A quantitative analysis of the reactions involved in stratospheric ozone depletion in the polar vortex core

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Abstract. We present a quantitative analysis of the chemical reactions involved in polar ozone depletion in the stratosphere, and of the relevant reaction pathways and cycles. While the reactions involved in polar ozone depletion are well known, quantitative estimates of the importance of individual reactions or reaction cycles are rare. In particular, there is no comprehensive and quantitative study of the reaction rates and cycles averaged over the polar vortex under conditions of heterogeneous chemistry so far. We show time series of reaction rates averaged over the core of the polar vortex in winter and spring for all relevant reactions and indicate which reaction pathways and cycles are responsible for the vortex-averaged net change of the key species involved in ozone depletion, that is ozone, chlorine species (ClOₓ, HCl, ClONO₂), bromine species, nitrogen species (HNO₃, NOₓ) and hydrogen species (HOₓ). For clarity, we focus on one Arctic winter (2004/2005) and one Antarctic winter (2006) in a layer in the lower stratosphere around 54 hPa, and show results for additional pressure levels and winters in a supplement.

1 Introduction

After the discovery of the ozone hole (Farman et al., 1985), the chemistry of polar ozone depletion in the stratosphere has been the subject of ongoing research for the last 30 years (see e.g. articles, review papers and textbooks by Solomon et al., 1986, Wayne et al., 1995, Portmann et al., 1996, Brasseur et al., 1999, Solomon, 1999, Brasseur and Solomon, 2005, WMO, 2011, Müller, 2012, Solomon et al., 2015). In general, the chemistry of polar ozone depletion is understood very well (see e.g. the recent overview in Müller, 2012, or von Hobe et al., 2013). Remaining issues, such as uncertainties in the formation pathways of Polar Stratospheric Clouds (PSCs) or uncertainties in the contribution of the different cloud types to chlorine activation (e.g. Lowe and MacKenzie, 2008; Peter and Grooß, 2012; Wohltmann et al., 2013), do not pose a serious challenge to the generally accepted basic theory.

While the reaction pathways and reaction cycles that are involved in ozone depletion are well known (e.g. Portmann et al., 1996; Solomon, 1999; Müller, 2012), quantitative estimates of the importance of single reactions or reaction cycles are rare, and are limited to case studies or certain aspects of the chemical system (e.g. Portmann et al., 1996; Grenfell et al., 2006; Frieler et al., 2006) or apply mainly to conditions undisturbed by heterogeneous chemistry (e.g. Brasseur and Solomon, 2005).
Here, we give a comprehensive overview of the temporal evolution of the vortex-averaged reaction rates and mixing ratios and associated reaction pathways and cycles for one Arctic winter (2004/2005) and one Antarctic winter (2006) in a layer in the lower stratosphere around 54 hPa. Results for additional winters and pressure levels can be found in a supplement. Vortex-averaged mixing ratios and reaction rates are obtained from runs of the Lagrangian (trajectory-based) ATLAS Chemistry and Transport Model (Wohltmann and Rex, 2009; Wohltmann et al., 2010). Since results cannot be based on direct observations due to a lack of measurements of the mixing ratios of minor species and reaction rates, only a model-based approach is feasible. The most important model parameters that influence the vortex-averaged rates are the initial mixing ratios, the laboratory measurements of the rate coefficients of the reactions (taken from Sander et al., 2011) and the meteorological data that drive the model, which are taken from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA Interim reanalysis (Dee et al., 2011). A detailed study of the uncertainties is outside the scope of this paper, but some of the major uncertainties will be noted. For a study of the uncertainties, see e.g. Kawa et al. (2009).

Reasonable agreement of the modeled and observed mixing ratios for many species gives us confidence that our results represent the real atmosphere well in most cases (see the basic validation against observations in Section 6 and extensive additional comparisons of the model to measurements in the supplement). There was however a significant overestimation of HCl compared to measurements in our original model runs, which has also been observed in other models like SD-WACCM (Brakebusch et al., 2013; Wegner, 2013; Solomon et al., 2015) or MIMOSA-CHIM (Kuttippurath et al., 2015). Interestingly, the SLIMCAT CTM shows a discrepancy of the same order of magnitude, but with opposite sign in a recent publication (Santee et al., 2008).

Hence, we consider the differences in HCl between models and observations as “state-of-the-art” in current CTMs and apply an empirical correction to bring the HCl mixing ratios in closer agreement to observations. The correction is based on changing the HCl solubility, which is a possible cause for this discrepancy. This introduces some uncertainty in our results, which are explored in Section 6 and by comparisons to the uncorrected runs shown in the supplement. However, a detailed discussion of this important issue would deserve its own study.

While it is easy to identify the gross production or loss of a species by single reactions, it is difficult to identify what causes the net loss or production of a species (e.g. Lehmann, 2002). Often, the reactions that directly produce or remove a species of interest are part of a complicated chain of reactions, frequently involving reaction cycles. Many of these cycles will consume as much of a species as they produce (null cycles, equilibria), others will lead to a net loss of the species (loss cycles). Numerous loss cycles in the stratosphere are catalytic cycles, i.e. they destroy the species of interest with the help of another species that is constantly recycled in the loss cycle. To complicate things further, cycles may share species and reactions. Hence, it is of crucial importance to know the reaction cycles and pathways in the chemical system to actually understand the underlying reasons for the loss or production of a species, and these relationship are discussed in the following.

In Section 2, the methods and the model are introduced. The section contains a description of the ATLAS model and the setup of the runs and explains the method to obtain vortex-averaged reaction rates. In Section 3, we give an overview of the typical evolution of polar ozone chemistry in winter to provide a foundation for the following discussion. Section 4 discusses reactions and partitioning sorted by chemical families, that is nitrogen, hydrogen, chlorine, bromine and oxygen compounds.
Section 5 explores in how far the results can be generalized to other years. Section 6 discusses the HCl discrepancy and gives a short discussion of validation of other species with observations. Section 7 contains the conclusions.

Results of this study are extensively used in a companion paper (Wohltmann et al., 2017) to develop a fast model for polar ozone chemistry.

5 2 Model and methods

2.1 Model overview

ATLAS is a global Chemistry and Transport Model (CTM) based on a Lagrangian (trajectory-based) approach. A detailed description of the model can be found in Wohltmann and Rex (2009) and Wohltmann et al. (2010). Updates to the chemistry module and polar stratospheric cloud module are described in Wohltmann et al. (2013). The model includes a gas-phase stratospheric chemistry module, heterogeneous chemistry on polar stratospheric clouds, a particle-based Lagrangian denitrification module and a dehydration parameterization. The chemistry module comprises 47 active species and more than 180 reactions. Absorption cross sections and rate coefficients are taken from recent JPL recommendations (Sander et al., 2011). In addition to the binary background aerosol, the model simulates three types of Polar Stratospheric Clouds, that is supercooled ternary HNO$_3$/H$_2$SO$_4$/H$_2$O solutions (STS), solid clouds composed of nitric acid trihydrate (NAT), and solid ice clouds.

2.2 Model setup

Model runs are driven by meteorological data from the ECMWF ERA Interim reanalysis (Dee et al., 2011). The initial horizontal model resolution is 150 km. The runs use the hybrid pressure-potential temperature coordinate of the model, which is a pure potential temperature coordinate above 100 hPa. The vertical range of the model domain is 350 K to 1900 K. Vertical motion is driven by diabatic heating rates from ERA Interim. Two model runs are started, one for the northern hemispheric winter, and one for the southern hemispheric winter. The run for the northern hemisphere starts on 1 October 2004 and ends on 31 March 2005. The run for the southern hemisphere starts on 1 April 2006 and ends on 30 November 2006. Model data before 15 November 2004 or 1 May 2006 are not used for analysis to allow for a spin up of the mixing in the model and for a sufficiently stable vortex in the northern hemisphere. Results for two additional winters (2009/2010 in the northern hemisphere and 2011 in the southern hemisphere) are only shown in the supplement.

The number density of NAT particles in the runs is set to 0.1 cm$^{-3}$, the number density of ice particles is set to 0.01 cm$^{-3}$ and the number density of the ternary solution droplets is set to 10 cm$^{-3}$. A supersaturation of HNO$_3$ over NAT of 10 (corresponding to about 3 K supercooling) is assumed to be necessary for the formation of the NAT particles. A detailed discussion of the rationale behind these choices can be found in Wohltmann et al. (2013). For ice particles, a supersaturation of 0.35 is assumed based on Microwave Limb Sounder (MLS) satellite measurements of H$_2$O (Waters et al., 2006) and ECMWF temperatures.
The treatment of conditions where both NAT and STS clouds are allowed to form has changed compared to Wohltmann et al. (2013). Since mixed NAT/STS clouds are commonly observed (e.g. Pitts et al., 2011), they can now form in the model to allow for a more realistic behaviour, see Nakajima et al. (2016) for details.

The settings for the polar stratospheric cloud parameterizations largely favor the formation of liquid clouds (binary liquids and STS clouds) over the formation of NAT clouds and activation of chlorine predominantly occurs on liquid clouds in the model runs.

The Lagrangian particle model is used to simulate the nucleation, growth, sedimentation and evaporation of large NAT particles. These particles are formed with a nucleation rate of \(7.8 \times 10^{-6}\) particles per h and cm\(^3\) and an initial radius of 0.1 µm, wherever a supersaturation of 10 for HNO\(_3\) is exceeded. Dehydration by falling ice particles is simulated by a simple algorithm that irreversibly removes all ice above a given supersaturation, which is set to 0.7 here (note that the value 0.35 given above is only used in the chemistry module, the value in the dehydration module is set separately).

H\(_2\)O, N\(_2\)O, HCl, O\(_3\), CO and HNO\(_3\) are initialized from all measurements of the MLS instrument performed during 1 November 2004 and 1 May 2006 for the northern and southern hemispheres, respectively. CH\(_4\) is initialized from a monthly mean HALOE (Halogen Occultation Experiment) climatology (mean of the years 1991–2002) as a function of equivalent latitude and pressure (Grooß and Russell III, 2005). NO\(_x\) is initialized from the monthly mean HALOE data set by putting all NO\(_x\) into NO\(_2\). ClONO\(_2\) is calculated as the difference between Cl\(_y\) and HCl. Cl\(_y\) is taken from a Cl\(_y\)-N\(_2\)O tracer-tracer correlation from ER-2 aircraft and Triple balloon data (Grooß et al., 2002). As in Wohltmann et al. (2013), we increase the amount of ClONO\(_2\) by 10 % at the expense of HCl, see the discussion there. BrONO\(_2\) is assumed to contain all Br\(_y\), which is taken from a Br\(_y\)-CH\(_4\) relationship from ER-2 aircraft and Triple balloon data in Grooß et al. (2002). All Br\(_y\) values are scaled with a constant factor to give maximum values of 19.9 ppt.

2.3 Production and loss rates

Reaction rates are calculated for every reaction separately in the ATLAS model. For this purpose, one artificial species per reaction is introduced to the model, which is produced at the same rate as the other products of the reaction. For instance, a reaction of the type

\[A + B \rightarrow C + D\]

is modified to

\[A + B \rightarrow C + D + P\]

where P is an artificial product species. The mixing ratio of the artificial product species is reset to zero every 24 hours. This way, P directly gives the 24 h averaged rates of production of the species C and D by this reaction and the 24 h averaged loss rates of the species A and B. The 24 h time period is used to capture the diurnal cycle of the photochemically active species. For technical reasons, heterogeneous reaction rates with the same chemical equation, but on different surface types are added together in a single reaction rate.
2.4 Vortex averages

Production and loss rates are averaged over all air parcels of the model inside the polar vortex that are situated in a layer between 61.3 hPa and 47.4 hPa. The logarithmic mean level of this layer is at 54 hPa. The vortex edge is assumed to be situated at the 36 PVU contour of modified potential vorticity (PV) in the northern hemisphere and at the −36 PVU contour in the southern hemisphere. Modified PV is calculated from the potential vorticity field of the ERA Interim reanalysis according to Lait (1994), with \( \theta_0 = 475 \) K. The supplement shows results for the additional pressure levels 32 hPa, 42 hPa and 70 hPa.

In order to obtain a more consistent picture of the ozone chemistry in the vortex, we exclude air parcels that experience a too high amount of mixing with extra-vortex air during the course of the model run. This basically limits our results to the core of the vortex. For that purpose, we initialize a “vortex tracer” as an artificial chemical species near the start of the model run (15 November in the northern hemisphere, 1 May in the southern hemisphere), which is set to 1 inside the vortex and to 0 outside the vortex. The vortex tracer is then transported and mixed like any other species in the model and can take any value between 0 and 1. We only include in the vortex mean those parcels for which the vortex tracer has a value greater than 0.7. This value was chosen as a compromise between obtaining a sufficient number of trajectories for averaging and a set of trajectories that is chemically sufficiently homogenous.

Without the vortex tracer, analysis would get much more complicated. An example may illustrate this: In the southern vortex, air masses rich in NO\(_y\) are mixed into the edge region of the vortex during the course of the winter, while in the core of the vortex, air masses are depleted of nitrogen species due to denitrification. This leads not only to differences in mixing ratios of the nitrogen species over the vortex, but also to different reactions being important in different parts of the vortex. For example, while very low ozone values are reached in the core of the vortex since deactivation into ClONO\(_2\) is hindered, ozone values at the edge are higher and chlorine is also deactivated into ClONO\(_2\) (ClONO\(_2\) “collar”, see e.g. Douglass et al., 1995).

2.5 Different air masses

The 54 hPa level is not a material surface and we look at different air masses at different points of time, due to mixing, the movement of the isentropes relative to the pressure level and the additional sinking of the air masses relative to isentropes due to diabatic cooling. A similar caveat applies to the definition of the vortex edge. As a consequence of these transport effects, the temporal derivative of the vortex-averaged mixing ratio of a chemical species may deviate from the vortex-averaged chemical net change of this species.

Unfortunately, it is difficult to look at the same air mass over the course of several months, since an air mass with a well defined extent at the beginning of the winter will completely lose its identity due to mixing and transport during the course of the winter.
2.6 Equilibria

In many cases, two species X and Y with short lifetimes are in a fast equilibrium with each other, and will not change their mixing ratios if the external conditions do not change. For example, consider two fast reactions of the form

\[ X + Z_1 \rightarrow Y + \ldots \]

\[ Y + Z_2 \rightarrow X + \ldots \]

Then, assuming that the change of the mixing ratios is zero

\[
\frac{d[X]}{dt} = -k_1[X][Z_1] + k_2[Y][Z_2] = 0
\]

(1)

where \([X]\) is the concentration of X and \(k_1\) and \(k_2\) are the reaction coefficients, we obtain equilibrium conditions like

\[
\frac{[X]}{[Y]} = \frac{k_2[Z_2]}{k_1[Z_1]}
\]

(2)

Similar equations can be derived for more complex reaction systems. We use the equilibrium conditions in the following not only to show relationships between mixing ratios of different species, but also to determine the reactions involved in an equilibrium: For a given set of species, we start with a small set of reactions involved in the equilibrium derived from the vortex-averaged reaction rates, and calculate the mixing ratios of the species from the equilibrium conditions. Then, we compare these mixing ratios with the mixing ratios in the output of the model. If the mixing ratios determined by the two methods do not agree with each other, we add reactions until we reach good agreement with a set of reactions as small as possible.

3 Short overview of the chemical evolution

The evolution of the chemistry of polar ozone depletion can be divided into several phases (first defined by Portmann et al., 1996). For orientation, Figure 1 shows the evolution of temperature and sunlight in both hemispheres.

After the polar vortex forms in late autumn and early winter in response to the cooling of the atmosphere in the beginning of polar night, air masses in the vortex are well isolated from mid latitudes (e.g. Solomon, 1999). In the first phase in early winter, chlorine, the main player in the chemistry of lower stratospheric ozone depletion, is present mainly in the form of species that do not contribute to ozone depletion. These passive reservoir gases are mainly HCl and ClONO\(_2\) (e.g. Solomon, 1999). Chemical activity is low due to the lack of sunlight in the polar night. This period lasts from the forming of the vortex until the first polar stratospheric clouds form.

In the second phase, HCl and ClONO\(_2\) are transformed from passive reservoir gases to Cl\(_2\) through heterogeneous reactions on the surface of polar stratospheric clouds (e.g. Solomon et al., 1986), which condense when temperatures in the polar night get cold enough. A second effect of the clouds can be the removal of large quantities of HNO\(_3\) and H\(_2\)O by sedimentation (denitrification and dehydration, e.g. Toon et al., 1986, Fahey et al., 1990), which can prolong ozone loss later in spring (e.g. Portmann et al., 1996). In the model setup used here, activation predominantly occurs on liquid STS clouds and only to a lesser
Figure 1. Vortex-averaged temperature (blue), vortex minimum temperature (dashed blue) and fraction of the vortex in sunlight (yellow) for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. The upper thin blue line shows the threshold temperature for the formation of NAT clouds used in the model (based on vortex mean mixing ratios and considering supersaturation), and the lower thin blue line shows the same for ice clouds. The vortex tracer criterion described in the text was not applied (in contrast to all other figures).

Part on solid NAT clouds. Unfortunately, observations of PSCs are not detailed enough to sufficiently constrain the ratio of activation on STS versus NAT clouds, and there is also uncertainty in other parameters like size distribution, number densities and required supersaturation. However, chlorine activation and ozone loss are robust quantities with respect to most changes in PSC parameterizations (Wohltmann et al., 2013).

The third phase starts when sunlight comes back, enabling catalytic cycles to destroy ozone in large quantities. Cl₂ is transformed to Cl and ClO by photolysis. Then ozone is removed mainly by the catalytic ClO dimer cycle and the catalytic ClO–BrO cycle (e.g. Solomon, 1999). The second cycle makes ozone destruction sensitive to the amount of bromine and to bromine chemistry. At the same time, reactions that deactivate active chlorine and that are dependent on sunlight gain importance. Hence, there is a constant competition between activation and deactivation in this phase, as long as it is cold enough (e.g. Portmann et al., 1996; Solomon et al., 2015).

In the fourth phase, when temperatures rise in spring and PSCs evaporate, chlorine is deactivated into the reservoir gases again and ozone loss significantly decreases.

4 Budgets of the chemical families and chemical evolution arranged by families

4.1 Nitrogen species

NO_y is defined as the sum of all nitrogen containing species except for the long-lived source gas N₂O, which is the source of all other nitrogen species in the stratosphere. N₂O has both natural and anthropogenic sources in the troposphere (e.g. WMO, 2011; Montzka, 2012). Figure 2 shows the partitioning of NO_y at 54 hPa. The majority of NO_y is in the form of HNO₃ in
Figure 2. Vortex-averaged partitioning of \textit{NO}_y species for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Species \textit{NO}_3, BrONO\textsubscript{2}, ClONO\textsubscript{2} and N are not shown due to their small mixing ratios.

the considered altitude range. This is due to the fact that removal of HNO\textsubscript{3} by photolysis and OH is not very efficient at these altitudes. The initial level of HNO\textsubscript{3} is about 13 to 14 ppb in both hemispheres and the initial \textit{NO}_y is about 1 ppb higher. Figure 2 shows that the mixing ratio of HNO\textsubscript{3} (and \textit{NO}_y) declines to about 3 ppb in February 2005 in the northern hemisphere and increases again in March due to transport (see Figure 3, which shows that the rate of change by chemistry is small in this time period). In contrast, HNO\textsubscript{3} decreases to about 0.5 ppb (with \textit{NO}_y at 1 ppb) in the southern hemisphere after June 2006. The rate of change of HNO\textsubscript{3} is not dominated by chemical changes, but by denitrification, i.e. the irreversible removal of HNO\textsubscript{3} by sedimenting cloud particles (e.g. Toon et al., 1986; Fahey et al., 1990), as shown in Figure 3. Denitrification is much more severe in the southern hemisphere due to the lower temperatures (see Figure 1, e.g. Solomon, 1999), leaving almost no \textit{NO}_y. The amount of \textit{NO}_y present in the other important nitrogen reservoir ClONO\textsubscript{2} is limited by the mixing ratio of Cl\textsubscript{v} (less than 3.5 ppb). Thus, ClONO\textsubscript{2} never contributes more than about 25% to \textit{NO}_y. The supplement shows that \textit{NO}_y increases with height from 10 ppb at 70 hPa to 18 ppb at 32 hPa, but that the majority is always in the form of HNO\textsubscript{3}.

\textit{NO}_x is defined as the sum of the short-lived and reactive species NO, NO\textsubscript{2}, NO\textsubscript{3} and 2 N\textsubscript{2}O\textsubscript{5}, which only form under sunlit conditions from the longer lived HNO\textsubscript{3}. Only a small part of \textit{NO}_y is in the form of \textit{NO}_x under sunlit conditions, typically less than 1.5 ppb in the northern hemisphere and less than 0.5 ppb in the southern hemisphere at 54 hPa. The supplement shows that \textit{NO}_x increases from about 0.5 ppb at 70 hPa to 2.5 ppb at 32 hPa due to increasing radiation. For the following discussion, it is reasonable to define an “extended” \textit{NO}_x, which also includes ClONO\textsubscript{2} (and, much less importantly, BrONO\textsubscript{2} and HO\textsubscript{2}NO\textsubscript{2}), since these species are in very fast equilibria with the “classical” \textit{NO}_x species under sunlit conditions.

The extended \textit{NO}_x mixing ratios can be changed by a number of reactions. Vortex-averaged reaction rates of these reactions for the northern and southern hemisphere are shown in Figure 4. Extended \textit{NO}_x is almost exclusively produced from HNO\textsubscript{3}
Figure 3. Vortex-averaged net chemical reaction rate of HNO₃ (red) for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa and sum of the vortex-averaged change by sedimentation and the net chemical reaction rate (blue).

Figure 4. Vortex-averaged chemical reaction rates of reactions changing extended NOₓ (NO + NO₂ + NO₃ + 2N₂O₅ + ClONO₂ + BrONO₂ + HO₂NO₂) for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative. The net change of extended NOₓ is shown as a green line. Reactions with rates which cannot be distinguished from the zero line at plot resolution are not shown.
under sunlit conditions by photolysis and reaction with $\text{OH}$

$$\text{HNO}_3 + h\nu \rightarrow \text{NO}_2 + \text{OH} \quad (\text{R1})$$

$$\text{HNO}_3 + \text{OH} \rightarrow \text{H}_2\text{O} + \text{NO}_3 \quad (\text{R2})$$

see e.g. Portmann et al. (1996). The OH reaction contributes about 60%–80% of the production (cf. Figure 7 in Portmann et al., 1996). Extended NO$_x$ is lost to HNO$_3$ mainly by the heterogeneous reactions

$$\text{N}_2\text{O}_5 + \text{H}_2\text{O} \rightarrow \text{HNO}_3 + \text{HNO}_3 \quad (\text{R3})$$

$$\text{ClONO}_2 + \text{HCl} \rightarrow \text{Cl}_2 + \text{HNO}_3 \quad (\text{R4})$$

and by the gas-phase reaction

$$\text{OH} + \text{NO}_2 + \text{M} \rightarrow \text{HNO}_3 + \text{M}. \quad (\text{R5})$$

The reaction $\text{N}_2\text{O}_5 + \text{H}_2\text{O}$ proceeds on the background aerosol, not only when polar stratospheric clouds are present. When sunlight comes back to the vortex in spring, reactions R1–R2 are not in equilibrium with R3–R5 and net production of extended NO$_x$ from HNO$_3$ occurs (e.g. Portmann et al., 1996). Due to the denitrified conditions in the southern hemisphere, much less NO$_x$ is produced there (e.g. Douglass et al., 1995). While the rates of the reactions changing extended NO$_x$ increase by a factor of 4 from 70 hPa to 32 hPa (see supplement), the relative importance of the reactions does not change.

Note that the introduction of an extended NO$_x$ does not work very well under polar night conditions, since ClONO$_2$ is not in equilibrium with the classical NO$_x$ then, which can be seen in the fact that extended NO$_x$ is lost in polar night due to heterogeneous reaction of ClONO$_2$, although no NO or NO$_2$ is present.

Figure 5 shows the partitioning of extended NO$_x$ during daytime and at night. Daytime averages are defined over the parts of the vortex where the solar zenith angle is smaller than 80° and nighttime averages are defined over parts of the vortex where the solar zenith angle is larger than 100° (note that this means that the areas for the daytime average and the nighttime average may not be coincident). Except for early winter, NO$_x$ proper is only present in appreciable quantities after the start of March 2005 or October 2006 in the northern and southern hemispheres, respectively. During daytime, the partitioning between NO and NO$_2$ is so fast that steady state conditions can be assumed (e.g. Douglass et al., 1995). The three reactions that determine the equilibrium are

$$\text{O}_3 + \text{NO} \rightarrow \text{NO}_2 + \text{O}_2 \quad (\text{R6})$$

$$\text{ClO} + \text{NO} \rightarrow \text{Cl} + \text{NO}_2 \quad (\text{R7})$$
Figure 5. Vortex-averaged partitioning of extended NO$_x$ species for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Top row: Daytime averages (parts of the vortex where the solar zenith angle is smaller than 80°). Bottom row: Nighttime averages (parts of the vortex where the solar zenith angle is larger than 100°). Days without sufficient data for averaging are not shown (grey bars).
\[
\text{NO}_2 + h\nu \rightarrow \text{NO} + \text{O} \quad \text{(R8)}
\]

Figure 6 shows the formation and loss rates of NO to illustrate that these are the relevant reactions (the corresponding plots for NO\textsubscript{2} look identical, but mirrored). The equilibrium condition derived from these reactions is

\[
\frac{[\text{NO}]}{[\text{NO}_2]} = \frac{k_{R8}}{k_{R6}[\text{O}_3] + k_{R7}[\text{ClO}]}.
\quad \text{(3)}
\]

In the southern hemisphere during daytime, the fraction of NO compared to NO\textsubscript{2} is much higher than in the northern hemisphere (80\%–90\% NO and 10\%–20\% NO\textsubscript{2} in the southern hemisphere, and 20\%–40\% NO and 60\%–80\% NO\textsubscript{2} in the northern hemisphere, see Figure 5 and percentage plots in the supplement). The relative partitioning is approximately constant between 70 hPa and 32 hPa (see supplement). The higher NO levels are caused by the much lower ozone levels in the southern hemisphere (see Figure 19), which shifts the equilibrium by hindering the \text{O}_3 + \text{NO} reaction (e.g. Douglass et al., 1995). At night, no NO is present, since the reaction \text{NO}_2 + h\nu does not take place. NO\textsubscript{3} does not significantly contribute to the budget of NO\textsubscript{x} due to the fast reaction into NO\textsubscript{2} by

\[
\text{NO}_3 + h\nu \rightarrow \text{NO}_2 + \text{O} \quad \text{(R9)}
\]

during daytime and into \text{N}_2\text{O}_5 at night (see discussion of R13).

Despite the relatively low mixing ratios, NO\textsubscript{x} plays an important role in the chemistry of polar ozone depletion (the gas phase catalytic NO\textsubscript{x} cycle is in general an important contributor to stratospheric ozone chemistry, e.g. outside of the vortex and in higher altitudes than discussed here). In addition to the equilibrium between NO and NO\textsubscript{2}, there is a fast equilibrium
between NO$_2$ and ClONO$_2$ under sunlit conditions, so that all three species are coupled. ClONO$_2$ is to a good approximation in an equilibrium between

$$\text{ClONO}_2 + h\nu \rightarrow \text{Cl} + \text{NO}_3$$
$$\rightarrow \text{ClO} + \text{NO}_2$$ \hfill (R10)\hfill (R11)

and

$$\text{ClO} + \text{NO}_2 + \text{M} \rightarrow \text{ClONO}_2 + \text{M}$$ \hfill (R12)

see e.g. Portmann et al. (1996). Reaction R10 is the dominant branch of the ClONO$_2$ photolysis. The equilibrium condition for ClONO$_2$ can be written as

$$[\text{ClO}][\text{NO}_2] = \frac{k_{R10} + k_{R11}}{k_{R12}}[\text{ClONO}_2].$$ \hfill (4)

Alternatively, since NO and NO$_2$ are in equilibrium, this can also be expressed in terms of NO

$$[\text{ClO}][\text{NO}] = \frac{(k_{R10} + k_{R11})k_{R8}[\text{ClONO}_2]}{k_{R12}(k_{R6}[\text{O}_3] + k_{R7}[\text{ClO}])}.$$ \hfill (5)

Production of NO$_x$ from HNO$_3$ in spring will increase NO$_2$. In turn, ClONO$_2$ will increase almost instantly at the expense of NO$_2$ to match the equilibrium condition again. In this sense, ClONO$_2$ is produced from HNO$_3$ via NO$_x$ in spring (e.g. Portmann et al., 1996). This is an important deactivation pathway for active chlorine in the northern hemisphere, since it consumes ClO.

There is a striking difference in ClONO$_2$ production in spring between the northern and southern hemisphere. While ClONO$_2$ increases to 2 ppb in the northern hemisphere, it stays below 0.1 ppb in the southern hemisphere due to the strongly denitrified conditions there. The hindrance of the deactivation path via ClONO$_2$ under denitrified conditions prolongs the period of ozone loss, since the other deactivation path via the reaction Cl + CH$_4$ is only effective under low ozone levels (e.g. Douglass et al., 1995, Portmann et al., 1996), see Section 4.3.2. ClONO$_2$ stays relatively constant in March 2005 in the northern hemisphere after the initial increase, since the decrease in ClO is compensated by an increase in NO$_2$.

The mixing ratio of N$_2$O$_5$ is governed by

$$\text{NO}_2 + \text{NO}_3 + \text{M} \rightarrow \text{N}_2\text{O}_5 + \text{M}$$ \hfill (R13)

$$\text{N}_2\text{O}_5 + h\nu \rightarrow \text{NO}_2 + \text{NO}_3$$ \hfill (R14)

$$\text{N}_2\text{O}_5 + \text{H}_2\text{O} \rightarrow \text{HNO}_3 + \text{HNO}_3.$$ \hfill (R3)

At night N$_2$O$_5$ is produced slowly by reaction R13, with NO$_3$ produced by

$$\text{NO}_2 + \text{O}_3 \rightarrow \text{NO}_3 + \text{O}_2.$$ \hfill (R15)
During daytime, no N$_2$O$_5$ is produced, since the NO$_3$ produced by reactions R10 and R15 is easily photolyzed. However, N$_2$O$_5$ is only slowly depleted during daytime, so that N$_2$O$_5$ levels during daytime and at night are comparable in the northern hemisphere (up to 30% of classical NO$_x$ is in N$_2$O$_5$). In the southern hemisphere, low ozone levels hinder the production of N$_2$O$_5$. The produced N$_2$O$_5$ can react back to the reservoir HNO$_3$ via reaction R3 (see Figure 4).

5  4.2  Hydrogen species

The sources for HO$_x$ = OH + HO$_2$ are mainly H$_2$O, CH$_4$ and HNO$_3$. The source for stratospheric water is humid tropospheric air “freeze-dried” at the tropopause and sources for CH$_4$ are both natural (e.g. wetlands) and anthropogenic (e.g. WMO, 2011; Montzka, 2012). CH$_4$ is slowly oxidized to H$_2$O in the stratosphere (e.g. Hanisco, 2003, see also below). Figure 7 shows that the mixing ratio for H$_2$O is between 2 ppm and 6 ppm at 54 hPa and that CH$_4$ has a mixing ratio of about 1 ppm. The decrease of H$_2$O in July 2006 in the southern hemisphere is caused by dehydration by sedimenting cloud particles, similar to the situation for HNO$_3$. The northern hemisphere in 2005 is not cold enough for the formation of a significant amount of ice clouds (Figure 1).

Production and loss processes of HO$_x$ are fairly complicated (Hanisco, 2003). Figure 8 shows the production and loss rates of an extended HO$_x$ at 54 hPa, where we have included some species that are not a net source or sink of HO$_x$ over a diurnal cycle (extended HO$_x$ = OH + HO$_2$ + H + HOC$_l$ + HOBr + HO$_2$NO$_2$). Particularly in the southern hemisphere, production from CH$_4$ oxidation, which can be initiated by

$$X + CH_4 \rightarrow XH + CH_3$$  \hspace{1cm} (R16)

with X = Cl, O($^1$D), OH plays an important role. It then continues with a complicated chain of reactions involving CH$_2$O (see Hanisco, 2003, for more details). The maximum yield of this reaction chain is 4 HO$_x$ per CH$_4$, but the yield is normally
Figure 8. Vortex-averaged chemical reaction rates of reactions changing extended HO\textsubscript{x} (OH + HO\textsubscript{2} + H + HOCl + HOBr + HO\textsubscript{2}NO\textsubscript{2}) for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative. The net change of extended HO\textsubscript{x} is shown as a green line. Methane oxidation is modelled by simplified net reactions in ATLAS, the reactions denoted as methane oxidation in the legend are Cl + CH\textsubscript{4} → HCl + CH\textsubscript{2}O + HO\textsubscript{2} and Cl + CH\textsubscript{2}O → HCl + CO + HO\textsubscript{2}.

lower (Hanisco, 2003). For example, the HCl formed by R16 with X = Cl lowers the yield of the reaction chain starting with this reaction because of the reaction

\[ \text{HCl} + \text{OH} \rightarrow \text{H}_2\text{O} + \text{Cl}. \]  

(R17)

Reaction of water with O(\textsuperscript{1}D)

\[ 5 \text{H}_2\text{O} + \text{O(\textsuperscript{1}D)} \rightarrow \text{OH} + \text{OH} \]  

(R18)

and photolysis of HNO\textsubscript{3}

\[ \text{HNO}_3 + h\nu \rightarrow \text{NO}_2 + \text{OH} \]  

(R1)

can also produce HO\textsubscript{x}. Sinks are the recombination into water

\[ \text{OH} + \text{HO}_2 \rightarrow \text{H}_2\text{O} + \text{O}_2 \]  

(R19)

and the reactions

\[ \text{OH} + \text{NO}_2 + \text{M} \rightarrow \text{HNO}_3 + \text{M} \]  

(R5)

\[ \text{HNO}_3 + \text{OH} \rightarrow \text{H}_2\text{O} + \text{NO}_3 \]  

(R2)
ClO + OH → HCl + O₂. \hspace{1cm} (R20)

 HOₓ is in equilibrium with HOCl under sunlit conditions and when chlorine is activated (e.g. Portmann et al., 1996)

\[
\text{HOCl} + h\nu \rightarrow \text{Cl} + \text{OH}
\] \hspace{1cm} (R21)

ClO + HO₂ → HOCl + O₂. \hspace{1cm} (R22)

Hence, the heterogeneous reaction

\[
\text{HOCl} + \text{HCl} \rightarrow \text{Cl}_2 + \text{H}_2\text{O}
\] \hspace{1cm} (R23)

is also a sink for HOₓ. The rates of the reactions changing extended HOₓ increase by a factor of 4 from 70 hPa to 32 hPa (see supplement). The relative importance of the reactions does not change with the exception of methane oxidation, which becomes less important with increasing height.

HOₓ levels are about 1 to 4 ppt at 54 hPa under sunlit conditions in both hemispheres (Figure 9). Maximum values increase from 3 ppt at 70 hPa to 7 ppt at 32 hPa. At night, no HOₓ is present, since there is no production and HOₓ recombines into water and HNO₃ (Figure 8). HOCl can reach mixing ratios of up to 0.15 ppb at 54 hPa in both hemispheres, as long as chlorine is activated (see Figure 11). Figure 9 shows that HOₓ does not simply scale with the amount of sunlight, HNO₃, H₂O and CH₄: In the southern hemisphere, HOₓ shows a peak under conditions of both chlorine activation and sunlight, related to the fact that reactions like Cl + CH₄ and HOCl + HCl play a role in HOₓ chemistry (Figure 8, see also the discussion in Section 4.3.2 related to HCl).

Figure 9 also shows the partitioning of HOₓ. Similar to NOₓ, there is a fast equilibrium between OH and HO₂. The partitioning between the HOₓ species is determined mainly by

\[
\text{HOCl} + h\nu \rightarrow \text{Cl} + \text{OH}
\] \hspace{1cm} (R21)

\[
\text{ClO} + \text{HO}_2 \rightarrow \text{HOCl} + \text{O}_2
\] \hspace{1cm} (R22)

\[
\text{ClO} + \text{OH} \rightarrow \text{Cl} + \text{HO}_2
\] \hspace{1cm} (R24)

\[
\text{O}_3 + \text{HO}_2 \rightarrow \text{OH} + 2\text{O}_2
\] \hspace{1cm} (R25)

\[
\text{O}_3 + \text{OH} \rightarrow \text{HO}_2 + \text{O}_2
\] \hspace{1cm} (R26)
Figure 9. Vortex-averaged partitioning of HOx species for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Daytime averages (parts of the vortex where the solar zenith angle is smaller than 80°). Nighttime averages are near zero and not shown. Days without sufficient data for averaging are not shown (grey bars).

\[ \text{NO} + \text{HO}_2 \rightarrow \text{NO}_2 + \text{OH}. \]  

(R27)

Figure 10 shows the rates of HO2 production and loss (the corresponding plots for OH look similar, but mirrored, with ClO + HO2 replaced by HOCl + hν).

The rates of the reactions partitioning HOx increase by a factor of 4 from 70 hPa to 32 hPa again (see supplement), the relative importance of the reactions does not change except for the reaction R24 getting less important with increasing altitude.

The production and loss rates of HOx proper are dominated by the reactions R21 and R22 which form the equilibrium with HOCl (not shown).

The equilibrium is mainly on the side of HO2 in both hemispheres (see Figure 9, e.g. Hanisco, 2003). The fraction of OH is somewhat higher in the southern hemisphere (about 20%–40%) than in the northern hemisphere (about 10%–20%), see also percentage plots in the supplement. The relative partitioning is approximately constant with altitude (see supplement). Absolute OH levels are of relatively similar magnitude in both hemispheres (up to 1 ppt in the southern hemisphere and up to 0.5 ppt in the northern hemisphere).

Equations for the equilibrium values of OH, HO2 and HOCl can be derived from reactions R21 to R27. The ratio of OH and HO2 under sunlit conditions in the northern hemisphere and in the southern hemisphere before mid-September is to a good approximation given by

\[ \frac{[\text{OH}]}{[\text{HO}_2]} = \frac{k_{R25}[\text{O}_3] + k_{R22}\text{[ClO]} + k_{R27}\text{[NO]}}{k_{R26}[\text{O}_3] + k_{R24}\text{[ClO]}}. \]  

(6)
Figure 10. Vortex-averaged chemical reaction rates of reactions changing HO$_2$ for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa to illustrate HO$_x$ partitioning. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative. The net change of HO$_2$ is shown as a green line.

That is, the ratio depends only on O$_3$, ClO and NO. The equation can be simplified to

$$\frac{[\text{OH}]}{[\text{HO}_2]} = \frac{k_{R25}[\text{O}_3] + k_{R22}[\text{ClO}]}{k_{R26}[\text{O}_3] + k_{R24}[\text{ClO}]}$$

under conditions when chlorine is activated and no NO is present (see Figure 5, before March 2005 and mid-September 2006 in the northern and southern hemisphere, respectively) and to

$$\frac{[\text{OH}]}{[\text{HO}_2]} = \frac{k_{R25}[\text{O}_3] + k_{R27}[\text{NO}]}{k_{R26}[\text{O}_3]}$$

in spring in the northern hemisphere (end of March 2005, after chlorine is deactivated). In the southern hemisphere, the change between conditions rich in ClO$_x$ and rich in NO$_x$ is marked by a change in the relative partitioning of OH from about 20% to about 40% at the start of October (see percentage plots in the supplement). A similar discussion and expressions for mid-latitudes can be found in Cohen et al. (1994).

Under sunlit conditions and when chlorine is activated, the equilibrium of HOCl is given by

$$[\text{HOCl}] = \frac{k_{R22}}{k_{R21}} [\text{ClO}][\text{HO}_2].$$

At night, HOCl remains constant (except for heterogeneous processing), since neither the photolysis reaction nor the ClO + HO$_2$ reaction can proceed.

HO$_x$ is relevant for ozone depletion, since the reactions HCl + OH → H$_2$O + Cl (R17) and ClO + OH → HCl + O$_2$ (R20) play a role in chlorine activation and deactivation (note that the second reaction differs from reaction R24 in the products). In addition, chlorine can be activated by the heterogeneous HOCl + HCl (R23) reaction.
4.3 Chlorine species

4.3.1 Overview and partitioning

When the polar vortex forms, the majority of chlorine is present in the form of HCl and the remainder is present in the other important reservoir gas ClONO$_2$ (e.g. Solomon, 1999; Santee et al., 2008). Figure 11 shows the partitioning between the various inorganic chlorine species (Cl$_y$). The available amount of Cl$_y$ is about 2.7 to 3.3 ppb at 54 hPa. The increase of Cl$_y$ is due to transport from above. Due to the stronger descent in the northern hemisphere, Cl$_y$ increases to values that are about 0.3 ppb higher in the northern hemisphere in 2005 than in the southern hemisphere in spring 2006, although the initial values are similar. This increases the chlorine potentially available for ozone depletion in the northern hemisphere compared to the southern hemisphere. In both hemispheres, the initial mixing ratio of HCl is about 2 ppb at 54 hPa (75% of Cl$_y$) and the initial mixing ratio of ClONO$_2$ is about 0.7 ppb (see also Santee et al., 2008). Cl$_y$ is produced by photolysis and reaction with O($^1$D) from chlorofluorocarbons (CFCs), hydrochlorofluorocarbons (HCFCs) and similar species of mainly anthropogenic origin (e.g. WMO, 2011; Montzka, 2012). The relative partitioning inside Cl$_y$ is approximately constant between 70 hPa and 32 hPa (with only a slight increase in Cl$_y$ with altitude, see supplement), so that most of the following discussion applies to the entire altitude range.

In early winter, passive reservoir gases HCl and ClONO$_2$ are transformed to Cl$_2$ through heterogeneous reactions on the surface of polar stratospheric clouds (e.g. Solomon et al., 1986), starting in December 2004 in the northern hemisphere and in May 2006 in the southern hemisphere. Since the major reaction that transforms the reservoir gases to Cl$_2$ is HCl + ClONO$_2$ (e.g. Solomon et al., 1986, see also Section 4.3.2), the amount of chlorine that can be activated is limited by the mixing ratio of the less abundant ClONO$_2$ (e.g. Portmann et al., 1996). While some ClONO$_2$ can be regenerated by the reaction ClO + NO$_2$ + M and some HCl can be activated by the reaction HOCl + HCl, this is not sufficient to remove all HCl over the course of the winter, and total HCl mixing ratios consistently stay larger than 0.5 ppb in our model runs, while ClONO$_2$ decreases to near zero values. Due to the applied correction to the HCl solubility (see Section 6), up to 0.5 ppb of HCl is dissolved in STS droplets in the southern hemisphere at 54 hPa (more with increasing altitude).

When sunlight starts to come back, Cl$_2$ is quickly transformed to the active chlorine species ClO and its dimer Cl$_2$O$_2$ by photolysis of Cl$_2$

\[
\text{Cl}_2 + h\nu \rightarrow \text{Cl} + \text{Cl}
\]  \hspace{1cm} (R28)

immediately followed by the reaction

\[
\text{Cl} + \text{O}_3 \rightarrow \text{ClO} + \text{O}_2
\]  \hspace{1cm} (R29)

to produce ClO. The time between first activation and the first sunlight becomes apparent in a Cl$_2$ peak in December 2004 and May 2006 in Figure 11.
Figure 11. Vortex-averaged partitioning of inorganic chlorine species (Cl, below) for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Top row: Daytime averages (parts of the vortex where the solar zenith angle is smaller than 80°). Bottom row: Nighttime averages (parts of the vortex where the solar zenith angle is larger than 100°). Days without sufficient data for averaging are not shown (grey bars). Species ClNO2 and BrCl are not shown due to their small mixing ratios. The area labeled “HCl (cloud)” shows HCl dissolved in STS droplets due to the applied correction to the HCl solubility (see section 6).
Under polar conditions, the dimer of ClO plays an important role. Reactions that determine the ratio of ClO and its dimer Cl₂O₂ are

\[ \text{ClO} + \text{ClO} + \text{M} \rightarrow \text{Cl}_2\text{O}_2 + \text{M} \]  \hspace{1cm} (R30)

\[ \text{Cl}_2\text{O}_2 + h\nu + \text{M} \rightarrow 2\text{Cl} + \text{O}_2 + \text{M} \]  \hspace{1cm} (R31)

\[ \text{Cl}_2\text{O}_2 + \text{M} \rightarrow \text{ClO} + \text{ClO} + \text{M} \]  \hspace{1cm} (R32)

Reactions R30 and R31 are part of the ClO dimer ozone loss cycle and play an important role in ozone depletion (see Section 4.5). Note that reaction R31 includes an intermediate step over ClOO not shown here. Figure 12 shows the partitioning of ClOₓ = ClO + 2Cl₂O₂ at 54 hPa. Active chlorine is mainly present in the form of Cl₂O₂ at night and in the form of ClO during daytime (e.g. Stimpfle et al., 2004; WMO, 2011). At night, only small levels of ClO (10 % of ClOₓ in our model) are maintained by the interplay between the forward reaction R32 (thermal dissociation) and backward reaction R30 (e.g. Stimpfle et al., 2004; WMO, 2011). The nighttime equilibrium is highly temperature dependent (e.g. Stimpfle et al., 2004), and higher temperatures shift the equilibrium towards ClO at night (see Figure 12). During daytime, most Cl₂O₂ is photolyzed into Cl which reacts to ClO and there is an equilibrium between ClO and Cl₂O₂ by the photolysis reaction R31 and reaction R30 (e.g. WMO, 2011). About 70 % of ClOₓ is present as ClO during daytime in all altitudes in our model (Figure 12 and supplement). This is in good agreement with the 60 %–70 % inferred from direct aircraft observations of ClO and ClOOCl by Stimpfle et al. (2004). Modelled ClO values are also in reasonable agreement with MLS measurements in the northern hemisphere in 2005 and in the southern hemisphere in 2006 (see supplement).

There is a long history of studies investigating the reactions R30 to R32, which are central for polar ozone depletion and have been uncertain for a long time (e.g. Molina and Molina, 1987; Burkholder et al., 1990; Stimpfle et al., 2004; Frieler et al., 2006; Schofield et al., 2008; Kremser et al., 2011; Canty et al., 2016). Recently, a study of the photolysis cross section of Cl₂O₂ challenged the understanding of polar ozone depletion and caused a thorough reinvestigation of the chemistry of ozone depletion (see WMO, 2011, 2.2.2 and references therein). The uncertainties have been resolved (Kawa et al., 2009; WMO, 2011; von Hobe and Stroh, 2012; Canty et al., 2016), but the Cl₂O₂ photolysis is still one of the reactions that causes the largest uncertainties in ozone depletion (Kawa et al., 2009). In addition, there is also some uncertainty in the nighttime equilibrium between ClO and Cl₂O₂ (Kawa et al., 2009; Canty et al., 2016), but this is not relevant for the amount of ozone depletion.

When chlorine is activated (from December 2004 to the beginning of March 2005 in the northern hemisphere and from May 2006 to September 2006 in the southern hemisphere), ozone is removed by the ClO dimer cycle and the ClO–BrO cycle (e.g. Solomon, 1999, see also Section 4.5). ClOₓ peaks at about 2.4 ppb in the northern hemisphere and at about 2.0 ppb in the southern hemisphere at 54 hPa, with little variation from 70 hPa to 32 hPa (see supplement).
Figure 12. Vortex-averaged partitioning of ClOx for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Top row: Daytime averages (parts of the vortex where the solar zenith angle is smaller than $80^\circ$). Bottom row: Nighttime averages (parts of the vortex where the solar zenith angle is larger than $100^\circ$). Days without sufficient data for averaging are not shown (grey bars).
The higher peak values of ClO$_x$ in the northern hemisphere are caused not only by the higher Cl$_y$ values (Figure 11), but also by the more efficient activation of HCl into ClO$_x$ in the northern hemisphere, which provides about 0.1–0.2 ppb additional chlorine from HCl compared to the southern hemisphere (see discussion in the section “HCl loss” below).

4.3.2 Reservoir HCl

Figure 13 shows the time evolution of the vortex-averaged reaction rates of all relevant reactions that change HCl at 54 hPa. The supplement shows that the rates and relative importance of the reactions are similar from 70 hPa to 32 hPa, so that the following discussion applies to the entire altitude range. In early winter (up to the end of December 2004 in the northern hemisphere and up to the end of July 2006 in the southern hemisphere), HCl is removed by heterogeneous reactions on polar stratospheric clouds and active chlorine is produced. When sunlight returns, this is followed by a phase of competition between HCl removal by heterogeneous reactions and deactivation of ClO$_x$ into HCl by gas-phase reactions (e.g. Portmann et al., 1996; Solomon et al., 2015), most pronounced in August 2006 in the southern hemisphere. In the southern hemisphere, this is followed by a phase of deactivation of active chlorine into HCl by the Cl + CH$_4$ reaction in September to October (e.g. Douglass et al., 1995). Finally, when sunlight comes back, the gas-phase loss reaction HCl + OH becomes important and competes with the production by Cl + CH$_4$. In the southern hemisphere, HCl is near equilibrium during this time, while in the northern hemisphere, chlorine that was initially deactivated into ClONO$_2$ is slowly transformed into HCl.

HCl loss (chlorine activation)

The most important heterogeneous loss reaction for HCl is

\[
\text{ClONO}_2 + \text{HCl} \rightarrow \text{Cl}_2 + \text{HNO}_3 \quad (R4)
\]

Another heterogeneous activation channel for HCl is

\[
\text{HOCl} + \text{HCl} \rightarrow \text{Cl}_2 + \text{H}_2\text{O}. \quad (R23)
\]

While this reaction plays a smaller role in the northern hemisphere in 2005, the reaction is important in the southern hemisphere in the months August and September 2006. In the southern hemisphere, the HOCl + HCl reaction accounts for about 70% of the HCl activation by heterogeneous reactions (integrated over the winter) and in the northern hemisphere it accounts for about 30%. HOCl only exists in significant amounts when chlorine is activated. In the southern hemisphere, there is a shift from activation by ClONO$_2$ + HCl in early winter to activation dominated by HOCl + HCl later in spring. The reason for this is the deactivation of chlorine by the reaction Cl + CH$_4$ in the southern hemisphere (caused by the low ozone values, see next section below), which provides HCl and produces HO$_x$ and in turn HOCl (for a detailed discussion, see Crutzen et al., 1992, Portmann et al., 1996). This can clearly be seen in the rising HCl levels in September in the southern hemisphere (Figure 11) and in the increased HO$_x$ levels (Figure 9). In contrast, deactivation is mainly into ClONO$_2$ in the northern hemisphere, keeping HCl levels low. Since the southern hemisphere is more denitrified, activation by ClONO$_2$ + HCl is hindered. Note that the net effect of Cl + CH$_4$ and HOCl + HCl is a net chlorine deactivation in the southern hemisphere (green line in Figure 13).
Figure 13. Vortex-averaged chemical reaction rates of reactions involving HCl for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative. The net change of HCl is shown as a green line. Reactions with rates which cannot be distinguished from the zero line at plot resolution are not shown.

It is evident that initially about 0.1–0.2 ppb more HCl is removed in the northern hemisphere in 2005 than in the southern hemisphere in 2006 (Figure 11), although the temperatures are lower in the southern hemisphere and the initial values of HCl and ClONO$_2$ are similar. Figure 13 shows that the reason is that the ClONO$_2$ + HCl reaction activates more HCl in the northern hemisphere. This is caused by reformation of ClONO$_2$ after the initial depletion in the northern hemisphere (see Figure 17 below), which is much smaller in the southern hemisphere.

The only other reaction that removes HCl in relevant quantities is the gas-phase reaction with OH, which is only important under sunlit conditions:

\[ \text{HCl} + \text{OH} \rightarrow \text{H}_2\text{O} + \text{Cl} \]  

(R17)

In the northern hemisphere, the reaction is relevant in March in 2005. For the southern hemisphere, the reaction is relevant in October 2006 and later.

**HCl production (chlorine deactivation)**

HCl is only produced by reactions that are indirectly dependent on sunlight. Deactivation of active chlorine occurs mainly into HCl in the southern hemisphere but predominantly into ClONO$_2$ in the northern hemisphere. The reason for this is that deactivation into ClONO$_2$ is hindered in the southern hemisphere by strong denitrification (absence of NO$_2$), while normally it would be the preferred pathway of deactivation (e.g. Douglass et al., 1995; Portmann et al., 1996).

In the long term, the partitioning of HCl and ClONO$_2$ in spring and summer favors HCl (e.g. Portmann et al., 1996; Santee et al., 2008). This can be seen in the fact that in the northern hemisphere, HCl is produced from ClONO$_2$ in spring after ClO$_x$
has been deactivated into ClONO$_2$ some time earlier (see also Section 4.3.3 and Figure 17), while in the southern hemisphere, active chlorine is deactivated mainly into HCl. The chemical change rates of both HCl and ClONO$_2$ decrease to zero in November in the southern hemisphere (green lines in Figures 13, 16 and 17). The reaction

\[
\text{Cl} + \text{CH}_4 \rightarrow \text{HCl} + \text{CH}_3 \tag{R33}
\]

is the main production reaction for HCl in both hemispheres (e.g. Portmann et al., 1996). In addition, it is responsible for deactivation under ozone hole conditions in the southern hemisphere. The reaction

\[
\text{ClO} + \text{OH} \rightarrow \text{HCl} + \text{O}_2 \tag{R20}
\]

becomes similar in importance to Cl + CH$_4$ as a production process for HCl around late February 2005 in the northern hemisphere, and plays a smaller role earlier and later in winter. In the southern hemisphere, the reaction is not negligible around September and October 2006. The reaction is responsible for part of the chlorine deactivation in both hemispheres. The reaction ClO + OH has two product channels. The channel into HCl yields about 8% of the products (the other channel is into Cl).

**The role of Cl in HCl production**

The rate of the Cl + CH$_4$ reaction is proportional to the mixing ratio of Cl. Figure 14 shows the mixing ratio of Cl for both hemispheres. Figure 15 shows that Cl is determined by the two source reactions

\[
\text{Cl}_2\text{O}_2 + h\nu + \text{M} \rightarrow 2\text{Cl} + \text{O}_2 + \text{M} \tag{R31}
\]

\[
\text{ClO} + \text{NO} \rightarrow \text{Cl} + \text{NO}_2 \tag{R7}
\]

and a reaction that removes Cl

\[
\text{Cl} + \text{O}_3 \rightarrow \text{ClO} + \text{O}_2. \tag{R29}
\]

Reaction R31 is coupled to the catalytic ClO dimer cycle. Cl levels are considerably higher in the southern hemisphere due to a lack of ozone, which hinders the recombination to ClO (e.g. Douglass et al., 1995). This favors the deactivation of active chlorine via the Cl + CH$_4$ reaction (e.g. Portmann et al., 1996). Cl shows a very distinct behaviour in the southern hemisphere. It does not just increase with the amount of sunlight, but shows a peak in September, followed by a near constant plateau. This curve shape is approximately repeated in the curves of the reaction rate of the Cl + CH$_4$ reaction (Figure 13), since CH$_4$ is relatively constant. Figure 15 shows that the peak in Cl mixing ratios in the southern hemisphere is related to reaction R31, i.e. the catalytic ozone destruction. In contrast, the plateau is related to the NO$_x$ reaction R7 (see also Wayne et al., 1995, p. 2836).

**4.3.3 Reservoir ClONO$_2$**

Figure 16 shows the reaction rates of the most important reactions changing ClONO$_2$. The reaction rates increase by a factor of 2 from 70 hPa to 32 hPa in 2004/2005, but are more constant in 2009/2010 (see supplement). As discussed in Section 4.1,
Figure 14. Vortex-averaged Cl mixing ratios for the Arctic winter 2004/2005 and the Antarctic winter 2006 at 54 hPa.

Figure 15. Vortex-averaged chemical reaction rates of reactions involving Cl for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative.
Figure 16. Vortex-averaged chemical reaction rates involving ClONO₂ for the Arctic winter 2004/2005 and the Antarctic winter 2006 at 54 hPa. Production reactions are shown positive and are separated by a line in the legend from the loss reactions, which are shown negative. The net change of ClONO₂ is shown as a green line. Reactions with rates which cannot be distinguished from the zero line at plot resolution are not shown.

The gross change rates of ClONO₂ are dominated by a near equilibrium between the reactions

\[
\text{ClONO}_2 + h\nu \rightarrow \text{Cl} + \text{NO}_3 \tag{R10}
\]

\[
\rightarrow \text{CIO} + \text{NO}_2 \tag{R11}
\]

and

\[
\text{CIO} + \text{NO}_2 + M \rightarrow \text{ClONO}_2 + M \tag{R12}
\]

under sunlit conditions (e.g. Portmann et al., 1996). Net changes of ClONO₂ are induced by changes in NOₓ which shift the equilibrium (e.g. Portmann et al., 1996).

The net change of ClONO₂ is one order of magnitude smaller than the gross rates. Figure 17 shows the net effect of R10–R12 and all other reactions that play a role in changing ClONO₂. These reactions, which are all loss reactions, remove ClONO₂ either by heterogeneous reactions or gas-phase reactions of ClONO₂ with a radical. The most important heterogeneous reaction is the same as for HCl

\[
\text{ClONO}_2 + \text{HCl} \rightarrow \text{Cl}_2 + \text{HNO}_3. \tag{R4}
\]

The reactions of ClONO₂ with a radical are of the type

\[
\text{ClONO}_2 + X \rightarrow \text{XCl} + \text{NO}_3 \tag{R34}
\]
Figure 17. Vortex-averaged chemical reaction rates involving ClONO$_2$ for the Arctic winter 2004/2005 and the Antarctic winter 2006 at 54 hPa. In contrast to Figure 16, the net production rate of the fast cycle ClONO$_2$ + hv → Products / ClO + NO$_2$ + M → ClONO$_2$ + M is shown. This cycle is separated by a line in the legend from the loss reactions. The green line shows the net change of ClONO$_2$ by chemistry. Reactions with rates which cannot be distinguished from the zero line at plot resolution are not shown.

where X is X = O, Cl or OH. The reaction with Cl is the dominant reaction. The relative importance of these reactions is approximately constant from 70 hPa to 32 hPa (see supplement).

In early winter in the polar night (December 2004 in the northern hemisphere, May and June 2006 in the southern hemisphere), the loss of ClONO$_2$ by heterogeneous reactions dominates, leading to the activation of chlorine. After the complete depletion of ClONO$_2$, this is followed by a phase with only small production and loss, due to near-zero levels of both NO$_x$ and ClONO$_2$. The peak in production in early spring in the northern hemisphere (end of February and beginning of March 2005) is caused by the net production of extended NO$_x$ from HNO$_3$ and is the main deactivation pathway for active chlorine in the northern hemisphere (e.g. Douglass et al., 1995; Portmann et al., 1996). At that time, ClONO$_2$ is out of equilibrium and more ClONO$_2$ is produced by ClO + NO$_2$ than is lost by photolysis. In the southern hemisphere, the rates are much lower due to the strongly denitrified conditions. In late March 2005, ClONO$_2$ is lost in the northern hemisphere and finally converted to HCl, which is the favored reservoir under summer conditions (e.g. Portmann et al., 1996; Santee et al., 2008). In the southern hemisphere, the rates of change are low in late October and November, since the deactivation already occurred directly into HCl.

4.4 Bromine species

Sources of inorganic bromine, which has both natural and anthropogenic sources, are mainly halons and methyl bromide, but also some short-lived species (e.g. WMO, 2011; Montzka, 2012). Inorganic bromine (Br$_y$) levels in the stratosphere are about 20 ppt at maximum (e.g. WMO, 2011), with marginally lower levels in the altitude range considered here. Bromine chemistry is still somewhat uncertain due to uncertainties in the reaction coefficients (Sander et al., 2011; von Hobé and Stroh, 2012).
Atomic bromine is released from the source gases mainly by photolysis. Under sunlit conditions, it is in equilibrium with BrO. The relevant reactions are

\[
\text{Br + O}_3 \rightarrow \text{BrO + O}_2 \quad \text{(R35)}
\]

\[
\text{5} \quad \text{BrO + h}\nu \rightarrow \text{Br + O} \quad \text{(R36)}
\]

\[
\text{BrO + NO} \rightarrow \text{NO}_2 + \text{Br} \quad \text{(R37)}
\]

\[
\text{BrO + ClO} \rightarrow \text{Br + Cl} + \text{O}_2. \quad \text{(R38)}
\]

Reactions R35 and R38 are directly involved in the ClO–BrO cycle, see Section 4.5.

Figure 18 shows the partitioning between the various inorganic bromine species at 54 hPa, which is very similar from 70 hPa to 32 hPa (see supplement). Bromine is mostly present in the form of HOB\text{r} and Br\text{ONO}_2 at night before chlorine activation starts. In contrast to the less reactive chlorine, these are not real reservoir gases, since they easily photolyze into Br and react further to BrO during daytime, which gives bromine a great potential to destroy ozone despite the low mixing ratios (e.g. Lary, 1996; Solomon, 1999). BrO is a dominant species during daytime (e.g. Lary, 1996), except in October and November in the southern hemisphere, when the reaction Br + O_3 is hindered by the low ozone levels and Br mixing ratios are significant. Heterogeneous reactions play only a minor role and are not needed for activation (e.g. Lary et al., 1996; Wayne et al., 1995). As long as chlorine is activated, almost all bromine is in the form of Br\text{Cl} at night (e.g. Lary et al., 1996; von Hobe and Stroh, 2012). Br\text{Cl} is produced by the reaction

\[
\text{BrO + ClO} \rightarrow \text{BrCl} + \text{O}_2. \quad \text{(R39)}
\]

During daytime, most of this Br\text{Cl} is transformed to Br by photolysis

\[
\text{BrCl + h}\nu \rightarrow \text{Br + Cl} \quad \text{(R40)}
\]

followed by transformation to BrO by reaction R35.

4.5 Oxygen species

Figure 19 shows the vortex-averaged mixing ratios of ozone for the northern and southern hemisphere. Ozone stays more constant in the northern hemisphere 2005 compared to the southern hemisphere 2006 (1.5 ppm to 2.5 ppm at 54 hPa), since the change by ozone depletion is nearly cancelled by transport of air rich in ozone from above and over the vortex edge. In
Figure 18. Vortex-averaged partitioning of inorganic bromine species for the Arctic winter 2004/2005 (left) and the Antarctic winter 2006 (right) at 54 hPa. Top row: Daytime averages (parts of the vortex where the solar zenith angle is smaller than 80°). Bottom row: Nighttime averages (parts of the vortex where the solar zenith angle is larger than 100°). Days without sufficient data for averaging are not shown (grey bars).
Figure 19. Vortex-averaged ozone mixing ratios for the Arctic winter 2004/2005 and the Antarctic winter 2006 at 54 hPa.

In contrast, ozone values decrease from 2.5 ppm to less than 0.5 ppm at 54 hPa in the southern hemisphere in 2006, both due to weaker transport and larger ozone depletion. While the initial ozone value of 2.5 ppm in the northern hemisphere in 2004/2005 is relatively constant with height, the value at the end of the winter varies from 1 ppm at 70 hPa to 3 ppm at 32 hPa due to changes in ozone depletion, production and transport (see supplement). In the warmer winter 2009/2010, values are more similar at the end of the winter and less ozone depletion is observed (see supplement).

$O_3$ is in a very fast equilibrium with $O$. The dominant cycle is part of the well-known Chapman chemistry (Chapman, 1930)

$$O_3 + h\nu \rightarrow O + O_2$$  \hspace{1cm} (R41)

$$O + O_2 + M \rightarrow O_3 + M.$$  \hspace{1cm} (R42)

We follow the usual convention here to treat $O_3$ and $O$ together as odd oxygen $O_x$ (e.g. Brasseur and Solomon, 2005; Solomon, 1999). Since $O$ mixing ratios are low, the chemical rate of change of $O_x$ is nearly the same as the rate of change of ozone.

Odd oxygen is destroyed by several catalytic cycles. It is well known (e.g. Solomon, 1999) that the two dominant cycles in anthropogenic polar ozone depletion by halogens are the ClO dimer cycle (first proposed by Molina and Molina, 1987)

$$2\left( Cl + O_3 \rightarrow ClO + O_2 \right)$$  \hspace{1cm} (R29)

$$ClO + ClO + M \rightarrow Cl_2O_2 + M$$  \hspace{1cm} (R30)

$$Cl_2O_2 + h\nu + M \rightarrow 2Cl + O_2 + M$$  \hspace{1cm} (R31)

$$2O_3 \rightarrow 3O_2$$
Figure 20. Vortex-averaged net chemical change of odd oxygen. The green line shows the net chemical change rate of ozone, which nearly equals the change rate of odd oxygen at this altitude. The contribution of different catalytic cycles to the ozone loss is shown by the reaction rates of their rate limiting step. Only the three most important cycles are shown, the contribution of other cycles is small. Ozone production, which is almost exclusively by the $O_2 + h\nu$ reaction, is shown in cyan.

and the ClO–BrO cycle (first proposed by McElroy et al., 1986)

$$\begin{align*}
\text{Cl} + O_3 & \rightarrow \text{ClO} + O_2 \\
\text{Br} + O_3 & \rightarrow \text{BrO} + O_2 \\
\text{BrO} + \text{ClO} + M & \rightarrow \text{Br} + \text{Cl} + O_2 + M
\end{align*}$$

$$2O_3 \rightarrow 3O_2.$$

The ClO dimer cycle is effective, because the Cl$_2$O$_2$ photolysis, which produces Cl and not ClO, is fast compared to the competing Cl$_2$O$_2$ loss reaction Cl$_2$O$_2$ + M (Wayne et al., 1995, p. 2836). The uncertainty of the rate coefficient of the Cl$_2$O$_2$ photolysis (R31) and the rate coefficient of the BrO + ClO reaction (R38) are two of the parameters which produce the largest uncertainties with respect to ozone loss in models (e.g. Frieler et al., 2006; WMO, 2011), although much of the uncertainty in the photolysis rate coefficient has been resolved recently (Kawa et al., 2009; von Hobe and Stroh, 2012; Canty et al., 2016).

Figure 20 shows the contribution of the different catalytic cycles to the net chemical rate of change of ozone. For this purpose, the reaction rates of the rate limiting step of the reaction cycles have been used. This is possible here without ambiguities, since all rate limiting reactions are only involved in one cycle.

Typical peak loss rates vary between 40–60 ppb per day and are similar in both hemispheres in the altitude range 70 hPa to 32 hPa for these two winters (compare also the peak values of ClO$_x$ discussed earlier and the supplement). Somewhat higher values of up to 70 ppb are only seen in 42 hPa and 32 hPa in the southern hemisphere. Hence, the higher ozone depletion in the southern hemisphere is mainly caused by the much longer time period in which ClO$_x$ is activated. Photochemical production
of ozone is evident in March 2005 in the northern hemisphere and in October and November 2006 in the southern hemisphere and increasing with altitude (see supplement).

As long as appreciable amounts of ClO exist (January to February 2005 in the northern hemisphere, June to September 2006 in the southern hemisphere), the ClO–ClO cycle contributes about 50% to the net ozone loss and the ClO–BrO cycle contributes about 30–40% in all altitudes (see supplement). These values compare well with the results of Kuttippurath et al. (2010) at 475 K and for 2005 (50% ClO–ClO to 30% ClO–BrO) and with the results of Frieler et al. (2006) (50% ClO–ClO to 30%–50% ClO–BrO, but for years different from 2005). See also Grenfell et al., 2006, for case studies of the partitioning.

Loss cycles of the form

\[ X + O_3 \rightarrow XO + O_2 \]
\[ XO + O \rightarrow X + O_2 \]
\[ O + O_3 \rightarrow O_2 + O_2 \]

where \( X = \text{OH, H, NO, Cl, Br} \), which dominate in most latitudes and seasons, only play a minor role in the polar lower stratosphere due to the relatively low production of atomic oxygen by photolysis (e.g. von Hobe and Stroh, 2012). The ClO–O cycle contributes about 10% at 54 hPa and increases in importance with altitude, in agreement with e.g. Kuttippurath et al. (2010).

The most important null cycle acting on odd oxygen apart from reactions R41 and R42 is the nitrogen cycle

\[ O_3 + NO \rightarrow NO_2 + O_2 \] \hspace{1cm} (R6)
\[ NO_2 + h\nu \rightarrow NO + O \] \hspace{1cm} (R8)
\[ O_3 \rightarrow O_2 + O. \]

5 Interannual variability

In addition to the runs for the Arctic winter 2004/2005 and the Antarctic winter 2006, we performed model runs for the Arctic winter 2009/2010 and the Antarctic winter 2011 to get at least an idea of the interannual variability of the reaction rates and partitioning and to assess if the results can be generalized. Figures for the additional years can be found in the supplement.

Since meteorological conditions are usually very similar in the southern hemisphere from year to year, the results do not change significantly for the Antarctic winter 2011 compared to 2006, and the same discussion as for the winter 2006 applies. Except for rare cases as the vortex split in 2002, the dates, magnitudes of the values and the relative partitioning will also be applicable to other southern winters.

The northern winters 2004/2005 and 2009/2010 were taken as examples for a colder and a warmer winter and reflect the larger meteorological variability in the northern hemisphere. The main differences are in the timing of the changes in reaction rates and mixing ratios. The first main difference is a later activation of chlorine (end of December 2010 compared to start of December 2005), reflected e.g. in the timing of the heterogenous reaction rates. The second main difference is an earlier deactivation by an earlier stratospheric warming in February. That means an earlier increase in e.g. ClONO\(_2\), NO\(_x\) or HO\(_x\)
and the associated reaction rates, and an earlier decrease in ClO$_x$. The timing of activation and deactivation can vary greatly between different northern winters, from winters with no chlorine activation and heterogeneous ozone depletion at all to winters that stay cold until April.

The absolute values of the reaction rates and mixing ratios, the relative partitioning inside the chemical families and the relative importance of reactions remain similar for most species for these two northern winters. One of the notable exceptions is of course ozone, which shows less depletion in a warm winter like 2009/2010. This comparison suggests that the relative partitioning and the relative importance of reactions can be generalized to other northern winters, but that the absolute values of ozone depletion and of several reaction rates will differ.

6 Model validation

It is desirable that the results of the ATLAS model agree well with observations to increase the confidence in the model results for minor species and reaction rates which cannot be confirmed by observations. It is out of the scope of this study to give a comprehensive model validation against observations, but we include an extensive set of comparisons to observations of the MLS (e.g. Santee et al., 2008) and ACE-FTS (e.g. Bernath, 2017) satellite instruments in a supplement to provide additional information for the interested reader. In addition, the reader is referred to Wohltmann et al. (2010) and Wohltmann et al. (2013) for a detailed validation.

Since there is reasonable agreement between measurements and the ATLAS model for many species, we will restrict discussion on the notable differences between ATLAS and observations and show only some selected results in the main text. The focus is on a prominent disagreement between modeled mixing ratios of HCl and observations.

6.1 HCl discrepancy between model and measurements

Figure 21 and 22 show a comparison of the vortex-averaged mixing ratios of some important species (O$_3$, H$_2$O, HNO$_3$, HCl and N$_2$O, blue and black lines) with corresponding vortex averages measured by the MLS satellite instrument (e.g. Santee et al., 2008, red circles with bars for the accuracy, version 3 data). Note that the vortex averages do not take into account the vortex tracer criterion (as in the preceding sections) to facilitate comparison with MLS (the supplement contains comparisons with MLS at individual locations for more detailed information). The blue lines denote the runs actually used in this paper (with a 5 K offset to HCl solubility), and the black lines denote the uncorrected model runs (without any change to HCl solubility).

It is obvious that the runs without a change to the HCl solubility significantly overestimate HCl, a behaviour also observed in other models like SD-WACC (Brakebusch et al., 2013; Wegner, 2013; Solomon et al., 2015) or MIMOSA-CHIM (Kuttippurath et al., 2015). Interestingly, the SLIMCAT CTM shows a discrepancy of the same order of magnitude, but with the opposite sign in a recent publication (Santee et al., 2008), while in an older version, species mixing ratios depend on the model resolution (Chipperfield et al., 1997).

Apart from the discrepancy in HCl, the agreement of model and observations for other species like ozone, water vapour or HNO$_3$ is quite satisfactory. The reason for the HCl discrepancy is unknown, but several solutions are possible:
Figure 21. Vortex-averaged mixing ratios of O₃, H₂O, HCl and HNO₃ for the Arctic winter 2004/2005. The red dots show MLS satellite measurements, the blue lines show the ATLAS runs used in this paper and the black line shows the original runs which are not empirically corrected for the HCl discrepancy.
Figure 22. Comparison of vortex-averaged mixing ratios for the Antarctic winter 2006
- The initial amount of ClONO$_2$ in the model is underestimated, which hinders the HCl + ClONO$_2$ reaction (discussed in Santee et al., 2008; Brakebusch et al., 2013; Wegner, 2013). ClONO$_2$ only represents about 30% of inorganic chlorine at the start of the winter and is clearly the limiting species of the reaction. An underestimation in ATLAS is however unlikely, since it is not supported by measurements of ClONO$_2$ by ACE-FTS (see supplement) and would require increasing the ClONO$_2$ mixing ratios by more than 100%.

- Less NO$_x$ is transported across the vortex edge in the model compared to the real atmosphere, which impedes the reformation of ClONO$_2$ and HCl depletion via the HCl + ClONO$_2$ reaction (discussed in Solomon et al., 2015). This is likely very dependent on the transport and mixing scheme used in the model, and Eulerian (grid-based) models will show differences to Lagrangian (trajectory-based) models here due to the lower numerical diffusion in the latter. The underestimation of HCl in SLIMCAT (Santee et al., 2008) and the dependency of the species abundances on model resolution in the same model (Chipperfield et al., 1997) may point into this direction.

While it cannot be excluded that less NO$_x$ is transported across the vortex edge in ATLAS compared to the real atmosphere, the good agreement of most other species in ATLAS with measurements, including the tracer N$_2$O (Figure 21 and 22), suggests that this option is unlikely. As long as errors in subsidence do not compensate for errors in mixing, this also means that the discrepancy cannot be caused by differences in subsidence between model and real atmosphere.

Note that there is a marked discrepancy in N$_2$O between MLS and the model in the southern hemisphere in October and November, which is however after the period of interest here. The reason for this discrepancy is unknown, but a possible reason could be overestimation of mixing across the vortex edge in ERA Interim in this time period.

- HCl is taken up in PSCs and sediments out of the observed layer (discussed in Wegner, 2013). This is not supported by either the temporal or the spatial evolution of HCl.

- An unknown heterogeneous reaction is depleting HCl. This cannot be excluded, but some boundary conditions need to be fulfilled, e.g. the reaction needs to involve HCl and it must not change the mixing ratios of other observed species too much.

- There is a temperature bias in the ERA Interim data compared to the real atmosphere or a water vapor bias in the model (discussed in Brakebusch et al., 2013; Wegner, 2013; Solomon et al., 2015). The solubility of HCl in STS droplets is a strong function of both temperature and water vapor. The discussion in Simmons et al. (2014) and our own comparisons at the station Ny Ålesund suggest that the bias between ERA Interim and radiosonde data is less than 0.3 K in the considered altitude range and cannot explain the differences between MLS and ATLAS. The water vapor data is in good agreement between MLS and ATLAS (Figure 21 and 22) and is also not likely to cause the differences in HCl.

- The parameterized solubility of HCl in STS droplets is underestimated (discussed in Brakebusch et al., 2013; Wegner, 2013). This is a promising possibility (since it does not affect most other species, consistent with the reasonable agreement of the other species to measurements), but it requires changes to the solubility that are above the stated uncertainties of the solubility parameterization by Luo et al. (1995) that is used in the model.
Since there is not enough evidence to narrow down this list to a likely candidate, we decided on an empirical approach: We introduced a temperature offset for the calculation of the Henry constant of HCl and changed the offset until we obtained a good agreement between measured and modeled HCl, which was the case for an offset of $-5\,\text{K}$. The effect of the changed Henry constant is twofold: First, it increases the amount of HCl dissolved in STS droplets. Second, it increases the rates of the heterogeneous reactions (even if only small amounts of HCl are dissolved) by changing the $\gamma$ values. Brakebusch et al. (2013) and Solomon et al. (2015) did take the approach to introduce a temperature bias, but we opted here for a different solution, since introducing a temperature bias would have an additional direct effect on reaction rates in the model.

The HCl mixing ratios of the model runs with the changed Henry constant agree considerably better than before with the satellite measurements in both hemispheres. A larger discrepancy remains in December in the northern hemisphere. In addition to the results for the corrected runs, the supplement shows the same figures for the uncorrected runs to assess the impact of the changes in HCl solubility. The main difference is considerably less chlorine activation in both hemispheres (e.g. the peak value of ClO for the northern winter 2004/2005 at 54 hPa changes from 2.5 ppb to 1.5 ppb). This corresponds to higher mixing ratios of the reservoir gases and lower reaction rates of HCl and ClONO$_2$. In the northern hemisphere, the reduced chlorine activation leads to reduced ozone loss (e.g. the mixing ratio of ozone at 70 hPa at the end of March 2005 increases from 1 ppm to 1.5 ppm), However, in the southern hemisphere, chlorine activation is still sufficient to deplete ozone to near zero values. Most other findings remain largely unaffected, including: the relative importance of reactions, the relative partitioning and the results for the NO$_x$ and HO$_x$ chemistry.

### 6.2 Other notable differences

The supplement shows maps of comparisons to MLS satellite observations at 46 hPa for the northern winter 2004/2005 and the southern winter 2006. The maps show all measurements on the 15th of every month for the species ClO, HCl, H$_2$O, HNO$_3$, N$_2$O and O$_3$. Additionly, comparisons to ACE-FTS measurements are shown for the southern winter 2006 as a function of equivalent latitude and altitude. Species are H$_2$O, O$_3$, CH$_4$, HCl, ClONO$_2$, N$_2$O, NO, NO$_2$, N$_2$O$_5$ and HNO$_3$. In general, agreement is reasonable both for the spatial patterns and the absolute values, with the following exceptions:

- An underestimation of HNO$_3$ outside of the vortex by the model compared to both MLS and ACE-FTS, which is however outside of the region of interest here and may be related to the initialization.

- An overestimation of O$_3$ outside of the vortex by the model, which may be related to the underestimation of HNO$_3$ (less ozone depletion by NO$_x$).

- A significant overestimation of ClONO$_2$ around 600 K by the model compared to ACE-FTS in the southern hemisphere, very likely caused by the initialization (there is only a limited amount of measurements of ClONO$_2$ and initialization is based on a tracer relationship).
7 Conclusions

We have given a quantitative analysis of the reactions involved in polar ozone depletion in the stratosphere. For clarity, this study focuses on vortex averages in a layer around 54 hPa for one specific winter in each hemisphere, with additional winters and pressure levels shown in a supplement. The chemistry of polar ozone depletion is well known, but quantitative estimates of the partitioning of the chemical families or the importance of single reactions and reaction cycles are rare. To our knowledge, this is the first comprehensive study providing quantitative results averaged over the core of the polar vortex under conditions perturbed by heterogeneous chemistry. The main aim of this study is to quantify the partitioning of HO\textsubscript{x}, NO\textsubscript{x} and ClO\textsubscript{x}, the relative importance of production and loss reactions and the timing of the reactions. Some selected findings are:

- Our results suggest that the relative partitioning inside chemical families and the relative importance of reactions in the northern winter 2004/2005 and the southern winter 2006 can be generalized to other winters. The absolute values of ozone depletion, the absolute values of the reaction rates and the timing of changes will differ, particularly in the northern hemisphere.

- The ClO dimer cycle contributes about 50\% to the vortex-averaged ozone loss at 54 hPa in both hemispheres, while the ClO–BrO cycle contributes about 40\%, in good agreement with earlier studies (e.g. Frieler et al., 2006; Kuttippurath et al., 2010). The ClO–O contributes about 10\% at 54 hPa and increases in importance with altitude at the expense of the ClO dimer and ClO–BrO cycle.

- In the southern hemisphere, there is a clear shift from chlorine activation by the ClONO\textsubscript{2} + HCl reaction in early winter to activation by the HOC\textsubscript{I} + HCl reaction later in winter. HOC\textsubscript{I} + HCl accounts for about 70\% of the activation of HCl in the southern hemisphere at 54 hPa in 2006 (integrated over time), while it accounts for only 30\% of the activation in the northern hemisphere in 2005. The reason for this is the deactivation of chlorine by the reaction Cl + CH\textsubscript{4} in the southern hemisphere (caused in turn by the low ozone values), which provides HCl and produces HO\textsubscript{x} (see e.g. Crutzen et al., 1992, Portmann et al., 1996). This can clearly be seen in rising HCl levels in September in the southern hemisphere and in increased HO\textsubscript{x} levels.

- ClO\textsubscript{x} peaks at 2.0 ppb at 70 hPa and at 2.5 ppb from 54 hPa to 32 hPa in the northern hemisphere in both 2005 and 2010. The peak values in the southern hemisphere in 2006 and 2011 are lower (2.0 ppb at 54 hPa and 42 hPa) due to less transport from above and less reactivation of ClONO\textsubscript{2} than in the northern hemisphere. Hence, the higher ozone depletion in the southern hemisphere is mainly caused by the much longer time period in which ClO\textsubscript{x} is activated. About 70\% of ClO\textsubscript{x} is present as ClO during daytime at 54 hPa, in good agreement with e.g. Stimpfle et al. (2004).
– HOx levels peak at 4 ppt. HOx is mainly produced from CH₄ oxidation in the southern hemisphere, while in the northern hemisphere, production by HNO₃, CH₄ and H₂O play comparable roles. The partitioning between OH and HO₂ results in 20%–40% OH in the southern hemisphere and in 10%–20% OH in the northern hemisphere.

– NOx levels are smaller than 2 ppb in the northern hemisphere and smaller than 0.75 ppb in the southern hemisphere due to the denitrified conditions there. The partitioning between NO and NO₂ during daytime results in 80%–90% NO in the southern hemisphere and 20%–40% NO in the northern hemisphere. The higher NO levels are caused by the much lower ozone levels in the southern hemisphere (e.g. Douglass et al., 1995).

– 60%–80% of the production of NOx in spring is caused by the HNO₃ + OH reaction at 54 hPa in the northern hemisphere, the remainder is caused by the HNO₃ + hν reaction (cf. Portmann et al., 1996). Deactivation of ClOₓ by the formation of ClONO₂ in the northern hemisphere is caused by a shift in the fast equilibrium between ClO, NO₂ and ClONO₂, which in turn is caused by the production of NOx (e.g. Portmann et al., 1996).

Results of this study are extensively used in a companion paper (Wohltmann et al., 2017) to develop a fast model for polar ozone chemistry.

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