Lidar Temperature Measurements During the SOLVE Campaign and the Absence of PSCs from Regions of Very Cold Air

John Burris, Thomas McGee, Walt Hoegy and Paul Newman
Laboratory for Atmospheres, Code 916, Goddard Space Flight Center, Greenbelt, Md. 20771

Leslie Lait, Laurence Twigg and Grant Sumnicht
Science Systems and Applications, Inc., 5900 Princess Garden Parkway, Lanham, Md. 20706

William Heaps
The Instrument Technology Center, Goddard Space Flight Center, Greenbelt, Md. 20771

Chris Hostetler
Radiation and Aerosols Branch, Mail Stop 435, Langley Research Center, Hampton, Va. 23681-2199

Roland Neuber
Alfred Wegner Institute for Polar Research, Telegrafenberg A43, D-14473 Potsdam, Germany

Klaus F. Künzi
Institute for Environmental Physics, University of Bremen, FB1, P. O. Box 330 440, D-28334 Bremen, Germany

Abstract. NASA Goddard Space Flight Center's Airborne Raman Ozone, Temperature and Aerosol Lidar (AROTEL) measured extremely cold temperatures during all three deployments (December 1-16, 1999, January 14-29, 2000 and February 27-March 15, 2000) of the Sage III Ozone Loss and Validation Experiment (SOLVE). Temperatures were significantly below values observed in previous years with large regions regularly below 191 K and frequent temperature retrievals yielding values at or below 187 K. Temperatures well below the saturation point of Type I polar stratospheric clouds (PSCs) were regularly encountered but their presence was not well correlated with PSC observations made by NASA Langley Research Center's Aerosol Lidar co-located with AROTEL. Temperature measurements by meteorological sondes launched within areas traversed by the DC-8 showed minimum temperatures consistent in time and vertical extent with those derived from AROTEL data. Calculations to establish whether PSCs could exist at measured AROTEL temperatures and observed mixing ratios of nitric acid and water vapor showed large areas favorable to PSC formation but lacking PSCs. The flight on December 12th, 1999 encountered large regions having temperatures up to 10 K below the NAT saturation temperature but only small, localized regions that might be identified as PSCs.

Introduction

During the winter of 1999/2000 the Sage III Ozone Loss and Validation Experiment (SOLVE) campaign was carried out concurrently with the Third European Stratospheric Experiment on Ozone (THESEO) to study the processes controlling stratospheric ozone levels from the northern mid-latitudes to the pole. A second SOLVE objective was to provide the correlative measurements necessary to validate the Stratospheric Aerosol and Gas Experiment (SAGE III) satellite instrument. Temperature was a key measurement for both the expected satellite validation and its impact upon Polar Stratospheric Clouds (PSCs). PSCs play a central role in the loss of Arctic ozone by providing surfaces on which inactive reservoir species are converted to active species capable of destroying ozone. Their formation depends critically upon an air parcel's temperature and time history, its altitude and the mixing ratio profiles of nitric acid and water vapor [*Solomon,* 1999, and references within; *World Meteorological Organization*, 1999] . Because of temperature's pivotal role in the formation and duration of PSCs, NASA Goddard Space Flight Center flew a new lidar on the DC-8 to provide temperature profiles at high vertical resolution. Lidar retrievals from an aircraft platform offer distinct advantages over a satellite by providing high precision temperature profiles coincident with measurements of ozone, aerosols, clouds and water vapor throughout the region of interest. These measurements were designed to help understand under what conditions PSCs form and persist by identifying regions in the lower Arctic stratosphere where temperatures dropped below the local saturation point. In this paper we present profiles of extremely cold temperatures made by the Airborne Raman Ozone, Temperature and Aerosol Lidar (AROTEL) from the DC-8. Comparisons are made between these cold temperature measurements and those retrieved by balloon sondes. Results comparing PSC observations by the Langley Research Center's (LaRC) Aerosol Lidar with calculations showing locations favorable to the existence of PSCs as a function of T_{arotel} - T_{nat} and measured values of nitric acid and water vapor are presented.

Instrument

AROTEL is a combined aerosol, ozone and temperature lidar designed for operation on an aircraft platform such as the DC-8. AROTEL's design utilized a hardware configuration that has been successfully implemented on other Goddard ground based and airborne lidars. AROTEL employed two lasers, a xenon chloride excimer operating at 308 nm and a Nd:YAG with outputs at 1064, 532 and 355 nm. A differential absorption measurement of stratospheric ozone was made using the 308 and 355 nm wavelengths [*McGee et al.*1993, *McGee et al.* 1995; *Steinbrecht*, 1994]. Temperatures were retrieved using elastic (Rayleigh) molecular scattering at 355 nm and inelastic (Raman) scattering at 387 nm (from 355 nm). The Raman signal was derived from the inelastic scattering of 355 nm radiation by molecular nitrogen and had an energy shift of ~2331 cm^{-1}. Signals were captured using a telescope with a 40 cm diameter primary mirror. They were then separated into their constituent wavelengths using beam splitters and bandpass filters with high out-of-band blocking. Gated photomultipliers (PMTs) were used for signal detection on all UV channels because of their sensitivity, linearity and low noise. Both the non-linear behavior in the PMT's gain and signal induced noise were minimized by limiting the detected signal's dynamic range associated with any one PMT to less than five orders of magnitude. Analog signal detection was implemented for returns originating within 1 km of the aircraft while photon counting was employed to beyond 60 km. This configuration permitted both a compact design and high detector throughput but, because of the relatively large filter bandpass, limited temperature retrievals to solar zenith angles

(SZAs) 85 degrees. Temperature retrievals to 60 km were made for SZA's 95 degrees, for SZA's less than 95 degrees the maximum reportable altitude was progressively reduced as the SZA decreased due to interference from the sun. Besides ozone and temperature, AROTEL measured the lidar scattering ratio at 355 nm which provides a picture of aerosol scattering properties in the UV. The LaRC Aerosol lidar retrieved aerosol backscattering at 1064 nm and depolarization at 532 nm using the output from AROTEL's YAG. A more detailed description of the AROTEL instrument can be found in *McGee et al* [submitted to *Opt. Eng.*].

Measurement Technique

Temperature profiling via lidar is a well established technique that has been successfully employed for many years [*Evans et al.*, 1997; *Gross et al.*, 1997*; Hauchecorne and Chanin* , 1980; *Keckhut et al.*, 1990; *Leblanc et al.*, 1998; *Steinbrecht*, 1994]. For the temperature measurement, AROTEL used a Nd:YAG laser that produced between 200-300 millijoules (per pulse) at 355 nm at a pulse repetition rate of 50 Hz. The large UV backscattering cross sections maximized signal returns without interference from ozone absorption. Data was acquired using both elastic and inelastic scattering. Inelastic scattering enhanced AROTEL's retrieval capabilities in the presences of clouds and aerosols due to the absence of Mie backscattering at the detected wavelength. Temperatures both within and below clouds, PSCs and aerosol layers have been successfully derived from the inelastic data. Currently, lidar ratios <1.4 can be utilized for temperature retrievals. When compared to ground based lidar systems, an aircraft platform at flight altitude experiences less signal attenuation by not having to traverse the troposphere and a major enhancement in return signals by minimizing the range^{-2} loses. Temperature retrievals were made within a much shorter time period than is possible from a much larger, ground-based, lidar system.

Both elastic and inelastic returns were directly proportional to atmospheric number density. Removal of the signal's range^{-2} dependence provided a relative measurement of atmospheric density when tied to a model; absolute densities could be recovered by normalizing the measured profile against either a balloon sonde measurement of density at a reference altitude or a model (CIRA86). Because Rayleigh scattering introduces significant signal attenuation in the UV, a correction for it was made using the calculated density profile. In recent years stratospheric lidar retrievals have benefitted from the natural removal of volcanic aerosols injected by the eruption of Mt. Penatubo in 1991; the presence of these volcanic aerosols had made retrievals difficult below ~30 km.

Temperature profiles are derived from the measured atmospheric densities utilizing the ideal gas law [PV = nkT] and the assumption that local hydrostatic equilibrium (dP/dz = g) prevails. It is important to note that temperature profiles are retrieved using changes in relative, not absolute, number density. The resulting temperature derivation is thus insensitive to the absolute values for a number of instrumental parameters including detector transmission efficiency, photocathode quantum efficiency, laser output power and size of the telescope's primary mirror. These parameters, however, do affect count rates and hence measurement uncertainties. The derived temperature at a geometric altitude z_i is given by

$$T(z_i) = T(z_0) * \frac{n_0}{n_i} + \frac{M}{R} \sum_{j}^{i} \frac{g_j n_j \Delta z_j}{n_0} \quad (1)$$

where z_0 is the tie-on altitude, n_j the measured atmospheric number density, M the average molecular mass, R the gas constant, g the gravitational constant and z the measurement interval. The temperature algorithm must be initialized at the highest measurement point (typically between 50-60 km for AROTEL) employing either a pressure or temperature tie-on; temperature was employed by AROTEL.

Errors

Errors in retrieved temperature arise from a number of sources. For reported temperatures above ~40 km, uncertainties associated with the tie-on temperature exert a significant impact. The retrieval algorithm is initialized using either the CIRA86 model (at 60 km geometric altitude) or the Goddard Data Assimilation Office (DAO) temperature product (at the 0.5 mbar pressure level). DAO temperatures could be obtained in advance of a flight for use in real time retrievals once the flight track had been determined. The errors for this data set using a tie-on at the 0.5 mbar pressure level (~55 km) are reported to be 5K [*Steve Pawson, private communication*]. The CIRA86 model is based on climatology with both monthly and zonal average temperatures, pressures and densities from 0 to 120 km in 5 km intervals. Observed differences between DAO and CIRA86 were generally less than 10 K at the tie-on altitude. This error is negligible three scales heights (~20 km) below the tie-on point. A tie-on at 60 km with an uncertainty of T generated a residual error of ~ T*exp(-3) at 40 km. Since the DAO tie-on at 0.5 mbar was generally better than 5 K, residual tie-on errors at 40 km were typically 0.3 K. For the CIRA86 model an uncertainty of 15 K at 60 km introduces a tie-on error of ~0.7 K at 40 km. Counts derived from other than molecular backscattering are another error source. These include dark counts from the PMTs and those generated by light sources whose output was within the bandpass of AROTEL's detectors. Background counts were subtracted from the acquired data before deriving temperatures. AROTEL's data channels were operated in a regime where the background count rate was linear with time; background was calculated by averaging counts in channels sufficiently distant in time such that no backscattering is observed. Statistical uncertainties associated with photon counting can exert a major impact for the low signal levels typically seen at the maximum altitude range of each data channel. These errors are reduced by integrating returns over a longer time period, decreasing the vertical resolution and smoothing the resulting data. Photon counting at high data rates can saturate both discriminators and counting electronics [*Donovan et al.*, 1993]. A correction for these induced non-linearities in the counting rate was employed in the retrieval algorithm. Additional error sources include uncorrected signal extinction due to aerosols and clouds. Representative measurement uncertainties for AROTEL temperature retrievals as a function of altitude during the SOLVE mission are given in Figure 1. Comprehensive treatments of both the temperature retrieval technique and associated errors can be found in *Gross et al.*[1997]; *Keckhut et al.,*[1993]; *Leblanc et al.*, [1998] and *Steinbrecht*, [1994].

Intercomparison

AROTEL temperatures were compared to those retrieved by both balloon sondes and the Meteorological Measurement System (MMS) [*Scott et al.*, 1990] during SOLVE [*Burris et al.*, this issue]. For the sonde comparisons a coincidence was defined as occurring for DC-8 flight tracks approaching within 100 km of the launch site within 12 hour of the retrieval (Over all three deployments the mean difference in time between sonde launch and the DC-8's closest approach was 3.8 hours, the mean separation was 35 km. The maximum difference in time was 11 hours.) . There were 28 separate elastic temperatures profiles that meet this criterion, for inelastically derived temperature 23 profiles were used. Elastically derived temperatures were ~1.1±0.6 K colder than sonde values while inelastic retrievals averaged ~1.2±1.1 K colder than sonde values [see *Kirkhut*, 1993 for a discussion of sonde measurement errors].

MMS made continuous in-situ temperature measurements at the ER-2's flight altitude. Coincidences between AROTEL and MMS existed for measurements acquired within 30 km of each other on the same flight date. Analysis of data from three separate flights showed AROTEL was from 2 to 3 degrees colder than reported MMS values for both elastic and inelastically derived temperatures. These differences are believed to be primarily due to an incomplete correction of aerosol extinction on the data channels.

AROTEL Temperature Measurements During SOLVE

AROTEL temperature profiles are calculated from the aircraft to ~60 km in 150 meters intervals with an effective vertical resolution of ~0.5-0.6 km [see *Burris et al.*, this issue of JGR]. The horizontal footprint ranged from 4 to 70 km depending on the integration time and signal to noise ratio. Temperatures are reported in geometric altitude and can be retrieved at URL: http://hyperion.gsfc.nasa.gov/Public/Ground_based/arotel/solve1. During the first deployment, the DC-8's geometric altitude was provided by the DADS (Data Acquisition and Distribution System) GPS system with a stated precision of ±50 meters. During the second and third deployments a differential GPS altitude was available via post processing of the flight data with a reported accuracy of ±1 meter over most of the DC-8's flight track [*Carl Sorenson, private communication*]. Temperatures derived from elastic backscatter data are reported for altitudes from the aircraft to approximately 45 km, inelastic scattering data was employed for retrievals from the aircraft to as high as ~26 km altitude depending on local conditions. During the winter of 1999-2000, low temperatures were observed throughout the Arctic polar region (Figure 2). Regions of very cold air were repeatedly observed by AROTEL during all three SOLVE deployments. Temperatures at or below 192 K were retrieved on many dates with extensive regions being observed during flights on December 7, 1999, December 10, 1999, December 12, 1999, January 14, 2000, January 16, 2000, January 20, 2000, January 23, 2000, January 25, 2000, January 27, 2000 and March 5, 2000. Localized temperatures at or below 188K were also frequently observed with particularly widespread regions occurring during flights on December 12, January 20, January 23 and January 25. In regions having only background stratospheric aerosols, as determined by AROTEL and Langley Aerosol data, low temperatures were derived utilizing both elastic and inelastic backscatter signals (Figure 3 displays elastic derived temperature for December 12 when minimal PSC activity was observed by the LaRC Aerosol Lidar). Temperatures below 191 K existed over a large area for altitudes ranging from ~18 km to almost 30 km on flight date December 12. The DC-8's flight track on December 12th (Figure 4) is shown over a PV map at 480 K and displays the large area over which the temperature soundings were made on this date. Figure 5 displays elastically derived temperatures between 185 and 195 K, at 2 K resolution, with significant structure evident in the temperature field. Broad areas between 187 and 189 K are seen from 20-26 km altitude and, in some cases, extending over 800 km (~1 hour flight time) horizontally. Differences between temperatures retrieved employing elastic and inelastic returns as a function of altitude were calculated for December 12 and are displayed in Figure 6; the mean differences between 20 and 25 km was ~0.3%. These small differences were due to the absence of PSCs and its resulting Mie interference on the elastic channel.

The strong influence exerted by PSCs on retrieved temperature is displayed in Figure 7 for the flight on December 7. Between 1230 and 1430 GMT the DC-8 flew beneath a cloud observed by the Langley Aerosol Lidar. Both inelastic and elastically retrieved temperatures within the PSC are shown. 1-sigma uncertainties are presented for both retrieved temperatures. The vertical extent of the cloud, as determined from its impact on the temperature retrieval, ranged from ~21 to 23 km and is consistent with the data acquired by Langley. The maximum temperature difference within the cloud between elastic and inelastic scattering was ~12 K. The backscattering ratio (BSR - defined as ($_{rayleigh}$ + $_{aerosol}$)/ $_{rayleigh}$)) at 355 nm attained a peak value of ~1.07 inside the cloud which corresponds to an optically thin PSC. The cloud's optical properties as measured by the LaRC Aerosol Lidar are consistent with its identification as a Type 1a PSC. Elastic and inelastically retrieved temperature for the same date but just outside the cloud generally had differences of less than ~2 K for altitudes ranging between 18 and 28 km. The mean temperature difference between inelastic and elastically determined temperatures was -0.2 ± 1.5 K. Slightly warmer temperatures were observed on this flight and they lacked the extensive vertical and horizontal structure seen on December 12 (Figure 5).

Temperature retrievals are currently possible for backscattering ratios (at 355 nm) that are less than approximately 1.4. Within clouds and aerosol layers having larger BSR values, retrievals are not feasible with this technique unless a correction for extinction is included. The issue of using inelastically scattered returns to derive temperatures within optically thin clouds and aerosols is addressed in a paper currently in preparation.

Balloon Sonde Temperature Measurements

AROTEL repeatedly encountered temperatures at or below 187 K during SOLVE.

Temperatures this cold are infrequently observed in the Arctic because the polar vortex is neither as stable nor long lived as in the Antarctic. Although few satellite measurements were available, numerous balloon sondes were launched during SOLVE that provided a detailed, continuous, record of temperatures to beyond ~25 km altitude. The cold temperatures encountered in the Arctic winter caused the balloons to burst before reaching their maximum altitude and permitted only sporadic sampling above ~25 km even when special techniques were employed to protect the balloons. These measurements occurred both within and without the polar vortex and generally covered the critical band between 18 and 25 km where PSCs were most frequently observed by the LaRC Aerosol Lidar. This study encompassed a region regularly traversed by the DC-8 during SOLVE although these sonde measurements were not generally coincident with those of AROTEL. Nonetheless, these sonde temperature records permit an independent assessment of the frequency and extent of the very cold temperatures observed by AROTEL throughout the SOLVE deployment. These meteorological sondes were launched from Ny Ålesund (78.9N, 11.9E), Danmarkshaven, Greenland (76.8N, 18.8W), Scoresbysund, Greenland (70.5N, 22.0W), Jan Mayen, Norway (70.9N, 8.7W) , Bjornoya Island, Norway (74.5N, 19.0E) and Keflavik, Iceland (64.0N, 22.6W). All six sites recorded low temperatures throughout the period. Figure 8 presents temperature retrievals by a sonde and AROTEL for inelastically derived temperatures during a coincident measurement; the sonde recorded a minimum temperature of ~189 K at 23 km, AROTEL measured ~188 K at the same altitude approximately 1.5 hours after the sonde's launch. Figure 9 plots the minimum temperature reported by these six sites as a function of date and altitude between November 30th and March 15th. Temperatures below 190 K were regularly seen from about December 15 through early February, 2000. The minimum temperature, 183 K, was observed on January 18 at Danmarkshaven, Greenland. As shown in Figure 9 there was the expected steady downward progression in minimum temperature as the winter progressed. The coldest temperatures were noted for the sites located at Danmarkshaven and Scoresbysond, Greenland and Ny Ålesund. The temperatures from mid-December into late January at Danmarkshaven, Scoresbysond and Ny Ålesund show very cold temperatures existing from ~20 to 25 km altitude for a considerable period of time. The picture provided by the sonde data is thus broadly consistent with the AROTEL retrievals in vertical extent, duration and lowest minimum value.

Correlations between observed PSCs and measured low temperatures.

Although temperatures significantly below T_{nat} (Figure 10) were repeatedly encountered, these cold regions did not correlate well with PSC observations made by the Langley Aerosol Lidar. Localized PSCs were observed on December 7, December 10, January 16 and January 29; the flight on December 12 showed minimal activity while data from January 20th, 23rd and 25th showed extensive PSC development. During the December 7th flight, regions with temperatures below 192 K were observed at altitudes ranging from 20 to 30 km for most of the 9 hour flight with PSCs being observed only sporadically. On the December 12th flight, the Langley Aerosol Lidar reported minimal activity during six hours of observations while AROTEL observed large regions below 191 K between 20 and 30 km altitude. Temperatures derived using both elastic and inelastic returns frequently showed areas at or below 187 K but with little obvious evidence for Type 1 or Type 2 PSCs in either the temperature or PSC record. These observations led to a study to determine where PSCs could exist as a function of $T-T_{nat}$ and the local mixing ratios of nitric acid and water.

To locate PSCs in the data, backscattering (BSCT) values from the 1064 nm channel of the Langley Aerosol lidar that were co-located with DAO temperatures greater than 200 K were identified. The calculated mean of these warm-air scattering ratios plus four times their standard deviation was used as the threshold value below which scattering was considered to have originated from background aerosols. In the plots of scattering ratios versus temperature, each measurement is color-coded by whether it is expected to be associated with a PSC. The NAT condensation temperature and the frost point are computed from the *Hanson and Mauersberger* [1988] criteria using (a) the pressures obtained by converting the analyzed geopotential heights (on pressure surfaces) to geometric heights and interpolating to the lidar profiles, (b) a quadratic best fit to the HNO_3 profile data from the ASUR instrument [*Mees et al.,* 1995] for the flight, and (c) a constant H_2O value of 5 ppmv (consistent with in situ measurements of stratospheric water vapor made by the Harvard H_2O instrument on board the ER-2). When temperatures (AROTEL or DAO) are less than the NAT condensation temperature, the region is colored red. Temperatures below the frost point are identified in yellow. All other points are colored gray.

The results for December 12th, January 16th and January 29th are presented in Figures 11-13. Regions where both Type 1 and 2 PSCs could exist but were not observed for temperatures up to 10 K colder than T_{nat} are identified in the data. For January 16th, PSCs were observed in over ~30% of the region where theory predicted they could exist. Theory and observations agreed that the clouds were not Type 2 PSCs. On January 29th agreement between observation and theory was better with PSCs being observed over approximately 70% of the area within which conditions supported the existence of PSCs. Theory predicted Type 2 PSCs could exist for the lowest measured temperatures however the backscattering ratios at 1064 nm were less than 5 which suggest that the observed clouds were not Type 2 PSCs. The observations on December 12th were interesting. Theory predicted that both Type 1 and 2 PSCs could exist for the temperature profiles retrieved by

AROTEL (figure 11). A small spike (~1.8) was identified in a region at a temperatures consistent with a type 2 PSC and another small spike (~2.3) suggestive of a type 1 PSC. In figure 11, the number of data points having a BSCT value of >1.5 totaled only 200 out of a possible 104,000 for temperatures less than or equal to T_{nat}. Using DAO instead of AROTEL temperatures gave results similar to AROTEL's for both January 16th and 29th (lower right panel in figures 12-13). The coldest DAO temperatures on December 12th were several degrees warmer then those measured by AROTEL. Calculations for this date showed two small spikes (BSCT of 1.8 and 2.2) that could be identified as type 1 PSCs and another spike (BSCT of 2.1) that was ~4 K above T_{nat} . DAO temperatures for all three dates are not consistent with the existence of Type 2 PSCs along the DC-8's flight track. The AROTEL temperature uncertainties for these data sets were typically less than 1 K and for December 12 were usually much less.

Discussion:

AROTEL temperature retrievals showed widespread areas of very cold air during all three deployments of the SOLVE campaign. Temperatures below 191 K were regularly encountered over large areas and localized regions at or below 189 K were frequently observed. Minimum temperatures retrieved by meteorological sondes launched during the mission are consistent with those derived from AROTEL data within the Arctic both in time and vertical extent. Comparisons between PSC observations by the Langley Aerosol Lidar and calculations designed to show whether PSCs could exist based on retrieved AROTEL temperatures were made. Agreement between observation and theory appears to be reasonably good for both January 16th and 29th. The data from the December 12th flight was unusual in that there were large regions having temperatures up to ten degrees below T_{nat} with minimal observational evidence for PSCs. The measured values for the 1064 nm backscatter ratio would rule out the presence of type 2 PSCs. In addition, the December 12th depolarization data acquired by the Langley Aerosol lidar showed no depolarization within the region reported here (Figure 14). This would indicate that both type 1a and type 2 PSCs were absent. It also implies that NAT rocks were absent. NAT rocks, large (>3 microns) HNO_3 hydrate particles in optically thin clouds extending over thousands of kilometers [*Luo et al.,*this issue of JGR; *Fahey et al.*, 2001], were first observed during SOLVE/THESEO. The peak backscatter value for background aerosols on 991212 was between 1.4-1.5. Except as noted above, this is also the case for temperatures below T_{nat} and suggest that background aerosols account for most of this scattering. A study of backscattering versus altitude for discrete 1 K temperature increments between 185 and 195 K was done to see if identifiable changes in the distribution of backscatter values (at 1064 nm) existed between regions that were/were not favorable to PSCs. Intense but extremely localized perturbations in the backscatter ratio were observed in the data at 18 and 22-23 km for temperatures that were 4-6 K below $T_{nat.}$ These perturbations are consistent with the data displayed in the lower left panel of figure 11. No such perturbations were observed for backscattering originating with background aerosols. This observation would be consistent with the existence of one or more small and high localized type 1b PSCs for 991212.

In conclusion, both the PSC and temperature data strongly suggest the absence of nat rocks and type 1a and type 2 PSCs for this date. Direct evidence in terms of actual data points above and below a BSCT of 1.5 (for $T < T_{nat}$) and indirect evidence in terms of the distribution in backscatter values for BSCT < 1.5 and $T < T_{nat}$ suggest that if type 1b PSCs exist they are confined to a very limited region.

The absence of PSCs within the Arctic under conditions believed favorable to their existence has been observed elsewhere. For the December SOLVE deployment, *Hitchman* [this issue of JGR] noted that, while cold temperatures existed over large areas with minimal denitrification, PSCs were not common within the cold pool. It was hypothesized that, in addition to cold temperatures, local, dynamically-induced upward motion is required for PSC formation. *Kawa* [1992] found that PSCs were not observed in the Arctic until the saturation ratio of nitric acid with respect to NAT was greater than ten, in the Antarctic they appeared at saturation ratios of one and above. He noted that this difference could not be accounted for in terms of known measurement uncertainties. Finally, *Voigt* [2000] using data acquired on a balloon flight from Kiruna, Sweden on January 25, 2000 found there was not a close correlation between PSCs and cold temperatures (as low as 185 K); the expected correlation was however observed in data acquired on flights in January, 1998.

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Figure Captions

Figure 1. Calculated temperature uncertainties for both Rayleigh (elastic / solid squares) and Raman (inelastic / open squares) temperatures derived on flight date 991207 as a function of altitude. The large changes observed in elastically derived temperature uncertainties at 26 and 39 km are due to changing between the low to mid-altitude and the mid to high altitude channels. Significantly more signal is directed towards the high altitude channel to extend its vertical range - at lower altitudes this translates into higher count rates (and lower measurement uncertainties).

Figure 2. Minimum reported temperatures within the Arctic for the 1999-2000 winter at 30 hPa (~23 km). The minimum temperature observed was 183 K in mid-January.

Figure 3. Elastic temperatures retrieved for flight date 991212. Temperatures were calculated from data acquired with five minutes of integration . After smoothing the effective vertical resolution is ~0.6 km.

Figure 4. Map of potential vorticity on the 480 K potential temperature surface for 12 UTC on December 12th, 1999. Data are from the UARS UKMO assimilation product and are in units of 10^6 K m^2/kg/s. The 195 K temperature contour is marked in red. The DC-8 flight path is shown in yellow. The 480 K potential temperature surface is at approximately 20 km altitude in this region on this date.

Figure 5. High resolution elastic temperatures between 185 K and 195 K for 991212. Large regions below 191 K are visible for most of the flight.

Figure 6. Differences between elastic and inelastically derived temperatures for 991212 calculated between 13 and 25 km altitude and averaged over the entire flight. From 20 through 25 km the differences averaged <0.3%.

Figure 7. Elastic (solid squares) and inelastically (solid triangles) derived temperatures acquired within an optically thin PSC on 991207. 1-sigma uncertainties are shown.

Figure 8. Balloon sonde and AROTEL temperatures (inelastically derived data) for 991212. The AROTEL profile was acquired at 17.36 GMT, the sonde was launched from Ny Ålesund at 16.03 GMT. The minimum sonde temperature was ~189 K at 23 km, AROTEL measured ~188 K for this altitude.

Figure 9. Plot of minimum temperatures versus date and altitude for sondes launched from Ny Ålesund (78.9N, 11.9E), Danmarkshaven, Greenland (76.8N, 18.8W), Scoresbysund, Greenland (70.5N, 22.0W), Jan Mayen, Norway (70.9N, 8.7W) , Bjornoya Island,

Norway (74.5N, 19.0E) and Keflavik, Iceland (64.0N, 22.6W) from November 30, 1999 through March 15, 2000.

Figure 10. T_{nat} (dashed line) and T_{sat} (dot-dash line) as a function of altitude for 5 ppmv of water and 6 ppbv of nitric acid. The solid line is a model temperature profile for 12/15, the dotted line is the US Standard Atmosphere's temperature.

Figure 11. The upper left plot is of LaRC backscatter data at 1064 nm used to identify PSCs on flight date 991212. The upper right plot is of AROTEL temperatures for the same altitude range as the LaRC data. The lower left plot displays the backscatter ratio data measured by LaRC as a function of T_{AROTEL} - T_{NAT} . The red area highlights conditions where Type I PSCs could exist for AROTEL temperatures and measured values of nitric acid and water. The yellow region highlights where Type II PSCs could exist. The dotted line delineates the regions where PSCs were observed from where they were not. The lower right plot is identical to that on the left except for the use of DAO, rather than AROTEL, temperatures for the PSC existence calculation.

Figure 12. The upper left plot is of LaRC backscatter data at 1064 nm used to identify PSCs for flight date 000116. The upper right plot is of AROTEL temperatures for the same altitude range as the LaRC data. The lower left plot displays the backscatter ratio data measured by LaRC as a function of T_{AROTEL} - T_{NAT} . The red area highlights conditions where Type I PSCs could exist for AROTEL temperatures and measured values of nitric acid and water. The yellow region highlights where Type II PSCs could exist. The dotted line delineates the regions where PSCs were observed from where they were not. The lower right plot is identical to that on the left except for the use of DAO, rather than AROTEL, temperatures for the PSC existence calculation.

Figure 13. The upper left plot is of LaRC backscatter data at 1064 nm used to identify PSCs 000129. The upper right plot is of AROTEL temperatures for the same altitude range as the LaRC data. The lower left plot displays the backscatter ratio data measured by LaRC as a function of T_{AROTEL} - T_{NAT} . The red area highlights conditions where Type I PSCs could exist for AROTEL temperatures and measured values of nitric acid and water. The yellow region highlights where Type II PSCs could exist. The dotted line delineates the regions where PSCs were observed from where they were not. The lower right plot is identical to that on the left except for the use of DAO, rather than AROTEL, temperatures for the PSC existence calculation.

Figure 14. Retrieved depolarization at 532 nm by the Langley Aerosol lidar on flight date 991212.