Latitudinal and interhemispheric variation of stratospheric effects on mesospheric ice layer trends

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[1] Latitudinal and interhemispheric differences of model results on trends in mesospheric ice layers and background conditions are analyzed. The model nudges to European Centre for Medium-Range Weather Forecasts data below ~45 km. Greenhouse gas concentrations in the mesosphere are kept constant. Temperature trends in the mesosphere mainly come from shrinking of the stratosphere and from dynamical effects. Water vapor increases at noctilucent cloud (NLC) heights and decreases above due to increased freeze drying caused by temperature trends. There is no tendency for ice clouds in the Northern Hemisphere for extending farther southward with time. Trends of NLC albedo are similar to satellite measurements, but only if a time period longer than observations is considered. Ice cloud trends get smaller if albedo thresholds relevant to satellite instruments are applied, in particular at high polar latitudes. This implies that weak and moderate NLC is favored when background conditions improve for NLC formation, whereas strong NLC benefits less. Trends of ice cloud parameters are generally smaller in the Southern Hemisphere (SH) compared to the Northern Hemisphere (NH), consistent with observations. Trends in background conditions have counteracting effects on NLC: temperature trends would suggest stronger ice increase in the SH, and water vapor trends would suggest a weaker increase. Larger trends in NLC brightness or occurrence rates are not necessarily associated with larger (more negative) temperature trends. They can also be caused by larger trends of water vapor caused by larger freeze drying, which in turn can be caused by generally lower temperatures and/or more background water. Trends of NLC brightness and occurrence rates decrease with decreasing latitude in both hemispheres. The latitudinal variation of these trends is primarily determined by induced water vapor trends. Trends in NLC altitudes are generally small. Stratospheric temperature trends vary differently with altitude in the NH and SH but add up to similar trends at mesospheric cloud heights.


1. Introduction

[2] Ice clouds in the summer mesopause region are very sensitive to the background status of the UMLT (upper mesosphere/lower thermosphere) and are considered to be indicators of long term trends caused by anthropogenic increase of greenhouse gases. These clouds are labeled ‘noctilucent clouds’ (NLC) if observed from the ground by lidar or by naked eye, and ‘polar mesospheric clouds’ (PMC) if observed from satellites. Since both phenomena rely on the same geophysical quantity (namely ice particles), we will occasionally use ‘NLC’ referring to both NLC and PMC.

[3] The longest record of PMC observations (28 years) comes from SBUV instruments (Solar Backscatter in the Ultraviolet) on various satellites. The data set has been intensively analyzed for trends and solar cycle variations (see DeLand et al. [2007] and Shettle et al. [2009] for some recent results). A solar cycle modulation and an increase of PMC albedo and occurrence rates was identified. The magnitude of the effects observed by SBUV increases with latitude which asks for trend studies at polar latitudes. Interhemispheric differences in ice layer morphology have been detected by SBUV and by other satellites [see, e.g., Petelina et al., 2006; Bailey et al., 2007; Hervig and Siskind, 2006]. Lidars and radars have also observed similar differences between the northern (NH) and the Southern Hemisphere (SH) [see, e.g., Chu et al., 2004; Morris et al., 2004; Lübken et al., 2009a]. Whether or not ice layers show trends is disputed in the literature [von Zahn, 2003; Thomas et al., 2003]. Some analysis of the same SBUV data set indeed shows a very small trend [see, e.g., Stevens et al., 2007, Figure 5]. A better
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In addition, the background water vapor which thereby leads to a redistribution of H$_2$O known as ‘freeze drying.’ In addition, the background water vapor is photolyzed applying actual Lyman-alpha fluxes updated once per day. Solar activity is expressed in terms of a daily Lyman-alpha flux time series from January 1960 until today which is regularly updated from the University of Colorado (ftp://lasftp.colorado.edu/pub/SEE-DATA/compositelya). The absorption in the solar Lyman-alpha line (121.6 nm, photolysis of H$_2$O and O$_3$) is calculated following Chabrillat and Kockarts [1997, 1998]. Various comparisons of results from LIMA/ice with ice cloud observations from lidars and from satellites have shown satisfactory agreement [Berger and Lübken, 2006; Lübken et al., 2008, 2009a]. Some comparison of water vapor profiles from LIMA with observations has been presented by LBB09 and Hartogh et al. [2010]. We realize that there are still some differ-

2. LIMA/Ice Model

[5] The LIMA model (Leibniz-Institute Middle Atmosphere model) is a general circulation model of the middle atmosphere which especially aims to represent the thermal structure around mesopause altitudes [Berger, 2008]. LIMA is a fully nonlinear, global, and three-dimensional Eulerian grid point model which extends from the ground to the lower thermosphere (0–150 km) taking into account major processes of radiation, chemistry, and transport. LIMA applies a triangular horizontal grid structure with 41,804 grid points in every horizontal layer ($\Delta x \sim \Delta y \sim 110$ km) and adapts to tropospheric and lower stratospheric data from ECMWF. This introduces short-term and year-to-year variability which influences, among other parameters, ice formation. In Figure 1, we show temperature trends in LIMA from the troposphere and lower stratosphere in comparison with a recent compilation of radio sonde analysis [Randel et al., 2009]. For this intercomparison the trend calculations in LIMA were restricted to the same latitude bands (60–90°N/S) and periods (1979–2007) as the radio sonde analysis. As can be seen from Figure 1, there is general agreement between LIMA and the radio sonde results. This confirms that ECMWF is reflecting real measurements and is correctly incorporated into LIMA.

[6] A 3-D Lagrangian ice transport model is superimposed on LIMA to study the formation and life cycle of ice particles in the polar mesopause region. The main idea and some recent improvements are described in more detail by LBB09. To cover the entire summer season LIMA/ice runs from 25 May to 15 August in the Northern Hemisphere, and from 25 November to 15 February in the Southern Hemisphere. Forty million condensation nuclei are transported in LIMA according to background winds, particle eddy diffusion, and sedimentation. Ice particles are formed if conditions are favorable. The formation and sublimation of ice is interactively coupled to background water vapor which thereby leads to a redistribution of H$_2$O known as ‘freeze drying.’ In addition, the background water vapor is photolyzed applying actual Lyman-alpha fluxes updated once per day. Solar activity is expressed in terms of a daily Lyman-alpha flux time series from January 1960 until today which is regularly updated from the University of Colorado (ftp://lasftp.colorado.edu/pub/SEE-DATA/compositelya). The absorption in the solar Lyman-alpha line (121.6 nm, photolysis of H$_2$O and O$_3$) is calculated following Chabrillat and Kockarts [1997, 1998]. Various comparisons of results from LIMA/ice with ice cloud observations from lidars and from satellites have shown satisfactory agreement [Berger and Lübken, 2006; Lübken et al., 2008, 2009a]. Some comparison of water vapor profiles from LIMA with observations has been presented by LBB09 and Hartogh et al. [2010]. We realize that there are still some differ-

Figure 1. Tropospheric and stratospheric temperature trends in the summer at high latitudes during the period 1979–2007. (top) Northern Hemisphere (60°N–90°N) and (bottom) Southern Hemisphere (60°S–90°S). Trends from radiosonde data (colored lines) are taken from Randel et al. [2009]. Black lines are trends used in LIMA in the same latitude bands and time periods as the radiosonde data.
Zonal mean occurrence rates (in percent) for ice clouds in the NH with a backscatter coefficient $\beta > 1 \cdot 10^{-10}/(m \cdot sr)$. Occurrence rates are determined for a time period of 41 days around central season (= 9 July). The straight line is fitted to the points representing an occurrence rate of 50%. The red dots indicate the southernmost extension of an ice cloud appearing at a single longitude.

Figure 2. Zonal mean occurrence rates (in percent) for ice clouds in the NH with a backscatter coefficient $\beta > 1 \cdot 10^{-10}/(m \cdot sr)$. Occurrence rates are determined for a time period of 41 days around central season (= 9 July). The straight line is fitted to the points representing an occurrence rate of 50%. The red dots indicate the southernmost extension of an ice cloud appearing at a single longitude.

ences between LIMA/ice and observations. For example, the PMC season in LIMA/ice starts ~5 days later than observed by SNOE [Bailey et al., 2007]. Since the analysis in our paper is mainly based on seasonal mean values our main results are not affected by such a delay.

[7] LIMA/ice results in the NH (1961–2009) and in the SH (1962–2009) are now available for 49 and 48 years, respectively (we label the SH season from, for example, November 1961 until February 1962 as ‘1962’). This allows us to study climatological behavior of NLC parameters. It is important to realize that we have used the same H$_2$O profile for all years at the beginning of the season (25 May and 25 November). This profile is then exposed to varying Ly-\(\alpha\) radiation, transport, and redistribution by freeze drying. The input water vapor profile at the beginning of the season is constant in time and does not exhibit any trends nor solar cycle variation. The observed solar cycle variations in H$_2$O are induced by varying photodissociation by Ly-\(\alpha\) radiation. As we will see later, we also observe trends in water vapor in the upper mesosphere in LIMA/ice. These are caused by temperature trends affecting the effectiveness of freeze drying. We therefore call them ‘induced water vapor trends.’

[8] When calculating mean values for ice parameters, we have used 41 days in the main season, namely 19 June to 29 July for the NH and 12 December to 21 January in the SH. This definition of ‘main season’ is centered at 9 July and 1 January, which follows the seasonal distribution of ice clouds in LIMA. Our definition is in agreement with seasonal variation of NLC and PMSE. For example, measurements from SNOE clearly indicate a later start of the PMC season in LIMA/ice starts from, for example, a detailed comparison with visual observations [see, e.g., Kirkwood et al., 2008]. Figure 3 of LBB09 gives a more detailed picture of occurrence rates varying with season and latitude. It is obvious from Figure 2 that occurrence rates generally increase in time for a fixed latitude, at least poleward of ~65°N. Equivalently, the latitude of a particular occurrence rate (e.g., 50%) decreases in time, here from ~80°N in the early 1960s to ~70°N at present. We also show the southernmost extension of the cloud in the season since some speculations about a southward extension are presented in the literature [see, e.g., Wickwar et al., 2002]. Our model does not show an unequivocal trend for ice clouds extending farther southward with time.

[9] The LIMA/ice model gives backscatter coefficients $\beta$ at a wavelength of $\lambda = 532$ nm in units of $10^{-10}/(m \cdot sr)$. In addition, we have determined albedo by Mie scattering calculations using the wavelength of SBUV instruments ($\lambda = 252$ nm) and typical scattering angles $\vartheta$ for a total of 7279 scattering segments appearing in LIMA. These segments are taken from a total of five single snapshots of the ice cloud as shown, for example, in Figure 2 of Berger and Lübken [2006]. We then determine albedos and $\beta$ in all segments given by latitude and longitude sections. The five snapshots are taken from the center of the PMC season where a mixture of weak and strong PMC occurs. In Figure 3 a comparison of maximum backscatter coefficients and albedo is shown for $\vartheta = 120°$. Although this is a typical value for SBUV, we note that scattering angles may vary substantially with season, from year to year, and from one instrument to another. From Figure 3, we have deduced a common function to convert $\beta_{\text{max}}$ values to albedo. We refrain from calculating albedo for each individual cloud in LIMA/ice since too many data points are involved (several trillion; see below) and detailed information required from SBUV instruments regarding
scattering angles, observation periods, latitudes, etc., is not readily available.

[12] In Figure 4, we show albedo based on $\beta_{\text{max}}$ values determined in LIMA/ice. We have grouped and averaged the albedo in certain latitude bins, namely 56°N–64°N, 65°N–74°N, and 74°N–82°N. These bins are chosen according to the selection in the data analysis presented by DeLand et al. [2007] and Shettle et al. [2009]. The sensitivity of SBUV instruments to detect PMC decreases with increasing latitude; that is, the albedo threshold increases with increasing latitude. For the latitude bands given above, typical albedo thresholds are $5.5 \times 10^{-6}$, $6.5 \times 10^{-6}$, and $7.5 \times 10^{-6}$ [DeLand et al., 2007]. Applying the conversion function introduced above (solid line in Figure 3), this corresponds to $\beta_{\text{max}} = 10.37, 12.52,$ and $14.74 \times 10^{-10}/(\text{m} \cdot \text{sr})$. The procedure to determine the albedo shown in Figure 4 is as follows: we start by calculating backscatter coefficients at every location (latitude, longitude, altitude) and at every time step, i.e., once per hour. Daily mean values are calculated using only those time bins where the backscatter coefficients exceed the $\beta$ thresholds mentioned above. Further averaging is performed over all longitudes, over all latitudes within a certain band (see above) and for the main NLC season from 19 June to 29 July. Finally, mean $\beta$ values are converted to albedo. This procedure aims at simulating the sampling procedure used for SBUV. We have performed a multiple regression analysis to the points shown in Figure 4 including a solar cycle signal and a linear trend. Considering the entire period from 1961 until 2008, this procedure results in trends of 0.030, 0.069, and 0.062 (units: $10^{-6}$/yr). This is in nice agreement with albedo trends derived from SBUV, namely 0.037, 0.063, and 0.067 (same units as above) [see DeLand et al., 2007, Table 4]. However, if we consider the same time period as SBUV, namely 1979–2006, we arrive at considerably smaller trends, namely 0.001, 0.044, and 0.026. Potential implications are discussed in section 5. Regarding solar cycle variations the fit in Figure 4 occasionally deviates substantially from the variations observed by LIMA/ice. We note that the current version of LIMA/ice considers solar cycle effects through photodissociation of water vapor only, whereas solar cycle effects on temperatures and winds are not yet taken into account. We expect that atmospheric heating during solar maximum will reduce some of the large albedos from LIMA/ice as shown in Figure 4.

[13] We have applied the albedo trend analysis in the 3 latitude bins and time periods mentioned above for various thresholds. The results are summarized in Table 1 and are shown in Figure 5. As mentioned above $\beta$ thresholds can be converted to albedo thresholds (and vice versa) applying the conversion function shown in Figure 3. We prefer to use mainly $\beta$ thresholds in our discussion since this is the prime quantity derived in LIMA/ice. We note that an increase of the threshold for $\beta$ decreases the occurrence of PMC...
is obvious from this plot that temperatures decrease until ~1995 but increase thereafter, similar to the NH. We have therefore determined trends in two periods: 1962–2009 and 1962–1995 in the Southern Hemisphere (1961–2009 and 1961–1994 in the Northern Hemisphere). The results for temperatures, water vapor, and NLC parameters are listed in Table 2. From the values listed in Table 2 it can be seen that in the SH short period NLC trends are significantly larger compared to long period trends, whereas the short/long period difference is generally smaller in the NH. Considering background trends, short/long period trends are less different between both hemispheres. We note that the mean temperature in 1976 is considerably higher compared to the general trend which is due to a bias in satellite temperatures in the stratosphere [Gleisner et al., 2005]. The fact that a bias in ECMWF temperatures leads to a clear bias in upper mesosphere temperatures (and ice layers) demonstrates the importance of the stratosphere for ice layers at mesopause altitudes. We noted (LBB09) that a similar bias exists in the NH for the years 1975 and 1976.

In Figure 8, we summarize trends of NLC parameters for 3 latitudes in the NH, and 2 latitudes in the SH. We do not show results at 54°S since there are hardly any NLC there. We show trends from the NH period 1961–1994 (SH: 1962–1995) but trends for the entire period are also considered and are also listed in Table 2. Trends in NLC parameters are closely related to trends in background parameters, most importantly by temperatures and water vapor. For comparison we therefore show trends of these parameters at typical NLC heights in Figure 9. The small arrows in the bottom of Figures 8 and 9 labeled ‘less ice’ and ‘more ice’ indicate where trends shown in these plots are associated with less and more ice, respectively. For example, brighter clouds (blue dots in Figure 8) and lower temperatures (red dots in Figure 9) are associated with more ice. As noted before, water vapor trends are indirectly

4. Interhemispheric Differences of Trends

In Figure 6, we show the seasonal and latitudinal extent of ice layers in the Southern Hemisphere, more precisely mean occurrence rates per day in percent as a function of latitude and time of year. A similar plot for the Northern Hemisphere and a definition of ‘occurrence rates’ is provided by LBB09. A total of 20 trillion single data points contribute to this plot, namely 40 million particles × 24 hours × 80 days × 48 years × 3 positions × 2 parameters (radii) × 2 hemispheres = 20 × 1012. As can be seen from Figure 6, there is some year-to-year variability in ice layer occurrence rates. This concerns not only occurrence rates but basically all ice layer parameters, namely altitude, brightness, mean radius, latitudinal extent, etc. There is a general tendency for occurrence rates to increase over the years.

We have determined trends of other NLC parameters, namely mean altitude of the NLC peak and brightness, where the latter is a synonym for the backscatter coefficient defined above. Again, zonal mean values of all events detected in the central NLC season are considered. As has been noted by LBB09 that some trends are not uniform in the entire period (1961–2009) but change around 1994. In Figure 7, we show zonal mean temperatures at 83 km at 69°S averaged in the period 12 December to 21 January. It

![Figure 5. Linear trends in albedo in units of 10-6/yr for the three latitude bands used by SBUV (see top right). Thresholds on the lower abscissa are given for backscatter coefficients in units of 10-10/(m · sr) and are converted to albedo (upper abscissa) using the mean curve from Figure 3. A rough approximation is β ~ 1.9× albedo (in units given in DeLand et al. 2007) at thresholds relevant for SBUV (vertical lines). Trends are given for a long period (1961–2008, squares and solid lines) and for a short period (crosses and dotted lines). The values are listed in Table 1.](image)

![Table 1. Linear Trends of Mesospheric Ice Cloud Albedo\(^a\)](image)

<table>
<thead>
<tr>
<th>Threshold</th>
<th>Period</th>
<th>NH:56°–64°N</th>
<th>NH:64°–74°N</th>
<th>NH:74°–82°N</th>
</tr>
</thead>
<tbody>
<tr>
<td>SBUV(^b)</td>
<td>1979–2006</td>
<td>0.037</td>
<td>0.063</td>
<td>0.067</td>
</tr>
<tr>
<td>βₘᵟᵣ 1</td>
<td>1979–2006</td>
<td>0.0158</td>
<td>0.0548</td>
<td>0.0766</td>
</tr>
<tr>
<td>βₘᵟᵣ 2</td>
<td>1979–2006</td>
<td>0.0194</td>
<td>0.0562</td>
<td>0.0785</td>
</tr>
<tr>
<td>βₘᵟᵣ 4</td>
<td>1979–2006</td>
<td>0.0273</td>
<td>0.0542</td>
<td>0.0784</td>
</tr>
<tr>
<td>βₘᵟᵣ 8</td>
<td>1979–2006</td>
<td>0.0302</td>
<td>0.0507</td>
<td>0.0730</td>
</tr>
<tr>
<td>βₘᵟᵣ 10.3</td>
<td>1979–2006</td>
<td>0.0015</td>
<td>0.00474</td>
<td>0.0689</td>
</tr>
<tr>
<td>βₘᵟᵣ 12.5</td>
<td>1979–2006</td>
<td>−0.0014</td>
<td>0.0439</td>
<td>0.0428</td>
</tr>
<tr>
<td>βₘᵟᵣ 13</td>
<td>1979–2006</td>
<td>−0.0024</td>
<td>0.0429</td>
<td>0.0419</td>
</tr>
<tr>
<td>βₘᵟᵣ 14.7</td>
<td>1979–2006</td>
<td>−0.0064</td>
<td>0.0405</td>
<td>0.0258</td>
</tr>
</tbody>
</table>

\(^a\)Units of albedo trends are 10⁻⁶/yr. Linear trends are taken from a fit which considers solar cycle variations. SBUV, Solar Backscatter in the Ultraviolet.

\(^b\)From Table 4 of DeLand et al. [2007]. Thresholds are given for backscatter coefficients, βₘᵟᵣ, in units of 10⁻¹⁰/(m · sr) and are converted to albedo using the mean curve from Figure 3. A rough approximation is β ~ 1.9× albedo, in units given in Figure 3.
Figure 6. Ice cloud distribution in the SH from 1962 to 2009. Note the following nomenclature for the SH: for example, “1995” refers to period from end of 1994 to beginning of 1995.
induced by temperature trends affecting the effectiveness of freeze drying.

5. Discussion and Summary

5.1. Albedo Trends, Detection Sensitivity, and Comparison With SBUV

[17] In Figures 4 and 5, we have shown albedo trends from LIMA/ice for two time periods in three latitudes bands applying various detection thresholds. When comparing these results with SBUV trends, we should note that several parameters influencing albedo cannot precisely be taken into account in our model. This concerns, for example, the variation of SBUV sampling with season and with latitude, the actual scattering angle for each SBUV instrument, etc. We note that tidal variations of ice clouds may influence PMC statistics as derived from SBUV [DeLand et al., 2007]. However, very little is known about tidal modulations of PMC properties and their variation with latitude. We have therefore not made an attempt to consider such an effect in our analysis.

[18] The trends shown in Figure 4 basically agree with SBUV but only if the long time period is considered. Considering the short SBUV period (1979–2006) LIMA/ice trends are generally smaller. This implies that trends are not uniform with time which is consistent with the time series of temperatures, etc. shown by LBB09 and in Figure 7. We note that the short period trends may have a tendency for too small values since the end of that period collocates with solar maximum and correspondingly small NLC values. As can be seen from Figure 5 trends for the short period are smaller compared to the long period, basically independent of thresholds (at least for low and moderate thresholds). At high thresholds, trends show a tendency to decrease. This implies that weak and moderate NLC are favored when background conditions improve for NLC formation, whereas strong NLC benefit less. Figure 5 also shows that trends increase with latitude but the difference between the two polar latitude bands disappears for large thresholds. This implies that the similarity in SBUV trends at high and

Table 2. Trends of Ice Layer and Background Parameters

<table>
<thead>
<tr>
<th>Period</th>
<th>Altitude (km/yr)</th>
<th>Occurrence (%/yr)</th>
<th>Brightness (10^{-10} (m · sr)/yr)</th>
<th>T (83 km) (K/yr)</th>
<th>H₂O² (ppm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>78°N</td>
<td>1961–2009</td>
<td>−0.019</td>
<td>0.667</td>
<td>0.200</td>
<td>−0.051</td>
</tr>
<tr>
<td>1961–1994</td>
<td>0.23</td>
<td>0.649</td>
<td></td>
<td>0.249</td>
<td>−0.070</td>
</tr>
<tr>
<td>69°N</td>
<td>1961–2009</td>
<td>−0.015</td>
<td>0.590</td>
<td>0.121</td>
<td>−0.051</td>
</tr>
<tr>
<td>1961–1994</td>
<td>−0.018</td>
<td>0.605</td>
<td></td>
<td>0.120</td>
<td>−0.076</td>
</tr>
<tr>
<td>54°N</td>
<td>1961–2009</td>
<td>−0.0030</td>
<td>0.0088</td>
<td>0.017</td>
<td>−0.036</td>
</tr>
<tr>
<td>1961–1994</td>
<td>−0.0049</td>
<td>0.0048</td>
<td></td>
<td>0.013</td>
<td>−0.049</td>
</tr>
<tr>
<td>78°S</td>
<td>1962–2009</td>
<td>−0.019</td>
<td>0.572</td>
<td>0.066</td>
<td>−0.076</td>
</tr>
<tr>
<td>1962–1995</td>
<td>−0.029</td>
<td>1.154</td>
<td></td>
<td>0.183</td>
<td>−0.112</td>
</tr>
<tr>
<td>69°S</td>
<td>1962–2009</td>
<td>−0.015</td>
<td>0.219</td>
<td>0.035</td>
<td>−0.076</td>
</tr>
<tr>
<td>1962–1995</td>
<td>−0.026</td>
<td>0.544</td>
<td></td>
<td>0.095</td>
<td>−0.123</td>
</tr>
</tbody>
</table>

*Trends of water vapor averaged in the period 1–10 July (NH) and 1–10 January (SH).
very high polar latitudes is at least partly caused by trends being reduced at very high polar latitudes and large thresholds.

[19] We repeat that we do not include mesospheric greenhouse gas trends in the current version of LIMA/ice. This may at least contribute to the difference between SBUV and LIMA/ice trends observed for the short time period. Including mesospheric greenhouse gas trends will presumably increase the trends shown in Figure 5.

5.2. Comparison of Ice Layer and Background Trends in NH and SH

[20] Considering the change of trends in the mid 1990s (see above), we concentrate on trends from the periods 1961–1994 in the NH and 1962–1995 in the SH. A summary of the results is presented in Figures 8 and 9. The abscissas are organized such that more ice (or lower NLC altitudes) is found in Figures 8 (left) and 9 (left), whereas less ice (or higher NLC altitudes) is found in Figures 8 (right) and 9 (right).

[21] Ice layer trends in both hemispheres decrease with decreasing latitudes, consistent with Figure 2. This is in agreement with SBUV observations, at least in the NH [DeLand et al., 2007]. Trends are very small at low latitudes in the NH. This is consistent with visual NLC observations performed at mid latitudes in the NH which indeed show negligible trends [Dalin et al., 2006; Kirkwood et al., 2008]. The variation of trends with latitude is similar for all three parameters considered here, namely brightness, occurrence rates, and altitudes. Trends of all three parameters decrease with decreasing latitude, both in the NH and SH. Trends in NLC altitudes are generally rather small, in particular at low latitudes. For example, at 54°N the NLC altitude changes by only ~150 m in the period 1961–1994. Figure 8 shows that brightness and occurrence trends tend to be larger in the NH compared to equivalent latitudes in the SH. SBUV albedo trends are also larger at 78°N compared to 78°S whereas the values at 69°N/S are rather similar [see DeLand et al., 2007, Table 4]. Interestingly, NLC altitude trends are smaller (i.e., less negative) in the NH compared to equivalent latitudes in the SH.

[22] In Figure 9, we show trends of temperatures at typical NLC altitudes (83 km) and of water vapor at the NLC peak. Temperature trends are generally larger (more negative) in the SH compared to the same latitudes in the NH. For water vapor it is the other way around: trends are generally smaller in the SH compared to NH. We recall that water vapor trends in LIMA are induced by the combination of temperature trends and freeze drying. In Figure 10, we show the

Figure 9. Trend of background conditions at NLC altitudes (83 km) at various latitudes in the northern and Southern Hemispheres as indicated on the ordinate. Red circles are temperature, blue diamonds are water vapor concentration. In the NH, only the years from 1961 until 1994 were considered in the linear fit (SH: 1962–1995), and the years 1975–1976 (SH: 1976) were ignored. Error bars were calculated considering variability of the data points.

Figure 10. Water vapor profiles in the (top) Northern Hemisphere (69°N) and (bottom) Southern Hemisphere (69°S) from before (green lines) and within (blue lines) the NLC season. The exact dates are 30–31 May and 1–10 July in the NH and 29–30 November and 1–10 January in the SH. Each colored line represents one NLC season. The black lines are the total means over the years 1961–1994 in the NH and 1962–1995 in the SH.
effect of freeze drying on the distribution of water vapor at 69°N/S by comparing profiles before and in the middle of the NLC season. The green profiles from before the NLC season are basically identical, whereas the blue profiles from within the NLC season vary due to freeze drying which depends on temperatures and also on the amount of water covered in the ice cloud. In more detail, the water vapor profiles are identical on 25 May, 0000 UT (NH) and 25 November, 0000 UT (SH). The green profiles shown in Figure 10 are taken from ~6 days later when year-to-year variation of dynamics and photodissociation by Ly-α has already introduced some small variability (see also Figure 6 of LBB09).

As can be seen from Figure 10, freeze drying is generally smaller in the SH compared to the NH. This is partly caused by temperatures which are generally larger in the SH and therefore lead to less water covered in the ice cloud. However, the amount of water vapor collected by ice particles during growth and released during sublimation is not simply proportional to temperature but depends on various parameters, e.g., the mean background water vapor concentration, the circulation, the effectiveness of forming ice. Regarding ice formation the hemispheric difference in water vapor trends counteracts the difference in temperatures trends: temperature trends would suggest stronger ice increase in the SH, water vapor trends a weaker increase. [25] As can be seen from Figure 10, freeze drying is generally smaller in the SH compared to the NH. This is partly caused by temperatures which are generally larger in the SH and therefore lead to less water covered in the ice cloud. However, the amount of water vapor collected by ice particles during growth and released during sublimation is not simply proportional to temperature but depends on various parameters, e.g., the mean background water vapor concentration, the circulation, the effectiveness of forming ice. Regarding ice formation the hemispheric difference in water vapor trends counteracts the difference in temperatures trends: temperature trends would suggest stronger ice increase in the SH, water vapor trends a weaker increase.

[24] Comparing latitude variations in ice layer trends and background trends (including interhemispheric variations) there are obviously similar structures in Figures 8 and 9. Induced water vapor trends show similar variations to brightness variations (and also occurrence trend variations, not highlighted). Temperature trends show similar variations to NLC altitude trends. Roughly speaking and ignoring details one can summarize as follows: water vapor trends cause trends in brightness and occurrence rates, whereas temperature trends cause trends in NLC altitudes. This also explains the latitudinal variation of NLC brightness and occurrence rates since water vapor trends clearly increase with latitude whereas temperature trends are less decisive.

[25] When comparing trends in NLC and background parameters, we should keep in mind that the mean values are different in both hemispheres and also vary with latitude. As has been shown by Lübken and Berger [2007] in a NH/SH comparison applying LIMA/ice, mean summer season temperatures at NLC altitudes are slightly larger in the SH compared to the NH which is enough to affect ice formation. The temperature difference is only 2–3 K, which is in agreement with falling sphere measurements at 69°N and 68°S [Lübken, 1999; Lübken et al., 2004]. The same temperature trends can therefore result in rather different effects on NLC, depending on the mean value. We recall in Figure 11 that ice particle growth can vary highly non-linearly with temperature and furthermore depends on ice particle radius, water vapor, etc. For example, the growth rate of an ice particle with a radius of 10 nm at an altitude of 85 km with a constant water vapor concentration of 3 ppmv nearly doubles if its temperature decreases from 139 K to 137 K, whereas it does not increase any further once temperatures are below ~130 K. Coming back to the NH/SH difference in mean temperatures, there is indeed significant evidence for less intense and frequent PMC in the SH compared to NH [Chu et al., 2006; Bailey et al., 2007].

[26] For some NLC parameters the change of trends with latitude in the NH seems to saturate for very high latitudes. This concerns, for example, occurrence rate trends at 69°N and 78°N. This is presumably caused by saturation effects in ice formation: if temperatures are very low the potential for forming NLC cannot be increased any further. For example, whereas particles generally grow faster if temperatures decrease, the growth rate is constant for temperatures smaller than approximately 130 K (see Figure 11). Furthermore, if occurrence rates are close to 100% they cannot increase any further. On the other hand, an increase of water vapor will tend to increase particle size and therefore brightness with little tendency for saturation.

[27] We summarize that larger trends in ice layer brightness or occurrence rates are not necessarily associated with larger (more negative) trends in temperatures. They can also be caused by larger trends of water vapor caused by larger freeze drying, which in turn can be caused by generally lower temperatures and/or more background water.

5.3. Stratospheric Origin for Ice Layer Trends

[28] It is important to remember that the trends shown in this paper do not include any potential impact of greenhouse gas trends in the upper stratosphere and entire mesosphere. Greenhouse gas concentrations in LIMA are kept constant (regarding water vapor: see above). However, any indirect influence of greenhouse gas trends below approximately 35 km (e.g., a cooling and shrinking of the stratosphere) is indirectly considered due to the adaption of LIMA to ECMWF described above. We have studied stratospheric influences on the UMLT and on PMC in detail by LBB09 and have shown that there is a strong link between trends in the stratosphere and trends in upper mesosphere temperatures. Roughly speaking, a cooling in the stratosphere leads to a shrinking and cooling at PMC altitudes and a
The subsequent intensification of PMC. It is tempting to associate the change of temperature trends in the mid 1990s to the ozone recovery which also started at the same time [World Meteorological Organization, 2007]. Unfortunately, little is known about ozone trends in the polar summer stratosphere and their potential role in temperature trends [Staehelin et al., 2001; Ramaswamy et al., 2001].

To elucidate the role of the stratosphere and its interhemispheric differences, we show in Figure 12 temperature trends for the entire atmosphere up to 120 km as a function of geometric altitude. As can be seen from Figure 12, trends in the stratosphere show a rather complicated variation with height and may be larger in the SH compared to the NH at particular altitudes (see also Figure 1). The integrated effect due to shrinking is indicated by trends of geometric heights and the associated temperature variation. It turns out that at NLC heights of (~83 km) the temperature trend is larger (more negative) in the SH compared to the NH (see also Figure 9). However, the contribution of stratospheric shrinking is nearly the same in both hemispheres which implies that other effects must be different. Since there are no trace gas trends in the mesosphere in the current version of LIMA the only remaining process is dynamical feedback, for example, a NH/SH difference of wave activity and/or wave filtering (see Berger [2008] for more details).

A recent update of temperature observations suggests that lower stratospheric trends in the summer season at polar latitudes are larger in the SH compared to the NH, but again varying substantially with altitude (see profiles from Randel et al. [2009] reproduced here in Figure 1). Interestingly, this compilation also shows a change of trends in the mid 1990s similar to what we have shown in Figure 7. This confirms the close connection between trends in mesospheric ice layers and stratospheric temperature trends. This also implies that the magnitude of trends (and in some cases even their sign) depends on the period chosen. Therefore, any comparison of trends must carefully consider the time period chosen for deducing trends. Furthermore, the change of trends with time means that applying too short time series in PMC trend analysis may result in spurious results. In addition, a rather large cyclic variation caused by, for example, solar cycle further complicates trend analysis if time series are too short.

There is evidence from theoretical and experimental studies that an active winter stratosphere may reduce ice layers in the summer hemisphere through interhemispheric coupling mechanisms [Becker and Schmitz, 2003; Karlsson et al., 2007; Körnich and Becker, 2010]. This effect was studied in detail in the NH summer of 2002 when the planetary wave activity in the SH stratosphere was unexpectedly high and, serendipitously, a comprehensive ground based and sounding rocket campaign took place in Northern Norway (see summary by Goldberg et al. [2004]). Indeed, LIMA also reproduced low ice layer activity in the NH summer of 2002 (see Figure 3 of LBB09). We should therefore expect that an enhanced planetary wave activity in the winter stratosphere during, for example, a sudden stratospheric warming will influence ice layers in the SH summer mesopause region. A detailed study of interhemispheric coupling effects in LIMA/ice is beyond the scope of this paper and will be done in the near future. A brief comparison did not show significantly less ice layers in the SH during NH major or Canadian stratospheric warmings. We note that enhanced planetary wave activity appearing stochastically over a time period of 30–40 years would not influence trends. There are some indications from trends in zonally asymmetric ozone that planetary wave activity in the winter stratosphere may have increased in the last decades [Gabriel and Schmitz, 2002]. However, it is not

Figure 12. Trends of temperatures (black lines) and geometrical heights (red lines) in the (top) Northern Hemisphere (69°N) and (bottom) Southern Hemisphere (69°S). The blue lines indicate the temperature trends in the mesosphere due to shrinking of the atmosphere below. Only the years from 1961 until 1994 (in the NH; SH: 1962–1995) were considered in the linear fit, and the year 1976 was ignored. Error bars indicate the uncertainty of the linear fit. The profiles were smoothed with a three-point running mean.
clear at the moment whether or not such an increase could influence ice layers in the summer hemisphere.

5.4. Summary

[32] In summary, this analysis confirms earlier results from LBB09 that stratospheric shrinking may influence temperatures at NLC heights and thereby the morphology of ice layers. We repeat that the current version of LIMA/ice does not include greenhouse gas trends in the mesosphere, whereas trends in the troposphere and stratosphere are implicitly taken into account. Ice layer trends decrease in magnitude with decreasing latitude, both in the NH and SH. In the NH, the southernmost extension of NLC does not show a significant trend which is consistent with ground based observations of NLC at midlatitudes. Temperature trends in the stratosphere (and therefore also in the mesosphere) are not uniform but change sign around 1995, both in the NH and SH. It is tempting to associate this effect to changing ozone trends and volcano eruptions appearing in the mid 1990s. In LIMA/ice any trend of temperatures at NLC heights induces a trend in water vapor due to freeze drying. This is an important mechanism which should be included in all models dealing with trends in the summer UMLT region. The latitudinal variation of water vapor trends determines the latitudinal variation of NLC brightness and occurrence rate trends.

[33] Even without mesospheric greenhouse gas trends, LIMA/ice reproduces trends in albedo observed by SBUV, but only if a time period is considered which is longer than covered by SBUV. Large albedo thresholds as applied in SBUV can influence trends, in particular at high polar latitudes.

[34] Interhemispheric comparison shows smaller trends of brightness and occurrence rates at equivalent latitudes in the SH compared to NH, and larger trends (more negative) of NLC altitudes in the SH. Very similar characteristics are found in induced trends of water vapor (smaller in SH compared to NH) and temperatures (larger in SH compared to NH). This implies that although temperature trends are larger in the SH, NLC brightness and occurrence rates are smaller. When comparing NLC trends with background trends, the absolute values have to be taken into account. For example, mean temperatures at NLC heights are only 2–3 K larger in the SH compared to the NH (consistent with observation) which is large enough to create significant NH/SW difference in NLC morphology.

[35] Trends in the stratosphere vary with altitude and are generally different in both hemispheres. However, the overall effect on the mesosphere due to shrinking is rather similar. This implies that larger temperature trends in the SH observed in LIMA/ice (and associated effects on NLC) are presumably caused by dynamical effects. In the future, we will run the LIMA/ice model including solar cycle influence on mesospheric temperatures, and also including increase of greenhouse gases.

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