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Stratospheric Ozone

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Chapter 6: Stratospheric Dynamics and the Transport of Ozone and Other Trace Gases

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3 -- THE BREWER-DOBSON CIRCULATION

As discussed in Chapter 5, most ozone production occurs in the tropical stratosphere as the overhead sun breaks apart oxygen molecules (O_2) into oxygen atoms (O), which quickly react with other O_2 molecules to form ozone (O_3). The problem with this simplified picture is that most ozone is found outside the tropics in the higher latitudes rather than in the tropics. That is, most of the ozone is found outside of its natural tropical stratospheric source region. This higher latitude ozone results from the slow atmospheric circulation that moves ozone from the tropics where it is produced into the middle and polar latitudes. This slow circulation is known as the Brewer-Dobson circulation, named after Brewer and Dobson.

The simple circulation model suggested by Brewer (1949) and Dobson (1956) consists of three basic parts. The first part is rising tropical motion from the troposphere into the stratosphere. The second part is poleward transport in the stratosphere. The third part is descending motion in both the stratospheric middle and polar latitudes, though there are important differences. The middle latitude descending air is transported back into the troposphere, while the polar latitude descending air is transported into the polar lower stratosphere, where it accumulates.

This model explains why tropical air is lower in ozone than polar air, even though the source region of ozone is in the tropics. However, we are getting a bit ahead of ourselves, and it is necessary to look at the big picture in more detail.

3.1 Stratospheric Circulation: the Big Picture

In order to get a simplified view of this large scale circulation in the stratosphere, it is useful to look at transport processes in a zonally averaged sense, that is, averaged around a latitude circle. **Figure 6.03** shows this zonally averaged circulation in the middle atmosphere superimposed on top of an annual average ozone density (in Dobson Units per kilometer). This circulation is what we call the Brewer-Dobson circulation. It is depicted by the black arrows. The figure also shows the seasonally averaged ozone density (red denotes a high density of ozone; blue denotes low ozone density). The ozone data is based on 1980-1989 Nimbus-7 SBUV data.

Brewer first proposed this slow circulation pattern to explain the lack of water in the stratosphere. He supposed that water vapor is "freeze dried" as it moves vertically through the cold equatorial tropopause (see Figure 6.02). Dehydration can occur here by condensation and precipitation as a result of cooling to temperatures below -80°C. The lowest values of water are found just near the tropical tropopause. Dobson later suggested that this type of circulation could also explain the observed high ozone concentrations in the lower stratosphere polar regions which are situated far from the photochemical source region in the tropical middle stratosphere. The Brewer-Dobson circulation additionally explains the observed latitudinal (i.e. north-south) distributions of long lived constituents like nitrous oxide and methane.

3.2 The Brewer-Dobson Circulation in the Tropics

The air that is slowly lifted out of the tropical troposphere into the stratosphere (see Figure 6.02 and 6.03) is very dry, with low ozone, and high CFC levels (see Figure 1.07 of Chapter 1). This tropical lifting circulation out of the lower stratosphere is quite slow, on the order of 20-30 meters per day. Most of the air rising into the stratosphere at the tropopause never makes it into the upper stratosphere. Between 16 and 32 km, the air density decreases by about 90%. This means that of the mass coming into the stratosphere at 16 km, approximately 90% of that mass will move towards the middle latitudes rather than be carried up to 32 km.

3.2.1 Ozone Source Region -- Air in the troposphere has relatively low ozone concentrations, except in highly polluted urban environments. Even polluted regions are relatively low when compared to stratospheric levels. As this "ozone clean" air moves slowly upward in the tropical stratosphere, ozone is being created by the slow photochemical production caused by the interaction of solar UV radiation and molecular oxygen.

Ozone is created in this region because it is here that the Sun, positioned high overhead during the day all year long, is most intense. There is enough of the necessary sufficiently energetic UV light to split apart molecular oxygen, O2, and form ozone (see Chapters 1 and 5 for discussion of ozone production). It typically takes more than 6 months for air at 16 km (near the tropical tropopause) to rise up to about 27 km.

Even though ozone production is small and slow in the lower tropical stratosphere, the slow lifting circulation allows enough time for ozone to build-up. Figure 6.03 shows this ozone density maximum up near 27 km. It is this that is commonly referred to as the "ozone layer".

3.2.2 CFC Transport -- Another result of this mass circulation is that most CFCs are carried from the troposphere into the stratosphere in the tropics, and are then recycled back into the troposphere in the middle-to high latitudes. Since it is the intense UV in the upper stratosphere that breaks down CFCs, and since very few CFC molecules makes it to the upper stratosphere, the lifetime of CFCs are quite long. It is estimated that the time scale needed to reduce CFC-12 by 63% is approximately 120 years. This lifetime results from the very slow circulation and the decrease of density, which both significantly impact the rate at which CFCs reach the upper stratosphere and are broken down by UV light.

3.3 The Brewer-Dobson Circulation in the Extratropical Latitudes

In the stratosphere, the Brewer-Dobson circulation carries air from the equator to the poles. Poleward of about 30°N and 30°S, the circulation becomes downward as well as poleward. This poleward and downward circulation tends to increase ozone concentrations in the lower stratosphere of the middle and high (i.e. extratropical) latitudes. In Figure 6.03, we see this increase of ozone at lower altitudes in the higher latitudes as a direct result of this circulation.

Another reason that ozone amounts increase in the lower stratosphere in the extratropical latitudes is that the lifetime of an ozone molecule gets longer here. As explained in Chapters 1 and 5, ozone is produced by molecular oxygen photolysis (producing two free oxygen atoms), and it is destroyed in catalytic reactions (generally utilizing free O atoms). Since there are very few O atoms in the lower stratosphere (because most of the UV necessary to produce them is absorbed at higher altitudes), the lifetime of ozone is very long. Thus, ozone is not easily destroyed in the lower stratosphere. As a result, ozone can accumulate as the Brewer-Dobson circulation moves air poleward from the tropical production region into higher latitudes and downward into lower altitudes.

3.4 Theory of the Brewer-Dobson Circulation: Why Does It Exist?

The mechanism behind the Brewer-Dobson circulation is both complex and quite interesting. At first glance, we might expect that the circulation results from solar heating in the tropics, and cooling in the polar region, causing a large equator to pole (meridional) overturning of air as warm (tropical) air rises and cold (polar) air sinks. While this heating and cooling does indeed occur, and while such a meridional overturning exists in the form of the so-called Hadley circulation (see section 3.8.1), it is not the specific reason for the existence of the Brewer-Dobson circulation. Rather, the Brewer-Dobson circulation results from wave motions in the extratropical stratosphere. In this section, we assume that the reader has some familiarity with the concept and types of atmospheric waves. For a more thorough discussion of waves, see section 5.3 of Chapter 2.

3.4.1 Standing Planetary Waves and Wave Breaking -- One type of atmospheric wave that exists is called the Rossby wave. Named for Carl G. Rossby, an early atmospheric research scientist, the Rossby wave exists due to a combination of meridional temperature gradients and the rotation of the planet (which produces the Coriolis force). The Rossby wave is a large-scale wave system whose size is thousands of kilometers in the horizontal and several kilometers in the vertical.

Large-scale topographical features, like the Rocky Mountains and the Himalaya-Tibet complex, together with the meridional temperature gradients and Coriolis deflection, create a variation of Rossby waves called standing planetary waves. These have very long wavelengths (up to 10,000 kilometers) and either remain stationary or move slowly westward (i.e., they move easterly). They eventually propagate vertically into the stratosphere.

3.4.2 Polar Night Jet Deceleration and Radiative Imbalance -- When a standing planetary wave reaches the stratosphere, it deposits its easterly momentum, decelerating the westerly wintertime stratospheric jet stream. This is the polar night jet we discussed in section 2.4.2-c and depicted in Figure 6.02. The polar night jet slows and can even be displaced, which has the effect of displacing the polar vortex region.

The deposition of easterly momentum into the polar stratosphere and the deceleration of the polar night jet is known as "wave breaking" (see section 4.1.2). It produces the phenomenon of the **stratospheric sudden warming** (see Chapter 2, section 4.2.2) as warmer middle latitude and even tropical air intrudes into the geographic polar region. This result is a situation that is thermodynamically imbalanced. Wintertime radiational cooling in the polar stratosphere quickly begins.

3.4.3 Sinking Air and Meridional Overturning -- This cooling of air is accompanied by sinking motions, since colder air is more dense and it sinks. It is this sinking motion that establishes the meridional overturning from equator to pole in the winter hemisphere. That is, the sinking air in the polar region must be balanced by a poleward flow of air into this region. By mass continuity requirements, this air must come from the tropics. Our Brewer-Dobson circulation cell is thus established as tropical air moving poleward to replace the sinking air at the poles is itself replaced by rising air in the tropics (see Figure 6.03).

A simple conceptual model for our Brewer-Dobson circulation is to use a paddle near the edge of a small circular pool. If you start paddling in one direction on the edge of the pool, in a short time, you'll set up a circulation that carries water fully around the pool (try this in your bathtub). You don't need to paddle everywhere in the pool, just at a single point and always in the same direction. The paddle in the stratosphere is the "wave activity" in the extratropical middle and upper stratosphere. This "wave activity" paddle causes the air to move poleward in the stratosphere, which causes the rising in the tropics, and the sinking in the polar region.

So while the Brewer-Dobson circulation cell is created due to mass continuity requirements, ultimately, its existence is due to the breaking of planetary waves into the winter hemisphere polar stratosphere. Note that the Brewer-Dobson cell is a winter time circulation. It almost nonexistent in the summer hemisphere. There the net mass flux is small and slightly downward.

3.4.4 Brewer-Dobson Circulation and Radiative Balance -- In the absence of any stratospheric waves and the consequent Brewer-Dobson circulation, the polar region in the middle of winter would be much colder than it actually is. Calculations show that without the waves and resulting Brewer-Dobson circulation, the polar stratosphere would be phenomenally cold. It is estimated that 30 km polar temperature would be about 160 K (-113°C or about -171°F), as opposed to the measured 200 K (-73°C or -99°F).

The temperature field derived without any waves or circulation is known as the "radiative equilibrium". In the winter lower stratosphere, the circulation in the polar region is downward. This adiabatic descent (compression) results in temperatures that are warmer than the corresponding radiative equilibrium value, producing a slightly warmer radiative balance.

3.5 Hemispheric Differences In Winter Transport

We have seen that the Brewer-Dobson circulation features rising motions around the equator, poleward transport, and sinking motions in the higher latitudes. This is the general pattern. However, there are significant hemispheric differences in the strength and behavior of the Brewer-Dobson cell. We see this by examining the large-scale features of the zonal mean distributions of ozone, trace gases, temperature, and zonal wind in the stratosphere.

3.5.1 Hemispheric Differences in Planetary Wave Activity -- The midwinter Brewer-Dobson circulation cells in the southern and northern hemispheres are quite different. The differences arise from the hemispheric differences in planetary wave forcing coming out of the troposphere (our "wave activity" paddle). Large-scale topographical features like the Rockies and the Himalaya-Tibet complex are mostly in the northern hemisphere. The southern hemisphere has significantly less land than the northern hemisphere and is almost entirely ocean from 55°S south to the Antarctic continent. The weak winter wave activity in the southern hemisphere means that the Antarctic polar vortex is much more isolated than its Arctic counterpart, and as a result, temperatures in the Antarctic polar vortex get extremely cold. These wave forcing differences have a profound influence on transport and result in the observed hemispheric differences in distribution of ozone and other stratospheric trace constituents.

Because of the prominent topography and land-ocean contrasts in the northern hemisphere, the northern stratosphere has more frequent and intense planetary wave activity during the northern winter than the southern hemisphere stratosphere during the southern winter. This stronger wave activity in the northern hemisphere leads to a stronger Brewer-Dobson circulation in the northern midwinter than during the southern midwinter. As will be discussed in a later section, these horizontal mixing processes in the southern hemisphere are confined to the subtropics and middle latitudes, seldom reaching the polar (Antarctic) region. By contrast, in the northern hemisphere, mixing processes often extend into the polar (Arctic) region, owing to the significant planetary wave activity and resultant Brewer-Dobson circulation.

In the section below, we consider the case of methane distribution and the seasonal hemispheric differences that result because of these differences in planetary wave activity.

3.5.2 Seasonal Hemispheric Differences in Methane Distribution -- This overall pattern of upward mass transport in the tropics and downward transport in the extratropics due to the Brewer-Dobson circulation cell is reflected in long-lived tracers such as methane (CH_4). **Figures 6.04a, b, c, and d** show methane distribution for January, March, July, and October respectively. Methane has its source region in the troposphere, and is lost in the stratosphere by a reaction with OH molecules and oxygen atoms (thus, it has a distribution opposite to ozone, see Chapter 5). Like with ozone density, the contours of methane density are displaced upward in the tropics and downward at higher latitudes, reflecting the influence of the Brewer-Dobson circulation as shown by the arrows. The ascending branch of the Brewer-Dobson circulation carries high-methane air from the tropical lower troposphere into the tropical stratosphere, while the descending branch of the Brewer-Dobson circulation carries low-methane air from the polar upper stratosphere into the polar lower troposphere.

Figures 6.04a and 6.04c show latitude-height distributions of CH_4 during January and July. The downward circulation during January (northern winter) between 60°N and the North Pole near 24 km (30 mb) is quite strong. In contrast, during July (southern winter) between 60°S and the South Pole near 24 km (30 mb), the circulation is noticeably weaker. Methane has a higher concentration in the Arctic in winter than in the Antarctic in winter.

The downward circulation is greater in the winter northern hemisphere than in the winter southern hemisphere, and so the amounts of CH_4 are greater in the Arctic than the Antarctic. This is because stronger wave activity in the northern hemisphere results in more meridional mixing of air from lower latitudes into the Arctic than into the Antarctic. Air descending into the Arctic lower stratosphere has undergone mixing with higher methane air from the northern middle latitudes stratosphere. By contrast, weaker wave activity in the southern hemisphere results in less meridional mixing. Air descending into the Antarctic lower stratosphere is mostly undiluted lower methane mesospheric air carried down by the descending branch of the Brewer-Dobson circulation. The result is lower wintertime CH_4 concentration in the Antarctic than in the Arctic.

3.6 Hemispheric Differences In Spring Transport

The spring seasons also exhibit hemispheric differences in temperature, wind, and trace constituent distributions. **Figures 6.05a and 6.05b** show the temperature and zonal wind fields for March and September, respectively.

The northern hemisphere spring (March) is characterized by rather flat meridional temperature gradients and weak winds in both hemispheres. The corresponding March CH_4 distribution and air circulation (Figure 6.04b) have upward motion in the tropics, and relatively strong downward motion throughout the stratosphere in the extratropical latitudes of both hemispheres. The southern hemisphere spring CH_4 distribution and circulation (Figure 6.04d) shows a qualitatively similar picture, but note that the southern hemisphere spring (September) polar temperatures shown in Figure 6.05b are colder and the zonal winds are stronger than the corresponding northern hemisphere spring (March) case shown in Figure 6.05a.

This difference is a direct result of the weaker planetary wave forcing during the southern winter and early spring compared to that in the northern hemisphere. The breakup of the southern polar vortex is delayed until late southern spring (November). The sharp meridional gradients in the long-lived tracer field across the southern polar vortex boundary (approximately 60°S) and the subsequent low tracer values within the wintertime polar vortex region (Figure 6.04c) also persist through the southern spring (Figure 6.04d).

3.7 The Quasi-Biennial Oscillation (QBO) and the Brewer-Dobson Circulation

Atmospheric weather and wave dynamics vary from year to year. These differences produce an interannual variability in the wave activity that affects the Brewer-Dobson circulation. One of the principal sources of year-to-year variability in the total ozone distribution is the **quasi-biennial oscillation** (QBO). (The QBO is described in Chapter 2, section 4.2.3, as well as in Chapter 8, section 5.3.1.)

Total ozone distribution is affected by the QBO for two reasons: (1) the QBO affects the stratospheric temperature structure, which in turn affects the photochemical balance of the upper stratosphere (see Chapter 5); and (2) the QBO directly modifies the Brewer-Dobson circulation.

3.7.1 What is the QBO? -- The QBO is an oscillation of the east-west wind in the tropical stratosphere. The QBO effect occurs throughout the tropics, but it is most often shown as a change in the direction of the stratospheric zonal wind at Singapore. **Figure 6.06**(also shown as Figure 2.16 in Chapter 2) shows these zonal winds at Singapore (1°N, 104°E) from 1978-1998 between 18-30 km altitude.

As we see in Figure 6.06, the wind direction over the tropical stratosphere changes sign (direction) about every year. However, because the QBO is due to the internal dynamics of tropical waves rather than the annual change of seasons cycle, the period of this wind oscillation is highly variable with periods ranging from 22 to 34 months. Hence the name, quasi-biennial oscillation, reflects the variable period of this phenomenon. The theoretical explanation of the QBO is given below in section 3.7.2-c. Note how winds blowing in one direction descend in altitude in time and are replaced by winds blowing in the opposite direction. We refer to winds blowing in a particular direction as a regime. There are westerly and easterly wind regimes associated with the QBO.

Through satellite measurements, we can get a more detailed picture of the QBO phenomenon throughout the tropics. The QBO phenomenon shows up in the overall equatorial zonal wind field. These changes in wind direction produce temperature anomalies, which in turn modify the Brewer-Dobson circulation. These three QBO-related features--zonal wind, temperature, and Brewer-Dobson circulation--are discussed in the next three subsections.

(a) Equatorial zonal wind QBO -- Satellite measurements of stratospheric winds from the High Resolution Doppler Imager (HRDI) instrument on board the Upper Atmosphere Research Satellite (UARS) have been recently compiled. Although the general characteristics of the equatorial zonal wind QBO are similar to previous Singapore balloon based radiosonde observations, the satellite measurements extent to higher altitudes (approximately 40 km) and sample a wide range of latitudes and longitudes as compared to the more sparsely spaced radiosonde network. This is especially true of the tropical latitudes which have very few radiosonde stations. **Figure 6.07** is a plot of equatorial belt zonal winds for 1992-1996 measured by the HRDI instrument aboard the UARS.

Both Figures 6.06 and 6.07 reveal the following unique characteristics of the tropical QBO. The QBO is generally observed between 20-35 km. The easterly winds are generally stronger than the westerly winds, persist longer at upper levels (approximately 30 km altitude), and have maximum wind speeds centered over the equator near 26 km. Westerly wind regimes descend faster in time and persist longer at lower levels than easterly wind regimes. Below 15 km, there is little evidence of the QBO, while above 35 km, the QBO coexists with another regular oscillation of the mesosphere known as the semiannual oscillation.

The QBO extends in latitude between about 15°N and 15°S. **Figure 6.08** shows this latitudinal extension in a latitude versus time cross-section of zonal wind at 25 km. From Figure 6.08 we see that there is a distinct difference in the onset of the easterly and westerly phases. Westerly accelerations (the transition from purple to orange) first appear at the equator, and in time slowly spread to higher latitudes, while the easterly accelerations are more uniform in latitude. Near the equator, the oscillation is fairly symmetric, while in the subtropics, the oscillation combines with the annually (seasonally) varying westerlies of the winter hemisphere.

(b) Temperature QBO -- Associated with the zonal wind QBO is a temperature QBO. The physical relationship between the zonal wind and the temperature is called the thermal wind relationship.

As explained in Chapter 2, thermal wind balance is a relationship between horizontal temperature gradients and vertical wind shear (i.e. the way that the wind is changing with height). The greater this horizontal temperature gradient, the greater the vertical shear of the geostrophic wind. A positive temperature gradient from pole to equator (cold to warm) produces increasing westerly winds with height, while a negative temperature gradient from pole to equator (warm to cold) produces decreasing westerly or increasing easterly winds with height. This physical relationship shows that the transition zone between westerly and easterly winds will be a warm temperature region. This is shown schematically in **Figure 6.09.**

The top panel of Figure 6.09 shows a descending westerly QBO phase in which a warm temperature area is associated with positive wind shear (increasing westerly winds with height), as we expect through the thermal wind relationship. The bottom panel of Figure 6.09 shows a descending easterly QBO phase. Here a cold temperature area is associated with negative wind shear (increasing easterly winds with height), again as we expect via thermal wind balance.

These temperatures can modulate ozone in two ways. First, in the tropical upper stratosphere, temperatures modulate photochemical reaction rates, such that warm temperatures are associated with lower ozone and colder temperatures are associated with higher ozone levels. Second, temperatures can directly impact the circulation by modifying the heating and cooling rates.

(c) QBO circulation -- As mentioned in the previous section, the temperature anomalies associated with the QBO winds induce a modification to the normal Brewer-Dobson circulation. This QBO circulation is superimposed on our normal Brewer-Dobson circulation. Depending on which phase, this circulation will either be speeded up or weakened.

The QBO descending easterly phase maintains colder temperatures between the overlying easterlies and underlying westerlies (see Figure 6.09). The result is that the infrared cooling to space will be smaller than normal in the QBO cold region. Because the heating from solar UV is approximately constant, the weakened cooling to space means that the total heating in the tropics is somewhat larger. This greater heating in the tropics results in a speeding up of the normal Brewer-Dobson lifting in the tropics.

Conversely, the QBO descending westerly phase maintains warmer temperatures between the overlying westerlies and underlying easterlies (again, see Figure 6.09). The result is that infrared cooling to space will be greater than normal in the QBO warm region. Again, because the heating from solar UV is approximately constant, the greater cooling to space means that the total heating in the tropics is somewhat smaller. This lesser heating in the tropics results in a slowing up of the normal Brewer-Dobson lifting in the tropics.

These downward and upward motions associated with the QBO at the equator are balanced by upward and downward motion in the subtropics, respectively. This circulation cell (pictured in Figure 6.09), which is connected by poleward or equatorward motions, is called the QBO-induced meridional circulation.

The subtropical branch of the QBO-induced circulation cell is located approximately between 15°N and 15°S. As one might guess, the QBO-induced circulation has an influence on trace gas constituents in the tropical stratosphere. QBO signals in ozone, methane, hydrogen fluorine and nitrous oxide have been reported from long term satellite observations.

3.7.2 Ozone Transport: Influence of the QBO -- *(a) Tropical ozone QBO* -- The QBO represents an important source of ozone variability throughout the lower stratosphere (see Chapters 8 and 9). The equatorial zonal wind QBO is confined to the equatorial stratosphere, a region where ozone is controlled by both transport and photochemistry. Below 30 km in the tropics, ozone is primarily under dynamical control, and thus is affected by the QBO-induced circulation that exists atop the Brewer-Dobson circulation. Above 30 km, ozone increasingly becomes under photochemical control, and thus responds to the QBO-induced temperature anomalies rather than transport effects.

As discussed above, the descending westerlies of the QBO are associated with a vertical circulation pattern that produces downward motion in the tropics and upward motion in the subtropics, weakening the normal Brewer-Dobson circulation in the tropics. Because the upward motion of air is slowed down and because the vertical gradient of ozone mixing ratio is positive in the lower stratosphere (i.e. increasing ozone with altitude), ozone production can proceed for longer periods. The result is a positive column ozone anomaly in the tropics and a negative anomaly in the subtropics. In the descending easterly phase of the QBO when the Brewer-Dobson circulation in the tropics is enhanced, ozone production has less time to occur, and the column ozone anomalies are reversed, resulting in a negative ozone anomaly in the tropics and a positive ozone anomaly in the subtropics.

Figure 6.10 shows these ozone anomalies versus altitude for the 4°S to 4°N region. The data is based on the UARS HALOE instrument, with reds denoting high ozone and the blues denoting low ozone. In this figure, the time average has been subtracted out. Superimposed on the ozone anomalies in the black contours are the HRDI zonal winds.

We see in Figure 6.10 alternating high and low ozone values in the 20-30 km altitude range, corresponding to the westerly and easterly shear zones of the QBO, respectively. Above 30 km, the ozone variability is controlled more by temperature dependent photochemical processes than transport. Thus, the ozone variation from 35-45 km is a response to the temperature variations due to both the QBO and the semiannual oscillation.

(b) Extratropical ozone QBO -- Observations of QBO signals in dynamical variables and constituents fields, such as column ozone and water vapor, exist in the extratropical stratosphere (see Chapters 8 and 9). However, unlike the situation in the tropical stratosphere where the mechanism for the observed ozone anomalies is fairly well understood, there is no well accepted explanation for how the equatorial QBO anomaly is transmitted to extratropical latitudes. In fact, there is even debate as to how large an extratropical QBO signal exists. Estimates of the magnitude of the midlatitude total ozone QBO range from 5-20 DU. Although this is still a area of active research, it is clear that the QBO has an important influence on the extratropical circulation and various constituent fields.

(c) Theoretical Explanation of the QBO -- Like the Brewer-Dobson circulation, the QBO depends on atmospheric waves. The three principal types of waves are gravity waves, mixed Rossby-gravity waves, and Kelvin waves. These types of waves are discussed in Chapter 2, section 5.3. The reader is referred to there for a further explanation of these waves.

The theory of the QBO was developed Prof. Richard Lindzen of the Massachusetts Institute of Technology and Prof. James Holton of the University of Washington during the early 1970s. They proposed that dissipation of vertically propagating equatorial waves is the source of momentum responsible for causing the wind QBO. They used a simple model to show that the dissipation of vertically propagating Kelvin and Rossby-gravity waves can produce a QBO-like circulation.

Although more recent two- and three-dimensional simulations of the QBO have supported their theory, the exact sources of the momentum remains unclear. For example, observations of Kelvin and Rossby-gravity waves suggest that these waves alone do not possess the momentum necessary to drive the observed circulation. Other explanations for alternate easterly momentum sources include equatorially propagating planetary waves, Rossby waves and westward propagating gravity waves. Gravity waves also have been suggested as a possible source of both easterly and westerly momentum, although this has not yet been verified. While it is clear that waves are responsible for the QBO, their ultimate source and characteristics remains elusive.

3.8 Circulation Patterns Outside the Stratosphere

We now turn to other meridional atmospheric circulation patterns that also play a role in distributing ozone. By meridional, of course, we mean that the circulations are between different latitudes (i.e. north-south). There are meridional circulations both below and above the stratosphere in the troposphere and mesosphere, respectively. We will look at these circulations briefly in this section.

3.8.1 Tropospheric Meridional Circulation -- While the Brewer-Dobson circulation is the most important circulation for understanding stratospheric ozone, other circulations impact the stratosphere in minor ways.

The Hadley circulation is a tropospheric circulation consisting of rising motion in the tropical troposphere and subsidence over the extratropical regions. Warm, moist air rises in the tropical troposphere along the Inter-Tropical Convergence Zone (ITCZ), where convectively driven cumulonimbus (thunderstorm) clouds tower into the atmosphere. These convective towers pump material from the surface into the tropical upper troposphere, where they can slowly be carried into the stratosphere. Air sinks back down in the cooler subtropical regions, resulting in subsidence and belts of semipermanent high pressure (and hence arid and semiarid climates). This circulation pattern is known as a Hadley cell.

The tropical branch of the Hadley circulation is not a continuous mechanism that occurs at all longitudes but rather a series of concentrated bands of "hot towers" of convection. The ITCZ seen on global visible cloud images on or around the equator as a thin line of thunderstorms within 10° of the equator. In addition, the Hadley circulation is responsible for the lower tropospheric easterlies and the upper tropospheric westerlies.

While we understand the processes that cause material to slowly move into the stratosphere, many of the details of this slow transport are currently areas of active research.

3.8.2 Mesospheric Mean Meridional Circulation -- In the upper stratosphere and mesosphere (30-90 km) during the solstices, the circulation is dominated by a single circulation cell with rising motion in the summer hemisphere and sinking motion in the winter hemisphere with a corresponding summer-to-winter hemisphere meridional drift required by mass continuity (see Figure 6.04a near 50 km). As a result, the summer polar mesosphere is much colder than its radiative equilibrium, while the winter polar mesosphere is much warmer. This circulation pattern is generally driven by small scale gravity waves. These waves occur during all seasons and at all latitudes in the mesosphere. **Figure 6.11** shows this circulation in the stratosphere and mesosphere, superimposed on the water vapor distribution for January (northern winter/southern summer).

Water vapor has a photochemical source in the upper stratosphere (approximately 50 km), and a sink in the upper mesosphere above about 80 km. Figure 6.11 shows the mean mesospheric circulation as it transports high water vapor values from the upper stratosphere to the upper mesosphere near 80 km in the middle to high latitudes of the summer (southern) hemisphere. Low water vapor values are subsequently transported from the lower thermosphere above 90 km down to the lower mesosphere/upper stratosphere at middle to high latitudes of the winter (northern) hemisphere.

3.9 Age of Air in the Stratosphere

Stratospheric ozone is destroyed by catalytic reactions with trace gases such as chlorine (see Chapter 5). The chlorine in the stratosphere is primarily released from human produced CFCs via UV photolysis (see Chapter 11). However, a CFC such as CFC-12 is photolyzed by wavelengths less than 240 nm. Since ozone and oxygen molecules absorb the overwhelming majority of this radiation, CFCs must reach very high altitudes before they can be UV photolyzed by radiation of this wavelength and release their chlorine into radical forms that are then free to attack ozone. The slow Brewer-Dobson circulation cell takes at least 1-2 years to move this air from the troposphere to the upper stratosphere, where this material remains for at least 1-2 years. The longer the residence time of CFCs in the stratosphere, the more chlorine can be liberated from the CFC into the radical species, and hence the more ozone that a radical chlorine atom can destroy.

Transport models can be used to estimate the average "age of air" for the stratosphere. This age is the average amount of time it takes for air parcels to be transported from the ground to a specific latitude and altitude region in the stratosphere. The modeled age is shown in **Figure 6.12.**

We note in Figure 6.12 that the air parcels in the troposphere are very quickly mixed due to convection and weather systems, so that the age of air distribution is relatively uniform in this region. This age diagram is based on an idealized tracer with a tropospheric mixing ratio that increases linearly in time. The age distribution reflects the influence of the Brewer-Dobson circulation. Air enters the stratosphere through the tropical tropopause, and ascends through the tropical stratosphere (see also Figures 6.04a-d). As noted in Section 3.2, nearly 90% of the air entering the stratosphere at 16 km never makes it to the top of the stratosphere around 32 km. Most of the air is from the tropics finds its way into the extratropical lower and middle stratosphere, where it enters the descending branch of the Brewer-Dobson circulation. Air that does make it to the top of the stratosphere, undergoes vigorous overturning in the mesosphere by the circulation that exists there (see section 3.8.2). This circulation is induced by gravity wave breaking (see Section 4.0 below).

It takes air parcels 4.5 to 5 years to be transported from the ground to the lower mesosphere. Air parcels return to the stratosphere via wintertime descent at middle and high latitudes (Figures 6.04a). The tropical lower stratospheric air is relatively "young", since it has been directly lifted out of the troposphere by the Brewer-Dobson circulation. This young tropical stratospheric air contrasts with lower stratospheric air at middle and high latitudes, where ages can exceed 5 years. Older air has a large abundance of chlorine radicals and small abundance of CFCs because it has had a longer period for the CFCs to be destroyed by UV photolysis. In comparison, tropical lower stratospheric air has low concentrations of chlorine radicals and higher CFC amounts. While it takes about 4-5 years for material to reach the upper stratosphere from the troposphere, only a small fraction of all tropospheric air ever makes it that high. The time needed to cycle most of the tropospheric air through the upper stratosphere is in fact many decades.