Validation of Temperature Measurements from the Airborne Raman Ozone Temperature and Aerosol Lidar During SOLVE

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Abstract. The Airborne Raman Ozone, Temperature and Aerosol Lidar (AROTEL) participated in the recent Sage III Ozone Loss and Validation Experiment (SOLVE) by providing profiles of aerosols, polar stratospheric clouds (PSCs), ozone and temperature with high vertical and horizontal resolution. Temperatures were derived from just above the aircraft to ~60 kilometers geometric altitude with a reported vertical resolution of ~0.6 km. The horizontal footprint varied from 4 to 70 km. This paper explores the measurement uncertainties associated with the temperature retrievals and makes comparisons with independent, coincident, measurements of temperature. Measurement uncertainties range from 0.1K to ~4K depending on altitude and integration time. Comparisons between AROTEL and balloon sonde temperatures retrieved under clear sky conditions using both Rayleigh and Raman scattered data showed AROTEL ~1 K colder than sonde values. Comparisons between AROTEL and the Meteorological Measurement System (MMS) on NASA's ER-2 show AROTEL being from 2-3 K colder for altitudes ranging from 14 to 18 km. Temperature comparisons between AROTEL and the United Kingdom Meteorological Office's model showed differences of approximately 1 K below ~25 km and a very strong cold bias of ~12 K at altitudes between 30 and 35 km.

1. Introduction

NASA Goddard Space Flight Center's Airborne Raman, Ozone, Temperature and Aerosol Lidar (AROTEL) participated in the SAGE III Ozone Loss and Validation Experiment (SOLVE) campaign during the winter of 1999/2000. This campaign, conducted jointly with the Third European Stratospheric Experiment on Ozone (THESEO), was based in Kiruna, Sweden and combined measurements from ground, balloon, aircraft and space-based platforms to study the processes governing ozone loss within the polar stratosphere. A second goal of the SOLVE campaign was to have been the validation of data products from the SAGE III instrument had it been launched. A key measurement for both satellite validation and ozone loss was temperature. Temperature plays a critical role in determining both the extent and duration of polar stratospheric clouds (PSCs) on whose surfaces inactive reservoir species are converted to reactive species that destroy ozone. Lidar temperature retrievals from an aircraft platform provide a unique data set that is acquired simultaneously with other onboard measurements. Retrievals can be made at times and locations inaccessible to a satellite instrument thereby affording a more comprehensive picture of atmospheric phenomena. AROTEL temperature retrievals offered significantly better vertical and horizontal resolution than was possible with a satellite instrument; the horizontal footprint can be varied between 4 and 70 kilometers with a reported vertical resolution of ~600 meters. In this paper we will present comparisons between AROTEL temperatures and those derived from coincident balloon sondes measurements. Comparisons between AROTEL temperatures and those retrieved by the Meteorological Measurement System (MMS) on the ER-2 will be made for dates on which the DC-8 and ER-2 flew coincident tracks. A study of temperature differences between AROTEL and those derived from the United Kingdom Meteorological Office (UKMO), National Centers for Environmental Prediction (NCEP) and NASA Goddard Space Flight Center's Data Assimilation Office (DAO) models will be made.

1.1 Instrument and Technique

AROTEL is a Rayleigh/Raman lidar that utilizes a hardware configuration previously validated on other Goddard Space Flight Center lidars. Temperature data was acquired using a laser operating at a wavelength of 355 nm. This wavelength permitted the use of strong Rayleigh and Raman UV backscattering while avoiding signal attenuation from the temperature dependent Huggins ozone band. Photon counting was utilized for data acquisition because of its large dynamic range and sensitivity to low signal levels. Photomultiplier tubes (PMTs), pulse height discriminators and multi-channel scalar cards constituted the detection system and were employed for all channels. Wavelength discrimination was provided by bandpass filters with ~1 nm bandpass and high transmission. Filters enabled the design of a compact detector with high detection efficiency but restricted temperature retrievals to solar zenith angles (SZAs) > 95 degrees. A complete discussion of the AROTEL instrument is provided in *McGee et al.*[submitted to *Appl. Opt.*]. AROTEL's operational characteristics relevant to the temperature measurement are provided in Table 1.

AROTEL temperatures were retrieved using an algorithm validated in numerous previous field campaigns. Temperatures were derived from data created by both elastic

(Rayleigh) molecular scattering and inelastic (Raman) scattering by molecular nitrogen(see *Ansmann et al.*, 1992; *Evans et al.*, 1997; *Gross et al.*, 1997; *Keckhut et al.*, 1990; *Hauchecorne et al.*, 1980; *Hauchecorne et al.*, 1992). These data are directly proportional to atmospheric number density. Temperatures are calculated using relative, not absolute, number densities thus making the retrieval insensitive to many instrumental parameters such as telescope size, photomultiplier tube quantum efficiency and detector transmission efficiency. Elastic scattering, with returns ~2000 times more intense than those from inelastic backscattering, provided numerous advantages such as more distant ranging, enhanced vertical resolution, better signal to noise and more rapid data acquisition. The sensitivity of molecular scattering to interference from Mie scattering by aerosols, thin cirrus and PSCs, however, generally limited its use to altitudes above 20-25 km (unless otherwise noted all altitudes are geometric). Because of the energy shift associated with inelastically scattered photons (~2331cm^{-1} for N_2), it is, to first order, not impacted by Mie scattering from clouds and aerosols. Reported inelastic temperature retrievals extend from the aircraft to ~26 km altitude, elastic retrievals are reported to ~45 km. Besides temperature, AROTEL provided several additional data products including ozone profiles to ~30 km (via differential absorption at 308 and 355 nm) and both backscatter and extinction data for aerosols at 355 nm (see *McDermid et al.*, 1991*, McGee et al.*1993, *McGee et al.* 1995; *Steinbrecht*, 1994). The Langley Aerosol lidar used data derived from scattering at 532 and 1064 nm to calculate aerosol backscattering and extinction coefficients and depolarization ratios (532 nm). These data were acquired simultaneously with the AROTEL aerosol (UV), ozone and temperature retrievals.

1.2 Measurement Uncertainties

Temperature uncertainties, in the absence of optically thick clouds and aerosols, are influenced by several mechanisms [*Ferrare et al.*, 1995; *Keckhut et al.,*1993; *Leblanc et al.*, 1998; *Steinbrecht,* 1994]. Statistical errors inherent in photon counting are the most important source of random errors. They depend critically upon whether elastic or inelastic returns are used because of the marked differences in scattering cross section between the two processes. Rayleigh scattering losses experienced by both the laser pulse and return signal can cause significant signal loss for both elastic and inelastic data, a correction is applied to all data. Since the instrument actually measures the relative atmospheric density profile this correction is made using the retrieved density. This is accomplished by setting the retrieved density at 30 km to a value derived from either a model or independent measurement at this altitude and using this value to renormalize the remainder of the data set. A study of the temperature algorithm was undertaken to determine how errors in the density normalization impact the retrieved temperature. Errors in the renormalized atmospheric densities are expressed through the Rayleigh extinction correction, a 10% error in the normalization density translates into a maximum 2 K error in temperature at 12 km. Signals are also impacted by the desired vertical and horizontal measurement resolution, the range between aircraft and scattering volume and how the signal is partitioned between the various detectors. The large dynamic range in returns from just above the aircraft to beyond 60 km altitude requires splitting the return signals among multiple PMTs. Systematic errors can arise from intense returns that introduce non-linearities into the PMT's background count rate thereby make it difficult to ascertain the true count rate for low signal levels [*Donovan et al.,*1993; *Steinbrecht*, 1994]. These non-linearities were minimized by gating (effectively setting the PMT gain to zero while the outward propagating laser pulse is still close to the lidar). Extremely high count rates also affect signal linearity because of discriminator saturation, a correction to convert observed into actual counts was made [*Donovan et al.*, 1993]. Except in the presence of optically thick clouds, extremely high count rates were generally not an issue because all PMTs were operated in a regime where the background counts remained linear. Systematic errors can also arise when extinction due to optical thin aerosols is not completely taken into account. They will manifest themselves as a slight cold bias over the affected altitude range.

Additional noise sources included dark counts associated with the PMTs (minimal) and counts that, depending upon operational conditions, originated with the sun or moon and fell within the detector's bandpass. Aurora were observed on several occasions but their emission was outside the instrument's detection bandpass. For temperature retrievals the full capability of the instrument was realized for solar zenith angles (SZA) 95 degrees, however, temperatures were derived for SZAs as low as 85 degrees. As the SZA decreased, the increased solar background restricted the maximum altitude of the measurement and significantly impacted the measurement uncertainty due to the high background count rates.

The retrieval algorithm must be initialized and this introduces an uncertainty above ~40 km. Temperatures were calculated using a routine initialized at the maximum altitude of the retrieval, typically between 54 and 60 km. Initialization was accomplished utilizing either a model (CIRA86) or temperature profile calculated using data from the Data Assimilation Office at Goddard Space Flight Center for the aircraft's projected flight track. Errors using DAO temperatures at the 0.5 mbar pressure level are 5 K (Steve Pawson - private communication). At 40 km the impact of the DAO tie-on uncertainty is < 0.7 K and at 30 km it falls below 0.1 K. Using a model to tie-on at 60 km can introduce larger uncertainties since it is based upon the general climatology. The CIRA86 model provided temperatures, pressures and densities from 0 to 120 km in 5 km intervals based on both monthly and zonal averages. CIRA86 and DAO generally differed by less than 10 K at the tie on altitude, this introduced an error in the retrieved temperatures at 35 km of < 0.5 K.

In addition to clear air, temperatures (using elastic and inelastically generated data) can be retrieved above optically thin clouds and aerosols (temperatures are retrieved within clouds using Raman returns for a lidar backscattering ratio of 1.4), however, the additional signal attenuation increases reported errors. Temperatures can occasionally be derived above optically thick clouds depending on signal levels. Temperature retrievals within PSCs or aerosol layers having lidar ratios greater than 1.4 are not considered here because of the inability to adequately correct for Mie scattering and signal attenuation. Neither can they be currently derived below clouds with ratios greater than ~1.4. A detailed discussion of temperature profiling within optically thin clouds and aerosol layers will be discussed in an upcoming paper currently in preparation.

Measurement uncertainties as a function of altitude for both the Rayleigh and Raman temperatures are given in Figure 1.

2. AROTEL and Balloon Sonde Temperatures

During SOLVE, the DC-8 made numerous overflights of the Network for the Detection of Stratospheric Change (NDSC) site at Ny Ålesund, on Spitsbergen island, as part of a correlative measurements effort. Balloon sondes were launched to coincide with DC-8 overflights of the island and thereby provide ozone and temperature profiles from within the same volume sampled by AROTEL. Other sonde sites used for validation included two on Greenland's east coast (Danmarkshavn and Scoresbvsund), Iceland (Keflavik) and two belonging to Norway (Jan Mayen and Bjornoya Island). Several issues developed in using sondes as part of a correlative measurement effort. The first arose because sonde observations were not always coincident the DC-8's flight track, either in time or space. For the comparisons reported here the mean time between sonde launch and the point of the DC-8's closest approach was 3.8 hours over all three deployments. It was 2.6 hours for the first deployment, 2.1 hours for the second deployment and 6.5 hours for the third deployment. The mean distance between the balloon launch site and closest approach of the DC-8 was ~35 km. Lidar temperatures within ±150 meters of a reported sonde altitude were averaged together for the comparison. The second issue concerned the relatively low bursting altitude of the meteorological sondes, 20 to 25 km, which prevented the acquisition of a good data set within the 25-35 km region. Figure 2 presents sonde versus lidar temperatures measured over Ny Ålesund for flight date December 7, 1999.

Mean differences between AROTEL and all balloon sonde temperature data used in the comparison as a function of altitude are given in terms of elastic and inelastically derived temperatures in Figures 3 and 4. Reported uncertainties in sonde temperatures by *Schmidlin*, [1988] at 100 mbar (~15 km) are 0.5 K, ~1 K at 20 mbar (~26 km) and ~4 K at 10 mbar (~30 km) [see also *Keckhut et al.*, 1993]. Sonde temperatures are reported over altitude intervals ranging from ~10 to ~1000 meters. For temperatures derived from elastically scattered returns, the AROTEL minus sonde difference was -1.1 ± 0.6 K between 12 and 28 km. The mean difference for inelastically determined temperature was -1.2 ± 1.1 K between 12 and 26 km altitude. No systematic differences were found between the lidar and sonde temperature measurements as a function of the sonde's distance from closest approach to the aircraft's flight track and for time differences of up to 11 hours (the maximum difference). As shown in figures 3 and 4, a strong bias of several degrees is noted at the base of reported AROTEL temperatures. This is believed due to an altitude registration problem associated with the channels used for data acquisition close to the aircraft.

3. AROTEL and MMS Temperature Comparison

Several opportunities for correlative measurements with the ER-2's MMS became available during SOLVE. MMS provides a number of data products including pressure, horizontal and vertical winds and in-situ temperature at the ER-2's flight altitude [*Scott et al.*, 1990]. Temperatures are derived from a resistive measurement on an open platinum wire calibrated against a NIST standard. Corrections to the static temperature were made for thermodynamic heating as a function of the aircraft's MACH number. On January 20, 2000 both the DC-8 and ER-2 followed essentially the same flight track from Kiruna towards the North Pole and back again. The criteria employed to define coincidence on this flight required the measurement sites differ by no more than 30 km horizontally and be separated by less than 10 minutes in time. This test provided numerous coincidence pairs between ~14.5 km and 18 km geometric altitude. Since PSCs were observed by the co-located Langley aerosol lidar during significant portions of the flight, inelastic temperatures were utilized because of their relative insensitivity to optically thin PSCs and aerosols. The data indicated that AROTEL had a cold bias of approximately 2.5 ± 0.9 K compared to reported MMS temperatures (Figure 5). AROTEL temperatures were calculated using a 5 minute data set and had reported errors ranging from ~0.4 to 1.0 K over this altitude range; the effective vertical resolution was ~0.5-0.6 km. MMS has a reported precision of 0.3K, an accuracy of 0.2% (~0.4K for this altitude) and one second resolution. On both January 27, 2000 and March 5, 2000 additional comparisons were made between MMS and AROTEL temperatures retrieved employing both elastic and inelastic returns. The criteria for coincidence used here was 30 km horizontally and 5 hours in time. Inelastic data for January 27th had a mean difference between AROTEL and MMS of -2.8 ± 0.5 K while the March 5th flight demonstrated a mean difference of -1.9 ± 0.2 K. Temperatures calculated using elastic returns also displayed a cold bias with respect to MMS of -2.0±1.8 K for January 27th, for March 5th a bias of -2.6±0.8 K was observed. All measurements demonstrated that both the inelastic and elastically derived AROTEL temperatures were from 2 to 3 K colder than reported MMS values. The cold bias observed on January 20th may reflect the impact of uncorrected signal attenuation due to the PSCs observed along the flight track on this date; the 5 hour offset between when the

ER-2 and DC-8 made measurements on for the January 27th and March 5th data is large and could account for some of the difference. Additional comparisons between MMS and AROTEL under more ideal conditions could help resolve these differences.

4. Ny Ålesund and Mauna Loa Temperatures: Lidar and Sondes

As a result of the observed cold temperature biases in AROTEL temperatures when compared against both MMS and the balloon sondes a study was undertaken of temperatures retrieved by Goddard's STROZ-LITE Ozone and Temperature lidar during two previous field campaigns. This is a mobile system installed inside a trailer with a 39" telescope, a xenon chloride excimer laser operating at 308 nm and a xenon fluoride excimer at 351 nm. Temperature data is derived from elastically scattered data at 351nm and inelastically scattered returns at 382 nm [*McGee et al.*, 1995]. The close physical similarity between the STROZ-LITE and AROTEL lidars and the utilization of a similar temperature algorithm by both instruments suggested a study utilizing STROZ-LITE's more extensive data base could help identity the origin of these biases. The data sets chosen included a winter campaign at the Arctic NDSC site on Spitsbergen Island, Ny Ålesund, during 1998 and a summer campaign at a site on Mauna Loa in Hawaii during 1995. Specific questions were whether biases had been observed in the data during these two campaigns and, if so, what were their magnitude and sign. An additional opportunity for an independent correlative study was provided by Stuart McDermid's JPL lidar also located on Mauna Loa during the same 1995 campaign [*McDermid et al.,* 1991].

The Ny Ålesund study consisted of ten dates in January and February of 1998 during which correlative measurements were made. A total of 1373 separate temperature differences were calculated. Lidar retrievals began at ~10 km with a reported altitude resolution of from 1 to 3 km depending on altitude. Reported uncertainties ranged from 0.1 to 0.5 K for data sets acquired with ~60 minutes of data. Data retrieved in the presence of PSCs was not used in the comparison. The comparison showed that the lidar temperatures had a small cold bias relative to the sondes of ~-1.4±1.9 K over altitudes between 15 to ~32 km (Figure 6).

A second correlative measurements opportunity involved the NDSC intercomparison campaign on the island of Hawaii during August and September, 1995. Numerous temperature retrievals were made by STROZ-LITE and compared to balloon sondes, another ground-based lidar (McDermid) and National Meteorological Center (NMC) values. Balloon sondes were launched from Hilo, Hawaii and timed to coincide with data acquisition by the lidars. As with the Ny Ålesund data, the Mauna Loa intercomparison showed that the STROZ-LITE lidar exhibited a slight cold bias relative to the sondes (Figure 7). A total of 16 dates were employed and the resulting biases ranged from ~0 K at 15 km to ~-4 K at 30 km. The reported altitude resolution varied with altitude from 1 to 3 km; measurement uncertainties ranged from ~0.1 K at 15 km to ~0.5 K at 30 km for approximately 90 minutes of data. Another temperature and ozone lidar on Mauna Loa, operated by Stuart McDermid of JPL, also provided temperature profiles available for comparison with balloon sonde profiles during 1995. An identical analysis was performed for the same time frame as above using data reported by McDermid and again a slight cold bias was observed in the lidar temperatures relative to the sondes (see figure 8). This bias ranged from ~2.1 K at 15 km to ~0 at 20 km and ~-2 K at 35 km. The reported altitude resolution varied between 1 and 6 km. Reported measurement uncertainties as a function of altitude were between 0.2 K and 1 K .

These results suggest that temperatures derived from lidar data have a small but real cold bias when compared to temperatures retrieved by balloon sondes regardless of whether the data was acquired in the tropics or Arctic. The differences observed from these ground-based measurements are consistent with those observed by AROTEL during the SOLVE campaign (-1.2±1.1 K for inelastically derived temperatures, -1.1±0.6 K for temperatures derived from elastically scatter data). All three ground-based lidar temperatures comparisons exhibited a strong bias at the lower end of the altitude range. In addition the two comparisons made in Hawaii displayed a definite slope in temperature versus altitude. *Leblanc's* [1998] analysis found that the observed bias could be due to the technique used to smooth the data and/or a slight altitude offset in the instrument's true altitude.

5. Model Temperatures

A correlative study was carried out using AROTEL temperatures and those calculated by three global grided meteorological analyses: the NASA Goddard Space Flight Center's Data Assimilation Office (DAO) product, the United Kingdom Meteorological Office (UKMO) product generated for the Upper Atmosphere Research Satellite (UARS) project, and data reanalysis generated by National Centers for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR). All three are generated by data assimilation procedures in which satellite, rawindsonde, and other measurements are merged into a running model so that the resulting data are consistent with both the measurements and atmospheric physics.

The DAO fields for the SOLVE period are obtained from the GEOS-3 assimilation system for EOS-Terra support. GEOS-3 is the successor to the GEOS-1 system documented in *Pfaendtner et al.* [1995]. These data grids lie on 36 pressure levels ranging from 1000 hPa to 0.2 hPa, have a horizontal resolution of one degree longitude by one degree latitude, and are produced at 0000, 0600, 1200, and 1800 UTC each day.

The UKMO UARS fields are supplemental correlative data for the UARS Project and are described in *Swinbank and O'Neil* [1994]. The version used here are grids which consist of 18 pressure surfaces from 1000 hPa to 0.4 hPa, have a horizontal resolution of 3.75 degrees longitude by 2.5 degrees latitude, and are produced at 1200 UTC daily.

The NCEP/NCAR reanalysis system's procedures are applied consistently to over 40 years of raw data, resulting in a dataset which is useful for long-term studies. (For further information, see *Kalney et al.*,1996). These data consist of 17 pressure surfaces from 1000 hPa to 10 hPa, have a horizontal resolution of 2.5 degrees longitude by 2.5 degrees latitude, and are produced at 0000, 0600, 1200, and 1800 UTC daily.

Differences between lidar and model were calculated for temperatures retrieved using elastic and inelastic returns on all flight dates for which data was reported. Model data was retrieved as a function of geopotential altitude and converted to geometric altitude for the comparison. Temperature retrievals employing elastic backscatter within PSCs were excluded as were retrievals using inelastic data within or below clouds with a lidar ration >1.4. The flights on December 7, 1999 and December 12th were representative of dates during which PSCs were and were not observed by the co-located Langley lidar. Differences were calculated for all three models and are qualitatively similar however we present only the UKMO results here. Between flight altitude (10-12 km) and ~25 km the mean difference between AROTEL inelastic temperatures and UKMO temperatures for December 7, 1999 was 0.7± 1.9 K, for December 12, 1999 AROTEL was slightly colder: -1.0 ± 1.3 K (Figures 9 and 10). Elastically derived temperatures to ~40 km for December 7 were marginally colder than the models, -2.4±4.4 K; for December 12, AROTEL was again slightly colder, -2.8±4.2 K. Elastically derived temperatures between 32 and 33 km altitude repeatedly showed large cold biases with respect to the models (Figures 11 and 12). For December 7th the mean difference between AROTEL and UKMO at 32.8 km was -12.6±1.4 K. The December 12th differences were -12.3±2.2 K at 32.6 km. When these altitudes are removed from the calculation the mean difference between AROTEL and UKMO for December 7th was -0.6±1.3 K, for December 12th it was -1.2±1.6 K which are consistent with differences derived for the inelastically derived temperatures.

The large differences between AROTEL and model temperatures in the region of from ~30 and 35 km were surprising given the much small differences observed between AROTEL and both MMS and balloon sonde temperatures. As was previously done for the sonde measurements, data sets acquired by the STROZ-LITE lidar during deployments to both polar and tropical sites were used for comparisons with model forecast made from ~10 km to over 50 km altitude. To provide a consistent model temperature product for previously acquired data only results from the UKMO model were used for these comparisons. Data from the Arctic site at Ny Ålesund provided temperatures on 17 dates between January 20, 1998 and February 9 1998. Differences between STROZ-LITE and UKMO were calculated between 10 and ~40 km altitude (Figure 13). The mean difference between STROZ-LITE and UKMO was -1.1±8.8 K for altitudes from 10.8 to 38.7 km. There was a slight cold bias of ~1 K from 10 to 22 km which increased to over 9 K at 32.6 km. Reported temperature errors for these data sets range from 0.1 at 10 km to 0.3 K at 38 km with an effective vertical resolution of between 1 and 3 km.

The STROZ-LITE data set acquired at Mauna Loa in 1995 and used in the sonde comparison above was also compared to UKMO temperature profiles for the specified dates. Here, unlike Ny Ålesund, the UKMO temperatures displayed much better agreement with the lidar as shown in Figure 14. Data was acquired on 16 dates from August 15, 1995 through September 1, 1995. A slight cold bias was noted between 30 and 40 km, at 31.3 km the mean difference between lidar and UKMO was -2.0 K and at 36.1 km it was -2.7 K. Differences were calculated from 14.2 km to 55.2 km with a mean difference of -0.9 K. Calculated uncertainties ranged from ~0.1 K to ~0.6 K at 55 km. The lidar's reported vertical resolution ranged from 1km at the lower end to ~ 3 km at 30 km and, because most of the signal was directed at the high altitude channel, ~2 km at 55 km.

Temperatures derived from data acquired by the JPL lidar were also compared to those from the UKMO model. For the previously reported data set, JPL reported a slight cold bias of -0.5 K averaged over data from 14.3 to 55.4 km (Figure 15). The largest biases were noted at 36.0 km at -5.0 K. Typical errors for the JPL measurements ranged from 0.2 to 1.0 K with a reported vertical resolution of from ~1.0 to 7 km.

Temperature measurements using the stellar occultation technique by the

Midcourse Space Experiment (MSX)/Ultraviolet and Visible Imagers and Spectrographic Imagers (UVISI) suite of instruments were also somewhat colder than UKMO by several degrees within the 30-35 km region [*Swartz et al. this issue*]. Comparisons between lidar temperatures acquired at the French lidar site, Observatoire de Haute Provence (at 44 N, 6 E) from 1979 through 1993 and NCEP temperatures showed differences between the 10 and 0.4 hPa pressure levels. At the 10 hPa pressure level a small systematic cold bias of several degrees was observed when data acquired within 1-2 years of the El Chichon and Mount Pinatubo eruptions was excluded. These biases were attributed to misalignment of the lidar, uncorrected aerosol loading and possible problems with the radiosondes [*Keckhut et al.,* 2001].

These results suggest that between 30 and 35 km altitude, systematic biases exist between the lidars and models. Both the JPL lidar and STROZ-LITE had, with respect to the UKMO temperatures, a slight cold bias in the tropics while STROZ-LITE and AROTEL displayed much larger temperature differences against UKMO in the polar regions. The observed cold bias with respect to the UKMO model for arctic temperatures would be consistent with uncorrected signal extinction originating with a thin aerosol layer however the absence of significant recent volcanic activity makes it difficult to understand the origin of such a layer. Neither of the two DC-8 aerosol lidars, the Langley Aerosol and DIAL lidar systems were able to provide data from this altitude to help resolve this problem. As noted earlier sonde temperature uncertainties are given as 4 K at 10 mbar which, when coupled to the limited number of soundings available to the community within this altitude range could explain some of the differences between the lidars and sonde measurements acquired in the polar regions. Better year round coverage exist outside the polar regions.

6. Conclusions

AROTEL temperatures, retrieved during the SOLVE campaign, have been compared with those from balloon sondes and the MMS instrument onboard the ER-2. A slight cold bias was found in both comparisons; when compared to MMS the difference varied between 2-3 K while temperatures derived from balloon sondes were 1.2 ±1.1 K warmer than inelastically derived AROTEL temperatures and 1.1 ±0.6 K warmer than temperatures derived from elastically scattered data. The differences between AROTEL and MMS are believed to be due principally to an incomplete correction for extinction due to aerosols and PSCs which would account for the slight cold bias. The small but consistent cold bias observed against sondes has been seen for both AROTEL and STROZ-LITE temperatures in the Arctic. The same algorithm is used by AROTEL and STROZ and although it has been tested extensively under numerous conditions a detailed investigation of the algorithm under Arctic type conditions is now underway to try and understand these differences. AROTEL temperatures, when compared to several temperature models (DAO, NCEP and UKMO), showed differences that were less than or equal to 1 K below ~30 km. Between 30 - 40 km AROTEL displayed a large cold bias of 12 K against the UKMO model. Significant temperature biases at these altitudes were also observed in data acquired by another Goddard lidar, STROZ-LITE, during a winter correlative measurements campaign at Ny Ålesund in 1998. These cold biases were significantly smaller in data acquired in Hawaii during a 1995 campaign. The good agreement between AROTEL and model below 30 km altitude suggest the discrepancy above 30 km is due to the limited number of retrievals available for use by the models in the polar region.

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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. Balloon sonde versus AROTEL inelastic temperatures acquired on flight date 991207 over Ny Ålesund. The sonde's maximum altitude was ~30 km. The minimum temperature was ~190 K at 25 km in good agreement with AROTEL at 189 K.

Figure 3. AROTEL (elastic) temperatures minus all sonde temperature retrievals that occurred within 100 km and 12 hours of a DC-8 flight track. The mean difference between AROTEL and the sondes was -1.1± 0.6 K between 12 and 28 km altitude.

Figure 4. AROTEL (inelastic) temperatures minus all sonde temperature retrievals acquired within 100 km and 12 hours of a DC-8 flight track. The mean difference between lidar and sonde was -1.2 ± 1.1 K between 12 and 26 km altitude.

Figure 5. AROTEL (inelastic) - MMS temperature differences for flight date 000120. Coincidences were for MMS retrievals made within 30 km and 10 minutes of the DC-8 flight track. The mean difference was -2.5 ± 0.9 K.

Figure 6. STROZ-LITE minus balloon sonde temperatures for Ny Ålesund deployment in 1998. The mean difference was -1.4 ± 1.9 K between 15 and 30 km.

Figure 7. STROZ-LITE minus balloon sonde temperatures for Mauna Loa campaign in 1995. The mean temperature difference was -1.5 ± 2.0 K. At 15 km the difference was +0.5 ± 1.3 K; at 20 km, -1.5 ± 1.0 K; at 25 km, -1.9 ± 0.7 K; at 30 km, -4.0 ± 1.4 K and at 35 km the difference was -0.2 ± 1.3 K.

Figure 8. JPL - sonde temperatures for Mauna Loa campaign in 1995. The computed mean difference was 0.0 ± 1.7 K.

Figure 9. Temperature difference plot of AROTEL minus UKMO for December 7, 1999. AROTEL temperatures were derived using inelastically scattered data. The mean difference was -0.7±1.9 K.

Figure 10. Temperature difference plot of AROTEL minus UKMO for December 12, 1999. AROTEL temperatures were derived using inelastically scattered data. The mean difference was -1.0±1.3 K.

Figure 11. Temperature difference plot of AROTEL minus UKMO for December 7, 1999. AROTEL temperatures were derived using elastically scattered data. The mean difference was -2.4±4.4 K. The mean difference at 32.8 km was -12.6±1.4 K. Removing the value at 32.8 km results in a mean difference between AROTEL and UKMO of -0.6±1.3 K.

Figure 12. Temperature difference plot of AROTEL minus UKMO for December 12, 1999. AROTEL temperatures were derived using elastically scattered data. The mean difference was -2.8±4.2 K. The mean difference at 32.6 km was -12.3±2.2 K. Calculating the temperature difference without including the large perturbation at 32. 6 km gives a value of -1.2±1.6 K.

Figure 13. STROZ-LITE minus UKMO model temperatures at Ny Ålesund during the 1998 NDSC campaign. Differences between STROZ-LITE and UMKO were calculated between 10 and ~40 km altitude. The mean difference between STROZ-LITE and UKMO was -2.7±4.2 K for altitudes from 10 to 40 km. There was a slight cold bias of ~1 K from 10 to 22 km which increased to over 9 K at 32.6 km. Reported temperature errors for these data sets range from 0.1 at 10 km to 0.3 K at 38 km with an effective vertical resolution of between 1 and 3 km.

Figure 14. STROZ-LITE minus UKMO model temperatures during the Mauna Loa NDSC correlative measurements campaign in 1995. Data was acquired on 16 dates from August 15th through September 1st, 1995. A slight cold bias was noted between 30 and 40 km, at 31.3 km the mean difference between lidar and UKMO was -2.0 ± 2.5 K and at 36.1 km it was -2.8 ± 2.2 K. Differences were calculated from 14.2 km to 55.2 km with a mean difference of -0.9 ± 2.7 K. Calculated uncertainties ranged from ~0.1 K to ~0.6 K at 55 km. The lidar’s resolution ranged from 1km at the lower end to ~ 3 km at 30 km and back to 2 km at 55 km.

Figure 15. JPL - UKMO model temperatures during the Mauna Loa NDSC campaign in 1995. The JPL data had a slight cold bias of -0.5 ± 3.0 K averaged between 14.3 and 55.4 km altitude. The largest bias was noted at 36.0 km at -5.0 ±

1.8 K. Typical errors for the JPL measurements ranged from 0.2 to 1.0 K with a reported vertical resolution of from ~1.0 to 7 km.