CHAPTER 8

# Impact of Climate Change on the Stratospheric Ozone Layer

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## 8.1 Introduction

The stratospheric ozone layer shows distinct variability on time-scales ranging from seasonal to decadal. Changes to the ozone layer are driven by both natural processes and human activities. In particular, long-term changes and trends need to be fully explained before robust assessments of the future evolution of the ozone layer can be made. It is necessary to answer questions like:

- What has happened in the past?
- Why did it happen?
- What is the likelihood of it happening again in the future?

Another key question has to be answered in this context:

- Are the observed fluctuations and changes driven by natural processes or are they dominated by human activities?

Climate change will affect the future evolution of the ozone layer through changes in temperature, chemical composition, and transport of trace gases and aerosols (*e.g.* Chapter 5 in WMO, 2007;[1] Chapter 4 in WMO, 2011).[2]

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Understanding of cause and effect relationships is often complex because many feedback processes are nonlinear.

The ozone layer is generally assumed to have been nearly unaffected by man-made ozone depleting substances (ODSs; *e.g.* chlorofluorocarbons: CFCs) prior to 1960. By the time concentrations of ODSs have returned to pre-1960 values, which could be the middle of this century (see Chapter 2), concentrations of greenhouse gases (GHGs) will have increased substantially (Chapter 9). An increase of long-lived GHG concentrations (carbon dioxide: $CO_2$, methane: $CH_4$, and nitrous oxide: $N_2O$) in the atmosphere leads to higher tropospheric temperatures (the greenhouse effect). On average, GHGs cool the stratosphere (Subsection 8.2.1–2), which modifies chemical reaction rates (Subsection 8.2.3) for ozone destruction and alters stratospheric circulation (Subsection 8.3.2), both influencing the state of the ozone layer. For those reasons a return to the same ozone layer that existed before 1960 will not occur. The rates of many chemical reactions are temperature dependent, and these reaction rates affect the chemical composition of the atmosphere. There are two types of ozone-destroying chemical reactions that are temperature dependent. Lower stratospheric temperatures lead to a slowing of most gas-phase reactions that destroy ozone (Chapters 1, 3, and 6), but on the other hand yield an intensified depletion of ozone in the lower polar stratosphere due to increased heterogeneous activation of halogens on the surfaces of particles in polar stratospheric clouds (PSCs; see Chapter 4). It is expected that as the stratosphere cools, the slowing of gas-phase reactions will dominate, resulting in a slightly thicker ozone layer (Chapter 9).

Since climate change also influences the dynamics of the troposphere and the stratosphere, and therefore affects the transport of trace gases and particles, dynamically induced temperature changes could locally enhance or weaken the temperature changes caused by radiative processes. So far, estimates of future changes of stratospheric dynamics are uncertain with some numerical models of the atmosphere projecting that stratospheric temperatures will increase in polar regions during the winter and spring seasons, while most models predict a further cooling (Subsection 8.3.2). Hence the net effect of stratospheric radiative, chemical, and dynamical processes and their interactions are poorly understood and quantified.

Due to the expected decrease of ODSs (Chapter 2), the future ozone layer will strongly depend on how increasing concentrations of GHGs affect the temperatures and circulation of the stratosphere. Such a prediction remains a challenge, in large part because the stratospheric circulation depends on how the tropospheric circulation and sea-surface temperatures (SSTs) evolve (Subsection 8.3.3), as well as the details of vertical coupling of the stratosphere and troposphere including the mass exchange and mixing between these atmospheric layers (Section 8.4).

Some aspects of the future evolution of the ozone layer can be projected with numerical models of Earth's atmosphere with reasonable confidence, in particular, with Climate-Chemistry Models (CCMs; see Chapter 9). Climate change is expected, on average, to accelerate the recovery of the ozone layer. A further cooling of the stratosphere will likely result in a slightly thicker ozone layer in the second half of this century ("super-recovery" of the ozone layer). However, such a recovery will not be uniform. Over the polar regions, there is much more uncertainty. Reduced winter temperatures in the lower stratosphere over the poles would be expected to create more PSCs (Chapters 4 and 5), which are needed for the rapid heterogeneous chemical reactions mainly responsible for the observed polar ozone loss in spring, especially in the Antarctic region, creating the most dramatic depletion of ozone, *i.e.* the ozone hole. By itself, this suggests the possibility of increased ozone loss in the northern polar lower stratosphere. However, most numerical models predict that circulation changes due to increasing GHGs will act in the opposite sense during late winter and spring. The net result is not yet clear.

This chapter will give a comprehensive overview about individual physical, dynamical and chemical processes and feedback mechanisms associated with climate change and their implication for ozone depletion, in particular the connection with the expected recovery of the ozone layer, which is discussed in Chapter 9.

## 8.2 Impact of Enhanced Greenhouse Gas Concentrations on Radiation and Chemistry

GHGs, mainly $CO_2$, water vapor ($H_2O$), $CH_4$, and $N_2O$, warm the troposphere by absorbing outgoing infrared (IR) radiation from the Earth. The dominant balance in the troposphere is between heating through release of latent heat and radiative cooling by GHGs. In the stratosphere, however, increased GHG concentrations lead to a net cooling as they emit more IR radiation into the upper atmosphere than they absorb. IR emission increases with local temperature. Therefore, the cooling effect increases with altitude, maximizing near the stratopause at around 50 km altitude, where temperatures of the stratosphere are highest. The stratospheric cooling effect of GHGs also varies with latitude, as it depends on the balance between absorption of IR radiation from below and local emission of IR radiation. The net cooling effect of GHGs extends to lower levels at high latitudes, roughly following the tropopause.

Any change in concentrations of radiatively active gases will alter the balance between incoming solar (short-wave) and outgoing terrestrial (long-wave) radiation in the atmosphere. For example, ozone absorbs both short- and long-wave radiation. To determine radiative forcing from stratospheric ozone changes, it is important to distinguish between immediate effects and those after the stratospheric temperature has adjusted, which takes some days (in the lower stratosphere) up to weeks (in the upper stratosphere). Depletion of ozone in the lower stratosphere induces an instantaneous increase in the short-wave solar flux at the tropopause and a slight reduction of the downwelling long-wave radiation. The net instantaneous effect is a positive radiative forcing. However, the decrease in stratospheric ozone causes less absorption of solar and long-wave radiation, yielding a local cooling. After the stratosphere has adjusted, the net effect of stratospheric ozone depletion is a negative radiative forcing (Chapter 1 in IPCC/TEAP, 2005).[3] In contrast, ozone depletion in the middle and upper stratosphere causes a slight positive radiative forcing (Chapter 1 in IPCC/TEAP, 2005). The maximum sensitivity of radiative forcing to stratospheric ozone changes is found in the tropopause region, and the maximum sensitivity of surface temperatures to these ozone changes also peaks near the tropopause.[4]

## 8.2.1 Past Temperature Changes

There is strong evidence for a large and significant cooling in most of the stratosphere since the 1960s[5–7] (see also Chapter 4 in WMO, 2011).[2] Figure 8.1 shows global average temperature anomalies (90°S–90°N) derived from the RICH radiosonde data set for the time period from 1960 to 2007, spanning a range of altitudes from the upper troposphere (300 hPa) to the middle stratosphere (30 hPa). Corresponding vertical profiles of near-global temperature trends during the period from 1979 to 2007 are shown in Figure 8.2. The radiosonde data sets indicate a warming of the troposphere and reveal an overall long-term cooling of the stratosphere, with trends increasing with altitude. There is reasonable agreement between the lower stratospheric trends derived from satellite data (since 1979) and radiosondes (see Figure 8.3). All data sets suggest that there has been a significant cooling in the stratosphere over the globe for recent decades, including the tropics (see middle part of Figure 8.2).

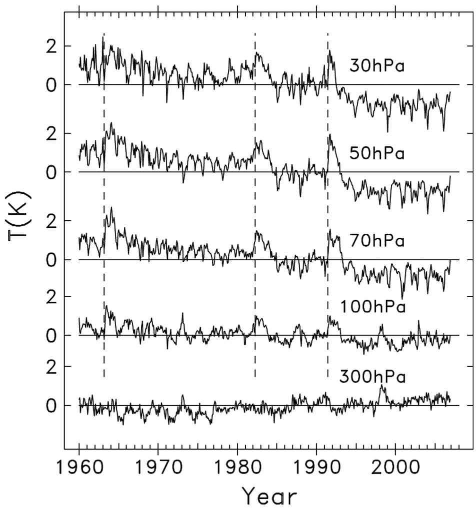


**Figure 8.1** Temporal evolution of global average temperature anomalies (K) at pressure levels spanning the upper troposphere to lower stratosphere derived from the radiosonde data set "RICH". The dashed lines denote the major volcanic eruptions of Agung (March 1963), El Chichon (April 1982) and Mt. Pinatubo (June 1991). Figure taken from Randel (2010).

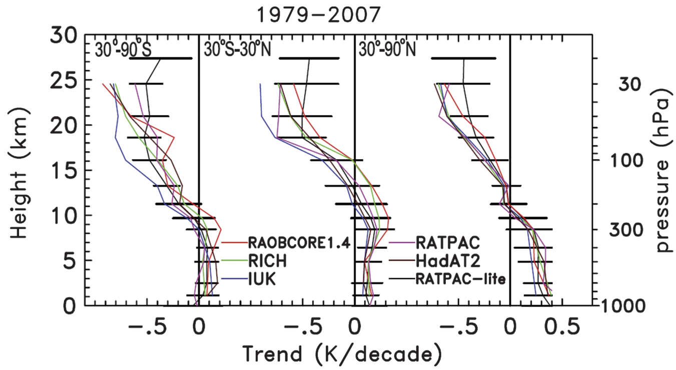


**Figure 8.2** Vertical profiles of annual mean temperature trends (K/decade) for 1979–2007 derived from the separate radiosonde data sets for latitude bands 30°–90°S, 30°N–S, and 30°–90°N. Here the calculated trends are based on a simple linear approach. Error bars show the two-sigma statistical uncertainty levels for the RATPAC-lite data. Figure taken from Randel *et al.* (2009).

Figure 8.3 shows the time series of global-mean stratospheric temperature anomalies as derived from satellite data (Microwave Sounding Unit/Stratospheric Sounding Unit) weighted over specific vertical levels (black lines). There is strong evidence for a large and significant cooling in the stratosphere during the last decades: It is about 0.5 K/decade in the lower stratosphere, whereas in the upper stratosphere the cooling trend increases to about 1.2 K/decade (Chapter 4 in WMO, 2011).[2] In addition, Figure 8.3 contains results from chemistry-climate model simulations (colored lines). The overall development of stratospheric temperature anomalies is mostly well reproduced by the majority of CCMs; so far, the obvious difference after 1998 between the SSU26 data series and results derived from CCMs is unexplained.

The stratospheric cooling has not evolved uniformly in recent decades (see Figures 8.1 and 8.3). A complete interpretation of these changes can only be given considering natural forcing affecting the behavior of the atmosphere, including the 11-year activity cycle of the sun, the quasi-biennial oscillation (QBO) of tropical zonal winds in the lower stratosphere, and large volcanic eruptions.[8,9]

The 11-year solar activity cycle is documented by obvious fluctuations in the intensity of solar radiation at different wavelengths. Eleven-year solar ultraviolet (UV) irradiance variations have a direct impact on the radiation and ozone budget of the middle atmosphere.[10] During years with maximum solar activity, the solar UV irradiance is clearly enhanced (near 200 nm, which is the wavelength range most important for the formation of ozone, with the difference between maximum and minimum activity amounts to 6%–8%),[11] which leads to additional ozone production and heating in the stratosphere and above. By modifying the meridional temperature gradient, the heating can alter the propagation of planetary and smaller-scale waves that drive the global

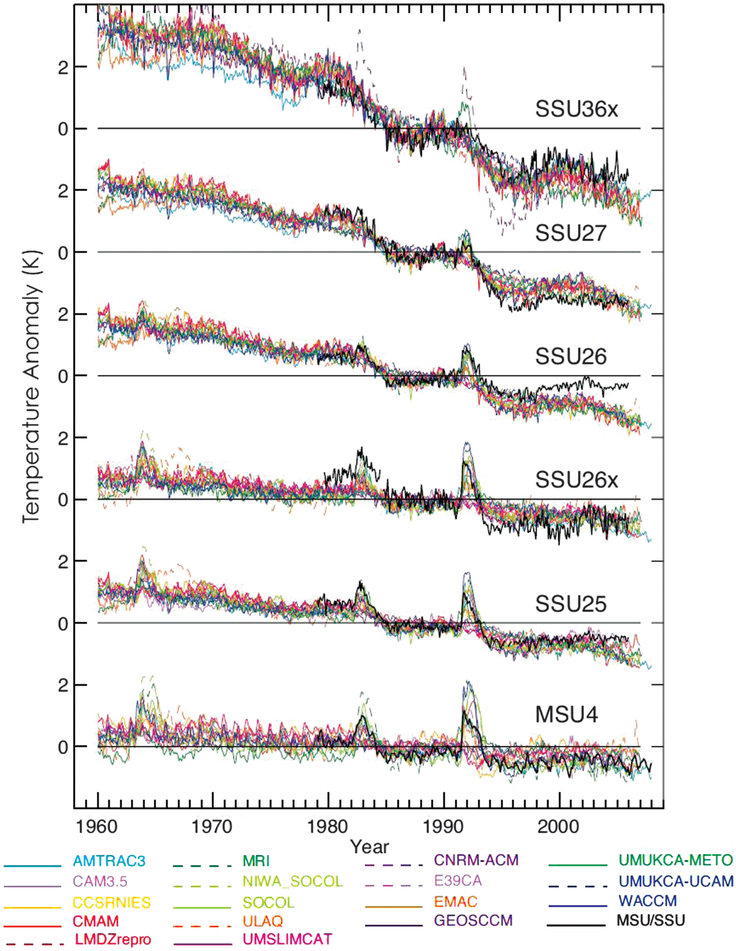


**Figure 8.3** Time series of global mean temperature anomalies (K) derived from satellite measurements (Microwave Sounding Unit: MSU; Stratospheric Sounding Unit: SSU; black lines) and chemistry-climate model calculations (colored lines), weighted for MSU/SSU weighting functions. The anomalies are calculated with respect to the period 1980–1994. SSU Channel 27 corresponds to ~34–52 km altitude, channel 36x to ~38–52 km, channel 26 to ~26–46 km, channel 25 to ~20–38 km, channel 26x to ~21–39 km, and channel MSU4 to ~13–22 km. Figure taken from Chapter 4 in WMO (2011).

circulation (see also Section 8.3). Although the direct radiative forcing of the solar cycle in the upper stratosphere is relatively weak, it could lead to a large indirect dynamical response in the lower atmosphere through a modulation of the polar night jet and the Brewer-Dobson circulation.[12] Such dynamical changes can affect the chemical budget of the atmosphere because of the temperature dependence of the chemical reaction rates and transport of chemical species.

The Arctic lower and middle stratosphere tends to be colder and less disturbed during west wind phases of the QBO (in the lower stratosphere near 50 hPa) while they are warmer and more disturbed during QBO east wind phases.[13–15] Further analyses[16,17] showed that this relationship is strong during solar minimum conditions, while during years around maximum solar activity the relationship does not hold. This solar-QBO relationship has remained robust in the observations since its discovery. Equatorial upper stratospheric winds (near 1 hPa) during the early winter appear to be relevant for the evolution of the northern hemisphere winter, especially the timing of stratospheric sudden major warmings.[18–20]

An obvious sign of transient warming (for about two years) is observed in the lower and middle stratosphere (Figures 8.1 and 8.3) following the large volcanic eruptions of Agung (1963), El Chichón (1982) and Mt. Pinatubo (1991). These warmings are mainly caused by the large amount of sulphur dioxide transferred into the lower stratosphere, leading to the formation of sulphuric acid aerosols, and an enhanced absorption of IR radiation.

In the lower stratosphere the long-term cooling manifests itself as more of a step-like change following the volcanic warming events.[9,21,22] The overall lower stratospheric cooling is primarily a response to ozone decreases, with a possible but much less certain contribution from changes in stratospheric water vapor.[23,24] Ramaswamy *et al.* (2006)[8] and Dall'Amico (2010)[9] suggested that the step-like time series behavior is due to a combination of volcanic, solar cycle and ozone influences. There is a substantial flattening of these temperature trends evident in Figure 8.1 after approximately 1995. The latter aspect agrees with small global temperature trends in the upper stratosphere and lower mesosphere observed in HALOE data for 1992–2004.[25] Although some flattening might be expected in response to the beginning recovery of the stratospheric ozone layer, the strength of this behavior is curious in light of continued increases in long-lived GHG concentrations during this decade.

### 8.2.2 Expected Future Temperature Changes

Long-term changes in radiative forcing over the coming decades are expected to continue to impact global mean temperatures in both the troposphere and stratosphere. As discussed in Subsection 8.2.1, over the past three decades increases in GHG concentrations and the decline in the amount of stratospheric ozone have been the primary forcing mechanisms affecting stratospheric climate.

Global concentrations of GHGs are expected to rise for at least the next half century, although significant uncertainties remain as to the exact rate of increase. It is expected that these changes will be dominated by increases of GHG concentrations and lead, on average, to a cooling of the stratosphere, but there can still be a seasonal warming, particularly at higher latitudes caused by modifications of planetary wave activity. Therefore, the assessment of the future evolution of polar temperatures is uncertain (see Chapter 9). Future changes in stratospheric water vapor are even more difficult to predict, in part because the changes observed over the last four decades are still not fully understood.[5,26,27] While declines in stratospheric ozone also yield a cooling of the stratosphere, within the recent decade (2000–2010) global ozone levels have begun to rise (Chapters 2 and 3 in WMO, 2011;[2] Chapter 9 of this book). Higher ozone levels will increase stratospheric ozone heating, which will at least partially offset the cooling due to increases in GHG concentrations. Because ozone concentrations are so sensitive to the background temperature field, understanding the complex interaction between changing constituent concentrations and temperature requires an evaluation of the coupling between chemistry, radiation and atmospheric dynamics.

### 8.2.3 Temperature Ozone Feedback

The assessment of the sensitivity of ozone-related chemistry to climate changes is complicated, since accompanying modifications occur in the dynamics, transport and radiation. In particular, temperature changes affect physical, dynamical, and chemical processes influencing the ozone content in the atmosphere in different ways. Moreover, the chemical state of the atmosphere changes as the concentration of trace species and ODSs change. This in turn alters the sensitivity of the stratospheric chemical system to temperature changes.

The sensitivity of ozone chemistry in the upper part of the stratosphere (about 35 to 50 km) to changes in temperature is well explained. There the chemical system is generally under photochemical control and is constrained by gas-phase reaction cycles that are well known. The largest stratospheric cooling, which is associated with increased GHG concentrations has been observed in the upper stratosphere and mesosphere (50 to 100 km). The most important ozone loss cycles in the upper stratosphere (via the catalysts $NO_X$, $ClO_X$, and $HO_X$) are slowing as temperatures decrease[28] (see Chapter 1), leading to higher ozone concentrations. In the lower mesosphere, enhanced ozone concentrations are primarily due to the negative temperature dependence of the reaction $O + O_2 + M \rightarrow O_3 + M$. The situation is more complex in the upper stratosphere and lower mesosphere with different ozone loss cycles having greater influence on ozone concentrations at different altitude ranges (*e.g.* $HO_X$ between about 45 and 60 km; $ClO_X$ between about 45 and 50 km; $NO_X$ between about 30 and 50 km). There the slower loss rates are controlled both by the temperature dependence of the reaction rate constants and by the reduction in the amount of atomic oxygen (change in $O_X$ partitioning). The rate-limiting reactions for all the ozone loss cycles are proportional to the atomic oxygen number density. The atomic oxygen number density, in turn, is also strongly appointed by the reaction $O + O_2 + M \rightarrow O_3 + M$.[29]

The situation is more complicated in the polar lower stratosphere (about 15 to 25 km) in late winter and spring. In addition to the gas-phase ozone loss cycles described above playing a similar role in determining the ozone concentration[30] there is an offset by chlorine- and bromine-containing reservoir species. These chemical substances are activated *via* heterogeneous processes on surfaces of polar stratospheric cloud (PSC) and cold aerosol particles, leading to markedly increased concentrations of $ClO_X$ and $BrO_X$. This in turn yields significant ozone losses *via* $ClO_X$ and $BrO_X$ catalytic cycles in the presence of sunlight. The rate of chlorine and bromine activation is strongly dependent on stratospheric temperatures, increasing significantly below approximately 195 K. In a polar lower stratosphere with enhanced (over natural levels) concentrations of ODSs, as it is currently observed, chlorine and bromine activation and consequent ozone losses at lower temperatures, counteract any ozone increase through temperature driven reduction in $NO_X$ and $HO_X$ gas-phase ozone loss.

Due to systematic differences in the winter and spring season temperatures in the Arctic and Antarctic lower stratosphere, ozone concentrations are developing different in the northern and southern polar stratosphere. In the Antarctic stratosphere temperatures are almost always below the threshold for heterogeneous activation of chlorine and bromine containing species during the winter and in early spring. In the case of elevated ODS concentrations this leads to a significant depletion of ozone and the formation of the Antarctic ozone hole. In contrast, the Arctic lower stratosphere is dynamically much more active and lower stratospheric temperatures are principally higher (about 10 to 15 K). Here temperatures lie close to the threshold value for activation of chlorine and bromine species and ozone depletion *via* heterogeneous chemical reactions are much smaller. Consequently, a significant change in Arctic stratospheric temperatures, for example, a cooling due to climate change would strongly influence springtime ozone concentrations in the Arctic region.

While it is expected that the reduction of ODSs in the next decades will lead to enhanced ozone concentrations, this increase could be affected by modifications in temperature, chemical composition and transport (see also Section 8.3). The future evolution of ozone concentrations is sensitive to changes in both chemical constituents *and* climate. A further cooling of the polar lower stratosphere could delay the recovery of the ozone layer in the Arctic and Antarctic region but on the other hand it could accelerate the ozone recovery in other parts of the stratosphere (see Chapter 9). As mentioned earlier, not only is temperature-dependent chemistry affecting the stratospheric ozone content, but so are dynamical processes. This topic is discussed in the following section.

## 8.3 Impact of Enhanced Greenhouse Gas Concentrations on Stratospheric Dynamics

Even in the absence of ODSs, climate change can alter the distribution of stratospheric ozone. Climate and circulation changes affect the transport of trace gases and particles within the stratosphere. But climate change also influences the air mass exchange between the troposphere and the stratosphere, for example, the entry of chemical substances into the stratosphere. Therefore, the lifetimes of long-lived chemical substances are determined by the exchange rate of air masses across the tropopause.[31]

Although the stratosphere and troposphere are different in many ways, the atmosphere is continuous, allowing vertical wave propagation and a variety of other dynamical interactions between these regions. A complete description of atmospheric dynamics requires a full understanding of both layers. The dynamical coupling of the stratosphere and troposphere is primarily mediated by the dynamics of atmospheric waves. A variety of such waves originates in the troposphere, propagates upward into the stratosphere and higher up and then dissipates, forming the spatial and temporal structure of stratospheric motions. Moreover, the stratosphere not only shapes its own temporal evolution but also that of the troposphere. Tropospheric and stratospheric dynamics as well as the dynamical coupling of both altitude regions are affected by climate change.

### 8.3.1 Importance of Atmospheric Waves

The activity of atmospheric waves is divided into three consecutive processes: the generation mechanisms (*i.e.* forcing of waves), the propagation through the atmosphere, and the dissipation of waves primarily due to wave breaking and thermal damping. Outside the tropical region the temperature structure of the stratosphere depends mostly on a balance between diabatic radiative heating and adiabatic heating from induced vertical motion due to dissipation of large-amplitude, planetary-scale Rossby waves.[32] These waves have typical wavelengths of several thousands of kilometres (wave numbers one to three). They either remain stationary or propagate very slowly from the east to the west (*i.e.* quasi-stationary waves) because they are generated due to a combination of overflows of large-scale orographic barriers (like the Rocky Mountains, the Andes, or the Himalaya), meridional temperature gradients, and the Coriolis force (as a consequence of Earth's rotation). Depending on background wind conditions these waves can become unstable somewhere in the stratosphere-mesosphere region (up to 100 km) while propagating upwards.[33] This so-called "wave breaking" leads to a deposition of thermal energy in the upper stratosphere and mesosphere, while planetary waves deposit easterly momentum in the lower stratosphere, which both decelerating the west wind of the stratospheric polar vortex in wintertime (*i.e.* the polar night jet). The weakening of the westerly jets has to be adjusted in the geostrophic balance and causes therefore a small meridional wind component that drives the residual circulation.[34] The stirring of air isentropically across larger distances of the winter stratosphere within a region is known as the "surf zone".[35] This region is bounded by sharp gradients of tracer concentrations in the winter subtropics and at the edge of the polar night jets. The deceleration of the polar vortex (*i.e.* weakening of the zonal wind speed) is accompanied by a warming of the stratosphere at higher latitudes resulting in a thermodynamic imbalance yielding a radiative cooling of the polar stratosphere back towards radiative equilibrium. Consequently, downwelling of polar air masses is enhanced, which must be balanced by a poleward flow of air from lower latitudes. This response pattern describes a meridional circulation which is called the Brewer-Dobson (BD) circulation (see Subsection 8.3.2).

The basic climatology of the extra-tropical stratosphere is mostly understood in terms of large-scale wave dynamics together with the seasonal cycle of radiative forcing. For example, the easterly winds of the summer stratosphere prevent upward propagation of planetary waves.[33] Therefore, in summer stratospheric variability is much smaller than in winter. Dissimilar distribution of the continental land masses between the northern and the southern hemisphere imply asymmetries in the efficiency of planetary wave generation mechanisms. Consequently, in the northern winter stratosphere, planetary wave disturbances are significantly larger than those in southern winter.

In the tropical lower and middle stratosphere between 100 and 10 hPa, the prevailing variability mode is the quasi-biennial oscillation (QBO) of the tropical zonal mean wind field, alternating between westerly and easterly winds with a mean period of approximately 28 months. The alternating wind regimes repeat at intervals that vary from 22 to 34 months (see Baldwin *et al.* (2001)[36] for a review). The peak-to-peak amplitude of the zonal wind speed is about 55 m/s near 20 hPa.[37] Maximum easterlies are generally stronger than westerlies, *i.e.* $\sim$35 m/s and $\sim$20 m/s, respectively. The QBO signal in temperature amplitude is approximately 8 K. The QBO affects the global stratospheric circulation and, therefore, influences a variety of extra-tropical phenomena including the strength and stability of the polar night jet, and the distribution of ozone and other gases (Baldwin *et al.*, 2001). The QBO is mainly driven by the dissipation of a variety of west- and eastward propagation large-scale equatorial waves.[38,39]

Although stratospheric variability has long been viewed as being caused directly by variability in tropospheric wave sources, it is now widely accepted that the configuration of the stratosphere itself also plays an important role in determining the vertical flux of wave activity from the troposphere because of the strongly inhomogeneous nature of the stratospheric background state, for example, the steep gradient of zonal wind and potential vorticity at the edge of the polar vortex.[40] Given a steady source of planetary waves in the troposphere, any modulation in stratospheric background, for example, gradients of temperature, winds or potential vorticity alter the vertical wave fluxes, giving rise to the possibility of internally driven variability of the stratosphere, as it was already demonstrated in an early numerical model study by Holton and Mass (1976).[41] Some recent idealized modeling studies even suggest that realistic stratospheric variability can also arise in the absence of tropospheric variability.[42,43]

Further, the tropospheric circulation itself is also influenced by the stratospheric configuration. Reflection of stationary planetary wave energy back into the troposphere can occur when the polar vortex exceeds a critical threshold in the lower stratosphere, yielding structural changes of the leading tropospheric variability patterns.[44–46]

Not only is the consideration of large-scale planetary and synoptic wave dynamics very important in determining the climatology of the stratosphere, but also the effects of smaller-scale waves (*e.g.* gravity waves forced by orography or convective events) must be considered for explanations of stratospheric dynamic variability. Any systematic change in processes affecting generation, propagation or dissipation of all waves results in systematic changes of the temperature structure of the stratosphere. The capability of numerical models of the atmosphere to simulate the climatology and space-time changes of stratospheric properties depends critically on the ability to simulate highly nonlinear wave dynamics in a robust way. One of the most challenging aspects of modelling the dynamical coupling of the troposphere and stratosphere is the parameterization of the effects of unresolved waves (in particular gravity waves) and their feedback on the resolved flow. Global atmospheric models mostly have insufficient horizontal and vertical resolution to resolve all necessary characteristics and effects, and therefore they must be prescribed, *i.e.* parameterized in the models. Another issue is that deep convection (an important excitation mechanism for waves that propagate into the stratosphere) is a sub-grid-scale process that must also be parameterized.

### 8.3.2 The Brewer-Dobson Circulation and Mean Age of Air

The meridional circulation in the stratosphere is called the Brewer-Dobson (BD) circulation, named after Alan Brewer and Gordon Dobson, to honor their fundamental research studies of stratospheric water vapor and ozone measurements.[48,49] The BD circulation is the major driver for transport of stratospheric air masses from tropical to higher latitudes. Its climatology is characterized by rising motion of air in the tropics from the troposphere into the stratosphere and poleward transport there (Figure 8.4). The BD circulation is more pronounced in the winter hemisphere. Due to mass conservation, descending motion of air occurs in stratospheric middle and higher latitudes, mixing stratospheric air back into the troposphere.

One reason could be that the BD circulation should result from solar heating in the tropical region and cooling at higher latitudes, leading to a large circulation cell reaching from the tropics to polar regions as warm tropical air ascends and cold polar air descends. Although this reasoning seems consistent with the observations, the BD circulation mainly results from planetary wave forcing observed in the extra-tropical stratosphere. Thus, the BD circulation has been likened to an enormous wave driven pump.[47]

In particular, the wintertime stratosphere is dominated by planetary (Rossby) waves propagating upward from the troposphere (see Subsection 8.3.1). The existence of the BD circulation is strongly linked to planetary wave activity (Figure 8.4). The BD circulation is different in the southern and northern hemisphere because of hemispheric differences in land–ocean distributions. This leads to more frequent and intense planetary wave activity and stronger BD circulation in the northern winter season. In wintertime, the horizontal mixing in the northern hemisphere often reaches the polar region, whereas in the southern

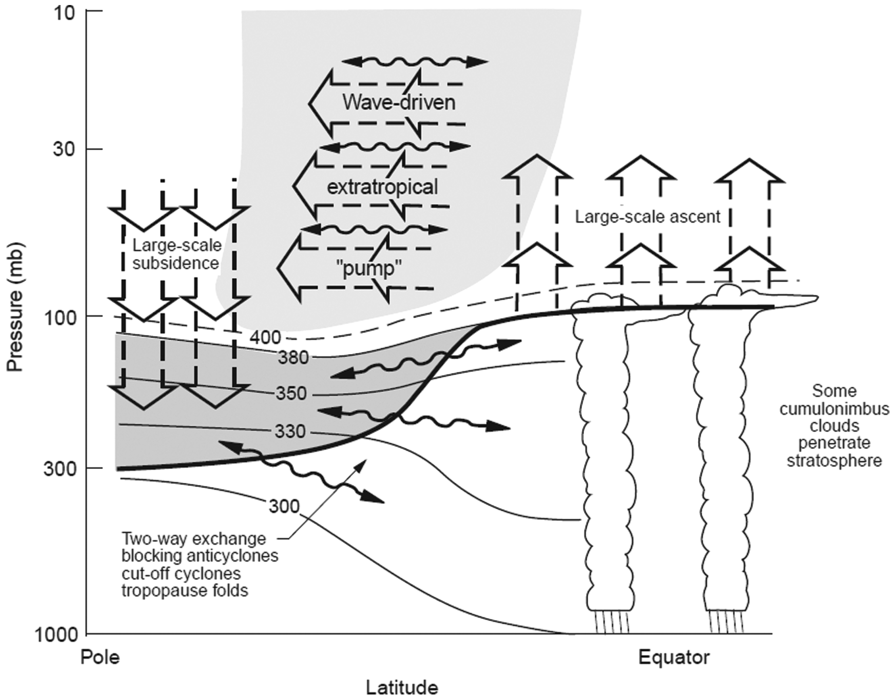


**Figure 8.4** Dynamical aspects of stratosphere-troposphere exchange. The tropopause is shown by the thick black line. Thin lines denote isentropic (*i.e.* constant potential temperature) surfaces (K). The light shaded area in the stratosphere denotes wave-induced forcing, *i.e.* the extra-tropical pump area, driving the Brewer-Dobson circulation (see text). The heavy shaded area indicates the so-called lowermost stratosphere. Figure taken from Holton *et al.* (1995).

hemisphere horizontal mixing is most of all confined to lower and middle latitudes and seldom reaches the Antarctic region.

Among other factors, the BD circulation determines the distribution of ozone (Figure 8.5) and water vapor in the stratosphere, but it also affects the lifetimes of ozone anthropogenic ODSs, and of some GHGs. It is expected that climate change will modify processes responsible for the generation of the BD circulation.

So far observations provide an unclear picture of trends in the strength of the BD circulation, as intrinsic variability is not well known and effects are difficult to measure. Analyses of the BD circulation based on chemical measurements (and estimates of the mean age of air, see below) are often very noisy and have error bars that exceed the amplitude of the observed trends[51] (see also Chapter 4 in WMO, 2011).[2]

Recent studies of General Circulation Model (GCM) and Climate-Chemistry Model (CCM) simulations consistently indicate an acceleration of the BD

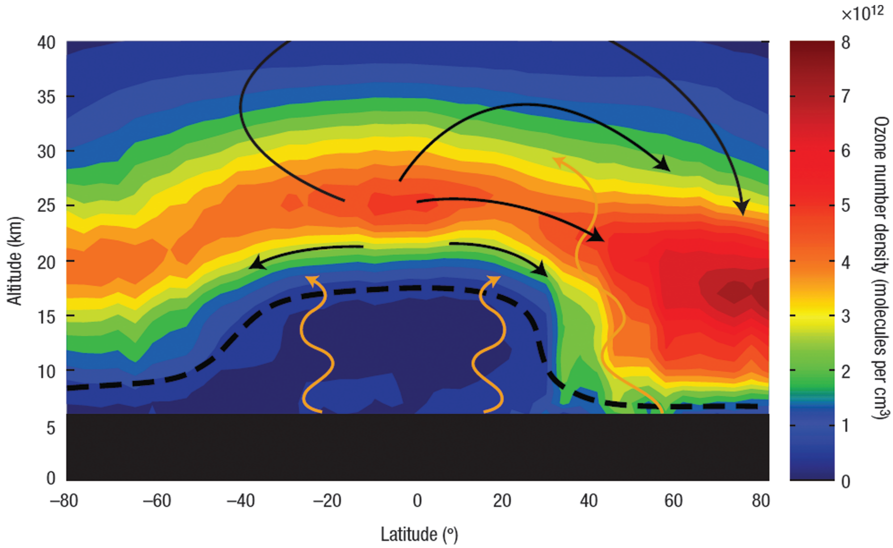


**Figure 8.5** Brewer-Dobson circulation and stratospheric ozone. A longitudinally averaged cross-section of the atmosphere shows a schematic of the meridional stratospheric circulation, i.e. the Brewer-Dobson circulation (black arrows), and the ozone distribution (molecules $cm^{-3}$) as measured by the OSIRIS satellite instrument in March 2004. The circulation is forced by waves propagating up from the troposphere (orange wiggly arrows), especially in the winter hemisphere, and it strongly shapes the distribution of ozone by transporting it from its source region in the tropical upper stratosphere to the high-latitude lower stratosphere. Consequently, ozone number densities are higher at polar latitudes than in the tropics. The dashed line represents the tropopause. Copyright OSIRIS Science Team. Figure taken from Shaw and Shepherd (2008).

circulation in response to increasing greenhouse gas concentrations in future with distinct consequences for the recovery of the ozone layer[52–55] (see also Chapter 4 in SPARC CCMVal, 2010[56] for an overview). As an example, Figure 8.6 presents results derived from a number of CCM simulations showing an almost steady increase of tropical upwelling in the future. The net upward mass flux of air into the tropical stratosphere is taken as a measure of the BD circulation. The implications for a future increase in the BD circulation are substantial. Such an increase would, for example, change the spatial distribution of stratospheric ozone, with increased total ozone at high latitudes and decreased total ozone in the tropics. It would also increase the stratosphere-to-troposphere ozone flux[57] and it would decrease the net age of stratospheric air.[53,58–60]

A stronger BD circulation would tend to warm the extra-tropical regions and cool the tropics. It would have direct implications for injection of tropospheric source gases (*i.e.* original gases transported into the stratosphere and then reacts there), product gases (*i.e.* intermediate or final products produced in the

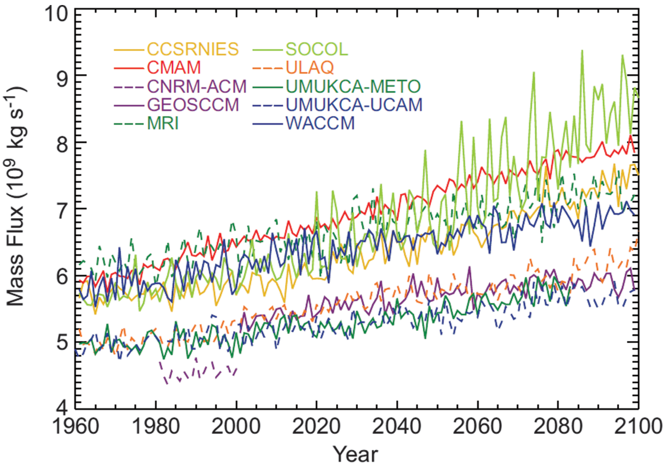


**Figure 8.6** Annual mean upward mass flux (kg s$^{-1}$) at 70 hPa, calculated from residual mean vertical velocity between the turnaround latitudes of the BD circulation; results derived from CCM simulations. Figure taken from Chapter 4 of SPARC CCMVal (2010).

troposphere), and water vapor into the stratosphere and for transport of stratospheric air mass with high ozone concentrations and very low water vapor content into the troposphere. The two-way coupling between troposphere and stratosphere is also demonstrated by the link between ozone depletion and changes in surface climate in the southern hemisphere[61,62] (Chapter 4 in WMO, 2011).[2] Details are discussed in Chapter 7.

Observational investigations of the variability and long-term changes of the BD circulation have not yet provided a clear picture. Nedoluha *et al.* (1998),[63] for example, reported that a slower BD circulation could explain the negative trend in upper stratospheric methane, whereas Waugh *et al.* (2001)[64] argued that a faster BD circulation could explain the observed trend in upper stratospheric chlorine which has been measured by the Halogen Occultation Experiment (HALOE) onboard the Upper Atmosphere Research Satellite (UARS). A weakening of the BD circulation was shown by Salby and Callaghan (2002),[65] consistent with the results presented by Hu and Tung (2003),[66] who found a reduction of planetary wave activity occurring only in late winter (1979–1999) and no obvious change before 1979. They proposed that ozone depletion could be the reason for this reduction due to radiative-dynamical feedback increasing ozone depletion. Conversely, a strengthening of the BD circulation by ozone depletion was suggested by Li *et al.* (2008).[67] An accelerated upwelling in the tropical region was identified in long-term observations by Thompson and Solomon (2005)[68] and Rosenlof and Read (2008).[69] Thompson and Solomon (2009)[7] demonstrated that the contrasting latitudinal structures of recent stratospheric temperature (*i.e.*, stronger cooling in the tropical lower stratosphere than in the extra-tropics) and ozone trends (*i.e.*, enhanced ozone reduction in the tropical lower stratosphere) are consistent with the assumption of an accelerated stratospheric overturning BD circulation.

The observed drop in the tropical lower stratospheric water vapor concentrations after 2001 (see Figure 8.7) is consistent with an enhanced tropical upwelling during that time[27] (see Chapter 4 in WMO, 2011)[2] and a step-like increase in the summed extra-tropical activity of planetary waves from both hemispheres[71] indicating an accelerated in BD circulation in both hemispheres. Other studies suggested that changes in the BD circulation can account for a large fraction of the long-term total ozone decline outside the polar regions (*i.e.* weakening of BD circulation) until the mid-1990s and recent increases in northern middle latitude spring (*i.e.* strengthening of BD circulation).[72–74] Similar results were observed in chemical-transport models (CTMs) using ECMWF reanalysis data.[75]

The mean stratospheric transport time from the tropical lower stratosphere, where tropospheric air enters the stratosphere to any point in the stratosphere

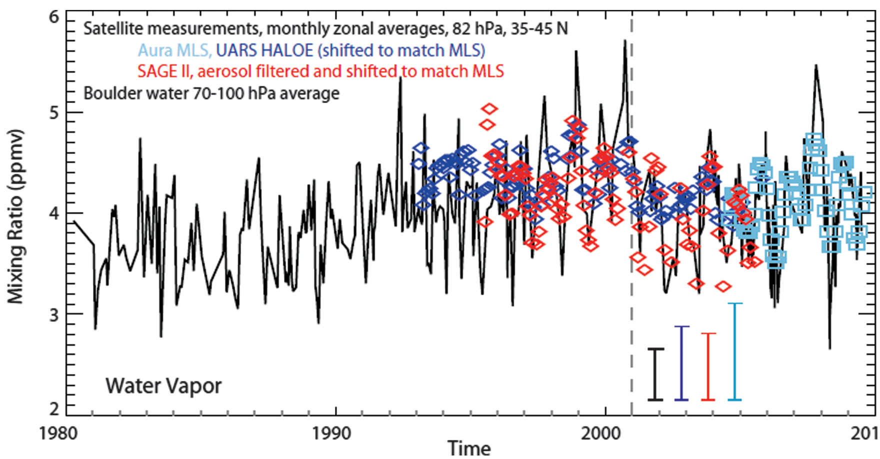


**Figure 8.7** Observed changes in stratospheric water vapor mixing ratio (ppmv). Time series of stratospheric water vapor mixing ratio (ppmv) averaged from 70 to 100 hPa near Boulder Colorado (40°N, 105°W) from a balloon-borne frost point hygrometer covering the period 1981 through 2009; satellite measurements are monthly averages, balloon data plotted are from individual flights. Also plotted are zonally averaged satellite measurements in the 35°N-45°N latitude range at 82 hPa from the Aura MLS (turquoise squares), UARS HALOE (blue diamonds) and SAGE II instruments (red diamonds). The SAGE II and HALOE data have been adjusted to match MLS during the overlap period from mid-2004 to the end of 2005, as there are known biases (Lambert *et al.*, 2007). Representative uncertainties are given by the colored bars; for the satellite data sets, these show the uncertainty as indicated by the monthly standard deviations, while for the balloon data set this is the estimated uncertainty provided in the Boulder data files. Figure adapted from Solomon *et al.* (2010).

can in principle be determined by observations of inert trace gases exhibiting a pronounced temporal trend in the troposphere, like sulphur hexafluoride ($SF_6$), another important GHG used by the electricity industry. The spatial distribution of these transport times, also called mean age of air, is a characterization of the BD circulation. For instance, tropical upwelling is inversely related to mean age of air so that the ''age of air'' changes as the stratospheric climate changes.[58,76] The global distribution of the mean stratospheric transport time of air can be assessed using observations of the GHG $SF_6$, for example, from ESA's MIPAS/ENVISAT instrument.[77] So far the available time series derived from satellite measurements are too short to estimate robust trends since interannual variability is very high. Assessments of ascent rates in the lower tropical stratosphere have been provided using different methods including satellite-based observation of water vapor and diabatic heating rates.[78–81] Several CTMs were compared with respect to their transport properties and strategies have been developed to use the meteorological re-analyses for multi-annual simulations and to improve the model performance.[82,83] The multi-annual CTM simulations show a slight acceleration of the BD circulation for the past 30 years, which seems to be in contrast with estimates derived from $CO_2$ and $SF_6$ balloon measurements executed in the last 30 years.[51]

While current observational data records are too short to derive statistically significant trends in the BD circulation, several independent studies with both General Circulation Models (GCMs) and Climate-Chemistry Models (CCMs) have indicated that the BD circulation will strengthen and that the mean age of stratospheric air will decrease in a future climate with enhanced GHG concentrations[31,52–55,59,60,67,84–89] (see also Chapter 4 in WMO, 2011).[2] Although this strengthening of the BD circulation has been identified as a robust feature of many climate change simulations, the underlying mechanisms are so far not sufficiently understood to explain the cause and effect relationship. The acceleration of the BD circulation may result from an increase of the extra-tropical generation of planetary waves[59] among others produced by an enhanced temperature gradient between the tropics and extra-tropical regions in the upper troposphere and lower stratosphere (UTLS) region. This is associated with global warming in the models, leading to an enhanced poleward eddy heat flux in the stratosphere.[90] The possible impact of enhanced generation of planetary waves in the (sub)-tropical region caused by higher tropical SSTs and its importance for an intensified upwelling in the tropical UTLS region is discussed for example in Deckert and Dameris (2008).[91] More details are presented in Subsection 8.3.3.

Due to the difficulties in obtaining long-term observations of dynamical parameters in the stratosphere (with the exception of temperature and ozone), a validation of past changes of stratospheric dynamics has not yet been performed in a sufficient way. Li and Waugh (1999),[92] for example, showed that the mean age of air in their two-dimensional model is mainly sensitive to changes in the BD circulation, while changes in mixing show a much weaker effect on mean age. This situation is reversed for tracers with chemical sinks in the stratosphere, for example, nitrous oxide ($N_2O$) and chlorofluorocarbons

(CFCs). Changes in the intensity of horizontal transport and mixing should influence distributions and correlations of tracers like $N_2O$ and CFCs, observable quantities which can be linked to changes in stratospheric dynamics. Correlations between long-lived tracers in the stratosphere are known to be very robust and are sensitive to mixing processes in the stratosphere[92] (see Plumb, 2007 for a review).[93]

Austin and Li (2006)[58] suggested that a long-term change of mean age of air, on the order of more than half a year, should already have occurred since the mid 1970s. This is in contrast to the findings of Engel *et al.* (2009).[51] As age of air can be derived from observations, this provides a quantity for the validation of modeled changes in the BD circulation. A strengthening of the BD circulation due to increased GHG concentrations would enhance stratosphere-troposphere mass fluxes, which would have important consequences, for example, a reduction of the lifetimes of ODSs, changes in high- and low-latitude temperature, or enhanced downward transport of stratospheric ozone into the troposphere. Such effects influence the future evolution of atmospheric composition, particularly concentrations of ozone and water vapor.

Further consequences will certainly depend on particular future changes in the BD circulation. While the sign of the predicted change in the BD circulation is consistent between atmospheric models, the magnitude and the detailed structure, and therefore the consequences are not. At the moment the differences between models are difficult to interpret because the reasons for the changes of the BD circulation remain unclear. As long as the causes and mechanisms for a possible enhancement of the BD circulation and the magnitude of these changes remain insufficiently explained, the consequences for stratospheric parameters cannot be predicted with the necessary robustness. This uncertainty limits the validity of prognostic studies regarding the future evolution of climate and atmospheric chemical composition, in particular of the stratospheric ozone layer (see Chapter 9).

### 8.3.3 The Role of Sea Surface Temperatures

With higher GHG concentrations leading to higher tropospheric temperatures, the oceans will absorb heat and sea surface temperatures (SSTs) will tend to increase. SSTs affect the activity of atmospheric waves, *i.e.* generation, propagation and dissipation of planetary as well as gravity waves, and hence the BD circulation.[59,91,94–97] Most of the recent numerical modeling studies that are based on simulations with GCMs and CCMs are imposing SSTs without atmospheric feedback in order to make multi-decadal integrations feasible[52–54,98,99] (see also SPARC CCMVal, 2010).[56] The SST fields are either taken from observations or climate model simulations (*i.e.* coupled Atmosphere-Ocean General Circulation Models, AOGCMs). To a large extent, this practice reproduces zonal-mean hydrological characteristics and interannual variability in stratospheric dynamics of the respective observations and AOGCM simulations.[94,96,100] These investigations indicate that there is an obvious distinction in tropospheric reaction to prescribed tropical and extra-tropical SST anomalies, with implications for stratospheric dynamics (*e.g.* tropical upwelling) and chemistry (*e.g.* polar ozone). The tropospheric reaction to SST anomalies in the tropics is mostly barotropic and deep-convection mediated. The reaction to SST anomalies outside the tropical region is weaker, shallower, and more complicated, because baroclinicity is involved and latent-heat release from convection is much weaker. This makes it more difficult to study atmospheric changes from SST modifications in middle and higher latitudes.[101,102]

In the case of prescribed higher SSTs in tropical regions, numerical simulations with GCMs and CCMs have indicated that deep convection is an important messenger transferring the SST signal into the stratosphere. Currently, there are two mechanisms discussed that may act simultaneously and that both involve SST-related an intensification of tropical deep convection.[103] The first mechanism is based on higher temperatures in the tropical and subtropical upper troposphere resulting from stronger latent-heat release due to intensified deep convection.[59,104] This increases the latitudinal temperature contrast between the tropical upper troposphere and extra-tropical lowermost stratosphere, strengthening locally the zonal wind velocity. The altered wind profile influences planetary wave activity, altering tropical lower stratospheric wave dissipation and hence tropical upwelling and the BD circulation.[97] However, it remains unclear whether a modification of wave generation mechanisms or modifications in wave propagation conditions dominate this impact on the BD circulation and whether waves generated in the tropics or extra-tropics are involved.

The second mechanism focuses on changes in tropical upper tropospheric pressure perturbations that are associated with intensified deep convection due to enhanced SSTs in the tropics. Deckert and Dameris (2008).[91] considered both SST-induced modifications to the convection-related eddy dissipation in the tropical lower stratosphere and the associated implications for the BD circulation. Comparable to a stone hitting a water surface, pressure perturbations excite tropical quasi-stationary planetary waves. They propagate upward as they dissipate but carry enough of the SST signal across the tropical tropopause into the lower stratosphere to affect the tropical upwelling *via* the principle of downward-control.[105] Additionally, Chen *et al.* (2001)[106] and Chen (2001)[107] demonstrated that these eddies can ascend through the tropical easterly winds and cross the tropopause. Rind *et al.* (2002)[108] and Fomichev *et al.* (2007)[89] investigated atmospheric conditions with doubled $CO_2$ concentrations. They inferred a pattern of amplified eddy dissipation occurring in the tropical lower stratosphere too, appearing to accelerate the upwelling in the tropics by stimulating an anomalous BD cell locally. According to these numerical studies, the strengthening of eddy dissipation is mainly referring to enhanced tropical SSTs, but the importance of deep convective quasi-stationary eddy generation was not yet proved in these investigations.

Both mechanisms discussed above are able to explain observational signs for an accelerated upwelling in the tropics across the tropopause. Satellite as well as radiosonde data indicate that a reduction in temperatures and ozone concentrations have occurred over the past four decades, particularly in the tropical lower stratosphere at all longitudes and during all seasons.[68,109] This finding is consistent with the hypothesis of intensified tropical upwelling. Although studies of stratospheric mass transport trends support this hypothesis, they have large uncertainties.[52,59] Radiative changes as a result of anthropogenic ozone depletion might account for similar modifications of tropical upwelling (Forster *et al.*, 2007).[110] Both convection-related mechanisms fulfil the requirement of enhanced planetary wave breaking at low latitudes. The observed decrease of temperature in the tropical tropopause region in 2001[27] is part of a close relationship between SSTs and lower stratospheric temperatures.[69] There is a clear anticorrelation between SSTs in the western tropical Pacific Ocean—the region on Earth with highest SSTs—and temperatures and ozone and water vapor concentrations in the tropical lower stratosphere. Anomalously high SSTs coincide with low temperatures and ozone concentrations, and *vice versa*. This anticorrelation is unlikely to result from the lifting of air due to convection, since the stratospheric signal occurs at altitudes well beyond the highest-reaching thunderstorms. Both mechanisms discussed above could contribute to this anticorrelation. So far, the response of planetary wave activity to anomalies in convection-related pressure perturbations seems to be a better candidate because it is more immediate than the planetary wave response to anomalies in latitudinal temperature contrast. For example, some of the mentioned numerical sensitivity studies indicated that there are various different latitudes where stratospheric wave breaking is sensitive to SST anomalies.[94,96]

Moreover, it was shown by Brönnimann *et al.* (2006)[111] that wave breaking in the extra-tropical northern stratosphere during winter responds to the El Niño/Southern Oscillation (ENSO) signal in tropical SSTs. Nevertheless, the sensitivity of planetary wave activity to extra-tropical SST anomalies relative to SST anomalies in tropical regions is unknown.[89,112–114] In theory, the tropospheric response to extra-tropical SST changes should influence the life-cycle of planetary waves *via* altered ocean-continent temperature contrast, changes in position and strength of storm tracks, and modified barotropic or baroclinic instability.[32]

## 8.4 Coupling of the Stratosphere and the Troposphere in a Changing Climate

The net mass exchange between the troposphere and stratosphere is mostly associated with the BD circulation[47,115] with a net upward flux in the tropics balanced by a net downward flux in the extra-tropical regions (Subsection 8.3.2). Air in the tropical lower stratosphere rises slowly (about 0.2 to 0.3 mm/s) and carries ozone-poor air from the troposphere higher up into the stratosphere. There, with increasing altitude, photochemical production of ozone becomes more effective. The upwelling in the tropics is clearly modulated by the seasonal cycle and the tropical QBO phase.[36] When the QBO is in its westerly phase the ascent rate is lower, and there is more time for ozone production, which enhances the tropical total ozone column. However, near the tropopause the picture is more complex, with two-way mixing across the extra-tropical tropopause at and below synoptic scales, and vertical mixing in the tropical-tropopause layer (TTL) resulting from convective processes. In the subtropics and extra-tropics there is not only transport of chemical species and particles from the troposphere into the lowermost stratosphere occurring through quasi-isentropic motion (associated with synoptic-scale and mesoscale circulations, *e.g.* baroclinic eddies, frontal circulations) but there is also substantial transport of air masses from the stratosphere into troposphere. Quantification of this two-way transport has improved significantly over recent years through analyses of observations and numerical modelling studies (see Stohl *et al.* (2003)[116] for a review). However, significant quantitative uncertainties remain about the role of small-scale circulations, for instance the importance of convective systems in transporting air from the troposphere into the stratosphere and *vice versa*. A further complication is that all related processes are affected by climate change, making it even more difficult to assess future changes of stratosphere-troposphere connections.

### 8.4.1 Stratosphere-troposphere Coupling

The stratosphere and the troposphere are connected by physical, dynamical and chemical processes. Changes in the concentrations of radiatively active gases in the stratosphere yield significant changes in stratospheric temperature[6,8,23,117] (see Subsection 8.2.1). In addition, especially in winter, the stratosphere is significantly affected by upward propagating tropospheric waves that dissipate in the stratosphere and mesosphere and slow the polar night vortex (Subsection 8.3.1). Statistical analyses of dynamical quantities have demonstrated a strong connection between stratospheric and tropospheric modes of variability.[118–121] For instance, during northern winter, a high correlation exists between the intensity of the stratospheric polar night jet and the North Atlantic Oscillation (NAO) in the middle troposphere,[122] which is a key parameter for weather and climate in Europe. A strong stratospheric polar night jet is associated with a positive phase of the NAO, corresponding to stronger westerlies in the North Atlantic region and positive temperature anomalies over central and northern Eurasia.

Moreover, Thompson and Wallace (1998; 2000)[123,124] identified vertically coherent patterns from the stratosphere down to Earth's surface with more zonally symmetric, quasi-annular anomalies characterized by geopotential anomalies of one sign over the polar cap, offset by anomalies of the opposite sign over lower latitudes. This concept of annular modes applies equally well in either hemisphere, and describes the leading mode of variability from the surface through the stratosphere.[121] This variability pattern is now referred to as the Northern Annular Mode (NAM) and the Southern Annular Mode (SAM). The surface NAM is similar to the NAO pattern, but is geographically broader in scale. For the purposes of stratosphere-troposphere coupling, it makes little difference whether tropospheric variability is described by the NAO or the NAM. While NAM and SAM variability exist in the troposphere throughout the year, the variability extends into the stratosphere during the winter and spring seasons, when stratospheric dynamical variability is enhanced (the stratospheric circulation in the summer is quiescent).

During northern winter, deep positive and negative NAM anomalies are associated with anomalously strong or weak polar zonal wind jets, respectively. Usually they appear first in the upper stratosphere and mesosphere and then propagate downward to the troposphere. There they are seen as anomalies of tropospheric meteorological fields with a time lag of several weeks.[125,126] Thompson *et al.* (2002)[127] found a high correlation between extreme weather events and the strength of the stratospheric polar vortices, hence implying that considering stratospheric anomalies in winter might improve extended-range weather forecasting. Baldwin *et al.* (2003; 2007)[128,129] emphasised the importance of persistent circulation anomalies in the lower stratosphere in winter for the phase of the tropospheric NAM. The annular mode variations are consistent with deep temperature anomalies that result from modulation of the residual circulation. Thus, anomalously strong wave driving (as during a sudden warming) leads to stronger downwelling and warming over the polar cap. The anomalous warming extends into the upper troposphere.

On climate change time scales, changes in tropospheric variability can be associated with stratospheric variability of either natural or anthropogenic origin. For example, Kodera (2002)[130] showed that during the maximum of the 11-year cycle of solar activity, the initial radiative solar signal leads to obvious modifications of the stratospheric zonal mean wind, which are correlated with the NAO index (see above). A similar connection between the stratosphere and troposphere was also found for the SAM index, but only during maximum solar activity.[131] These results, derived from observed data, were supported by investigations performed with GCMs. Matthes *et al.* (2006),[132] for example, detected significant differences in the near-surface geopotential height in the northern hemisphere between minimum and maximum solar activity resembling the signature of the AO, with more positive phases during solar maximum. Moreover, stronger polar night jets during solar maximum are associated with stronger tropospheric cyclone activity in the North Atlantic region and warmer and more humid winters in central Europe and Eurasia.

Temperature anomalies which were identified near Earth's surface following large volcanic eruptions resembled anomalies associated with the positive phase of the NAO.[133,134] Accompanying investigations with data derived from GCM simulations[135,136] showed that dynamical feedback processes initiated by the tropical stratospheric warming after volcanic major eruptions due to enhanced sulphur aerosol loading were responsible for the tropospheric response.[137,138]

In addition to naturally forcing mechanisms of stratosphere-troposphere coupling, man-made contributions are important. The stratosphere is strongly influenced by radiative perturbations, which are caused by reduced stratospheric ozone content and enhanced GHG concentrations including water vapor.[23,24,139] Schwarzkopf and Ramaswamy (2008)[140] analysed multi-decadal climate model simulations with varying (*i.e.* transient) boundary conditions. They found a sustained and significant global, annual mean cooling in the lower and middle stratosphere since about the 1920s, a global temperature change signal developing clearly earlier than in any lower atmospheric region that mostly results from carbon dioxide ($CO_2$) increases. Particularly since the beginning 1980s, stratospheric ozone depletion has strengthen the cooling in the stratosphere. Forster *et al.* (2007)[110] used a "radiative fixed dynamical heating" model to demonstrate that the effects of tropical ozone decreases at about 70 hPa and lower pressures can lead to significant cooling below which is comparable in magnitude to changes of other radiatively active trace gases (*e.g.* Chapter 5 in WMO, 2007).[1]

Thompson and Solomon (2002)[61] documented a clear increase of the tropospheric southern hemisphere circumpolar circulation since the 1970s with a warming of the Antarctic Peninsula and Patagonia and a marked cooling of the East-Antarctic region and the Antarctic plateau which is obviously associated with a shift to a more positive phase of the SAM. Similar tropospheric signatures were derived from GCM simulations with a prescribed stratospheric polar ozone loss. An induced cooling was shown to lead to more positive phases of the surface NAM and SAM[122,141,142] GCM simulations dealing with the influence of increasing GHG concentrations indicated a positive trend in the NAM in the troposphere.[62,143,144]

However, these studies came to contradictory conclusions about the relevance of the stratospheric contribution. Idealized numerical studies also demonstrated that parts of tropospheric variability can be explained by anomalies propagating downward from the stratosphere into the troposphere and that even climate near the Earth surface is affected.[145–148] So far the mechanisms which are important for the downward coupling of the stratosphere and troposphere remain unclear. There are several mechanisms worth considering: (1) Planetary wave activity (see Subsection 8.3.1): the vertical propagation of planetary waves from the troposphere to the stratosphere is affected by stratospheric dynamical conditions. Perlwitz and Harnik (2003)[149] suggested that wave reflection from the upper stratosphere could influence tropospheric circulation. (2) Planetary scale wave-mean flow interaction: the interaction between planetary waves with the zonal mean flow in the stratosphere may lead to a downward propagation of zonal wind and temperature anomalies that could reach the lower troposphere and Earth surface[137,150,151] (3) Direct responses to variances of potential vorticity: Hartley *et al.* (1998)[152] and Black (2002)[153] showed that changes in lower stratospheric zonal circulation can lead to obvious changes of tropopause height and tropospheric wind speed. (4) "Downward control": in case of adequately long anomalous wave driving, secondary equilibrium circulations develop in the stratosphere extending to the troposphere (Haynes *et al.*, 1991;[105] see also Subsection 8.3.2). (5) Influence of stratospheric conditions on baroclinic instability in the troposphere: for example, Wittmann *et al.* (2004)[154] found that the addition of a stratospheric jet to the tropospheric jet yielded a net near-surface geopotential height anomaly that is strongly similar to the AO. Synoptic-scale tropospheric responses to stratospheric changes were also identified by Charlton *et al.* (2004).[155] Moreover, simulations with simplified GCMs indicated that changes in both planetary wave propagation and planetary wave energy due to tropospheric climate change are important.[112,113] (6) Geostrophic and hydrostatic adjustment of the tropospheric flow to anomalous wave drag[105,156] and anomalous diabatic heating at stratospheric levels.[156]

It must be kept in mind that the mechanisms mentioned above are not independent from each other, and on the other hand it is also not clear so far how they act together to produce the observed stratosphere-troposphere coupling.

On the one hand the investigations by Thompson and Solomon (2002)[61] discussed above implied an active role of the stratosphere in the development of extreme tropospheric weather events. But on the other hand, Polvani and Waugh (2004)[157] pointed out that the stratosphere itself is particularly forced by tropospheric dynamics and rather responds passively or acts as a referrer transferring initial tropospheric anomalies *via* the stratosphere back into the troposphere. Moreover, Fyfe *et al.* (1999)[143] and Gillett *et al.* (2002)[62] demonstrated that effects of increasing GHG concentrations can be simulated adequately in GCMs without stratospheric dynamics. Scaife *et al.* (2005)[158] showed that the IPCC climate projections of the 20$^{th}$ century indeed revealed a positive trend in the NAO—even in models with low stratospheric resolution—however, the magnitude of the observed NAO trend was underestimated by these climate models.

But there are many other numerical model studies showing a different picture. The influence of stratospheric variability on the lower atmosphere seems to be covered qualitatively better by stratosphere-resolving models, *i.e.* GCMs and CCMs containing the complete stratosphere. For instance, the importance of the stratosphere in determining adequately tropospheric climate change patterns was demonstrated by Stenchikov *et al.* (2006)[159] and Miller *et al.* (2006).[160] They concluded that the tropospheric climate response due to large volcanic eruptions was underestimated by ocean-atmosphere GCMs (AOGCMs, *i.e.* climate models), most of which do not resolve the stratosphere. Shindell *et al.* (1999, 2001)[136,144] already emphasised the important role of the stratosphere for simulating realistic tropospheric AO trends in their GCM. They demonstrated that including stratospheric dynamics strongly improved the simulated magnitude of the observed NAO trend between 1960 and 1990. This finding is consistent with investigation of Rind *et al.* (2005a;b)[112,113] who found a larger impact of a stratospheric forcing on the NAO than on the AO.

These results indicate that stratosphere-troposphere coupling may play a specific role for suitable assessments of tropospheric climate change patterns. But so far it is unclear whether the under-representation of the vertical coupling in the climate models is related to the missing stratospheric resolution, or if it is related to reduced stratospheric variability in climate models.

So far there is a clear tendency in the scientific community to support the point of view that stratospheric response to climate change plays a potentially important role for tropospheric climate.[129] Signs of stratospheric response to tropospheric climate change are therefore of major interest. Current projection studies with GCMs, climate models and CCMs, however, do not reveal a coherent picture of the stratospheric change, ranging from a projected increase of the stability of the stratospheric polar night jets in a future climate (due to further radiative cooling) to a projected decrease (due to enhanced tropospheric wave forcing see Subsection 8.3.2). Stratospheric ozone is projected to recover significantly in the first half of the 21st century, leading to a weakening of the polar vortex, while rising carbon dioxide levels are expected to counteract this process (see Chapter 9).

## 8.4.2 The Tropical and the Extra-tropical Tropopause Layer

The tropopause region both in the tropics and in the extra-tropical regions is of specific importance in understanding the coupling of the troposphere and the stratosphere. It is not only the tropopause itself (*i.e.* defined as the lowest level above which the lapse rate of temperature with height becomes less 2 K/km) which characterize the transition from the troposphere and the stratosphere, but also the closely adjacent layers which must be considered. The status of the tropical tropopause layer (TTL) and the extra-tropical tropopause layer (ExTL) both determine the exchange of air masses from the troposphere into the stratosphere and vice versa. Moreover, anomalies near the tropopause are highly correlated with tropical surface temperature anomalies and with tropopause level ozone anomalies, less so with stratospheric temperature anomalies.[161] Tropopause temperature anomalies are correlated with stratospheric water vapor concentrations.[5,27]

The TTL is usually set as the height region extending from the level of the temperature lapse rate minimum around 11–13 km[162,163] (see Fueglistaler *et al.* (2009)[164] for a review) to the level of highest convective overshoot, slightly above the cold point tropopause (CPT) at about 16–17 km. The very low temperatures (regularly below 200 K) experienced by air propagating upward through this part of the atmosphere play a crucial role for dehydration, and thus for stratospheric humidity. Changes which the TTL has undergone within the last few decades are not fully understood.[161,165]

Therefore, assessments of the future evolution of the TTL are a complex matter. The TTL is environed by a warming troposphere below and a cooling stratosphere above, which makes it difficult to estimate the response of CPT and stratospheric humidity changes. The appreciation of the future evolution of TTL temperatures is complicated by the fact that there is a tropospheric amplification of surface warming.[166] Conversely, a strengthening of the BD circulation, as discussed in Subsection 8.3.2, would imply a lowering of TTL temperatures. For example, Seidel *et al.* (2001)[167] obtained an increase in the height of the CPT of approximately 40 m and a decrease in pressure by about 1 hPa during the years from 1978 to 1997. Furthermore, Seidel *et al.* (2001)[167] and Zhou *et al.* (2001)[168] have detected a cooling of tropical tropopause temperatures of approximately 1 K during this time period, resulting in a decrease in the saturation volume mixing ratio of water vapor of about 0.5 ppmv. These temperature trends in the tropical tropopause region are therefore opposite to tropospheric warming dominating the response of the CPT. The CPT seems to be largely affected by increases in the BD circulation (here mainly by the tropical upwelling) and by increased convection as suggested by Zhou *et al.* (2001)[168] (see also Deckert and Dameris, 2008).[103]

CCMs are able to reproduce the basic dynamical and chemical structures of the TTL.[161,165] Although they are able to simulate the historical trends in tropopause pressure which are obtained from reanalysis data, trends in cold point tropopause temperatures are not consistent across CCMs and reanalyses. The altitude of the tropical tropopause has increased, and the level of main convective outflow appears to have decreased (just as water vapor concentrations) in historical CCM simulations as well as in reanalyses.

Changes in the TTL may not only affect the water vapor content of the stratosphere but also influence the abundance of many other species in the stratosphere. This concerns very short-lived chemical substances (VSLS) mostly of natural origin, such as biogenic bromine compounds that are carried to the stratosphere *via* deep convection followed by transport through the TTL[169,170] (Chapter 2 in WMO, 2007;[1] Chapter 1 in WMO, 2011).[2] Changes in deep convection may further affect the transport of longer-lived chemical substances such as methyl bromide and aerosols produced by biomass burning.[171] Moreover, chemical species may be transported in particulate form across the tropical tropopause, for example, organic sulphur-containing substances.[172]

Little is known about these processes affecting the transport in the TTL, and even less is known about climate induced changes. Given the uncertainties in our understanding of mechanisms in the TTL and of their previous changes, assessments of changes of the future evolution of TTL processes and transport through the TTL are still difficult. A future atmosphere with increasing GHG loadings is expected to develop a warmer troposphere with enhanced deep convection. But for the reasons mentioned above it remains speculative that this will be reflected in a warmer tropopause layer, higher water vapor mixing ratios in the stratosphere with less rapid recovery of ozone.[173] Analyses of the recent decades using observations in combination with CCM results suggest the dominance of other processes, possibly related to changes in the BD circulation.[161,165]

The ExTL is a layer of air adjacent to the local extra-tropical tropopause, which has been interpreted as the result of irreversible mixing of tropospheric air into the lowermost stratosphere[174,175] or as the result of two-way stratosphere-troposphere exchanges.[176,177] It is a global feature with increasing depth towards high latitudes, and has been found to be different for different tracers.[178] The origin of ozone in the ExTL changes markedly with season, with photochemical production dominating in summer and transport from the stratosphere dominating the winter and spring seasons. A general increase of the extra-tropical tropopause height in recent years has been identified by Steinbrecht *et al.* (1998)[179] and Varotsos *et al.* (2004)[180] who related this raising to changes of the ozone column. This long-term change of the tropopause height provides a very sensitive indicator of human effects on climate.[181–183] Changes in the ExTL determine the influence of the stratosphere on the troposphere through: (1) transport of ozone from the stratospheric into the troposphere, (2) UV fluxes,[57] and (3) radiative forcing of the surface climate.[70] Therefore, an accurate representation of dynamical and chemical processes in the upper troposphere and lower stratosphere (UTLS) in CCM is a necessary prerequisite for a robust prediction of the ozone layer and climate change. Hegglin *et al.* (2010)[184] found that the main dynamical and chemical climatological characteristics of the ExTL are generally well represented by most CCMs. Moreover, it is shown that the seasonality in the distribution of lower stratospheric chemical tracers is consistent with the seasonality in the BD circulation.

Possible future changes in BD circulation are all likely to change ozone concentrations, even in the UTLS. So far, limitations of the detailed knowledge on ExTL processes prevent robust assessments of the future evolution of the ExTL state. In particular, the relative contribution of isentropic (quasi-horizontal) and convective (vertical) transport and mixing of tropospheric air into the lowermost stratosphere is not well explained. For example, if the frequency or intensity of mid-latitudinal deep convection would change in a future warmer climate, this would affect the chemical composition of the lowermost stratosphere and, thus, the mid-latitudinal ozone layer. Based on simulations with climate models (AOGCMs) there are indications for decreases in the total number of deep convective events and of extra-tropical storms, but an increase of the mean strength of a single event and in the number of the most intense storms.[185,186] Therefore, more reliable future assessments of implications for ExTL dynamics and chemistry need further investigations.

### 8.4.3 Expected Future Changes

Numerical model studies indicate that climate change will impact the mass exchange across the tropopause. For instance, Rind *et al.* (2001)[85] estimated a 30% increase in the mass flux due to a doubling of atmospheric $CO_2$ concentrations, and Butchart and Scaife (2001)[31] estimated that the net upward mass flux above the TTL would increase by about 3% per decade due to climate change. In both studies, the changes in the mass flux resulted from more intensive wave propagation from the troposphere into the stratosphere. Modeling studies of tropospheric ozone[30,187,188] also found that climate change caused a comparable percentage increase in the extra-tropical stratosphere-to-troposphere ozone flux.

For a doubled $CO_2$ concentration, all 14 climate-change model simulations in Butchart *et al.* (2006)[53] resulted in an increase in the annual mean troposphere-to-stratosphere mass exchange rate, with a mean trend of 11 Gg $s^{-1}$ $year^{-1}$, or about 2% per decade. The predicted increase occurred throughout the year but was, on average, larger during the boreal winter than during the austral winter.

Butchart and colleagues were unable to conclude whether stratospheric ozone changes or ozone feedbacks had a significant impact on the underlying trend in the mass exchange rate. Other simulations[189] suggest that the trend in tropical upwelling is not constant. Periods (over several years) of enhanced upwelling coincide with periods of significant ozone depletion.

Butchart *et al.* (2010)[53] analyzed the response of stratospheric climate and circulation to increasing amounts of GHG concentrations and ozone recovery in the 21st century. Therefore, simulations of 11 CCMs were investigated using nearly identical forcings and experimental set-ups. Among others they found that on average the annual mean tropical upwelling in the lower stratosphere (at about 70 hPa) increases by almost 2% per decade. 59% of this trend was attributed to parameterised orographic gravity wave drag in the CCMs. They concluded that this is a consequence of the eastward acceleration of the subtropical jets which increases the upward flux of (parameterized) momentum reaching the lower stratosphere in these latitudes.

The majority of CCMs simulate continued decreasing of tropopause altitude and convective outflow pressure by several hPa/decade in the 21st century, along with an approximate 1 K increase per century in cold point tropopause temperature and 0.5–1 ppmv per century increase in water vapor mixing ratio above the tropical tropopause. These changes indicate significant perturbations to TTL processes in a future climate with enhanced GHG concentrations, in particular to deep convective heating and humidity transport.[165]

## 8.5 Concluding Remarks

It is obvious that understanding of long-term changes of the stratospheric ozone layer is a complex problem which makes robust assessments of its future evolution difficult. On the one hand, the modulation of stratospheric ozone concentrations is driven by natural variability, like solar irradiance and volcanic eruptions, and internal variability of stratospheric circulation on different time-scales affecting the stratospheric thermal structure and the transport of air masses. Ozone production and destruction is controlled by photochemical processes, homogeneous gas-phase reactions and heterogeneous chemistry on surfaces of particles (aerosols, PSCs). It must be considered that the chemical depletion of ozone in the presence of volcanic aerosols or PSCs is of nonlinear nature. On the other hand, the whole story becomes even more complex within a changing climate with enhanced GHG concentrations. Climate change influences net ozone production (*i.e.* sum of ozone destruction and production) both in direct and indirect ways and, therefore, will affect the rate of ozone recovery, which will be different at various altitudes and latitudes. Cooling of the stratosphere due to enhanced GHG concentrations has opposite effects in the upper and lower stratosphere, slowing down the gas-phase ozone loss rate but increasing the heterogeneous ozone loss rate on PSCs. This will accelerate ozone recovery in the upper stratosphere and delay it in the lower stratosphere. Moreover, changes in the stratospheric circulation have the potential to modify the future evolution of the stratospheric ozone layer in the 21st century. For example, it is known that the strength of the BD circulation is directly related to dissipating planetary waves, which are forced in the troposphere, *i.e.* stronger wave forcing coincides with a weaker polar night jet and higher polar temperatures. Furthermore, circulation modes can affect the ozone distribution in the UTLS both directly and indirectly by influencing propagation of planetary waves from the troposphere into the stratosphere. Future changes in the generation of tropospheric waves and circulation modes will influence polar ozone abundance dynamically. Nevertheless, so far there is no consensus from numerical model studies on the sign of this change, making assessments of the rate of ozone recovery uncertain. Generally, a better understanding of stratosphere-troposphere coupling is a key issue for more reliable assessments of future climate change and recovery of the stratospheric ozone layer. Warming and expansion of the tropopause region in future climate could additionally obscure ozone recovery rates as the inverse relation between total ozone and tropopause height seems to hold for long time scales as well.[179,190]

Additionally, future changes of stratospheric water vapor concentrations are uncertain. Chemistry-climate models predict increases of stratospheric water vapor, but confidence in these predictions is low, because these models both have a poor representation of the seasonal cycle in tropical tropopause temperatures (which control global stratospheric water vapor abundances) and cannot reproduce past changes in stratospheric water vapour abundances;[5] (Chapter 4 in WMO, 2011).[2] In a warmer climate, numerical model studies suggest an increase in water vapor outflow to the tropical lower stratosphere. Stratospheric water vapor concentrations may also increase through enhanced methane ($CH_4$) concentrations. On the other hand, higher $CH_4$ concentration would remove reactive chlorine, particularly in the upper stratosphere. Numerical modeling studies suggest that increased water vapor concentrations will enhance odd hydrogen ($HO_X$) in the stratosphere and subsequently increase ozone depletion.[191] Increases in water vapor concentrations in the polar regions would raise the formation of PSCs, potentially increasing springtime ozone depletion. Moreover, if water vapor concentrations would increase in the future, there will be also radiative effects.

An increase in nitrous oxide ($N_2O$) emission (from extended use of artificial fertilizer) will enhance the amount of stratospheric nitrogen oxides ($NO_X$). This is expected to reduce ozone in the middle and high stratosphere which would make ozone destruction even worse. $N_2O$ will probably remain the largest ozone-depleting emission for the rest of the century.[192] Also, changes in $NO_x$ and non-methane hydrocarbon emissions are expected to affect the tropospheric concentrations of the hydroxyl radical (OH) and, hence, impact the lifetimes and concentrations of stratospheric trace gases such as $CH_4$ and organic halogen species.

Future climate change will seriously affect the amount of stratospheric ozone mainly through enhanced GHG concentrations, leading to a cooling of the stratosphere and changes in stratospheric circulation. Beside carbon dioxide changes in stratospheric concentrations of water vapor, methane or nitrous oxide must be also taken into account while influencing ozone chemistry and radiative effects in the stratosphere. Although large uncertainties exist, especially in vertical wave propagation into the stratosphere and stratospheric dynamics, the current consensus view is that the rate of ozone recovery will be accelerated by climate change in most parts of the stratosphere except the polar lower stratosphere in winter and spring (Chapter 9).

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