Multiwavelength Aerosol Scatterometer for Airborne Experiments to Study the Optical Properties of Stratospheric Aerosol

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1. Introduction

In the recent past the role of polar stratospheric clouds (PSCs) and of stratospheric aerosol in polar ozone depletion raised the attention of the scientific community (Solomon 1990; Fiocco et al. 1997). The understanding of PSCs in terms of concentrations of particles, sizes, optical parameters, formation processes, and microphysics is relevant for evaluating their contribution to chlorine activation, dehydration, and denitrification of the lower polar stratosphere.

In view of a better understanding of PSC properties, during the last few years joint lidar and optical particle counters (OPC) measurements were carried out in Antarctica (Deshler et al. 1991; Adriani et al. 1992). While the OPC was able to assess particle concentrations and sizing, the lidar could give a contemporary measurement of the aerosol volume cross section for backscattered signals and information on the depolarizing properties of the particles, hence their thermodynamical phase. In the approximation of the Mie theory, the two measurement techniques were compared to estimate the refractive index of the particles (Adriani et al. 1995). However, the comparison between lidar and balloon-borne measurements needs some care, due to differences in the sampling technique. The lidar is a ground-based instrument and the OPC is carried by balloons and moves with respect to the ground, so that the two measurements are not taken on the same air mass. Ancillary information about the wind speed profile have to be used to choose, among the lidar temporal sequence, the best lidar profile to be compared—at a given altitude—with data from the balloon-borne OPC, drifting with the wind. This technique enhances the reliability of the data comparison, although it still may be reduced when the clouds under investigation are connected to local phenomena like formation through orographic lee waves. In such cases, the assumption that the air mass does not change its properties over long temporal or spatial ranges may not be true (Adriani et al. 1992).

The development of a backscattering instrument that can fly in conjunction with particle counters, and therefore perform in situ measurements, affords a direct comparison of the two sets of data. A backscatter sonde was developed a few years ago by Rosen and Kjome (1991) and has been used quite extensively (Rosen et al. 1994; Larsen et al. 1995; Larsen et al. 1996; Rosen and Kjome 1997). This backscatter sonde is based on the use of a flash lamp as a light source. The Multiwavelength Aerosol Scatterometer (MAS) is conceptually similar to the backscatter sonde but uses laser diodes as a light source, which are very reliable, small-sized, and have low power consumption. MAS was designed for operating on the stratospheric research aircraft M-55 Geophysica (Borrmann et al. 1995), and it was used for the first time in the Arctic during the first Airborne Polar Experiment campaign (Rovaniemi, Finland, 1996–97). Its weight is about 25 kg and all electronics and optics are contained within a metal layered carbonium box, 654 × 360 × 350 mm size. The instrument was placed inside the aircraft basin, in the lower part of the fuselage, facing the outside air through a shuttered glass entrance window.

The primary objective of MAS was to perform scattering and depolarization measurements on aerosols and clouds that were also measured with a forward-scattering spectrometer probe (FSSP model 300) particle counter (Baumgardner et al. 1992). The FSSP measures particle size distributions for particles with size diameters between 0.4 and 23 μm in 31 size bins. Thus contemporary backscatter, depolarization, and size distribution...
measurements of particles could be obtained, and used to infer aerosol shape and composition (i.e., refractive index).

A simpler balloon-borne version of MAS has been flown several times from polar as well as midlatitude and tropical regions in the framework of the project Stratospheric Regular Sounding (Adriani et al. 1998). Those flights were performed along with the University of Wyoming optical particle counter (Hofmann and Rosen 1982; Hofmann and Deshler 1991), which performs observations of integral number density of particles with radii larger than 0.15, 0.25, 0.5, 1, 2, 3, and 5 μm. Some results from these joint balloon flights will be described to show how the measurements can be used in the study of the atmospheric particulate matter.

2. The instrument

a. Optics

The optical layout of the instrument is given in Fig. 1. All optics are mounted on a carbonium optical bench, to ensure low weight and good thermal stability. The scatterometer uses three cw laser diodes at 680, 780, and 830 nm (LaserMax MD type lasers). Emitted powers are 5, 10, and 30 mW, respectively, while laser light is polarized by a factor of 100:1 and has a bandwidth of 0.1 nm. Laser light is simultaneously transmitted into the atmosphere through a glass window via two plane mirrors and an on/off modulation of the laser emission provides sky background measurements. The laser beams are aligned to come out parallel and almost collinear with the telescope optical axis. The beam divergence is 0.25 mrad; hence all the beam path is within the telescope field of view. Atmospheric particles (aerosol and/or clouds) and molecules backscatter part of the laser light back to the instrument and, when the particles do not have spherical shapes, part of that light is depolarized. Optical calculations following Measures (1984) show that the beams enter the telescope field of view only 25 m away from the telescope’s aperture, since the second plane mirror partially shadows the telescope aperture, and are completely inside the field of view only after another 25 m. This configuration, while reducing the total amount of light collected by the telescope, ensures the collection of light effectively scattered back at 180° ± 0.1°. The 90% of the backscattered signal comes from the 75 m closest to the instrument. This has been verified by observing the instrument response to an LED light placed at various distances from the telescope.

The backscattered light is collected by an F2 telescope. The objective consists of a low aberration negative doublet with a 120-mm aperture that corrects the aberrations induced by the primary spherical mirror. The primary mirror focuses the light through reflection on the secondary plane mirror onto a pinhole acting as a field stop for the optical system. This field stop makes
aberration triplets. Interferential filters with 7-nm FWHM centered on the laser wavelengths give further spectral reduction of sky background.

b. Electronics

MAS works under the control of two PC104 standard CPU 80286 computers. One of these computers is entirely devoted to the acquisition and storage on a solid-state disk of the avionic data. These data come from the airplane unit for connection with scientific equipment, via serial communication, and give information on the aircraft speed, orientation, global positioning system (GPS) position, and UTC time, as well as data on atmospheric pressure and temperature, important for successive MAS data reduction. Resynchronization between the two computer’s clocks and the UTC time from GPS is performed every 10 min. The second computer is devoted to instrument housekeeping and data acquisition.

Instrument housekeeping consists essentially of controlling the temperatures of the lasers, the light detectors, the interferential filters, the telescope and the electronics by means of 10K3A BetaTHERM temperature sensors, resistor heaters, and monitoring laser power output. Temperature stabilization is specially important for the lasers, which show a temperature dependent wavelength drift of 0.25 nm/°C, for the light detectors, which should be operating between 10° and 35° C, for the optics.

The instrument is not sealed, and environmental conditions may range from −20° to 20° C at 1000 hPa to −80°C at 50 hPa (at about 20-km height). In such a wide range of conditions it was necessary to sufficiently heat all optics to prevent surface fogging and to cool light detectors and part of the electronics since in stratospheric conditions heat dissipation through the air is reduced. This was accomplished by cold fingers to adequate heat sinks. A plot of the thermal behavior of various parts of the instrument during real flight conditions may be seen in Fig. 2. After an initial warming, all points tend to maintain a fixed temperature or to cool. The cooling results are more pronounced during the last part of the flight when the descent forces cold air to enter the instrument. The more sensitive elements with respect to temperature—that is, lasers, interferential filters, and light detectors—remain well inside their range of operability during the whole flight. Laser output was controlled by TLS220 light-to-frequency converters from Texas Instruments, to give information about variation on the light power source.

The amount of light collected from the atmosphere is at photon counting rates. The light detectors are avalanche photodiodes able to work in photon counting mode. The collection of the counts and temperatures inside the instrument are carried out by means of an OEM board, working under computer control and a DM406 A/D–D/A converter from Real Time Devices.

Photons counts of sky background or backscattered signal and laser power output are collected at 0.5 Hz. Inside temperatures are measured every 10 s. The sky background is subtracted from the backscattered signal and reported every 2 s. At an average airplane speed of 200 m s⁻¹, this gives a spatial resolution of 400 m. The instrument can only be operated at night.

MAS has a maximum power consumption of 80 W.

3. The measurements

During December 1996 and January 1997, MAS joined the Airborne Polar Experiment campaign that took place in Rovaniemi, Finland. An electronics failure that could not be fixed in the time frame of the Rovaniemi campaign limited MAS deployment to three flights. No useful scientific data were collected during that campaign, and no measurements by MAS will be presented here. To illustrate how the measurement techniques can be applied to infer aerosol properties, some results from balloon flights with a simpler MAS and an optical particle counter will be described hereafter.

On 14 January 1997 a balloon carrying a condensation nuclei counter, a radiosonde, an ozonesonde, and MAS was launched from Andoya, Norway (69°N, 16°E). That version of MAS was equipped with one 680-nm laser and two receiving channels, for polarized and depolarized backscatter. Figure 3 shows the polarized (left panel) and depolarized (right panel) signal acquired during the flight. The presence of a tropospheric cloud, from 5–10 km, and a polar stratospheric
cloud around 25 km may be clearly seen. The thick black line is the signal expected from a purely molecular atmosphere. Note that since the two receiving channels have different efficiencies, there is not a proportionality of 1.4% between the reconstructions of the molecular polarized and depolarized signal.

The first parameter retrievable from MAS raw data is the backscattering ratio (BR). Neglecting the background contribution, the number of photons received in a fixed amount of time $N$ is proportional to the volume backscattering coefficients of aerosols $\beta_a$ and molecules $\beta_m$, to the number of photons emitted by the laser $E_0$, to the detectors efficiency $Q$, and to various instrumental parameters, which we indicate as $C$. That is, $N = CE_0Q(\beta_a + \beta_m)$. The backscattering ratio, $BR = (\beta_a + \beta_m)/\beta_m$, is a measure of the contribution to the backscattering from particles with respect to the one expected from molecules. The signal from a purely molecular atmosphere is proportional to the molecular density. It can be calculated using the perfect gas law by means of in situ measurements of temperature $T$ and pressure $p$. The number of photons $E_m$ expected from a pure molecular atmosphere is then estimated as $E_m = K(p/T)$. The proportionality factor $K$ is obtained by setting $E_m$ equal to the measured signal $E(z_c)$, at a proper calibration altitude $z_c$, where the atmosphere is free from aerosol, $K = E(z_c)[T(z_c)p(z_c)]$. The region for this matching can be easily identified when MAS is flying with a particle counter. This procedure, when applied to different altitudes, allows also the cancellation of a possible constant bias in the signal, caused by stray light due to unwanted scattering from parts of the instrument. The ratio BR can be calculated dividing the measured signal by the one expected from molecular scattering, $BR = E/E_m$. This is depicted in Fig. 4 (left panel). In aerosol free regions, BR is equal to 1, while enhanced backscattering from cloud particles or aerosols results in a BR greater than one.

Laboratory tests in low pressure, low temperature conditions and housekeeping data recorded during real flights showed that both laser emission, in terms of emitted power, wavelength, and detectors response, remain stable enough to neglect them as sources of error in the BR retrieval. The uncertainty attributed to BR is mainly associated with the fluctuations of the photon statistics in the signal. Other uncertainties come from errors in pressure and temperature measurements. These play a twofold role: their profiles are used to reconstruct the molecular density, and their values at a given calibration level are used in the assessment of the calibration factor.
OCTOBER 1999 1333
NOTES AND CORRESPONDENCE

Fig. 5. Processed data from the 23 Jan 1996 flight from Andoya, Norway. In the left panel the backscattering ratios at 680- and 830-nm wavelength and the depolarization ratio are depicted. A thick cloud is present between 18 and 26 km, mostly liquid with a solid core at 24 km. The right panel shows the CR profile along the cloud.

$K$, thus uncertainties in determining $p$ and $T$ affect, for instance, $E_m$ both directly and indirectly through $K$, since $K$ is also a function of $p$ and $T$.

Errors in $K$ determination are due primarily to uncertainties in pressure and temperature at the calibration level more than to inherent statistics on a given set of calibration altitudes. This uncertainty in $K$ increases as the calibration altitude $z_c$ increases due to the decrease in signal strength as air density decreases. Assuming a calibration altitude of 23 km, uncertainties of 0.2 K for temperature, 0.5 hPa for pressure, and adding all contributions from different error sources—as fluctuations in photon statistics—by ordinary error theory, an error lower than 8% can be assessed on BR values, up to the top of the sounding. This can be reduced if a lower altitude for calibration is chosen. Uncertainty increases with increasing altitude, since the percentage of error in pressure determination increases as pressure decreases.

Systematic errors may be induced by an incorrect choice of the calibration altitude. These errors are likely to be more pronounced when the BR at long wavelength is estimated, due to its greater sensitivity to aerosol loading. The choice of a calibration altitude that is not aerosol free leads to an overestimation of the molecular signal; then the calibration can be impossible on some occasions.

An important parameter that can be assessed from BR

is the aerosol volume backscattering coefficient $\beta_a$, defined as

$$\beta_a = \beta_a(BR - 1) = \beta_a \left( \frac{E}{E_m} - 1 \right)$$

where $N$ is the number of molecules in a unit volume and $d(\sigma_m)/d\Omega$ is the Rayleigh backscattering cross section (Collins and Russell 1976). Figure 4 shows the layered structure of the cirrus and the PSC, in term of $\beta_a$. Due to the difference in magnitude between the cirrus and PSC coefficients, two distinct horizontal scales have been used. Uncertainties in $\beta_a$ are due to uncertainties in the determination of BR and, again, in pressure and temperature needed to estimate $N$, using the gas law.

The depolarization ratio (DR), depicted in the left panel of Fig. 4, is the third measured parameter. It can be defined in different ways, but the most common is the ratio between the depolarized and polarized signals. This parameter gives mostly qualitative information about the asphericity of the scatterers. Spherical particles backscatter without changing the incident light polarization, while aspherical particles (such as aerosols in solid phase) depolarize a fraction of the backscattered light. High values of depolarization, as in the region between 5 and 10 km, thus suggest the presence of ice particles. The PSC in the region around 25 km, however, shows a low depolarization, which suggests the presence of spherical homogeneous particles.

The molecular atmosphere is expected to give a small depolarization around 1.4%. This value has been confirmed both experimentally (Cohen and Low 1969) and theoretically (Young 1980). In our measurements no absolute calibration was performed. The depolarization ratio was normalized to the molecular value assuming that the ratio between the two channels was 1.4% in aerosol free regions. Profiles of the uncertainties to be attributed to BR, $\beta_a$, and DR are depicted in the right panel of Fig. 4. Again those arise mainly from uncertainties in the measure of temperature and pressure.

Another polar stratospheric cloud observation occurred during a balloon flight that took place from Andoya, Norway, on 23 January 1996. In that flight MAS, an OPC, and a condensation nuclei counter were present. MAS was equipped with two lasers at 680 and 830 nm wavelength and three receiving channels, two for polarized backscattering and one for depolarization at 680 nm. The use of two distinct wavelengths afford the calculation of the aerosol color ratio (CR) defined as

$$CR = \frac{\beta_a \lambda=830\text{nm}}{\beta_a \lambda=680\text{nm}}.$$
composition (viz., refractive index) and, in general, increases with the particle size (Larsen et al. 1994).

In Fig. 5 the BRs at the two wavelengths DR and CR are depicted, in the range of altitudes where the PSC was present. The three-layer structure of the cloud can be clearly seen. The cloud appears to be mostly liquid with a solid core at 24 km, as revealed by DR values. A monomodal lognormal function of the form

\[ \frac{dn(r)}{dr} = \frac{N_0}{\sqrt{2\pi} \ln\sigma} \exp\left(-\frac{\ln^2(r/r_g)}{2 \ln^2\sigma}\right) \]

can be used to approximate size distributions for stratospheric aerosols in many conditions (Pinnick et al. 1976). There, \(dn(r)/dr\) is the number of particle in a unit volume and radii between \(r\) and \(r + dr\), \(N_0\) is the total number concentration, \(\sigma\) is the width of the lognormal curve, and \(r_g\) is the median radius. Figure 6 gives the results of modeling the color ratio, which does not depend on \(N_0\), varying median radius and standard deviation. The calculation has been performed using Mie theory (Bohren and Huffman 1983). Also, \(r_g\) and \(\sigma\) were allowed to vary respectively between 0.05 and 1 with steps of 0.005 \(\mu m\) and between 1.005 and 2 with steps of 0.005. The \(\beta_n\) coefficients were calculated by numerical integration over the size distribution from 0.01 to 10 \(\mu m\).

Inspection of Fig. 6, given the measured values of CR, puts some constraint on the acceptable values of \(\sigma\) and \(r_g\).

The use of two or more CRs would permit the assessment of \(r_g\) and \(\sigma\) for a normalized lognormal, and direct comparison with measured BRs would provide an estimation for \(N_0\). This has been done in the past (D’Altorio et al. 1993; Del Guasta et al. 1994; Larsen et al. 1994). However, in post volcanic conditions (Deshler et al. 1993) and in PSCs (Peter 1997) the bi-modal distributions are often a better approximation to the real size distribution. In such conditions, we may regard the use of a monomodal as producing only an “optically equivalent” size distribution and there is poor confidence in retrievable parameters as total surface area density or total volume density.

MAS reveals its best capabilities when used in conjunction with a particle counter, for the two instruments complement each other. This is particularly true in the assessment of the particle mean refractive index, since the reduction of OPC data in terms of particle size distribution requires an assumption on the value of this
parameter (Hofmann and Deshler 1991). Moreover, data reduction may be founded on the basis of Mie theory when no aspherical scatterers are present, as revealed by inspection of the depolarization data, otherwise the use of more sophisticated mathematical tools would be required.

An evaluation of the particle mean refractive index can be done by comparison of OPC and MAS data. A set of refractive indices, hence a set of different OPC calibrations, can be used to invert in term of particle size distributions the OPC raw data. The retrieved distributions can then be used to calculate the aerosol volume backscattering coefficients at the MAS wavelengths. Comparison of the set of calculated backscattering coefficients with the observed ones will provide the best estimation of the refractive index. Finally, the resulting refractive index is the one that ensures the best match between the calculated coefficient from the OPC data, and the one observed by MAS.

4. Conclusions

The Multiwavelength Aerosol Scatterometer (MAS), an instrument for studying particles in the atmosphere, has been conceived for use in conjunction with optical particle counters to improve the quality of the measurements on clouds and aerosol in the polar stratospheres. It can measure scattering ratios at 680, 780, and 830 nm as well as estimate the depolarization of the laser light at the same wavelengths consequent to the backscattering from solid particles. Intercomparison with optical particle counter results can give more information about the composition of the particles through an estimation of the refractive index. This can be achieved comparing the MAS optical measurements with the size distribution and concentration measurements taken by the particle counter. A detailed analysis of an intercomparison between the laser backscatter sonde and the optical particle counter has not been described here, being the subject of different works. The goal of the present paper is the presentation of the instrument itself. The sonde can be used on airborne platforms, and a simpler version of it can be used on balloons. Experiments have been performed by balloons since 1996.

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