross the equator at low levels, due to the location of the ITCZ.

In terms of global-scale transport, it may seem natural to divide the atmosphere into two parts at the equator (or perhaps at the seasonally varying ITCZ). Figure 2 clearly shows, however, that there is no barrier to transport at the equator or ITCZ (see Fig. 1 also). Instead, there are transport barriers in the subtropics of each hemisphere that divide the atmosphere into three regions: the Southern Hemisphere extratropics, the Tropics, and the Northern Hemisphere extratropics. Air tends to disperse relatively rapidly within each region (20 days is shown in this case), while exchange between the regions occurs at a slower rate. This can be inferred from the fact that there is little overlap between the Green’s functions for air originating in the Tropics and air originating in the extratropics. On the other hand, the Green’s functions for air originating at different locations within a single region, the Northern Hemisphere extratropics, for example, are very similar (not shown). On the winter hemisphere side, the boundary between the Tropics and extratropics is relatively distinct. There is more overlap between the extratropics and Tropics on the summer hemisphere side than on the winter side. The primary transport from the extratropics to the Tropics occurs in the lower branches of the Hadley cells where air moves from the subtropics toward the ITCZ,
from whence it is lofted into the tropical troposphere (see Figs. 1e,f). Transport from the Tropics back into the extratropics occurs through a broader zone at higher altitudes. These results are very similar to those of Bowman and Carrie (2002) for an idealized circulation with no geography or seasonal cycle.

A global view of the transport circulation is given in Fig. 3. Bowman and Carrie (2002) discuss some advantages and disadvantages of different ways to represent the transport circulation. Figure 3 shows the most probable transport path (the displacement of the peak of the particle distribution from its initial location) after 10 days for a large grid of initial conditions. The upper panels (a) and (b) show the transport by the NCEP winds, the middle panels (c) and (d) show the transport by the CCM3 winds, and the bottom panels (e) and (f) show the difference (CCM3 − NCEP). To avoid having the displacement arrows line up exactly head to tail (a result of the discrete grid used to estimate the displacement), which makes the figure difficult to read, a small random displacement is added to each vector.

The NCEP panels (a) and (b) show the expected general characteristics of the zonal-mean transport circulation: a single thermally direct cell in each hemisphere with ascent in the Tropics and subsidence in the extratropics. The two overturning cells are asymmetric during the solstitial seasons, with the larger and stronger cell extending from the cold hemisphere across the equator. The CCM3 results and the differences between the two datasets are discussed in the next section.

b. CCM3 − NCEP differences

This section discusses the differences between the transport in the two datasets. The large-scale patterns of transport with the CCM3 winds seen in Fig. 4 generally resemble those from NCEP shown in Fig. 1. The initial locations in Fig. 4 are the same as in Figs. 1a and 1b. Two features are notable in comparison with Figs. 1a and 1b. First, the particle dispersion in CCM3 is generally less than in NCEP, both in latitude and altitude. This will be shown in greater detail below. Second, during DJF CCM3 has stronger subsidence south of the equator in what is normally thought of as the ascending branch of the Hadley circulation (also seen in Fig. 3e). Both datasets show some tendency toward a bimodal distribution for particles originating in the mid troposphere south of the equator; but this tendency is much more pronounced in CCM3, and, in fact, the lower peak has a greater density of particles than the higher.

Quantitative comparison of CCM3 and NCEP transport from many different initial conditions (not shown) shows that the greatest differences in the transport between CCM3 and NCEP occur in the Tropics and subtropics. Figure 5 shows the differences between the Green’s functions for CCM3 and NCEP [(G)_{CCM3} − (G)_{NCEP}] for selected initial conditions and seasons. Positive values indicate that (G)_{CCM3} is greater than (G)_{NCEP}. In the NH during winter (a), NCEP winds tend to move more particles upward and into the Tropics than the CCM3, which shows greater subsidence. The pattern is similar, but not as well defined in the SH during winter (d). In the NH during summer (b), NCEP winds disperse the air more in the latitudinal direction [(G)_{CCM3} is smaller in the NH extratropics], and carry particles farther upward around the Hadley cell and into the SH tropical upper troposphere (dashed contours in tropical upper troposphere). This is also similar to the SH during summer (c), where the CCM3 dispersion is narrower, and particles are not carried as far around the upper branch of the Hadley circulation in a given amount of time. These differences seen in Fig. 5 also show up clearly in Fig. 3. During DJF, in the NCEP data the most likely particle paths are upward in the Tropics south of the equator (Fig. 3a). In CCM3 the most likely paths are downward (Fig. 3c). The difference between the two results is shown in Fig. 3e, which reveals a deep region...
of very different vertical motion south of the equator. There are similar differences of a slightly smaller magnitude in the NH tropics in JJA. During JJA, NCEP has a nearly uniform region of rising motion to the north of the equator. In CCM3, a narrow region of strong rising motion is surrounded by relatively strong subsidence, especially above 600 hPa. In the extratropics there appear to be smaller systematic biases between the two wind fields. In most cases CCM3 winds appear to move particles downward and equatorward more rapidly than NCEP, in contrast to the situation in the Tropics, although this is not true in the SH during JJA, when the bias has the opposite sign.

c. Box model

Although it is difficult to validate the NCEP and CCM3 transport calculations directly due to an absence of global chemical observations, long-term station measurements of long-lived trace species can be used to check the gross properties of the transport. Because anthropogenic trace species are usually released in greater amounts in the Northern than the Southern Hemisphere, there is typically an observable difference in the concentration of trace species between the two hemispheres. For long-lived trace substances, this difference is primarily determined by the rate at which air mixes be-
between the two hemispheres. To compare the transport calculations with trace species observations, the Green’s functions are used to compute the rate at which a trace substance released in the extratropics of one hemisphere mixes into the extratropics of the other hemisphere.

The existence of partial barriers to transport in the subtropics of each hemisphere, along with the relatively rapid mixing within, separately, the Tropics and extratropics, suggests that a box model like that of Bowman and Carrie (2002) could be a useful conceptual framework in which to compare the transport calculations and the trace species observations. The box model approach has the additional advantage that analytical solutions can be found under reasonable simplifying assumptions.

Following Bowman and Carrie (2002) we divide the atmosphere into three boxes: the SH extratropics (90°-30°S), the Tropics (30°S-30°N), and the NH extratropics (30°-90°N). The extratropical boxes each comprise 25% of the atmosphere; the Tropics makes up the remaining 50%. The boxes are assumed to be well mixed internally, and the tropical box exchanges air with both of the extratropical boxes at equal and constant rates. The two extratropical boxes are not directly connected.

The equations for the mass of a conserved trace species in each box are

\[
\frac{d(m_X)}{dt} = -fx_S + fX_T,
\]

\[
\frac{d(2m_X)}{dt} = fX_S - 2fX_T + fX_N,
\]

\[
\frac{d(m_X)}{dt} = +fX_T - fX_N,
\]

where \(X_S\), \(X_T\), and \(X_N\) are the mass mixing ratios of the tracer in the SH, tropical, and NH boxes, respectively; \(m\) is mass of either of the extratropical boxes; and \(f\) is the rate of mass exchange between the boxes.

Dividing by \(m\) (and by 2 in the second equation) gives equations for \(X_S\), \(X_T\), and \(X_N\):

\[
\frac{dX_S}{dt} = -rX_S + rX_T,
\]

\[
\frac{dX_T}{dt} = \frac{1}{2}rX_S - rX_T + \frac{1}{2}rX_N,
\]

\[
\frac{dX_N}{dt} = +rX_T - rX_N,
\]

where \(r = f/m\) is the fractional rate of mass exchange between the boxes relative to the mass of an extratropical box.

To allow a comparison with the NCEP and CCM3 results below, (9) is solved for the initial conditions \(X_S = 0\), \(X_T = 0\), \(X_N = 4\), that is, for a tracer initially contained entirely within the Northern Hemisphere box. The concentrations as a function of time are

\[
X_S(t) = 1 - 2e^{-rt} + e^{-2rt},
\]

\[
X_T(t) = 1 - e^{-2rt},
\]

\[
X_N(t) = 1 + 2e^{-rt} + e^{-2rt}.
\]
different values for the mass exchange rate between the boxes, \( r \). The values for the different datasets and seasons are summarized in Table 3. The estimated mass exchange rates range from 0.5% to 1.9% day\(^{-1}\), depending on season and the choice of box. In every case the mass exchange rates for NCEP are larger than CCM3 by a relative amount of about 10%–15%.

The three-box model agrees quite well with the results from NCEP and CCM3. The differences between the box model and the circulation models can be attributed to the following factors. First, the “box” boundaries used in the NCEP and CCM3 (vertical walls at 30\(^\circ\)N and S) are not the precise locations of the mixing barriers, which slope poleward and upward in each hemisphere. At short times, especially, there is rapid transport across the artificial box boundary due to rapid internal mixing in the extratropics as well as due to systematic exchange with the Tropics. The box-model solution has a hard time matching both the rapid initial adjustment between the NH and tropical boxes that takes places within a few days and the slower exchange at time scales of 1 month to 1 year. Second, contrary to the assumptions of the box model, the tracer does not mix instantaneously within each of the three boxes. Third, the exchange between the three boxes is not a simple pipeline as the box model assumes. It is noteworthy that the

**Fig. 5.** Differences between the Green’s functions for CCM3 and NCEP winds \((G)_{\text{CCM3}} - (G)_{\text{NCEP}}\) for (a) NH, DJF; (b) SH, DJF; (c) NH, JJA; and (d) SH, JJA. Solid contours indicate that CCM3 is greater than NCEP, while dashed contours indicate the reverse.

**Fig. 6.** (a) Analytical solution to Eq. (9), which is the tracer concentration in each of the three boxes of the idealized model as a function of time. (b) Tracer concentration in each of the three boxes computed from the Green’s function for the NCEP DJF data (solid lines). The dashed lines are the results of a nonlinear fit to the NCEP data using the analytical forms in Eq. (9).
calculated concentration in the SH box, which is less influenced by some of the spinup issues than the other two boxes, gives the best fit to the results from the theoretical box model.

It is also informative to include a source term in the three-box model to represent the emission of a long-lived trace species in the NH. The solution for a constant source in the NH box is given in Bowman and Carrie (2002). After the initial transients die away, the concentrations in each box increase linearly with time at the same rate with a constant difference in concentration between the three boxes. This is equivalent to a constant time lag in concentration, with the SH box lagging the tropical box, which in turn lags the NH box (the source region). The lag between the NH and SH boxes is \((1 + 2a)/a\). The lags corresponding to the estimated values of \(r\) from the curve-fitting procedure are given in Table 3.

A rough estimate of the actual interhemispheric mixing time scale can be obtained from measurements of long-lived trace gases taken in midlatitudes of the Northern and Southern Hemispheres. For this purpose we use Atmospheric Lifetime Experiment/Global Atmospheric Gases Experiment (ALE/GAGE) and chlorofluorocarbon (CFC-12) data from the late 1970s and 1980s, a period when CFC concentrations were increasing nearly linearly. Figure 7 shows the available monthly mean CFC-12 data from Adrigole and Macehead, Ireland, and Cape Grim, Tasmania (Prinn et al. 2000). It is apparent that CFC-12 concentrations increased nearly linearly during this time period. A least squares linear fit to each of the time series gives the results shown in Table 4. The interhemispheric lag computed using the average of the two slopes (17.4 pptv yr\(^{-1}\)) and the difference between the intercepts (31.1 pptv) is 1.8 yr, which agrees well with the NCEP and CCM3 estimates of 1.8 and 2.1 yr, respectively (Table 3).

### 4. Summary and conclusions

We use a Green’s function approach to compare the transport circulations of the NCEP reanalysis with a standard simulation by CCM3. The climatological Green’s function of the mass conservation equation for a conserved passive tracer is estimated from a large ensemble of long-term particle trajectories. The trajectories are computed using the large-scale three-dimensional velocity fields from the NCEP reanalysis and from CCM3.

The Green’s function approach provides a quantitative measure of the transport circulation, both the most likely motion of air parcels and their climatological dispersion characteristics. The transport circulation is dominated by thermally direct motion, with rising motion in the ITCZ and subsidence in the subtropics and extratropics of both hemispheres. The tropical circulation undergoes a substantial seasonal cycle as the rising branch of the Hadley circulation follows the solar heating into the summer hemisphere. During the solstitial seasons the Hadley cell in the summer hemisphere is very weak, while the wintertime Hadley cell carries air across the equator from the winter hemisphere to the ITCZ.

A schematic of the global-scale zonal-mean transport circulation is shown in Fig. 8 for the DJF season. This is similar to Fig. 11 in Bowman and Carrie (2002), but it includes the asymmetry in the circulation that results from CCM3.

### Table 3. (top two rows) Mass exchange rate between the boxes as a percentage of the mass in one of the extratropical boxes (25% of the total atmosphere). The three estimates for each region (NH, Tropics, and SH) use the time history of the tracer concentration in the respective box. (bottom two rows) Lag of the concentration of a conserved passive tracer in the SH extratropics relative to the NH extratropics. See the text for a discussion of the differences between the estimates.

<table>
<thead>
<tr>
<th></th>
<th>DJF</th>
<th>JJA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>NH</td>
<td>Tropics</td>
</tr>
<tr>
<td>Mass exchange rate (% day(^{-1}))</td>
<td>0.9</td>
<td>1.6</td>
</tr>
<tr>
<td>NH → SH tracer lag for constant source (yr)</td>
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<td>0.7</td>
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</table>

### Table 4. Coefficients of least squares linear fits to the ALE/GAGE time series of CFC-12 measurements at Macehead, Ireland, and Cape Grim, Tasmania. The “intercept” is evaluated at the beginning of 1978. The slope is in units of pptv yr\(^{-1}\). The time lag between the hemispheres is 1.79 yr.

<table>
<thead>
<tr>
<th></th>
<th>Ireland</th>
<th>Tasmania</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope</td>
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<td>17.3</td>
</tr>
<tr>
<td>Intercept</td>
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<td>245.0</td>
</tr>
<tr>
<td>Correlation coefficient</td>
<td>0.99</td>
<td>0.99</td>
</tr>
</tbody>
</table>

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Fig. 7. Monthly mean clean-air measurements of CFC-12 at Macehead, Ireland, and Cape Grim, Tasmania. The solid and dashed lines are linear least squares fits to the two time series. The coefficients of the linear regression are given in Table 4.

Fig. 8. Schematic of the transport circulation for the DJF season.
from the seasonal cycle. In the extratropics of both hemispheres air disperses relatively quickly throughout the region between ~30° latitude and the pole (gray arrows). Air from the extratropics moves into the Tropics at low levels in the trade wind circulation (long black arrows). Upon reaching the ITCZ, this air ascends into the middle and upper tropical troposphere. There is fairly rapid dispersion of air within the Tropics (gray arrows). As air spreads out in the upper branches of the Hadley cells and descends in the subtropics, some subsiding air mixes into the extratropics of both hemispheres (short black arrows). The dark gray bands indicate that there are partial barriers to transport between the Tropics and extratropics. Although the Hadley circulation varies considerably with the seasonal cycle, the transport barriers persist in all seasons, including the equinoctial seasons (not shown).

This picture of the zonal-mean transport of traces substances is quite different from either the conventional zonal-mean Eulerian-mean circulation (Peixoto and Oort 1992, Fig. 7.19) or the zonal-mean Lagrangian-mean circulation, whether calculated using isentropic coordinates or the residual-mean formalism (Townsend and Johnson 1985; Stone and Randel 1999; Held and Schneider 1999). Air parcels cannot be thought of as primarily circulating around either the single hemispheric cells of the Lagrangian-mean flow or the three cells per hemisphere of the Eulerian-mean flow. Instead, there is relatively rapid dispersion of air with the three atmospheric “boxes” (SH extratropics, Tropics, and NH extratropics) and slower transport around a single thermally direct zonal-mean cell in each hemisphere. In the process there is a slow exchange of air between the three boxes through the subtropics. An understanding of the precise processes that transport air between the three boxes cannot be obtained from the zonal-mean statistics presented here, but will require detailed study of individual parcel trajectories.

CCM3 generally does a good job of simulating the large-scale transport circulation, both qualitatively and quantitatively. The most notable differences between CCM3 and NCEP are the following. Climatologically, particles spread out more rapidly with the NCEP winds than with CCM3. This may indicate that CCM3 is not representing the full range of variability of atmospheric motion seen in NCEP or that the motion is slightly weaker than NCEP indicates. The Hadley circulation is slightly more vigorous in NCEP than CCM3. Thus transport in the Hadley cells is somewhat quicker in NCEP. In the Tropics there is a noticeable bias between the two results in the ITCZ, with CCM3 having a pronounced tendency for subsidence in the ascending branch of the Hadley circulation. Air does ascend in the ITCZ in CCM3, but there is a separate population of particles that descend. That population is distinctly larger than in NCEP.

It is, of course, very difficult to say which wind data are “better.” Although the NCEP reanalysis is tied to global observations through the assimilation process, the direct effects of convective transport have not been included in this study, and the indirect effects on the large-scale circulation may be sensitive to the choice of convective schemes, boundary layer parameterizations, etc. The results do indicate that the two-box (hemispheric) models of global tracers (e.g., Levin and Heshami 1996; Bowman and Cohen 1997) do not provide a good representation of the transport characteristics of the troposphere. The bulk behavior of trace substances can, however, be represented reasonably well by a simple three-box model. It is encouraging that the extratropical exchange times of the two datasets agree with each other, and with the ALE/GAGE data, to within about 10%. Future studies will compare the transport properties computed using Lagrangian methods with explicit calculations of trace substance dispersion calculated with Eulerian scheme built-in to CCM3.

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REFERENCES


