Remote Sensing of Aerosol Optical Depth in a global surface network

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Remote Sensing of Aerosol Optical Depth in a Global Surface Network

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presented by

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Abstract

Atmospheric aerosols provide, through their direct and indirect effects, a predominantly negative or cooling contribution to the global radiation budget. Their integral optical activity is summarized in the aerosol optical depth (AOD) that can be derived from measurements of transmitted sunlight.

In the early nineties, the World Meteorological Organization (WMO) deemed the quality of their long-time AOD observations insufficient for climate studies. A World Optical depth Research and Calibration Center (WORCC) was established at Davos in 1996 and given the task by WMO to develop stable instrumentation and improved methods of calibration and observation of AOD. These new developments should be demonstrated in a global pilot network.

The present thesis gives an account of the research and development work performed during implementation of these tasks.

After introducing the role of aerosols in climate change and the difficulties in their characterization, a personal review of the history of turbidity or AOD measurements is given. Algorithms for AOD, methods of calibration, and the construction of a Precision Filter Radiometer (PFR) are discussed. The world-wide AOD network of PFR instruments is presented. First results of AOD measurements at network stations are given as time series and climatological values.
Zusammenfassung

Atmosphärische Schwebstoffe oder Aerosole üben durch ihre direkten und indirekten Wechselwirkungen mit dem Sonnenlicht einen signifikanten, meist abkühlenden, Einfluss auf die Strahlungsbilanz der Erde aus. Die integrale optische Wirkung der Aerosole kann in ihrer optischen Dicke (AOD) zusammengefasst und aus Messungen der transmittierten Sonnenstrahlung abgeleitet werden.


In einer Einführung wird der Einfluss der Aerosole auf den Strahlungshaushalt in Zusammenhang mit der Klimaänderung gestellt. Ein historischer Rückblick zeigt die Entwicklung der AOD oder Trübungsmessungen bis zur Errichtung des WORCC aus persönlicher Sicht.

Die Algorithmen zur Bestimmung der AOD, Methoden der Kalibrierung und die Konstruktion des Präzisions-Filter-Radiometers (PFR) werden detailliert besprochen, und auf mögliche Fehler hin untersucht.

Das weltweite Messnetz von PFR wird vorgestellt und erste Resultate in Form von Zeitreihen und klimatologischen Werten angegeben.
Chapter 1  Introduction

Aerosols are an important constituent part of the atmosphere not only with respect to climate change, but also to air quality including transboundary transport of air pollution, human health and mortality, which is affected by respirable particles, or aesthetic aspects like impaired visibility range. Human activities are modifying the concentration and composition of atmospheric aerosols through combustion of fossil fuels, changing land cover, biomass burning, as well as urban and industrial emissions.

Growing recognition of the role of atmospheric aerosols in the Earth's radiation budget and hydrological cycle has led to a steady increase of scientific interest in aerosol physical, chemical and optical properties over the last decades. Aerosols provide through their direct and indirect effects a negative radiative forcing, or cooling. Anthropogenic aerosols could thus compensate the positive radiative forcing by increasing concentrations of greenhouse gases to a significant, but largely unknown extent. Despite the large numbers of studies on aerosol climate interaction, the uncertainty range of aerosol indirect effects has actually increased from the second to the third assessment of the Intergovernmental Panel of Climate Change (IPCC), and the level of scientific understanding was still classified as ‘very low’.

Aerosols modify the radiative flux in the atmosphere directly by scattering and absorption of sunlight. A certain fraction of the solar irradiance is scattered in the forward direction and reaches the surface as part of the diffuse irradiance, leaving the hemispheric energy flux unchanged. Another fraction of the incoming solar radiation is scattered back into space. Some types of aerosols, notably mineral dust and carbonaceous aerosols are also absorbing short and long wave radiation. Such aerosols are thus changing the vertical temperature profile of the atmosphere and modifying the radiative balance by emission of thermal radiation. According to the fourth IPCC report (IPCC, 2007), the radiative forcing by the aerosol direct effect is estimated at -0.5 [-0.9 to -0.1] W m$^{-2}$ with a medium to low level of scientific understanding.

Small aerosols act as cloud condensation (CCN) and ice nuclei and modify as such the optical properties of water and ice clouds and possibly the hydrological cycle. These mechanisms are known under the name of aerosol indirect effects. A review of these indirect effects and estimate of their global impact is given by Lohmann and Feichter (2005). Without going into details, a brief qualitative description of the main physical mechanisms is repeated here.

The first indirect or Twomey effect refers to enhanced cloud albedo due to the more numerous but smaller cloud droplets in a cloud of constant water content when the number of CCN is increased (Twomey, 1959, 1974). Current estimates of the global radiative forcing at the top of the atmosphere by the first indirect effect are between -0.5 and -1.9 W m$^{-2}$.

The second indirect or cloud lifetime effect is a consequence of the first effect, in that the precipitation of clouds containing smaller droplets is delayed and the

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1 Numbers in brackets indicate 90% uncertainty range of the underlying non-normal distribution
effective cloud coverage enhanced (Albrecht, 1989). The radiative forcing of the second indirect effect is estimated at a similar range as the Twomey effect.

The semi-direct effect covers absorption of solar radiation by aerosols leading to an extra heating of the atmosphere, which can inhibit condensation or result in evaporation of existing cloud droplets. (Grassl, 1979) This mechanism can partially compensate the cooling due to the indirect effects as the radiative forcing is estimated to lie between -0.3 to +0.1 W m\(^{-2}\).

According to the fourth IPCC report (IPCC, 2007), the radiative forcing, the total radiative forcing of these cloud albedo or indirect effects is estimated at -0.7 [-1.8 to -0.3] W m\(^{-2}\). Their level of scientific understanding has improved from ‘very low’ in 2001 to ‘low’ in 2007.

These IPCC global estimates are derived from a significant number of climate models. Quaas and co-authors (2008) have estimated 5-year average aerosol radiative forcing from satellite-based AOD retrievals and Earth radiation budget measurements. They find direct forcing over land (-1.8 W m\(^{-2}\)) to be about twice the forcing over ocean, where on the other hand indirect forcing is twice as strong as over land (-0.2 W m\(^{-2}\)). These observational results indicate that global climate models may overestimate the aerosol indirect effect.

Aerosol indirect effects encompass particle – cloud interactions and are thus closely related to cloud physics and the hydrological cycle in global climate. Both aerosols and clouds act mainly on the shortwave (solar) side of the radiation budget and their combined radiative forcing may become manifest in surface solar radiation data. A secular trend of ‘Global dimming’ was indeed detected in global surface radiation data by Ohmura and Lang (1989) for the second half of the 20th century. Since the late eighties this dimming seems to have reversed to a world wide ‘Brightening’ (Wild et al., 2005). Both trends can potentially be explained with first rising then declining SO\(_2\) emissions (Streets et al., 2006), which are a major source of anthropogenic aerosols.

Compared to lifetimes of greenhouse gases (decades to centuries), aerosols have shorter (days to weeks) residence times in the atmosphere, and their concentration thus shows much larger temporal and spatial variability. Any assessment of the trend in total atmospheric aerosol load and global distribution thus requires global coverage of observations as well as long-term monitoring, which is beyond the capabilities of any single institution or agency, and calls for international collaboration and coordination. To worsen the observational problem, aerosols, unlike greenhouse gases, cannot be characterized by their concentration alone, but have physical, chemical and optical properties. Moreover, the atmospheric aerosol burden represents a heterogeneous mixture of particles spanning a vast range in size. Therefore, a wide collection of measurement techniques (WMO, 2003a) is required to assess the aerosol composition of the atmosphere. Tropospheric aerosols are accessible for detailed characterisation by ground-based and occasional air borne in-situ measurements, but their vertical column integral and distribution can only be assessed through remote sensing from ground stations or satellite platforms.

The optical activity of aerosols in the atmospheric column can be summarized by the aerosol optical depth, which is a wavelength dependent measure of the total extinction of sunlight due to scattering and absorption by aerosols. Additional information about the light scattering properties of the aerosols is required for the determination of the single scattering albedo (SSA) and asymmetry parameters which are essential in the calculation of aerosol radiative forcing. Still AOD is the single most comprehensive variable to assess the
aerosol load of the atmosphere and represents the least common denominator by which ground based remote sensing; satellite retrievals and global modeling of aerosol properties are compared. AOD is thus one of four core parameters of the aerosol component in the Global Atmosphere Watch (GAW) program of the World Meteorological Organisation (WMO, 2003a).

Ground based AOD observations have high temporal, typically hourly, resolution, but are restricted to a limited number of sites. Although they have been made since at least 50 years at various locations and for different periods, an observational proof of changing atmospheric aerosol burden, as for the CO$_2$ concentration, on a global scale had not been established by mid 90ies (Andreae, 1996). Even worse, a system of sufficient coverage and accuracy did not exist that would allow scientists thirty years from then to assess whether aerosol burdens have changed since the end of the 20th century. Such a system to accurately monitor the global distribution of aerosols requires the combination of continuous observations from satellites, networks of ground-based instruments, and dedicated field experiments (Kaufman et al., 2002). Since the turn of the century AOD has been retrieved not only over oceans but also over land from measurements of several dedicated aerosol instruments (e.g. POLDER, MODIS, MISR) on a fleet of Earth observing satellites. MODIS can provide global coverage at moderate temporal, typically monthly, resolution with estimated errors (Kaufmann et al., 2002) of $±0.03$ $±0.05$AOD over the oceans and $0.05±0.20$AOD over land. Actual uncertainty of satellite retrievals is probably larger, as Myhre and co-authors (Myhre et al., 2005) found AOD over ocean from 9 algorithms to differ by at least a factor of 3, and identified cloud screening as the most probable main reason for this large disagreement.

Satellite AOD observations with near global coverage are therefore supplemented by ground based networks, notably AERONET (Holben et al., 1998) or EARLINET (Matthias et al., 2004). These networks can provide more accurate AOD observations in the order of 0.01 optical depths for satellite ground truthing. They are also contributing to the emerging aerosol climatology (Holben et al., 2001; Michalsky et al, 2001) at sites with multi-year observations. A WMO workshop (WMO, 2005a) has recently identified about 90 global surface stations with 4 or more years of continuous observations. Roughly half of these stations are well organized in the AERONET project while the remaining ones are operated by various national meteorological services. The AOD measurements at GAW stations play an important role as a backbone network that will provide traceability between these diverse national observation programs. Many of the GAW global stations also are running observation programs that include in-situ aerosol measurements of mass, number concentration, light absorption and scattering coefficients as well as chemical composition. These data are available at the World Data Center for Aerosols (WDCA) in Ispra (http://wdca.jrc.it).
CHAPTER 1

Introduction
Chapter 2  History of turbidity measurements

Scientific interest in the transparency of the atmosphere started at least 250 years ago with the work of Pierre Bouguer (1698 – 1758), a French mathematician and hydrographer, who in 1729 published his photometric observations of the Sun and the Moonlight and proposed an exponential dependency of the transmission of stellar light through the Earth’s atmosphere. The Swiss physicist Horace Bénédict de Saussure (1740 – 1799) developed and used a so called diaphanometer to determine the transparency of air as well as a cyanometer to judge the blueness of the sky, which he associated with the cleanliness of the air.

With the development of caloric radiometers and the introduction of the compensation principle in the second half of the 19th century, it became possible to replace these visual photometric observations by more accurate quantitative measurements and, by adding spectral filters, to obtain more specific information about the turbidity or transparency of the atmosphere and its extinction components. Claude Pouillet (1791 – 1868) made the first absolute measurements of solar radiation (1838) with his pyrheliometer. Through application of Bouguer’s empirical law, he was able to determine the intensity of solar radiation above the atmosphere, the so called solar constant, at 1.79 ly min^-1 (1249 W m^-2) or within 10% of today’s accepted value of 1.959 ly min^-1 (1367 W m^-2). Knowledge of the total extraterrestrial irradiance and its spectral distribution is a fundamental prerequisite for the study of atmospheric extinction processes. Until the advent of space flight, the solar constant could only be derived indirectly from ground based measurements in combination with models of these extinction processes.

Many of these processes were recognized and well understood by the end of the 19th century. The Irish physicist John Tyndall (1820-1893) had examined the optical properties of gases, including absorption spectra of water vapor, ozone and carbon dioxide. He noticed the importance of these atmospheric components for terrestrial climate by saying that “without water vapor, the Earth’s surface would be held fast in the iron grip of frost”. Lord Rayleigh (John William Strutt, 1842 – 1919) developed his theory for scattering of light by small particles between 1871 and 1899 whereby he could explain the blue color of the sky as the scattering of sunlight by the molecules of air itself (Rayleigh, 1899). Gustav Mie (1869 – 1957) expanded the electromagnetic scattering theory (Mie, 1908) to particles larger than optical wavelengths and laid the foundation for aerosol optical depth retrieval from radiance measurements.

Since the beginning of the 20th century, routine measurements of the solar radiation were made at a number of astronomical and meteorological observatories in Europe (e.g. Potsdam, Davos, and Uppsala) and USA (e.g. Mt. Whitney) in order to establish climatology of solar radiation or to determine the solar constant. Through the work of Langley and Fowle at the Smithsonian Institution, the solar constant and its spectral distribution became known with sufficient accuracy to permit the determination of atmospheric turbidity from terrestrial measurements of solar radiation. Although the limitation of the Beer-Lambert law to quasi-monochromatic radiation was well known, it was still applied to derive ‘effective’ or ‘complex’ extinction coefficients from panchromatic measurements with pyrheliometers. These simple turbidity coefficients suffered from strong virtual variations depending on solar zenith distance and atmospheric humidity. In order to reduce these diurnal variations, not considered to be part of turbidity, Linke and
Boda (1922) proposed their ‘Trübungsfaktor’, which relates the total extinction to the spectrally integrated Rayleigh extinction. This turbidity factor can be interpreted as the number of pure and dry atmospheres required to match the extinction of the actual atmosphere at a given zenith angle. In a later (Linke, 1942) revision of his Trübungsfaktor, Linke related the total extinction to that of dust-free atmosphere with fixed water vapor content of 1 atm-cm. This ‘Neuer Trübungs faktor’ would be representative for dust turbidity, but was still hampered by virtual variations.

Somewhat more consistent results were obtained from spectrally filtered measurements that allowed the separation of the infrared region dominated by water vapor absorption from total radiation. The infrared part of the solar radiation above 0.625 μm was measured through a standard colour glass filter, and subtracted from total measurements to obtain shortwave (Kurzstrahlung) radiation. In analogy to the original Linke factor, a ‘Kurztrübungsfaktor’ was then derived that was unaffected by water vapor absorption, but could not eliminate the virtual diurnal variation due to the broadband averaging of the dust extinction.

A different approach to quantify atmospheric turbidity was taken by Anders Ångström (Ångström 1929, 1930) by assuming that the extinction of dust can be modelled similar to the molecular scattering by a power law in wavelength of the form $\beta \lambda^{-\alpha}$. By analyzing the extensive spectrometric data set of the Smithsonian Institution, he could confirm this assumption and found an average value of 1.3 for the exponent $\alpha$, which he first called size coefficient and associated it with a typical size of the dust particles of about 1 μm. He assumed the coefficient $\beta$ to give a close measure of the total quantity of dust present in the atmosphere and derived its relation to the column number density $N$ experimentally as $\beta = 1.7 \cdot 10^{-9} N$. From the relative stability of $\alpha$ he concluded that the 1 μm class of particles dominates the optical extinction of all other size classes and therefore renamed $\alpha$ as wavelength exponent. In fact, the wavelength exponent is closely related to the aerosol number size distribution $n(r)$ which Junge (1952) had found experimentally to be inversely proportional to the particle volume $r^3$. From Mie theory it follows (Junge, 1952) that a power law aerosol size distribution $n(r) = C/r^\nu$ will lead to a power law for the extinction with an exponent $\alpha = \nu - 2$. Assuming a constant wavelength exponent $\alpha$, the turbidity coefficient $\beta$ could be determined from broadband filter measurements through pre-calculated lookup tables or graphical interpolation, similar to today’s satellite retrieval methods.

Schüepp (1949) introduced another model for dust turbidity, closely resembling the Ångström power law, but assuming a wavelength dependency of the wavelength exponent and expressing his coefficient $B$ in decadic form normalized at a wavelength of 0.5 μm. For an $\alpha$ exponent of 1.3, Ångström’s and Schüepp’s turbidity coefficients are related by $B = 1.07 \cdot 10^3$. Schüepp used a set of Schott filters with sharp cut-off wavelengths and devised a number of graphical procedures (nomograms) to derive his coefficient $B$ and the wavelength exponent $\alpha$. It appears, however, that his laborious method was very sensitive to observational errors and required utmost care in calibration of the radiometers and in cleaning the filters (Robinson, 1966). Valko (1961) later developed a simplified method to determine the Schüepp coefficients as well as the precipitable water content from actinometric measurements.

All of these early efforts based on broadband measurements failed to provide a consistent climatology of atmospheric dust turbidity. The lengthy measurements with filtered actinometers themselves were cumbersome, delicate due to the high precision required in obtaining spectral irradiances from differences of broadband measurements and
tedious to evaluate for the variety of corrections to be applied. Consequently, they were pursued at few stations or for short periods only.

It was not until Volz (1959) had developed a handheld sun photometer that turbidity measurements became more popular. The Volz instrument used a photoelectric selenium cell that is insensitive to infrared radiation, thus eliminating the need for water-vapour corrections and permitted rapid measurements through several filters. With the introduction of metallic interference filters, quasi monochromatic measurements became possible and the evaluation was greatly simplified.

A first turbidity network was established in 1961 by the U.S. Weather Bureau, collecting daily values of the Schüepp turbidity coefficient $B$ from initially 20 and later 40 stations using the Volz sun photometer with a single channel at 500 nm (Volz, 1969). In the early 70's, the US Environmental Protection Agency (EPA) introduced a more sophisticated sun photometer with dual channels at 380 and 500 nm and replacing the selenium elements with silicon detectors. Starting in 1963, a number of European stations contributed to this network and about 12 stations provided results from their pyrheliometer measurements. The purpose of this early network was (Flowers, 1969):

1. to determine the “clean” air or background turbidity and its geographical, seasonal, and long-term variations;
2. to determine the effects of cities on turbidity;
3. to detect any unusual air pollution events, including forest fires and volcanic eruptions; and
4. to provide information related to the aerosol and gaseous pollution of the atmosphere.

A turbidity network of global extent was initiated by the World Meteorological Organization (WMO) as a component of their Background Air Pollution Monitoring Network (BAPMoN) established in 1968. It was soon (WMO, 1975) recognized that the sun photometers available at that time lacked the calibration accuracy and stability necessary for long-term monitoring in routine network operations. Nevertheless, by 1992 more than 140 stations participated in this network with a variety of different sun photometers and up to 88 of them have reported 3 daily values to the US National Climatic Data Centre (NCDC) in Asheville, NC.

After 20 years of operation, increasing doubts about the quality and usefulness of these data led to a detailed review of the NCDC archive by a group of experts (WMO, 1993a), who concluded that:

1. The collected data appear not suitable for scientific analysis of either short or long term changes in global AOD.
2. Most of the instruments used were unstable and suffered from poor calibration.
3. The algorithms for data reduction and quality control were flawed.
4. The geographical and temporal distribution of turbidity observations was inadequate to obtain a global representation of turbidity. The group thus recommended that the turbidity archive should no longer be continued.

A few turbidity programs at single research sites (National Oceanic and Atmospheric Administration at Mauna Loa, Australian Bureau of Meteorology at Cape Grim, or German
Weather Service at Lindenberg) were started in the mid 80-ies and included regular calibration and data quality control. The longest continuous data series available today thus cover about 20 years of observations.

By 1990, BAPMoN was merged with the WMO Ozone observation system into the Global Atmosphere Watch (GAW) program with the intention to improve the quality, timeliness and completeness of observational data on the chemical composition and physical characteristics of the atmosphere related to climate change. Concerning the aerosol components in GAW a group of experts considering turbidity measurements came to the conclusion that "The lack of success with many previous efforts to accomplish this task routinely must be acknowledged and deficiencies corrected by any future effort to help assure satisfactory results" (WMO, 1991). This group made detailed recommendations for future turbidity measurements, but cautioned, "It should be noted that fulfilling these requirements is not a simple or easy task". In subsequent meetings the WMO turbidity experts identified the need for a World Calibration Center, similar to what has been established for solar radiation measurements. Such a World Optical depth Research and Calibration Center (WORCC, see http://www.pmodwrc.ch/worcc) was thus established in 1996 at the Physikalisch-Meteorologisches Observatorium Davos (PMOD/WRC) as one of the Swiss contributions to the GAW programme, and given the following initial tasks:

2. Development of procedures to ensure worldwide homogeneity of AOD observations.
3. Development of new instrumentation and algorithms for AOD observations, including quality control procedures.
4. Implementation of a trial phase network at GAW stations to test new instrumentation, methods and procedures.
5. Training of station operators in AOD observations.
Chapter 3  Sun photometry

Sun photometry is a passive remote sensing technique to infer aerosol optical depth (AOD) from atmospheric transmission measurements using the Sun as a light source (e.g. Volz, 1959; Shaw, 1983; WMO, 1993b). Sunphotometry is conceptually and technically simple and can achieve high accuracy in the order of a few 0.001 optical depths with well designed and well calibrated instrumentation. Sunphotometry makes no a priori assumptions about aerosol properties, and only limited additional information (such as particle size distribution) about the observed aerosol ensemble can be retrieved from multi-wavelength observations.

Although the classic method has been described by many authors and recommendations for its implementation were given in the CIMO Guide to Meteorological Instruments and Methods of Observation (WMO, 1996), a detailed discussion of the algorithms used in the context of this work is given in the following section. Uncertainties of these algorithms are estimated from comparison of different approximations. Constraints of their practical use can then be recommended based on the assumption that algorithmic uncertainty should be significantly smaller than the calibration uncertainty of the instrumentation.

3.1 Basic equation for AOD

A sun photometer measures the direct beam solar irradiance at one or several narrow wavelength bands in arbitrary units. The measured signal $S$ is assumed to obey the Bouguer-Lambert-Beer law:

$$S(\lambda, m, R) = S_0(\lambda) e^{-m\delta} R^2 + \varepsilon,$$

(3.1)

where $S_0$ is the exoatmospheric signal at wavelength $\lambda$ and standard Sun-Earth distance of 1 astronomical unit, $m$ is the optical optical air mass along the line of sight to the Sun, $\delta$ is the total optical depth, $R=r/r_0$ is the Sun-Earth distance in astronomical units, and the term $\varepsilon$ accounts for the circumsolar sky radiance in the field of view of the sun photometer.

The total optical thickness $m\delta$ includes several terms $\delta_i$ describing the extinction by different atmospheric components: molecular scattering, gas absorption and aerosol extinction. As these components have different vertical structures, their optical optical air mass along a refracted slant path through the atmosphere are usually slightly different. Therefore, the total optical thickness has to be written as $\tau = m\delta = \Sigma m\delta_i$.

At this point, a remark about the terminology used by different authors or remote sensing communities is appropriate: According to the American Meteorological Society, optical thickness means “the (dimensionless) line integral of the extinction coefficient along any path in an optical medium”, while optical depth indicates “the optical thickness measured vertically above a given altitude”. This definition is consistent with terminology

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2 According to AMS Glossary of Meteorology: A measure of the length of the path through the atmosphere to sea level traversed by light rays from a celestial body, expressed as a multiple of the pathlength for a light source at the zenith. To get a representative value at high elevation, the above values must be multiplied by the ratio of the actual atmospheric pressure to the sea level pressure.
used by the WMO (1996). In aerosol literature optical thickness is often used as a synonym of optical depth. Here, both terms are used according to above definitions.

Taking the logarithm of this expanded Bouguer-Lambert-Beer law and re-arranging terms leads to the basic equation for determination of aerosol optical depth $\delta_A$:

$$
\delta_A = \frac{\ln(S_0) - \ln(S - \varepsilon) - 2\ln(R) - \sum_{i=1}^{n-1} \delta_im_i}{m_A},
$$

(3.2)

where the subscripts $A$ indicate aerosol specific terms.

In equation (3.2) the signal $S$ is the only measured quantity, all other terms are based on models of atmospheric extinction or of the measurement process that are approximated by relatively simple expressions for practical use. A considerable variety of approximations to compute optical air masses $m$ and non-aerosol optical depths $\delta_i$ as well as methods to determine the calibration value $S_0$ were proposed in the literature dealing with Sun photometry of the last twenty years. Many investigators have implemented their preferred mixture of algorithms and methods to determine AOD from measurements. As long as these recipes were applied consistently within a regional network or in analyzing long-term measurements of a monitoring station, the choice of algorithms was of little consequence, but in order to establish a global database, differences in AOD due to processing should become minimal.

### 3.2 Path length or Optical air mass calculations

The optical air mass is the ratio of actual path length taken by the direct solar beam to the analogous path when the Sun is overhead from the top of the atmosphere to the surface.

In a plane parallel approximation, the optical air mass is equal to the secant of solar zenith angle, but for the spherical atmosphere this simple expression is accurate to 1% only for zenith angles smaller than about 70°. If better accuracy is required, as for Langley calibration of $S$, the path lengths must be calculated by ray tracing through a refracting atmosphere.

In Sun photometry, the optical air mass can also be interpreted as the ratio of optical thickness over optical depth, which is defined by line integrals of the extinction coefficient along the path of light. These integrals depend on the vertical profile of the scattering or absorbing coefficient, and the optical air mass will be (slightly) different for each extinction component in the same geometric path.

The formulations to calculate respective optical air mass for molecular scattering, ozone absorption and aerosol extinction are discussed in the following sections.
Molecular scattering optical air mass

Molecular scattering is proportional to the density $\rho$ of air, and the corresponding optical air mass thus proportional to the integral of $\rho(z)$ along the line of sight or for a vertical path, simply proportional to the atmospheric pressure.

The height $H$ of a homogeneous atmosphere is obtained by normalization to the standard density $\rho_0$ of dry air

$$H(\zeta = 0) = \frac{1}{\rho_0} \int_0^\infty \frac{\rho(z)}{\cos \zeta(z)} \, dz .$$

(3.3)

In the simple approximation of a plane and homogeneous atmosphere, the length of a slant path with solar zenith angle $SZA = \zeta$ would simply become $H(\zeta) = H(0) / \cos(\zeta)$ and the optical air mass is $m = H(\zeta)/H(0) = 1/\cos(\zeta)$. In a curved and inhomogeneous atmosphere the solar zenith angle varies with height according to the basic equation for astronomical refraction

$$n(R + z) \sin(\zeta) = \text{const} = n_0 R \sin(\zeta_0) ,$$

(3.4)

where $n$ is the index of refraction and $R$ is the radius of the Earth and the constant is defined by the corresponding values at the surface $\zeta=0$.

Figure 1 Path of a refracted Sun ray through the atmosphere, with $\zeta$ the true (astronomical) and $\zeta_0$ the apparent (refracted) solar zenith angle. $R$ is the Earth radius and $z$ the height in the atmosphere.

After some substitution and assuming that the refractive power $\delta = n - 1$ is proportional to the density of air (Kasten, 1966), the slant path length of molecular scattering can be expressed as a function of $\zeta$. 
CHAPTER 3 Sun photometry

As the actual vertical density profile is usually unknown, it must be approximated by an analytical expression or by a suitable standard atmosphere.

Bemporad (1904) has calculated his widely used table of optical air mass assuming an exponential approximation for the density.

Later in the 20th century, with the advent of the first standard atmospheres and improved values for the refraction of air, several recalculations of Bemporad’s table were undertaken, though the differences among them were usually small (<1%) for solar zenith angles up to 85° or optical air mass smaller than 10.

Kasten and Young (1986) have used the 1972 ISO standard atmosphere and a refractive index of air at 700nm to calculate absolute and optical air masses to 5 significant digits. They also fitted an analytical expression to calculate relative optical air mass for a given apparent solar elevation angle $\gamma$ between zero and 90 degrees

$$m_r(\gamma) = \frac{1}{\sin(\gamma) + 0.50572(\gamma + 6.07995)^{-1.6364}}.$$

This fit deviates by less than 0.5% from the numerical results, and by less than 0.1% for solar elevations above 5°. It should however be remembered that these results are valid for the chosen standard atmosphere and can be off by a few percents near the horizon whenever the real atmosphere deviates markedly (e.g. temperature inversion) from the assumed mean conditions, and that the refraction correction that needs to be applied to calculate true zenith angles depends on meteorological parameters like temperature, pressure and humidity.

Ozone optical air mass

Ozone has a significantly different vertical profile in that it is concentrated in a relatively thin layer of the stratosphere at a height of about 20 to 25km for which a plane parallel approximation is appropriate, if a purely geometric correction for the height of the layer is applied. The international Ozone community has adopted a formulation according to Komhyr (1989) for the calculation of the Ozone optical air mass that is routinely used for evaluation of total ozone from Dobson measurements:

$$m_{o3}(\gamma, h) = \frac{R + h}{\sqrt{(R + h)^2 - (R + r)^2 \times \cos^2(\gamma)}},$$

with $R=6370$km the mean Earth radius, $r$ the station height above sea level in km, and $b$ a
latitude dependent estimation of the height of the Ozone layer that varies linearly from 26 km over the equator to 17 km at the poles.

In polluted areas, the tropospheric ozone can contribute 10 to 20% to the total column ozone content \( c \). If a reliable estimate of the tropospheric ozone content \( c_T \) is available (e.g. from NASA Langley Research Center under \url{http://asd-www.larc.nasa.gov/TOR/data.html}), then a weighted average of two layers could be used to calculate a more accurate optical air mass for ozone

\[
m_{o3}(\gamma, h) = \frac{m_{o3}(\gamma, h)(c - c_T) + m_{o3}(\gamma, h = 0)c_T}{c}.
\] (3.7b)

Relative differences in optical air mass calculated by formulae (3.7a) and (3.7b) are, however, less than 1%, and thus smaller than the estimated (WMO, 2003b) uncertainty of 2% for total ozone measurements.

**Aerosol optical air mass**

While the vertical profiles of air density and ozone are well known and corresponding optical air masses are readily calculated with excellent or good accuracy, little is known about the aerosol profile. Aerosols from local sources may be trapped in the boundary layer, long range transport of aerosols occurs in the free troposphere, and aerosols may be injected to the stratosphere by explosive types of volcanic eruptions. Even the presence of cosmogenic aerosols in the upper stratosphere cannot be completely excluded.

As the majority of aerosols are generated at the surface (mineral dust, salt spray, fire and combustion) a simple geometrical \( 1/\sin(\gamma) \) approach may be considered a good approximation for aerosol optical air mass. Typical scale heights for aerosol concentration are in the order of a few km (Turner et al. 2001), comparable to the scale height for water vapor, thus the formulation for water vapor optical air mass by Kasten (1966) can be used during periods of low stratospheric aerosol load.

\[
m_a \approx m_{H2O}(\gamma) = \frac{1}{\sin(\gamma) + 0.0548 \times (\gamma + 2.65)^{-1.452}}.
\] (3.8)

In cases of high stratospheric aerosol loads, a formulation similar to (3.7b), where the Ozone contents are replaced with total and stratospheric AOD components, could be used. Actual aerosol profiles can be determined from simultaneous Lidar measurements such as the micro pulse lidar deployed in the NASA MPLNET network co-located at some AERONET sites (Welton et al., 2000).

---

3 Roughly 40 tons of extraterrestrial material falls to the earth each day.
Summary of optical air mass approximations

The three different optical air masses are shown graphically in the left panel of Figure 2. For solar elevations below 10°, the ozone optical air mass differs significantly from the optical air mass for molecular scattering and aerosol (water vapor) extinction, which are nearly identical down to the horizon. The right hand panel of Figure 2 shows relative differences between simple geometric and 3 elaborated approximations for the optical air mass. Considering the cautioning remarks above, the differences between old and modern approximations seem minor and potential errors are best avoided by restricting observations to optical air masses smaller than about 6 or solar elevations larger than 9 degrees.

Although the chosen aerosol (water vapor) optical air mass closely follows the geometric one, they still differ by 1 percent at optical air mass 6, thus the question of ‘correct’ optical air mass for aerosols remains subject to discussion.

Figure 2 (left panel) Optical air mass for molecular scattering (solid), Ozone (dashed) and water vapor absorption (dotted). Right panel: Relative differences of molecular scattering optical air mass related to Kasten1989 (solid) formulation: geometric (dotted), Bemporad 1904 (asterix), Young 1994 (dashed), and water vapor (dash-dotted) for comparison.

3.3 Solar elevation and refraction

All optical air mass calculations are functions of the apparent solar zenith or elevation angle which is usually derived from the time of observation via astronomical algorithms. (the Volz instrument had a pointer attached that could be read in geometrical optical air mass) The astronomical ‘true’ position of the Sun must be corrected for atmospheric refraction which makes the solar elevation to appear higher than the calculated true position.

True Solar elevation $\gamma'$ or zenith distance $\zeta$ is calculated from the spherical trigonometric relation

$$\sin \gamma' = \sin \phi \times \sin \delta + \cos \phi \times \cos \delta \times \cos H = \cos \zeta,$$  \hspace{1cm} (3.9)
where $\phi$ is the geographic latitude of the observer, $\delta$ is the solar declination and $H$ the hour angle measured from true local solar noon. Traditionally, the hour angle was obtained from mean local time at the observer’s longitude and the equation of time. Today, when the solar equatorial position angles are readily calculated, the hour angle can be found as $H = \theta_0 - \lambda - \alpha$ with $\theta_0$ the Greenwich sidereal time reduced by geographic longitude $\lambda$ (east negative) and $\alpha$ the solar right ascension.

From among the many available algorithms to calculate solar position, the one given by Michalsky (1988) is recommended by WMO (1996) for solar radiation measurements. It is based on trigonometric series approximations to the Astronomical Almanach for 1985 and gives apparent elevation, including refraction at sea level, to an accuracy of about 0.01° which is considered sufficient for meteorological applications. (At WORCC, a more complex algorithm by Montenbruck and Pfleger (1994) which achieves an angular accuracy of 1 second of arc and $5 \times 10^{-6}$ AU in Sun-Earth distance is used for solar position calculations, simply because it was available from other work.) Both algorithms were checked over a period of several years against the U.S. Naval Observatory Multiyear Interactive Computer Almanac MICA (see [http://aa.usno.navy.mil/software/mica](http://aa.usno.navy.mil/software/mica)) and were found to be within their specification.

When an astronomical algorithm for true angles is used, the correction for refraction must be applied explicitly.

Atmospheric refraction is determined from astronomical observation of catalogued stars near the horizon or approximated by ray tracing through a standard atmosphere using the above equation (3.4). The Pulkovo tables are based on the Soviet GOST-73 standard atmosphere and are the generally accepted astronomical standard for refraction. Several authors have provided numerical fits to these and other tables that approximate refraction as a function of true (sometimes of apparent) angles and possibly of pressure and temperature.

At WORCC, an approximation with corrections for temperature and pressure given by Meeus (1991) is used.

$$r = \frac{1.02}{\tan(\gamma + \frac{10.3}{\gamma + 5.11})} \times \frac{p}{1010} \times \frac{283}{273 + T}. \quad (3.10)$$

where $r$ is the correction for atmospheric refraction in minutes of arc, $\gamma$ is the true solar elevation in degrees of arc, $p$ the atmospheric pressure in hPa and $T$ the air temperature in degree Celsius.

As a word of caution: it should be avoided to apply a refraction correction to angles calculated by an algorithm that already includes refraction, such as in Michalsky (1988). At optical air mass 6, the resulting error amounts to 4% if the correction is applied twice.

**Error estimates and useful range of observations**

The accuracy of the optical air mass used in equation (3.1) depends on atmospheric and astronomical models and the approaches to fit simple expressions to numerical results of these models. Their accuracy is roughly proportional to their value, AOD observations...
taken at large optical air mass will suffer from a degraded accuracy due to optical air mass uncertainty. It is of some interest to find an upper optical air mass limit for reliable AOD observations.

If an upper limit of uncertainty is set at a relative error of 0.1% in optical air mass, the following questions can be posed:

A. What is the upper optical air mass limit if an astronomical uncertainty of 0.01° is assumed?
B. How accurate must the time and site of observation be known?
C. What is the error in refraction and when does it become dominant?

Answer A: Relative errors of the different optical air masses for an assumed error of 0.01° in solar elevation are plotted in Figure 3 versus optical air mass. For the uncertainty in true solar angle of the Michalsky algorithm, the relative error of 0.1% is exceeded at optical air mass $\approx 6$. Relaxing the constraint to a relative error of 0.2% opens the acceptable range of observations to optical air mass 10.

![Figure 3](image.png)

Figure 3 Relative errors of optical air masses for a SZA error of 0.01°. From top down, the curves are for Ozone, molecular, water vapor to the geometric optical air mass at the bottom. The horizontal line indicates the assumed error limit of 0.1%

Answer B: Accurate timing of the measurements is most critical when the solar elevation changes fastest. That is the case for tropical sun rise and set at equinox. Measurements in polar regions are more tolerant to timing errors.

In order to get an idea about the timing accuracy, we start from the equation (3.9) which may be read as the reciprocal of geometric optical air mass. Substituting $\sin \Phi \sin \delta = a$ and $\cos \Phi \cos \delta = b$ the optical air mass $m$ can be written as
and its derivative with respect to hour angle $H$ becomes

$$
\frac{dm}{dH} = \frac{b \sin H}{(a + b \cos H)^2}.
$$

Dividing by (3.11a) the relative optical air mass error is obtained as a function of $H$

$$
\frac{dm}{m} = \frac{b \sin H}{(a + b \cos H)} \times dH.
$$

Equation (3.12) gives a relation for the required accuracy of $H$ for a relative optical air mass error of 0.1%.

The dependence of $dm/m$ on $b = \cos \Phi \cos \delta$ is analysed graphically in Figure 4 by plotting the time accuracy required to achieve a precision of 0.1% in optical air mass. The figure is intended to show the typical range of timing tolerance. Numbers for the annual course at a given location can be calculated from equation (3.12). A steep asymptotic rise of the admissible timing error occurs around solar culmination when the elevation is changing very slowly.

Whenever the Sun is well above the horizon (for optical air mass <3), the timing requirements are rather loose, but in tropical and subtropical latitudes, time has to be known to about 3 seconds if measurements are taken up to optical air mass 6. As the timing requirement scales linearly with optical air mass precision, an accepted error of 1% would permit timing within 30 seconds. Systems equipped with filter wheels to read several wavelengths in sequence should add individual time stamps to each measurement. Integration times longer than a few seconds should be avoided.
CHAPTER 3 Sun photometry

Figure 4 Timing requirement for 0.1% precision in optical air mass as a function of optical air mass. The curves depict the range of clock accuracy at latitude of 0°, 30° and 60° for both solar solstices and the equinox. The single curve in top-right corner indicates admissible timing errors near winter solstice at high latitudes.

A clock accuracy of a few seconds is in principle quite easy to achieve, but in practice more difficult to maintain in a network of remote instruments. If the data acquisition system has access to an internet timeserver or a GPS receiver, time can be kept within 1 second. If the system depends on manual interaction, great care has to be taken to ensure frequent and accurate synchronization of clocks.

The geographic latitude $\Phi$ determines the culmination height $\Gamma = 90° - \Phi + \delta$, and thus the same accuracy requirements as for the solar elevation are applicable. An accuracy of 0.01° is good enough for measurements up to optical air mass 6 with a precision of 0.1%. This specification translates to a required position accuracy of 0.6 minutes of arc or about 1km in North-South direction.

The geographic longitude $l$ is linked with time, in that 15° of longitude correspond to 1 hour of time. The accuracy requirement for time is thus directly applicable to longitude when expressed in angular or distance units. An acceptable timing error of 4 seconds translates to 1 minute of arc or slightly more than 1km East-West at mid-latitudes. The geographical coordinates of a site should thus be known within 1 minute of arc.

**Answer C**: The potential error of refraction can be estimated from the spread of results obtained by Meeus (1991) approximation and by the formulation recommended by WMO (1996). In order to test how the underlying atmospheric model influences refraction, two model atmospheres (US STD76 and midlatitude winter) were used to calculate refraction with MODTRAN4 (Berk et al., 1999) at several elevation angles and 2 different wavelengths (450nm, 600nm). MODTRAN4 is a MODerate spectral resolution atmospheric TRANsmittance algorithm and computer model. Atmospheric refraction can
be derived from the line of sight parameters that are computed by ray tracing. The results are shown in Figure 5. For optical air mass less than 7, the spread in refraction calculations derived from Figure 5 remains below the estimated uncertainty of solar elevation of 0.01°. For larger optical air mass, the refraction approximation becomes the dominant error term.

From the above analysis, it may be concluded that the chosen algorithms for optical air mass calculations have a relative error of 0.001 up to optical air mass 6 and can be used to optical air mass 10.

![Figure 5 Refraction results obtained from 2 analytical approximations (Meeus solid line, WMO dashed line) and 2 different models in MODTRAN4 calculations (stars). The error bars show the spread of 2 atmospheric models and 2 wavelengths.](image)

### 3.4 Optical depths

In the right hand side of equation (3.1) several optical depths of additional extinction components other than aerosols are subtracted from the total optical depth. The wavelengths in Sun photometry are selected such as to minimize the influence of gas absorbing, but the extinction due to molecular scattering has to be taken into account at all wavelengths. As the optical depth of molecular scattering is typically larger than of aerosols for wavelengths shorter than 500nm, it should be determined with high accuracy.

**Rayleigh scattering**

Scattering by the air molecules can be described by Rayleigh’s scattering theory for particles much smaller than the wavelength of light. Under that assumption, molecules can be represented by ideal dipoles that are exited by an incident plane wave of intensity to emit – according to Huygens – a spherical wave, the scattered radiation. The scattering
amplitude is proportional to the square of the frequency and to the polarizability of the dipole, and its angular distribution depends on the polarization of the incident wave. It varies with the cosine of the scattering angle $\theta$ for light polarized parallel to the scattering plane and is omnidirectional for light polarized perpendicular to the scattering plane. If the incident light is unpolarized, the scattered light will be partially polarized.

For a non-absorbing, spherical particle of radius $a$ with a relative index of refraction $n$, illuminated by unpolarized light of wavelength $\lambda$, the scattered intensity at distance $r$ can be written as:

$$I(\theta, \lambda) = \frac{a^6}{r^2 \lambda^4} \left( \frac{n^2 - 1}{n^2 + 2} \right)^2 \left( 1 + \cos^2(\theta) \right) I_0,$$

(3.13)

and the polarization as

$$P = \frac{1 - \cos^2 \theta}{1 + \cos^2 \theta}.$$

(3.14)

The molecular, or Rayleigh, scattering intensity varies with the inverse forth power of the wavelength. It is symmetric in forward and backward direction and completely polarized at a scattering angle of 90°.

In the case of Rayleigh scattering in the atmosphere, the expression becomes a bit more complicated. First, the index of refraction is a function of wavelength and thus causes a slight deviation from the $\lambda^{-4}$ law. Furthermore, the air molecules are not ideal spheres (except for the rare gases) and their anisotropy increases Rayleigh scattering slightly by a factor $1 + \rho$, and decreases the degree of polarization to $P = (1 - \cos^2 \theta) / [1 + \cos^2 \theta + 2 \rho/(1 - \rho)]$, where $\rho$ is the so called depolarization factor. Different molecules have different depolarization factors, e.g. 0.037 for N₂ or 0.103 for CO₂. Penndorf (1957) has measured a value of about 0.03 for air.

Equation (3.13) can be rewritten to calculate the Rayleigh scattering cross section

$$\sigma(\lambda) = \frac{24 \pi^3}{\lambda^4} \left( \frac{n^2(\lambda) - 1}{n^2(\lambda) + 2} \right)^2 \left( 6 + 3 \rho(\lambda) \right) \left( 6 - 7 \rho(\lambda) \right),$$

(3.14)

where, $N_s$ is the molecular number density and $n(\lambda)$ is the index of refraction and $\rho(\lambda)$ the depolarisation factor for air. The right-most fraction $(6 + 3\rho)/(6 - 7\rho)$ is often called the King factor.

The Rayleigh scattering optical depth $\delta(\lambda)$ of the atmosphere can then be calculated as the path integral of the product of cross section and particle density.
\[ \delta(\lambda, z_0) = \int_{z_0}^{\infty} \sigma(\lambda) N(z) dz \quad (3.15) \]

Many authors have performed such integrations and given tabulations or approximation formula for Rayleigh optical depth, using different and improved formulations for the index of refraction and the depolarization factor as they became available. Teillet (1990) gives an overview of historic Rayleigh calculations.

Among the more recent calculations, Bucholtz (1995) has used a wavelength dependent depolarization factor from Bates (1984) and a numerical fit for the refractive index given by Peck and Reeder (1972) to calculate scattering cross section that differ by +1.7% to -1.4% from previous results. Arguing that the refractive index is a function of number density, he integrated the cross section using 6 different atmospheric models and presented a set of approximations that fit his calculations to better than 0.2% below and better than 0.1% above 500nm. Bréon (1998) gave arguments for the independence of Rayleigh optical depth from the numerical integration of the number density, which in fact will always be proportional to the surface pressure according to

\[ \int_{0}^{\infty} N(z) dz = \int_{0}^{p_0} \frac{-dp}{M_{\text{air}} g(z)} = \frac{p_0}{M_{\text{air}} g'} \quad (3.16) \]

where \( M_{\text{air}} \) is the molecular mass of air and \( g \) is the acceleration of gravity.

The introduction of an equivalent acceleration of gravity \( g' \) in the last expression accounts for the decrease of \( g \) with height. Numerical integration with typical atmospheric profiles showed that \( g' \) is smaller than \( g_0 \) by 0.23±0.01%.

According to Bréon, the Rayleigh optical depth can be found from surface pressure

\[ \delta(\lambda) = \frac{\sigma(\lambda) p_0}{g' M_{\text{air}}} \times k_{wv}, \quad (3.17) \]

where \( k_{wv} \) is an additional, small correction factor for the partial pressure of water vapor.

Bodhaine and co-authors (1999) proposed to calculate Rayleigh optical depth by the method of Bréon, using the same index of refraction and wavelength dependent depolarization factor as Bucholtz. In addition to these similarities, they use a latitude and height dependent formulation for the equivalent acceleration of gravity and a correction to the refractive index and depolarization factor accounting for increased CO\(_2\) concentration. Their calculation remains simple enough to be performed in a spreadsheet and is used in our evaluation of AOD.
Figure 6 gives a summary of several formulations of Rayleigh optical depth plotted as difference to the calculation by Bodhaine et al. (1999) in units of milli optical depth. The estimated uncertainty for recent calculations based on the same or similar values for the refractive index and depolarization factor is in order of 0.0025 optical depths for UV wavelengths, and less than 0.001 for wavelengths above 400nm. Hoyt (1977) has noted that the larger values by Penndorf may overestimated Rayleigh optical depth and be responsible for the negative aerosol optical depths occasionally found in historical data.

![Figure 6 Differences of several Rayleigh optical depth approximations (as indicated) to values given by Bodhaine.](image)

All Rayleigh optical depth values are given for a sea level pressure of 1013.5 hPa, but can be scaled proportionally for different atmospheric pressure. The pressure should be known with such accuracy that the error in pressure scaled Rayleigh correction does not exceed its estimated uncertainty of 0.001 to 0.0025 optical depths. This condition requires a pressure accuracy of about 3hPa or better for wavelengths below 400nm.

At Davos, the mean pressure around noon for 'clear days' is 2.7±6 hPa higher than the average noon value for all weather conditions or 3.9 hPa higher than the barometric value of a standard atmosphere at the height of Davos. From these considerations it follows that either simultaneous pressure measurements or at least daily pressure values at noon are needed to calculate Rayleigh optical depth with a desired accuracy of 0.002.

**Ozone absorption**

The ozone optical depth \( \delta_{O3}(\lambda) = a_{O3}(\lambda) \epsilon \) can be calculated from the spectral absorption coefficient \( a_{O3}(\lambda, T) \) and the atmospheric column content \( \epsilon \).

Gueymard (1995) has compiled recent measurements published by several authors in a continuous table of the ozone absorption coefficients at a common reference temperature of 228 K. As the actual effective temperature of the ozone layer is usually not known, the temperature dependence of the absorption coefficient is only moderate (<0.5%/K) and the
ozone correction is relatively small in the Chappuis band, we use the absorption coefficients as given by Gueymard.

Among the standard AOD wavelengths, only the channels at 500 and 675 nm need to be corrected for Ozone optical depth. The correction amounts to about 0.01 optical depths for an average concentration of 0.3 atm-cm or 300 Dobson units. Thus, for an OD uncertainty of ±0.001, the total ozone needs to be known to ±30 Dobson units or 10% of the measured concentration. Day to day variation can be up to 20%, thus daily measurements of concentrations are required to achieve a 0.001 level of uncertainty for the ozone correction.

Total ozone data are often available for a local or nearby Dobson or Brewer UV spectrometer at the World Ozone and Ultraviolet Data Centre (http://www.woudc.org) in Toronto or from satellite observations of the Total Ozone Mapping Spectrometer (TOMS) or the Ozone Monitoring Instrument (OMI) that may be found under http://toms.gsfc.nasa.gov/ozone/ozone.html on the internet.

**Nitrogen dioxide absorption**

NO₂ is a highly variable atmospheric constituent which plays a key role in the complex ozone cycle, both in the stratosphere where it is naturally present with a concentration in the order of \(10^{-4}\) atm-cm, and in the troposphere, where its concentration may reach \(10^{-2}\) atm-cm in heavily polluted areas. Although these concentrations are typically two orders of magnitude lower than ozone, the corresponding optical depth at wavelengths around 400 nm is comparable to those for ozone at 500 nm, and should therefore be corrected in AOD measurements. Absorption coefficients are given by Gueymard (1995) or by Burrows et al. (1998).

Starting from April 1996, total and tropospheric NO₂ columns are retrieved from satellite observations with GOME and SCIAMACHY instruments. Data are available as daily or monthly means from the Tropospheric Emission Monitoring Internet Service TEMIS under http://www.temis.nl.

**3.5 Diffuse stray light**

Sun photometers inevitably receive some diffuse stray light from atmospheric forward scattering in their finite field of view (FOV). As a result, the measured direct beam signal \(S\) is increased by a diffuse component \(\varepsilon\) leading to an overestimation of the transmission and thus to an underestimation of the optical depth. Under hazy conditions, this diffuse stray light is easily noticed as whitish circumsolar radiance of the sky, while under low turbidity conditions, the solar disk contrasts clearly from the blue sky. The circumsolar radiation is proportional to the aerosol optical depth and strongly dependent from the aerosol scattering phase function. As the latter is not retrievable from classic Sun photometry, the circumsolar stray light \(\varepsilon\) must be estimated by model calculations based on assumed aerosol characteristics. The aerosol phase function is strongly peaked in the forward direction, exceeding the Rayleigh phase function at scattering angles close to zero by 2 orders of magnitude. Thus, the contribution of Rayleigh scattering to diffuse radiation can be neglected, even at shorter wavelengths, where \(\delta_R > \delta_A\).

Grassl (1971) has calculated the circumsolar radiation as a function of aerosol type (maritime or continental), optical depth, wavelength, and instrument field of view. He
reported relative stray light contributions for a 1° FOV between 0.1% and 2% of the incoming radiation at 500nm, depending on the aerosol model and optical depth.

The Ångström wavelength exponent $\alpha$ can be derived from multi-wavelength Sun photometry and gives a rough indication of the dominant mode of the aerosol size distribution, with smaller values of $\alpha$ indicative of larger particles. As large particles exhibit a more pronounced forward peak of the scattering phase function than small particles, and hence a brighter circumsolar radiances, an inverse relationship between $\alpha$ and $\varepsilon$ may be expected.

Russel et al. (2004) have calculated diffuse light correction factors for a wide range of realistic aerosol types observed by AERONET. Their analysis of these correction factors revealed a good correlation with the Ångström wavelength exponent $\alpha$. They have provided a best-fit equation to approximate correction factors in the form of

$$C_f = \frac{\delta}{\delta^*} = A \cdot e^{B\alpha}$$

where $\delta$ and $\delta^*$ are true and apparent aerosol optical depths and $A$ and $B$ are wavelength dependent factors given for 3 different FOV at 14 wavelengths. These corrections are negligible (<1% of apparent AOD) for narrow ($<\approx 1^\circ$) half angle FOV, but can become as large as 10% of AOD at 350nm for a FOV $\approx 2^\circ$.

### 3.6 Cloud screening

Evaluation of AOD is severely hampered by the presence of optically thin water or ice clouds in the path of observation. Cloud contamination can be avoided in manual measurements by visual judgment of trained personnel, though this may introduce some subjective bias to the results of data analysis. Unattended observations gathered by automated equipment need to be screened for clouds by algorithmic methods, which should provide reliable results under varying degrees and types of cloud interference. Several authors (e.g. Harrison and Michalsky, 1994; Smirnov et al., 2000; Alexandrov et al., 2004, Kaufman et al., 2006) have proposed such algorithms based on various assumptions about the physical, temporal, or spectral differences between aerosols and clouds. In the quality control for PFR measurements, three different methods of cloud-screening based on adaptations of published algorithms are applied.

The “objective” method of Harrison and Michalsky (1994) was primarily developed for Langley calibrations using measurements at optical air mass larger than 2. It is based on the fact that the first derivative of the measured irradiance with respect to optical air mass $\delta S / \Delta m$ is always negative under clear sky conditions. For any paired samples, where $\delta S=(S_{n+1}-S_n)/(m_{n+1}-m_n) > 0$, the sample $S_{n+1}$ is probably affected by a cloud, either intruding the path of observation before noon or leaving the path in the afternoon. Cloud perturbations 'towards noon', that is rising irradiance level in the morning due to a thin cloud leaving the line of sight, or falling irradiance in the afternoon, cannot be discerned by looking at the sign of the derivative alone, which is also negative in these cases. Harrison and Michalsky thus identify intervals where $\delta S > 0$ and, assuming cloud perturbations to be symmetric in time, eliminate an additional interval of equal duration before or after the perturbation. In a second step, they eliminate measurements where the first derivative is less than twice the mean over all remaining derivatives. This method has the advantage that it requires no instrument specific information like wavelength or calibration value except the solar elevation, and is thus applicable to any, even a broadband radiometer. Such
differential filtering does however not work very well near solar culmination, because the division by small optical air mass changes between consecutive samples strongly overrates any signal fluctuations. Thus clouds around noon are not reliably filtered by this method. This is no big restriction for Langley calibrations, but when instantaneous optical depths are retrieved, the number of samples at small optical air mass may be significantly reduced by false cloud detection.

The AERONET “triplet” cloud screening algorithm (Smirnov et al. 2000) is based on the assumption that the temporal variation of AOD is typically smaller than of cloud optical depth. The short-term variability is derived from so called "triplets" of 3 observations \( \tau_1, \tau_2, \tau_3 \) taken 30 seconds apart. Measurements are excluded if the triplet variability \( \Delta \tau = \max(\tau_{1,2,3}) - \min(\tau_{1,2,3}) \) exceeds an empirical threshold value of 0.02 or 0.03 times the triplet average whatever is larger at any wavelength. In a second and third step, the stability and smoothness of daily AOD variation is analyzed by statistical tests on daily sets of cloud-free triplets. An iterative process eliminates samples until the desired stability is achieved or the number of remaining samples falls under a specific limit. Unlike the objective method, additional information about the radiometer is required for the calculation of AOD. Quality assurance in AERONET Level2 data includes manual cloud screening, too.

Cloud screening is applied as part of the PFR data quality control and recorded by 3 individual binary flags indicating the possible presence of optically thick clouds, thin clouds detected by a differential filter, and thin clouds detected by a modified triplet filter for each observation.

1) Optically thick (\( \tau > 2 \)) clouds are detected by relating the PFR signal from the IR channel (862nm) to a simple model \( S(m) = S_0 \cdot \exp(-\tau \cdot m) \) and flagging samples where the measured signal is less than the modelled value. This rule is applicable to individual samples but does not work so well for large optical air mass \( m > 6 \), when the modelled signal falls below the dark signal level.

2) Thin clouds are detected by analyzing the first derivative of the PFR signal at 412nm, similar to the objective method, but instead of relating derivatives to twice their daily average, they are compared to a multiple of the instantaneous derivatives of the Rayleigh extinction. This modification allows the filter to work sequentially on paired samples and makes it slightly more robust for data taken at small optical air mass.

3) A second filter for thin clouds is based on the AERONET triplet method, which becomes a moving window filter for the rate of change of total optical depth when it is applied to the continuous, equidistant series of PFR measurements. This filter sets a cloud flag if the spread of total optical depth at 412nm for 3 neighboring samples exceeds an adjustable threshold value. Additional statistical tests to eliminate potential cloud contamination are applied during the calculation of hourly mean.

None of the filter algorithms is able to unambiguously identify the presence of a cloud perturbation for individual measurements, but their combination is expected to provide a more reliable cloud identification than any given single method. The cloud-filters for PFR measurements were tested statistically on measurements taken in 2003 at the following stations in the PFR network:
In Table 1, the percentage of cloud contamination, identified by different filter algorithms and combinations thereof, is listed with respect to the total number of measurements taken in 2003. The number of samples considered cloud-free by all 3 filters is less than 45% at all stations, and even less than 10% at HPB, but none of the individual filters reject more than 55% of the measurements as cloudy. Thus the large percentage of suspected cloud contamination in the measurements is due to non-overlapping cloud detection by the different filter algorithms. Indeed, when looking at the thin cloud filters #2 and #3 alone, their mutual agreement (row 6) is smaller than their disagreement (row 7) at three stations and equally large at HPB only. When looking at the cases where a thin cloud is detected by one, but not by the other filter (rows 8 and 9), a clear distinction can be made between the high altitude station JFJ, where filter #3 has a significantly lower probability (5.8%) to indicate an unconfirmed cloud than filter #2 (28.5%), and the mid altitude station HPB, where the inverse situation occurs. At the low altitude station BLO, the unconfirmed probabilities are about equal. It can be concluded that the performance of both filters depends rather strongly on the dominant type of cloudiness at a given station.

Table 1 Cloud filter statistics for 2003 at 4 PFR stations given as percentag of cloud affected samples. Filter #1 acts on single samples (static signal level), Filter #2 uses paired samples (first derivative), and Filter #3 is based on triplet variation.

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<thead>
<tr>
<th>GAWPFR station</th>
<th>HPB</th>
<th>ASP</th>
<th>JFJ</th>
<th>BLO</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Number of samples</td>
<td>146664</td>
<td>281113</td>
<td>138450</td>
<td>294980</td>
</tr>
<tr>
<td>2 Cloud-free samples</td>
<td>9.4%</td>
<td>41.4%</td>
<td>18.0%</td>
<td>22.4%</td>
</tr>
<tr>
<td>3 Filter #1</td>
<td>61.3%</td>
<td>24.1%</td>
<td>61.0%</td>
<td>41.7%</td>
</tr>
<tr>
<td>4 Filter #2</td>
<td>47.5%</td>
<td>42.1%</td>
<td>43.4%</td>
<td>47.8%</td>
</tr>
<tr>
<td>5 Filter #3</td>
<td>63.4%</td>
<td>28.4%</td>
<td>20.7%</td>
<td>44.8%</td>
</tr>
<tr>
<td>6 Filter #2 AND #3</td>
<td>37.8%</td>
<td>21.1%</td>
<td>14.9%</td>
<td>29.8%</td>
</tr>
<tr>
<td>7 Filter #2 XOR #3</td>
<td>35.4%</td>
<td>28.3%</td>
<td>34.3%</td>
<td>32.9%</td>
</tr>
<tr>
<td>8 Filter #2 exclusive</td>
<td>9.8%</td>
<td>21.1%</td>
<td>28.5%</td>
<td>18.0%</td>
</tr>
<tr>
<td>9 Filter #3 exclusive</td>
<td>25.7%</td>
<td>7.3%</td>
<td>5.8%</td>
<td>14.9%</td>
</tr>
</tbody>
</table>

The relative performance of the different PFR cloud screening algorithms with respect to optical air mass is shown in Figure 7. As expected, the modified objective method (top left panel) performs best (least number of unconfirmed flags) at optical air masses larger than 3 and should be preferred method in Langley calibrations, for which it was originally designed.
Figure 7 Frequency distribution of thin cloud filter flags at Bratt's Lake in 2003. Filter actions are normalized by the number of measurements in optical air mass interval shown in bottom right panel.

The Triplet filter (top right) works best for optical air mass smaller than 3 where more than 75% (bottom right) of the observations occur at mid-latitude stations. It becomes the preferred method in AOD evaluation. Coincident cloud detection (bottom left) works relatively well over a wide range of optical air mass and is a good compromise for routine cloud screening of the PFR measurements.
CHAPTER 3 Sun photometry
Chapter 4 Calibration of sun photometers

Aerosol optical depth is a dimensionless quantity that is not directly measurable, but must rather be retrieved from observed atmospheric transmission, and can therefore not be related to any reference. Atmospheric transmission itself is a relative value relating the (spectral) irradiance $S(\lambda)$ measured at ground level to the solar irradiance $S_0(\lambda, r)$ at the top of the atmosphere. Due to this relative nature of transmission, the signal of a sun photometer can be measured in arbitrary units, and the signal $S_0$ it would read on top of the atmosphere, often called its extraterrestrial value (ETV), can be considered its calibration coefficient for AOD.

Because optical depth $\delta$ is a logarithm quantity, its error $\Delta \delta$ is proportional to the relative calibration error $\Delta S_0 / S_0$.

$$\delta + \Delta \delta = \frac{\ln \left[ S_0 (1 + \Delta S_0 / S_0) \right] - \ln(S)}{m} = \delta + \frac{\ln(1 + \Delta S_0 / S_0)}{m}$$

or

$$\Delta \delta = \frac{\Delta S_0}{S_0 m} \quad (\Delta S_0 << S_0) \quad (4.1)$$

Thus a calibration error of 1% results in a maximum error in optical depth of 0.01 at optical air mass $m=1$. Under clean air conditions, where AOD is often less than 0.1, this would represent a rather large relative error in AOD. The current GAW specification (WMO, 2005a) calls for an AOD uncertainty of $\Delta \delta=0.005+0.01/m$ requiring a calibration accuracy of 1%. This specification is similar to the accuracy specification for satellite retrievals of $\Delta \delta=0.015$ over land and $\Delta \delta=0.010$ over the ocean (Chylek et al, 2003).

The ETV calibration coefficient $S_0$ for individual sun photometers is usually determined by atmospheric extrapolation methods. Networked instruments can also be calibrated by comparison with a reference radiometer. Sun photometers can also be calibrated radiometrically, so that their measurements are given in units of Watt/m$^2$/nm, and their AOD calibration coefficient then determined from a measured solar spectrum. Finally, their ETV can be measured in-situ, when a sun photometer is brought to the top of the atmosphere on a suitable platform. These different options are summarized below and will be described in the following paragraphs:

1. Extrapolation through the atmosphere
2. Comparison to reference radiometer
3. Radiometric calibration combined with solar spectrum
4. in-situ measurements from balloon or satellite platforms

4.1 Atmospheric extrapolation

According to the Beer-Lambert-Bouguer law $S(m) = S_0 \exp(-\delta m)$, the logarithm of $S_0$ can be obtained from two radiometer readings $S_i$ taken at different optical air mass $m_i$ by plotting the logarithm of measured signals $S_i$ versus optical air mass $m_i$ and fitting a straight
line through these points. The slope of that line represents the optical depth $\delta$, and the ordinate intercept at optical air mass zero the logarithm of $S_0$. This method is commonly known as Langley plot in honour of Samuel P. Langley of the Smithsonian Institution who first used it around 1910 to determine the solar constant with his spectrobolometer.

While two measurements at different optical air masses are in principle sufficient to find the 2 unknowns in a Langley plot, it is common practice to use a larger number of measurements and to find the unknowns as maximum likelihood estimations of an over determined system of equations. The underlying assumption is that the variations in optical depth are small and randomly distributed.

Great care must be taken in calibrating spectral radiometers by the Langley plot method that the assumption of constant optical depth during measurements can be safely assumed. Unfortunately, this condition cannot be verified by the Langley plot method alone, because the optical depth variation itself is minimized by the maximum likelihood process. When the distribution of optical depth fluctuations is assumed to be normal, common least squares fitting techniques can be used to solve the system of normal equations. Suitable algorithms can be found in many textbooks (e.g. Press et al., 1986) and are provided by various software packages.

However, certain cases of systematic variations of the optical depth can lead to systematic errors in the calibration constant as Shaw (1976) has pointed out: “If, for instance, the optical depth increases in proportion to $\cos(SZA) \approx 1/m$ or proportional to a parabolic function in hour angle, the measured signal can be modelled as

$$S(m) = S_0 e^{-\delta_1 + \frac{\delta}{m}}.$$  \hspace{1cm} (4.3)

the extrapolation of these signals will result in a reduced calibration constant”

$$S_0' = S_0 e^{-\delta}.$$  \hspace{1cm} (4.4)

From this example, it becomes obvious, that the potential relative error in calibration will be roughly proportional to the absolute value of the variation in optical depth, that is a $\delta_1$ of 0.01 will produce a -1% error in calibration. It is also obvious, that the logarithmic signal

$$\ln(S) = \ln(S_0) - \delta_0 m - \delta_1 = \ln(S_0') - \delta_0 m$$  \hspace{1cm} (4.5)

is a linear function in $m$ and that the Langley plot would thus show a perfect fit to a straight line.

Systematic variations of the optical depth are dominated by variations in the aerosol characteristics during measurements. Their concentration, composition and size spectrum can be modified by advection, convection and evaporation processes that are driven by insolation, which is proportional to $\cos(SZA)$ or $1/m$, thus Shaw’s assumed diurnal variation of optical depth is based on plausible causes. Local aerosol sources, like industrial areas or changes in surface coverage (vegetation, soil, water) around the observing site may
introduce spatial variations in optical depth as the line of sight sweeps different azimuthal
directions during observations at different solar zenith angles.

Considering these aspects, it becomes clear that reliable results from Langley plots can be expected only at sites and in periods where potential variations in optical depth are minimal. A first criterion would require small overall aerosol optical depths that can be found at remote sites, preferably above the turbulent mixing zone and far away from strong sources. A second criterion would call for rapid variations of the optical air mass \( m \approx \sec(SZA) \) during measurements in order to limit the time span for potential AOD variation. The SZA changes fastest near the horizon, but the rate of change depends also on solar declination (season) and geographical latitude, being quickest around equinox in the tropics. At the equator, it always takes less than 1.30 hours to cover the optical air mass interval between \( m=2 \) and \( m=6 \), at 45° latitude it takes 2 to 3:45 hours, and at 60° latitude more than 3 hours during summer only. The coincident meeting of both criteria at Mauna Loa, in combination with the available infrastructure, makes that place the widely recognized, world-prime site for Langley calibrations. At other, less fortunate locations, the procedure to obtain reliable calibration constants by atmospheric extrapolation methods can be very tedious or even impracticable.

The classic form of Langley extrapolation, where the logarithmic signals are extrapolated versus optical air mass \( m \)

\[
\ln(S) = \ln(S_0) - \delta m, \quad (4.6)
\]

was, and still is current practice (Shaw et al., 1973; Holben et al, 1998; Michalsky et al, 2001).

Due to different vertical structures of Rayleigh scattering and aerosol extinction the respective optical air masses become noticeably different near the horizon with the aerosol optical air mass \( m_A \) being larger than the Rayleigh optical air mass \( m_R \). Beer's law should thus more correctly be written as

\[
\ln(S) = \ln(S_0) - (\delta_R m_R + \delta_A m_A), \quad (4.7)
\]

where the total optical depth \( \delta \) is separated into components \( \delta_R \) and \( \delta_A \). The actual aerosol optical air mass is depending on the vertical distribution of aerosols, which may vary considerably. Schotland and Lea (1986) have estimated the bias error of the classic Langley extrapolation to be of the order of 1% by numerical integration of different vertical profiles of tropospheric aerosols to obtain the aerosol air mass \( m_A \). Using Lidar observations at Mauna Loa of the stratospheric aerosol component after the El Chichon volcanic eruption, they showed that it is possible to reduce this bias error below 0.1%. While actual aerosol profiles are rarely available, their result clearly shows the necessity to use a specific aerosol air mass for accurate Langley calibrations.

As an improvement over the classic method, some authors (Bruegge et al, 1992; Russel et al, 1993) have proposed a refined Langley method where a single regression of measurements versus aerosol optical air mass \( m_A \) is performed and the signal extinctions due to Rayleigh scattering and gas absorption are compensated by adding approximations of the corresponding optical depths.
\[
\ln(S) + \delta_R m_R + \delta_{O_3} m_{O_3} + \delta_{N_2} m_{N_2} = \ln(S_0) + \delta_A m_A.
\] (4.8)

Solving this equation for \( \delta_A \) is the common way to derive momentary aerosol optical depths from a calibrated instrument. Thus both the instrument calibration and data reduction process are based on the same model of atmospheric transmission. This ‘refined’ Langley plot has become the preferred method at WORCC for calibration of PFR instruments.

When assuming that random fluctuations or systematic variations \( \Delta \delta = | \delta_i - \delta_0 | \) of the optical depth are the dominant source of error, it is easily seen that measurements at large optical air mass \( m_i \) are given more weight in the calibration error \( | \Delta \log(S_0) | = m \Delta \delta \). On the other hand, automated measurements are usually taken at regular intervals and are thus denser at small, and sparse at larger optical air mass values. This uneven distribution of samples in Langley plots gives more statistical weight to measurements at small optical air masses. Various schemes were developed to account for this variable weight of samples in Langley plot regressions. AERONET, for example, employs a sampling scheme of taking measurements at regular optical air mass intervals for optical air mass larger than 2.

Both classic and refined Langley plots find the calibration constant \( S_0 \) (and the mean optical depth \( \delta_0 \)) as least squares solutions to the minimization problem of the form

\[
\chi^2(S_0, \delta_0) = \sum_i (\ln(S_i) - \ln(S_0) - \delta_0 m_i)^2.
\] (4.9)

Herman et al. (1981) have derived an alternative Langley method that minimizes the variance \( \Sigma[\delta_i - \delta_0]^2 \) in optical depth. The least squares solution becomes then the familiar linear regression coefficients of \( \log(S)/m \) versus \( 1/m \). Herman et al. have shown that the normalized standard deviations \( \sigma(\log(S_0))/\sigma(\delta) \) of their weighted fit are always smaller than in the classic Langley plot.

**Synthetic Langley plots**

In order to compare results from different Langley techniques, a synthetic dataset of atmospheric transmissions was calculated with MODTRAN4 code using the US standard 1976 atmosphere model. The dataset comprises transmission values at three wavelengths 862, 500, 368 nm, covering optical air masses between 2 and 6 in equidistant steps of 0.5, and 4 different configurations of observation height and aerosol types labeled M1 to M4 in Table 2.

**Table 2 Configuration parameters for MODTRAN4 model calculations of synthetic transmission data. Ångström turbidity parameters \( \alpha \) and \( \beta \) are calculated from corresponding AOD results at 862, 500 and 368 nm.**

<table>
<thead>
<tr>
<th>Model</th>
<th>Height</th>
<th>Aerosol model</th>
<th>Visibility</th>
<th>( \alpha )</th>
<th>( \beta )</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>50 m</td>
<td>Standard</td>
<td>23 km</td>
<td>1.19</td>
<td>0.13</td>
</tr>
<tr>
<td>M2</td>
<td>1600 m</td>
<td>Standard</td>
<td>23 km</td>
<td>1.20</td>
<td>0.10</td>
</tr>
<tr>
<td>M3</td>
<td>1600 m</td>
<td>Moderate, aged volcanic</td>
<td>23 km</td>
<td>1.14</td>
<td>0.12</td>
</tr>
<tr>
<td>M4</td>
<td>3600 m</td>
<td>Background stratospheric</td>
<td>50 km</td>
<td>1.33</td>
<td>0.03</td>
</tr>
</tbody>
</table>
Table 3 lists the results and their statistical error for different Langley methods applied to these synthetic datasets. The classic method always overestimates the extraterrestrial value by up to 4%. Acceptable errors of less than 1% are obtained only for configuration M4 confirming the widespread opinion that sun photometers need to be calibrated at high altitudes. AOD values derived at sea level by the WMO recommended algorithm with specific optical air mass from measurements with a sun photometer calibrated by the classic method at high altitude will be 0.01 to 0.03 optical depths too large. In a model run with aerosol extinction switched off, these errors vanish altogether, thus they are related to the problem that the aerosol extinction is not well approximated by the Rayleigh optical air mass.

The refined method can produce calibration values within 0.1%, except in configuration M3 when the assumed aerosol optical air mass does not represent the aged volcanic aerosols in the stratosphere well. Even then, the error is less than 0.5% and in all cases the statistical errors never exceed 0.1%, suggesting that the refined method can be successfully used for calibrations at low altitudes. These results also support the use of the Kasten water vapor air mass as a good approximation for the aerosol air mass as an underestimation of about 1% is obtained when just the simple expression $1/\sin(\gamma)$ is used.

The alternative method is always underestimating the calibration values by up to -1.7% in configuration M3.

Table 3 Extraterrestrial transmissions with statistical errors resulting from different Langley methods applied on synthetic data.

<table>
<thead>
<tr>
<th></th>
<th>862nm</th>
<th>500nm</th>
<th>368nm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>M1</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Classic</td>
<td>1.014±0.0027</td>
<td>1.026±0.0050</td>
<td>1.038±0.0075</td>
</tr>
<tr>
<td>Refined</td>
<td>1.000±0.0001</td>
<td>1.000±0.0002</td>
<td>0.999±0.0003</td>
</tr>
<tr>
<td>Alternative</td>
<td>0.999±0.0001</td>
<td>0.993±0.0017</td>
<td>0.981±0.0048</td>
</tr>
<tr>
<td><strong>M2</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Classic</td>
<td>1.012±0.0022</td>
<td>1.021±0.0040</td>
<td>1.031±0.0061</td>
</tr>
<tr>
<td>Refined</td>
<td>1.001±0.0002</td>
<td>1.001±0.0004</td>
<td>1.001±0.0006</td>
</tr>
<tr>
<td>Alternative</td>
<td>1.000±0.0001</td>
<td>0.995±0.0013</td>
<td>0.985±0.0037</td>
</tr>
<tr>
<td><strong>M3</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Classic</td>
<td>1.010±0.0018</td>
<td>1.018±0.0035</td>
<td>1.029±0.0057</td>
</tr>
<tr>
<td>Refined</td>
<td>0.998±0.0005</td>
<td>0.997±0.0005</td>
<td>0.997±0.0003</td>
</tr>
<tr>
<td>Alternative</td>
<td>0.998±0.0004</td>
<td>0.993±0.0017</td>
<td>0.983±0.0042</td>
</tr>
<tr>
<td><strong>M4</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Classic</td>
<td>1.002±0.0004</td>
<td>1.003±0.0007</td>
<td>1.008±0.0017</td>
</tr>
<tr>
<td>Refined</td>
<td>1.000±0.0000</td>
<td>1.000±0.0000</td>
<td>1.000±0.0002</td>
</tr>
<tr>
<td>Alternative</td>
<td>0.999±0.0001</td>
<td>0.995±0.0012</td>
<td>0.988±0.0031</td>
</tr>
</tbody>
</table>

Day to day variations in $S_0$ are often much larger than the standard error for any given day due to subtle variations of the atmospheric stability on individual days. Therefore, a reliable calibration with an uncertainty of 1% has to be found as a statistical average over a sufficient number, typically about 20-30, of ‘good’ Langley extrapolations.

Depending on atmospheric conditions, e.g. continental summer, the required number of Langley plots may span over several months, during which a sun photometer channel may show degradation. It may become necessary under such circumstances to interpolate 'good'
Langleyplots by a time dependent regression of the form $S_0(t)=S_0(t_0) + c(t-t_0)$ referring to a given date of reference $t_0$. As with the primary Langleyplots, various forms of least-squares techniques, including weighted or robust methods (Street, 1988; e.g. in Matlab Statistics Toolbox), can be used and should include estimates of standard errors for $S_0(t_0)$ and the drift coefficient $c$.

**Robust Langley calibrations**

The calibration method used at WORCC is based on a linear regression of a series of automated Langley calibrations and subsequent interpolation to daily extraterrestrial coefficients.

On each sunny day, the measurements in the morning and afternoon are subjected to separate Langley calibrations if more than 10 cloud-free measurements can be identified in a range between optical air masses 2 and 6. A refined Langley regression is then performed in three iterations, successively eliminating measurements with potential cloud contamination based on upper and lower constraints on the residuals of the linear regression. The last step provides final estimates of the calibration value and its standard error that are collected in a calibration history file for further analysis.

This calibration history is then analysed over periods of typically 6 to 12 months in order to find representative calibration values and their drift rates including estimated uncertainties. Langley extrapolations within the chosen period are screened for a standard error of less than 1% of the calibration value. The frequency distribution of such calibrations is shown on the left panel of Figure 8. A robust linear regression versus time fitted through calibrations that are within ±2 standard deviations of the mean is shown on the right panel of Figure 8. Finally, this regression can be used to obtain daily calibration values for the period where its 95% confidence interval remains below a required limit (e.g. ±1%) of the calibration value.

![Figure 8 Results of Langley calibrations over 1 year. Left panel: Frequency distribution of N=88 calibrations around a mean value of 3.763±0.278 (2σ). Right panel: Robust regression with a slope of +0.5%/year (solid line), 95% confidence interval (dotted) and 1% limits (dashed).](image)

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This statistical method provides an objective way for validating the calibration of field instruments at network stations where a minimal number of 20 – 30 Langley extrapolations per year are possible.

Figure 9 shows the long-term evolution of the calibration coefficients for 4 wavelengths of a PFR instrument at a low-altitude station in the GAW network obtained over 6-month intervals by the method described above. The overall drift of the instrument is smaller than 0.4% per year. In April 2004, a high altitude calibration campaign at Mauna Loa confirmed the adjacent in-situ calibrations to within better than 0.4%. At high altitude stations, this level of uncertainty can be achieved in much shorter time spans.

Figure 9 Temporal evolution of semi-annual calibrations of a PFR instrument at a low-altitude site in the GAW network. Error bars on lower series indicate 95% uncertainty of the Langley calibrations. The specific PFR was originally calibrated at ‘Davos’, then re-calibration at ‘Mauna Loa’.

4.2 Comparison to reference radiometer

A filter radiometer can be inter-calibrated by comparison to a second, well calibrated radiometer, provided the two instruments match closely in their centre wavelength. In a simple comparison of a calibrated radiometer with known calibration $S_{0R}$ the unknown calibration coefficient $S_{0X}$ of a test radiometer can be determined from a number $N$ of simultaneous measurements $S_{iX}$ and $S_{iR}$ as

$$S_{0X} = S_{0R} \frac{1}{N} \sum_{i=1}^{N} \frac{S_{iX}}{S_{iR}}.$$  \hspace{1cm} (4.10)  

The assumption of matching wavelengths is easily verified by plotting the ratios versus time or optical air mass should show no curvature. This condition is readily fulfilled by all instruments deployed in the GAWPFR network as their interference filters were
manufactured at the same time. These instruments are thus routinely calibrated by this method against a set of master instruments maintained at WORCC.

For multi-channel instruments, the variability of Langley calibrations is often correlated among channels, indicating that atmospheric rather than instrumental effects are governing the day-to-day variations. When diurnal variations of Rayleigh scattering and Ozone absorption are properly accounted for by the refined Langley method, the atmospheric variability is dominated by aerosol extinction. Assuming a constant aerosol size spectrum, represented by the Ångström exponent $\alpha$, and just the AOD expressed as Ångström coefficient $\beta$ to be variable, the AOD spectrum may be represented by a time dependent form of Ångström's law

$$\delta_A(\lambda, t) = \beta(t) \lambda^{-\alpha}$$ (4.11)

Under this assumption, the optical depths at different wavelengths are strongly correlated and their diurnal variations lead to correlated variations of the Langley calibrations for different channels of a sun photometer. The General Method devised by Bruce Forgan (1994) takes advantage of this by performing the regression against aerosol optical thickness instead of optical air mass. For the determination of momentary aerosol thickness $\delta_A m_A$, a 'well calibrated' reference channel can be used.

### 4.3 Radiometric calibration

Historically, the Langley method of atmospheric extrapolation was used to determine the extraterrestrial solar spectrum with a calibrated spectroradiometer at wavelengths where the atmospheric extinction can be described by the Bouger-Lambert-Beer law. By reversing this method, the extraterrestrial value of a spectral radiometer can be determined by a radiometric calibration of its response and relating it to a solar irradiance spectrum. Such a laboratory calibration of sun photometers can eliminate expensive campaigns at a high altitude station and avoids the dependence on weather conditions. Absolute calibrations are traceable to SI units and can be provided by national standard institutes or other laboratories equipped for optical metrology. Absolute calibration of a spectral radiometer can be performed either by measurements of a calibrated source, such as a spectral irradiance standard lamp, or by comparison with a detector based representation of the spectral irradiance scale.

A radiance spectrum of the solar disk centre was taken by Labs and Neckel (1968) at the Sphinx high altitude research station on Jungfraujoch using a spectrometer that was carefully calibrated against a black body standard in Heidelberg and tungsten ribbon lamps as transfer standards. In a later publication (Neckel and Labs, 1984) they converted the original radiance measurements to irradiance data by additional measurements of the solar limb darkening using the original spectrometer. The resulting NL84 spectrum has become one of the most widely used work for two decades.

At the same period, Arvesen and co-authors (1969) have published a solar irradiance spectrum taken from a research aircraft flying at 12km and thus avoiding the larger part of water vapour absorption. Brasseur and Simon (1981) have measured the ultraviolet part of the solar spectrum from stratospheric balloon flights at 30 to 40km height and Smith and Gottlieb (1974) had prepared a synthetic infrared spectrum.
From the above mentioned spectra, a composite spectrum covering the range from 200nm to 10μm was constructed by Wehrli (1985) and scaled such that its integral equalled the solar constant of 1367W m⁻². This spectrum was later adopted by WMO as a standard and has found broad application, e.g. was included in the LOWTRAN7 code.

New reference spectra have since become available based on satellite measurements that overcome the limitations imposed on previous observations take within the Earth’s atmosphere. Between 1992 and 1994 Thuiller et al. (2003) have measured the solar spectrum between 200nm and 2400nm on three ATLAS space-shuttle flights and on the EURECA satellite mission. Based on these measurements, a new reference spectrum was proposed in 2004 (Thuiller et al., 2004). Over the spectral range (350 – 1050nm) used for Sun photometry, this new spectrum and the old WMO spectrum agree to within ±2% to ±3% when they are degraded to 5nm resolution (Fröhlich and Wehrli, 2006).

Lamp Calibration

Lamp calibrations of filter radiometers were used in several rocket and stratospheric balloon experiments of PMOD for the determination of the solar constant and its variation. Three rocket flights in December 1983, ’84 and ’86 reached peak heights of about 300km and the balloon of the September 1985 experiment floated at 40km height. These in-situ measurements of the extraterrestrial values at three wavelengths of 368, 500 and 778nm (Wehrli and Fröhlich, 1991) were found to agree with their laboratory calibration to within 1.5%. These results were significantly better than the estimated uncertainty of about 3% based on an uncertainty of 2% for the lamp and an estimated uncertainty of 1% for the solar spectrum.

A comparison of the lamp calibration method with Langley extrapolations at Jungfraujoch was made in 1993 (Schmid and Wehrli, 1995). A sun photometer with six wavelengths between 500 and 1024nm was calibrated twice in the laboratory in December 1992 and December 1993, and by Langley calibrations on 11 carefully selected days at Jungfraujoch from September to November 1993.

From the relative differences (shown in Figure 10) it was concluded first that Langley calibrations at a mountain station can reach a precision of 0.25% with an estimated uncertainty smaller than 1%. Second, that while the lamp calibration method has an estimated error of about 3%, the actual error seems to be only about half of that. It was further concluded that a long term stability of better than 1% can be achieved even for sun photometers with degradations rates of several percents per year. Schmid and Wehrli thus recommended that lamp calibrations should mainly be employed to monitor the stability of sun photometers that were calibrated more accurately by Langley extrapolations.
CHAPTER 4  Calibration of sun photometers

In a follow-up publication, Schmid and co-authors (Schmid et al., 1998) extended the comparison between Langley extrapolation and lamp method to 13 channels between 313 and 1024nm and incorporated calibrations by additional lamps and three different solar spectra. Their results largely confirmed those of the previous study and led to the additional conclusions, that the lamp method offers a possibility to calibrate sun photometers in the UV-A spectral range where the Langley method failed, and that the choice of standard solar spectrum has a profound effect on the spectral distribution of the differences to Langley calibrations.

**Trap calibration**

Standard lamps have the advantage that their irradiance is just about 5 to 100 times lower than the Sun, but still within the normal measuring range of a filter radiometer. However, their absolute uncertainty is limited to about 2% and their long term stability is subject to sudden jumps. Lamp based calibrations require at least a triad of lamps that need recalibration against a primary standard every 50 to 100 hours of operation.

Radiometric standards based on quantum detectors were developed in the 70ties (Geist, 1979) that have smaller uncertainties in the order of 0.1%, and are more stable and robust than lamps. Several silicon diodes can be geometrically arranged in a so called “trap” configuration that has a quantum efficiency of nearly 100% (Fox, 1991). The absolute response of a trap detector can be expressed as a function of wavelength

$$R(\lambda) = \frac{\lambda}{hc} \approx \frac{\lambda}{1239.5} \frac{A}{W}.$$  \hspace{1cm} (4.12)

Actual trap detectors do not quite reach quantum efficiencies of 1 due to physical limitations of the silicon diodes. They can however be related to primary standards such as a cryogenic radiometer and become accurate and stable secondary standards.
A trap detector was built from three large area silicon diodes (Hamamatsu S3411, 18x18mm²) and calibrated at 12 laser wavelengths between 350 and 900 nm against the cryogenic radiometer at the Physikalisch-Technische Bundesanstalt (PTB) in Berlin. Figure 11 shows the spectral response of a single silicon diode and theoretical response of the trap detector on the left panel, the right panel shows its calibrated response plotted as quantum efficiency. The calibration points are interpolated at any wavelength between the end points by cubic splines. The uncertainty is estimated at 0.1% for wavelengths above 400nm and 0.25% at 350nm. This may not necessarily represent the absolute accuracy, as Köhler and co-authors (Köhler et al., 1996) compared 18 detector standards in an international comparison of spectral scales and found the agreement between different trap detectors to be ±0.5% in the visible and ±1% in the near UV and near infrared.

This trap detector is used to calibrate the absolute response of PFR in a spectral flux comparator facility at WORCC, similar to the method employed by Friedrich et al. (1995) at PTB as schematically shown in Fig. 12. The responses of the PFR and the trap detector to monochromatic light are measured alternatively during a wavelength scan over the filter bandpass of the PFR. The spectral response of the PFR can then be expressed in absolute units of Volts per Watt. Further details about the optical setup and calibration method can be found in Wehrli (2000).
CHAPTER 4  Calibration of sun photometers

The integral of the absolute response represents a convenient measure of the radiometric sensitivity of the PFR that can be performed any time in the laboratory. The reproducibility was determined by repeated setups and alignments of the spectral comparator facility and is estimated at about 0.33%. Figure 13 depicts the results of roughly semi-annual trap calibrations of the PFR instrument that was used in the stratospheric balloon experiment.

Figure 13 Relative variation of the absolute response of a PFR instrument. Individual points are plotted as ratio to their average value, the colors indicate spectral channels at 862nm (squares), 500nm (triangles), 412nm (diamonds) and 368nm (circles). Error bars indicate the estimated repeatability of ±0.33%.

From these trap calibrations, an annual drift for the 500nm channel of -0.08±0.03% and for the 412nm channel of -0.13±0.03% can be determined; no significant drifts can be
deduced for the channels at 862 and 368nm. These drift rates are approximately confirmed by routine Langley calibrations of PFR instruments at high altitude stations Mauna Loa or Jungfraujoch.

### 4.4 Stratospheric balloon flight

On October 22\textsuperscript{nd}, 1998 the PMOD/WRC stratospheric balloon experiment SIMBA98 was launched from Aire-sur-l’Adour in southern France and reached a ceiling altitude of 39.5km, where measurements of the solar total and spectral irradiance were taken during 70 minutes before the gondola was separated from the balloon. This experiment carried, among other instruments, a set of Precision Filter Radiometers (PFR) recently developed at WORCC, with the aim to measure quasi in-situ their extraterrestrial calibration coefficients. A more detailed description of the SIMBA98 experiment is given in Anklin et al. (1999).

The extinction of the residual atmosphere above the balloon was calculated with MODTRAN 3.7 code using the 1976 US standard atmosphere model in a modified version that provides finer altitude resolution above 30 km. Averaged transmission values were obtained that ranged between 0.9966 at 368nm to 0.9997 at 862nm. The corrected signals still showed systematic, negative trends of the order of a few parts in 10\textsuperscript{-3}, hinting at an accelerated ageing effect owing to the high ultraviolet flux at 40 km. The uncertainty of these in-situ calibration coefficients was estimated at 0.2% (Wehrli, 2000).

### 4.5 Summary of calibration methods

The PFR that was flown on the stratospheric balloon was also calibrated radiometrically by both the lamp and trap methods. It was also calibrated by comparison with another PFR that was calibrated by Langley extrapolations at Jungfraujoch. This gave the opportunity to compare all calibration methods presented in the sections above. Table 4 summarizes the results of these calibrations made in 1998 and 1999 related to the stratospheric balloon flight in October 1998.

<table>
<thead>
<tr>
<th>Method</th>
<th>862 nm</th>
<th>500nm</th>
<th>412 nm</th>
<th>368 nm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Langley calibrations</td>
<td>1.007</td>
<td>1.001</td>
<td>1.007</td>
<td>1.008</td>
</tr>
<tr>
<td>Lamp calibration</td>
<td>1.053</td>
<td>1.028</td>
<td>1.032</td>
<td>1.068</td>
</tr>
<tr>
<td>Trap calibration</td>
<td>1.003</td>
<td>0.971</td>
<td>0.977</td>
<td>1.023</td>
</tr>
</tbody>
</table>

Langley calibrations of a PFR at Jungfraujoch (3580m) were made on 16 clear days between December 1998 and February 1999. A side-by-side comparison between the Jungfraujoch and the balloon instruments gave an agreement within +0.8%. This difference is about twice as large as the combined (single) standard deviation for Langley calibrations of 0.25% (Schmid and Wehrli, 1995) and the estimated uncertainty of 0.2% for the balloon measurements (Wehrli, 2000). Possible explanations for this difference are that either the Langley calibrations were overestimated due to small systematic diurnal variations of AOD, or that the stratospheric transmission is overestimated by MODTRAN due to some trace gases missing in the model atmosphere.

Both laboratory calibrations differ by several percent from the balloon measurements, with a general overestimation of the calibration coefficients by the lamp and an.
underestimation by the trap method. Differences between lamp and trap detector standards used at WORCC are in the order of 5%, independent of wavelength. The consistent over-estimation obtained by the lamp calibration may hint at an ageing effect of the lamp standards, though that would lead to a wavelength dependent error. The overall result of the trap calibration of -0.6±3% may indicate a wavelength dependent error in the WMO solar spectrum. This suspicion is supported by a recent comparison (Fröhlich and Wehrli, 2006) of the WMO spectrum with the RSSV1 spectrum from the ATLAS 1 and 3 missions (Thuillier et al., 2004). An underestimation of the solar spectral irradiance of 1 to 2% below 500nm, and an overestimation of about 2% at 850nm were found for the WMO spectrum. When a corresponding correction is applied to the laboratory calibrations, the spectral errors of the trap calibration above 400nm become constant at about -1.5%, but the error at 368nm rises to +4%. The discrepancy between lamp and trap calibration remains unresolved by a simple change of the solar spectrum.

From this comparison of calibration methods it is concluded, that the Langley extrapolation gives significantly better results than the laboratory methods. This conclusion is supported by several intercomparisons of sun photometers (see chapter 6.3) indicating a consistency of the Langley method in the order of 1% between independent calibrations. The radiometric calibrations at WORCC suffer from a systematic difference between lamp and trap based radiometric standards used. Even if this discrepancy is eventually resolved, the uncertainty of this method will still be limited by differences and uncertainties of available solar reference spectra in the order of 2% or more.
Chapter 5  Design of the PFR instrument

The design of sun photometers is of utmost simplicity and has barely changed since the development of the hand-held “Volz” type instrument around 1960. They basically consist of a collimator that defines a narrow field of view, a well blocked narrow band interference filter of suitable wavelength, and a semiconductor detector – amplifier system to convert the optical to an electrical signal that is fed to a readout system. Modern semiconductor components permitted the construction of sophisticated filter radiometers for meteorological applications that incorporate digital voltmeters or micro processors for on-line data evaluation even in hand held units.

When WMO recommended abandoning manual turbidity measurements, further instrument development focused on automation, weather hardening and adding Sun tracking equipment. Two scientifically interesting extensions of the sun photometer principle took place around 1990. One was the sky scanning radiometer that complements the extinction measurements by simultaneous measurements of aureole and sky radiances. Sky radiance measurements in the solar vertical or almucantar contain information about the aerosol phase function that can be used to retrieve extended microphysical (size distribution) and additional optical properties (single scattering albedo, asymmetry parameter) of aerosols through inversion techniques (Nakajima et al., 1983; Dubovik and King, 2000). The Sky scanning radiometer CIMEL CE 318 has found wide use in the global AERONET network (Holben et al., 1998) coordinated by NASA.

The second new development was the rotating shadow band radiometer that measures hemispherical global and diffuse spectral irradiances quasi simultaneously by temporary occultation of the Sun with a rotating shadow band (Harrison et al., 1994). By reconstruction of the direct component this radiometer can be used as a classic sun photometer without the need for a two axis Sun tracker. Additional information like the single scattering albedo that are contained in the direct to diffuse ratio can be retrieved (Kassianov et al., 2007) under clear sky conditions which are difficult to assert based on hemispherical measurements alone. This type of instrument has also found widespread use in several national networks within the US, as well as by independent research groups.

The technological progress is reflected in the price of a typical sun photometer: a “Volz” type instrument was offered around 1970 for about 100$, advanced hand held multi wavelength units as deployed by WMO in the BAPMoN network around 1980 were priced at 1000$ and the automated systems in use since about 1990 cost 10'000$ without trackers. At least for ground based application, there seems to be neither need nor market for more complex instruments like scanning spectroradiometers.

In the mid 1990ies, these new developments were considered not yet mature enough for deployment at GAW stations, and an ad hoc group of experts recommended the construction of a classic multi-wavelength sun photometer. A new Precision Filter Radiometer (PFR) was thus designed with 4 independent channels in a single, compact tube. Figure 14 shows a PFR mounted on a commercial solar tracker at the GAW Global observatory at Danum Valley in Malaysia. General specifications of the PFR are listed in Table 5. The custom mounting flange pictured in Figure 14 will accept a second PFR for on-site calibration by a traveling standard.
As improved instrument stability was a major design goal, special attention was given to minimize filter degradation. The air tight tube can maintain an internal atmosphere of dry nitrogen gas over extended periods, sunlight enter through a quartz front window that can be readily cleaned as part of the daily routine maintenance. An active Peltier controller maintains the core detector system at 20°C over an ambient temperature range from -20°C to +35°C. An internal shutter limits filter exposure to brief moments of actual measurements.

Figure 14 A PFR instrument mounted on its Sun tracker on the roof of the GAW station in Danum Valley, Malaysia. The larger apertures behind the front window correspond to the 4 independent photometric channels. Behind the smaller, central aperture an electronic sensor is used to monitor the pointing of the PFR.

Measurements of the PFR are performed by a commercial data logger (Campbell Scientific CR10X) built into a separate electronic box hosting also the power supply, serial data interface and a barometric sensor. Details of the PFR design are described in the following sections.
### Table 5 General optical and mechanical specifications of the PFR instrument.

<table>
<thead>
<tr>
<th></th>
<th>Ch1</th>
<th>Ch2</th>
<th>Ch3</th>
<th>Ch4</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Center wavelength</td>
<td>861.6</td>
<td>500.5</td>
<td>411.4</td>
<td>367.6</td>
<td>nm</td>
</tr>
<tr>
<td>FWHM bandwidth</td>
<td>5.5</td>
<td>5.0</td>
<td>4.5</td>
<td>3.8</td>
<td>nm</td>
</tr>
<tr>
<td>Field of view</td>
<td>±1.25</td>
<td>degree</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope angle</td>
<td>0.72</td>
<td>degree</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dimensions</td>
<td>Ø 89 x 390</td>
<td>mm</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mass</td>
<td>3</td>
<td>kg</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Power consumption</td>
<td>&lt;20</td>
<td>W</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### 5.1 Wavelength selection and number of channels

A set of 5 standard wavelengths were recommended by WMO for use in their BAPMoN turbidity network. For historical reasons, a channel at 500nm was mandatory and formed together with channels at 368 and 778nm the primary set of wavelengths; two additional wavelengths at 675 and 862nm were recommended as optional channels. In order to facilitate comparison of results and calibration transfers, optical filters or dispersive elements should localize these wavelengths within ±2nm and have a full width half mean (FWHM) bandwidth of at most 5nm.

Figure 15 shows the position of these wavelengths in relation to atmospheric gas absorption (H\(_2\)O, O\(_2\), O\(_3\), and NO\(_2\), Rayleigh scattering and Ångström aerosol extinction spectra. The assumed concentrations are typical values taken from model atmospheres (column ozone 350 Dobson units, water vapor 1.5 atm-cm, nitrogen dioxide 0.001 atm-cm, sea-level pressure for Rayleigh scattering and oxygen absorption, and a global average optical depth of 0.125 at 500 nm for aerosol extinction with an Ångström exponent of 1.3. From this plot, it becomes obvious that just wavelengths between 850 and 880 nm are
practically free of gaseous absorption, while the broad absorption bands of ozone and nitrogen dioxide contaminate all wavelengths shorter than 850 nm.

Because the aerosol extinction spectrum is determined by a heterodisperse mixture of particles, it shows rarely much wavelength structure and a small number of 3 to 6 channels are sufficient to characterize the extinction spectrum by a power law.

A set of 4 WMO standard aerosol wavelengths at 368, 412, 500 and 862nm was selected for the PFR instrument. It was assumed, that total ozone content required for correction of measurements at 500nm would be provided by the GAW stations in the context of their observational program.

### 5.2 Interference filters

The stability of interference filters, which hindered the successful use of sun photometers for decades, did not quite follow the technological progress in detectors and associated electronics over the last 20 years. Annual degradation rates in the order of 5 to 10% were common in the early 1990ies and are still reported today. (Wehrli and Fröhlich, 1991; Holben et al., 1998; Alexandrov et al., 2002)

The filter stability is difficult to assess as it depends strongly on filter design, selection of optical materials and manufacturing process, all of which are proprietary of commercial companies and performance reports are seldom published. Dielectric coatings are hygroscopic and can change their characteristics when exposed to elevated humidity even if they are sealed in metallic mounting rings. They are mechanically soft and may delaminate from their glass substrate under thermal stress. Broadband blocking is achieved by classic colored filters, which are optically bonded to the interferences coatings. Both the glass filters and bonding materials can change in transmission with age or under exposure to shortwave radiation. In the mid nineties, several manufacturers introduced ion assisted deposition (IAD) techniques by which the dielectric materials are compacted by a beam of heavy ions during the deposition process, thus rendering the coating less susceptible to structural change.

All PFR instruments in the GAW network are equipped with matching IAD filters from the same manufacturing batch, and thus well suited for side by side comparison. The filters are placed immediately behind the radiometric aperture; stray light from the aperture will not be reflected off the glass. The silicon detectors are slightly tilted off the optical axis in order to prevent inter-reflections with the backside of the filter that may afflict measurements.

**Filter response measurements**

All filters were individually measured by the manufacturer and their central wavelengths are within 0.7nm from their nominal values.

The spectral pass bands for all PFR instruments for the GAW network were characterized for effective central wavelength and bandwidth, determined as average wavelength weighted by the spectral response and equivalent width of a rectangular bandpass with equal throughput as the filter. These measurements were made with the PFR illuminated by a grating monochromator (Jobin Yvon HR640) with 0.6nm spectral resolution in steps of 0.5nm; the PFR response is related to the optical power measured by a calibrated silicon detector. Figure 16 shows typical filter response curves on a logarithmic scale with median wavelength and equivalent bandwidth. Table 6 lists the central
wavelengths and FWHM bandwidth, showing the close spectral matching of the PFR instruments.

Table 6 Results of spectral characterization central wavelength / bandwidth of a number of PFR instruments.

<table>
<thead>
<tr>
<th>PFR</th>
<th>Red</th>
<th>Green</th>
<th>Blue</th>
<th>Violet</th>
</tr>
</thead>
<tbody>
<tr>
<td>N01</td>
<td>862.7 / 5.5</td>
<td>500.6 / 5.1</td>
<td>411.9 / 4.3</td>
<td>367.6 / 3.7</td>
</tr>
<tr>
<td>N05</td>
<td>862.7 / 5.6</td>
<td>500.7 / 5.1</td>
<td>412.6 / 4.4</td>
<td>368.2 / 3.7</td>
</tr>
<tr>
<td>N09</td>
<td>863.0 / 5.6</td>
<td>500.9 / 5.1</td>
<td>411.9 / 4.4</td>
<td>367.6 / 3.7</td>
</tr>
<tr>
<td>N14</td>
<td>861.3 / 5.6</td>
<td>500.9 / 5.1</td>
<td>411.8 / 4.4</td>
<td>367.6 / 3.7</td>
</tr>
<tr>
<td>N19</td>
<td>862.7 / 5.6</td>
<td>501.5 / 5.1</td>
<td>412.5 / 4.4</td>
<td>368.3 / 3.5</td>
</tr>
<tr>
<td>N27</td>
<td>862.4 / 5.6</td>
<td>501.3 / 5.1</td>
<td>412.1 / 4.4</td>
<td>367.7 / 3.6</td>
</tr>
</tbody>
</table>

Figure 16 Response curves for the 4 channels of a typical PFR instrument covering a range of about 5 decades.

**Out of band rejection**

Spectral leakage of light outside the nominal passband can cause significant errors in Sun photometry because of the variation in the terrestrial solar spectrum with solar zenith angle would cause artificial curvature of Langley plots. In order to keep the calibration error due to spectral leakage below 0.1%, the interference filters of the PFR should be blocked to $10^{-4}$ (862nm) to $10^{-7}$ (368nm) from 300 to 1100 nm depending on their wavelength and bandwidth.

The manufacturer had designed the filters for a blocking of better than $10^{-6}$, and has provided measured transmission data showing the absence of any significant leakage. Nevertheless, measurements of the out-of-band response of the prototype PFR at WORCC revealed unexpected spectral leakage in the channels at 500 and 862nm as shown...
in Figure 17. These leaks were neither reported by the manufacturer nor observed in our own transmission measurements of individual filters, but showed up only in the completed PFR instrument.

Detailed experiments with another series of filters have shown that a broad leakage in the visible range of the 862nm channel can be explained by fluorescence of colored blocking glass components used in the interference filters. Physical separation of filter and detector has reduced this leak by 2 orders of magnitude, and by a simple reversal of the filter a reduction by a factor of ten was achieved.

![Figure 17 Out-of-band rejection for a typical PFR instrument](image)

In order to estimate the errors introduced by the spectral leaks observed in the two long wave channels of the PFR, two Langley extrapolations using simulated data were compared. MODTRAN4 transmission spectra of the US standard 1976 atmosphere were calculated at 6 solar zenith angles. PFR measurements were simulated by weighting these spectra with measured filter transmissions, once including spectral leaks, and once using just the bandpass range of central wavelength ±10nm. In the longwave channels, signals were increased by about 1% due to spectral leaks, in the shortwave channels, the increase was less than 0.1%.

Extraterrestrial transmissions (nominally 1.00) were then determined by Langley extrapolations from these two datasets and are compared in Table 7. The spectral leaks lead to a slight underestimation of the nominal transmission in the order of a few $10^{-4}$, roughly doubling the error of the extrapolation method. For the shortwave channels, no difference was found at the $10^{-4}$ level. It may be concluded that the measurement error due to spectral leakage is eliminated by the calibration process or remains negligibly small.
Table 7 Extraterrestrial transmission determined by Langley extrapolation of synthetic measurements; 'contaminated' including spectral leaks; ‘pure’ using just filter bandpass.

<table>
<thead>
<tr>
<th>Channel</th>
<th>862nm</th>
<th>500nm</th>
<th>412nm</th>
<th>368nm</th>
</tr>
</thead>
<tbody>
<tr>
<td>dS</td>
<td>0.0082</td>
<td>0.0102</td>
<td>0.0007</td>
<td>0.0005</td>
</tr>
<tr>
<td>Contaminated</td>
<td>0.9995</td>
<td>0.9990</td>
<td>0.9991</td>
<td>0.9991</td>
</tr>
<tr>
<td>Pure</td>
<td>0.9998</td>
<td>0.9996</td>
<td>0.9991</td>
<td>0.9991</td>
</tr>
</tbody>
</table>

5.3 View geometry and circumsolar stray light

The field of view (FOV) of a sun photometer has to be large enough to accept the full solar disk, and as small as possible to minimize the amount of diffuse sky light in the measurements, but a narrow FOV also requires high pointing accuracy that can be difficult to maintain in an operational environment. The current recommendation of WMO calls for a full opening angle of 2.5° and a slope angles of 1°.

The viewing geometry of the PFR is defined by 2 circular apertures of 7 and 3 mm diameter for each channel that are separated by 160mm and define a FOV of ±1.25° and a slope angle of 0.72°. With this viewing geometry, the stray light correction factor (see chapter Sunphotometry) for the PFR becomes to ≈1.01 or 1% of the measured AOD at all 4 wavelengths when an Ångström exponent of $\alpha=1.35$ is assumed. This correction is not routinely applied, but should be considered when results from filter radiometers with different viewing geometries are compared.

5.4 Pointing monitor

In order to ensure that the full solar disk remains in the FOV, a sun tracking equipment is required to point the PFR within the effective FOV=(slope angle – solar radius) ≈ 0.45° or 27 arcminutes of the Sun’s center. A quick geometric calculation shows that an error in optical depth of 0.01 results from a vignetting of just 6% or 1 arc minute.

![Figure 18 4-Quadrant detector pixels with pinhole image of the Sun (central dark spot with grey circle corresponding to penumbra and diffraction) seen from front side of the PFR. Formula to calculate pointing directions left-right and up-down, indices corresponding to pixel numbers.](image-url)
With some care, a PFR can readily be aligned to the Sun within the required accuracy and a properly set sun tracker should maintain pointing stable to a few arcminutes. Nevertheless, it was deemed useful to include a solar pointing monitor in the PFR in order to get a remote diagnostic tool to test the solar tracking and flag mispointed measurements. This decision later proved to be a valuable measure for quality control.

The pointing monitor consists of a 4 quadrant silicon detector that is illuminated through a pinhole of 1 mm diameter at a distance of 70 mm. Due to finite angular diameter of the Sun and a small diffraction effect, the actual spot diameter is about 1.8 mm, matching neatly the 1.5 mm square pixel size of the detector. When the light spot is centered on the detector, all 4 pixels will produce equal signals. By subtracting signals from paired pixels, the spot of sunlight can be localized on the 4 quadrant detector by a smooth function that is almost linear near the point of symmetry and asymptotically approaching ±1 when the spot lies completely on either half-side of the detector.

These pointing monitor devices were individually characterized to determine their sensitivity \( k \) and offset constants \( LR_0 \) and \( UD_0 \). To that purpose, the PFR is mounted on a parallactic Sun tracker with one of the pointing monitor’s directions in parallel to the hour angle axis. While the tracker is stopped a little in advance of the Sun position, measurements are taken at 10 seconds intervals, letting the Sun pass through the FOV of the instrument. From these measurements, the centre of the PFR FOV, the pointing offset and the angular sensitivity of the monitor can be determined graphically. A second set of measurements, with the PFR rotated by 90°, gives the offset for the complementary direction.

![Figure 19 Calibration of a pointing monitor. The trapezoid curve shows the radiometer signal versus Sun angle, the step curve is the pointing monitor up-down signal. The offset is close to zero, and the radiometer slope angle ±0.66°.](image)

Figure 20 shows monthly overviews of the pointing accuracy at two different GAW sites: the left panel shows a station where the tracker was working stable, and a small misalignment was corrected by the station operator. At the station shown on the right panel, the tracker was not working well and the PFR was slightly misaligned. A significant
number of measurements were thus flagged for pointing errors by the PFR quality control algorithm.

![Figure 20 Two examples of monthly pointing checks. Left side: after mechanical adjustment the PFR is well aligned on the solar disk, and tracker stability is good. Right side: overall alignment is OK, but an unstable performance of active tracker and mechanical mounting are clearly visible.]

### 5.5 Design for radiometric stability

Special considerations were made in order to achieve a high instrumental stability of the new precision filter radiometer. These design measures are based on the long experience at PMOD/WRC with earlier sun photometers in terrestrial and space applications. The basic concept of keeping the delicate interference filters in a dry and dark environment at constant temperature was implemented by the following design elements.

**Internal shutter**

Ageing or degradation of filter radiometers is exposure or dose dependent, at least for the shorter wavelengths (Wehrli et al., 1995). The PFR is equipped with a mechanical shutter that blocks the Sun light while no measurements are taken. The rate of exposure is thereby reduced to 1.6% with respect to permanent exposure. The shutter can also be used to verify photometric dark signals, although these are easily found from night time measurements.

**Temperature Stabilization**

Both the detectors and the optical filters of spectral radiometers usually show some temperature dependence of their characteristics. The temperature coefficient of the quantum efficiency for semiconductor detectors is a function of wavelength that remains small for photon energies larger than the band-gap energy, but increases steeply as this limit is approached. In case of silicon with a band-gap of 1.14eV, the temperature coefficient becomes significant only for wavelengths above 1 μm. Typical temperature coefficients of the photodiodes used in the PFR are about $10^{-2}/°C$ at 1100 nm, well below $+10^{-3}/°C$ at
wavelengths between 950nm and 350nm and about -10^{-3}/°C below 350nm. Narrow band interference filters show only minor, reversible shifts in central wavelength with temperature, typically of the order of 0.02 nm/°C and insignificant changes in transmission.

For a filter radiometer that is operated outdoors and may experience changes in ambient temperature of several tens of degrees, a temperature correction is still required for precision measurements. For best results, thermostatically controlled detectors are the preferred choice over empirical corrections based on measured temperature. A simple heating system that maintained a constant temperature above the expected ambient temperature maximum was used in an older design (Wehrli and Fröhlich, 1991), but because of accelerated filter aging at elevated temperatures, a thermostatic system with cooling capability was chosen for the new design. Thus the PFR incorporates an active Peltier system that maintains its filters and detectors at a constant temperature of 20°C over an ambient temperature range from -20° to 35°C.

The Peltier element has been the only point of failure for the PFR, so far. It is susceptible to fracture upon mechanical shock which may occur during shipment of the instrument. A kind of fatigue or gradually degrading heating and cooling power was also observed. Any malfunction of the Peltier system does not directly hamper the photometric measurements, but may require the application of a temperature correction. Sensor temperature and Peltier system performance are constantly monitored by housekeeping measurements.

Temperature coefficients were determined by simultaneous measurements with two PFRs, one had its temperature set at 10, 15, 20 and 25°C, and the second was operated at the reference temperature of 20°C. Their signal ratio, shown in Figure 21, is independent of the solar irradiance, and can be used to derive typical temperature coefficients listed in Table 8.

<table>
<thead>
<tr>
<th>TC</th>
<th>R862</th>
<th>R500</th>
<th>R412</th>
<th>R368</th>
</tr>
</thead>
<tbody>
<tr>
<td>slope/°C</td>
<td>0.97E-04</td>
<td>-6.85E-04</td>
<td>-2.56E-04</td>
<td>-2.12E-04</td>
</tr>
<tr>
<td>std. Error</td>
<td>1.31E-05</td>
<td>1.45E-05</td>
<td>1.30E-05</td>
<td>2.55E-05</td>
</tr>
<tr>
<td>Correlation coef.</td>
<td>0.1334</td>
<td>0.8619</td>
<td>0.5168</td>
<td>0.1614</td>
</tr>
<tr>
<td>ETV/°C</td>
<td>3.3E-05</td>
<td>-5.4E-04</td>
<td>1.0E-06</td>
<td>-2.5E-04</td>
</tr>
</tbody>
</table>

The sign of the temperature coefficient corresponds to what is expected from the coefficients of the silicon detector which are positive in the infrared, negligible in the visible and slightly negative in the ultraviolet. The somewhat stronger temperature dependence at 500nm can be attributed to the variation of the solar spectrum when a wavelength shift of 0.02nm/°C is assumed for the interference filter. These temperature coefficients are applied to measurement where data quality control indicates sensor temperature out of limits.
Humidity control

Problems with the stability of interference filters due to adsorption of humidity have traditionally been minimized by a sealed housing and use of a desiccant cartridge. The PFR is mounted in a vacuum tight tube that is permanently filled with dry nitrogen gas at slight overpressure. Inspection of instruments returning to Davos for recalibration has shown that this overpressure is maintained over several years.

The combination of above design features proved to be very successful as experience with more than 30 instruments over several years has shown. Annual drift rates of calibration coefficients are below 1%, which represents a ten-fold improvement over the stability of previous instruments. This high stability relieves to a certain extent the need for frequent calibration that hindered the successful use of sun photometers in many places.

![Figure 21 Temperature dependencies of PFR irradiance signals determined from simultaneous measurements by two radiometers operating at different temperature. The slopes of the linear regression lines are estimates for the temperature coefficients.](image)

5.6 Data logging

The four photometric channels are measured simultaneously by a commercial data logger system (Campbell Scientific CR10X) with 13 bit resolution. Automatic signal ranging within the PFR is used to increase the dynamic range to 16 bits. A complementary set of housekeeping data (temperatures, currents, ambient pressure, etc.) is measured together with the photometric signals in order to monitor proper functioning of the PFR system. These housekeeping data include simultaneous readings of the barometric pressure for an accurate determination of the Rayleigh correction.

Measurements are taken every minute and stored in the data logger memory, which has a capacity of about 3 weeks of continuous operation. The CR10 logger communicates with software hosted on a Windows PC for data downloading or real time monitoring. Once
configured and operating, it will recover from a power failure without loss of data and continue to measure automatically.

Additional information on technical details and operation of the PFR system can be found in the User Manual that is available under ftp://ftp.pmodwrc.ch/pub/worcc/.

### 5.7 Summary

PFR instruments are operated in a global network of currently 10 stations since up to 10 years. In more than 700 months of cumulative operation in a wide range of climates, they have proven to be very reliable and stable sun photometers. Several national meteorological services in Europe and Japan have started new or have renewed existing programs for AOD observations with PFR instruments. Due to continuing demand, PMOD has built and sold many additional PFR, so that now a total of well over 50 instruments are operated world-wide.

In the frame of a classic sun photometer there is little room for technical improvement of the PFR. User requests for additional or different wavelengths were fulfilled by special versions including water vapor or UV channels. Addition of scientifically relevant new capabilities like Aureol or sky radiance measurements would need a complete redesign of the instrument, including the addition of a dedicated, steerable Sun tracker. Such systems are already available commercially and successfully used in other networks like AERONET and SKYNET.
Chapter 6  Global PFR network

A key task of WORCC was the implementations of a global trial network at 12 selected GAW stations with the objective to 'demonstrate that the new generation of instruments together with new calibration techniques and quality assurance procedures is indeed able to determine AOD with a precision adequate for the fulfillment of the objectives of GAW' (WMO, GAW143). The realization of this task was made possible by collaboration, arranged through intermediation of the WMO secretariat, between the participating GAW observatories, which had to provide a sun tracking facility and limited amount of manpower for routine maintenance, and Meteo Swiss which sponsored the manufacturing of the PFR systems by PMOD.

The following paragraphs describe the global and temporal extent, operational procedures and dataflow of the PFR network.

6.1 Global and temporal extent

After successful tests of a prototype PFR at Jungfraujoch in 1998, the first units were shipped in 1999 to the GAW global observatories at Hohenpeissenberg (Germany), Mauna Loa (Hawaii) and Mace Head (Ireland). In the following years more stations could be added until in 2007, the PFR network included 10 stations, shown in Figure 22 and listed in Table 9, located from the tropics to the polar region and from sea level to above 3500 m.

![Figure 22 Map of the 10 GAWPFR network in 2007. Station names in italic are not yet operational.](image)

Most stations are located in pristine areas where AOD values are indicative of background levels. The mountain stations at Izaña, Jungfraujoch, Mauna Loa and Mount
Waliguan are situated well above most of the tropospheric aerosols and have the advantage of providing excellent opportunity for Langley calibrations.

Six sites are co-located with other AOD networks. The GAW network provides thus a path to link different national networks through intercomparison with PFR instruments that are traceable to WORCC.

Lack of sun trackers was, and remains, a major obstacle to overcome at some of the global stations. No partnership could be established in South America, and just two stations are located in the southern hemisphere. While the extent of the GAW network still remains below its goal of 12 stations, more than 36 additional PFR systems are operated by national weather services in Europe and Japan.

Table 9 Identification, localization and start of measurements date for 12 GAW stations in PFR network. Note: acronyms in rightmost column will be used for further referencing stations.

<table>
<thead>
<tr>
<th>Station, Country code</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Time Zone</th>
<th>Elevation</th>
<th>Start date</th>
<th>Acronym</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alice Springs, AU</td>
<td>23.80° S</td>
<td>133.87° E</td>
<td>+9.5</td>
<td>547 m</td>
<td>2002/04</td>
<td>ASP</td>
</tr>
<tr>
<td>Bratt’s Lake, CA</td>
<td>50.28° N</td>
<td>104.70° W</td>
<td>-6</td>
<td>586 m</td>
<td>2001/05</td>
<td>BRA</td>
</tr>
<tr>
<td>Cape Point, ZA</td>
<td>34.35° S</td>
<td>18.49° E</td>
<td>+2</td>
<td>230 m</td>
<td>pending</td>
<td>CPT</td>
</tr>
<tr>
<td>Danum Valey, MY</td>
<td>4.98° N</td>
<td>117.84° E</td>
<td>+8</td>
<td>426 m</td>
<td>2007/05</td>
<td>DMV</td>
</tr>
<tr>
<td>Hohenpeissenberg, DE</td>
<td>47.48° N</td>
<td>11.01° E</td>
<td>+1</td>
<td>995 m</td>
<td>1999/08</td>
<td>HPB</td>
</tr>
<tr>
<td>Izaña, ES</td>
<td>28.30° N</td>
<td>16.50° W</td>
<td>-1</td>
<td>2307 m</td>
<td>2001/08</td>
<td>IZO</td>
</tr>
<tr>
<td>Jungfraujoch, CH</td>
<td>46.55° N</td>
<td>7.99° E</td>
<td>+1</td>
<td>3580 m</td>
<td>1999/01</td>
<td>JFJ</td>
</tr>
<tr>
<td>Mace Head, IE</td>
<td>53.33° N</td>
<td>9.90° W</td>
<td>0</td>
<td>10 m</td>
<td>1999/07</td>
<td>MHD</td>
</tr>
<tr>
<td>Mauna Loa, US</td>
<td>19.50° N</td>
<td>155.57° W</td>
<td>-10</td>
<td>3397 m</td>
<td>1999/11</td>
<td>MLO</td>
</tr>
<tr>
<td>Ny Alesund, NO</td>
<td>78.91° N</td>
<td>11.88° E</td>
<td>+1</td>
<td>50 m</td>
<td>2002/05</td>
<td>NYA</td>
</tr>
<tr>
<td>Ryori, JP</td>
<td>39.03° N</td>
<td>141.83° E</td>
<td>+9</td>
<td>230 m</td>
<td>2002/04</td>
<td>RYO</td>
</tr>
<tr>
<td>Mt. Waliguan, CN</td>
<td>26.28° N</td>
<td>100.90° E</td>
<td>+8</td>
<td>3810 m</td>
<td>pending</td>
<td>WLG</td>
</tr>
</tbody>
</table>

Table 10 Temporal coverage of network stations in data months per year. At NYA solar measurement are possible only during polar summer (21.Feb. – 21.Oct.). At JFJ, the solar dome was out of operation from August 2006 until October 2007 due to remodeling of the station.

<table>
<thead>
<tr>
<th>Station</th>
<th>1999</th>
<th>2000</th>
<th>2001</th>
<th>2002</th>
<th>2003</th>
<th>2004</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASP</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>BRA</td>
<td>8</td>
<td>12</td>
<td>12</td>
<td>10</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>HPB</td>
<td>5</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>IZO</td>
<td>7</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>JFJ</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>MHD</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>MLO</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>NYA</td>
<td>5</td>
<td>7</td>
<td>6</td>
<td>7</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>RYO</td>
<td>9</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>data months</td>
<td>707</td>
<td>21</td>
<td>43</td>
<td>63</td>
<td>94</td>
<td>100</td>
<td>100</td>
<td>99</td>
<td>97</td>
</tr>
</tbody>
</table>

The temporal extent of measurements is shown in Table 10 as number of annual data months collected until 2007. The station NYA covers less than 12 months per year because
it lies above the polar circle; the annual number of data points however is comparable to other stations.

In 9 years of operation, the GAWPFR network has generated more than 700 consecutive months of data with few gaps, some due to tracker or power failures, and others due to removal of the PFR for off-site calibration or repair. Good temporal coverage is essential for detecting trends in long-term data sets with large short term variability.

### 6.2 Network Operation

Measurements of solar spectral irradiance are taken by the PFR continuously at 1 minute intervals, which seems a rather high sampling rate considering that just three daily observations were prescribed under the BAPMoN protocol. These observations were performed with handheld instruments, where a trained operator was able to judge the sky conditions visually for cloud contamination and to verify the proper functioning of the instrument for each observation. With un-manned observations, some form of data screening has to be applied either by post processing at the archive site or else by sophisticated instrument software during data acquisition. The former approach offers the advantage that screening algorithms can be refined or adapted with growing experience; the latter is more efficient in terms of data transmission, but both are based on quasi-continuous measurements. For the PFR network the first solution was adopted, mainly because of the limited processing capabilities of the data logger, but also because of the added flexibility in data screening.

Figure 23 Schematic representation of the data flow in the GAWPFR AOD network.
The raw photometric measurements and simultaneous readings of the atmospheric pressure and a number of instrument parameters are retrieved from the data logger storage in so called level-1 files. Depending on the facilities and resources available at each station, this retrieval may occur automatically or manually within one month since the last download.

A first data quality control is applied during the conversion of level-1 data to daily level-2 files which are the main data units in the network and are composed of a meta data header describing station and instrument parameters and a data body of normalized and quality controlled measurements. This level-2 file structure is flexible enough to accommodate data from different sun photometers with multiple wavelengths and each file represents a self-contained dataset that is suitable for further evaluation. Being flat ascii files, they are easily transported between different computers and can be stored and retrieved by their file name convention in any directory structure, e.g. are not restricted to a specific data base architecture or programming language.

Level-2 files are produced at the stations by standardized software which also performs the initial quality control (QC) tests that include cloud screening, monitoring of Sun pointing and instrument health. The results of these tests are represented in a binary coded QC flag for each measurement. Appropriate warning messages are issued if a technical malfunction is diagnosed by the program.

No measurements are discarded in this process, but the QC flags will be used to exclude invalid or doubtful measurements from further processing. In addition to these flags, the true (unrefracted) solar elevation angle is calculated and added to each data record.

This software is written in FORTRAN90 computer language which permits license free distribution to the stations and portability to all hardware platforms. Further technical details about the structure of data files and use of the FORTRAN software can be found in the PFR manual that is available from WORCC under ftp://ftp.pmodwrc.ch/pub/worcc

The level-1 and level-2 files are then either made accessible on an ftp-server at the PFR station home institution or else sent to WORCC by email.

In addition to the PFR raw data, the station should provide daily total ozone values whenever available for the correction of ozone absorption at 500nm. Some stations (BRA, HPB, and JFJ) provide this information from a local or nearby Dobson or Brewer spectroradiometer. A look-up table of TOMS monthly latitudinal means with a 5° spacing averaged over the years from 1989 to 1999 was implemented in the level-2 program and is used to estimate ozone correction at the other stations.

A comparison of this satellite climatology table with daily Dobson measurements at Arosa and Hohenpeissemberg, and daily Brewer values at Bratt’s Lake during several years showed that the monthly means of TOMS retrievals are about 25 Dobson units (DU) larger than the measurements at Arosa and Hohenpeissemberg, and 4 DU larger at Bratt’s Lake. We cannot decide whether this is an instrumental effect (according to WMO-GAW report 149, differences between Dobson and Brewer instruments are usually smaller than 5 DU) or due to the more homogenous topography at Bratt’s Lake that permits more accurate satellite retrievals than in the Alps. The monthly standard deviation of the differences is about 30DU at all three stations and would result in a 1σ uncertainty of the ozone correction of 0.001 optical depths, adequate for the purpose of correcting AOD at 500nm.
Level-2 data are evaluated to level-3 daily files at WORCC on a monthly basis. These level-3 files are organized similar to level-2 files with a meta data header and a data body containing AOD results and Ångström coefficients. Additional quality control flags are generated to indicate data records where physically impossible or unexpected results \( \text{AOD}<0, \text{AOD}(\lambda_1)<\text{AOD}(\lambda_2) \) where \( \lambda_1<\lambda_2 \), or Ångström \( \alpha <0 \) were detected, and finally the results are visually inspected for unusual signatures. Housekeeping data control includes monthly overviews of the instrument and tracking performance in order to detect technical problems or unusual situations and to resolve or clarify them in collaboration with station personnel. These monthly results are considered preliminary until a final quality assurance has been applied.

Quality assurance (QA) specifically includes a posterior calibration of the PFR and interpolation of calibrations and reprocessing of the AOD results when required. This post calibration is achieved by either statistical analysis of in-situ Langley calibrations at sites where reliable results are possible or else by an exchange of the instrument and its re-calibration at Davos.

Finally, hourly data records are prepared from quality assured level-3 data into annual files in NARSTO format and submitted to the World Data Center for Aerosols (WDCA). This data center is operated by the Institute for Environment and Sustainability at the Joint Research Centre of the European Commission in Ispra, Italy and collects measurements of aerosol properties taken in the GAW program. AOD results for 9 PFR stations up to 2005 can currently be obtained from WDCA.

The hourly data records contain mean and median values of AOD at the 4 wavelengths of the PFR, together with corresponding standard deviations and number of 1-minute samples. The high frequency AOD data are further screened per wavelength as follows:

1) A minimum of \( N=50 \) ‘cloud-free’ samples per day are needed to start processing. This requirement will eliminate days with less than 1 hour of sunshine.

2) A minimum of \( M=6 \) ‘cloud-free’ samples per hour are needed to calculate hourly statistics. This requirement will eliminate records with insignificant coverage.

3) Samples larger than 2 standard deviations off an initial hourly mean are eliminated as outliers, however, samples smaller than the initial mean are included in the next screening step. This screening rule tries to capture measurements affected by residual cloud contamination.

4) Hourly intervals where the standard deviation either exceeds 20% of the cleaned mean or 0.05 optical depths are set invalid.

Step 4) assumes that aerosol concentrations are changing ‘slowly’, which may not apply for stations close to strong local sources. However, as the GAW stations are located at remote or background sites, this rule may be applicable. A more sophisticated method screening for smoothness could analyze the temporal behavior of the samples, but is also much more complex to implement in a robust fashion suitable for routine processing of large data sets.
6.3 Network intercomparisons

Comparability between different national and international networks is required in order to build up a global AOD monitoring data base and fill in observational gaps. Homogeneity within a given network can be achieved through standardization of instrumentation and procedures in combination with a data quality control system, as it is for instance implemented in the AERONET program (Holben et al., 1998). Given the differences in instrumentation characteristics, calibration strategies and processing algorithms used by different networks, the effective equivalence of AOD observations needs to be estimated through Intensive Observation Periods (Schmid et al., 1999) or extensive field comparisons (McArthur et al., 2003, Mitchell and Forgan, 2003) of co-located instruments representing different networks.

Filter radiometer comparisons were held twice at WORCC on occasion of the International Pyrheliometer Comparisons (IPC) in 2000 (FRC-I) and 2005 (FRC-II). In FRC-I, measurements of 16 instruments of 6 different types representing 11 institutions from 7 countries were evaluated by uniform software in order to discriminate between instrument specific results. Albeit weather conditions were far from favorable, 340 measurements were collected in two ‘clear-sky’ periods on 27th September and 9th October 2000. Nine instruments were calibrated within less than a year before FRC-I, and one instrument was evaluated using a calibration obtained shortly after FRC-I by the participating institution. The other 6 instruments were provided with calibration coefficients based on PFR-N01 during the comparison.

Table 11 Daily means of AOD on 9 October 2000 measured by 10 filter radiometers participating in the FRC-I using original calibration coefficients. Instruments 1 – 8 agree within a range of less than 0.02 optical depth at 7 wavelengths compared.

<table>
<thead>
<tr>
<th>#</th>
<th>Wavelength</th>
<th>368</th>
<th>412</th>
<th>500</th>
<th>610</th>
<th>675</th>
<th>778</th>
<th>862</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>PFR-N01(^4)</td>
<td>0.049</td>
<td>0.042</td>
<td>0.031</td>
<td>0.013</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>PFR-N20</td>
<td>0.048</td>
<td>0.042</td>
<td>0.032</td>
<td>0.013</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>PFR-N20</td>
<td>0.056</td>
<td>0.051</td>
<td>0.031</td>
<td>0.011</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>CS1010</td>
<td>0.039</td>
<td>0.031</td>
<td>0.021</td>
<td>0.012</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>CSEM2016</td>
<td>0.046</td>
<td>0.029</td>
<td>0.041</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>CSEM2000</td>
<td>0.053</td>
<td>0.051</td>
<td>0.028</td>
<td>0.013</td>
<td>0.016</td>
<td>0.013</td>
<td>0.012</td>
</tr>
<tr>
<td>7</td>
<td>MFRSR 378</td>
<td>0.052</td>
<td>0.033</td>
<td>0.023</td>
<td>0.021</td>
<td>0.019</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>MFRSR 914</td>
<td>0.049</td>
<td>0.042</td>
<td>0.032</td>
<td>0.028</td>
<td>0.023</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>SPI-03</td>
<td>0.073</td>
<td>0.062</td>
<td>0.045</td>
<td>0.038</td>
<td>0.026</td>
<td>0.022</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>MS110</td>
<td>0.033</td>
<td>0.016</td>
<td>0.004</td>
<td>0.003</td>
<td>0.003</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

| Mean AOD | 0.051 | 0.048 | 0.032 | 0.025 | 0.017 | 0.014 | 0.014 |
| Sdev. 1 – 8 | 0.012 | 0.007 | 0.008 | 0.010 | 0.010 | 0.008 | 0.007 |
| Sdev. 1 – 10 | 0.004 | 0.005 | 0.004 | 0.008 | 0.006 | 0.001 | 0.005 |
| Range 1 – 8 | 0.010 | 0.013 | 0.014 | 0.019 | 0.012 | 0.002 | 0.012 |

\(^4\) Instrument manufacturers

PFR PMOD/WRC, Switzerland, [http://www.pmodwrc.ch](http://www.pmodwrc.ch)
CS Carter-Scott Design, Australia, [http://www.middletonsolar.com](http://www.middletonsolar.com)
CSEM Swiss Center for Electronics and Microtechnology, the SPM 2000 is no longer produced
MFRSR Yankee Environmental Systems, USA, [http://www.yesinc.com](http://www.yesinc.com)
SP1 Dr. Schulz & Partner GmbH, Germany, [http://www.drschulz.com](http://www.drschulz.com)
MS110 EKO Company, Japan. The MS110 is no longer produced
Daily mean AOD values from 10 filter radiometers with original calibration coefficients listed in Table 11 show agreement to within about 0.02 optical depths or less, when the last two instruments are not taken into account. PFR instruments, N01 and N26, agree even down to 0.001. These two master instruments used at Davos were calibrated on a stratospheric balloon (N01 in 1998) and at Mauna Loa (N26 in 2000).

The results of FRC-I have demonstrated that different types of current filter radiometers can actually achieve an agreement within ±0.01 optical depths, which is twice as good as the accuracy goal set for the GAW program by WMO (WMO, 1993a).

This encouraging result from a limited data set only was confirmed by an extended field comparison of network sun photometers during summer 2001 (McArthur et al., 2003), where four types of instruments co-located at Bratt’s Lake were operating under AERONET, GAWPFR and U.S. Dept. of Agriculture network protocols. The results of this 3-month comparison indicate that individual AOD observations from the direct-pointing instruments were indeed comparable within ±0.01 optical depths and their overall root mean squared difference at 500nm was 0.007.

A second filter radiometer comparison was held at WORCC in autumn 2005 between 15 instruments from 9 countries. FRC-II was based on AOD results provided by participants using their preferred algorithms. During 10 days with excellent weather conditions more than 3000 clear Sun were collected at 1 minute intervals with daily mean AOD between 0.025 and 0.110 at 500nm. Excellent agreement within better than ±0.01 at the common wavelengths of 870±5nm and 500±3nm was found for all instruments, indicating that this level of uncertainty found in previous comparisons is routinely achieved by operational observation programs. An example of measured AOD and the differences between instruments on a typical day is shown in Figure 24.
In order to estimate the uncertainties in AOD due to algorithms, the raw measurements of a PFR were processed simultaneously by three experts. Besides measurements taken during FRC-II, they were given data taken earlier in 2005 at the instruments network site, Bratt’s Lake, and data taken in 2004 at Mauna Loa to derive calibration coefficients. Table 12 lists the daily average values of AOD as they were obtained in 5 different ways of processing. The spread of results is smaller than 0.005 at all 4 wavelengths; calibration site and epoch seem to have negligible effect, indicating a high stability of the instrument over time and during transport.

On the right side of Table 12, equivalent results are given for a CIMEL radiometer that was processed by version 1 and 2 of the direct Sun algorithm of AERONET. This instrument was calibrated briefly before FRC-II at the European calibration facility of AERONET. The difference between processing versions is similar to the spread in the PFR results at 500nm, but significantly larger at 380 nm and 870nm. Version 2 results are within 0.005 to the PFR results with a notable improvement at 380 nm with respect to version 1. The calibration at 870 nm was apparently off by more than 1%, and the
corresponding data would be corrected or eliminated in AERONET quality assurance process, which includes a re-calibration.

This exercise shows that the uncertainties in AOD due to algorithms are of the same order as those associated with calibration. It also confirms that AOD derived from well calibrated instruments running under different network protocols can achieve agreement to 0.005 optical depths. It must be emphasized here, that this level of uncertainty may be spoiled by operational defaults like contamination of the optical path (thin cirrus clouds, dirty optics) or degraded pointing of the instrument.

6.4 First Results of the GAW PFR network

Starting with hourly means, as they are available from WDCA from 1999 to 2005, time series of daily, monthly and annual means were constructed and are presented here either in graphical or tabular form. Until 2003, screening of the high frequency data for hourly means was limited to steps 1 and 2 described above in 6.2 Network operations. Therefore additional screening of hourly data was applied to eliminate few (typically less than 10) outliers from each dataset before calculating the coarser averages.

Aerosol optical depths are usually reported as arithmetic means and associated standard deviations over a selected averaging period, based on the hypothesis of an underlying normal distribution. However, AOD is often better characterized by a lognormal distribution and described by geometric mean and standard deviation. Based on a statistical analysis of skewness and kurtosis in a multi-year and multi-station AOD data set, O’Neill and co-authors have shown (O’Neill et al., 2000) that a lognormal distribution systematically provides a more robust base for reporting AOD statistics than the normal distribution. The graphical overviews of AOD results shown here are confirming this suggestion, and thus geometric means and standard deviations will be given in addition to the usual arithmetic parameters.

Daily means from 9 stations are presented in graphical overviews (Fig. 25 to 34) showing the time series of AOD at 500nm (top panel), its distribution as a histogram (bottom left panel) and (bottom right) a scatter plot of the Ångström wavelength exponent \( \alpha \) versus AOD. The wavelength exponent summarizes the results of the 3 other PFR channels at 368, 412 and 862nm under the assumption that the empirical Ångström law \( \text{AOD} = \beta \lambda^{-\alpha} \) is applicable. Under the same assumption, the Ångström turbidity coefficient \( \beta \) is readily obtained as \( \beta = \text{AOD}(500) \times 0.5^{\alpha} \).

Monthly values are listed as geometric means in tabular form in Appendix A. No results are given for months where a minimal number of 30 quality controlled hourly values was not available, either because of unfavourable weather conditions, polar night or due to technical limitations like misalignment of the solar tracker. The ‘climatological values’ CLMT in the bottom rows of each table are calculated from at least 90 hourly values during the period of observations, not as the average of monthly means given above. The geometric standard deviation STDV applies to the same data set and gives an estimate for the range of the monthly variability of hourly mean AOD. This monthly climatology of AOD is illustrated in Figure 35.

Annual means of AOD and Ångström wavelength exponent are listed in Tables 13 to 21 together with the number of hourly means. (The total number of observations is actually 6 to 60 times larger). Angle brackets \( < > \) are used to indicate arithmetic means and
slashes / / to indicate geometric means. Caption of Figure 25 and Table 13 are applicable to all following figures and tables for the other station overviews as well.

Brief descriptions listed under each station entry are based on information obtained from the GAW Station Information System (http://www.empa.ch/gaw/gawsis) and personal communications by station managers.
**Alice Springs (ASP)**

The Bureau of Meteorology station at Alice Springs is located in a dry desert climate zone in the Northern Territory of Australia. It is a station of the World Climate Research Program (WCRP) Baseline Surface Radiation Network (BSRN) since 1995, as well as a contributing station of the Global Atmospheric Watch program of the Atmospheric Research and Environment Division of WMO.

The annual variation of AOD shows the standard pattern of low values in Austral winter and higher values in summer. Minimal daily means are comparable to high altitude stations in the northern hemisphere. Maximal values might be due to local bush fires (B. Forgan, private communication), but also due to long-range transport of biomass burning smoke from Indonesia.

![Graphical overview of AOD daily means at Alice Springs, Australia. Top: Annual variation of AOD at 500nm; Bottom left: Histogram and lognormal fit, right: Ångström wavelength exponent α versus AOD at 500nm.](image)

**Figure 25**

![Table 13 Annual statistics of AOD at 500nm and Ångström exponent at Alice Springs. Arithmetic means are listed under <AOD> and <Alpha>, /AOD/ is the geometric (or logarithmic) mean, and N gives the number of hourly means per year. The ‘overall’ results in the last row are based on hourly means for the whole observation period.](image)

**Table 13**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2002</td>
<td>0.091 ± 0.132</td>
<td>0.052 /* 2.66</td>
<td>1.03 ± 0.40</td>
<td>2339</td>
</tr>
<tr>
<td>2003</td>
<td>0.050 ± 0.044</td>
<td>0.039 /* 2.00</td>
<td>0.85 ± 0.35</td>
<td>3037</td>
</tr>
<tr>
<td>2004</td>
<td>0.052 ± 0.043</td>
<td>0.039 /* 2.16</td>
<td>0.94 ± 0.38</td>
<td>2448</td>
</tr>
<tr>
<td>2005</td>
<td>0.056 ± 0.047</td>
<td>0.046 /* 1.80</td>
<td>1.02 ± 0.43</td>
<td>2652</td>
</tr>
<tr>
<td>overall</td>
<td>0.066 ± 0.075</td>
<td>0.047 /* 2.18</td>
<td>0.96 ± 0.37</td>
<td>10468</td>
</tr>
</tbody>
</table>
Bratt's Lake (BRA)

Bratt's Lake is situated approximately 23 km south of the city of Regina, Saskatchewan. The original observatory was established in 1995 for the WCRP BSRN program. The observatory is on the Canadian prairie and is situated on a working grain farm. The topography is flat within ±5 meter over a radius of 20 km at an altitude of 578 meters above sea level. The term ‘lake’ refers to an intermittent body of water of less than 0.5 square kilometers to the east of the site. Major inorganic aerosol components are measured by filter sampling method. The station runs a Cimel radiometer for AEROCAN and a Brewer ozone spectrometer.

![Figure 26 as Fig. 24, for Bratt's Lake, Canada](image)

![Table 14 as Table 13, for Bratt's Lake, Canada](image)

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001</td>
<td>0.097 ± 0.076</td>
<td>0.079 /* 1.87</td>
<td>1.25 ± 0.37</td>
<td>1367</td>
</tr>
<tr>
<td>2002</td>
<td>0.113 ± 0.117</td>
<td>0.083 /* 2.05</td>
<td>1.30 ± 0.41</td>
<td>1859</td>
</tr>
<tr>
<td>2003</td>
<td>0.139 ± 0.153</td>
<td>0.096 /* 2.29</td>
<td>1.26 ± 0.44</td>
<td>1879</td>
</tr>
<tr>
<td>2004</td>
<td>0.111 ± 0.141</td>
<td>0.075 /* 2.25</td>
<td>1.24 ± 0.41</td>
<td>1221</td>
</tr>
<tr>
<td>2005</td>
<td>0.079 ± 0.051</td>
<td>0.066 /* 1.80</td>
<td>1.40 ± 0.32</td>
<td>848</td>
</tr>
<tr>
<td>overall</td>
<td>0.113 ± 0.109</td>
<td>0.085 /* 2.03</td>
<td>1.24 ± 0.37</td>
<td>7091</td>
</tr>
</tbody>
</table>
Hohenpeissenberg (HPB)

Hohenpeissenberg is an isolated mountain 50 km to the south west of Munich. It rises 300 m above the surrounding area that is populated to an extent typical for Central Europe and partly covered with meadows (~70%) and forests (~30%). The observatory is situated on top of the mountain at an altitude of 989 meters above sea level. It has a long history of meteorological and climatological observations (since 1781). HPB has been a GAW global observatory since 1993 with a comprehensive measurements program covering aerosol, ozone, trace gases and precipitation chemistry.

The station lies frequently above a winter inversion layer that keeps local atmospheric pollutant contained to the valleys below, thus wintertime AOD is thus often comparable to small values found in Davos.

![AOD 500nm vs. Date](image)

![Number of cases vs. AOD 500nm](image)

![Ångström α vs. AOD 500nm](image)

**Figure 27 as Fig. 24, for Hohenpeissenberg, Germany**

**Table 15 as Table 13, for Hohenpeissenberg, Germany**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>1999</td>
<td>0.106 ± 0.080</td>
<td>0.082 /* 2.12</td>
<td>1.53 ±0.29</td>
<td>276</td>
</tr>
<tr>
<td>2000</td>
<td>0.101 ± 0.071</td>
<td>0.078 /* 2.08</td>
<td>1.47 ±0.33</td>
<td>823</td>
</tr>
<tr>
<td>2001</td>
<td>0.111 ± 0.091</td>
<td>0.085 /* 2.08</td>
<td>1.54 ±0.44</td>
<td>682</td>
</tr>
<tr>
<td>2002</td>
<td>0.121 ± 0.089</td>
<td>0.091 /* 2.22</td>
<td>1.47 ±0.36</td>
<td>598</td>
</tr>
<tr>
<td>2003</td>
<td>0.106 ± 0.087</td>
<td>0.076 /* 2.35</td>
<td>1.41 ±0.37</td>
<td>472</td>
</tr>
<tr>
<td>2004</td>
<td>0.085 ± 0.065</td>
<td>0.065 /* 2.15</td>
<td>1.33 ±0.37</td>
<td>493</td>
</tr>
<tr>
<td>2005</td>
<td>0.141 ± 0.098</td>
<td>0.109 /* 2.14</td>
<td>1.40 ±0.43</td>
<td>1208</td>
</tr>
<tr>
<td>overall</td>
<td>0.118 ± 0.083</td>
<td>0.092 /* 2.10</td>
<td>1.43 ±0.35</td>
<td>4552</td>
</tr>
</tbody>
</table>
**Izaña (IZO)**

The Izaña station is located on the Island of Tenerife, Spain, roughly 300 km west of the African coast. The meteorological observatory is situated on a mountain platform at an altitude of 2367 meters above sea level, 15 km north-east of the volcano Teide. Local wind field at the site is dominated by north-westerly winds. A predominant meteorological attribute of the Canary Islands region is the presence of the trade wind inversion that persists through most of the year and is well below the altitude of the station. The ground in the vicinity around Izaña is loosely covered with light volcanic soil. The vegetation in the surrounding area is sparse, consisting mainly of broom. IZO is a GAW global observatory with a broad range of atmospheric measurements including aerosol light scattering coefficient and number concentrations, ozone and greenhouse gases.

![Graph of AOD 500nm vs Date]

**Table 16 as Table 13, for Izaña, Spain**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001</td>
<td>0.057 ± 0.073</td>
<td>0.032 /* 3.00</td>
<td>0.47 ±0.69</td>
<td>490</td>
</tr>
<tr>
<td>2002</td>
<td>0.043 ± 0.071</td>
<td>0.022 /* 3.14</td>
<td>1.12 ±0.92</td>
<td>2523</td>
</tr>
<tr>
<td>2003</td>
<td>0.028 ± 0.038</td>
<td>0.017 /* 2.77</td>
<td>1.27 ±0.82</td>
<td>830</td>
</tr>
<tr>
<td>2004</td>
<td>0.062 ± 0.082</td>
<td>0.034 /* 2.86</td>
<td>0.88 ±0.60</td>
<td>1068</td>
</tr>
<tr>
<td>2005</td>
<td>0.111 ± 0.111</td>
<td>0.063 /* 3.12</td>
<td>0.72 ±0.64</td>
<td>766</td>
</tr>
<tr>
<td>overall</td>
<td>0.061 ± 0.086</td>
<td>0.031 /* 3.13</td>
<td>0.98 ±0.78</td>
<td>5677</td>
</tr>
</tbody>
</table>
The Ångström $\alpha$ versus AOD plot at Izaña reveals a distinct branch where AOD is significantly larger than expected from the station altitude, and corresponding $\alpha$ values are very low. Figure 29 shows the histogram of the Ångström exponent $\alpha$ and a fitted bi-modal distribution with parameters $\alpha_1=1.388\pm0.425$ representing clean air conditions, and $\alpha_2=0.216\pm0.210$ indicative of Sahara dust events.

![Figure 29 Histogram of Ångström wavelength exponent $\alpha$ at Izaña observatory showing a bi-modal distribution composed of a broad normal distribution peak centered at $\alpha\approx1.39\pm0.42$, and a second mode at $\alpha\approx0.22\pm0.21$, reflecting frequent intrusions of Sahara minral dust over the station.](image-url)
**Jungfraujoch (JFJ)**

The high altitude research station Jungfraujoch is situated on a mountain saddle (3580m) between the two mountains Jungfrau (4158m) and Mönch (4099m). The GAW global station is located in the central Alps, far from major emission sources and is often representative of the lower free troposphere above a continental region. During summer months, thermal convection may break up the planetary boundary layer and up polluted air from the industrialized regions below.

A full suite of gas phase components is monitored by several research groups, in-situ aerosol measurements are performed by the Laboratory of Atmospheric Chemistry of the Paul Scherrer Institute, and an atmospheric radiation monitoring program (CHARM) is run by Meteo Swiss.

![Graph](image_url)

**Figure 30 as Fig. 24, for Jungfraujoch, Switzerland**

**Table 17 as Table 13, for Jungfraujoch, Switzerland**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>1999</td>
<td>0.025 ± 0.019</td>
<td>0.020 /* 1.90</td>
<td>1.28 ±0.34</td>
<td>762</td>
</tr>
<tr>
<td>2000</td>
<td>0.024 ± 0.016</td>
<td>0.020 /* 1.76</td>
<td>1.32 ±0.37</td>
<td>802</td>
</tr>
<tr>
<td>2001</td>
<td>0.025 ± 0.021</td>
<td>0.020 /* 1.76</td>
<td>1.15 ±0.35</td>
<td>752</td>
</tr>
<tr>
<td>2002</td>
<td>0.025 ± 0.018</td>
<td>0.020 /* 1.86</td>
<td>1.49 ±0.41</td>
<td>852</td>
</tr>
<tr>
<td>2003</td>
<td>0.029 ± 0.022</td>
<td>0.024 /* 1.74</td>
<td>1.23 ±0.33</td>
<td>1035</td>
</tr>
<tr>
<td>2004</td>
<td>0.022 ± 0.013</td>
<td>0.019 /* 1.65</td>
<td>1.21 ±0.29</td>
<td>629</td>
</tr>
<tr>
<td>2005</td>
<td>0.031 ± 0.025</td>
<td>0.025 /* 1.83</td>
<td>1.19 ±0.46</td>
<td>1256</td>
</tr>
<tr>
<td>overall</td>
<td>0.027 ± 0.019</td>
<td>0.023 /* 1.78</td>
<td>1.25 ±0.34</td>
<td>6088</td>
</tr>
</tbody>
</table>
Mace Head (MHD)

Mace Head Research Station is located on the west coast of Ireland, County Galway. The site offers excellent exposure to the North Atlantic, and is representative of relatively clean background marine air. The nearest major conurbation approximately 90 km to the east of Mace Head is Galway. The hilly area around Mace Head is wet and boggy with a lot of exposed rock and vegetation which consists mainly of grasses and sedges. The facilities at the site consist of three laboratory buildings, equipped with a wide range of state of the art in-situ gaseous, aerosol particulate and radiative instrumentation. The two shore laboratories, ca. 90m from the shore, house gas and aerosol in-situ measurement equipment; meteorological and solar radiation parameters are measured at the cottage laboratory, ca. 300m from the shore. MHD is a GAW global station.

Figure 31 as Fig. 24, for Mace Head, Ireland

Table 18 as Table 13, for Mace Head, Ireland

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001</td>
<td>0.100 ± 0.061</td>
<td>0.088 /* 1.60</td>
<td>0.85 ± 0.41</td>
<td>396</td>
</tr>
<tr>
<td>2002</td>
<td>0.103 ± 0.057</td>
<td>0.090 /* 1.65</td>
<td>0.93 ± 0.45</td>
<td>240</td>
</tr>
<tr>
<td>2003</td>
<td>0.133 ± 0.093</td>
<td>0.107 /* 1.96</td>
<td>0.91 ± 0.41</td>
<td>209</td>
</tr>
<tr>
<td>2004</td>
<td>0.094 ± 0.058</td>
<td>0.082 /* 1.64</td>
<td>0.92 ± 0.44</td>
<td>262</td>
</tr>
<tr>
<td>2005</td>
<td>0.103 ± 0.065</td>
<td>0.089 /* 1.70</td>
<td>0.74 ± 0.53</td>
<td>186</td>
</tr>
<tr>
<td>overall</td>
<td>0.106 ± 0.063</td>
<td>0.092 /* 1.67</td>
<td>0.84 ± 0.44</td>
<td>1293</td>
</tr>
</tbody>
</table>
Mauna Loa Observatory (MLO)

MLO is located on the north flank of Mauna Loa Volcano, on the Big Island of Hawaii. Due to its remote location in the Pacific Ocean, high altitude (3397m), and great distance from major pollution sources, MLO is a prime spot for sampling the Earth's background air in the well mixed free troposphere. The observatory protrudes through the strong marine temperature inversion layer present in the region, which separates the more polluted lower portions of the atmosphere from the much cleaner free troposphere.

![Graph of AOD vs Date for Mauna Loa](image)

**Figure 32 as Fig. 24, for Mauna Loa, USA**

**Table 19 as Table 13, for Mauna Loa, USA**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2000</td>
<td>0.018 ± 0.013</td>
<td>0.016 /* 1.52</td>
<td>1.58 ± 0.30</td>
<td>1829</td>
</tr>
<tr>
<td>2001</td>
<td>0.020 ± 0.015</td>
<td>0.017 /* 1.62</td>
<td>1.46 ± 0.35</td>
<td>2218</td>
</tr>
<tr>
<td>2002</td>
<td>0.019 ± 0.013</td>
<td>0.017 /* 1.57</td>
<td>1.49 ± 0.36</td>
<td>1998</td>
</tr>
<tr>
<td>2003</td>
<td>0.017 ± 0.014</td>
<td>0.014 /* 1.61</td>
<td>1.20 ± 0.27</td>
<td>1867</td>
</tr>
<tr>
<td>2004</td>
<td>0.016 ± 0.014</td>
<td>0.013 /* 1.61</td>
<td>1.15 ± 0.29</td>
<td>1991</td>
</tr>
<tr>
<td>2005</td>
<td>0.025 ± 0.021</td>
<td>0.020 /* 1.78</td>
<td>1.31 ± 0.36</td>
<td>2182</td>
</tr>
<tr>
<td>overall</td>
<td>0.020 ± 0.013</td>
<td>0.017 /* 1.59</td>
<td>1.35 ± 0.31</td>
<td>12085</td>
</tr>
</tbody>
</table>
Ny Ålesund (NAS)

Ny-Ålesund in the European Arctic is one of the world's northernmost human settlements. The small town has about 30 permanent inhabitants and hosts up to 150 visiting scientists during summer season. During the polar night from 26th October to 16th February no measurements are possible at all, but the dark period is well used for annual re-calibration of the PFR at Davos.

![Graph showing aerosol optical depth (AOD) variations in Ny Ålesund](image)

**Figure 33 as Fig. 24, for Ny Ålesund, Norway**

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2002</td>
<td>0.074 ± 0.054</td>
<td>0.063 /* 1.75</td>
<td>1.21 ±0.35</td>
<td>880</td>
</tr>
<tr>
<td>2003</td>
<td>0.106 ± 0.081</td>
<td>0.086 /* 1.91</td>
<td>1.31 ±0.40</td>
<td>861</td>
</tr>
<tr>
<td>2004</td>
<td>0.108 ± 0.087</td>
<td>0.087 /* 1.91</td>
<td>1.28 ±0.44</td>
<td>920</td>
</tr>
<tr>
<td>2005</td>
<td>0.108 ± 0.100</td>
<td>0.085 /* 1.95</td>
<td>1.19 ±0.50</td>
<td>1027</td>
</tr>
<tr>
<td>overall</td>
<td>0.101 ± 0.077</td>
<td>0.082 /* 1.87</td>
<td>1.21 ±0.38</td>
<td>3688</td>
</tr>
</tbody>
</table>

When the Sun rises above the horizon, fog and low clouds, as well as the shadow of surrounding mountains severely limit the number of measurements. Aerosol optical depth at Ny Ålesund is usually higher during spring than in summer. This is due to the Arctic haze phenomenon when aerosol removal mechanisms are suppressed and the cold and stable winter Arctic boundary layer may extend over industrial sources in the south. Summer precipitation results in a fast removal of accumulated aerosols and AOD values fall well below continental values.
Ryori (RYO)

Ryori regional GAW station is sited on a hilly cape at 230m on the Pacific coast in the northern part of the Japanese main island, Honshu. The surrounding area is sparsely populated, about 120 km apart from Sendai, the largest city of the region with a population of one million. The area has a temperate climate with distinct four seasons, with annual temperature of about 10 °C and annual precipitation is about 1,300 mm. The measurement program at Ryori covers mainly greenhouse gas concentrations and includes vertical aerosol profiles measured by an operational LIDAR system.

![Graph showing AOD 500nm vs. Date for Ryori, Japan from Jan 02 to Jan 06](image)

![Graph showing Number of cases vs. AOD 500nm](image)

![Graph showing Ångström α vs. AOD 500nm](image)

Figure 34 as Fig. 24, for Ryori, Japan

Table 21 as Table 13, for Ryori, Japan

<table>
<thead>
<tr>
<th>YEAR</th>
<th>&lt;AOD&gt;</th>
<th>/AOD/</th>
<th>&lt;Alpha&gt;</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>2002</td>
<td>0.186 ± 0.131</td>
<td>0.149 /* 1.98</td>
<td>1.26 ±0.34</td>
<td>713</td>
</tr>
<tr>
<td>2003</td>
<td>0.221 ± 0.236</td>
<td>0.148 /* 2.36</td>
<td>1.28 ±0.26</td>
<td>869</td>
</tr>
<tr>
<td>2004</td>
<td>0.182 ± 0.130</td>
<td>0.142 /* 2.06</td>
<td>1.25 ±0.32</td>
<td>1497</td>
</tr>
<tr>
<td>2005</td>
<td>0.190 ± 0.156</td>
<td>0.143 /* 2.10</td>
<td>1.27 ±0.32</td>
<td>1312</td>
</tr>
<tr>
<td>overall</td>
<td>0.196 ± 0.162</td>
<td>0.150 /* 2.07</td>
<td>1.26 ±0.28</td>
<td>4391</td>
</tr>
</tbody>
</table>


6.5 Monthly climatology

Most stations of the GAWPFR network have been operating since at least 4 years, and thus qualify as long-term monitoring sites according to GAW. It now becomes possible to establish an initial climatology for these stations that may serve as reference values against which aerosol events like boreal fire smoke, desert dust storms or volcanic eruptions can be detected.

For GAW Global stations it is required that “they are situated in remote locations with very low background levels of pollutants and are representative of large geographic areas” (WMO, GAW No. 143). This condition implies that measurements at these locations should be unaffected by local sources of pollution most of the time. Monthly means will smooth out episodic variability in AOD on hourly to weekly time scales and represent the basic data unit for climatology. A table of all monthly means of the stations in GAWPFR is given in Appendix A.

Geometric means and standard deviations were calculated from the hourly values for each month. Figure 35 shows an overview of this monthly climatology with the geometrical standard deviation indicated by vertical bars. A corresponding monthly climatology of the Ångström exponent is shown in Figure 36. The individual diagrams are arranged with station height increasing from top to bottom. Coastal stations Mace Head (MHD), Ny Ålesund (NAS) and Ryori (RYO) are below 250m; continental stations Alice Springs (ASP), Bratt’s Lake (BRA) and Hohenpeissenberg (HPB) are below 1'000m. The mountain stations Izaña (IZO), Mauna Loa (MLO) and Jungfraujoch (JFJ) are above 2'300m.

All stations show larger AOD values in the summer than in winter months, with maximal values typically between spring and summer. This behaviour was already described by Ångström (1964) for a number of historic records as a general rule. He gave an obvious cause “that on account of the rapid heating of the ground in spring and early summer, the atmosphere shows its greatest vertical instability, and that consequently dust and combustion products are then carried most effectively from ground to the atmosphere”. The vegetation cycle (biogenic aerosols like pollen, DMS secondary aerosols) or agricultural activities (plowing, harvesting) may further contribute to enhanced summer aerosol loads or introduce additional peaks during biomass burning periods. Such a pattern is clearly visible at the continental stations Bratt’s Lake and Alice Springs (inverted calendar cycle of southern hemisphere) and at the coastal station Ryori. (The value in August at Bratt’s Lake is biased due to boreal fires in 2003 when the monthly mean was about twice that of other years)

In continental Hohenpeissenberg, AOD values remain rather constant from April to August and at a lower level from November to February, with transition phases in March and September. Solar elevation or radiative energy seems to be the dominant factor driving AOD levels. A possible reason for this may be that due to its location on top of a hill, the observatory stays, at least partly, isolated from aerosols originating from the surrounding plane some 500m below through the mechanism suggested by Ångström.

A seasonal pattern with peak values between spring and summer is also observed at the high altitude stations Mauna Loa and Jungfraujoch. Both are even more isolated from surrounding aerosol sources than Hohenpeissenberg and frequently located in the free troposphere. The pattern observed at these two sites would thus represent an average over large regions like Western Europe or the Pacific including long-range transport from Asia.
(Dutton, 1994). AOD levels above 3'000m are about 4 to 8 times smaller than at the lower stations, with the exception of Alice Springs where AOD is just about twice the values observed at high altitudes.

At Izaña, higher values are found from July to October, probably due to a seasonal pattern in Sahara dust events, indicated by significantly higher monthly variability and smaller Ångström exponents (coarse mode aerosols) for these months.

The polar station Ny Ålesund shows a different pattern with largest AOD values occurring in March followed by a steady decline to smallest values in September. This behaviour may be explained by the ‘Arctic Haze’ phenomena, where pollution from Europe transported to high latitudes is accumulated in a stable inversion layer and removal processes are reduced during polar night (see e.g. Barrie, 1986 or Shaw, 1995).

### 6.6 Annual and multi-year climatology

With just 4 to 7 years of data, the time series are still too short to allow for a reliable trend analysis (Weatherhead et al, 1998) thus only static annual means were given in tables 13 to 21.

Table 22 gives a summary of the different averaging methods over the whole observation periods at the 9 GAW stations. In addition to arithmetic mean <AOD> and geometric mean /AOD/, median values are given under {AOD}. Geometric means are much closer to the medians than arithmetic means. Their geometric standard deviation is of the order of 2, which means that AOD varies between 50% and 200% of the observed geometric mean. At Izaña, frequent Sahara dust events are responsible for a larger variability by a factor of three.

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<th>/AOD/ ±</th>
<th>{AOD} ±</th>
<th>&lt;Alpha&gt; ±</th>
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Long-term geometric mean values of AOD at 6 stations below 1km elevation (MHD, NYA, RYO, ASP, BRA, HPB) average to 0.09 (arithmetic means to 0.117) and vary by a factor of 3 between ASP /0.047/ and RYO /0.150/. This ‘global’ average is somewhat smaller than the generally accepted global (arithmetic) mean of 0.150 that is derived from
ground based observations and satellite retrievals. This may be partly explained by the fact that the distribution of GAWPFR stations is far from representing global conditions, as most of them are located at background sites.

Long-term geometric mean values of AOD at stations above 3km elevation (IZO, JFJ, MLO) average to 0.024 (arithmetic mean 0.036). The PFR result for Mauna Loa is in perfect agreement with the corresponding result \(<\text{AOD}> = 0.02 \pm 0.01\) of the AERONET climatology.

The long-term mean of Ångström \(\alpha\) exponents average to 1.33 for the high altitude stations and to 1.16 for the surface stations, the former is very close to Ångström’s classic value of 1.3 for natural aerosols. Coarse mode aerosols indicated by small \(\alpha < 1\) exponents are dominating at Mace Head (maritime salt spray), Alice Springs and Izaña (mineral dust).
Figure 35 Overview of monthly climatology of AOD at 500nm for 9 stations in the GAWPFR network. Columns indicate the geometric mean of hourly averages; geometric standard deviation is indicated by vertical bars. Diagrams are arranged top to bottom in ascending station height.
Figure 36 Overview of monthly climatology of the Ångström α coefficient for 9 stations in the GAWPFR network. Columns indicate the arithmetic mean of hourly averages; the standard deviation is indicated by vertical bars.
Chapter 7 Conclusions and Outlook

The main objectives of this work were to develop a system of instrumentation, calibration and data reduction algorithms including data quality control to provide AOD observations of high quality for the Global Atmosphere Watch program of WMO. The major achievements of this work can be summarized as follows:

1. A Precision Filter Radiometer for AOD measurements was designed with special emphasis on instrumental stability and built-in features for data quality control. Multiyear calibration sequences have demonstrated an excellent stability of the PFR with annual drift rates of less than 1%, which represents a ten-fold improvement over instrument performance around 1990. This high stability significantly reduces the need for frequent calibration.

2. AOD algorithms, based on current WMO recommendations and scientific literature, were discussed and their limitations estimated. If an upper limit for relative algorithmic errors is set at 0.1%, or a tenth of the realistic calibration uncertainty, it was found that AOD measurements should be limited to optical air mass less than 6, that timekeeping needs to be accurate to better than 3 seconds, and that atmospheric pressure should be recorded with each measurement to an accuracy of 3hPa.

3. Several calibration techniques for filter radiometers, including variants of the classic Langley plots, two different laboratory methods, and measurements from a stratospheric balloon were explored. From this study, we conclude that Langley plot techniques remain the only method to derive a spectral calibration with adequate accuracy for AOD measurements.

4. In a numerical simulation, the refined Langley method was shown to avoid a systematic overestimation of calibration coefficients at low altitude sites. Based on this finding, an ‘objective’ procedure was developed that permits a reliable in-situ calibration also for low altitude sites to an accuracy previously only achieved at high altitude stations.

5. Two international instrument intercomparisons and a field comparison of 4 networks have shown that agreement of surface based AOD measurements better than 0.01 optical depths are achieved. It has also been shown that the choice of different algorithms may result in AOD differences that are of the same order as between different, well calibrated instruments.

6. A growing set of high quality AOD measurements is now available for long-term monitoring in the GAW network. These AOD measurements can also complement the co-located in-situ measurements of aerosol properties, solar radiation and meteorological parameters available at many GAW stations for case studies.
A review of WORCC by the SAG/Aerosols in 2004 on behalf of Meteo Swiss acknowledged that the originally given tasks were completely reached and their implementation was considered a major, cost-effective contribution towards a global AOD network. The SAG/Aerosol strongly recommended the transition of the PFR network from its demonstration phase into a fully operational mode, which would become the apex of the measurement traceability pyramid in a global network. The group acknowledged also that this new role of WORCC would require additional manpower and funding.

Based on the favorable review of the Swiss GAW program, including WORCC, by an international team of experts, and the endorsement by the Commission of Atmospheric Science working group on Environmental Pollution and Atmospheric Chemistry (CAS/EPAC) (WMO, 2005b) of these SAG/Aerosol recommendations for WORCC, additional funding was granted by Meteo Swiss in 2007.

The increase of funding will be invested in additional manpower and operational improvements of the PFR network towards faster turn-around of AOD data and a regular calibration schedule. The high quality of the AOD data obtained from the PFR network will permit to monitor additional aerosol parameters like fine and coarse mode fraction.
Appendix

Tables of monthly mean AOD

Monthly geometric means of hourly averages of AOD at 500nm for the GAWPFR sites. Missing values indicate months where the minimal number of 30 hourly averages was not reached due to weather conditions, polar night, or technical reasons. Climatological values (CLMT) and geometric standard deviation (STDV) are based on original hourly, not monthly averages.

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