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Key Points:
- Understanding of the Brewer-Dobson circulation is reviewed
- Climate change speeds up the Brewer-Dobson circulation
- Processes driving the Brewer-Dobson circulation trends are not fully understood

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The Brewer-Dobson circulation

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Abstract One of the more robust results of greenhouse gas-induced climate change to emerge from chemistry-climate and climate model projections in the last decade is, depending on the greenhouse gas scenario, an ∼2.0–3.2% per decade acceleration of the global mass circulation of tropospheric air through the stratosphere. This circulation is generally known as the Brewer-Dobson circulation and is characterized by tropospheric air rising into the stratosphere in the Tropics, moving poleward before descending in the middle and high latitudes. The circulation is, however, poorly constrained by observations, and many fundamental questions about it remain. In this review, historical developments in observations, theory, and models describing the Brewer-Dobson circulation are presented along with a reexamination of the basis of the current understanding of the Brewer-Dobson circulation and the mechanisms driving it and its response to climate change. Impacts of anthropogenically driven changes in the Brewer-Dobson circulation are also reviewed.

1. Introduction

This paper reviews recent advances and the current status and challenges in the understanding of the Brewer-Dobson circulation and its response to climate change. A dynamical/meteorological perspective is taken reflecting the emphasis of much of the recent research on this topic though a few particularly relevant studies of the observed transport of material tracers are alluded to as well.

The first use of the term "Brewer-Dobson circulation" in a peer-reviewed journal was by *Newel* [1963] in an article published over 50 years ago in the *Quarterly Journal of the Royal Meteorological Society*. A few earlier papers had, however, used the equivalent term "Dobson-Brewer circulation" with the first of these published in *Nature* by *Goldsmith and Brown* [1961]. *Newel* [1963] also refers to a "Dobson-Brewer circulation" in addition to a "Brewer-Dobson circulation," though both terms had the same meaning. These studies used these terms to refer to a global mass circulation in which tropospheric air enters the stratosphere in the Tropics and then moves upward and poleward before descending in the middle and high latitudes. This basic physical model had been proposed by Dobson and Brewer to explain observations of ozone [*Dobson et al.*, 1929; *Dobson*, 1956] and water vapor [*Brewer*, 1949] though their physical reasoning was not fully supported by a theoretical explanation until the advances made, independently, by *Andrews and McIntyre* [1976, 1978a, 1978b, 1978c] and *Boyd* [1976]. A précis of the key historical developments leading to the modern-day concept of a Brewer-Dobson circulation follows in section 2. Nowadays, as outlined in section 3, the term "Brewer-Dobson circulation" tends to be used in a generic sense to label the gross features of the mean mass transport within the stratosphere without reference to any particular model. These gross features are, nonetheless, essentially those originally identified by *Dobson et al.* [1929], *Brewer* [1949], and *Dobson* [1956].

In the last decade, there has been a surge of interest in the Brewer-Dobson circulation mainly resulting from the development of stratosphere-resolving general circulation models (GCMs) [e.g., *Pawson et al.*, 2000; *Gerber et al.*, 2012] and chemistry-climate models (CCMs) [e.g., *Eyring et al.*, 2005; *SPARC CCMVal*, 2010], though improvements in observations and reanalyses [e.g., *Iwasaki et al.*, 2009; *Seviour et al.*, 2012], and theoretical developments [e.g., *Plumb*, 2002; *Waugh and Hall*, 2002] have all contributed too. In agreement with the pioneering results of *Rind et al.* [1990], the new stratosphere-resolving GCMs and CCMs consistently predict a speeding up of the Brewer-Dobson circulation in response to greenhouse gas-induced climate change [*Butchart and Scaife*, 2001; *Butchart et al.*, 2006; *Garcia and Randel*, 2008; *Li et al.*, 2008; *Calvo and Garcia*, 2009; *McLandress and Shepherd*, 2009; *Butchart et al.*, 2010a, 2010b; *Okamoto et al.*, 2011; *Bunzel and Schmidt*, 2013; *Oberländer et al.*, 2013]. On the other hand, it is difficult to observe any such changes in the strength of the Brewer-Dobson circulation in the available measurements [*Engel et al.*, 2009; *Bönisch et al.*, 2011; *Diallo et al.*, 2012; *Seviour et al.*, 2012; *Stiller et al.*, 2012]. Part of the problem is that while the Brewer-Dobson circulation can be analyzed directly from model output, the observed changes in the circulation are only inferred indirectly from the behavior of material tracers and are, therefore, plagued by large uncertainties. Section 5 examines these issues.

Although an acceleration of the Brewer-Dobson circulation is now established as one of the more robust features of *model* climate projections, there is still considerable uncertainty over the details of the driving mechanisms and, in particular, the mechanisms responsible for the secular changes and variability [e.g., *Forster et al.*, 2011]. The underlying concept is a wave-driven circulation [*Holton et al.*, 1995] though in the model simulations there is a rather large spread (uncertainty) in the partitioning of the driving among the different wave types (e.g., resolved planetary and synoptic-scale Rossby-waves, and unresolved (parameterized) buoyancy waves [*Butchart et al.*, 2010b]). Current understanding of the role of the different wave types in the driving of the Brewer-Dobson circulation and the processes causing the long-term secular changes are reviewed in sections 4 and 5.2, and section 6, respectively.

A changing Brewer-Dobson circulation will impact on many aspects of the stratosphere though, arguably, the most significant impacts will be observed in the recovery of stratospheric ozone [e.g., *Bekki et al.*, 2011], in changes in the lifetimes of ozone-depleting substances and some greenhouse gases [e.g., *Butchart and Scaife*, 2001], and in the exchange of mass between the stratosphere and the troposphere [e.g., *Zeng and Pyle*, 2003; *Hegglin and Shepherd*, 2009]. Section 7 reviews these impacts.

Overall concluding remarks appear in section 8.

2. History

The concept of a global stratospheric mass circulation almost certainly originates from *Dobson et al.* [1929] who, despite some concerns, noted that "the only way in which we [*Dobson et al.*] could reconcile the observed high ozone concentration in the Arctic in spring and the low concentration in the Tropics, with the hypothesis that the ozone is formed by the action of sunlight, would be to suppose *a general slow poleward drift in the highest atmosphere with a slow descent of air near the Pole*." *Brewer* [1949] likewise noted that "all the observed phenomena [in upper air water vapor measurements] can be explained if it is assumed that air circulates by *a slow mean motion into the stratosphere at the equator, moves poleward in the stratosphere and sinks into the troposphere in temperate and polar regions"* [*Brewer*, 1949, Figure 5]. *Brewer* [1949] also had his concerns due to the apparent violation of the conservation of angular momentum as the air moves poleward. Nonetheless, the basic physical concept was confirmed by *Dobson* [1956]. Further refinements then followed from studies (see *Sheppard* [1963] for a review) of stratospheric transport deduced from radioactive fallout, a consequence of the USA's proliferation in atmospheric nuclear testing over the northern tropical Pacific in the late 1950s. The USSR (i.e., the former Soviet Union) also carried out a large number of atmospheric nuclear tests in the late 1950s though their monumental explosions were at middle and high latitudes. The United Kingdom contributed too, in a small way, with most of their tests in the Southern Hemisphere. The longevity of the tropical maxima in the concentrations of radioactive debris and also aerosol from tropical volcanic eruptions [*Dyer and Hicks*, 1968] provided the first evidence for weak transport out of the Tropics and, hence, the existence of subtropical transport barriers.

The first estimate of the stratospheric meridional circulation from diabatic forcing was by *Murgatroyd and Singleton* [1961] who used radiative heating rates to deduce rising motion at the tropical tropopause, descent across the extratropical tropopause and strong transport in the mesosphere from the summer to winter hemisphere [*Murgatroyd and Singleton*, 1961, Figure 2]. Qualitatively, this agreed remarkably well with the circulation deduced from the transport of tracers though *Murgatroyd and Singleton* [1961] again noted a possible nonconservation of angular momentum in their model which would imply the existence of important eddy processes that they had neglected. In contrast, when *Vincent* [1968] calculated the Eulerian-mean meridional stratospheric circulation and included the rectified eddy contribution he found, instead of a single hemispheric circulation cell, two cells with a reverse cell in the high latitudes that had ascent in polar regions and descent in the midlatitudes [*Vincent*, 1968, Figure 1].

A resolution of this paradox between the two-cell Eulerian-mean circulation and the single-cell mass-transporting circulation was not obtained until *Andrews and McIntyre* [1976, 1978a, 1978b, 1978c] reasoned that instead of using the Eulerian mean, it was more appropriate to use either the alternative generalized Lagrangian mean or, indeed, their rather ingenious transformed Eulerian mean (see sections 3.3

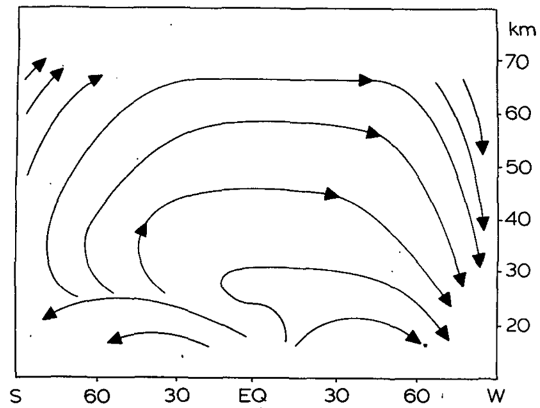

Figure 1. Streamlines of the Lagrangian-mean velocities. The figure represents the radiatively driven estimate to the generalized Lagrangian-mean flow (section 3.3) for three equinoctial months of the year. For further details, see *Dunkerton* [1978]. Figure 4 from *Dunkerton* [1978]. ©American Meteorological Society. Used with permission.

and 3.2, respectively, for the definitions). *Dunkerton* [1978] followed this approach to derive a dynamically consistent picture of the mean-transport streamlines for the stratosphere and mesosphere (Figure 1) which is now generally considered as defining the basic concept of the Brewer-Dobson circulation, despite subsequent refinements in detail [e.g., *Plumb*, 2002] and the different models used to describe the circulation (see section 3). Advective air parcel trajectory calculations by *Kida* [1983a] later confirmed this defining picture. Furthermore, by inspiring the formulation of the "downward-control principle" [*Haynes and McIntyre*, 1987; *Haynes et al.*, 1991], the theoretical work of *Andrews and McIntyre* [1976, 1978a, 1978b, 1978c] also led to a better understanding of the underlying mechanisms driving the Brewer-Dobson circulation through wave breaking and the concomitant "gyroscopic pumping" (see *Holton et al.* [1995] for a comprehensive review of these developments and section 4.1 for an explanation of gyroscopic pumping).

3. Models Describing the Brewer-Dobson Circulation

Because the Brewer-Dobson circulation has no formal definition, the term has been used to describe a broad range of stratospheric overturning circulations even when some of these perhaps do not conform to the original physical model postulated by *Dobson et al.* [1929] and *Brewer* [1949] or, indeed, the corresponding theoretical model of *Dunkerton* [1978] shown in Figure 1. Here only the single-cell time-averaged equator-to-pole stratospheric mass circulation will be referred to as the "Brewer-Dobson circulation." This mass circulation can, nonetheless, be described by a wide range of models and their various refinements. A survey of the models most commonly used and the relationships among them is presented below. Further details can be found in several lucid reviews of stratospheric transport phenomena written by *Matsuno* [1980], *Holton et al.* [1995], *Plumb* [2002], *Waugh and Hall* [2002], *Plumb* [2007], and *Shepherd* [2007].

3.1. Mass and Diabatic Circulations

The zonally averaged "mass circulation" of the stratosphere is defined simply by the transport of (materially) conservative tracers such as ozone below ~48 km, where it is long lived [e.g., *Dobson*, 1956], or radioactive isotopes [e.g., *Newel*, 1963]. In practice, the tracers are usually only "quasi-conservative" or "long-lived" and have weak sources and sinks, or decay rates, in the sense that the resultant changes from these over the relevant advection timescales can be considered small.

For the time average, such as the seasonal mean, the mass circulation is also well approximated by the zonally averaged "diabatic circulation" (e.g., section 2). Following the original approach of *Murgatroyd and Singleton* [1961], the diabatic circulation $(\overline{v}_d, \overline{w}_d)$ is obtained diagnostically from the zonal-mean thermodynamic equation

$$\frac{\partial \overline{\theta}}{\partial t} + \frac{\overline{v}_d}{a}\frac{\partial \overline{\theta}}{\partial \phi} + \overline{w}_d \frac{\partial \overline{\theta}}{\partial z} = \overline{Q} \tag{1}$$

and mass-continuity equation

$$\frac{1}{a\cos\phi}\frac{\partial}{\partial \phi}(\overline{v}_d \cos\phi) + \rho_0^{-1}\frac{\partial}{\partial z}(\rho_0 \overline{w}_d) = 0, \tag{2}$$

using radiative heating rates and temperatures, though here potential temperature is used instead of temperature (definitions of the notation used in the equations are given at the end of the review).

However, *Murgatroyd and Singleton* [1961] acknowledged that their approach omitted the eddy heat-flux convergence term

$$-\frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}(\overline{v'\theta'}\cos\phi)-\rho_0^{-1}\frac{\partial}{\partial z}(\rho_0\overline{w'\theta'}) \tag{3}$$

that would appear on the right hand side of (1) if the conventional Eulerian-mean meridional circulation $(\overline{v},\overline{w})$ was used instead of $(\overline{v}_d,\overline{w}_d)$. Consequently, in regions where the eddy heat-flux convergence (3) is significant, such as the Northern Hemisphere winter stratosphere, the diabatic and Eulerian-mean meridional circulations differ considerably. Indeed, it is this difference that accounts for the physically spurious two-cell structure of the Eulerian-mean meridional circulation noted in section 2, which does not correspond, even approximately, to the mean Lagrangian motion of air parcels in the meridional plane.

3.2. Transformed Eulerian Mean Circulation

The discrepancy between the diabatic and Eulerian-mean circulations arises from a nonlinear rectification of the eddy transport processes contributing to the mean transport with this additional contribution to the mean often referred to as the Stokes drift [e.g., *Dunkerton*, 1978; *McIntyre*, 1980a]. With a simple but ingenious transformation of the Eulerian-mean equations, *Andrews and McIntyre* [1976, 1978c] derived the alternative "residual-mean-circulation" which, for a wide variety of circumstances, combines the contributions of eddy and mean transport into a single quantity, to a good approximation. In this "Transformed Eulerian-Mean (TEM)" formulation, the residual-mean circulation $(\overline{v}^*,\overline{w}^*)$ is defined by

$$\overline{v}^* = \overline{v}-\rho_0^{-1}\frac{\partial}{\partial z}\left(\frac{\rho_0\overline{v'\theta'}}{\overline{\theta}_z}\right)=-\frac{1}{\rho_0\cos\phi}\frac{\partial\psi}{\partial z}, \tag{4}$$

and

$$\overline{w}^* = \overline{w}+\frac{1}{a\cos\phi}\frac{\partial}{\partial\phi}\left(\frac{\cos\phi\,\overline{v'\theta'}}{\overline{\theta}_z}\right)=\frac{1}{a\rho_0\cos\phi}\frac{\partial\psi}{\partial\phi}. \tag{5}$$

This residual-mean circulation satisfies a mass-continuity equation similar to (2) but of particular relevance is the corresponding transformation of the thermodynamic equation to

$$\frac{\partial\overline{\theta}}{\partial t}+\frac{\overline{v}^*}{a}\frac{\partial\overline{\theta}}{\partial\phi}+\overline{w}^*\frac{\partial\overline{\theta}}{\partial z}=\overline{Q}-\rho_0^{-1}\frac{\partial}{\partial z}\rho_0\left(\frac{\overline{v'\theta'}\,\overline{\theta}_\phi}{a\overline{\theta}_z}+\overline{w'\theta'}\right). \tag{6}$$

For quasi-geostrophic flows under nonacceleration conditions, the rectified eddy forcing terms on the right-hand side of (6) are negligible [*Andrews and McIntyre*, 1978c] and (1) and (6) become isomorphic. Hence, under these assumptions, the residual-mean circulation is a good approximation to the diabatic circulation [i.e., $(\overline{v}^*,\overline{w}^*)\approx(\overline{v}_d,\overline{w}_d)$] and it is generally not necessary to distinguish between the two when diagnosing the Brewer-Dobson circulation, at least for the ascending and descending branches (see also section 5.1).

3.3. Generalized/Modified Lagrangian-Mean and Transport Circulations

Dunkerton [1978] argued that the diabatic and residual-mean circulations should also provide reasonable estimates of the "generalized Lagrangian-mean circulation" $(\overline{v}^L,\overline{w}^L)$ introduced by *Andrews and McIntyre* [1978a]. Roughly speaking, this involves averages over sets of moving fluid parcels rather than averages over, say, latitude as would be the case with the conventional Eulerian mean. The generalized Lagrangian mean is also referred to as the "modified Lagrangian mean" [e.g., *McIntyre*, 1980a] if the sets of moving fluid parcels initially have identical quasi-conservative physical properties, such as potential vorticity (or long-lived trace-gas concentrations) and potential temperature. Although this generalized or modified Lagrangian-mean circulation provides a much better conceptual representation of the mean mass transport within the stratosphere, there are significant practical problems with its application [*McIntyre*, 1980b].

One of the problems is the generalized Lagrangian-mean circulation $(\overline{v}^L,\overline{w}^L)$ does not satisfy a mass-continuity equation [*Andrews and McIntyre*, 1978a], and when it was diagnosed from numerical models by *Kida* [1983a] and also *Plumb and Mahlman* [1987, Figure 9b], it was found to include a large divergent component and to actually differ somewhat from the Lagrangian-mean flow estimated by *Dunkerton* [1978]. By defining a "transport-circulation" $(\overline{v}_T,\overline{w}_T)$ that essentially represents the advective transport by

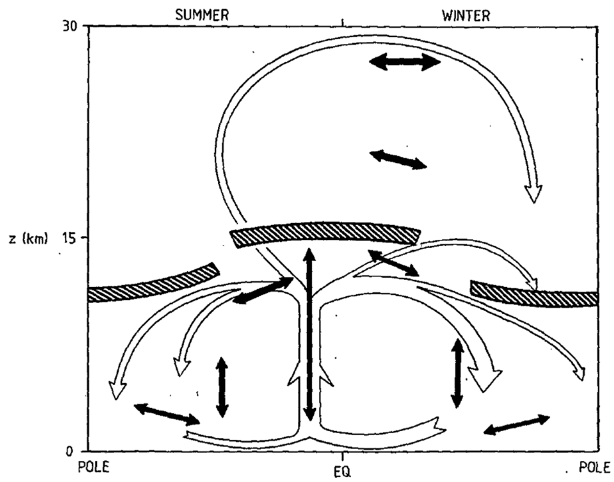

Figure 2. Schematic representation of the transport characteristics of a GCM at solstices. Broad arrows: advective transport. Thin double-headed arrows: locations and directions of diffusive transport. Figure 18 from *Plumb and Mahlman* [1987]. ©American Meteorological Society. Used with permission.

the nondivergent component of the Lagrangian-mean meridional circulation, *Plumb and Mahlman* [1987] overcame this problem and obtained a circulation [*Plumb and Mahlman*, 1987, Figure 5b] that better resembles that in Figure 1. Moreover, this transport circulation equals the residual mean circulation, at least for small-amplitude adiabatic disturbances. However, *Plumb and Mahlman* [1987] also found that the overall transport in their GCM at solstices included important additional contributions from diffusive processes as summarized in Figure 2.

3.4. Mixing and Transport Barriers

Two-way mixing (including the diffusive processes of *Plumb and Mahlman* [1987] discussed in section 3.3) and the rapid stirring of air parcels is now recognized as an important component of stratospheric transport [*Plumb*, 2002; *Shepherd*, 2007]. It results from the same waves that drive, for instance, the residual-mean meridional circulation through wave breaking and dissipation (see section 4) and is quasi-horizontal for planetary and synoptic-scale Rossby waves. Stirring by planetary waves occurs predominantly in winter in a huge midlatitude "surf zone" [*McIntyre and Palmer*, 1983, 1984], while that for the synoptic-scale waves occurs throughout the year in the subtropical lower stratosphere, and above and poleward of the subtropical jets. This stirring by the synoptic scale waves above the subtropical jets extends as high as 25 km in summer [*Haynes and Shuckburgh*, 2000]. In general, the transport associated with the stirring is rapid compared to the time taken for horizontal advection by the diabatic or residual-mean circulations. Hence, to a good approximation, the mean stratospheric mass transport can be represented by a combination of vertical advection by the diabatic or residual-mean circulations plus the two-way horizontal (isentropic) mixing [e.g., *Plumb*, 2007].

Mixing within the surf-zone creates sharp potential vorticity gradients at its edges which act as transport barriers around the winter polar vortex and in the Subtropics. A second subtropical barrier occurs in the summer hemisphere though the reasons for this are unclear [*Plumb*, 2002]. These barriers divide the stratosphere into four distinct regions: the summer hemisphere, the Tropics, the surf-zone and the winter polar vortex [*Plumb*, 2002]. *Plumb* [2007, Figure 3] summarizes the relative transport contributions of the meridional circulation and mixing in the four regions. Briefly, the meridional circulation is weak in summer but strong in the other regions (i.e., there is strong tropical upwelling followed by poleward and then downward motion in the winter hemisphere). Mixing or stirring is strong within the surf zone and dominates over the horizontal transport by the meridional circulation. Consequently, within this region, *Pendlebury and Shepherd* [2003] found, at least for a mechanistic model, that the meridional component of the Lagrangian-mean velocity deduced from air parcel motions bears little relation to the meridional component of the TEM velocity though, in general, there was a much better correspondence between the vertical components. Conversely, outside the surf zone, where the mixing and wave dissipation is relatively weak, the Lagrangian-mean and TEM (residual-mean) velocities were in good quantitative agreement [*Pendlebury and Shepherd*, 2003].

3.5. Mean Age of Air

A single transport diagnostic that combines the effects of the transport by the slow overturning diabatic or residual-mean circulations with the two-way mixing (including numerical and, in some cases, explicit diffusion in model simulations) and the rapid stirring of air parcels is the "mean age-of-air" [e.g., *Hall and Plumb*, 1994]. Age of air is generally defined as the transit time for the air to have traveled to a particular location from an initial location, usually taken to be its stratospheric entry point at the tropical tropopause

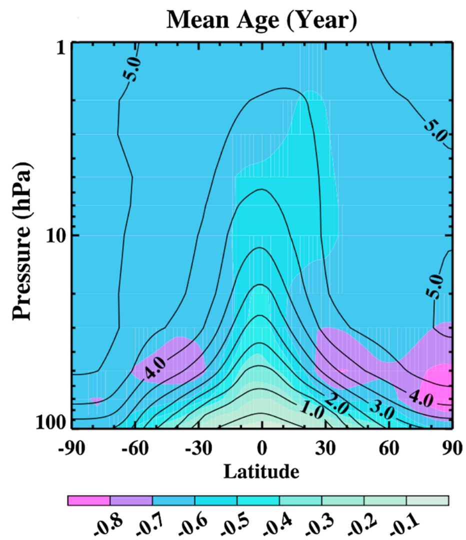

Figure 3. Annual mean age of air in years simulated by a CCM for the year 2000 (contours) and the simulated change in age from 2000 to 2080 (colors). Figure 2a from *Li et al.* [2012]. ©American Geophysical Union. Used with permission.

[*Waugh and Hall*, 2002]. However, because of mixing, an air parcel will be made up of a mixture of air that has traveled over many different transport pathways each with a different transit time. Hence, there is no single age for the parcel but, instead, an age spectrum [*Kida*, 1983b] and associated mean age. A characteristic feature of the combination of mixing, stirring, and advection by the meridional circulation is a mean age distribution with isopleths bulging upward in the Tropics and sloping down toward high latitudes with the oldest air found at the highest altitudes at all latitudes (e.g., Figure 3). Without the two-way transport due to the quasi-isentropic stirring, the oldest air would simply be in the polar lower stratosphere as a result of the transport by the overturning circulation.

4. Driving Mechanisms and Turnaround Latitudes

4.1. Wave Driving and Gyroscopic Pumping

Qualitatively, the underlying mechanism for the persistent poleward mass flow in the middle and upper winter stratosphere is the "extratropical pump" [*Holton et al.*, 1995] or, perhaps, more appropriately the "Rossby-wave pump" [*Plumb*, 2002] since it is now known that even small wave forcing close to the Equator can be as significant as the extratropical wave driving [*Plumb and Eluszkiewicz*, 1999]. As discussed in *Holton et al.* [1995], the pumping is the nonlocal effect of the wave drag from dissipating upward propagating waves from the troposphere. Drag from the dominant planetary-scale Rossby waves can only be westward, and consequently, the pumping action is one way with the air driven poleward to conserve angular momentum. This, in turn, sucks up air in the Tropics and pushes it down in the middle and high latitudes, at least in the steady state limit. From simple kinematic considerations, the flow has to be upward in the Tropics and downward in middle and high latitudes; otherwise, the circulation would require a reverse pole-to-equator flow at higher levels and there is no corresponding eastward Rossby-wave drag to balance the angular momentum budget. As the mechanism for the poleward flow involves a westward force causing air to move poleward due to the Earth's rapid rotation, some leading researchers [e.g., *McIntyre*, 2000] also refer to this as "gyroscopic pumping" though, arguably, it is the wave dissipation and forcing rather than the gyroscopic mechanism that is the most important aspect. In particular, the wave force addresses the original concerns of *Dobson et al.* [1929], *Brewer* [1949], and *Murgatroyd and Singleton* [1961] regarding the possible nonconservation of angular momentum (see section 2) as the waves themselves transport and deposit the momentum required to balance the angular momentum budget [e.g., *Andrews et al.*, 1987].

Conflicting results have been obtained for the latitudes where the Rossby-wave forcing drives the intraseasonal and interannual variations in the Brewer-Dobson circulation. For instance, the tropical upwelling was found by *Zhou et al.* [2012] to be well correlated with the subtropical wave force, whereas *Ueyama and Wallace* [2010] and *Ueyama et al.* [2013] showed a significant correlation with the high latitude forcing. Possible reasons for this discrepancy, suggested by *Ueyama et al.* [2013], were a poor representation of the low latitude wave force in the data sets they used and an over emphasis of the relative importance of the high latitude wave force due their use of correlation coefficients rather physically based diagnostics.

The single-cell poleward transport in the winter hemisphere that extends into the middle and upper stratosphere has become known as the "deep branch" of the Brewer-Dobson circulation [e.g., *Birner and Bönisch*, 2011]. In addition separate, faster, "shallow branches" are observed in both hemispheres throughout the year. Again, these almost certainly result from Rossby-wave pumping though now the synoptic-scale waves that are responsible are present throughout the year in the subtropical lower stratosphere and drive a poleward flow there and in the upper troposphere in both hemispheres (Figure 2 in *Plumb* [2002]). Notably in the *Birner and Bönisch* [2011] study, this separation into two branches was made using diagnostics based on the residual circulation alone.

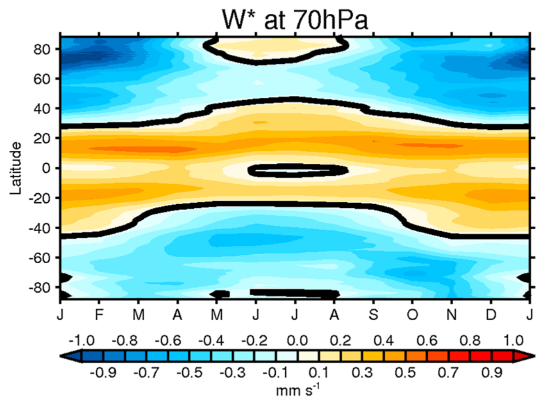

Figure 4. Mean residual vertical velocities $\overline{w}^*$ at 70 hPa from ERA-Interim for 1989–2009. The thick black lines in the Subtropics denote the position of the turnaround latitudes where the residual-mean vertical velocities, $\overline{w}^*$, change sign from upward to downward. Note that the turnaround latitudes should not be confused with the subtropical transport barriers that form the latitudinal boundaries of the "tropical-pipe" (section 3.4) [*Plumb*, 1996; *Plumb and Eluszkiewicz*, 1999]; i.e., the tropical upwelling region and tropical-pipe are not the same. Adapted from *Seviour et al.* [2012, Figure 4a].

In contrast to the stratospheric Brewer-Dobson circulation, the mesospheric mass transport is from the summer to winter pole. This is because at these levels the dominant wave pumping is from buoyancy or gravity waves which can exert both an eastward and westward drag on the mean flow [*Plumb*, 2002; *Alexander et al.*, 2010]. The underlying stratospheric winds selectively filter the upward propagating gravity waves leaving just a net eastward drag and equatorward pumping in the summer mesosphere and a net westward drag and poleward pumping in the winter hemisphere [*Plumb*, 2002]. Hence, the resultant mesospheric transport is from pole to pole [e.g., *Murgatroyd and Singleton* [1961, Figure 2].

While the wave-driving paradigm provides the underlying reason for the observed Brewer-Dobson and mesospheric circulations, there are also important features of these circulations, particularly in the Tropics, that cannot be wholly explained by a response to wave drag and continue to remain poorly understood. One example is the observed maximum in the tropical upwelling on the summer side of the Equator (e.g., Figure 4) which, based on results from simple numerical experiments, *Plumb and Eluszkiewicz* [1999] argue can only be properly explained if the annual cycle in the diabatic heating, $\overline{Q}$, is considered too. On the other hand, the main features of the morphology of the annual mean tropical upwelling in the lower stratosphere, such as the local minimum at the Equator (e.g., Figure 4), appear, at least, to be robust in both models and analyses [e.g., *Geller et al.*, 2008]. Further discussion of the limits to the applicability of the wave-driving paradigm, particularly in the deep Tropics, can be found in the excellent reviews of *Plumb* [2002] and *Haynes* [2005].

4.2. The Downward-Control Principle

A rather useful approach for quantifying the wave driving or pumping when analyzing the Brewer-Dobson circulation is the classic "downward-control principle" of *Haynes et al.* [1991]. Using the TEM formulation of *Andrews and McIntyre* [1976, 1978c], which was introduced here in section 3.2, *Haynes et al.* [1991] derived an expression for the extratropical residual-mean vertical velocity, $\overline{w}^*$, in terms of the vertical integral of the mean zonal forces (e.g., wave drag), $\overline{\mathcal{F}}$, acting above z. In the steady state limit, this takes the form

$$\overline{w}^*(\phi, z) = \frac{1}{\rho_0 \cos\phi} \frac{\partial}{\partial\phi} \left[\int_z^\infty \left(\frac{\rho_0 a \overline{\mathcal{F}} \cos^2\phi}{\overline{m}_\phi} \right) dz' \right], \tag{7}$$

where the integration is up a line of constant zonal mean absolute angular momentum (again see the Notation section for the symbol definitions). Near to the Equator (e.g., within $\sim \pm 15°$), these lines no longer span the vertical [*Haynes et al.*, 1991; *Rosenlof and Holton*, 1993], and hence, the formula can only be applied sufficiently far from the Equator. Nonetheless, the total tropical upwelling mass flux, for instance, can be deduced from the total extratropical downwelling mass flux by mass continuity (section 5.2) [*Holton*, 1990; *Rosenlof and Holton*, 1993; *Rosenlof*, 1995].

Haynes et al. [1991] considered a range of time-dependent solutions for their downward-control principle (see *Haynes* [2005] for a succinct review), and based on this, *Rosenlof and Holton* [1993] concluded that for sufficiently long time-scales and sufficiently large horizontal scales, the expression (7) can be used to estimate the mean meridional residual circulation of the lower stratosphere. Also using the results of *Haynes et al.* [1991], *Rosenlof and Holton* [1993] argued that the seasonal mean is sufficient for the steady state assumption to be valid.

The value of the downward-control principle is that the mass flow across a particular level, z, can be determined solely from a knowledge of the zonal forces $\overline{\mathcal{F}}$ above that level. In the stratosphere and mesosphere,

$\overline{\mathcal{F}}$ essentially results from the momentum deposition from dissipating Rossby and gravity waves. Therefore, insofar as $\overline{\mathcal{F}}$ can be decomposed into contributions from different wave types (e.g., planetary and synoptic-scale Rossby waves, and smaller-scale buoyancy or gravity waves), the downward-control principle and, in particular, equation (7) can be used to quantify the relative contributions of these different wave types to the driving of the Brewer-Dobson circulation [e.g., *Li et al.*, 2008] (see also section 5.2).

4.3. Turnaround Latitudes and Tropical Upwelling

An important legacy of the studies of *Rosenlof and Holton* [1993] and *Rosenlof* [1995], apparent in recent research, is the widespread use of the integrated tropical upwelling mass flux in the lower stratosphere as a standard (scalar) metric for quantifying the overall strength of the Brewer-Dobson circulation [e.g., *Butchart and Scaife*, 2001; *Austin and Li*, 2006; *Li et al.*, 2008; *McLandress and Shepherd*, 2009; *Butchart et al.*, 2010b; *Li et al.*, 2010; *Deushi and Shibata*, 2011; *Garny et al.*, 2011; *Okamoto et al.*, 2011; *Seviour et al.*, 2012; *Bunzel and Schmidt*, 2013; *Oberländer et al.*, 2013]. This metric is strictly only physically meaningful, at least for quantifying the strength of the mass circulation, if the mass flux is calculated over all tropical latitudes where the residual-mean vertical velocities are positive (i.e., upward) and not a fixed latitude band centered on the Equator as in some studies [e.g., *Garcia and Randel*, 2008; *Randel et al.*, 2008; *Yang et al.*, 2008; *Calvo and Garcia*, 2009]. *Rosenlof* [1995] referred to the boundaries of the tropical upwelling region as "turnaround latitudes" as the residual-mean vertical velocities, $\overline{w}^*$, change sign there from upward to downward. Figure 4 (adapted from *Seviour et al.* [2012, Figure 4a]) shows the significant seasonal variations in the position of these turnaround latitudes at 70 hPa (∼19 km) and, hence, why using an average following the seasonal movement of the upwelling region is more appropriate for calculating the annual mean mass flux entering the stratosphere than an average based on fixed latitudes. The 70 hPa level is generally chosen for this metric [e.g., *Rosenlof*, 1995; *Seviour et al.*, 2012] since it is high enough to exclude any two-way vertical exchange in the tropical tropopause layer but low enough to include most of the deep branch of the Brewer-Dobson circulation. On the other hand, most of the shallow branch is, arguably, below 70 hPa, and therefore, the tropical upwelling at this level only really quantifies the strength of the circulation in the deep branch. Above 70 hPa, in-mixing from the Extratropics [*Neu and Plumb*, 1999; *Plumb and Eluszkiewicz*, 1999] also complicates the simple physical interpretation of the upwelling as a measure of the overall strength of the mass circulation.

5. Diagnosing the Brewer-Dobson Circulation

5.1. Observations

Prior to the development of the TEM theory by *Andrews and McIntyre* [1976, 1978c], diagnosing the Brewer-Dobson circulation from observed meteorological variables (i.e., winds and temperatures) had proved to be conceptually problematic. The conventional two-cell Eulerian-mean circulation [*Vincent*, 1968] was not physically consistent with the observed mass circulation, and while the diabatic circulation of *Murgatroyd and Singleton* [1961] better represented the mass transport, the angular momentum budget could not be fully explained (see sections 2 and 3). Although both these issues were resolved with the TEM theory (again see section 3), another serious problem was the lack of global measurements of stratospheric winds and temperatures. Even when, soon after the development of the TEM theory, global analyses of the stratosphere became possible through the availability of data from operational satellites [e.g., *Bailey et al.*, 1993]; the mean meridional circulation, $(\overline{v}, \overline{w})$, and also $(\overline{v}^*, \overline{w}^*)$, remained poorly constrained, and noisy [*Iwasaki et al.*, 2009], until the development of the latest state-of-the-art reanalyses such as the European Centre for Medium-range Weather Forecasts (ECMWF) Interim Reanalysis (ERA-Interim) [*Dee et al.*, 2011] or the Japanese 25-Year Reanalysis (JRA-25) [*Onogi et al.*, 2007]. Indeed, when *Iwasaki et al.* [2009] compared the Brewer-Dobson circulation in five reanalysis data sets—JRA-25 [*Okamoto et al.*, 2011], ECMWF ERA-40 [*Uppala et al.*, 2005], ERA-Interim [*Dee et al.*, 2011], National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) [*Kalnay et al.*, 1996], and NCEP/Department of Energy (NCEP/DOE) [*Kanamitsu et al.*, 2002]—they found that apart from those obtained from ERA-Interim or JRA-25, the zonal-mean vertical velocities in the lower stratosphere were very noisy. They also concluded that ERA-Interim had a more realistic thermodynamic-dynamic equilibrium as a result of the use of a four-dimensional variational method and bias correction scheme. Therefore, arguably, amongst the current best estimates of the observed TEM, residual mean circulation for the stratosphere are almost certainly those shown in Figure 5 inferred from ERA-Interim (see *Seviour et al.* [2012] for details). However, no complete

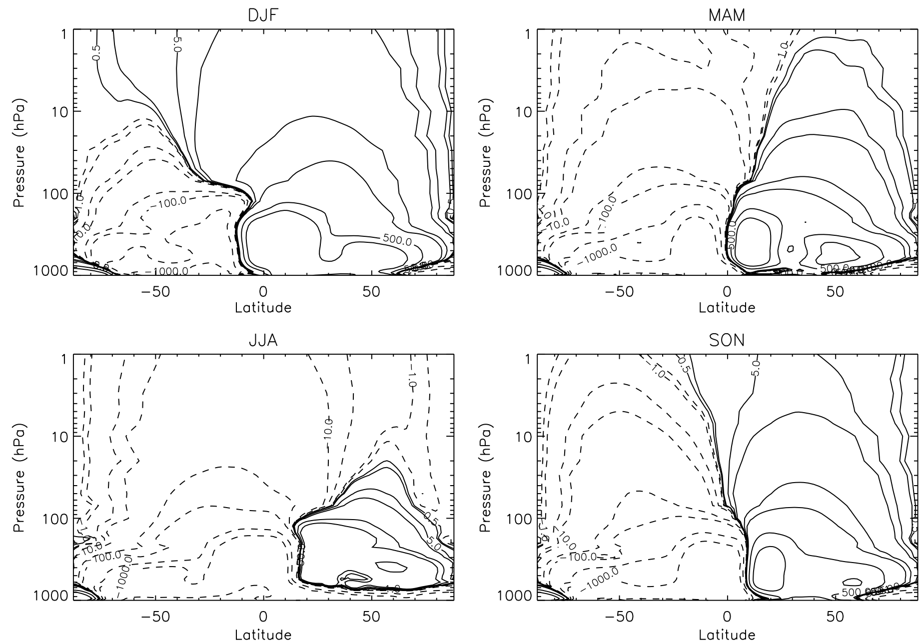

Figure 5. Seasonal mean TEM stream function, ψ, from ERA-Interim for 1989–2009. Contours have units of $kg\,m^{-1}\,s^{-1}$ and are spaced logarithmically. Dashed contours represent negative values. Figure 3 from *Seviour et al.* [2012].

assessment or intercomparison of the Brewer-Dobson circulation has been published including all the currently available analyses and reanalyses [*Fujiwara et al.*, 2012].

An alternative to calculating the residual mean circulation $(\overline{v}^*, \overline{w}^*)$ directly from equations (4) and (5) is to use the iterative method described by *Murgatroyd and Singleton* [1961] to solve the TEM thermodynamic equation (6) for $\overline{v}^*$ and $\overline{w}^*$, assuming quasi-geostropic dynamics which allows the rectified eddy forcing terms on the right-hand side of (6) to be neglected (see section 3.2). *Solomon et al.* [1986], *Gille et al.* [1987], and *Eluszkiewicz et al.* [1996] applied this approach successfully to satellite data though numerical model results have since shown that the method is not quite so successful for constructing a *quantitative* representation of the global residual-mean circulation (section 5.2) [*Beagley et al.*, 1997]. Further approximations can, however, be made to the TEM thermodynamic equation (6) in the tropical lower and middle stratosphere as now both the rectified eddy forcing terms on the right-hand side of (6) *and* the horizontal advection term are negligible [*Rosenlof*, 1995] and in the steady state limit

$$\overline{w}^* \frac{\partial \overline{\theta}}{\partial z} \approx \overline{Q}, \tag{8}$$

where $\overline{Q}$ can be determined using a radiative-transfer algorithm [*Yang et al.*, 2008]. Residual-mean vertical velocities calculated this way are usually referred to as "diabatic residual-mean vertical velocities." Importantly, *Rosenlof* [1995] found that these diabatic residual-mean vertical velocities do provide reliable estimates of the zonally averaged ascent rates in the tropical lower stratosphere, though *Rosenlof* used the iterative method of *Murgatroyd and Singleton* [1961] rather than equation (8) to calculate $\overline{w}^*$ and included a correction (of about 15% in the tropics) constraining the global average $\overline{w}^*$ on a pressure surface to be zero (see *Rosenlof* [1995] for details).

The residual-mean circulation can also be obtained indirectly from eddy heat and momentum fluxes by downward control (i.e., equation (7)) as pioneered by *Holton* [1990] when he estimated the global mass exchange between the stratosphere and troposphere. As would be expected from the steady state limit of the downward-control calculation, *Rosenlof and Holton* [1993] found the method worked better in the solstice seasons than in equinox seasons, though they also concluded that to accurately estimate the residual mean circulation would require the additional knowledge of the unresolved zonal forcing from gravity waves [see also *Seviour et al.*, 2012].

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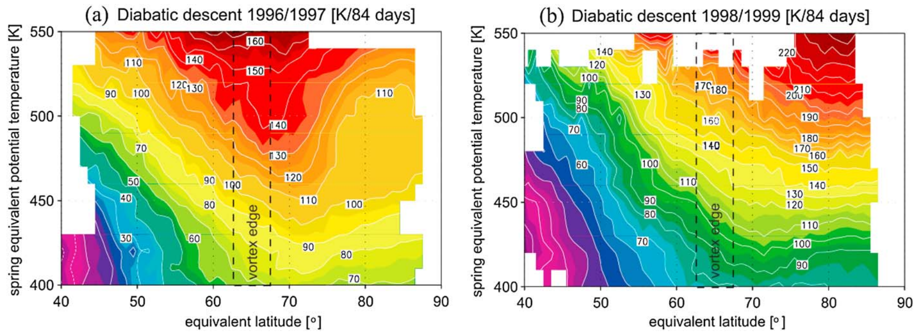

Figure 6. The Lagrangian descent (K per 84 days) as a function of equivalent latitude and spring equivalent potential temperature for the time period from 15 December to 10 March for (a) winter 1996/1997 and (b) winter 1998/1999. Note the different color scales. Figure 6 from *Tegtmeier et al.* [2008]. ©American Geophysical Union. Used with permission.

Mean mass transport within the stratosphere has also been deduced from observed winds and temperatures using trajectory calculations [e.g., *Rosenfield et al.*, 1994; *Krüger et al.*, 2008; *Tegtmeier et al.*, 2008]. Typically, these calculations are based on a combination of quasi-horizontal advection along isentropic surfaces by the observed winds plus cross isentropic advection by $\dot{\theta}$ ($\equiv \frac{d\theta}{dt}$), where $\dot{\theta}$ is obtained from a diabatic heating model using observed temperatures. Using this method, *Tegtmeier et al.* [2008] calculated the mean Lagrangian diabatic descent within the polar vortex for 47 Arctic winters from 1957/1958 to 2003/2004 by averaging the descent obtained from trajectories at, more or less, the same "equivalent latitude" (i.e., the latitude that encloses the same area as a circumpolar isoline of potential vorticity on an isentropic surface; see *Butchart and Remsberg* [1986] for the complete definition of equivalent latitude). This led to a description of the meridional structure of the downwelling in the deep branch of the Brewer-Dobson circulation with the strongest descent occurring near the vortex edge in cold undisturbed winters and in the vortex core during disturbed winters [see also *Manney et al.*, 1994]. Examples from *Tegtmeier et al.* [2008] for an undisturbed (1996/1997) and a disturbed (1998/1999) winter are shown in Figure 6, with the disturbed winter 1998/1999 showing stronger vortex-averaged diabatic descent than the undisturbed winter 1996/1997. Moreover, *Tegtmeier et al.* [2008] found the disturbed winters showed stronger vortex-averaged diabatic descent, in general, which is perhaps not surprising given the higher level of wave activity in those winters. For the Southern Hemisphere, *Manney et al.* [1994] found that for the winters of 1992 and 1993, the strongest diabatic descent above the 900 K isentropic level (~50 km) was in the center of the vortex while below that level the strongest descent was near the vortex edge as in the undisturbed northern winters.

Diallo et al. [2012] used a different approach to diagnose the Brewer-Dobson circulation based on the age of stratospheric air, though again this involved trajectory calculations using analysed winds from ERA-Interim plus heating rates (see *Diallo et al.* [2012] for details of their method). Unlike the residual-mean circulation diagnostics obtained from ERA-Interim by *Seviour et al.* [2012], the age-of-air diagnostics take into account mixing and stirring, though *Diallo et al.* [2012] also noted that by using in their trajectory calculations diabatic heating rates instead kinematic vertical velocities (i.e., the actual ERA-Interim vertical velocities) they obtained younger ages, apart from in the middle and upper Northern Hemisphere stratosphere where the diabatic trajectories produced older ages. The reasons for this are not yet properly understood though using the Modern-Era Retrospective Analysis for Research and Applications (MERRA) [*Rienecker et al.*, 2011], *Schoeberl and Dessler* [2011] also found that the ages obtained from kinematic trajectories were older than those obtained from diabatic trajectories throughout the stratosphere. The distribution of ages calculated by *Diallo et al.* [2012] using the diabatic trajectories were, nonetheless, consistent with the existence of the shallow and deep branches of the Brewer-Dobson circulation [*Birner and Bönisch*, 2011]. Using the age of air diagnostics, *Diallo et al.* [2012] showed further that both branches were modulated by the variability in tropical upwelling, and also by the subtropical transport barriers for the shallow branch, with a change in the sign of the overall modulation at ~25 km (see *Diallo et al.* [2012] for more details). Another benefit of the age-of-air diagnostic is that it allows the Brewer-Dobson circulation inferred from meteorological reanalyses to be compared, in principle, with that inferred from the observed transport of long-lived tracers (e.g., section 6.1).

Historically, the observed transport of long-lived tracers such as ozone [e.g., *Dobson et al.*, 1929; *Dobson*, 1956], water vapor [e.g., *Brewer*, 1949], or radioactive isotopes [e.g., *Newel*, 1963; *Sheppard*, 1963] provided only qualitative information about the Brewer-Dobson circulation. In more recent studies, quantitative information has been obtained in the form of empirical estimates of the age of air [e.g., *Hall and Plumb*, 1994] derived from observations of carbon dioxide [e.g., *Boering et al.*, 1996; *Andrews et al.*, 2001] or sulphur hexafluoride, SF_6 [e.g., *Hall and Waugh*, 1998; *Stiller et al.*, 2008]. However, these really are only quantitative *estimates* since the mean age derived this way can have very large uncertainties as discussed, for instance, in section 2 of *Ray et al.* [2010]. Furthermore, most of the studies used in situ measurements and therefore are unable to provide a complete global description of the stratospheric transport. One study by *Stiller et al.* [2008] has presented estimates of the global distribution of the mean age derived from satellite measurements of SF_6, but the reliability of these results is questionable as the effects of the known mesospheric losses of SF_6 were not taken into account in the estimate. The neglect of the mesospheric losses can lead to a mean age that is up to 1.5 years too old, at least compared to the mean age derived from measurements of CO_2 [*Neu et al.*, 2010].

Apart from the age of air diagnostics of *Stiller et al.* [2008], the observations of tracers generally do not provide global seasonally averaged quantitative estimates of the Brewer-Dobson circulation. Nonetheless, useful quantitative estimates of vertical transport can be derived from trace-gas observations in the regions and seasons where and when there is relatively unmixed ascent or descent. For instance, a number of studies have used trace-gas observations to estimate ascent rates for the tropical lower stratosphere [e.g., *Mote et al.*, 1996; *Niwano et al.*, 2003] and descent rates within the polar vortex for Antarctic [e.g., *Abrams et al.*, 1996a; *Allen et al.*, 2000, *Kawamoto and Shiotani*, 2000] and Arctic [e.g., *Abrams et al.*, 1996b; *Greenblatt et al.*, 2002, 2003; *Ray et al.*, 2002] winters. The estimates of ascent rates in the equatorial lower stratosphere from water vapor by *Mote et al.* [1996] are particularly interesting as they depend on the existence of an annual cycle in the Brewer-Dobson circulation causing an annual cycle in the tropical tropopause temperatures [e.g., *Yulaeva et al.*, 1994] which, in turn, imprints a similar cycle on the water vapor entering the stratosphere. *Mote et al.* [1996] and many subsequent studies refer to this as the "water vapor tape recorder," and it is the "speed" of this tape recorder or, more precisely, the rate of upward propagation of the annual cycle signal in the water vapor that provides a quantitative estimate of the ascent rate.

5.2. Models

Although the Brewer-Dobson circulation is two dimensional (i.e., some form of zonal average such as the Lagrangian or transformed-Eulerian mean as defined in section 3), the underlying wave-driving mechanism is three dimensional (3-D) and therefore only 3-D models are considered in this review. It is also worth noting that a novel new 3-D residual-mean circulation has recently been defined using time means instead of spatial means [*Kinoshita and Sato*, 2013; *Sato et al.*, 2014], but this will not be assessed any further here since, again, as outlined at the beginning of section 3, the definition of the Brewer-Dobson circulation used in this review is restricted to the two-dimensional equator-to-pole mean stratospheric mass circulation.

As mentioned in sections 2 and 3, *Kida* [1983a, 1983b] was almost certainly the first to study stratospheric transport characteristics in a 3-D GCM, albeit a hemispheric model. Global studies soon followed and several [e.g., *Plumb and Mahlman*, 1987; *Boville*, 1995; *Beagley et al.*, 1997; *Butchart and Austin*, 1998] used the dynamically consistent model winds and temperatures to diagnose the seasonally averaged TEM residual mean circulation. *Plumb and Mahlman* [1987] used their results to better understand the different circulation models used to describe the Brewer-Dobson circulation (see section 3), while *Boville* [1995], *Beagley et al.* [1997], and *Butchart and Austin* [1998] each demonstrated that their models were able to qualitatively simulate all the main features of the residual mean circulation. In addition, the accuracies of the diabatic and downward-control approximations (see section 5.1) to the TEM residual stream function (defined by the terms on the right-hand side of equations (4) and (5)) were examined by *Beagley et al.* [1997], and while their attempt to reconstruct the residual circulation directly from the diabatic heating was only partially successful, the downward-control estimate was found to work quite well, at least outside tropical regions. *Beagley et al.* [1997] also applied downward control in the manner discussed in section 4.2 to separate the planetary and parameterized gravity wave contributions to the driving of the midstratospheric polar downwelling though, as discussed below, subsequent multimodel studies [*Butchart et al.*, 2010a, 2010b, 2011] suggest that these particular calculations are probably somewhat model dependent.

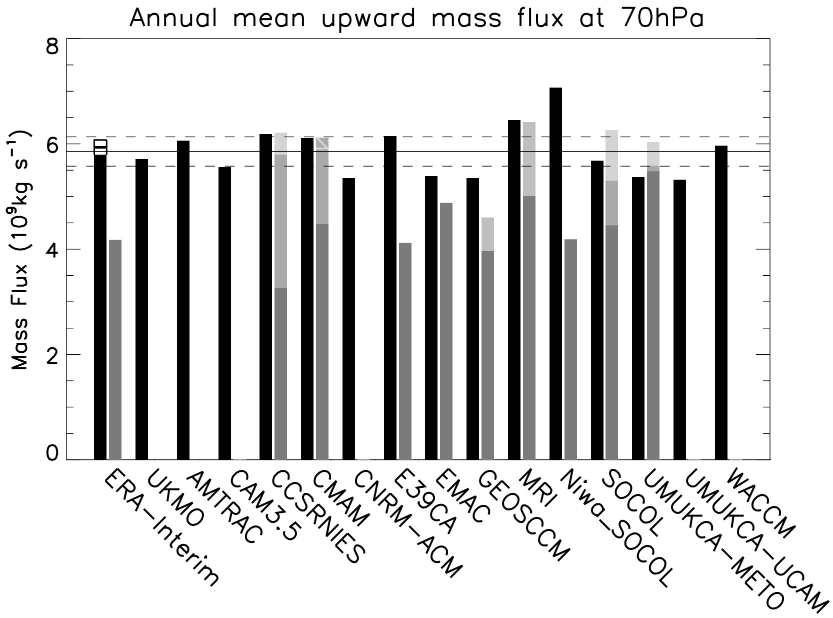

Figure 7. Annual mean simulated upward mass flux at 70 hPa from 11 different CCMs for 1980 to 1999, ERA-Interim reanalysis for 1989 to 2009 and UKMO analyses [*Swinbank and O'Neill*, 1994] for 1992 to 2001. Upwelling calculated from $\overline{w}^*$ is shown by black bars. Upwelling calculated by downward control (when drag diagnostics were available for the model) is split into contributions from resolved waves (dark grey), orographic gravity wave drag (grey), and nonorographic gravity wave drag (light grey). Gravity wave contributions are shown combined for the GEOSCCM and the MRI model. For some models and also the ERA-Interim reanalysis, gravity wave drag diagnostics were not available and therefore only the contribution from planetary (resolved) wave drag is shown. In practice, the downward-control estimates of the tropical upwelling were obtained from the total extratropical downwelling using mass continuity (see *Butchart et al.* [2011] for further details). The black horizontal lines show the multimodel mean and the intermodel standard error. The interannual standard error for the ERA-Interim reanalysis is shown by the unshaded part of the bar with the horizontal line at the midpoint being the multiyear mean. Figure 10a from *Butchart et al.* [2011]. ©American Geophysical Union. Used with permission.

During the last decade, or so, significant increases in computing power together with a recognition of the importance of the two-way interactions between climate change and ozone recovery [e.g., *Bekki et al.*, 2011; *Forster et al.*, 2011] have led to the development of a large pool of stratosphere-resolving CCMs [*Morgenstern et al.*, 2011] and their coordinated evaluation and intercomparison [*SPARC CCMVal*, 2010]. In particular, *Butchart et al.* [2011] compared the annual mean tropical upwelling between the turnaround latitudes (section 4.3) in near-identical 1980 to 1999 historical simulations from 11 of the CCMs described in *Morgenstern et al.* [2011]. Compared to ERA-Interim *Butchart et al.* [2011] found that all but one of the models accurately reproduced the locations of the turnaround latitudes with the annual cycle in the integrated upward mass flux between these turnaround latitudes generally well reproduced. On average, the annual mean tropical upwelling at 70 hPa agreed remarkably well with that obtained from ERA-Interim and there was reasonably good agreement among the individual models (see black bars in Figure 7).

Again *Butchart et al.* [2011] found that on average, the downward-control estimates of the upwelling (total height of the grey bars in Figure 7) agreed well with the directly calculated upwelling, with the parameterized orographic and nonorographic gravity wave drags contributing 21.1% and 7.1%, respectively, to the driving of the tropical upwelling at 70 hPa. However, these percentages varied considerably among the individual models (e.g., from 2.0% to 40.9% for the orographic gravity waves) though the spread generally does not appear to be related to model formulation or resolution. Understanding why the good agreement in the tropical upwelling between all the models and ERA-Interim is rather insensitive to the differences in the partitioning of the wave driving among the models is one of the key outstanding challenges in modeling the TEM residual mean circulation with obvious ramifications for identifying the causes of the projected trends in the Brewer-Dobson circulation (see section 6).

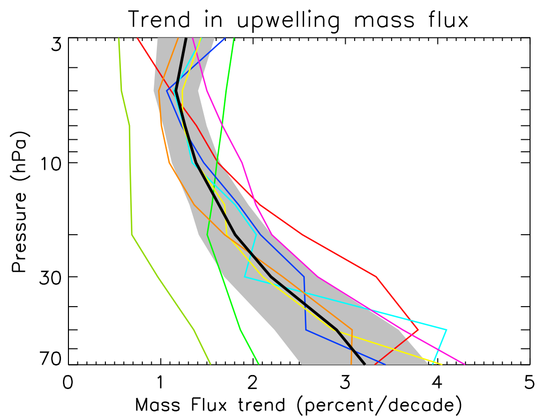

Figure 8. Projected trends in tropical upwelling in percent per decade based on a linear fit to the years 2006–2099 from RCP8.5 scenario simulations of eight stratosphere-resolving GCMs (see *Hardiman et al.* [2013] for details of the models and simulations). Tropical upwelling is calculated for each year, as the mass upwelling between the turnaround latitudes and takes into account the seasonal movement of these turnaround latitudes. The black line is the multimodel mean with the shading showing the intermodel standard error, scaled to represent a 95% confidence interval. Adapted from *Hardiman et al.* [2013, Figure 2, lower-right panel].

6. The Brewer-Dobson Circulation in a Changing Climate

6.1. Response to Climate Change

The first evidence of the Brewer-Dobson circulation responding to climate change was almost certainly that presented by *Rind et al.* [1990] based on a doubling of CO_2 amounts in short 3 year simulations of a GCM. The CO_2 doubling resulted in a stronger residual-mean circulation, broadly consistent with the latest multimodel projections [e.g., *Butchart et al.*, 2010b; *Hardiman et al.*, 2013]. Moreover, the interpretation given by *Rind et al.* [1990] involving a response to changes in the driving from *both* planetary and parameterized gravity waves is consistent with current understanding (see section 6.2). *Rind et al.* [1990] also noted some possible important ramifications of the stronger circulation, such as the faster removal of chlorofluorocarbons (CFCs) from the atmosphere and increased midlatitude column ozone. Indeed such an increase in midlatitude column ozone was obtained by *Mahfouf et al.* [1994] when they doubled CO_2 in a CCM and thereby obtained further circumstantial evidence for a faster Brewer-Dobson circulation. However, no quantitative projections of the response of the residual-mean circulation to climate change appear to have been published prior to *Butchart and Scaife* [2001].

Following *Rosenlof and Holton* [1993] and *Rosenlof* [1995], *Butchart and Scaife* [2001] used the net upward mass flux between the turnaround latitudes at 70 hPa to quantify the strength of the Brewer-Dobson circulation or, more particularly, the strength of (mostly) the deep branch of the Brewer-Dobson circulation (see section 4.3) in two 60 year transient simulations of a GCM and found an increase of roughly 3% per decade in response to increasing greenhouse gas (GHG) concentrations. The projected change in the tropical upwelling in the lower stratosphere and, more generally, the residual-mean circulation has now been analyzed [e.g., *Austin*, 2002; *Land and Feichter*, 2003; *Kodama et al.*, 2007; *Garcia and Randel*, 2008; *Li et al.*, 2008; *Calvo and Garcia*, 2009; *McLandress and Shepherd*, 2009; *Oman et al.*, 2009; *Li et al.*, 2010; *Deushi and Shibata*, 2011; *Garny et al.*, 2011; *Okamoto et al.*, 2011; *Bunzel and Schmidt*, 2013; *Lin and Fu*, 2013; *Oberländer et al.*, 2013; *Schmidt et al.*, 2013] and compared [e.g., *Butchart et al.*, 2006, 2010a; *Hardiman et al.*, 2013] in many GCMs and CCMs. Significantly all the models project, at least for the annual average, an acceleration of the residual-mean circulation making this, potentially, one of the more robust manifestations of GHG-induced climate change. On average, the projections for 70 hPa are converging toward a trend of roughly 2% per decade [*Butchart et al.*, 2010a] for the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions (SRES) GHG scenario A1B (medium) [see *Nakicenovic and Swart*, 2000] and ~3.2% per decade (e.g., Figure 8) for the more extreme Representative Concentration Pathway (RCP) 8.5 scenario [*Moss et al.*, 2010; *Riahi et al.*, 2007]. A response due to secular changes in ozone concentrations has been found too [e.g., *Li et al.*, 2008; *Oman et al.*, 2009; *McLandress and Shepherd*, 2009; *Oberländer et al.*, 2013].

Figure 8 shows that for the latest multimodel projections [*Hardiman et al.*, 2013], the trends in tropical upwelling occur throughout the depth of the stratosphere [see also *Kodama et al.*, 2007; *Garcia and Randel*, 2008; *Li et al.*, 2008; *Butchart et al.*, 2010a] but with a decrease in the percentage trend from the lower to middle stratosphere. This decrease has been found in other studies [e.g., *Li et al.*, 2008; *Butchart et al.*, 2010a, 2010b; *Oberländer et al.*, 2013], and it, too, is probably a ubiquitous feature of the projected changes in the residual-mean circulation. In particular, *Lin and Fu* [2013] have decomposed the trends in the Brewer-Dobson circulation in 12 of the CCMs described in *Morgenstern et al.* [2011] into the transition, shallow, and deep branches which they defined as having vertical extents of 100–70, 70–30, and above 30 hPa, respectively. An acceleration occurred in all three branches, although it was smaller in the deep branch,

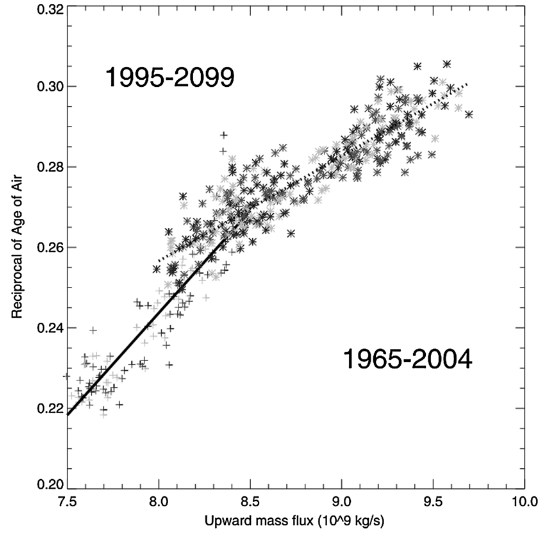

Figure 9. Stratospheric age of air, averaged poleward of 60°N at 47 hPa plotted as a function of tropical upwelling in two three-member ensembles of CCM simulations. The plus symbols denote results from a 1960–2005 ensemble using observed SST forcing, with a grey scale denoting the ensemble member. The asterisks denote similar results from a 1990–2099 ensemble using modeled SST forcings. The solid and dotted lines are linear regressions through the results for the 1960–2005 and 1990–2099 ensembles, respectively. Figure 4 from *Austin and Li* [2006]. ©American Geophysical Union. Used with permission.

consistent with the decrease in the percentage trend with altitude in Figure 8. Another notable feature of the projected changes in the Brewer-Dobson circulation is a narrowing of the upwelling branch in the upper troposphere and lower stratosphere [*Li et al.*, 2010; *Hardiman et al.*, 2013]. This contrasts with a widening of the tropical belt, which is also evident in the tropical tropopause layer in a changing climate [e.g., *Seidel and Randel*, 2007; *Seidel et al.*, 2008], despite some uncertainty over the details of the tropical expansion [e.g., *Birner*, 2010; *Davis and Rosenlof*, 2012]. Above ~20 hPa, the upwelling region widens in response to climate change [*Hardiman et al.*, 2013].

The seasonality also appears to be a ubiquitous characteristic of the response of the tropical upwelling at 70 hPa to climate change, with the largest increases in the mass flux generally projected to occur in the Boreal winter [*Butchart et al.*, 2006; *Li et al.*, 2008; *McLandress and Shepherd*, 2009; *Butchart et al.*, 2010a; *Deushi and Shibata*, 2011; *Garny et al.*, 2011; *Bunzel and Schmidt*, 2013]. Furthermore, because the tropical upwelling at 70 hPa mainly results from the deep branch of the Brewer-Dobson circulation (see section 4.3), which is predominately a winter hemisphere phenomenon (e.g., Figure 5), an important manifestation of the seasonality in the upwelling trends is a hemispheric asymmetry in the corresponding annual mean trends in the extratropical downwelling, with the largest increases in downwelling projected for the Northern Hemisphere. This would imply both an increase in the amplitude of seasonal cycle in the tropical upwelling mass flux and increased hemispheric asymmetry in the annual mean extratropical downwelling mass flux (e.g., see section 7.3), in response to climate change. However, since the deep branch of the circulation is also stronger in the Northern Hemisphere, both the seasonality and hemispheric asymmetry in the responses are less apparent when the trends are expressed as percentage changes as, for instance, in *Butchart et al.* [2010a, Figure 8]. It is also possible that the seasonality and hemispheric asymmetry of the circulation itself contributes to the corresponding seasonality and asymmetry in the trends though this cannot be properly verified due to the absence of a comprehensive understanding of how the wave drag, driving the circulation, responds to climate change (see section 6.2).

Support for the projected strengthening of the Brewer-Dobson circulation has been obtained from the evolution of the mean age-of-air tracer included in most CCMs. The first published results were probably those of *Austin and Li* [2006] reporting, for an ensemble of 1960 to 2100 CCM simulations, a linear relationship between the reciprocal of the mean age and the tropical upwelling mass flux averaged over the previous four years (Figure 9). *Austin and Li* [2006] argued persuasively that the high, statistically significant correlation between these two variables in Figure 9 is dominated by the long-term secular trends in both variables and suggest that the change in slope seen around 2005 is simply a consequence of a switch from using observed, to using modeled, sea surface temperature and sea ice (SST) forcing in the CCM. *Oman et al.* [2009] also concluded that the decreases in the mean age-of-air they obtained from a suite of CCM simulations in response to climate change were primarily caused by increases in upwelling in the tropical lower stratosphere. Likewise, *Deushi and Shibata* [2011] considered the seasonality and structure of the age response in 1960 to 2100 CCM simulations and again concluded that these were primarily related to details of the residual-mean circulation response, apart from a local maximum decrease in age over the Antarctic in December which they attributed to an earlier breakdown of the polar vortex.

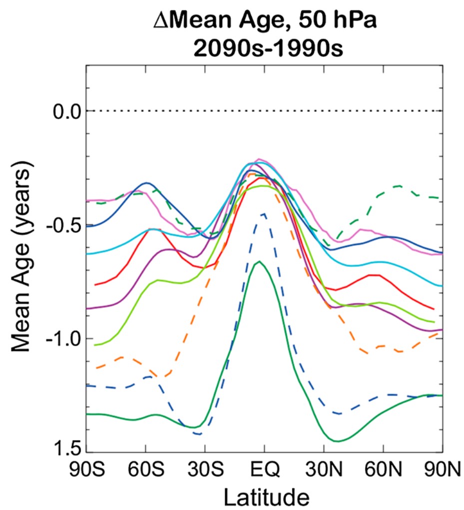

Figure 10. Mean age changes at 50 hPa during 21st century simulations from ten of the CCMs described in *Morgenstern et al.* [2011]. The mean age difference is the difference between an average of the last 10 years of the simulations (usually, 2090–2099), and an average over 1990–1999. All of the CCMs predict younger age at all latitudes at the end of the 21st century. Figure 5.18 (left panel) from *Neu et al.* [2010].

To better identify the processes that cause the decrease in mean age in a changing climate, *Li et al.* [2012] considered, in 21st century simulations of a CCM, long-term changes in the age spectra as well as the mean age. In these simulations, it turned out that the decrease in mean age (e.g., colors in Figure 3) was due to both an acceleration of the residual-mean circulation and a weakening of the recirculation of stratospheric air between the Tropics and midlatitudes (i.e., mixing through the subtropical transport barriers). The weaker recirculation decreased the tail of the age spectra and correspondingly reduced the amount of really old air masses [*Li et al.*, 2012]. Notably in this particular CCM, there was a very strong correlation between this weakening of the stratospheric recirculation and the increase in the residual-mean circulation. In the subtropical lower stratosphere mixing increased in response to climate change but the impacts of this were smaller than those from the acceleration of the residual-mean circulation [*Li et al.*, 2012].

As with the increase in tropical upwelling, a decreasing mean age at all latitudes in the lower stratosphere (50 hPa) appears to be a robust model response to GHG-induced climate change (Figure 10) [*Butchart et al.*, 2010b; *Neu et al.*, 2010]. On the other hand, when *Engel et al.* [2009] estimated the mean age from balloon-borne measurements of carbon dioxide and sulphur hexafluoride (SF_6), they concluded that if anything, the mean age had increased over the 30 years from 1975 to 2005, at least in the northern midlatitudes between 24 and 35 km (Figure 11). However, *Waugh* [2009] has pointed out that this result was obtained from just 27 balloon flights over the 30 year period and further emphasized the huge uncertainty in estimating mean age from CO_2 and SF_6 observations (see also section 5.1).

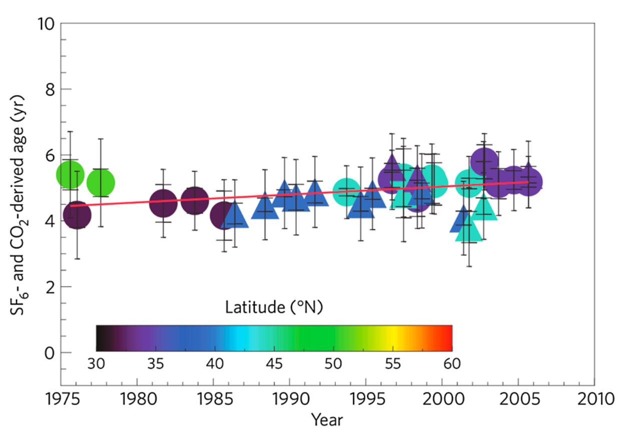

Figure 11. Mean age derived from balloon-borne measurements of SF_6 (circles) and CO_2 (triangles). The color code shows the (northern) latitude of the measurements. The inner error bars show the 1 sigma standard deviation of the mean-age values between 24 and 35 km for each individual balloon flight. The outer (larger) error bars denote the overall uncertainty of the mean-age value, including an assessment of the representativeness of a single profile observation. The overall uncertainty has been taken into account for the calculation of the long-term trend. Figure 3 from *Engel et al.* [2009].

The complexities of determining age of air trends from atmospheric trace species such as CO_2 or SF_6 have now been further assessed by *Garcia et al.* [2011] in a detailed modeling and theoretical study. A particular issue highlighted by *Garcia et al.* [2011] is the sparse spatial and temporal sampling of CO_2 and SF_6 used by *Engel et al.* [2009]. Indeed, age trends roughly consistent with those reported by *Engel et al.* [2009] can be obtained with similar sampling of CCM output even when the same CCM projects an overall decrease in the mean age [*Garcia et al.*, 2011]. On the other hand, *Garcia et al.* [2011] were unable to explain why in that CCM the midlatitude estimates of mean age derived from CO_2 were about 8 months younger than those derived from SF_6 [*Garcia et al.*, 2011]. In contrast, the mean ages derived

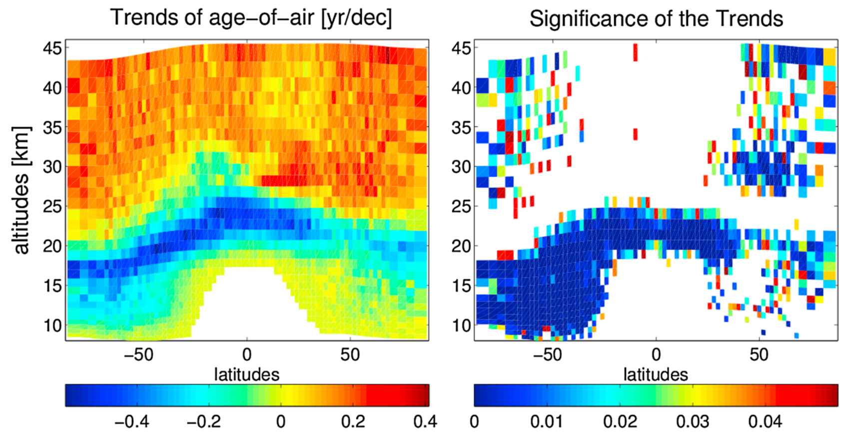

Figure 12. (Left) age trends (years per decade) derived from ERA-Interim for 1989 to 2010 using diabatic trajectories (see *Diallo et al.* [2012] for more details). (Right) *P*-value of its associated Student's statistical test of significance where *P*-value is less than 0.05. Figure 13 from *Diallo et al.* [2012].

from the observations of these two tracers by *Engel et al.* [2009] were very similar. Overall, *Garcia et al.* [2011] concluded by suggesting that as proxies for trends in the stratospheric mass circulation, the observations of trace species are rather ambiguous and to properly calculate the mean age would require observations of tracers that have strictly linear growth. Such observations are not available.

Although *Stiller et al.* [2012] eliminated the problem of the sparse data coverage of *Engel et al.* [2009] by using global vertical profiles of SF_6 observed from an orbiting spacecraft, the zonally averaged mean-age trends obtained were found to be spatially inhomogeneous. This is perhaps not surprising given the very short data record from September 2002 to January 2010. Nonetheless, *Stiller et al.* [2012] argue that their inhomogeneous spatial distribution of mean-age trends is, at least, consistent with the increased tropical upwelling projected by models *and* the results of *Engel et al.* [2009] for middle latitudes and can be explained with the hypothesis that the intensification of the overturning circulation is accompanied by the simultaneous weakening of the subtropical mixing barriers allowing increased recirculation of stratospheric air which can change the shape of the age spectra [e.g., *Li et al.*, 2012]. This hypothesis is, however, not consistent with the model results of *Li et al.* [2012] noted above which indicated a weakening of the recirculation. It also fails to explain why, in general, models project a decreasing mean age more or less everywhere in the stratosphere (e.g., Figures 3 and 10), nor does it explain the correlation in model results between tropical upwelling trends and mean-age trends (e.g., Figure 9). On the other hand, the positive age-of-air trends found in the northern midlatitudes above 25 km by *Stiller et al.* [2012], and also *Engel et al.* [2009], are consistent with the age trends obtained from ERA-Interim for 1989 through 2010 by *Diallo et al.* [2012] though, again, the positive ERA-Interim age trends are only marginally statistically significant (Figure 12). *Monge-Sanz et al.* [2013] also inferred from ERA-Interim a small but statistically significant positive age trend over the northern midlatitudes above 24 km, for 1990 to 2009.

In the lower stratosphere, the ERA-Interim age trends calculated by *Diallo et al.* [2012] were both negative and statistically significant (Figure 12), consistent with an acceleration of the shallow branch of the Brewer-Dobson circulation [*Bönisch et al.*, 2011], but not the −5% per decade trend in tropical upwelling at 70 hPa inferred from ERA-Interim by *Seviour et al.* [2012]. However, this upwelling trend was considered to be unreliable by *Seviour et al.* [2012] because it was inconsistent with the negative temperature trend, assuming a mainly adiabatic temperature response at that level (70 hPa) to the changes in upwelling. Thus, the most reliable trends inferred from ERA-Interim for the Brewer-Dobson circulation are, arguably, those obtained from the age of air diagnostics by *Diallo et al.* [2012] or the alternative but similar results of *Monge-Sanz et al.* [2013]. Importantly, trends obtained this way include both the effects of changes in the residual-mean circulation and changes in mixing and stirring. *Diallo et al.* [2012] conclude by suggesting that the shallow and deep branches of the Brewer-Dobson circulation have evolved in different directions in the period 1989 to 2010 [see also *Bönisch et al.*, 2011], but they also add a caveat that reliable long-term trends are somewhat difficult to estimate from only 22 years of data.

Despite the absence of a consensus, and the unreliability of the trends in the observed residual-mean circulation and the age-of-air diagnosed from either reanalyses or the transport of observed tracers, indirect evidence for a strengthening of the Brewer-Dobson circulation has been observed in other stratospheric parameters. For instance, radiosonde observations indicate a cooling of the tropical lower stratosphere from 1979 to 2003 which *Thompson and Solomon* [2005] argue is probably an adiabatic response to increased tropical upwelling. Likewise, *Young et al.* [2012] found that satellite and radiosonde stratospheric temperature changes from 1979 to 2005 were statistically consistent with a strengthening of the Brewer-Dobson circulation but only after the effects of the interannual variability were removed by regression (see *Young et al.* [2012] for further details). The contrasting latitudinal structure of recent (1979–2006) stratospheric temperature and ozone trends was also found to be consistent with large-scale increases in the stratospheric overturning Brewer-Dobson circulation [*Thompson and Solomon*, 2009], while lower concentrations of stratospheric water vapor after 2001 have been shown to be consistent with enhanced tropical upwelling [*Randel et al.*, 2006]. However, perhaps the most novel evidence is the observed weakening of the amplitude of zonal wind quasi-biennial oscillation in the tropical lower stratosphere from 1959 to 2006 which *Kawatani and Hamilton* [2013] suggest can be most reasonably explained by a trend for enhanced upwelling near the tropical tropopause.

6.2. Mechanisms for the Secular Changes

Studies that have artificially separated the effects of tropospheric warming from those of stratospheric cooling [e.g., *Olsen et al.*, 2007; *Oman et al.*, 2009; *Oberländer et al.*, 2013] have shown fairly convincingly that changes in the Brewer-Dobson circulation are mainly a response to the tropospheric warming, including the concomitant SST changes and not the direct radiative effect of increasing GHG amounts cooling the stratosphere. Also, since the Brewer-Dobson circulation is primarily a wave-driven circulation, its response can be directly attributed by downward control to the wave drag responding to climate change. It further follows from the downward-control theory that only changes in the wave drag in the vicinity of the turnaround latitudes (section 4.3) are relevant [e.g., *Plumb and Eluszkiewicz*, 1999; *Semeniuk and Shepherd*, 2001; *Haynes*, 2005] and changes in the drag elsewhere will simply drive anomalous secondary circulations which, in turn, can only alter the structure of either the upwelling in the ascending branch or the downwelling in the descending branch [*Shepherd and McLandress*, 2011]. Different wave types (e.g., Rossby or gravity waves) respond differently to climate change but by applying downward control (section 4.2) at the turnaround latitudes, the contribution of the response from each wave type to the overall response of the Brewer-Dobson circulation to climate change can be conveniently quantified [e.g., *Butchart et al.*, 2006; *Li et al.*, 2008; *McLandress and Shepherd*, 2009; *Butchart et al.*, 2010a, 2010b; *Okamoto et al.*, 2011; *Shepherd and McLandress*, 2011; *Bunzel and Schmidt*, 2013; *Oberländer et al.*, 2013].

Beginning with the pioneering work of *Rind et al.* [1990], virtually all subsequent studies that have investigated the role of wave drag have concluded that both resolved and parameterized waves (i.e., unresolved buoyancy waves) contribute to driving a stronger Brewer-Dobson circulation in the model climate projections. Again, like the split among wave types driving the circulation (section 5.2; Figure 7), the split among wave types contributing to the climate response is highly model dependent [e.g., *Butchart et al.*, 2006, 2010a, 2010b], though whether there is any connection between these two model dependencies is unclear. In a multimodel comparison, *Butchart et al.* [2010a] found that on average, the percentage contribution of the parameterized waves to the driving of the trends in the upwelling is much larger than that contributing to driving the upwelling itself, at least at 70 hPa (~59% and ~30%, respectively). Moreover, during the Boreal winter, the percentage contribution of the parameterized waves to the climate response can be even more significant [*Li et al.*, 2008; *Butchart et al.*, 2010a; *Bunzel and Schmidt*, 2013].

The underlying mechanisms determining the response of the parameterized wave drag and, in particular, the orographic gravity wave drag to climate change are much better understood than the mechanisms involving the resolved waves. In part, this is because the parameterizations themselves include the inherent simplifying assumption of vertical wave propagation on which current understanding of the role of these waves depends [e.g., *Li et al.*, 2008; *Butchart et al.*, 2010a; *Okamoto et al.*, 2011]. As noted above, trends in the overall strength of the Brewer-Dobson circulation can only be determined by changes in the wave drag in the vicinity of the turnaround latitudes which, in general, are situated in the latitude bands of the subtropical jets (e.g., Figure 4). Most parameterizations of orographic gravity waves typically deposit momentum in the lower stratosphere in the westward vertical wind shear on the upper side of these jets, especially in the Northern Hemisphere (e.g., see Figure 13a and also *Okamoto et al.* [2011, Figure 7]). An eastward

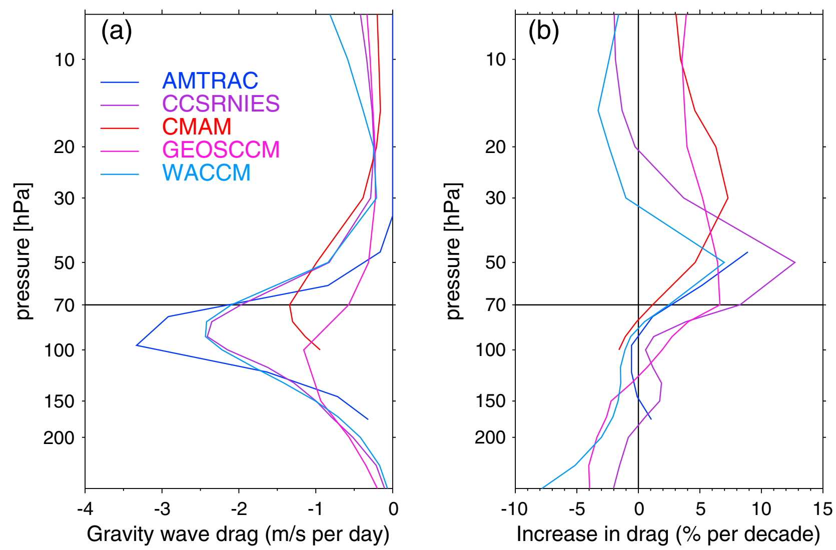

Figure 13. (a) Vertical profiles of the annual mean orographic gravity wave drag [ms^{-1}(day)$^{-1}$] at the northern turnaround latitude for 70 hPa (i.e., the latitude north of the Equator where $\overline{w}^* = 0$ at 70 hPa) for 21st century simulations from five CCMs. For all five CCMs, the location of the turnaround latitude was broadly in agreement with that inferred from reanalyses (e.g., Figure 4) and lies within the latitude band of the subtropical jet [*Butchart et al.*, 2010a]. The annual mean is the average of the four seasonal means. The results shown are for the year 2000 based on a linear fit to the time series of annual mean vertical profiles. (b) Change in drag (% per decade) based on the linear fit to the time series. Figure 12 from *Butchart et al.* [2010a]. ©American Meteorological Society. Used with permission.

acceleration and upward penetration of the jets under climate change then causes the parameterized gravity waves to break higher in the atmosphere resulting in a trend for more drag above 70 hPa (Figure 13b and also *Okamoto et al.* [2011, Figure 7]). In turn, the extra parameterized drag above 70 hPa drives stronger tropical upwelling at that level.

The situation for the resolved waves is more complicated since these waves propagate both horizontally and vertically and their explicit sources are more strongly linked to the tropospheric climate than, say, the sources employed in the parameterizations. Currently, there is no consensus on the mechanism causing the increase in stratospheric wave drag from resolved waves or, more specifically, from the planetary and synoptic scale Rossby waves. Although a number of GCM studies [e.g., *Butchart and Scaife*, 2001; *Garcia and Randel*, 2008] have considered changes in the generation and propagation characteristics of these waves, their findings were not conclusive. On the other hand, results from a simplified GCM were more conclusive with an imposed tropospheric warming increasing the baroclinicity in the subtropical troposphere and thereby generating more synoptic and planetary waves leading to an increase in the flux of wave activity entering the stratosphere in the region of the turnaround latitudes [*Eichelberger and Hartmann*, 2005]. However, *Shepherd and McLandress* [2011] have since pointed out that the acceleration of the Brewer-Dobson circulation in this simple model was much weaker (∼25%) than that found in the more comprehensive models. Also, *Shepherd and McLandress* [2011] argue that the change in meridional temperature gradient imposed by *Eichelberger and Hartmann* [2005] in the subtropical troposphere was more extreme than that generally predicted by climate models.

Other studies, such as *Deckert and Dameris* [2008], have identified a specific mechanism explaining a localized strengthening of the tropical upwelling (i.e., within the upwelling region itself) as a response to tropospheric warming and SST changes. However, since there is no change in the wave drag in the vicinity of the turnaround latitudes, the mechanism suggested by *Deckert and Dameris* [2008] cannot contribute directly to the overall strengthening of the Brewer-Dobson circulation.

A study that does consider the response of the resolved wave drag in the vicinity of the turnaround latitudes is *Shepherd and McLandress* [2011]. Building on an earlier study of *McLandress and Shepherd* [2009], *Shepherd and McLandress* [2011] decomposed the waves in 21st century simulations of a CCM into stationary and transient waves, which were further separated into planetary-scale and synoptic-scale components. As with most climate projections, the eastward subtropical jets in the simulations analyzed by *Shepherd and*

McLandress [2011] responded to climate change by becoming stronger and moving upward. This, in turn, shifted the critical layer for Rossby wave breaking upward and equatorward. A similar movement of the critical layer can also be seen in the multimodel climate response analyzed by *Hardiman et al.* [2013]. *Shepherd and McLandress* [2011] argued rather convincingly that this displacement of the critical layer allows more wave activity to penetrate into the subtropical lower stratosphere and that since the critical layer or wave breaking region lies within the vicinity of the turnaround latitudes [e.g., *Hardiman et al.*, 2013, Figure 3], the extra wave breaking drives a stronger Brewer-Dobson circulation. The mechanism is referred to by *Shepherd and McLandress* [2011] as "critical-layer control of subtropical wave breaking." Interestingly, *Shepherd and McLandress* [2011] found that only transient waves were important in this process as, in their CCM, the changes in the stationary planetary wave drag occur mainly in the middle and high latitudes and not the Subtropics. The proposed mechanism is, nonetheless, rather compelling since if it is confirmed by the analysis of other models, then the response of both the resolved and parameterized wave drags to a changing climate can be attributed to the same mechanism based on an upward shift in wave breaking resulting from the eastward acceleration and upward movement of the subtropical jets.

7. Consequences of a Changing Brewer-Dobson Circulation

7.1. Modulated Ozone Recovery

Since the basic concept of the Brewer-Dobson circulation evolved from the mass circulation deduced, in part, from the observed distribution of stratospheric ozone (section 2) [*Dobson et al.*, 1929; *Dobson*, 1956], it is perhaps not surprising that long-term changes and interannual variability in the Brewer-Dobson circulation have a significant impact on the ozone distribution [e.g., *Fusco and Salby*, 1999; *Randel et al.*, 2002; *Jiang et al.*, 2007; *Salby*, 2008; *Weber et al.*, 2011]. *Rind et al.* [1990] recognized this when they first discovered that climate change accelerates the residual-mean circulation and they suggested that a possible important consequence of their discovery would be increased ozone transport to midlatitudes (see section 6.1). *Mahfouf et al.* [1994] subsequently found that midlatitude column ozone increased in a simulated doubled CO_2 climate (again see section 6.1) though it is unclear if their increased ozone resulted from the circulation response or from the temperature response with the GHG-induced cooling of the stratosphere slowing down photochemical ozone destruction or, indeed, a combination of both processes. Also, the two processes are not completely independent as the circulation response can further affect the temperature response through adiabatic heating or cooling. Nonetheless, other studies [e.g., *Sigmond et al.*, 2004; *Fomichev et al.*, 2007; *McLandress et al.*, 2010] suggest that to a first approximation, the two effects (i.e., the direct radiatively driven diabatic temperature response and the circulation response including the associated adiabatic temperature response) are additive, and therefore, the impacts of the two effects on ozone can be considered separately, at least from a conceptual point of view. Above about 10 hPa, ozone is under photochemical control with a transport and chemically driven region lying below this. Hence, the temperature effect is likely to be most important in the middle to upper stratosphere while any effects of changes in transport are more likely to be seen in the lower stratosphere.

By considering the vertical and latitudinal structure of projected changes in ozone for the 21st century, *Shepherd* [2008] and *Li et al.* [2009] confirmed that the speeding up of the Brewer-Dobson circulation due to climate change had a significant role in ozone recovery in their respective CCMs. In both studies, upper stratospheric ozone increased more or less uniformly with latitude as would be expected from the photochemical response to GHG-induced cooling. Conversely, lower stratospheric ozone decreased in the Tropics but increased in the Extratropics consistent with enhanced poleward transport by a faster Brewer-Dobson circulation.

The influence of GHG-induced circulation and temperature changes on ozone recovery has now been analyzed in a coordinated multimodel ensemble of simulations [*Morgenstern et al.*, 2011] by *Austin et al.* [2010] and *Oman et al.* [2010]. In agreement with *Shepherd* [2008] and *Li et al.* [2009], *Oman et al.* [2010] again found that in the upper stratosphere most of the GHG influence was through the radiative decline in temperatures, while in the lower stratosphere the main impact was from the acceleration of the Brewer-Dobson circulation. *Oman et al.* [2010] convincingly demonstrated this by showing that across the multimodel ensemble of simulations, the changes in tropical lower stratospheric ozone from 1960 to 2100 (i.e., the changes from the period before to the period after the time when stratospheric ozone is significantly affected by chemical depletion resulting from the presence of man-made halogens) were well correlated with the corresponding changes in tropical upwelling (Figure 14). *Oman et al.* [2010] argue further that

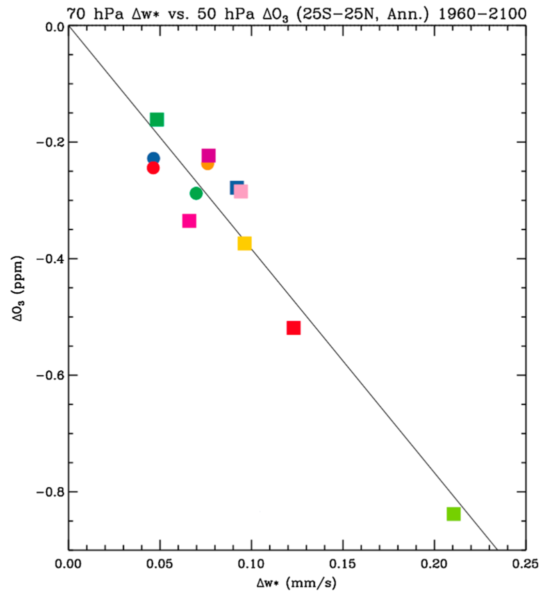

Figure 14. Scatter plot showing the change (from 1960 to 2100) in 70 hPa $\overline{w}^*$ and 50 hPa ozone. The values are annual averages over 25°S–25°N for 12 CCMs (see *Oman et al.* [2010] and *Morgenstern et al.* [2011] for details of the individual models), with the black line showing the linear fit. Adapted from *Oman et al.* [2010, Figure 6].

because the linear fit shown in Figure 14 nearly intersects the origin, the decreases they found in tropical ozone at this level (50 hPa) were almost certainly dominated by the changes in tropical upwelling in the lower stratosphere. The decreases in ozone in the tropical lower stratosphere resulted in the total tropical column ozone not returning to 1960 levels by end of the 21st century in most models. In contrast, in the middle and high latitudes, *Oman et al.* [2010] found that lower stratospheric ozone, and also the total column ozone, increased during the 21st century, mainly as a result of the speeding up of the Brewer-Dobson circulation. Consequently, column ozone at these latitudes was projected by most models to return to 1960 levels well before the end of the century, and at a significantly faster rate than would be expected solely from the declining amounts of ozone-depleting substances. Moreover, *Austin et al.* [2010] found that for the same model simulations, the hemispheric asymmetry in the speeding up of the Brewer-Dobson circulation (see section 6.1) led to column ozone returning to 1980 levels about 10 years sooner in the northern Extratropics than in southern midlatitudes. Interestingly, however, the faster Brewer-Dobson circulation appears not to have an impact on ozone recovery in the Antarctic, at least for the annual average [*Austin et al.*, 2010]. Possibly, this is because the recovery of Antarctic ozone itself also has an impact on the Brewer-Dobson circulation and induces reduced downwelling in polar latitudes and confines the GHG-induced increases in downwelling to the midlatitudes [e.g., *Hardiman et al.*, 2013, Figure 7].

7.2. Faster Removal of CFCs

Ozone-depleting substances or more specifically CFCs, and some GHGs such as nitrous oxide (N_2O), are inert in the troposphere but are destroyed in the stratosphere by, for example, photodegradation. Therefore, their removal rates from the atmosphere and concomitant (instantaneous) atmospheric lifetimes (i.e., the atmospheric burden of the substance divided by the removal rate; e.g., see *Prinn et al.* [1999, section 1.4]) are largely determined by the time taken for all the tropospheric air (i.e., ~90% of the atmosphere) to have passed through the stratosphere at some time. This, in turn, depends on the speed of the stratospheric Brewer-Dobson circulation. Different lifetimes then arise from differences in the spatial and temporal distribution, and strengths, of the stratospheric sinks of the individual species. *Butchart and Scaife* [2001] refer to this as the "efficiency of the stratospheric photodegradation of the species" and this, too, has an additional dependence on the strength *and* structure of the Brewer-Dobson circulation. For example, because photodegradation is generally altitude dependent it requires air to be transported above a certain altitude, depending on the species (i.e., for the overturning circulation to be sufficiently deep), before significant photodegradation occurs [e.g., *Douglass et al.*, 2008, Figure 1].

Rind et al. [1990] were, again, almost certainly the first to link a faster removal of CFCs from the atmosphere to a speeding up of the Brewer-Dobson circulation under climate change. Using a rather simple two-box stratosphere-troposphere model, *Butchart and Scaife* [2001] estimated that with a 3% per decade strengthening of the tropical upwelling, the faster removal would result in CFC amounts returning 5 and 10 years earlier to the levels otherwise expected in 2050 and 2080, respectively (Figure 15). In obtaining this estimate, *Butchart and Scaife* [2001] assumed that the efficiency of the stratospheric photodegradation of the CFCs was not affected by either the changing climate or the recovery of stratospheric ozone.

More sophisticated CCM calculations including both the effects of a faster circulation and the effects of changes in the photodegradation efficiency also predict that in response to climate change, the stratospheric loss rates of the CFCs will increase leading to a corresponding decrease in their lifetimes

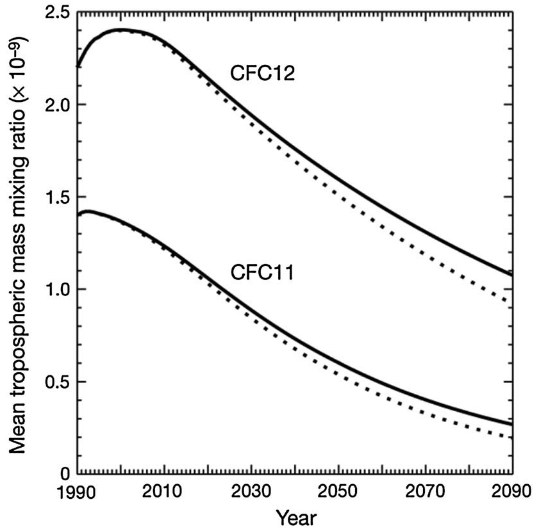

Figure 15. The solid curves show the predicted amounts of CFC11 and CFC12 assuming the current specified lifetimes of 45 and 100 years, respectively. The dotted curve shows amounts after the lifetimes have been adjusted to allow for a 3% per decade increase in the upward mass flux at 68 hPa. Figure 3 from *Butchart and Scaife* [2001].

[*Douglass et al.*, 2008, Figure 8]. Although the fractional release of CFCs (i.e., the fraction of an air parcel's CFC concentration photolyzed since the parcel first entered the stratosphere) decreases throughout the lower and middle stratosphere in these simulations due to shorter resident times this is more than offset by the direct effect of the faster Brewer-Dobson circulation increasing the flux of tropospheric air circulating through the stratosphere. *Douglass et al.* [2008], however, point out a significant model dependency in the compact relationship between the fractional release of CFCs and the speed of the Brewer-Dobson circulation or, more precisely, the mean age of air, and hence, this balance between the effects of the reduced fractional release and a faster circulation could vary considerably between models. Another weakness of these CCM calculations is the concentrations of the CFCs rather than their emissions are specified at the lower boundary and therefore changes in the stratospheric loss rates have no influence on the atmospheric burden of the CFCs in the simulations. This is because with the concentration boundary conditions, the variations in the CFC loss rates due to stratospheric change are effectively compensated for by an artificial surface flux of CFCs constraining the tropospheric loading to the specified boundary conditions [*Douglass et al.*, 2008]. *Douglass et al.* [2008] therefore conclude that the impact of a faster Brewer-Dobson circulation on the decline of CFCs in the atmosphere can only be fully tested in CCMs if emission (flux) boundary conditions are used for the long-lived species. Conversely, the majority of recent CCM ozone projections apply concentration boundary conditions [e.g., *Austin et al.*, 2010; *Bekki et al.*, 2011; *Morgenstern et al.*, 2011] and, consequently, exclude the influence of a faster removal of CFCs, irrespective of how significant this influence is.

7.3. Global-Scale Stratosphere-Troposphere Exchange

A particularly significant development in the understanding of stratosphere-troposphere exchange was the authoritative review of the subject by *Holton et al.* [1995]. This review prompted a paradigm shift away from conceptual models based on synoptic and small-scale processes occurring near the tropopause (see *Gettelman et al.* [2011] for an excellent review of these processes) to the current understanding in which the seasonally and annually averaged mass exchange between the stratosphere and troposphere is believed to be largely controlled by the Brewer-Dobson circulation, at least from a global perspective. Therefore, how the stratosphere-troposphere exchange responds to climate change will depend on how the Brewer-Dobson circulation responds.

One of the more influential manifestations of stratosphere-troposphere exchange is the flux of ozone from the stratosphere into the troposphere. This is an important source of ozone throughout the troposphere where the ozone acts to enhance the greenhouse effect and initiates the chemical removal of methane and other hydrocarbons from the atmosphere. In model simulations, *Collins et al.* [2003] and *Zeng and Pyle* [2003] found that this ozone flux increased under climate change but even though both studies attributed the increase to a speeding up of the Brewer-Dobson circulation, neither model properly resolved the stratospheric circulation and, moreover, used prescribed values of stratospheric ozone. Likewise, although the multimodel study of *Stevenson et al.* [2006] confirmed the increasing ozone flux in a changing climate, the models, again, did not properly represent the circulation or evolution of ozone in the stratosphere. *Hegglin and Shepherd* [2009], on the other hand, obtained quantitative estimates of the stratosphere-to-troposphere ozone flux from a single stratosphere-resolving CCM with self-determined stratospheric ozone, and these, too, confirmed an increasing flux due to the changing climate (see Figure 16). The much larger long-term trend calculated by *Hegglin and Shepherd* [2009] for the Northern Hemisphere than for the Southern

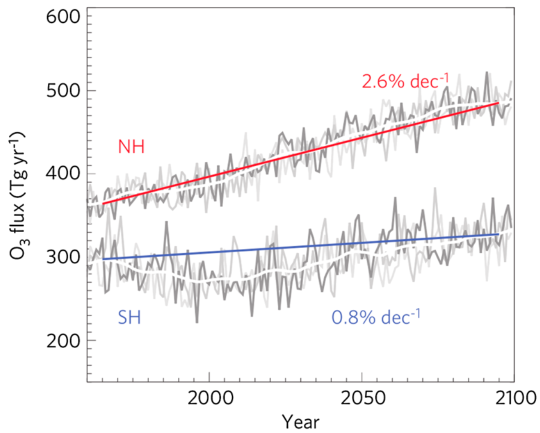

Figure 16. Long-term time evolution of annual mean stratosphere-to-troposphere ozone flux for both the Northern Hemisphere (NH) and the Southern Hemisphere (SH) from an ensemble of three stratosphere-resolving CCM simulations (grey lines). The white line shows the 5 year running ensemble mean. The red and blue lines are a linear fit to decadal averages for 1960–1970 and 2090–2100 for the Northern and Southern Hemispheres, respectively and therefore exclude any modulation of the long-term secular trend due to the effects of anthropogenic ozone depletion followed by recovery during the intervening period. Figure 2a from *Hegglin and Shepherd* [2009].

Hemisphere (red and blue lines in Figure 16, respectively), which excludes the shorter-term (decadal) variations due to ozone depletion and recovery is, again, consistent with the known hemispheric asymmetry in the response of the Brewer-Dobson circulation to climate change (section 6.1). In both hemispheres, the results of *Hegglin and Shepherd* [2009] show that this long-term increase in the ozone flux resulting from the speeding up of the Brewer-Dobson circulation is, however, modulated by the effects of ozone depletion and recovery (i.e., the deviation of the white lines from the red and blue lines in Figure 16; see *Hegglin and Shepherd* [2009] for more details.) In particular, in the Southern Hemisphere the effects of ozone depletion in the period prior to about 2000 more than outweigh the effects of the faster Brewer-Dobson circulation (Figure 16) [*Hegglin and Shepherd*, 2009]. Conversely, in the 21st century the projected ozone-flux increase results from the combined effect of recovering stratospheric ozone and a speeding up of the Brewer-Dobson circulation [*Hegglin and Shepherd*, 2009].

8. Concluding Remarks

The Brewer-Dobson circulation is one of the few truly global-scale phenomena observed in the Earth's atmosphere below ~50 km. It is particularly prominent because of its widespread controlling influence on the stratosphere. For instance, it has important roles in determining the thermodynamic balance of the stratosphere, the lifetimes of CFCs and some greenhouse gases, the temperature of the tropical tropopause, the water vapor entry into the stratosphere, the period of the tropical quasi-biennial oscillation, and the transport and redistribution within the stratosphere of aerosols, volcanic and radioactive debris, and chemical tracers such as ozone. Clearly, it is therefore not possible to consider all aspects of the Brewer-Dobson circulation in a single review. Instead, this review has focused primarily on the methods used to diagnose the Brewer-Dobson circulation from observations and chemistry-climate and climate models, plus current understanding of the driving mechanisms and the response of the circulation to climate change. One aim is to reconcile the apparent robust model projections with the difficulties in obtaining supporting evidence from the currently available measurements and observations.

A combination in the late twentieth century of theoretical developments (e.g., the transformed Eulerian mean formulation) and the emergence of high-quality self-consistent global stratospheric data sets (e.g., state-of-the-art reanalyses and output from stratosphere-resolving GCMs and CCMs) was identified as providing the basis of current knowledge and understanding of the Brewer-Dobson circulation. More importantly, a synergy of these two key developments during the last decade, or so, has allowed the Brewer-Dobson circulation, its driving mechanisms, and its response to climate change to be studied quantitatively for the first time. Prior to this, knowledge about the Brewer-Dobson circulation was mainly conceptual and qualitatively based.

Arguably, this more quantitative approach is the reason for the shift in emphasis in recent studies of the Brewer-Dobson circulation away from the more traditional transport diagnostics inferred from material tracers toward either meteorologically based diagnostics constructed from winds and temperatures or simulated theoretical tracers such as the age of air. Indeed, as discussed in sections 5.1 and 6.1, attempts to derive quantitative information about the mean stratospheric mass circulation from observations of trace species have proved to be problematic. Part of the problem is the paucity of measurements of suitable tracers though the methods used for assimilating the information from the available tracer observations are not nearly as highly developed as, say, meteorological reanalyses (e.g., ERA-Interim). As argued by *Garcia et al.* [2011], by using the tracers observations in conjunction with numerical models, it should, in future, be possible to make better quantitative use of these observations. Also, it should be emphasized that some of the most important effects of anthropogenically driven changes in the Brewer-Dobson circulation, such as the impacts on ozone recovery, the lifetimes of CFCs, and stratosphere-troposphere mass exchange (section 7), are all associated with the transport of chemical tracers.

Almost certainly the most important outcome of research into the Brewer-Dobson circulation over the last decade is the broad consensus on how the circulation will respond to future changes in climate. Depending on the greenhouse gas scenario considered, model projections are converging toward a 2.0–3.2% per decade increase in the net upwelling mass flux in the tropical lower stratosphere which, as discussed in section 4.3, is now used almost universally as a standard measure of the overall strength of the Brewer-Dobson circulation or at least the deep branch of the circulation. This stronger upwelling is entirely consistent with projections for the theoretical age of air tracer which indicate that in a changing climate the stratospheric air will, on average, be younger (i.e., the mean time taken for the air to reach a particular location after entry into the stratosphere will be shorter). Apart from that in the subtropical lower stratosphere, the younger mean age most likely results from both a speeding up of the residual-mean (overturning) circulation and weaker mixing or recirculation of the stratospheric air between the Tropics and midlatitudes (e.g., H. Garny et al., The effects of mixing on age of air, submitted to *Journal of Geophysical Research*, 2014). The responses of both these processes to climate change appears to be well correlated too.

Although the 2.0–3.2% per decade acceleration of the Brewer-Dobson circulation is remarkably robust across models, attempts to quantify the contributions from different waves types (e.g., Rossby and gravity waves) in driving this trend have proved less successful or at least the results are highly model dependent. A similar model dependency is seen in the decomposition of the driving of the circulation itself into different wave types. In both cases, the model dependency does not appear to be obviously related to model formulation and is a major weakness in current understanding of the Brewer-Dobson circulation. Better observational constraints on the circulation (e.g., section 5.1) and also the wave momentum fluxes and, in particular, the momentum fluxes from small-scale buoyancy waves [*Alexander et al.*, 2010] are therefore essential for making further progress in understanding in this area.

Another important gap in the current knowledge about the Brewer-Dobson circulation considered here is a comprehensive understanding of how the wave drag, driving the circulation, responds to climate change. For the small-scale buoyancy waves, the mechanisms are reasonably well understood. Possibly, this is merely a consequence of the simplifications and similarities of the parameterization schemes used in the models though at least for the gravity waves with orographic sources, the parameterizations are derived from basic fluid dynamical principles. On the other hand, there is currently no consensus on a mechanism for the resolved waves and although critical-layer control of Rossby wave breaking offers some promise. [*Shepherd and McLandress*, 2011] further detailed studies will be required to confirm this as an explanation for the response of the transient planetary and synoptic wave drags.

And finally, evidence is now emerging that a changing Brewer-Dobson circulation also has a significant role in the dynamical coupling between the stratosphere and troposphere with implications for surface climate and weather [e.g., *Baldwin et al.*, 2007; *Karpechko and Manzini*, 2012; *Scaife et al.*, 2012]. Therefore, it appears that the influences of the Brewer-Dobson circulation and its response to climate change are not solely confined to the stratosphere but are almost certainly omnipresent throughout the Earth's atmosphere below ~50 km.

Notation

$\overline{()}$ zonal mean.
$()'$ departure from zonal mean (eddies).
a radius of the earth.
$\overline{\mathcal{F}}$ mean zonal forcing.
g gravitational acceleration.
H scale height [$\equiv RT_s/g$].
$\overline{m}$ angular momentum per unit mass [$\equiv a\cos\phi(\overline{u} + a\Omega\cos\phi)$].

p pressure.
p_s a constant reference pressure.
Q diabatic heating.
T temperature.
T_s a constant stratospheric mean temperature.
u eastward velocity.
v northward velocity.
w a measure of "vertical velocity" [$\equiv \mathrm{d}z/\mathrm{d}t$].
z a measure of "height" [$\equiv -H\ln(p/p_s)$].
θ potential temperature.
ϕ latitude.
ψ stream function for the TEM residual-mean circulation.
Ω Earth's angular velocity.

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