Gravity wave events from mesoscale simulations, compared to polar stratospheric clouds observed from spaceborne lidar over the Antarctic Peninsula

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Abstract

This article compares Gravity Waves (GW) and Polar Stratospheric Clouds (PSC) above the Antarctic Peninsula for winters (June, July and August) between 2006 and 2010. GW activity is inferred from stratospheric temperature and vertical winds from the Weather and Research Forecast mesoscale model (WRF), and documented as a function of time and geography for the studied period. Significant GW activity affects 32% of days and follows the Peninsula orography closely. Volumes of PSC, composed of ice and Nitric Acid Trihydrate (NAT), are retrieved using observations from the spaceborne lidar CALIOP (Cloud-Aerosol Lidar with Orthogonal Polarization). They are documented against GW activity as a function of time and longitude. 65% of ice PSC are observed during GW events, when the average volume of PSC per profile doubles. Maximum ice PSC volumes are seen directly over the Peninsula (65°W), while maximum NAT PSC volumes appear downstream further East (~35°W). Effects of GW events on NAT PSC are felt as far East as 40°E. Our results support the importance of gravity waves as a major mechanism driving the evolution of ice PSC in the area, but the effects on NAT PSC are harder to detect. After a GW event ends, volumes of ice PSC get back to their usual levels in less than 24h, while this process takes more than 48 hours for NAT PSC. Daily profiles of H2O and HNO3 mixing ratios, retrieved from MLS observations, are used to derive ice and NAT frost points with altitude and time. Combining these frost points with modeled stratospheric temperatures, the volumes of air able to support ice and NAT crystals are quantified and compared with PSC volumes. Correlation is high for ice crystals, but not for NAT, consistent with their much slower nucleation mechanisms. Observations of ice PSC over the domain are followed by a strong increase (+50-100%) in NAT PSC formation efficiency 2 to 6 hours later. This increase is followed by a steep drop (6-10h later) and a longer period of slow decline (10-24h later), at the end of which the NAT PSC formation efficiency is less than half its initial value. The fact that these effects tend to cancel each other out, coupled to the important lag in NAT PSC reaction to GW activity, suggest why it is especially difficult to quantify how GW activity impacts NAT PSC cover.

1. Introduction

Polar Stratospheric Clouds (PSC) form during polar winter nighttime, May to September in Antarctica. They play an active part in the formation of the seasonal ozone hole, as sun-activated reactions on PSC particles transform passive species (such as HCl and HBr) into active chlorine and bromine that cause ozone loss [*Solomon*, 1999]. Moreover, during sedimentation PSC particles scavenge nitric acid, slowing down the reconversion of active chlorine into passive species necessary to ozone recovery [*Jensen et al.*, 2002]. Climate-change related drops in stratospheric temperatures could increase PSC formation, leading to enhanced polar ozone depletion [*Hitchcock et al.*, 2009].

The formation of PSC particles is driven by stratospheric temperature, pressure and available mixing ratios of nucleating species (i.e. water vapor, HNO3 and H2SO4). Depending on these factors, nucleation processes lead to the formation of solid NAT (Nitric Acid Trihydrate, water and HNO3), liquid STS (Sulfate Ternary Solution, water, HNO3 and H2SO4) or solid ice particles. PSC made of these particles are often described, respectively, as Type Ia, Ib and II (roughly from the most to least frequent). This classification originates in the distinct optical properties of each particle type as seen from lidar [*Poole and McCormick*, 1988], but is imprecise as PSC often contain a mixture of all types. It is however convenient to describe a PSC whose composition is dominated by a given particle type. Secondary particle types have been described, e.g. Enhanced NAT [*Tsias et al.,* 1999] or NAT rock [*Fueglistaler et al.,* 2002], but are more rare.

Temperature is the main factor driving the extent and composition of PSC cover. It needs to be colder than specific thresholds for ice, NAT or STS formation, depending on the abundance of source species [*Hopfner et al.*, 2009]. At average levels of stratospheric water vapor and HNO3, ice form in air colder than ~192K at 100hPa and ~183K at 20hPa [*Marti and Mauersberger*, 1993], while the NAT threshold is 6-7K warmer [*Tabazadeh et al.*, 1994]. In the Arctic, stratospheric temperatures generally hover around those thresholds [*Pitts et al.*, 2011], and external processes acting on temperatures are crucial in driving the presence and eventual cover of PSC. In the Antarctic, during most of the polar winter stratospheric temperatures are generally colder than required for PSC formation. External processes still play a role, especially during the early, warmer stages of polar winter (i.e. May-June, [*McDonalds et al.*, 2009]) but also later into the season, depending on atmospheric conditions. Gravity waves (GW) are among the most important of such mechanisms: they generate intense temperature fluctuations that propagate up to the mid-stratosphere [*Plougonven et al.*, 2008], which lead to the accelerated formation of stratospheric clouds. The

Antarctic Peninsula is a noticeable "hot spot" of intense GW activity [*Alexander and Teitelbaum*, 2007], but GW happen in all wintertime polar areas [*Alexander and Shepherd*, 2010].

The spaceborne lidar CALIOP (Cloud-Aerosol Lidar with Orthogonal Polarization) on the CALIPSO satellite (Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observation) monitors Earth continuously and autonomously since 2006 [*Winker et al.*, 2007]. CALIOP's high sensitivity to weakly opaque atmospheric features, and its ability to document their vertical variability, allows a very accurate retrieval of the location and altitude of optically thin clouds [*Martins et al.* 2011], among them PSC. CALIOP's sensitivity to light polarization allows the identification of PSC types [*Pitts et al.*, 2007], while its frequent overpasses of polar regions (~14 times a day) insures the sampled PSC are representative of the cloud cover. Thanks to these advantages, CALIOP observations have led to breakthrough insights into PSC properties and processes, and provided a comprehensive overview of the seasonal evolution of PSC population across several years over Antarctica [*Noel et al.*, 2008; *Pitts et al.*, 2009].

A previous paper described an optically thick ice PSC observed by CALIOP over the Antarctic Peninsula during at least 2 days [*Noel et al.*, 2009]. High-resolution mesoscale simulations showed it formed following intense, quasi-stationary temperature drops (~15K) linked to orographic gravity waves. These brought stratospheric temperatures well below the ice frost point, leading to fast generation of ice crystals. Nearby observations revealed a NAT PSC downstream with respect to the polar vortex, while upstream was clear-sky. Back-trajectories confirmed that the air masses holding the NAT PSC went through the GW-affected area. This suggests NAT crystals nucleated on melting ice particles sedimenting from the GW-generated ice PSC and advected by the strong winds of the polar vortex, a process known as mountain-wave seeding [*Eckermann et al.*, 2009]. The details of this process, its generality and importance for the general PSC population over Antarctica, remain poorly known and hard to quantify.

Here several years of data are used to investigate how GW over the Antarctic Peninsula influence ice and NAT PSC. First, atmospheric properties from a mesoscale model document stratospheric GW activity during south hemisphere winters between 2006 and 2010 (Sect. 2). CALIOP observations then document volumes of ice and NAT PSC during the same period (Sect. 3), which are interpreted against GW activity from the previous section. Finally, model output are combined to observations of PSC and mixing ratios, to document the effect of GW activity on PSC formation efficiency from the intersecting dataset (Sect. 4). Results are summarized and discussed in Sect. 5.

2. Gravity Wave activity

2.1. Model setup

The Weather Research and Forecast model [*Skamarock et al.*, 2005], WRF v. 3.2.1, ran on a 100x100 domain centered on the Peninsula with 20 km horizontal resolution and 120 vertical levels. Minimum pressure was requested as 5hPa with a 3 km damping layer to avoid reflections of atmospheric waves on the domain lid. Topography was derived from the database provided with WRF. 38 vertical levels were available in the 120-10hPa vertical range where PSC are most frequent. Initial and boundary conditions for the model were provided by 6-hourly ERA-Interim reanalyses at 0.75° resolution [*Simmons et al.*, 2006]. Stratospheric state was snapshotted every 3 hours. Simulations were run along consecutive 4-days periods, overlapping by one day. The first day of each period was discarded as warm-up. Simulations cover June 1st to August 31st, between 2007 and 2010. For 2006 simulations began June 14th (i.e. when CALIOP observations start). In the output, only the 90 x 90 cells (1800 x 1800 km) in the domain center were considered to avoid border effects.

2.2. A Gravity Wave event: July 17th, 2006

Gravity Waves lead to intense spatial fluctuations of vertical wind. Figure 1 shows its maximum variation along the vertical between 120 and 10hPa, from the WRF output for July 17th, 2006 at 12:00UTC. This is roughly the peak activity during a GW event (chosen randomly in the dataset) that began on July 16th near 12:00UTC, when wind and temperature fluctuations appeared in simulated fields above the Peninsula orography (65°W). The event lasted until July 20th, when fluctuations disappeared near 06:00UTC.

Large variations in vertical wind are observed directly above the Peninsula, North of 74°S between 73°W and 55°W. Vertical profiles of vertical wind speed and temperature (Fig. 2) were extracted at the white (63°W, 70°S) and grey (45°W, 70°S) spots in Fig. 1, respectively strongly affected and unaffected by the GW event. In the GW-affected area (blue lines), vertical winds (Fig. 2a) go from strongly positive (upward motions) to strongly negative (downward motions), and often get faster than 1 m/s. Associated temperature fluctuations of roughly ±10K (Fig. 2b) cool air to 180K at 40 hPa, considerably colder than required for ice crystal formation at such pressure levels (~186K). In the unaffected area (green lines), vertical motions are slower than 0.05 m/s (Fig. 2a), thus almost negligible, while temperatures get close to the frost point (near ~30hPa) but do not cross it (Fig. 2b). This shows the importance of GW as triggers of ice PSC formation. As an aside, note how the thermal tropopause inversion (Fig. 2b, near 130 hPa) is negligible compared to the temperature fluctuations created by the gravity waves (blue line).

Fig. 3 presents zonal cross-sections of stratospheric vertical winds (left) and temperature (right) as a function of longitude and pressure, for latitudes near 70°S. Orography is faintly visible in white at the bottom. Strong vertical wind perturbations appear from 70°W to 69°W and especially 65°W to 63°W. Those line up with the East front of two Peninsula peaks, respectively the Douglas Range and Mt. Jackson, downstream with respect to the polar vortex. This alignment confirms the orographic origin of GW. Atmospheric waves follow the orientation of the vortex and propagate East; their influence shows as far as 45°W. Looking back at the start of the event (July 16th, not shown), it took roughly 24 hours for the waves to extend that far East from 65°W. Vertically, waves affect the entire vertical column up to 10 hPa. Temperature fluctuations (Fig. 3, right) are less visually dramatic since their stratified decrease with altitude dominates the figure, but obviously overlap wind fluctuations.

Based on this case and numerous others, GW-affected domain cells were identified as vertical winds faster than 1 m/s between 10 and 120 hPa. Attempts were made to filter out non-GW perturbations based on their typical spatial frequencies through Fourier transforms, but these did not improve the results and were not pursued further.

2.3. GW frequency maps

Fig. 4 shows how frequently horizontal domain cells are GW-affected during Antarctic winters (June-July-August) between 2006 and 2010, applying the above criteria based on vertical wind fluctuations on the entire model output. The geographical distribution of GW-affected cells is roughly similar to the case study (Fig. 1), but follows more closely the shape of the Peninsula. Frequencies reach 25% directly above the tallest mountains, consistent with their orographic origin. The shape of affected areas is similar for all months (not shown), although areas which are GW-affected more than 25% of the time are larger in July and smaller in August. A second, smaller local maximum of GW frequencies is slightly visible on Fig. 4 near mountain peaks at 70°W (probably Mt. Stephenson in the Douglas Range, as in Fig. 3). This mountain leads to more GW activity during July (not shown). The most intense GW activity is in 2006 and 2010, the weakest in 2007 and 2009.

Fig. 5 shows the monthly frequencies of cells affected by GW within 5° longitude bins. GW activity is maximum in July (up to 12% of cells affected), then June (up to ~9%), and minimum in August (up to 6%). In all months GW activity peaks near 65°W, right above the

Peninsula mountains. It decreases rapidly eastwards and westwards, and is almost negligible outside of the 75°W-55°W longitude range. The secondary local maximum observed on the frequency map (Fig. 4), west of the Peninsula, also shows up here (thin green line, near 70°W). It is especially visible in July.

2.4. Vertical wind and temperature distributions

Fig. 6 shows histograms of speed for vertical winds faster than 1 m/s above the simulated domain, between 10 and 120 hPa, every 3 hours. The height of peaks describes the fastest wind on a given date, while the colors inform on the number of domain cells affected by winds faster than 1 m/s - i.e. the geographical extent of the GW-affected area at a given time. For instance, in 2006, GW effects were felt during more than 14 consecutive days (June 24th to July 8th), with wind speeds faster than 6 m/s. The optically thick, GW-induced ice PSC described in *Noel et al.* [2009] happens in this period. Small peaks are sporadically observed, for instance during the second half of August 2006. They barely cross the 1 m/s threshold and generally last for less than a day.

Fig. 7 shows histograms of temperature using the same conventions as Fig. 6. Side by side, these figures show that fast vertical winds are often accompanied by the apparition of roughly symmetrical temperature drops and increases (caused by vertical fluctuations in temperatures as in Fig. 3b). For instance, the 2006 period of uninterrupted GW activity (June 24th - July 8th) reaches temperatures as cold as 160K. Such an impressive cooling (~20K) is however rare, it is reached roughly once per year. Wind and temperatures are not always correlated: for instance, the fast, short-lived vertical winds on August 2nd, 2007 or the last fortnight of August 2008 (Fig. 6) are not linked to noticeable temperature drops (Fig. 7). It can be hypothesized that, during such events, winds are not strong or widespread enough to generate significant wave activity and temperature drops. To avoid taking those into account, in the rest of the article a GW event is defined as an uninterrupted period of at least 24 hours during which at least 10 000 km2 of the domain are GW-affected. As in the previous section, a 400 km2 cell is considered GW-affected when its vertical winds reach 1 m/s.

Table 1 describes GW events identified during Antarctic winters 2006-2010. Out of 437 simulated days, 143 are GW events (32%). 2006 presents a rather small number of GW events (3) compared to the other years, but includes the longest event (15 days) reaching the fastest vertical winds (10.7 m/s), and the largest volume with fast vertical winds during events (4.5%). On the opposite side of the spectrum, 2008 presents a large number of events (8), spanning the longest period (43 days, 47% of the season). However, these events are short, weak, and affect a small part of the domain (1.1 to 2.8%). As a result, 2008 presents the smallest volume with fast vertical winds during GW events (2%). 2009 has the least GW activity: 3 events and 15 days in total, or 16% of the season. Among the 143 days of GW events, 34 fall in June, 69 in July and 40 in August, following the same monthly hierarchy of GW activity found in Fig. 5.

3. GW events and PSC observations

3.1. PSC observations from CALIOP

As in *Noel et al.* 2008 and 2009, PSC were detected in CALIOP observations of total attenuated backscatter (TAB) through the following steps: 1) averaging TAB profiles horizontally over 10 km to improve the signal-to-noise ratio, 2) thresholding the TAB above the molecular backscatter (normalized in PSC-free regions between 34 and 38 km of altitude) to retrieve PSC altitudes, 3) removing spurious detections due to nongeophysical signal fluctuations, by imposing specific geometries typical of PSC, such as minimal horizontal and vertical extensions. In addition to previous efforts, we added an extra step of threshold-based detection from perpendicular backscatter, as in *Pitts et al.* [2009]. Ice and NAT classification was based on layer-integrated depolarization and scattering ratios, following the boundaries in *Pitts et al.* 2009. For simplification, a single class of NAT was considered, in contrast with *Pitts et al.* [2009] and [2011] who considered respectively two and three NAT classes. From these retrievals, we calculated volumes of NAT and ice PSC. We supposed a CALIOP point describes a tridimensional section of atmosphere, defined by CALIOP's vertical resolution (60 to 300 meters depending on altitude) and the horizontal footprint of profiles. To simplify the comparison with simulations (Sect. 2), this footprint was supposed rectangular with dimensions defined by the horizontal averaging used here (10 km x 10 km). A volume was thus calculated for each CALIOP point identified as NAT or ice PSC; PSC volumes were then summed and binned along time and space coordinates for analysis purposes.

PSC retrieval and classification using the present algorithm were compared with the dataset from *Pitts et al.* [2011] (hereafter called NASA dataset) for August 2007 over the domain defined in Sect. 2.1 (not shown). In contrast with the present detection, which considers a single horizontal averaging distance of 30 km, cloud detections in the NASA dataset are based on a multi-resolution horizontal averaging scheme (5, 15, 45 and 135 km). The present dataset agrees well with the NASA one for 15 and 45 km horizontal averaging distances, consistent with the 30-km averaging resolution selected here. For such detections, the difference between datasets in average daily PSC volume per profile (considering all PSC types) is less than 1%. This variation is also less than 1% considering only PSC classified as NAT and less than 4% considering only PSC classified as ice in both datasets. Including PSC detected at a 135 km averaging distance in the NASA dataset increases the total PSC volume by a few percents. These percents contain the optically thinnest PSC, requiring the largest averaging for detection: primarily young STS PSC emerging from sulfate aerosols, or NAT PSC with low number densities. Overall, the present dataset contains slightly fewer PSC than the NASA dataset considering all averaging resolutions. Since the present article studies ice and NAT PSC, an accurate detection of optically thinner STS PSC was not attempted (as in the more comprehensive NASA dataset) and lower detection rates on these clouds should be expected. For the present purposes of analyzing ice and NAT PSC, we consider the difference between both datasets to be small enough for them to be statistically equivalent.

3.2. PSC detections and volumes

PSC Observations were analyzed between 60°S and 70°S, for consistency with the analysis of *Alexander et al.* [2011], and between 100°W and 100°E, to cover the influence zone of Peninsula gravity waves. Due to its limited horizontal field of view, CALIOP observes less than 25% of this domain each day; moreover, this sampling is irregular, i.e. the number of CALIOP profiles in the domain changes from one day to the next. To account for this imperfect sampling, PSC volumes are normalized by the number of available CALIOP profiles in the studied region over the time period of aggregation. Days with less than 500 profiles were discarded for fear they would bias results. Hence we present PSC volumes per profile, which should be independent of sampling fluctuations. For the sake of brevity, these will be referred to simply as PSC volumes from now on.

Over the 2006-2010 Antarctic winters (JJA), 397 days were considered adequately sampled (over 437). Ice PSC appear in 37% of them (i.e. an ice PSC was identified in at least one CALIOP profile). GW events contain 138 well-sampled days, 52% of which showed ice PSC presence. Calm periods contain 98 well-sampled days with ice PSC 21% of the time. NAT PSC were almost ubiquitous as they were detected 93% of the time, with no difference between GW events or calm periods. Thus, in this domain, ice PSC are more frequent than average during GW events; this is not true for NAT PSC. Overall, 63% of ice PSC and 46% of NAT PSC volume were observed during GW events, while GW events affected only 32% of days. Average volumes are 4.3 km^3 per profile for ice PSC and 45.1 km^3 for NAT PSC (roughly 10 times larger). This very large difference is mostly due to ice PSC being much less frequent than NAT PSC, a large number of unaffected profiles are thus counted in the average. Sect. 3.4 will show that this difference decreases significantly when only cloudy periods are considered. The largest volumes of ice PSC (Table 2) are observed in 2008 and 2006 (6.5 and 6.1 km^3 per profile), and the largest volumes of NAT PSC in 2006 and 2008 (68.6 and 55.2 km^3 per profile). Table 1 showed that 2006 had the longest and most intense GW events, while 2008 had the most days affected by GW events. The smallest volume is observed in 2009 for ice PSC (the year of minimum GW activity, Table 1) and 2010 for NAT PSC. These results suggest that, first and foremost, the intensity of GW activity directly affects the observed volume of PSC in the domain. Second, long and intense GW events are correlated with more NAT PSC, while short and numerous GW events are correlated with more ice PSC. Ice and NAT PSC volumes are small in June and large in July. This monthly ordering follows the trend in GW events found in Fig. 5 and Sect. 2.4.

3.3. PSC volumes during the 2006, 2008 and 2009 Antarctic winters

Figure 8 shows how daily PSC volumes change during the 2006 (top), 2008 (center) and 2009 (bottom) Antarctic winter seasons, chosen for the specificity of their GW activity. The influence of GW events (thick lines) is not particularly obvious. For example, the extended 2006 GW event (June 25th to July 10th) contains two spikes in ice PSC volume (June 28th and July 8th), but these are separated by 7 days with zero ice PSC. A local maximum of NAT PSC is observed close to the beginning of the event (June 28th), but the following drop in NAT PSC volume (until July 3rd) also happens during the GW event. This specific GW event therefore includes both increases and decreases in ice and NAT PSC volumes.

However, considering the three years shows GW events likely correlate with increases in ice PSC volume. First, it is worth noticing again the limited ice and NAT PSC volume during 2009 (year of minimum GW activity) compared to 2006 and 2008. Moreover, ice and NAT PSC volumes are more stable during periods without GW - this is especially noticeable by comparing 2009 with other years, or focusing on August 2006 and 2008. It suggests that GW events are linked to fluctuations in PSC volume. Second, all significant increases in ice PSC volumes happen during GW events, except two (July 24th, 2006 and July 4th 2009). On the other hand, ice PSC and large volumes of NAT PSC were also observed out of GW events - e.g. August 2006 or June-July 2009. GW events thus appear to enhance ice PSC formation, but are not a necessary condition. Besides, ice PSC increases seem correlated with simultaneous, or slightly delayed, increases in NAT PSC volume. This is however not specific to GW events, see e.g. ice PSC volumes increase on July 4th, 2009 and NAT PSC follow the next day.

All these remarks generally stand for 2007 and 2010 (not shown). They do not apply to every case, but CALIOP's sampling of the domain is imperfect: on a given day a NAT PSC can be observed and the nearby ice PSC missed, or vice versa. This could be mitigated by using longer averaging times, but then the slight delay in NAT PSC increase would be missed.

3.4. Distribution of daily PSC volumes

Fig. 9 shows the distribution of daily averaged PSC volume, considering the whole observation period (2006-2010). Only days with PSC observations were considered, i.e. 64% of days for ice PSC and 93% of days for NAT PSC (Sect. 3.2) - in other words, the most frequent daily volume of ice PSC is 0 km3, but is not shown in the distributions. All days are described in the left column, GW events in the middle, and calm periods on the right, for ice (top) and NAT (bottom) PSC. GW events are from Sect. 1.4 (Table 1). Calm periods are defined as limited GW activity for at least 48 hours straight, occurring at least 24 hours after the end of a GW period, to let GW-related effects dissipate. Limited GW activity is defined as less than 2000 GW-affected km2 in the domain.

PSC volumes are roughly 4 times larger for NAT than ice. Daily PSC volumes stay below 100 km3 per profile for ice and 200 km3 per profile for NAT. In calm periods, ice PSC volumes are small: 11 km^3 per profile on average, with a 4.9 km^3 median. The median might be more significant, as calm days with ice PSC are rare and a single day with ice PSC volume of 45 km^3 skews the mean (Fig. 9, top right). During GW events, ice PSC volumes are large: 20.4 km^3 on average, with a 12.5 km^3 median (roughly twice the one in calm periods). Volumes larger than 50 km^3 make an appearance. NAT PSC volumes are also considerably larger during GW events: 70.3 km^3 on average, more than twice the average of calm periods (28.4 km^3). NAT volumes larger than 200 km^3 are frequent in GW events. These results show a strong impact of GW events on PSC volumes.

3.5 PSC volume as a function of time

We extracted ice and NAT PSC volumes in CALIOP observations as a function of time after the end of each GW-affected period (Table 1) followed by at least 48 GW-free hours. Along this 48-hours period, PSC volumes (Fig. 10) transition between the averages for GW events and calm periods (Fig. 9). The transition takes a different time for each PSC type. The ice PSC volume (blue line in Fig. 10) drops down quickly, taking 12 to 24 hours to get back to the near-zero volumes found in GW-free periods. The fastest decline in ice PSC volume is right after GW events end. NAT PSC volumes (green line) stay affected much longer, taking as long as 48 hours to drop down to the 20-30 km^3 level typical of volumes observed during calm periods. The steepest decline in NAT PSC volumes occurs between hours 36 and 48, when a ~50% drop is recorded.

3.6. PSC volumes as a function of longitude

Fig. 11 shows ice and NAT PSC volumes as a function of longitude in 10° bins. During GW events (top), ice PSC volume (blue line) is maximum near 75°W (~75 km^3), right above the GW-triggering orography of the Peninsula. Large volumes (> 25 km^3) extend far sideways (80°W-30°W), especially East. Outside of this longitude range, the volume of ice PSC is smaller, down to levels observed during calm periods. Thus the very small ice PSC volume reported in Sect. 3.2 (4.3 km^3 per profile) is explained by 1) the rarity of ice PSC in time (Sect. 3.2 and 3.3), and 2) their acute geographic locality. However, ice PSC reach large volumes punctually (during GW events) and locally (above the Peninsula), as shown here. NAT PSC reach much larger volumes (green line in Fig. 11), but are also locally dependent. The local maximum NAT volume (~150 km^3/profile) is much larger than the average, in calm periods (28.4 km^3) and GW events (70.3 km^3). The maximum is East of the Peninsula (~35°W), which suggests these NAT PSC form primarily through mountain-wave seeding triggered by ice PSC westward. The seeding effect on NAT PSC volume starts near 65°W, and extends as far as 40°E. GW-generated ice PSC therefore affect NAT PSC on several thousands kilometers, as in *Alexander et al.* [2011]. Outside this longitude range, the NAT PSC volume is stable around 50 km^3. This is larger than the average volume in GW-free periods, for reasons which will be discussed in Sect. 4.

During calm periods (Fig. 11, bottom), ice PSC almost disappear, as in Sect. 3.4 (Fig. 9). A few PSC still appear near 100°W-80°W, quite far upwind of the Peninsula. They form a geographically consistent cluster and cannot be attributed to data noise; we lack an explanation for their rather puzzling presence at this point. By contrast, a relatively stable background level of NAT PSC fluctuates between 50 km^3 above the Peninsula to ~20 km^3 East of 20°W. This is close to the average volume in calm periods (Sect. 3.4) and to the volume 48 hours after the end of GW events (Sect. 3.5). Even during calm periods (unaffected by GW temperature fluctuations), volumes of NAT PSC fluctuate with longitude and are larger between 80°W and 20°W. This could be due to weak GW activity creating limited atmospheric perturbations above the Peninsula. These would not be identified as GW events, as they would fail to cross the 1 m/s wind speed threshold, but could still manage to cool temperatures below the NAT frost point (easier to reach than the ice frost point).

The two vertical dotted lines on Fig. 11 limit the modeled domain (Sect. 2) used to identify GW. Fig. 11 suggests that comparing model output with observations in that domain, as in the next section, should capture well enough the main effects of gravity waves on PSC, although some information (mostly about NAT PSC) might be lost East.

4. GW activity, frost points and PSC

Sections 2 and 3 document effects of gravity waves on PSC seen by CALIOP, thus adding to the significant evidence from literature. However, trying to quantify these effects raises two problems: 1) GW affect PSC formation when they help to bring temperatures below frost points. Even strong GW have zero impact if they fail to bring temperatures down enough, or if the stratosphere was already cold enough to begin with. This explains why GW events do not always affect PSC volumes (Fig. 8, Sect. 3.3); 2) Large-scale planetary waves condition pre-existing atmospheric conditions, and can influence the probability of PSC formation. When such waves are advected by the polar vortex, their effects will interact with those of any present GW. This is apparent on Hovmoller diagrams in *Alexander et al.* [2011], where increased NAT PSC volumes above the Peninsula (linked to GW), propagating eastward (downwind), are actually included in larger NAT PSC systems originating west of the Peninsula (upwind) linked to planetary wave activity.

These two effects, combined, make it very difficult to isolate and quantify the impact of gravity waves and mountain-wave seeding on PSC. In an attempt to work around these problems, in this section we relate observed ice and NAT PSC volumes to volumes colder than the frost points. Considering the nucleation of PSC particles occurs homogeneously, this relation should be independent of changes in stratospheric conditions (temperature, pressure, concentrations). Changes in this relation could identify a qualitative change in the PSC particle formation process, such as the triggering of the seeding effect.

4.1. Ice and NAT frost points

Frost points were retrieved using observations from the Microwave Limb Sounder (MLS). Onboard Aura in the A-Train constellation, it is well colocated with CALIOP, especially at the Poles where the orbital shift between the two satellites is minimum. MLS vertical profiles of mixing ratios for water vapor and HNO3 (Level 2 data v2.2, H2O and HNO3 products) were averaged daily over the considered domain (Sect. 1.1). Profiles with Quality flag below 0.9 (for H2O) and 0.4 (for HNO3) were removed to avoid observations affected by tropospheric clouds [*Livesey et al.*, 2007]. Since the formation of PSC decreases gas phase HNO3 and H2O, their presence can bias MLS measurements and underestimate the actual abundance of either species. This would give a cold bias to our retrieved frost points [*Schoeber et al.,* 2006]. To avoid this, MLS profiles were discarded if colocated CALIOP profiles (same A-Train overpass, closer than 180 km) revealed an average PSC thickness greater than 500 meters. MLS profiles with less than 4 colocated CALIOP profiles were also discarded. For water vapor, averaging was performed on log(H2O) as recommended by the MLS documentation (see also *Vomel et al.* [2007]).

In the 10-120 hPa range under study, the minimum MLS concentrations are 0.1 ppmv (H2O) and 0.7 ppbv (HNO3), and the resolutions ~3.5 km vertically, 200-400 km along-track and 7-10 km cross-track. As an example, Fig. 12 shows water vapor mixing ratios retrieved for 2006. In a strong drying event between July 24th and August 8th, water vapor concentrations between 10 and 40 hPa fall from ~5 to 4 ppmv. This event is characteristic of the wintertime Antarctic stratosphere (e.g. *Randel et al.*, 2004). A similar drying is observed in HNO3 mixing ratios (not shown). The mixing ratios and the timing of the stratospheric drying agree with *Nedoluha et al.*, 2002.

Ice and NAT frost points were calculated as a function of time and altitude by applying the formulas of *Hanson and Mauersberger* (1988), for HNO3, and *Murphy and Koop* (2005), for ice, to daily averaged mixing ratio profiles (as in Fig. 12) and full-resolution pressure and temperature fields from WRF. As an example, Fig. 15 shows ice frost points retrieved for the 2006 winter season, using mixing ratios from Fig. 12 and simulations. Frost points are stratified and decrease regularly with increasing altitude, from ~192K at 100 hPa to ~183K at 10 hPa. The drying event in the July-August transition translates to a 4K frost point cooling, mostly noticeable at 20 hPa. Other levels are also affected though. The NAT frost points (not shown) evolve similarly, with the same drop during the July-August drying event, but are generally 5 to 7K warmer (consistent with *Tabazadeh et al.*, 1994).

4.2. GW activity and volume colder than the frost points

Ice and NAT frost points are here compared to WRF temperatures, to quantify the volume of stratospheric air able to sustain PSC formation.

Table 3 shows volumes colder than ice and NAT frost points (T_{ice} and T_{NAT}, respectively), for each simulated year and on average for June, July and August. The largest volumes with T<Tice are in 2010 and 2008, the smallest in 2007. The largest volumes with T<TNAT are in 2010 and 2006, the smallest in 2009. Over all years, the minimum volume with T<Tice is observed in June and the maximum in August (when volumes are ~4 times as large). This follows the seasonal stratospheric evolution, which gets globally colder as the austral winter progresses. Volumes with T<TNAT follow the same monthly evolution (minimum in June, maximum in August), but the relative increase is much less dramatic (+22%) than for ice. These volumes should not be directly compared to the average PSC volumes found in Sect. 3 (Table 1), as they document different domains (a valid comparison is conducted later in the section). They can however be compared with GW events described in Table 1. The absence of any clear correlation between cold volumes and GW properties (either their total length of events, the affected volume or the maximum wind speed) suggests that volumes with T<Tice

and T<TNAT are most likely driven by large-scale synoptic changes, rather than by intense but small-scale fluctuations from GW.

Fig. 14 shows the evolution of volume with T<Tice during the 2006, 2008 and 2009 Antarctic winters. GW events (identified with thick lines) correlate poorly to increases in volume with T<Tice. Large volumes with T<Tice are seen during non-GW periods. This is especially noticeable in 2009, when most of temperatures colder than the ice frost point happens in August. August 2009 shows only limited GW activity, and its temperature is driven by the large-scale stratospheric cooling of the winter season. Moreover, small volumes with T<Tice are seen during GW events. Theses remarks confirm that the impact of GW on frost point crossing is extremely dependent on the initial, pre-GW atmospheric conditions.

Fig. 15 shows the distribution of daily volumes with T<Tice, during GW events days (left) or not (right) over the 2006-2010 period. GW-affected days show more small volumes of T<Tice, but also more large volumes, i.e. the distribution is broader. An explanation is that without GW, the volume with T<Tice is primarily driven by large-scale seasonal temperature change, meaning these volumes are either non-existent (early winter, warm stratosphere) or large (late winter, cold stratosphere). When present, gravity waves create small-scale temperature drops leading to lots of small volumes with T<Tice in periods warmer than Tice (i.e. June). Later in the season (i.e. August), GW temperature drops provide a slight increase to the already-large volumes with T<Tice. These two effects combine to increase small and large volumes with T<Tice in GW events. Daily average volumes with T<Tice are slightly larger during non-GW days.

Previous sections show that ice PSC are clearly much larger in GW events (Sect. 3.4), and are almost inexistent without GW. However, large volumes colder than the ice frost point appear both during and outside of GW events. The cooling ability of GW events therefore cannot be solely responsible for their influence on PSC formation: even if GW do impact stratospheric temperatures, they are not their main influence and are not necessary to drive them below the frost point in large volumes. Such temperatures are routinely reached in large volumes even during calm periods. It is possible the speed of the GW temperature drops, or their intensity, are responsible for the enhanced PSC formation.

4.3. PSC volumes and volumes below frost points

As noted previously, PSC volumes from CALIOP (Sect. 3) cannot be compared directly to frost point volumes from WRF and MLS measurements (Sect. 4.2). CALIOP's sampling is local and irregular in time, while the simulated domain is geographically restricted. Thus, to compare PSC observations and simulations, the following steps were followed: 1) for each CALIOP orbit overpassing the model domain, the closest simulation in time was identified. The worst-case time delta with observations is therefore 3 hours (the simulation time-step); 2) the output fields of the identified simulation were extracted along CALIOP's orbit coordinates and regridded on its altitude levels, respecting the daily vertical pressure structure. In other words, simulated stratospheric fields were extracted at the time, location and altitude of CALIOP observations to produce a space-time intersection of observed PSC and simulated frost points volumes. These volumes, discussed in the rest of this section, should not be compared with those from Sect. 2 and 3, as they do not describe the same areas and periods. Information about GW activity is still relevant though, as it pertains to the simulation domain as a whole.

Reassuringly, intersecting volumes of ice PSC and $T< T_{ice}$ are highly correlated. As an example, Fig. 16 shows their daily evolution for the 2006 Antarctic winter. For most of the season, daily fluctuations and extent of observed ice PSC volumes are consistent with simulated volumes with $T<T_{ice}$. The correlation breaks down after August 8th, when the ice PSC volume drops significantly, several days before a drop in cold volumes. By contrast, the NAT PSC volume (not shown) is always much smaller than the $T<T_{NAT}$ volume, and their daily fluctuations show no apparent correlation. This is probably due to the slow nucleation of NAT particles, which requires air to stay colder than the NAT frost point for at least ~24 hours [*Tabazadeh et al.*, 1996]. Here, GW temperature fluctuations seem too fast to significantly affect the NAT PSC volume, and the formation of NAT PSC depends more on the air mass temperature history than on the instantaneous local temperature. The fact that volumes colder than the NAT frost point are much larger than NAT PSC volumes also suggests that the possible cold bias in NAT frost points (discussed in Sect. 4.1) does not impact the present results significantly.

Fig. 17 shows the distribution of collocated and simultaneous ice PSC volumes and volumes with $T<T_{ice}$ in individual profiles (no averaging) for the entire studied dataset (JJA 2006-2010). As with the daily averages (Fig. 16), instantaneous profile values are highly correlated (correlation coefficient ~0.54, average ratio between both 0.98). Agreement seems best in profiles with large volumes. This linear correlation is not found between volumes of NAT PSC and volumes with $T<T_{NAT}$ (not shown). The average ratio between both volumes is close to 0.46. Again, this is not surprising given the slower nucleation of NAT crystals. These results show that temperature drops have a strong visible impact on ice PSC, but not on

NAT PSC. Since the NAT PSC cover is considerably larger during GW events (Sect. 3), an external mechanism must be responsible, such as mountain-wave seeding.

4.4 Ratios of PSC volumes on volumes colder than the frost point

From here on, ratios of PSC volumes on volumes colder than the relevant frost point shall be referred to as ice ratio (ice PSC volume on volume with $T<T_{ice}$) and NAT ratio (NAT PSC volume on volume with $T<T_{NAT}$). These ratios describe PSC nucleation efficiency, and thus should diagnose the impact of external factors on PSC formation, independently of stratospheric temperature or mixing ratios. Fig. 18 shows the evolution of the ice and NAT ratios as a function of longitude, for the entire period (full lines) and days affected by GW events (dashed lines). The ice ratio (blue lines) oscillates around 1, consistently with the behavior observed in Fig. 17 : the ice PSC volume is approximately equal to the volume colder than the ice frost point. Ice ratios are however lower west of 60°W and greater east, a tendency that slightly intensifies East during GW events. The NAT ratio (green lines) also stays approximately constant with longitude, and stays in the 0.4-0.5 range. During GW events, the NAT ratio is slightly smaller than average west of 55°W and greater east. Enhanced NAT ratios (East of 55°W) appear East of enhanced ice ratios, consistent with NAT formation downstream from the Peninsula.

Near-unity ice ratios (Fig. 17 and 18) are consistent with a fast formation of crystals. Water vapor reacts fast to temperature drops, be they slow synoptic-scale atmospheric changes or fast local fluctuations caused by gravity waves. As a consequence, there is an almost one-to-one correspondence between simulated volumes colder than the ice frost point and observed volumes of ice PSC. Ice ratios greater than 1, as observed eastward of the Peninsula, are a priori nongeophysical; they can either be due to 1) the model underestimating the extent of cold volumes, or 2) important sedimentation of crystals below the areas colder than the ice frost point, which would survive long enough to be observed by CALIOP and identified as ice PSC before they sublimate. These two possibilities would allow the volume of observed ice PSC to get larger than the volume of temperatures able to sustain them. Both possibilities would get worse during GW, which is consistent with the results. The absence of significant change in NAT ratio during GW events apparently contradicts the mountain-wave seeding effect. According to it, ice PSC should lead to an accelerated formation of NAT crystals. This should record as a clear NAT ratio increase, which is absent in Fig. 17. There is an increase east of 55°E, but extremely weak, and smaller than the ice ratio increase east of 60°W which is potentially due to uncertainties in model output. The observed fluctuations of NAT ratio therefore cannot be reliably considered physically meaningful.

Fig. 19 shows the NAT ratio function of the delay after a given volume of ice PSC was observed. After no or little ice PSC are observed (< 15km3 per profile), the NAT ratio stays more or less constant with time (dashed line) between 0.2 and 0.25. After large volumes of ice PSC are observed (full lines), the NAT ratio starts increasing rapidly, and stays above 0.25 two to six hours later. This translates to a 50-100% increase in NAT PSC formation efficiency above average. Afterwards, the NAT ratio decreases slowly, reaching 0.1 twenty hours after ice PSC observation (less than 50% of the average NAT PSC formation efficiency). The intensity of the initial increase and its following drop depends on the volume of ice PSC observed: larger volumes lead to a stronger increase in NAT ratio (more than 0.4 for volumes greater than 250 km3/profile), followed by a stronger drop (down to 0.07). We suggest the strong initial increase in NAT PSC formation efficiency is due to the use of crystals from ice PSC as nucleation seeds. This is supported by the correlation between volumes of ice PSC and increases in NAT ratio - larger ice PSC mean more crystals available for NAT nucleation. The formation of NAT crystals depletes the available HNO3, which leads to a below-average NAT ratio in the following hours. When the NAT ratio is shown as a function of longitude (Fig. 17), these two effects cancel each other out, leading to zero effect on average.

5. Discussion

This article compares mesoscale simulations of stratospheric conditions with volumes of PSC observed by spaceborne lidar above the Antarctic Peninsula, to better understand how Gravity Waves affect their formation and properties.

First GW activity was simulated to identify intense and long-lived GW events (Sect. 2) above the Antarctic Peninsula during five Antarctic winters (2006-2010). Main results involve the documentation of GW activity in this domain and period, including daily fluctuations of fastest winds and coldest temperatures. We present a calendar of GW events, and the geographic distribution of GW activity over the studied area. GW activity was especially strong in 2006 and 2008, and weak in 2009. GW affect ~32% of days and are generally stronger in July and weaker in August. Maximum GW activity is right above the high mountain of the Peninsula, with a second, much weaker maximum above the Douglas Range in Alexander island.

We then correlated observations of ice and NAT PSC over Antarctica with GW activity, to investigate their potential effect on PSC properties (Sect. 3). We document the evolution and distribution of daily ice and NAT PSC volumes during GW events and calm periods. Even though GW only affect ~32% of days, 63% of ice PSC volume were observed during GW events. Volume of ice PSC doubled during GW events, reaching 20km3/profile. Since GW activity is unrelated to seasonal cooling, this change in ice PSC volume is not due to large-scale temperature changes, and is related to GW activity instead. On the other hand, 46% of NAT PSC were observed during GW event. This suggest the effect of GW on these clouds is weaker, yet significant. Our results are consistent with *Alexander et al.* [2011], who found ~50% of ice and NAT PSC were due to GW during 2007. Years with long GW events show larger NAT PSC volumes, and years with short events larger ice PSC volumes. This suggests the length of events, and more generally timing issues, influence how GW events affect PSC. Volumes of ice PSC are strongly longitude-dependent and are largest above the Peninsula (~65°W), while volumes of NAT PSC are largest near ~35°W, far downstream with respect to the polar vortex. We showed that after a GW event, ice PSC volumes get back to the near-zero volumes of calm periods in 12 to 24 hours, with a steep decline during the first 12 hours. Volumes of NAT PSC take as long as 48 hours to get back to non-GW levels. These results strongly support mountain-wave seeding as important for NAT PSC formation above Antarctica. The impact of this mechanism is felt as far East as 40°E, several thousand kilometers downstream of the GW.

Finally, we analyzed the intersection of observations with model output, combined with water vapor and HNO3 mixing ratios observed from space, to investigate changes in PSC formation efficiency due to gravity waves (Sect. 4). We documented volumes colder than ice and NAT frost points above the Peninsula, an information important for studies of stratospheric clouds. We showed that daily and instantaneous volumes of ice PSC are closely correlated with volumes colder than the ice frost point, irrespectively of GW events. Large volumes colder than the ice frost points are observed even without GW, thus the ability of GW to bring temperatures down is not solely responsible for the increased ice PSC observed during them. We propose that the speed and intensity of GW temperature drops plays a role in the enhanced ice PSC formation. On the other hand, observed volumes of NAT PSC do not correlate with volumes colder than the NAT frost point. This can be explained by the slower nucleation of NAT crystals. Finally, the observation of ice PSC leads to a strong increase (+50-100%) in NAT PSC formation efficiency 2 to 6 hours later. This increase can be attributed to enhanced NAT crystal nucleation, using ice crystals as seeds. It is followed by a sharp drop 6h to 10h after ice PSC observation, and a longer slow decline (10h-24h), that we attribute to local HNO3 depletion following the initial increase. After this period, NAT PSC formation efficiency is less than half its initial value. The amplitude of the initial increase and following decline in NAT PSC formation efficiency depends on the volume of ice crystals available as NAT nucleation seeds.

Among other things, our results show that 1) after the end of GW events, NAT PSC volumes take up to 48 hours to go back to non-GW levels; 2) during GW events, a short period of increased NAT formation efficiency is followed by a long decline to lower efficiency levels. The influence of these two phenomenons, which imply significant delays in mountain-wave seeding, might explain why it is difficult to quantitatively correlate GW activity to changes in NAT PSC volumes.

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References

Alexander, M. J., and H. Teitelbaum (2007), Observation and analysis of a large amplitude mountain wave event over the Antarctic peninsula, *J. Geophys. Res.*, *112*, D21103.

Alexander, S. P., A. R. Klekociuk, M. C. Pitts, A. J. McDonald, and A. Arevalo-Torres (2011), The effect of orographic gravity waves on Antarctic Polar Stratospheric Cloud (PSC) occurrence and composition, *J. of Geophysical Research*, *116*, D06109, doi: 10.1029/2010JD015184.

Alexander, S. P., and M. G. Shepherd (2010), Planetary wave activity in the polar lower stratosphere, *Atmos. Chem. Phys.*, *10*(2), 707–718.

Eckermann, S. D., L. Hoffmann, M. Hopfner, D. L. Wu, and M. J. Alexander (2009), Antarctic NAT PSC belt of June 2003: Observational validation of the mountain wave seeding hypothesis, *Geophys. Res. Lett.*, *36*, L02807.

Fueglistaler, S., et al. (2002), Large NAT particle formation by mother clouds: Analysis of SOLVE/THESEO-2000 observations, *Geophys. Res. Lett.*, 29(12), 1610, doi: 10.1029/2001GL014548.

Hanson, D., and K. Mauersberger (1988), Laboratory studies of the nitric acid trihydrate: implications for the south polar stratosphere, *Geophys. Res. Lett.*, *15*(8), 855–858.

Hitchcock, P., T. G. Shepherd, and C. McLandress (2009), Past and future conditions for polar stratospheric cloud formation simulated by the Canadian Middle Atmosphere Model, *Atmos. Chem. Phys.*, *9*(2), 483–495.

Hopfner, M., M. C. Pitts, and L. R. Poole (2009), Comparison between CALIPSO and MIPAS observations of polar stratospheric clouds, *J. Geophys. Res.*, *114*, D00H05.

Jensen, E. J., O. Toon, A. Tabazadeh, and K. Drdla (2002), Impact of polar stratospheric cloud particle composition, number density, and lifetime on denitrification, *J. Geophys. Res.*, *107*(D20), 8284.

Livesey, N. J., W. G. Read, A. Lambert, R. E. Cofield, D. T. Cuddy, and L. Froidevaux (2007), *MLS v2.2 level 2 data quality and description document*, 2nd ed. Jet Propulsion Laboratory.

Marti, J., and K. Mauersberger (1993), A survey and new measurements of ice vapor pressure at temperatures between 170 and 250K, *Geophys. Res. Lett.*, *20*(5), 363–366.

Martins, E., V. Noel, and H. Chepfer (2011), Properties of cirrus and subvisible cirrus from nighttime Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP), related to atmospheric dynamics and water vapor, *J. Geophys. Res.*, 116(D2), doi:10.1029/2010JD014519.

McDonalds, A., S. George, and R. Woollands (2009), Can gravity waves significantly impact PSC occurrence in the Antarctic?, *Atmos. Chem. Phys.*, *9*, 8825–8840.

Murphy, D. M., and T. Koop (2005), Review of the vapour pressures of ice and supercooled water for atmospheric applications, *QJRMS*, *131*, 1539–1565.

Nedoluha, G. E., R. M. Bevilacqua, and K. W. Hoppel (2002), POAM III measurements of dehydration in the Antarctic and comparisons with the Arctic, *J. Geophys. Res.*, *107*(D20), 8290, doi:10.1029/2001JD001184.

Noel, V., A. Hertzog, H. Chepfer, and D. Winker (2008), Polar Stratospheric Clouds over Antarctica from the CALIPSO spaceborne lidar, *J. Geophys. Res.*, *113*, D0225.

Noel, V., A. Hertzog, and H. Chepfer (2009), CALIPSO observations of wave-induced PSCs with near-unity optical depth over Antarctica in 2006-2007, *J. Geophys. Res.*, *114*, D05202.

Pitts, M. C., L. W. Thomason, L. R. Poole, and D. M. Winker (2007), Characterization of Polar Stratospheric Clouds with spaceborne lidar: CALIPSO and the 2006 Antarctic season, *Atmos. Chem. Phys.*, *7*(19), 5207–5228, doi:10.5194/acp-7-5207-2007. [online] Available from: http://www.atmos-chem-phys.net/7/5207/2007/

Pitts, M. C., L. R. Poole, and L. W. Thomason (2009), CALIPSO polar stratospheric cloud observations: second-generation detection algorithm and composition discrimination, *Atmos. Chem. Phys.*, *9*, 7577–7589.

Pitts, M. C., L. R. Poole, A. Dornbrack, and L. W. Thomason (2011), The 2009–2010 Arctic polar stratospheric cloud season: a CALIPSO perspective, *Atmos. Chem. Phys.*, *11*(5), 2161–2177, doi:10.5194/acp-11-2161-2011.

Plougonven, R., A. Hertzog, and H. Teitelbaum (2008), Observations and simulations of a large-amplitude mountain wave breaking over the Antarctic peninsula, *J. Geophys. Res.*, *113*, D16113.

Poole, L. R., and P. McCormick (1988), Airborne lidar observations of arctic polar stratospheric clouds: indications of two distinct growth stages, *Geophys. Res. Lett.*, *15*, 21–23.

Randel, W., F. Wu, O. S. J, K. Rosenlof, and G. E. Nedoluha (2004), Interannual changes of stratospheric water vapour and correlations with tropical tropopause temperatures, *J. Atmos. Sci.*, *61*, 2133–2148.

Schoeberl, M. R., S. R. Kawa, A. R. Douglass, T. J. McGee, E. V. Browell, J. Waters, N. Livesey, W. Read, L. Froidevaux, M. L. Santee, H. C. Pumphrey, L. R. Lait and L. Twigg (2006), Chemical observations of a polar vortex intrusion, *J. Geophys. Res.*, 111(D20), doi: 10.1029/2006JD007134.

Simmons, A., S. Uppala, D. Dee, and S. Kobayashi (2006), ERA-Interim: New ECMWF reanalysis products from 1989 onwards, *ECMWF Newsletter*, 1–53.

Skamarock, W. C., J. B. Klemp, J. Dudhia, D. O. Gill, D. M. Barker, W. Wang, and J. G. Powers (2005), *A description of the advanced research WRF version 2*, NCAR.

Solomon, S. (1999), Stratospheric ozone depletion: A review of concepts and history, *Rev. Geophys.*, *37*(3), 275–316.

Tabazadeh, A., R. P. Turco, K. Drdla, M. Z. Jacobson, and O. B. Toon (1994), A study of Type I polar stratospheric cloud formation, *Geophys. Res. Lett.*, *21*(1), 1619–1622, doi: 10.1029/94GL01368.

Tabazadeh, A., O. B. Toon, B. L. Gary, J. T. Bacmeister, and M. R. Schoeberl (1996), Observational constraints on the formation of Type Ia polar stratospheric clouds, *Geophys. Res. Lett.*, *23*(1), 2109–2112, doi:10.1029/96GL01998.

Tsias, A. et al. (1999), Aircraft lidar observations of an enhanced type Ia polar stratospheric clouds during APE-POLECAT, *J. Geophys. Res.*, 104(D19), 23,961–23,969.

Vomel, H., S. Oltmans, D. Hoffmann, T. Deshler, and J. Rosen (1995), The evolution of the dehydration in the Antarctic stratospheric vortex, *J. Geophys. Res.*, *100*(D7), 13,919–13,926.

Vomel, H. et al. (2007), Validation of Aura Microwave Limb Sounder water vapor by balloon-borne Cryogenic Frost point Hygrometer measurements, *J. Geophys. Res.*, *112*(D24), doi: 10.1029/2007JD008698.

Winker, D., W. Hunt, and M. McGill (2007), Initial performance assessment of CALIOP, *Geophys. Res. Lett.*, *34*, L19803.