Small scale structures of NLC observed by lidar at 69°N/69°S and their possible relation to gravity waves


*Leibniz-Institute of Atmospheric Physics at the Rostock University, Schlosstraße 6, 18225 Kühlungsborn, Germany
bAustralian Antarctic Division, Kingston, Tasmania 7050, Australia

Abstract

Lidar measurements of noctilucent clouds (NLC) were conducted by the Davis Rayleigh-/Raman-lidar in Antarctica (68.58° S, 77.97° E) and by the Rayleigh-/Raman-lidar at the ALOMAR observatory in northern Norway (69.28° N, 16.01° E). We compare southern and northern hemisphere NLC at time scales of 10 min to several hours using multi-year datasets (four seasons at ALOMAR, 2008–2011, and nine seasons at Davis, 2001/2002 to 2009/2010). NLC characteristics studied include the vertical structure of NLC layers, the duration of NLC layers as well as the apparent downward motion of NLC layers with time. We find multiple layers during 9% of all NLC observations with vertical separations of double layers between 1.5 and 3 km. The mean downward progression of NLC with measurement time is ~0.3 km/h and comparable at Davis and ALOMAR. We find no general spatial tilt of the layer at ALOMAR but individual layers show up to 2 km altitude difference at 40 km horizontal separation. Typical NLC observations at both stations last about 5 hours, hinting at horizontal extents of about 700 km, and reoccur after approximately 10 hours. This is in the range of mid-frequency gravity waves (GW). On short-time scales NLC characteristics are presumably impacted by small scale processes in the vicinity of the clouds, generated by e.g. breaking GW. In addition, we discuss a possible relation to GW by looking at the influence of stratospheric wind conditions on NLC layer characteristics at 69° S.

1. Introduction

Noctilucent clouds (NLC) consist of ice particles forming in the cold summer mesopause region at polar latitudes. First observed in north-west Europe in 1885 by at least three independent observers, their intriguing fine scale structure was known from the very beginning (Leslie, 1885; Jesse, 1885; Backhouse, 1885). The first quantitative studies of the wave-like features in NLC were performed using ground-based cameras, finding wavelengths of 50 km for long parallel bands transported by wind and smaller billows with wavelengths of 5–10 km (Witt, 1962). Those structures were soon attributed to the passing of internal atmospheric gravity waves with wavelengths of 5–10 km (Witt, 1962). Those structures were soon attributed to the passage of internal atmospheric gravity waves (GW) (Hines, 1968) and the instabilities associated with the breaking of those waves in the mesopause region (Fritts et al., 1993). An extensive review of atmospheric GW is given by Fritts and Alexander (2003). The deposition of energy and momentum through the wave-breaking process is considered the main driver of the residual circulation (Holton, 1983) that causes the cold temperatures at the summer mesopause that allow for the formation of NLC in the first place. Since their routine observations from satellites these clouds are also called polar mesospheric clouds (PMC) (Thomas and McKay, 1985). Regarding long-term changes, NLC/PMC are controversially discussed as tracers for climate change (Thomas, 2003; von Zahn, 2003; Lübken et al., 2012).

The large scattering cross section of NLC particles combined with its delicate dependence on background conditions like temperature makes the phenomenon an ideal tracer for GW and small-scale processes in this region. Since Witt (1962) NLC were studied to infer GW or turbulence generated by breaking GW both theoretically and experimentally (e.g. Hines, 1968; Fritts et al., 1993; Dalin et al., 2004; Chandran et al., 2009; Baumgarten et al., 2009; Dalin et al., 2010; Chandran et al., 2010; Pautet et al., 2011). The formation of these particles is mainly controlled by temperatures below the frost point that were shown to exist at the summer mesopause using rocket measurements (Lübken, 1999). Given the presence of water vapour, ice particles nucleate and grow to sizes larger than 20 nm. NLC form thin layers with concentrations of a few to hundred/cm³ (Baumgarten et al., 2008; Hervig et al., 2009). Variations in the background atmosphere may either enhance or destroy the clouds (e.g. Rapp et al., 2002) depending on the period of the waves. Active remote sensing of NLC using lidars first succeeded in 1989 (Hansen et al., 1989). Today some lidar stations capable of sounding NLC in daylight conditions are operated at polar latitudes in both
hemispheres (Thayer et al., 2003; Stebel et al., 2000; Gerdig et al., 2010; Chu et al., 2009, 2011; Hoffner et al., 2003). NLC are also routinely observed with ground-based camera networks (e.g. Dalin et al., 2008).

For the first time, we have combined lidar NLC observations from the southern and northern hemisphere to study small-scale variations at scales from minutes to a few hours. We use multi-year datasets from the Rayleigh-/Raman-lidar at Davis station in Antarctica (69° S) and the Rayleigh-/Mie-/Raman-lidar (RMR) at the ALOMAR observatory in northern Norway (69° N) at a temporal resolution of 10 min. We analyze NLC structures at vertical, temporal and horizontal scales. First, the occurrence and properties of multiple layers are analyzed. Then the apparent descending of NLC layers that is observed by lidar as measurement time proceeds is studied. To relate this observation to spatial scales, we analyze ALOMAR RMR twin lidar measurements at two sounding volumes separated by about 40 km with the highest resolution of 30 s. Finally the duration during which NLC is observed and the time elapsing between reoccurring NLC is determined. The observations are also interpreted with respect to stratospheric wind conditions.

2. Instruments

2.1. ALOMAR RMR lidar

The ALOMAR Rayleigh-/Mie-/Raman-lidar is a twin lidar system located at 69.29° N, 16.01° E in northern Norway that has been used for noctilucent cloud research since 1994 (Nussbaumer et al., 1996). It uses two 1.8 m diameter Cassegrain telescopes that can be pointed in directions up to 30° off-zenith (named north-west telescope NWT and south-east telescope SET) to record the backscatter of two Nd:YAG lasers at 532 nm that operate with a repetition frequency of 30 Hz and an energy of \( \approx 400 \text{ mJ/pulse} \) (532 nm) (von Zahn et al., 2000). Statistics of NLC properties observed at ALOMAR and NLC retrieval algorithms are reported by Fiedler et al. (2009).

We analyze data obtained in the summer months (NLC season June 1st–August 15th) of the years 2008–2011 that were recorded with a resolution of 30 s and 40 m. The data are averaged for 10 min, smoothed by a binomial filter with 475 m FWHM and resampled to 30 s and 40 m. The data are averaged for 10 min, June 1st–August 15th) of the years 2008–2011 that were recorded reported by Fiedler et al. (2009).

We analyze data obtained in the summer months (NLC season June 1st–August 15th) of the years 2008–2011 that were recorded with a resolution of 30 s and 40 m. The data are averaged for 10 min, smoothed by a binomial filter with 475 m FWHM and resampled to 30 s and 40 m. The data are averaged for 10 min, June 1st–August 15th) of the years 2008–2011 that were recorded reported by Fiedler et al. (2009).

We analyze data obtained in the summer months (NLC season June 1st–August 15th) of the years 2008–2011 that were recorded with a resolution of 30 s and 40 m. The data are averaged for 10 min, smoothed by a binomial filter with 475 m FWHM and resampled to 30 s and 40 m. The data are averaged for 10 min, June 1st–August 15th) of the years 2008–2011 that were recorded reported by Fiedler et al. (2009).

2.2. Davis lidar

The Davis Rayleigh-/Raman-lidar is located at Davis station, 68.58° S, 77.97° E in Antarctica. It is run by the Australian Antarctic Division, and has collected mesopause region data since early 2001 (Klekociuk et al., 2003). The site is comparable to ALOMAR in latitude but different regarding orography and dynamics in the stratosphere (see Section 4). The lidar detects NLC at 532 nm with a Nd:YAG laser of 600 mJ/pulse at 532 nm and a repetition frequency of 50 Hz. The backscattered photons are collected using a 0.29 m (1.0 m before 2005) telescope with a resolution of 10 s. NLC analysis and properties are reported by Klekociuk et al. (2008). In the austral summer seasons 2001/2002 to 2009/2010 (NLC season November 15th to February 28th), 2463 h of measurements were acquired, and NLC detected during 312 h at a significance of 3.5\( \sigma \) relative to the distribution of the background. The resolution for data analysis is 10 min and 112 m.

2.3. NLC parameters

For NLC retrieval, we analyze the volume backscatter coefficient of NLC particles at altitude \( z \) that is given by

\[
\beta(z) = n(z) \sigma(z) \sin \theta = 180° / \lambda = 532 \text{ nm}
\]

where \( n \) is the number density and \( \sigma \) is the scattering cross-section of the particles at wavelength \( \lambda \) in direction \( \theta \). \( \beta \) is also termed brightness in the following. From each altitude profile \( \beta(z) \), that is at a resolution of 10 min, we derive \( \beta_{\text{peak}} = \max \beta(z) \), the corresponding altitude \( z_{\text{peak}} \) as well as the integrated column brightness

\[
\beta_{\text{int}} = \sum_{z_{\text{bottom}}} \beta(z) \; dz
\]

\( z_{\text{top}} \) and \( z_{\text{bottom}} \) denote the minimum and maximum altitudes, respectively, with significant \( \beta \). \( dz \) is the width of one altitude bin. The NLC layer is further characterized by the centroid altitude

\[
z_c = \frac{\sum_{z_{\text{bottom}}} z \beta(z)}{\sum_{z_{\text{bottom}}} \beta(z)}
\]

and the layer width \( \delta z = z_{\text{top}} - z_{\text{bottom}} \).

To study the temporal evolution of NLC layers we define the duration \( t_{\text{NLC}} \) by the time between the first significant detection of one NLC layer to the last significant detection, according to the definitions in Sections 2.1 and 2.2. The cloud-to-cloud time \( \Delta t_{\text{NLC}} \) on the other hand is the time distance between the center of two consecutive NLC. To accept a NLC layer for this analysis, we require 20 min of measurements before and after the NLC to make sure that \( t_{\text{NLC}} \) is not affected by the start or stop of the instrument. We tolerate measurement gaps up to 10 min and gaps in NLC up to 10 min.

NLC measured by lidar often exhibit an apparent descending of the layer as measurement time proceeds. We term this height-time gradient “backward progression” realizing that this motion is caused by a superposition of sedimentation and advection of NLC structures through the lidar beam. The downward progression is derived as the slope \( \alpha \) of the linear regression fit \( f(t) = \alpha t + a_0 \) to the centroid altitude \( z_c(t) \) of each brightness profile.

Multiple layers are vertically stacked layers, where in between those layers the signal falls below the detection threshold, which means that the density of NLC particles is either very low or that particles are very small or non-existent. We define a multiple layer as a lidar profile where the signal vanishes in at least one altitude bin in between the NLC layer. We count such layers starting at low altitudes and assign a minimum and maximum altitude to each layer \( l \), from which we derive the vertical extent \( \delta z_{l} = z_{\text{top},l} - z_{\text{bottom},l} \) of layer \( l \) counting from below, a peak brightness \( \beta_{\text{peak},l} \) and its corresponding altitude \( z_{\text{peak},l} \). To exclude single detections that are restricted to one time-altitude bin we additionally require that a layer lasts for at least 50 min (20 min before and after) and varies no more than two bins in altitude.

Fig. 1 shows a simplified scheme of a NLC lidar measurement indicating the parameters defined here that will be used in the next section.
3. Data analysis

3.1. Multiple layers

We first quantify the occurrence of multiple layers before continuing with the temporal evolution of NLC layers. Two datasets displaying multiple layers from Davis and ALOMAR, respectively, are shown in Fig. 2. Using the algorithm described in Section 2.3, we find that 9% of all layers are multiple at both Davis and ALOMAR. The statistics for single and multiple layers are given in Table 1. Differences between the ALOMAR NWT and SET telescopes can be due to different sampling, as there was no operation of NWT during 2011. Single layers at ALOMAR are on average about 1 km lower than at Davis and appear slightly wider (1.54 km vs. 1.06 km). The vertical extent of each of the multiple layers at both stations is slightly smaller than the vertical extent of single layers. At ALOMAR, the top layer of a double layer NLC is considerably less bright (by a factor of 2–3), which is not observed at Davis. The mean vertical distance between the two lowest multiple layers at Davis is 3.0 km, larger than at ALOMAR ($\Delta z > 1.6$ km).

To check the dependence of these results, especially the single layer width, on a possibly different sensitivity of the instruments, we re-analyze both datasets with an artificial detection threshold $b_{\text{min}}$ of $4 \times 10^{-10}$ m/sr by setting all $b < b_{\text{min}}$ to zero. This method is comparable to Fiedler et al. (2003) who use a threshold of $b > 4 \times 10^{-10}$ m/sr for long-term analysis of NLC. This method reduced the Davis and ALOMAR datasets by about 21% and 43%, respectively. The results are given in Table 2. At ALOMAR, where we already observed lower $b$ at higher altitudes, the mean layer altitude decreases by 0.4 km, increasing the difference between both locations to 1.4 km. The mean layer width however adjusts to 1 km and agrees at both locations. The occurrence of multiple layers at ALOMAR decreases to 3%, half the value compared to

Fig. 1. Schematized lidar NLC observations and derived parameters regarding vertical structures and temporal evolution of the layers. Top left: single layer parameters: brightness profile $b(z)$, artificial threshold $b_{\text{min}}$, peak brightness $b_{\text{peak}}$, peak altitude $z_{\text{peak}}$, centroid altitude $z_c$, altitude bin width $\Delta z$ and layer width $\Delta z$. Top right: multiple layer parameters for layers $L = 1,2$: $z_{\text{top},1}, z_{\text{bottom},1}, z_{\text{top},2}, z_{\text{bottom},2}$, centroid altitude $z_c$, layer widths $\Delta z$, as well as layer separation $\Delta z$. Bottom: temporal evolution of NLC layers: two consecutive layers $L = 1,2$ during a measurement of length $T$. Indicated are the linear fits for the downward progression derived over different time scales $t_{\text{fit}}$, the layer duration $t_{\text{NLC},L}$ and the cloud-to-cloud time $\Delta t_{\text{NLC}}$.

Fig. 2. Detected NLC layers at Davis (top) and ALOMAR NWT (bottom). All colors show significant detections of NLC. Multiple layers occur in these examples around 14:30 UT at Davis and 23 UT and 9 UT at ALOMAR. $z_{\text{peak}}$ is indicated by black dots. (For interpretation of the references to color in this figure caption, the reader is referred to the web version of this article.)
3.2. Downward progression

The descending of NLC layers is quantified by linear regression fits to the centroid altitude $z_c(t)$ of each brightness profile as shown in Fig. 3. This analysis is applied to single layers only. $z_c$ is used instead of $z_{peak}$ as in the multiple layer analysis because it reflects the mean altitude of a single layer more robustly than $z_{peak}$. The fit is applied to NLC events of length $t_{fit}$ with $t_{fit}$ between 30 min and 4 h. $t_{fit}$ is measured from the first observation of the NLC and repeated if the duration of the NLC is larger than $t_{fit}$.

Using an example: a NLC of 5 h duration allows for 10 fits for $t_{fit} = 0.5$ h but only one fit for $t_{fit} = 4$ h. By this method, we can determine 400 fits of length 30 min to 10 fits of length 4 h at Davis. The mean downward slope is shown in Fig. 4 as a function of $t_{fit}$ together with the number of fits that contributed to that mean value. The downward progression is $d_z = \Delta z / t_{fit}$ of both Davis and ALOMAR independent of $t_{fit}$. The deviation of the mean of 0.05–0.1 km/h accounts for wave motions that are superposed on this linear motion.

As it is known that NLC tends to be brighter at lower altitudes (e.g., Fiedler et al., 2003; Chu et al., 2006), one can expect that brightness increases with time if the layer descends on average. For each fit that was done we averaged the column-integrated brightness $\beta_{int}$ during the first hour and during the last hour. The difference of both brightness values is also shown in Fig. 4. As by design, for short-time scales the brightness difference is zero but then slightly increases with time; although this is significant only at ALOMAR.

3.3. Duration of NLC

NLC at both stations are not continuously observed but appear for a limited time, fade away and then reoccur. NLC durations $t_{NLC}$ and cloud-to-cloud times $\Delta t_{NLC}$ were derived as described in Section 2.3. Especially lasting NLC (large $t_{NLC}$) and cloud-to-cloud times can only be derived if the measurement period was sufficiently long. Experimentally, long durations are more difficult to observe because of the weather restrictions that ground-based lidars are subjected to. However 36 and 15 measurements are longer than 24 h and the longest measurement runs covered 144.3 h and 106.8 h at Davis and ALOMAR, respectively (Fig. 5).
We therefore normalize the observed durations to the length of the measurements \( T \) by

\[
OF_x = \frac{\sum f_{\text{NLC},x}}{T} \quad \forall x < T
\]

Observations of \( t_{\text{NLC}} = 1 \text{ h}, 1 \text{ h}, 2 \text{ h} \) during a 10 h measurement lead to occurrence frequencies \( OF_{1 \text{ h}} = 10\% + 10\% = 20\% \) and \( OF_{2 \text{ h}} = 20\% \). Another advantage of this definition is that \( \sum_x OF_x = OP_x \).

Histograms of \( D_{\text{NLC}} \) and \( \Delta D_{\text{NLC}} \) for Davis (total 2463 h) and ALOMAR SET (total 1495 h) are shown in Fig. 6. Here we show ALOMAR SET only as NWT and SET give the same results. The longest durations captured were 12.3 h (Davis) and 10.7 h (ALOMAR) and we find cloud-to-cloud times \( \Delta t_{\text{NLC}} \) up to 20.6 h (Davis) and 17.2 h (ALOMAR). There seems to be no preferred value for these quantities. Mean durations of NLC are 4–6 h with Davis tending to shorter durations and mean cloud-to-cloud times are 9.5 h at both stations.

4. Discussion

We have evaluated lidar measurements of NLC at two locations at conjugate latitudes. First we discuss inter-hemispheric differences and the relation of dynamical processes at global scale on the occurrence of NLC. To find evidence for intra-hemispheric coupling we study stratospheric winds and their influence on the date of winter to summer transitions in the mesopause region. Coming to smaller scales, we discuss the observation of multiple layers of NLC and typical NLC durations and compare with e.g. satellite observations. As last point, we evaluate the downward progression of NLC at both locations. To prove that this downward progression is temporal, we analyze high-resolution multi-beam NLC data obtained at ALOMAR.

Davis is located on the Antarctic coast between the ocean and the Antarctic plateau east of the Amery ice shelf in Princess Elizabeth land. ALOMAR is situated on the island of Andøya on Ramnankan mountain (380 m) roughly 100 km west of the Scandinavian mountain ridge and 2000 km east of the Greenland ice sheet. Although located on the same latitude circle (69°S), the atmospheric background conditions at the two stations, Davis in the southern hemisphere (SH) and ALOMAR in the northern hemisphere (NH), can differ regarding sources of GW, dynamics in the stratosphere and thermal structure. Especially temperature has a strong influence on NLC altitudes (e.g. Lübken et al., 2009, their Fig. 8). In the SH, NLC have been observed at higher altitudes, with lower occurrence frequencies and lower brightnesses compared to the NH (Chu et al., 2006; Bailey et al., 2007). This is in agreement with higher climatological temperatures in the mesopause region in the SH summer as observed by satellites (Xu et al., 2007) and as expected from model studies (Lübken and Berger, 2007). We also observe a lower NLC occurrence rate at Davis and find that the brightness is about 20% lower than at ALOMAR (Table 2). It was argued by Siskind et al. (2003) that the inter-hemispheric differences are mainly driven by dynamical effects rather than differences in radiative forcing due to the orbital eccentricity of the Earth orbit. The cooling of the summer mesopause region is caused by the meridional circulation which is driven by breaking GW (see e.g. review by Smith, 2012). The passage of GW from the lower to the upper atmosphere provides for intra-hemispheric coupling that is mainly modulated by the stratospheric wind conditions through GW filtering (Becker et al., 2004). Gravity waves are excited in the troposphere by orography (mountain waves), fronts or jet streams and grow in amplitude as they propagate upward in the thinning atmosphere (Hines, 1968; Tsuda et al., 1989). Eastward propagating GW are filtered by the winter eastward jet in the stratosphere, but reach mesospheric altitudes during summer. At the mesopause, these waves break due to the zonal wind reversal from westward to eastward and thereby drive the meridional flow and cool the mesopause. The SH winter stratospheric polar vortex is more stable than in the NH which also leads to changed planetary and GW forcings. Due to the late breakdown of the polar vortex in the SH, the transition to the summerly flow at mesopause altitude is significantly later than in the NH, leading to a delay of the onset of the NLC season.
Experimental evidence of this mechanism including analysis of MF radar winds and the influence on mesospheric ice particles above Davis station is reported by Morris et al. (2012). In addition to this intra-hemispheric coupling, the warmer and more variable NH winter stratosphere leads to a decreased meridional mesospheric circulation and adds to the warming of the SH summer mesopause by inter-hemispheric coupling (Karlsson et al., 2007; Murphy et al., 2012).

4.1. NLC and stratospheric winds

To study the influence of the stratospheric conditions on our NLC measurements more closely, we compare the stratospheric zonal winds from ECMWF for our two stations in the observed period. Fig. 7 shows the zonal wind at 50 hPa centered around summer solstice at both locations. Solstice is here defined as day of year 355 (December 21) at Davis and day of year 172 (June 21) at ALOMAR. At ALOMAR, the conditions are stable during the summer period with zonal winds mostly between $-5$ and 0 m/s; whereas at Davis the winter to summer transition is delayed, leading to much more variable stratospheric wind conditions. Particularly, in the observed period, there are three seasons with especially late winter to summer transitions at Davis (seasons 2001/2002, 2007/2008 and 2008/2009). This offers the opportunity to study parameters of NLC that are subjected to different GW forcings. We restrict the Davis dataset to the early summer season (November/December) and examine the years with the late transition separately. Table 3 shows the occurrence rates of NLC and Table 4 the NLC layer statistics for the seasons with late and early transitions. We find that during seasons with late transitions the NLC occurrence frequency was lower, the altitudes higher and the brightness lower compared to seasons with early transitions. Also, multiple layers seemed to occur more frequently
in years with early transition. This is in accordance with higher temperatures at NLC altitudes that would be caused by the changed GW forcing. This effect is in addition to other influences on NLC characteristics, such as mesospheric winds, tides, planetary waves and inter-hemispheric coupling.

4.2. NLC and GW

Besides the effect on overall occurrence rates, it is likely that the internal structure of NLC at hourly time scales is also influenced by GW and we discuss our findings in this regard. Experimentally, vertical wavelengths of GW can be derived from density or temperature fluctuations measured by Rayleigh lidars, however, this technique is only able to resolve GW well below NLC altitudes, unlike resonance lidars or radars (Chu et al., 2011; Tsuda et al., 1989). In this work, we therefore did not derive stratospheric GW parameters from our lidar data, but want to employ NLC characteristics to infer GW parameters, though. Rauthe et al. (2006) reported on vertical wavelengths between 5 and 50 km (peak at 20 km) in the altitude range 40–60 km for summer conditions in mid-latitudes. Studies of GW above Davis and ALOMAR are unfortunately limited to winter conditions or case studies (Alexander et al., 2011; Blum et al., 2006). Rapp et al. (2002) have shown using model studies that the temperature variation in wave crests and troughs of GW are sufficient to enhance or destroy NLC ice particles on time scales of hours depending on the period of the GW involved. Stratospheric GW activity has been directly linked to the occurrence of mesospheric clouds above Antarctic as well as Arctic stations (Thayer et al., 2003; Gerrard et al., 2004; Chu et al., 2009), however, such a link could not be established at Davis (Innis et al., 2008) and ALOMAR (Schöch, 2007). These differences were attributed to the longitudinal structures of GW activity (Chandran et al., 2010).

Here, we want to interpret NLC characteristics derived from 10 min resolution lidar data at both sites with respect to GW. The observation of multiple layers could provide a direct link to wave activity in the mesosphere. Multiple layers can represent ice particle populations with different age or origin, or can result from one wider layer that splits, e.g. due to a local temperature variation or wind changes. The scales observed are smaller than typical vertical GW wavelengths of the order of 10–16 km (see e.g. Tsuda et al., 1989); however, the breaking of these waves might still cause smaller scale local structures, such that we attribute the occurrence of multiple layers to pronounced wave activity with the vertical distance of the layers corresponding to a vertical wavelength. We observe multiple layers with 9% probability. The vertical separation is 1.5–3 km at ALOMAR and Davis, respectively. Although possible that the larger separation at Davis hints a different GW spectrum reaching mesospheric altitudes above Davis, the error is too large to state a significant difference. A similar analysis of PMSE was performed by Hoffmann et al. (2005, their Fig. 2), who found vertical distances between PMSE layers of 2.9 km, the lower layer corresponding to mean NLC altitudes. Vertical wavelengths in this order of magnitude were also reported by Beatty et al. (1992) and Dalin et al. (2004). Hoffmann et al. (2008) have studied the influence of GW on layering processes in PMSE and NLC using microphysical simulations. NLC at 69° north or south are not permanent phenomena in the summer mesosphere, rather cloud patches are transported through the lidar beam by mean winds, with times of no significant backscatter ratio and times of NLC detections. Those NLC patches seem to correspond in general to satellite images of PMC, e.g. from CIPS on the AIM satellite (e.g. Carbary et al., 2003; Chandran et al., 2009) and are also considered when interpreting data of satellite instruments with coarser resolution, e.g. regarding temperature estimations from SOFIE (Hervig and Gordley, 2010). Recently, Baumgarten et al. (2012) have found corresponding observations by lidar and satellite up to 1 h after the coincidence using advection by a constant wind from radar measurements. We find a spectrum of mean duration of NLC of about 5 h at both stations (Fig. 6). This is in agreement with model studies (Kiliani et al., 2013). Assuming that the cloud patch is advected with a mean wind of about 40 m/s (Fiedler et al., 2011), a duration of 5 h corresponds to a distance of 720 km. This might be related to the horizontal wavelength of GW or tides with the subsequent temperature changes large enough to allow for the growth of NLC particles in its minimum and sublimation of particles during its maximum. NLC would appear periodically with a given time span and a separation in time that is dominated by the background temperature as indicated in Fig. 1. We find a mean cloud-to-cloud time of 9.5 h (Fig. 6) corresponding to a cloud separation of about 1400 km. The conversion to spatial scales are only to be understood as a first estimate, as the three-dimensional morphology of NLC, the propagation of waves against the mean wind flow and the actual wind patterns as can be obtained by radar are not considered in this simple pure advection model.

4.3. Vertical NLC layer progression

The downward progression of NLC layers is caused by an interplay of temporal effects like particle sedimentation by gravity, vertical winds, and advection of the NLC layer through the lidar sampling volume by strong wind and temperature structures induced by waves. The observations have shown that the progression is directed downward with a mean of ~0.3 km/h. During a NLC observation time $t_{\text{obs}}$ of 5 h, the particles apparently move down 1.5 km. The downward progression is coupled to the NLC duration by the vertical extension of the saturated zone with cold enough temperatures for ice formation. We deduce that this zone is of about equal vertical extent at both Davis and ALOMAR, because no significant differences regarding NLC duration and downward progression was found. However, it is shifted to higher altitudes at Davis.

An apparent downward progression measured by lidar could be caused by tilted NLC structures that prefer a specific orientation, provided by gravity and atmospheric turbulence. The direction of NLC progression is therefore related to the direction of the background winds. This is in accordance with higher temperatures at NLC altitudes that would be caused by the changed GW forcing. This effect is in addition to other influences on NLC characteristics, such as mesospheric winds, tides, planetary waves and inter-hemispheric coupling.

### Table 3


<table>
<thead>
<tr>
<th>Seasons</th>
<th>November/December</th>
<th>November–February</th>
<th>OF ratio (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late transition</td>
<td>32.3 h of 513.8 h</td>
<td>6.2% 171.0 h of 1387.2 h</td>
<td>12.3% 50.4</td>
</tr>
<tr>
<td>early transition</td>
<td>44.8 h of 366.7 h</td>
<td>12.2% 143.2 h of 1152.8 h</td>
<td>12.4% 98.3</td>
</tr>
</tbody>
</table>

### Table 4

Layer statistics of NLC at Davis for years with late transition and years with early transition only using data from November/December.

<table>
<thead>
<tr>
<th>NLC parameter</th>
<th>Davis late transition</th>
<th>Davis early transition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Single layers</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$t_{\text{peak}}$ (km)</td>
<td>33.5 h</td>
<td>36.2 h</td>
</tr>
<tr>
<td>$I_{\text{peak}}$ (10^-15/m/s)</td>
<td>85.00 ± 2.27 (0.16)</td>
<td>83.08 ± 2.87 (0.19)</td>
</tr>
<tr>
<td>$\Delta Z$ (km)</td>
<td>7.5 ± 3.9 (0.3)</td>
<td>10.5 ± 7.7 (0.5)</td>
</tr>
<tr>
<td>Multiple layers</td>
<td>3.3 h 9.1%</td>
<td>6.8 h 15.9%</td>
</tr>
</tbody>
</table>
beginning of the NLC season and by an influence on temperature cause NLC to appear, fade away and re-appear at an hour scale. We have compared lidar data of NLC observed from the sites Davis and ALOMAR at conjugate latitudes in the northern and southern hemisphere at scales of several minutes to hours. Both stations are located at the same latitude of 69° but differ regarding thermal structure and GW forcing at the mesopause region. The differences and similarities regarding NLC characteristics presented here are shortly summarized: Multiple layers, that are thought to be related to GW, are observed during 9% of the NLC observation time at both stations, the vertical distance varying between 1.5 km (ALOMAR) and 3 km (Davis) which might hint at a different spectrum of waves reaching mesospheric altitudes. At both locations NLC layers are 1 km wide, and at Davis on average 1–1.4 km higher. The variations of the seasonal onset in the SH leave distinct GW signatures in NLC, e.g. multiple layers occur more frequent during years of early seasonal onset and vortex breakdown, i.e. during years with reduced GW filtering. NLC tend to decrease in altitude as measurement time proceeds, with an average speed of –0.3 km/h. Using twin lidar measurements at ALOMAR, it was shown that the downward motion was not caused by the advection of tilted structures. A spectrum of durations of NLC up to 10–12 h is observed at both stations with a mean NLC duration of about 5 h. On average, NLC reoccur after 9.5 h at both stations. Using mean horizontal winds a duration of 5 h corresponds to 720 km which is in the range of horizontal wavelengths of mid-frequency GW.

Acknowledgments

We thank the ALOMAR staff for operating the ALOMAR RMR lidar as well as several students that have operated the lidar on a campaign basis. The Davis measurements used here were obtained under Australian Antarctic Science project 737, and we thank the Davis wintering scientists for collecting these data. The European Centre for Medium-Range Weather Forecasts (ECMWF) is acknowledged for providing stratospheric zonal wind data.

References


Backhouse, T.W., 1885. The luminous cirrus cloud of June and July. Meteorological Magazine 20, 133.


