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**Atmospheric Aerosol Properties from Ceilometer and Photometer
Observations at the L'Aquila Site**

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Dottorando

Andrea Balotti

Coordinatore

Prof. Massimiliano Aschi

Tutor

Prof. Vincenzo Rizi

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Andrea Balotti

Abstract

The ground-based instruments are still a backbone of high-quality atmospheric observation systems, providing high temporal resolution observations and a reference for satellite instruments validation. Some of them are relative cheap if compared to other research systems, and work automatically requiring little maintenance efforts from operators. This is the case of photometers and ceilometer, which are able to measure atmospheric constituents like aerosols, water vapor and clouds. In this work these instruments are used jointly with the aim to retrieve aerosol properties, analyzing the data collected at L'Aquila site. For each system, limitations and type of observable that is possible to derive are shown. Notably, using the specific ceilometer model installed at L'Aquila, the measured backscatter profile is influenced by the water vapor content. The column of water vapor at L'Aquila is investigated and compared against photometer, GNSS and radiosounding, at different time scales, to assess the influence on ceilometer profiles. An important advancement in photometry field is the use of lunar observations to partially cover the nighttime gap. The workflow for lunar photometry is developed and compared with the method applied in AERONET network. Finally, the algorithm GRASP for jointly inverting photometer and ceilometer observations is shown. Despite some inherent limitations and assumption, the photometer and ceilometer synergy is way to retrieve micro-physical and optical properties of atmospheric aerosol, requiring relative low-cost and low-maintenance instruments that can be applied extensively on the globe.

Keywords: atmospheric aerosol, ground-based remote sensing, water vapor content, lunar photometry, aerosol inversion.

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Introduction

A robust understanding of the Earth’s climate system requires accurate knowledge of atmospheric constituents that regulate radiative transfer, cloud processes, and the hydrological cycle. Among these, atmospheric aerosols and water vapor play a central role due to their strong influence on the planetary energy balance and their pronounced variability across spatial and temporal scales. Despite sustained progress in observations and modelling, both components remain major sources of uncertainty in climate assessments, underscoring the importance of their continued monitoring and characterization (IPCC 2023).

Atmospheric aerosols affect climate through both direct and indirect mechanisms. They interact directly with incoming solar and outgoing terrestrial radiation by scattering and absorbing energy, and indirectly by modifying cloud micro-physical properties, cloud lifetime, and precipitation processes. The Sixth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR6) identifies aerosol–radiation and aerosol–cloud interactions as the largest contributors to uncertainty in estimates of anthropogenic effective radiative forcing since pre-industrial times (IPCC 2023). This uncertainty reflects the complexity of aerosol physical and chemical properties, their vertical distribution, and the non-linear response of clouds to aerosol perturbations.

Observational evidence indicates that aerosol abundances have undergone substantial changes over both natural and anthropogenic timescales. Paleoclimate records from ice cores and marine sediments reveal large variations in aerosol loadings between glacial and interglacial periods, while industrial-era observations show a pronounced increase in aerosol concentrations until the late twentieth century, followed by declines in several Northern Hemisphere regions (IPCC 2023). Satellite observations and ground-based networks indicate predominantly negative trends in aerosol optical depth over Europe and North America since the early 2000s, contrasted by increasing or sustained high aerosol levels over parts of South and East Asia. These regional trends imply a reduction in the magnitude of the negative aerosol forcing and, consequently, an increase in net positive radiative forcing (IPCC 2023).

Water vapor, in contrast, is the most important greenhouse gas in the atmosphere and constitutes the dominant positive feedback in the climate system. Its concentration is tightly constrained by temperature through the Clausius-Clapeyron relationship, increasing by approximately 7 % per degree of warming. Consistent with this theoretical relationship, the IPCC AR6 reports that total column water vapor has very likely in-

creased since the 1970s, with particularly robust positive trends over the global oceans and tropical regions (IPCC 2023). These conclusions are supported by multiple independent observing systems, including satellites, radiosondes, reanalyses, and GNSS measurements, although uncertainties related to observing-system changes and regional data gaps remain.

Near-surface humidity observations further reveal the complexity of atmospheric moisture changes. Since the 1970s, near-surface specific humidity has very likely increased over both land and oceans, while relative humidity over land has decreased since around 2000, particularly across Northern Hemisphere mid-latitudes (IPCC 2023). These changes have important implications for evapotranspiration, drought occurrence, heat stress, and land-atmosphere coupling, highlighting the need for integrated analyses of temperature, circulation, and moisture processes.

Taken together, aerosols and water vapor exert a profound influence on the climate system through distinct but interconnected mechanisms. Aerosols provide a spatially heterogeneous and temporally evolving cooling effect with large associated uncertainties, while water vapor amplifies warming through a robust thermodynamic feedback. Reducing uncertainties in the quantification of both components is therefore essential for assessing past and future climate change, improving climate model performance, and supporting reliable climate projections.

Reducing the bias between models and observations, and addressing the uncertainties associated with aerosol climate effects, critically depends on the availability of high-quality, long-term observations capable of resolving key atmospheric properties in a vertically consistent manner (e.g., Stier et al. 2007; Wiegner et al. 2014). This concept trivially holds also for water vapor. Ground-based remote sensing techniques provide a fundamental observational pillar in this context, by offering high temporal resolution, well-characterized calibration, and enhanced sensitivity to local and regional processes. Then, ground-based observations can be viewed as a fundamental tool to validate and complementing satellite measurements.

Sun-sky photometers provide accurate measurements of column-integrated aerosol optical depth (AOD) and related optical and micro-physical properties, forming the basis of several global networks used to monitor aerosol variability and trends. However, photometric observations alone provide no information on the vertical distribution of aerosols, which can strongly influence aerosols radiative impact and their interaction with clouds (e.g., Frey et al. 2021; Kalashnikova et al. 2011). Lidar systems, by contrast, offer vertically resolved profiles of aerosol backscatter and extinction, enabling the identification of aerosol layers, transport pathways, and mixing processes, but typically require assumptions to retrieve intensive aerosol properties (e.g., Fernald 1984; Lopatin et al. 2013).

The complementary nature of lidar and photometer observations motivates their combined exploitation through joint inversion techniques. By simultaneously constraining column-integrated and vertically resolved measurements, a joint inversion approach reduces ambiguities inherent to single-instrument retrievals and enables a more physically consistent characterization of aerosol properties (Lopatin et al. 2013). This capability is particularly relevant in light of the dominant contribution of aerosol-related processes to

the uncertainty in anthropogenic radiative forcing highlighted by the IPCC AR6 (IPCC 2023).

Thesis outline

Within this context, the Department of Physical and Chemical Sciences of the University of L'Aquila, together with the Center of Excellence Telesensing of Environment and Model Prediction of Severe events (CETEMPS), manage two observation sites aimed at the long-term and continuous monitoring of aerosols and clouds. These two sites contribute to various global and European networks for atmospheric research. In these doctoral years, I am actively contributing in the process required to join to these networks and to the regular maintenance of the instruments at both sites.

Most of the work I have done in this doctoral thesis is aimed at the maintenance and study of the two sun-photometers operated by the University of L'Aquila. The second of them was installed in 2024, and is capable to measure during nighttime using the reflected light from the Moon. To keep constant in time the quality of observations, the photometers need to be annually calibrated, in addition to the regular on-site maintenance. The calibration is done at specific calibration centers, like the one at University of Lille in France. As person in charge to manage the sun-photometers at L'Aquila, I had the opportunity to spend a short period at the University of Lille to learn about advanced maintenance procedures.

The long-term photometric observations at L'Aquila site has been used to study the behavior of the total column of precipitable water vapor (PW) at different time scales, and compared with the PW observations from other instruments. This analysis is preparatory for the last part of this work since the PW in the atmosphere may influence the profile observed by the single-wavelength lidar (i.e, ceilometer) that is operated at L'Aquila site.

As output from workshops and training schools done during the doctoral period, I developed the processing chain to retrieve the AOD from photometer's lunar observations. The algorithm uses a free implementation of a lunar irradiance model with additional correction, and it provides comparable results respect to the lunar AOD available through AERONET network since August 2024.

The information obtained from photometry can be extended through joint lidar-photometer inversion to retrieve more aerosol properties in respect using a single instrument. By applying and evaluating the use of this synergistic strategy using the GRASP algorithm, the last part of this work aims at analyzing the characterization of aerosol using the photometer and ceilometer at L'Aquila site, including also the use of nighttime photometric observations.

The doctoral thesis is organized along these chapters:

- Chapter 1 contains a theoretical background about the most important aspects of radiative transfer, and explains some properties of atmospheric aerosols and water vapor that will be analyzed in the following chapters.

- Chapter 2 contains a description of the L'Aquila site, where this study has been conducted, and an overview of the major atmospheric observation networks at global/european level to which the L'Aquila site actively participate.
- Chapter 3 contains a general description of the instruments used in this work, including the theoretical background describing the operating principle, and the main quantities measured. Moreover, this chapter includes a description of the specific instruments model at L'Aquila site.
- Chapter 4 contains an assessment of precipitable water vapor at L'Aquila comparing several instruments. The study spans various time scales, from average climatological value to the diurnal behavior.
- Chapter 5 contains an analysis of nighttime AOD retrieved using the sun-sky-lunar photometer installed at L'Aquila site in 2024. The methodology has been developed from literature, and it was compared against AERONET nighttime observations for several example cases.
- Chapter 6 contains a description of the GRASP algorithm and the technical background which describe how it works. This chapter is aimed to implement the synergy of ground-based photometer and ceilometer to retrieve aerosol properties through the GRASP inversion algorithm. A specific example case at L'Aquila site is used to show the capabilities and limitations of this instrument's combination. This methodology has been developed from literature, but with modification to account for AOD observed at nighttime, and the use of a ceilometer instead of a multi-wavelength lidar.
- Chapter 7 contains a summary of findings shown in the previous chapters, and will exploit some aspects that was not possible to investigate in this doctoral work serve as a brief concluding chapter, summarizing all the findings presented in the thesis and exploring potential future projects that may stem from this research.

In summary, this doctoral thesis wants to show how jointly using different instruments may improve the characterization of atmospheric aerosols from ground-based observations, and the importance of a flexible inversion algorithm to perform such synergy.

Chapter 1

Atmospheric aerosols and water vapor

1.1 Introduction

The atmospheric medium is composed by a large variety of particles, that may interact each other depending on the domain of interest (e.g., radiation, chemistry, etc.). Because of the gravitational field of the Earth, most of the atmosphere is concentrated near the surface, with 99 % of its mass within the first 50 km from the surface (Wallace and Hoobs 2006).

The gases constituting most of the atmosphere mass (i.e., N_2 , O_2 , Ar , CO_2) are well-mixed and their mole fraction is constant with height up to an altitude of about 60 km (Liou 2002). Notable exceptions are the water vapor and ozone that are highly variable constituents in space and time, as well as other trace gases. Largest particles, as aerosol and cloud droplets, also have a very variable concentration. In Table 1.1 are reported the main constituents of the atmosphere as well as some of the most important greenhouse gases.

The radiation passing through the atmosphere is subject to phenomena of *absorption*, *scattering* and *emission*, all of them depending both on radiation's wavelength and chemical/thermodynamic characteristics of the atmosphere. For wavelengths commonly used by photometer and low-power lidar, the emission has often a negligible effect and will not be further considered from now on.

The combined effect of absorption and scattering is a change in magnitude and/or direction of the radiation, that ends in the attenuation of initial radiation beam, commonly defined as the *extinction* phenomena. In the atmosphere, gases and aerosols are the particles that contribute to the extinction, which magnitude depends linearly on (Wallace and Hoobs 2006):

- Intensity of radiation impinging on gases/aerosols.
- Concentration of gases/aerosols.

Table 1.1: Fractional Concentration by volume of the main gases in atmosphere from Wallace and Hoobs (2006). In bold the gases responsible for the greenhouse effect. For CO_2 and CH_4 , between round brackets, values observed in IPCC (2023).

Constituent	% by volume	Well-mixed
Nitrogen (N_2)	78.08 %	
Oxygen (O_2)	20.95 %	
Argon (Ar)	0.93 %	
Water vapor (H_2O)	0-5 %	No
Carbon dioxide (CO_2)	380ppm (415ppm)	
Methane (CH_4)	1.75ppm (1.87ppm)	
Nitrous oxide (N_2O)	0.3ppm	
Ozone (O_3)	0.1ppm	No

- Effectiveness of gases/aerosols in absorbing and scattering radiation.

All these factors must be accounted for each wavelength and for each type of gas or aerosol in the air volume.

The monochromatic radiation intensity (or monochromatic radiance, I_λ) is the energy flux per unit solid angle per unit area normal to the direction of the radiation beam, and is expressed in $W m^{-2} sr^{-1} \mu m^{-1}$. The monochromatic flux density (or monochromatic irradiance, F_λ) is defined as the integral of I_λ over all directions in a hemisphere above a unit of flat surface, and is expressed in $W m^{-2} \mu m^{-1}$.

In the case of extinction only (i.e., no emission), changes in magnitude of the monochromatic radiance are defined by the following (monochromatic) radiative properties:

$$\text{absorptivity} \quad A_\lambda = \frac{I_\lambda(\text{absorbed})}{I_\lambda(\text{incident})} \quad (1.1)$$

$$\text{reflectivity} \quad R_\lambda = \frac{I_\lambda(\text{reflected})}{I_\lambda(\text{incident})} \quad (1.2)$$

$$\text{transmissivity} \quad T_\lambda = \frac{I_\lambda(\text{transmitted})}{I_\lambda(\text{incident})} \quad (1.3)$$

These quantities are dimensionless and represent, respectively, the effectiveness of absorption, scattering and transmission, across a layer of absorbing/scattering particles. Because of the conservation of energy, it holds that:

$$A_\lambda + R_\lambda + T_\lambda = 1 \quad (1.4)$$

The decrease of radiance (I_λ) across an infinitesimal layer (ds) of absorbing/scattering atmosphere is:

$$dI_\lambda = -k_{e,\lambda} I_\lambda ds \quad (1.5)$$

where the $k_{e,\lambda}$ term is the *volume extinction coefficient* in m^{-1} . The $k_{e,\lambda}$ represents the probability per unit path length that a photon will be removed from the beam along the direction of incoming radiation, and it is defined as:

$$k_{e,\lambda} = N\sigma Q_{e,\lambda} \quad (1.6)$$

where N is the particle number density in m^{-3} , $\sigma = \pi r^2$ is the areal cross-section in m^{-2} for a single particle with radius r , and $Q_{e,\lambda}$ is the dimensionless extinction efficiency. In case of different kinds of particles in the air volume, their contributions to extinction can be summed up.

Substituting $Q_{e,\lambda}$ in Equation (1.6) with the absorption ($Q_{a,\lambda}$) or scattering ($Q_{s,\lambda}$) efficiencies, to get the *volume absorption coefficient* ($k_{a,\lambda}$) and the *volume scattering coefficient* ($k_{s,\lambda}$) respectively.

Moreover, because absorption/scattering efficiencies and coefficients are additive, it is trivial to obtain that:

$$\begin{aligned} Q_{e,\lambda} &= Q_{a,\lambda} + Q_{s,\lambda} \\ k_{e,\lambda} &= k_{a,\lambda} + k_{s,\lambda} \end{aligned}$$

Because the scattering process emits radiation in all directions around the particle, the scattered radiance, as well as the scattering and extinction coefficients, are directional quantities.

The angular dependency is explained by the *asymmetry factor* (g_λ) that is equals to 0 for isotropic scattering or ± 1 for completely forward/backward scattering. The *single-scattering albedo* (SSA, $\omega_{0,\lambda}$) parameter expresses how much of the radiation will be scattered given a single extinction event.

The $k_{e,\lambda}$, g_λ and $\omega_{0,\lambda}$ are the basic parameters used to characterize the optical properties of atmospheric particles (Wallace and Hoobs 2006).

1.1.1 Scattering

The scattering is a physical process by which a particle, in the path of an electromagnetic wave, continuously abstracts energy from the incident wave and radiates that energy in all directions. It is usually assumed the case of *independent scattering*, in which atmospheric particles are separated widely enough so that each particle scatters light in the same way as if all other particles did not exist (Liou 2002).

The scattering phenomena can be categorized in terms of changes to the radiation's wavelength, into:

- *Elastic scattering*: when scattered radiation has the same wavelength as the incident radiation.
- *Inelastic (or Raman) scattering*: when scattered radiation has a different wavelength respect to the incident radiation.

The observation of Raman scattering is useful to retrieve information about profiles of atmospheric temperature and concentration of water vapor, so that it is possible to infer

the aerosol extinction and backscatter coefficients without any kind of assumption, as it is needed in the case of elastic-only observations (see Section 3.3).

Another possible classification is related to the number of scattering events that should be considered for the propagating radiation:

- *Single scattering*: is an acceptable approximation for optically thin layers. The radiation is scattered only once before exiting from the air layer considered.
- *Multiple scattering*: needs to be accounted for optically thick and/or strongly scattering layers. The radiation is scattered more than once before exiting from the air layer considered.

The multiple scattering is an important process for the transfer of radiant energy in the atmosphere, especially when aerosols and clouds are involved (Liou 2002). Indeed, in presence of a sufficient number of multiple scattering events, a beam of radiation tends to be converted into isotropic radiation (Wallace and Hoobs 2006).

Because of the wide range of sizes and shapes in particles, different types of scattering can be observed, also called scattering regimes, that are identified through the dimensionless *size parameter* (x). This parameter is valid in case of a spherical particle, but this assumption is generally a good approximation except for very irregular particles (e.g., snowflakes, ice crystal, mineral dust, etc.) whose scattering is often computed directly using more sophisticated techniques. The size parameter is proportional to the ratio between particle radius (r) and wavelength (λ) of the incident radiation:

$$x = \frac{2\pi r}{\lambda}$$

The *Rayleigh scattering* regime holds for $x \ll 1$. In this case the scattering efficiency ($Q_{s,\lambda}$) is roughly proportional to λ^{-4} and the incident radiance is almost equally scattered between forward and backward directions (Figure 1.1.a). For $x \gtrsim 1$ the *Mie scattering* regime is used. In this case the scattering efficiency has not a fixed relationship with wavelength and the scattered radiance has a more complex angular distribution but mainly concentrated in the forward direction (Figure 1.1.b-c). For $x \gg 1$, the scattering tends to the *geometric optics* regime. In left plot of Figure 1.2, the scattering regimes for typical atmospheric particles are shown as function of wavelength.

The quantitative derivation of $Q_{s,\lambda}$ involves both the size parameter and the complex *refraction index* (m) of the particle. The real part of the refraction index (m_r) gives the strength of the scattering and the imaginary part (m_i) gives the strength of the absorption, with $m_i = 0$ meaning no absorption. In right plot of Figure 1.2, taken from Liou (2002), is reported the behavior of $Q_{s,\lambda}$, as a function of the size parameter, for a real part of the refraction index of $m_r = 1.5$ and several values of the imaginary part. The scattering efficiency shows a major oscillation coupled with minor ripples that approaches an asymptotic value for large values of the size parameter. The scattering efficiency between $x = 0.1$ and up to the asymptotic value, is related to the *Mie scattering* regime. Instead, from the asymptotic value onward, the scattering efficiency is related to the *geometric optics* regime. Both the oscillation and the minor ripples damp out as the particle's absorption increases (i.e., increasing m_i).

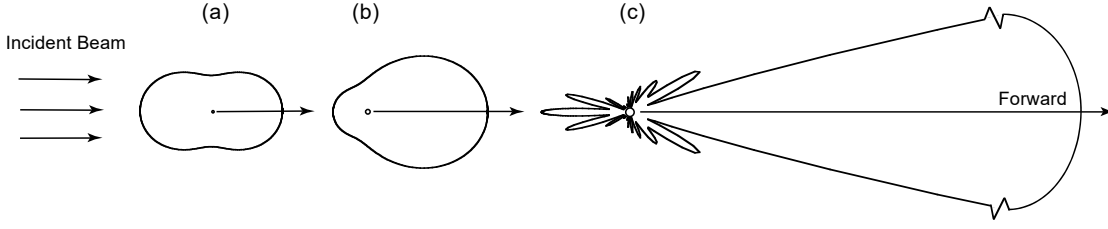


Figure 1.1: Angular distribution of the scattered radiance for a spherical aerosol with radius of (a) $10^{-4} \mu\text{m}$, (b) $0.1 \mu\text{m}$, or (c) $1 \mu\text{m}$, and illuminated by visible light of $0.5 \mu\text{m}$ wavelength. The size parameter is about 10^{-3} , 1.2 and 12.5 respectively. Adapted from Liou (2002).

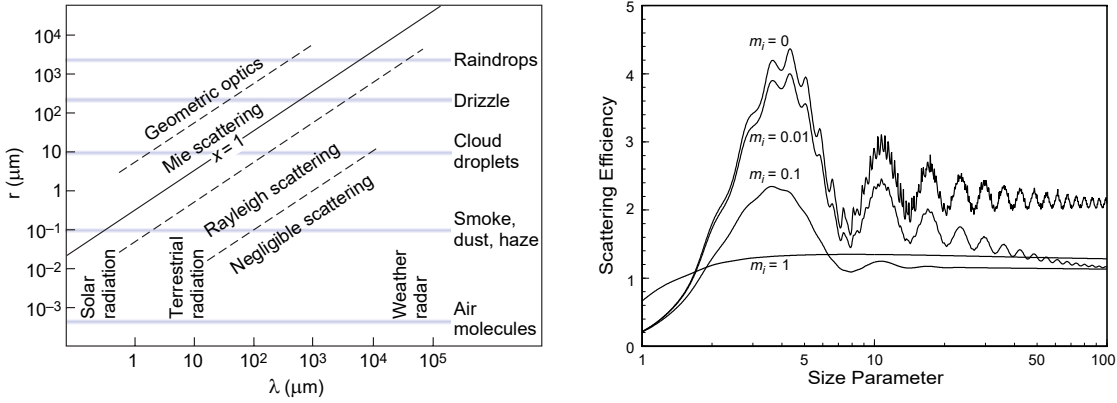


Figure 1.2: *Left:* Scattering regime as function of wavelength (λ) of the incident radiation and particle radius (r). Adapted from Wallace and Hoobs (2006). *Right:* Scattering efficiency as a function of size parameter, for different values of the imaginary part of the refractive index (m_i). The real part of refractive index is fixed to $m_r = 1.5$. Adapted from Liou (2002).

1.1.2 Absorption

The absorption is a process of attenuation of the incident radiation, and it depends on what happens to the structure of atoms or molecules after the interaction with the incoming radiation. The information provided next mostly comes from Liou (2002) and Wallace and Hoobs (2006).

The radiation transports a discrete quantity of energy through the well-known quantization rule of photons, and also atoms and molecules have quantized energy states. Then, the energy of impinging radiation is absorbed only if it is equal to the energy required to jump from a lower energy state to a higher energy state. The absorption is strictly correlated with the emission because the higher energy state is often unstable, and this causes a sudden jump again into the lower energy state with the consequent release of a photon.

The (quantized) phenomena that can happen to an atom or a molecule after the absorption is, in decreasing order of energy required: an electronic transition, a vibrational transition and a rotational transition. The electronic transition involves the promotion of

electrons to a higher energy state. This transition requires high-energy photons and is usually associated with absorption/emission of ultra-violet and visible radiation. In case of very energetic radiation, photo-ionization of atoms or photo-dissociation of molecules may happen. If radiation's energy is not enough to reach a higher energy state, vibrational and rotational transitions are still possible for molecules only. Vibrational modes are associated with near-infrared and medium-infrared radiation, and rotational modes with far-infrared and microwave radiation. In most of the molecules can be found also coupled roto-vibrational modes in the infrared region.

Given a particle, to each available transition is associated a narrow *absorption line* in the line spectrum. The radiative active particles have many possible transitions, then showing a very dense spectrum of absorption lines. It is important to highlight that it is possible to have a broadening of these narrow absorption lines due to Doppler effect (i.e., particle's movement relative to radiation, called *Doppler broadening*) and particle's collisions (i.e., related to ambient pressure, called *pressure broadening*), leading to a complex line spectrum with overlapping absorption lines.

These concepts are important to assess the Raman scattering in which absorption and emission change the wavelength of received radiation.

Most of the common wavelengths used in lidars are not coincident with molecular absorption lines, so that molecular absorption is not an issue (Kovalev and Eichinger 2004). However, some instruments, like the ceilometer used in this study, use a wavelength in which atmospheric water vapor molecules can have a strong absorption and a proper correction may be necessary for some type of analysis (e.g., Wiegner and Gasteiger 2015; Wiegner et al. 2019). In Figure 1.3 is shown the decrease of solar radiation reaching the surface, due to the absorption of various atmospheric gases and to the molecular scattering.

1.1.3 Polarization

The state of radiation is described by the Stokes vector:

$$S = (I, Q, U, V)$$

where I is the total radiance intensity, Q is the radiance difference between horizontal and vertical polarization, U is the radiance difference between 45° to -45° linear polarization, and V is the circular polarization term. This latter term is equal to zero in pure Rayleigh scattering, and usually negligible in respect to linear polarization (e.g., Gassó and Knobelspiesse 2022). The terms in the Stokes vector follow the rule:

$$I^2 \geq Q^2 + U^2 + V^2 \quad (1.7)$$

with the equality holding for monochromatic coherent radiation (Liou 2002). The radiation's polarization follows from Equation (1.7) as:

$$p = \frac{\sqrt{Q^2 + U^2 + V^2}}{I} \quad (1.8)$$

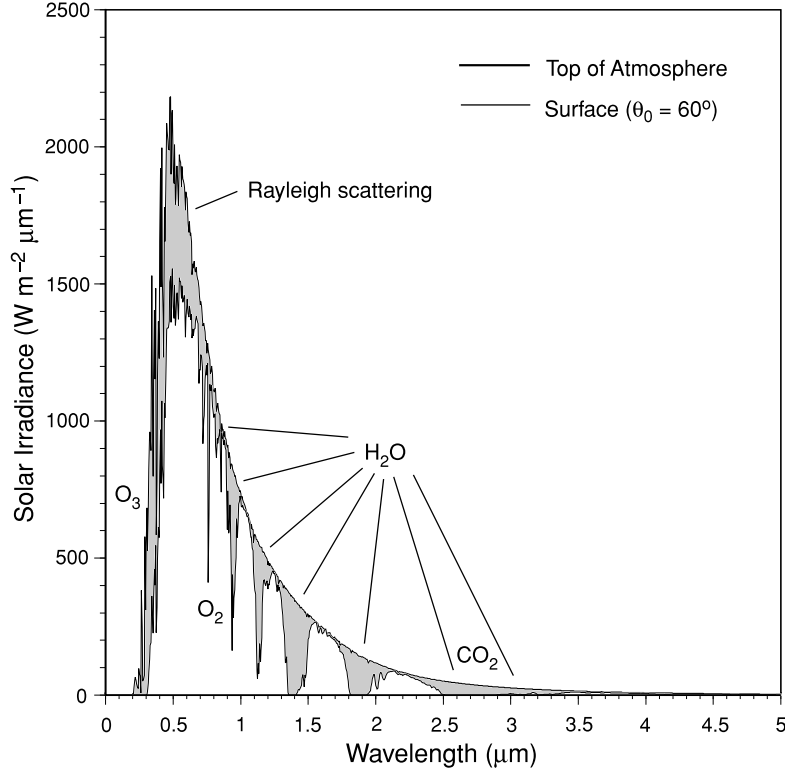


Figure 1.3: Extinction of solar radiation for an atmosphere without aerosols and clouds. Absorption happens in the whole solar spectra due to various atmospheric gases. The visible region extinction is dominated by Rayleigh scattering. Adapted from Liou (2002).

with $p = 0$ holding for completely unpolarized radiation, as is the case for natural light (Liou 2002).

Once radiation is emitted, the vector S is constant unless there is any interaction with an optical element (e.g., filters, molecules or particles). If radiation interacts with an optical element, the Stokes vector changes accordingly to:

$$S_{out} = M \cdot S_{in} \quad (1.9)$$

where M is the 4x4 scattering phase matrix (or Mueller matrix) defined as:

$$M(\theta) = \begin{pmatrix} M_{11} & M_{12} & 0 & 0 \\ M_{21} & M_{22} & 0 & 0 \\ 0 & 0 & M_{33} & M_{34} \\ 0 & 0 & M_{43} & M_{44} \end{pmatrix} \quad (1.10)$$

In the approximation of an isotropic and symmetric medium it holds $M_{21} = M_{12}$ and $M_{43} = -M_{34}$.

The scattering matrix M depends upon the scattering angle θ between the incident as scattered directions. These two directions define also the scattering plane.

The degree of linear polarization (DoLP) is defined starting from Equation (1.8) and discarding the negligible V component, as (L. Li et al. 2014):

$$DoLP = \frac{\sqrt{(Q^2 + U^2)}}{I} = -\frac{M_{12}(\theta)}{M_{11}(\theta)} \quad (1.11)$$

and the angle of polarization (AoP) is (L. Li et al. 2014):

$$AoP = \frac{1}{2} \arctan \left(\frac{U}{Q} \right) \quad (1.12)$$

These two derived quantities are widely used both for calibration (Z. Li et al. 2010; Z. Li et al. 2018) and for inversion (dos Santos Oliveira et al. 2023). Moreover, for most atmospheric aerosols, the dominant polarized scattering term is the DoLP. This quantity contains high sensitivity to aerosol size, refractive index, and shape (Mishchenko and Travis 1997), representing the most sensitive observable for micro-physical retrievals.

1.2 Atmospheric aerosols

The atmospheric aerosols are defined as suspension of fine solid or liquid particles in a gas (i.e., atmosphere) that range in size from a few nanometers (nm) to tens of micrometers (μm) (Seinfeld and Pandis 1998). The aerosols can be emitted by *natural* or *anthropogenic* sources. Moreover, depending on their history, they can be divided into *primary*, if they are directly emitted from the source, or *secondary*, if they are obtained from modification happening to the primary aerosols.

Considering the way in which aerosols interact with atmospheric radiation, either short-wave or long-wave, different behaviors affecting the Earth's energy budget can be distinguished. There is a *direct* effect due to scattering of incoming solar radiation, an *indirect* effect due to modification to cloud processes (i.e., formation, lifetime, etc.) and surface albedo (i.e., snow darkening), but also a *semi-direct* effect due to absorption of shortwave radiation (IPCC 2023).

In general, the aerosols provide a negative radiative forcing, helping to mitigate the positive radiative forcing due to greenhouse gases. The aerosol effective radiative forcing (AERF) can be quantified in -1.3 W m^{-2} (medium confidence), of which -1 W m^{-2} due to aerosol-cloud interactions and -0.3 W m^{-2} due to aerosol-radiation interactions (IPCC 2023). At the same time, the AERF is one of the energy budget's components with the higher uncertainty, although in the 6th IPCC report the uncertainty decreased respect the previous report, as shown in Figure 1.4.

The aerosol absorption by black carbon is known to be one of the largest contributors (Brasseur and Jacob 2017), together with carbon dioxide and methane, for atmospheric heating, but significant challenges still exist to assess the effects of absorbing aerosols. The IPCC AR6 reports about 50 % reduction of black carbon warming effects, but polarimetric observations from satellite suggest an underestimation of aerosol absorption in models (C. Chen et al. 2022).

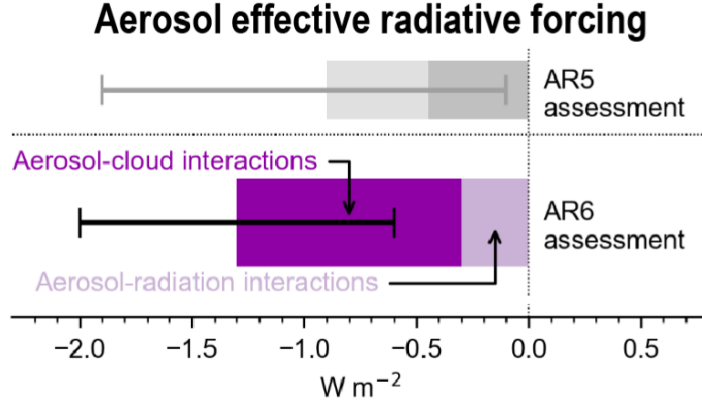


Figure 1.4: Aerosol effective radiative forcing (AERF) in the 6th IPCC report (AR6). Total AERF uncertainty decreased respect previous AR5 report, but the total AERF negative feedback increased. Extracted from Figure TS.15 in IPCC (2023).

The classification of the aerosols can be done on different basis. Aerosols can be classified by species, with the common classes being sulfates, nitrates, black carbon, organic aerosol, mineral dust, and sea salt, each one associated to a different response in terms of size, source, lifetime and interaction with radiation (Seinfeld and Pandis 1998). Alternatively, aerosols can be classified by their diameter size based on the main modes found in the particle size distribution (see Section 1.2.1), into Aitken mode ($<0.1\ \mu\text{m}$), accumulation mode ($0.1\ \mu\text{m}$ to $1\ \mu\text{m}$), and coarse mode ($>1\ \mu\text{m}$) (Seinfeld and Pandis 1998).

Is important to remarks that, in real cases, observing aerosols means observing a mixture of particles, so it makes sense to talk about predominant characteristics of the aerosol. As shown in Figure 1.5, different type of aerosols are associated to different micro-physical and optical properties, which are described next.

1.2.1 Particle size distribution

Particle size is one of the most critical parameters for characterizing aerosol behavior, as many physical and optical properties depend on it. To describe the size distribution quantitatively, let $n(r)$ denote the *particle size distribution* (PSD), where $n(r)dr$, represents the number of particles per unit volume with radii between r and $r + dr$. The average and integral properties of the PSD can be expressed in terms of the raw i^{th} moment of the distribution (Grainger 2025):

$$M_i = \int_0^\infty r^i n(r) dr \quad (1.13)$$

Because aerosols radius typically span over a range of five order of magnitude (i.e., $10^{-3}\ \mu\text{m}$ to $10^2\ \mu\text{m}$) (Brasseur and Jacob 2017), a log-normal distribution is usually the

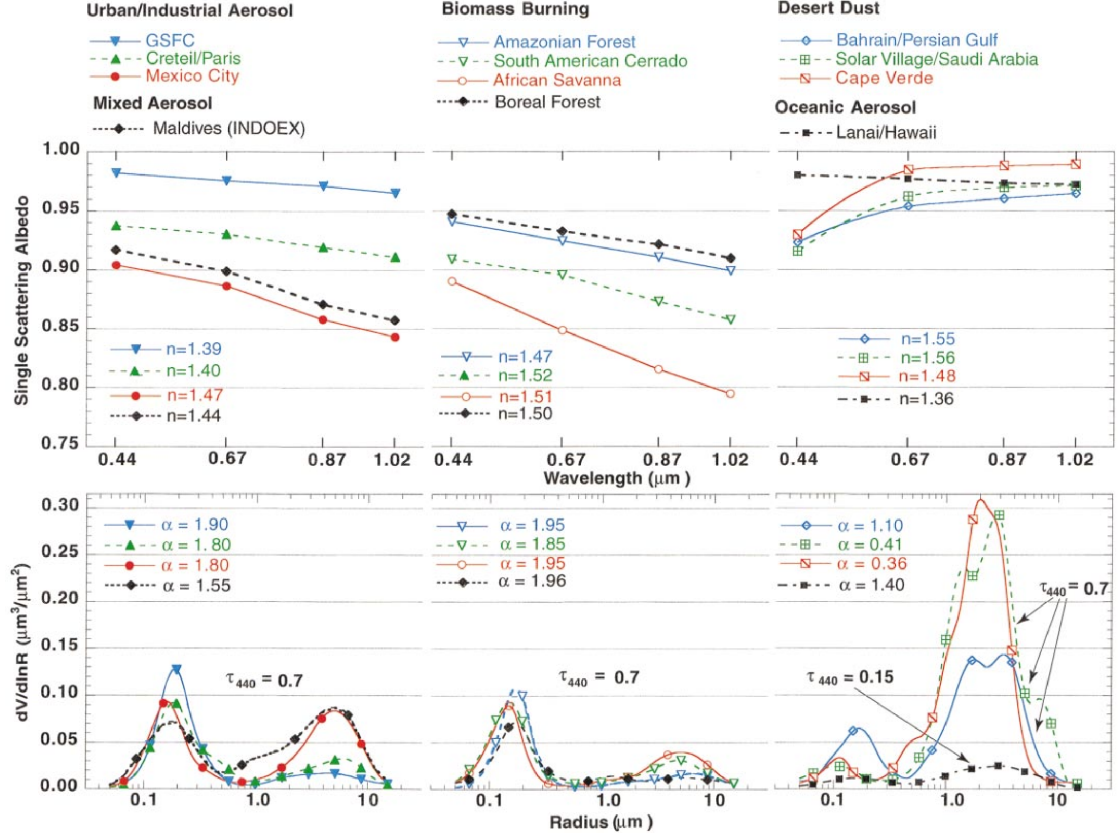


Figure 1.5: Characteristics of various aerosol types: urban, biomass smoke, desert dust, mixed, and oceanic. Colors and line styles show different sites around the globe highlighting the variability in aerosol properties inside each aerosol type. *Upper:* single-scattering albedo at different refractive index (n) values. *Lower:* particle size distribution at different Angstrom exponent (α) values. To ensure comparability, the size distribution is retrieved when optical depth (τ) is about 0.7, except for the oceanic type. Adapted from Dubovik et al. (2002).

best suited distribution to represent aerosols, and has the form (Grainger 2025):

$$n(r) = \frac{N_0}{\sqrt{2\pi}} \frac{1}{\ln s} \frac{1}{r} \exp -\frac{\ln^2(r/r_0)}{2 \ln^2 s} \quad (1.14)$$

where N_0 is the number of particles per unit volume, r_0 is the median radius and s is the geometric standard deviation. s is equal to 1 for *monodisperse* aerosol (i.e., only one particle size), and 1.5 to 2 for the typical *polydisperse* case found in atmosphere (Grainger 2025). Other commonly used distributions are the Junge power law, the Gamma distribution, and the modified Gamma distribution.

The PSD expresses the predominant characteristics observed in the aerosol. Typically, PSD can be described by a multi-modal log-normal distribution, combining fine and coarse modes (Torres et al. 2014), or by a binned distribution as in AERONET. In Figure 1.5

(lower plot) there are examples of binned PSD for different type of aerosols.

Finally, some important properties of the PSD can be derived accounting for Equation (1.13) and Equation (1.14) as described in Grainger (2025). The total particle number concentration per unit of volume:

$$N_0 = M_0 = \int_0^\infty n(r)dr \quad (1.15)$$

the total surface area concentration (for spherical particles):

$$A_0 = 4\pi M_2 = 4\pi \int_0^\infty r^2 n(r)dr \quad (1.16)$$

the total volume area concentration (for spherical particles):

$$V_0 = \frac{4}{3}\pi M_3 = \frac{4}{3}\pi \int_0^\infty r^3 n(r)dr \quad (1.17)$$

and the effective radius (for spherical particles):

$$r_{eff} = \frac{M_3}{M_2} = \frac{\int_0^\infty r \cdot \pi r^2 n(r)dr}{\int_0^\infty \pi r^2 n(r)dr} \quad (1.18)$$

The importance of the effective radius comes from the fact that energy removed from a light beam by a particle is proportional to the particle's area (provided the radius of the particle is similar to, or larger than, the wavelength of the incident light) (Grainger 2025).

1.2.2 Refractive index

The complex *refraction index* (m) gather the scattering (real part, m_r) and absorption (imaginary part, m_i) properties of a particle (Liou 2002), and can be written with the common notation:

$$m = m_r + im_i$$

The complex refractive index depends on the composition of aerosol particles (Dubovik et al. 2002).

The real part (RRI) influences the scattering efficiency and the shape of phase function. In general, m_r changes slightly with wavelength, so it is retrieved with less uncertainty by means of inversion (Dubovik and King 2000; Dubovik et al. 2006). The imaginary part (IRI) influences the magnitude of the single-scattering albedo (Dubovik et al. 2002). In this case, m_i can change significantly depending on the wavelength and aerosol type. The imaginary part by means of inversion has a bigger uncertainty respect the real part (Dubovik and King 2000).

In case of hygroscopic aerosols, the particle grows in radius and the refractive index tends to the one of water (i.e, $m_{H_2O} = 1.33 + i0$) as:

$$m_{wet} = \frac{m_{dry} + m_{H_2O}(g_{RH}^3 - 1)}{g_{RH}^3}$$

where g_{RH} is the hygroscopic growth factor (Zieger et al. 2013).

1.2.3 Single-scattering albedo

The single-scattering albedo (SSA) represents the percentage of a light beam that undergoes the scattering event (Liou 2002). SSA (ω_0) is defined as the ratio of scattering coefficient over extinction coefficient:

$$\omega_0 = \frac{k_s}{k_e}$$

Using the single-scattering albedo is it possible to infer if radiation undergoes mainly scattering or absorption processes, so that SSA determines if aerosol radiative forcing is positive (i.e., warming) or negative (i.e., cooling) (Dubovik et al. 2002). In Figure 1.5 (upper plot) there are examples of spectral SSA for different type of aerosols.

1.2.4 Phase function

The phase function defines the angular distribution of the scattering process. It is defined as the ratio of local-to-total scattering coefficients (Wallace and Hoobs 2006):

$$p(\theta, m, \lambda) = \frac{dk_s(\theta)}{k_s} \quad (1.19)$$

where θ is the scattering angle, m is the refractive index and λ is the wavelength. Moreover, the phase function is nothing but the term M_{11} in Equation (1.10). As the size parameter increases, the radiation become more forward scattered as depicted in Figure 1.1. Then, the phase function is obtained plotting the intensity of radiation as a function of the scattering angles, as in Figure 1.6.

The phase function, together with the AOD, determines the magnitude of the aerosol radiative forcing (Dubovik et al. 2002), and its knowledge is fundamental to retrieve correctly the optical properties in the inversion procedure explained in Section 6.3.

1.3 Atmospheric water vapor

Water vapor is a trace atmospheric constituent exerting a significant influence on both direct and indirect radiative forcing. Its distribution, however, exhibits pronounced spatial and temporal variability. The atmospheric water vapor content is primarily governed by temperature dynamics, as temperature controls processes such as evapotranspiration and turbulent mixing. The fact that approximately 99 % of the total atmospheric water resides below the tropopause (Sommer et al. 2023) highlights the strong coupling between water vapor and surface processes. Additionally, long-range transport and the air mass history substantially contribute to determining water vapor concentrations, particularly over continental regions (Wagner et al. 2006).

The contribution to the atmospheric absorption by water vapor is clear in Figure 1.3. In the near-infrared region the water vapor is the main absorbing molecules, with contributions from CO₂ around 2.7 μm (Liou 2002).

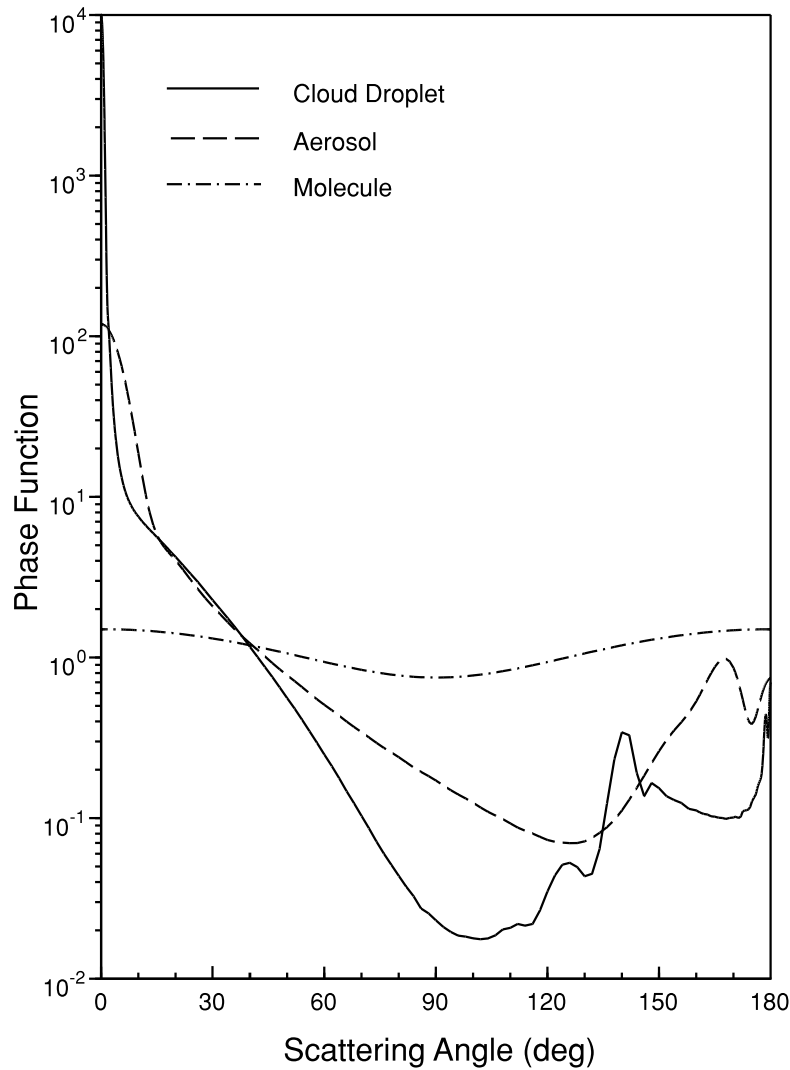


Figure 1.6: Normalized phase function at 500 nm wavelength and different scattering angles, for main atmospheric particles: cloud droplets ($10 \mu\text{m}$), aerosols ($1 \mu\text{m}$), and molecules ($10^{-4} \mu\text{m}$). Adapted from Liou (2002).

The content of atmospheric water vapor can be assessed by a variety of active/passive instruments either ground/space based. Given the water vapor high variability, the key point for a proper analysis is the ability to continuously and frequently monitor changes of water vapor in the atmosphere. In Chapter 4 the total atmospheric water vapor content is assessed at various time scales, ranging from inter-annual to daily variability, and using different instruments.

Chapter 2

Site and Observation networks

2.1 L'Aquila Atmospheric Observatory

The L'Aquila Atmospheric Observatory (LAO) is located west of L'Aquila town in the bottom of Middle Aterno valley, with an average height of 700 m above sea level (ASL). The whole area extends within the central part of the Italian Apennines mountain range, with mountains surrounding the site reaching up to 2000 m ASL, and the highest peak in Central Apennines (i.e., about 3000 m ASL) at just 20 km from the site. From a climatic point of view, the area is classified as a *temperate* climate (ranging between *Cfa* and *Cfb* classes of Köppen–Geiger climate classification), with cold winters and warm summers (CCCS 2019; Curci et al. 2021). Because of its orography, the area is often affected by thermal inversions, both on summer and winter. The precipitations are higher in autumn and winter respect to the summer but without very large differences in the amount along the year. The annual mean precipitation is about 650 mm/year (1991-2020 average) (Prof. Gabriele Curci, University of L'Aquila, personal communication), but with a higher amount in the northern part of the valley. In general, the climate within the Apennines valleys is drier respect other sites outside the Apennines mountain range (Scorzini and Leopardi 2018), and because of the general atmospheric circulation, the westward ridges are even drier respect to the eastward ones. Moreover, the area is affected by the Mediterranean circulation in the upper levels with intrusion of Saharan dust, more frequently between spring and summer (Mona et al. 2006), or when synoptic conditions favor the northward transport of dust from North Africa (Gaetani et al. 2012). The predominant land cover types in the valley are pastures and crops, except for the urban area of L'Aquila city which comprises also industrial and commercial warehouses in its surroundings.

The LAO consists of two separated sites located about 3 kilometers apart. The first site is more focused to aerosols observation and is located at the University of L'Aquila (42.37° N, 13.35° E, 680 m ASL), in a suburban context surrounded by commercial and industrial warehouses. In this site there are a multi-wavelength Raman lidar, a sun-sky-lunar photometer (see Section 3.2), UV sensors, and a weather station. The second site



Figure 2.1: Instruments at LAO which are used in this work, notably the photometer CimeL CE318-TP9 (left) and the ceilometer Vaisala CL51 (right).

is more focused on clouds observation and is located at Casale Calore place (42.38° N, 13.31° E, 683 m ASL), in a rural context surrounded by pastures and fields. In this site there are a cloud Doppler lidar, a ceilometer (see Section 3.3), a Doppler wind lidar, a weighting rain-gauge, a disdrometer, and a weather station. In addition to them, this site is also used to perform ozone-soundings, about twice a month. All the instruments deployed at those sites and their management, are the joint effort of University of L'Aquila, CETEMPS, and the ongoing collaboration with the Italian National Research Council (CNR). In Figure 2.1 are shown the instruments at LAO which observations are used in the following chapters.

It is worth pointing out that the distance between the two sites may pose an issue to the assumption of spatial co-location used to develop the synergy of photometer and ceilometer explained in Chapter 6. Especially in complex orography, the distance between sites may cause an uneven distribution of aerosol and clouds because of local breeze. Indeed, in the context of ACTRIS (see Section 2.2.2) the spatial co-location assumption is valid only up to about 1 km of distance between the instruments (Dr. Doina Nicolae, National Institute for Research and Development for Optoelectronics, personal communication).

2.2 Observational networks

Currently, many networks exist for the atmospheric observation, either at European or global scale. These networks may have different focus, either on providing high quality data for long-term studies, or near-real time observations to support weather forecasts. Participating in such networks allow taking advantage of the specific protocols and data processing algorithms developed by each of them. In the following two sections some of these networks are introduced, highlighting which of them include instruments from LAO site.

2.2.1 Photometer-related networks

Focusing on photometers, networks usually include only instruments from a specific manufacturer.

One of the most known network is the AErosol RObotic NETwork (AERONET) ¹ (Holben et al. 1998), developed by the National Aeronautics and Space Administration (NASA) in the '90s relying on the CE318 photometer built by the *Cimel Electronique* manufacturer (see Section 3.2.4). The AERONET network was born with the aim to provide standardized ground-based observations for validation of satellite remote sensing programs, and is able to provide near real-time data AOD data (Holben et al. 1998). Along the years, the data processing applied to raw data increased, improving the cloud screening algorithm (Giles et al. 2019; Smirnov et al. 2000), quality control and quality assurance (Giles et al. 2019; Holben et al. 2006), maritime operations (Smirnov et al. 2009), lunar processing (A. Barreto et al. 2013; Berkoff et al. 2011) among them. Photometers installed at LAO (see Section 3.2.4) belong to AERONET network since 2015.

Another network is SKYNET ² (Nakajima et al. 2020; Takamura et al. 2004), developed by the joint effort of different Japanese institutions, that become a network in the early 2000s. In SKYNET it is used the POM photometer built by the *Prede Co. Ltd.* manufacturer. The operating principle of the POM photometer is the same as the CE318, but differences arise in the wavelength and number of filters, and in the hardware design (e.g., the POM optical head is temperature controlled); moreover, it also exists specific versions of the photometer that allow for lunar and ship-based observations. The most important differences lie in data processing algorithms. The original code from Nakajima et al. (1996) was continuously developed along years, with the inclusion of improved on-site calibration (Campanelli et al. 2004), specific cloud screening (Khatri and Takamura 2009; Momoi et al. 2020), improved retrieval scheme (Hashimoto et al. 2012; Kudo et al. 2021), and lunar processing (Uchiyama et al. 2019).

Finally, the Global Atmosphere Watch Precision Filter Radiometer (GAW-PFR) network ³ (WMO 2005), developed under the recommendations of the Global Atmosphere

¹<https://aeronet.gsfc.nasa.gov/>, accessed February 1, 2026.

²<https://www.skynet-isdc.org/>, accessed February 1, 2026.

³<https://www.pmodwrc.ch/en/world-radiation-center-2/worcc/gaw-pfr/>, accessed February 1, 2026.

Watch (GAW) programme established by the World Meteorological Organisation (WMO) in the early '90s, is a network specifically aimed to ensure long-term stable and reference observations of AOD, through the use of the Precision Filter Radiometer (PFR) photometer built by the *World Radiation Centre* (WRC). These instruments are deployed at twelve GAW core-sites around the world, in addition to other sites belonging to other networks or institutions (Kazadzis et al. 2018). A tight quality control/quality assessment is a key point of this network. In the calibration center at PMOD World Radiation Centre in Switzerland, there is a reference set of three PFRs (i.e., called *PRF Triad*) managed by the World Optical Depth Research and Calibration Centre (WORCC), that is used to compare AOD against operational instruments, and check that the difference is within the tolerance established by WMO (WMO 2005). In 2014, a specific version of PFR was designed to allow also for lunar observations.

2.2.2 Ceilometer-related networks

In contrast to photometer case, the networks which include ceilometers are not stick to a specific instrument, except for smaller national-level networks.

E-PROFILE ⁴ is the European network for continuous atmospheric profiling developed within the framework of EUMETNET, the consortium of European National Meteorological Services, and it represents a core component of the EUMETNET Composite Observing System (EUCOS). The primary objective of E-PROFILE is to provide harmonized, near-real-time vertical profiles of key atmospheric variables to support numerical weather prediction (NWP) across Europe through a network of ground-based atmospheric profilers composed by automatic lidar and ceilometer (ALC), radar wind profilers and Doppler wind lidars, and microwave radiometers (Illingworth et al. 2019). The ceilometer at LAO (see Section 3.3.1) contributes to E-PROFILE network since 2023.

The Aerosols, Clouds and Trace Gases Research Infrastructure (ACTRIS) ⁵ is a European distributed research infrastructure dedicated to the long-term, high-quality and harmonized observation of atmospheric aerosols, clouds and short-lived trace gases. Inside ACTRIS, the *Centre for Aerosol Remote Sensing* (CARS) and the *Centre for Cloud Remote Sensing* (CCRES) play a key role in the coordination and harmonization of ground-based remote sensing observations. CARS is responsible for aerosol remote sensing activities, with a focus on lidar, ceilometer, and photometer systems. Instead, CCRES focuses on the remote sensing of clouds and cloud-related atmospheric processes, supporting both process studies and model evaluation. Within CCRES, ceilometers are primarily used to detect cloud base height, cloud occurrence and vertical cloud structure, providing essential information to complement radar radar-based observations. Although the University of L'Aquila is not yet officially affiliated with ACTRIS, the instruments at LAO are part of ACTRIS network since 2025.

⁴<https://eumetnet.eu/observations/vertical-profile-observations/>, accessed February 1, 2026.

⁵<https://www.actris.eu/about/actris-brief>, accessed February 1, 2026.

Chapter 3

Instruments

3.1 Introduction

Remote sensing instruments, such as photometers and ceilometers, offer complementary information: photometers provide column-integrated aerosol observations, while ceilometers give vertically-resolved backscatter profiles. The synergy between these instruments can be enhanced through advanced inversion algorithms. In Chapter 6 is assessed the retrieval of aerosol properties using a synergy of these instruments through an inversion algorithm, as shown in Román et al. (2018), especially with a focus on the integration of lunar photometer observations.

3.2 Sun-sky-lunar photometer

This type of instrument is a ground-based interference filter radiometer. The term photometer comes from the fact that the radiation observed is centered in the visible spectrum. The photometer is equipped with multiple narrow-band filters to collect radiation at specific wavelengths, and with step motors to point accurately a source of light and tracking its movement in the sky. With this kind of instrument is possible to measure both the direct irradiance and the diffused radiance. In the first case, what is observed is the attenuation due to a medium between the photometer and the source of light, in the second case, the distribution of the atmospheric radiance at different scattering angles. For what concern the Earth's atmosphere, the source of light can be the Sun or the Moon (see Section 5.3), and the extinction is caused by aerosol and atmospheric gases.

When observing the direct irradiance, the total attenuation of the medium can be obtained from the *Beer-Lambert-Bouguer* law:

$$F_{\lambda} = F_{0,\lambda} \cdot \exp^{\tau_{T,\lambda} \cdot m} \quad (3.1)$$

where F is the monochromatic irradiance at ground, F_0 is the monochromatic exoatmospheric irradiance, τ_T is the monochromatic *total optical depth* (TOD), and m is the

Table 3.1: Relevant contributions to extinction in each filter for Cimel photometers. Adapted from Giles et al. (2019).

Wavelength [nm]	Bandpass [nm]	Extinction by
340	2	Rayleigh, NO_2 , O_3
380	2	Rayleigh, NO_2
440	10	Rayleigh, NO_2
500	10	Rayleigh, NO_2 , O_3
675	10	Rayleigh, O_3
870	10	Rayleigh
940	10	Rayleigh, aerosol
1020	10	Rayleigh, H_2O
1640	25	Rayleigh, H_2O , CO_2 , CH_4

optical air mass accounting for a slanted path. The monochromatic irradiance is a flux density in unit of $W m^{-2} nm^{-1}$, from a physical point of view.

For each wavelength is necessary to account for all contributions to extinction of radiation due to particles other than aerosols. Indeed, the TOD in Equation (3.1) is the sum of different contribution, namely the gaseous absorption τ_g , the molecular (i.e., Rayleigh) scattering τ_R , and the aerosol scattering and absorption τ_a , so that:

$$\tau_T = \tau_R + \tau_a + \tau_g \quad (3.2)$$

Depending on the wavelength, one or more of these contributions take part to the extinction process, as shown in Table 3.1.

Moving from the Beer-Lambert-Bouger law to the real instrument, the irradiance is measured by the instrument's signal (V), expressed in terms of digital counts (DC), so that the Equation (3.1) become:

$$V_\lambda = \frac{V_{0,\lambda}}{d_{SE}^2} \cdot \exp^{\tau_{T,\lambda} \cdot m} \quad (3.3)$$

where d_{SE} is the current-to-average ratio of Sun-Earth distance in astronomical units (au) that account for variations of solar irradiance due to the orbit.

The $V_{0,\lambda}$ is the instrument's calibration constant obtained by means of the Langley plot calibration method (Shaw 1983; Shaw et al. 1973), which involve plotting the logarithm of observed V_λ against the aerosol optical air mass, and extrapolating $V_{0,\lambda}$ at air mass $m = 0$, so that Equation (3.3) can be written as:

$$\ln(V_\lambda \cdot d_{SE}^2) - \tau_{atm} \cdot m_{atm} = \ln(V_{0,\lambda}) - \tau_{a,\lambda} \cdot m_a \quad (3.4)$$

where subscript *atm* account for all contributions except aerosols, and subscript *a* account for aerosols. Given Equation (3.4), the term $\ln(V_{0,\lambda})$ is nothing but the intercept of the

linear regression over the points in the Langley plot. This kind of calibration requires stable conditions along the time required to encompass a number of air mass values (i.e., commonly between 7 and 2, or 5 and 2) (Toledano et al. 2018), so it can be performed with high accuracy only at high-altitude and low latitude sites, to ensure pristine conditions and fast changing air mass.

To avoid these constrain, usually the field instruments are calibrated by means of intercomparison against a master instrument calibrated as above. The spectral field instrument's calibration constant is obtained by the ratio of measured field (V_{λ}^{field}) and master (V_{λ}^{master}) signals:

$$V_{0,\lambda}^{field} = V_{0,\lambda}^{master} \cdot \frac{V_{\lambda}^{field}}{V_{\lambda}^{master}} \quad (3.5)$$

The uncertainty of the calibration is estimated in 0.5 % and 1-2 % for master and field constants respectively (Holben et al. 2006).

Finally, it is necessary to consider also the spectral *temperature response function* of the instrument and apply a suitable correction. In general, this response function is relevant for the 1020 nm filter with silicon detector, producing an uncertainty in AOD of ± 0.01 with extremely cold or hot temperatures (Giles et al. 2019), but in S. Chen et al. (2021) is shown how the 1640 nm InGaAs channel requires even higher correction than 1020 nm channel. In AERONET, starting from algorithm Version 3, the temperature correction is applied to all filters except for UV channels; this correction depends both on the sensitivity of the detector and the material of the filter, and must be characterized uniquely for each photometer (Giles et al. 2019).

3.2.1 Aerosol optical depth

Considering the relevant absorbing gases and their columnar abundance, it is possible to retrieve the monochromatic extinction due to aerosols along the vertical path, called *aerosol optical depth* (AOD, $\tau_{a,\lambda}$), as:

$$\tau_{a,\lambda} = \frac{\ln(V_{0,\lambda}) - \ln(V_{\lambda}) - m_{atm} \cdot \tau_{atm,\lambda}}{m_a} \quad (3.6)$$

where $m_{atm,\lambda}$ and τ_{atm} are the air mass and monochromatic optical depth for atmospheric particles other than aerosols, respectively, and m_a is the aerosol air mass. The computations for m_{atm} and τ_{atm} , for each specific particle contribution, and for m_a , are described in Giles et al. (2019) and not discussed further.

Because of the angular characteristics of aerosol scattering discussed in Section 1.1, it is worth to notice how high values of AOD tend to introduce a bias due to forward scattering in AOD observations. This bias is usually below the instrument absolute uncertainty of 0.01, but may be not negligible when high AOD from large diameter aerosol is observed at large zenith angles (Sinyuk et al. 2012).

The AOD has a spectral dependency that reflects the different aerosols type and their optical characteristics. This spectral dependency is described by the *Angstrom exponent*

(AE, α) (Ångström 1929) with the following expression:

$$\alpha = -\frac{\ln\left(\frac{\tau_{a,\lambda_2}}{\tau_{a,\lambda_1}}\right)}{\ln\left(\frac{\lambda_2}{\lambda_1}\right)} \quad (3.7)$$

The value of α is an indicator of the predominant mode in the size distribution of aerosols, with values near to 0 for coarse aerosols, like desert dust, and values above 2 for fine aerosols, like fresh smoke (Eck et al. 1999).

3.2.2 Precipitable water vapor

For what concern the optical depth due to the water vapor column, the things become a little bit more complicated. The filters sensitive to water vapor are 940 nm, 1020 nm and 1640 nm, with the former used to derive the water vapor optical depth (τ_w) required to correct the other two channels. At 940 nm, because of the strong water vapor absorption band, the Equation (3.2) become:

$$\tau_T = \tau_R + \tau_a + \tau_w$$

Given the transmissivity as $T_i = \exp(-m_i\tau_i)$, the Beer law (Equation (3.3)) can be written as:

$$V_\lambda = \frac{V_{0,\lambda}}{d_{SE}^2} \cdot (T_R \cdot T_a \cdot T_w) \quad (3.8)$$

The water vapor transmissivity (T_w), or equivalently the water vapor optical depth (τ_w), can be obtained once AOD and Rayleigh contributions at 940 nm are known. The AOD_{940} is obtained using the Angstrom exponent (α) value and the AOD at known wavelength (λ_i) as:

$$AOD_{940} = AOD_{\lambda_i} \cdot \left(\frac{940}{\lambda_i}\right)^{-\alpha}$$

On the other hand, the water vapor transmissivity is related to the vertical column of water vapor (W) by a nonlinear relation (Halthore et al. 1997):

$$T_w = \exp[-a \cdot (m_w W)^b] \quad (3.9)$$

where a and b are two coefficients specific for each instrument obtained by means of radiative transfer simulations (Halthore et al. 1997). From Equation (3.9), the water vapor column abundance, in units of g cm^{-2} , is obtained trivially as:

$$W = \frac{1}{m_w} \left[\frac{-\ln(T_w)}{a} \right]^{1/b} \quad (3.10)$$

The water vapor column is usually expressed in units of cm and defined as *precipitable water vapor* (PW), by means of dividing Equation (3.10) by the water density value.

3.2.3 Sky radiance

In addition to the direct observation of the source of light, it is possible to measure also diffuse radiance scattered by the atmosphere, provided an appropriate scan geometry and detector sensitivity is used. Measuring the diffuse radiance at different scattering angles allow obtaining information about the average *phase function* in the air mass observed, and retrieving its optical and micro-physical properties using an inversion algorithm.

The scan geometry must be a compromise between encompassing as many as possible scattering angles, and a fast scanning to avoid changes in the air mass. The *almucantar* (ALM) geometry consists in fixing the zenith along the line of sight between the photometer and the source of light, then scanning along the azimuth angles. This geometry allows the detection of sky inhomogeneities simply comparing the pair of symmetric angles from the left and right branches. As example, in AERONET if the radiance pair exhibits a difference higher than 20 % between right and left branches, the pair is eliminated (Holben et al. 2006).

The *principal plane* (PPL) geometry consist in fixing the azimuth along the line of sight between the photometer and the source of light, then scanning along the zenith angles. This geometry does not present any evident symmetry and, therefore, the detection of sky inhomogeneities is not straightforward as for almucantar case (Torres et al. 2014).

The maximum scattering angle of these geometries depends on the zenith angle of the source of light (θ_0), and is equal to $\Theta_{max} = 2\theta_0$ for ALM and $\Theta_{max} = \theta_0 + 90^\circ$ for PPL (Nakajima et al. 1996). Using the ALM geometry when the source's zenith angle is small, greatly limits the range of scattering angles observed. Therefore, in AERONET the ALM is used only when $\theta_0 > 50^\circ$, that is a strong limitation in certain regions and/or seasons (Sinyuk et al. 2020). Despite the advantage of PPL for small zenith angles, it is necessary to account that any obstacle on the horizon may prevent any observation at very low elevation, so that the theoretical maximum view angle is not 90° as stated above, but it practically reduces to $75\text{--}80^\circ$ (Sinyuk et al. 2020; Torres et al. 2014).

To overcome this issue, the Cimel CE318-T is capable to perform a dynamic scan called *hybrid* (HYB) geometry (Figure 3.1), that is a mix of ALM and PPL geometries. Indeed, the HYB scan starts from zenith and azimuth of the source of light, then increase monotonically the scattering angle changing either these angles until reaching 15° of elevation, then continue at fixed zenith angle like a standard ALM.

3.2.4 Cimel CE318

At L'Aquila site there are currently two photometer: one is a sun-sky Cimel CE318-NE-DP, and the other is a newer sun-sky-lunar Cimel CE318-TP9, both with polarization capability. The CE318-NE-DP has been part of AERONET from April 2015 until May 2025 (instrument #856), when it was replaced by the CE318-TP9 (instrument #1452). Since then, the old CE318-NE-DP is used as a spare instrument activated only to cover the gaps due to the annual calibration of the new CE318-TP9.

The two model of instruments share the same setup, but the CE318-TP9, developed after the work of A. Barreto et al. (2013), has a better accuracy, both in pointing and

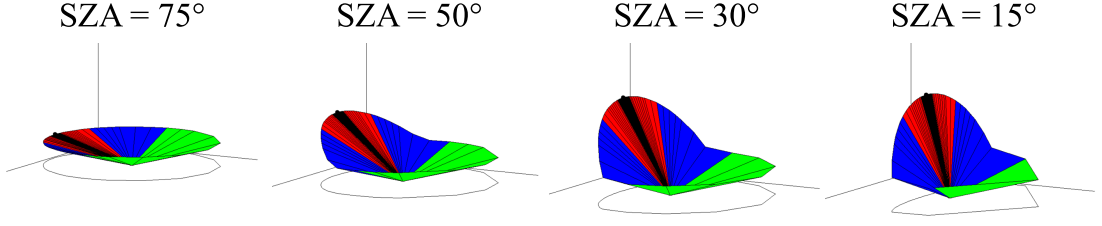


Figure 3.1: Hybrid scan geometry for Cimel CE318-T at different Sun zenith angles (SA). The colors represent the range of scattering angles (Θ) encompassed by the scan. Black: $0^\circ \leq \Theta < 6.5^\circ$, red: $6.5^\circ \leq \Theta < 31^\circ$, blue: $31^\circ \leq \Theta < 81^\circ$, green: $\Theta > 81^\circ$. Adapted from Sinyuk et al. (2020).

sensitivity, and the ability to track Moon's position.

The Cimel photometer is a multi-wavelength passive instrument capable to measure the atmospheric extinction in terms of total optical depth (TOD), and the sky radiance. The instrument is equipped with a wheel containing multiple filters at 340, 380, 440, 500, 675, 870, 937, 1020 and 1640 nm, with a bandwidth between 2 nm to 25 nm depending on the filter. The radiation passing through the filters is collected either by a silicon (340 nm to 1020 nm filters) or a InGaAs detector (1020 nm and 1640 nm filters). The double observation in the 1020 nm channel with both detectors allows tracking the consistency of observations made with the different detectors. In the polarized instruments, there is also an additional wheel containing three polarized filters, each at 120° respect the others.

In addition to the direct irradiance from Sun/Moon, the Cimel photometer is able to perform sky scanning with almucantar, principal plane, and hybrid geometries, among others. The hybrid geometry, as well as the Moon observation, are available only since the CE318-TP9 model. The direct irradiance measurement consists of a triplet of observations, separated by 30 seconds. The triplet method ensures an effective way for AOD's cloud screening as explained in (Giles et al. 2019). The sky scanning scenarios provide the radiance at different scattering angles as explained in the previous section.

The raw data from the instrument are processed by the Cimel PHOTOGETDATA software provided by the manufacturer. This software collects the flux of raw data, and it is able to save them as binary (.k7/.k8 format) or as text ASCII files. These files contain the observations for each measurement scenario that is enabled in the instrument's software.

To ensure homogeneous observations through time, as required by long term datasets, the instruments must be regularly calibrated. The calibration method depends on the observable. For irradiance, the Cimel photometers in AERONET are calibrated by means of intercomparison. The field instruments do co-located observations together with a master instrument. The calibration constant for the field instrument (C_{field}) is readily obtained from the ratio of the two observations and knowledge of the calibration constant for the master instrument (C_{master}) as already shown in Equation (3.5).

The master instruments at the calibration centers are regularly calibrated by means

of absolute calibration and intercomparison with reference photometers, at selected high altitude sites at Hawaii (USA) and Canary Island (Spain). Instead, the radiance from angular sky observations is calibrated using a calibrated integrating sphere as target. In such way, a calibration coefficient for each channel, and eventually each polarizer, is obtained.

3.3 Ceilometer

A ceilometer, commonly referred as *automatic lidar and ceilometer* (ALC), is a low-power lidar system that is eye-safe and can be operated unattended. It shares the same technology of a lidar, but with a simplified design accounting for only one wavelength.

Ceilometers usually operate in near-infrared band because it avoids strong Rayleigh scattering (Wiegner et al. 2014). In such way the backscatter signal is mainly due to larger particles like clouds and aerosols.

The ceilometer points to the zenith, and emits a laser beam that may be eventually scattered back from reflectors with an efficiency that depends on the reflector properties and the wavelength of the laser. The backscatter signal received (P) by the ceilometer is described by the lidar equation (Wiegner et al. 2014):

$$P(z) = C_L \frac{\beta(z)}{z^2} \exp \left[-2 \int_0^z \alpha(z') dz' \right] \quad (3.11)$$

where C_L is the calibration constant accounting system and optic characteristics, β is the backscatter coefficient in unit of $\text{m}^{-1} \text{sr}^{-1}$ and α is the extinction coefficient in unit of m^{-1} . The exponential term is the atmospheric transmission, and accounts for the double attenuation in path back and forth. Both backscatter and extinction coefficients come from the contribution of molecular and aerosol particles:

$$\begin{aligned} \beta &= \beta_m + \beta_p \\ \alpha &= \alpha_m + \alpha_p \end{aligned} \quad (3.12)$$

Because the signal decrease quadratically with the range, usually the return power in Equation (3.11) is range normalized. After such normalization, the signal is called *range corrected signal* (RCS), and it is defined as:

$$\text{RCS} = P(z) \cdot z^2 \quad (3.13)$$

With this quantity is easier to visualize the signal, but has the drawback to emphasize also the noise.

It is important to remark that the signal received at the instrument ($\hat{P}$) inherently includes also the background noise (P_{bg}) coming from the instrument electronics ($P_{bg,i}$) and ambient ($P_{bg,a}$). Then, it is necessary to characterize such contributions to retrieve the effective signal (P) as:

$$P(z) = \hat{P}(z) - P_{bg}(z) = \hat{P}(z) - [P_{bg,i}(z) + P_{bg,a}] \quad (3.14)$$

The instrumental noise can be range dependent, and is usually determined by means of dark measurements. The ambient noise is usually a constant shift of the signal, and can be corrected for each observation adjusting the signal in an aerosol-free region to the theoretical signal from molecules only.

The solution to Equation (3.11) can be written in respect to β_p or α_p using Klett-Fernald-Sasano method (Fernald 1984; Klett 1981; Sasano et al. 1985), with assumption on the Lidar ratio ($S_p = \alpha_p/\beta_p$) and computing the molecular terms from air density profiles (observed or assumed) (Wiegner et al. 2014). The solution using β_p is known to be more accurate, in particular in cases of low aerosol concentration and/or long wavelengths, because it is less sensitive to errors on the assumed lidar ratio (Wiegner et al. 2014).

Finally, the solution in terms of β_p (or α_p) can be found starting from a boundary value and proceeding iteratively (Wiegner et al. 2014):

$$\beta_p(z) = \frac{Z(z)}{N(z)} - \beta_m(z) \quad (3.15)$$

with

$$Z(z) = z^2 P(z) \exp \left[-2 \int_0^z (S_p(z') - S_m) \beta_m(z') dz' \right]$$

and

$$N_{forw}(z) = C_L - 2 \int_0^z S_p(z') Z(z') dz' \quad (3.16)$$

or

$$N_{back}(z) = \frac{z_0^2 P(z_0)}{\beta_m(z_0) + \beta_p(z_0)} + 2 \int_z^{z_0} S_p(z') Z(z') dz' \quad (3.17)$$

Hence, exists two possible ways to retrieve β_p , depending on the use of Equation (3.16) or (3.17), both still requiring the knowledge of the lidar ratio for aerosol (S_p) and molecules (S_m).

One way is the *forward* method, which use Equation (3.16), and it requires the knowledge of the instrumental calibration constant C_L . If such constant is known, the full overlap height between transmitted and received beams is small, and the α below full overlap height is small as well, then the β_p at overlap height is readily known, and it is possible to integrate upward to obtain the profile (Wiegner et al. 2014).

A second way is the *backward* method, which use Equation (3.17). Starting from the $\beta(z_0)$ value in a layer without aerosol, so that $\beta_p(z_0) \simeq 0$ and $\beta(z_0) = \beta_m(z_0)$ (e.g., Fernald 1984; Klett 1981). If the layer at z_0 is cloud-free and aerosol-free for enough time to obtain $\beta_m(z_0)$ by means of temporal integration, then it is possible to integrate downward to obtain the β_p profile (Wiegner et al. 2014).

A third option is available, and is based on the integrated backscatter (B) (Hopkin et al. 2019; O'Connor et al. 2004):

$$B = \int_{z_{low}}^{z_{top}} \beta_p(z) dz$$

In presence of non-iced fully-attenuating stratiform clouds, the lidar ratio at the cloud base is known to be $S_{strat} = 18.8 \pm 0.8$ sr, so the integrated backscatter can be defined as (O'Connor et al. 2004):

$$B = \frac{1}{2\eta \cdot S_{strat}} \quad (3.18)$$

where η is a factor accounting for multiple scattering. Using Equation (3.18), it is possible to estimate the calibration constant producing that B with an iterative procedure. Once the calibration constant is known, one of the forward methods shown above can be applied.

In research-grade multi-wavelengths lidars, using Raman technique and depolarization capability, is it possible to directly measure the lidar ratio without any kind of assumption to invert the problem. Instead, given the ceilometer characteristics, it is necessary to assume the lidar ratio and density profiles, or use observations from other instruments, to find the solution.

Another issue related to the ceilometer is the low signal-to-noise ratio (SNR) due to the low-power feature. Because of that and low sensitivity to Rayleigh scattering, observing the reference molecular backscatter $\beta_m(z_0)$ may be complex, and it requires long time (i.e., some hours) of stable conditions. Moreover, the noise from sunlight reduces further the SNR, so the signal can be used to retrieve aerosol properties profiles only in the first 4-5 km during daytime, and some kilometers more in nighttime (e.g., Heese et al. 2010).

3.3.1 Vaisala CL51

At L'Aquila site there is one ceilometer Vaisala CL51 manufactured by *Vaisala Oyj*. This instrument is a single-wavelength pulsed lidar, with an eye-safe low-power laser. This setup, allows for continuous and automatic operation.

The CL51 ceilometer emits a laser beam at 910 nm, with a pulse repetition frequency of 6.5 kHz and low energy pulses at 3 μ J. The nominal power of 19.5 mW is enough to detect strong reflectors, like clouds, up to 15 km. However, the aerosol signal is weaker and signal-to-noise ratio degrades rapidly with height, especially during day when there is the added noise from sunlight. Such kind of atmospheric background noise is adjusted automatically by the system and is transparent to the final user, but is still affect observations reducing the signal-to-noise ratio (Kotthaus et al. 2016). Although this correction, some ambient noise is still present and should be necessary to adjust the signal in aerosol-free regions (Bedoya-Velázquez et al. 2021). From our experience, aerosol layer can be detected well up to 3-4 km and 7 km, in daytime and nighttime, respectively.

The optical system is monoaxial, meaning that there is a single telescope to send and receive the laser signal. This setup allows for lower overlap height, so that the full overlap between the emitted beam and the receiver field-of-view is reached at around 200 m (Madonna et al. 2018; Wiegner et al. 2014). However, as the previous CL31 model, the CL51 has artifacts in the nearest range gates, so that the signal below 100 m AGL is commonly discarded (Kotthaus et al. 2016).

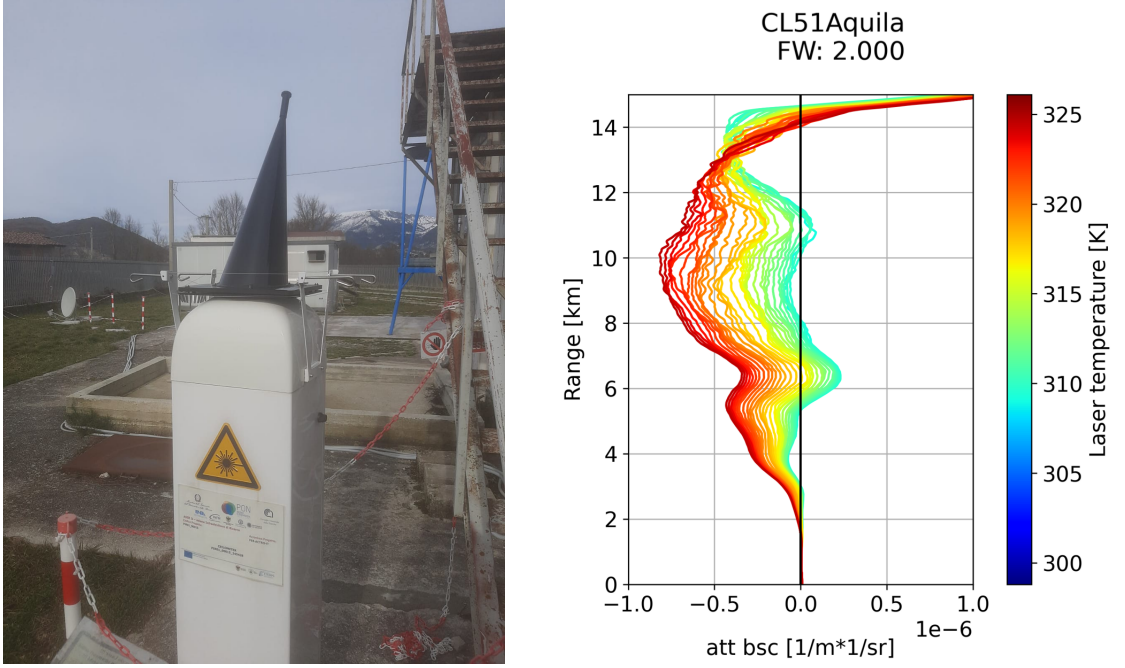


Figure 3.2: *Left:* Vaisala optical termination hood (black hat) over the L'Aquila CL51. *Right:* Attenuated backscatter signal from dark measurements at various laser temperatures for L'Aquila CL51 (from Mattis et al. (2024), courtesy of Dr. Alexander Geiß, LMU).

The range resolution is 10 m, and the current temporal resolution is 30 s. The instrument is tilted by 2° - 3° off zenith northward, as suggested in the manual and literature to avoid specular reflections from horizontal ice plate crystals.

One of the most critical point for CL51 is the water vapor absorption at 910 nm. Despite no being an issue for cloud base and mixing layer detection, it may cause large error if the CL51 is used to retrieve the backscatter coefficient profile. This error depends on the water vapor content and can up to 20 % at mid-latitudes, and even 50 % in the tropics (Wiegner and Gasteiger 2015). As a consequence of water vapor absorption, the extinction coefficient in Equation (3.12) must account also for that contribution:

$$\alpha = \alpha_m + \alpha_p + \alpha_w$$

Then, it is necessary to find a way to measure the effective transmission due to water vapor and true instrument wavelength, and account for such term in Equation (3.11) before moving to search for a solution. Different methods are available in literature, either using profiles from radiosonde (Wiegner and Gasteiger 2015) or microwave radiometer (Bedoya-Velásquez et al. 2021; J. Chen et al. 2025).

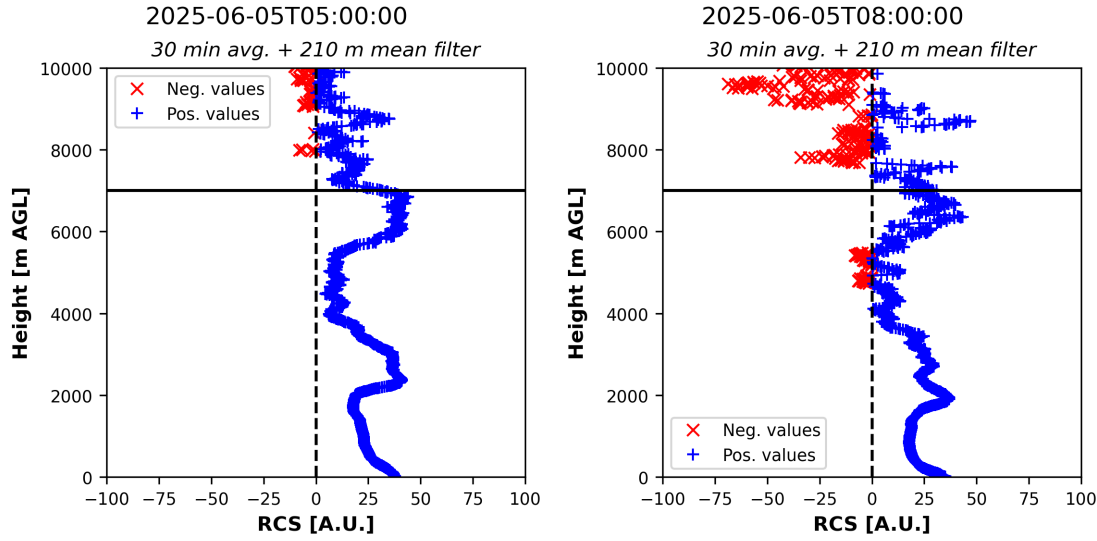


Figure 3.3: RCS signal on June 5, 2025 at two times in the morning (left: 5am UTC, right: 8am UTC; local time is UTC+2). The increase of temperature during daytime may eventually generate negative (i.e, unphysical) RCS values also in the lowest portion of atmosphere. The vertical dotted line shows the threshold between positive and negative RCS values. The horizontal solid line shows the height limit of 7000 m ASL used in Section 6.3.

3.3.1.1 Temperature sensitivity

From Kotthaus et al. (2016) was clear some temperature dependency in the background signal of Vaisala CL31 instruments. In ACTRIS-CARS there is an ongoing study about the temperature sensitivity of the signal in Vaisala CL51 instruments. The University of L'Aquila has been performing dark measurements (DM) along the years using a specific optical termination hood provided by the manufacturer (Figure 3.2, left). The termination hood fully attenuates the signal simulating a dark-sky condition, highlighting any noise or systematic bias in the signal due to the instrument itself.

The University of Munich (LMU), as the principal investigator, has obtained some preliminary results showing that higher temperature induces a range-dependent modification in the DM. In right Figure 3.2 (courtesy of Dr. Alexander Geiß, LMU) there is an example of such dependency for L'Aquila ceilometer. The instrument at LAO shows large changes in the signal, also thanks to the large range of temperature that can be observed at that site. Between 2 km to 13 km, the increase of laser temperature shifts the signal toward lower values.

In Figure 3.3 an example of how this sensitivity can affect RCS profiles. During daytime, the increase of temperature, hence of the laser, induces a shift of the profile toward negative values at about 5000 m. In aerosol-free condition, when the profile rapidly falls toward zero, the negative values may appear already at about 2–3 km (not shown).

For effective retrieval of aerosol optical properties, the correction of such background

is critical. At the moment no operative method to account for the temperature sensitivity exists, but one possible solution would be to perform DM for each instrument and generate a look-up table with a correction value depending on range and temperature (Mattis et al. 2024), or use a climatological monthly value from clear-sky nighttime observations (J. Chen et al. 2025; Kotthaus et al. 2016). Trivially, this correction is strictly bound to the specific hardware and firmware of the instrument, so any change to them will require different correction values.

In the analysis which follow, no background correction, neither for instrumental and atmospheric noises, has been applied to ceilometer data, so that the input profile used in Section 6.3 is just the smoothed RCS.

Chapter 4

Precipitable water vapor at LAO

4.1 Introduction

As described previously, the ceilometer at LAO operates in the near-infrared band. In particular, the 910 nm channel lies into a water vapor absorption band. As been demonstrated that errors in aerosol backscatter coefficient retrieval can be up to several percents for mid-latitudes, and even higher in the tropics, if water vapor absorption is ignored (Wiegner and Gasteiger 2015).

Because of lack of time, the water vapor correction described in Wiegner and Gasteiger (2015) has not been implemented. However, because the error due to water vapor absorption depends on the water vapor content in the atmosphere, it is possible to roughly estimate the order of water vapor influence on profiles from CL51 ceilometer once the average water vapor column at LAO is known.

The total column of atmospheric water vapor, or *precipitable water vapor* (PW), is defined as the total atmospheric water vapor contained in a vertical column of unit cross-sectional area extending between the Earth’s surface to the top of the atmosphere, and commonly expressed in terms of the height to which that water substance would stand if completely condensed and collected in a vessel of the same unit cross-section (AMS 2025). Because of that definition, the precipitable water vapor is usually expressed in units of kg m^{-2} or cm.

L’Aquila site is known to be relatively dry in respect to other sites outside the Apennines mountain range, as reported in Section 2.1, so it is possible to speculate about a low influence of the water vapor absorption. At first glance, the total water vapor content can be assessed from the climatological average PW provided by AERONET and reported in Table 4.1. This table shows different sites in Italy, and highlights how lower PW values are associated with higher altitude and higher distance from the sea.

Because of sun-sky photometer characteristics, the observations have a negative bias, the so called *dry-bias*, due to the fact that photometer measurements are done mostly in fair atmospheric conditions (i.e., not too many clouds, no rain) which are associated with lower PW values (e.g., Van Malderen et al. 2014). Regarding the AERONET algorithm

Table 4.1: Average value of precipitable water vapor from various site in Italy, ordered by type and altitude. For each site is reported the location (classified *inland* if more than 50 km from coast), the range of years used to compute the climatology, and the climatological precipitable water vapor (PW) in centimeters with its standard deviation (SD). Data from AERONET L2.0 V3 climatological tables.

Station	Type	Coordinates	Altitude mASL	Years	PW cm	SD
Lecce	Coastal	N 40°20', E 18°6'	30	2003-2024	1.81	0.61
Naples	Coastal	N 40°50', E 14°18'	50	2016-2024	1.85	0.72
Rome	Coastal	N 41°50', E 12°38'	130	2001-2025	1.68	0.61
Ispira	Inland	N 45°48', E 8°37'	235	1997-2025	1.57	0.77
L'Aquila	Inland	N 42°22', E 13°21'	683	2015-2025	1.27	0.54
Potenza	Inland	N 40°36', E 15°43'	770	2004-2025	1.21	0.46

for PW retrieval (see Section 3.2.2), despite the coefficients used to compute the water vapor transmissivity should be insensitive to different atmospheric profiles (Halthore et al. 1997), in Almansa et al. (2020) is reported how site-specific coefficients may help in reducing the systematic bias when comparing photometric PW observations against other reference observations.

Since the sun-photometer is the main instrument at LAO to derive continuous PW observations, in this chapter will be assessed the PW at different time scales and with different instruments, in order to evaluate how much photometric observations can be trusted.

4.2 Methods

The analysis of PW, was carried out using three independent observational techniques: sun-sky photometry, Global Navigation Satellite System (GNSS) observations, and radiosonde soundings. The use of these instruments allows the assessment of PW variability across different temporal scales, exploiting high-frequency continuous measurements in the photometry and GNSS cases, and vertically resolved atmospheric profiles for radiosonde case.

4.2.1 Instruments and datasets

Sun-photometer measurements are obtained using a Cimel CE318-NE instrument operating at the L'Aquila site since 2015 and belonging to the AERONET network. Columnar PW is retrieved from direct solar irradiance measurements in the 937 nm band using the Direct Sun Algorithm Version 3 which is explained in Section 3.2.2. Only Level 2.0 data are used in this study, ensuring cloud-screened, calibrated and quality-controlled

observations (Giles et al. 2019; Holben et al. 1998; Sinyuk et al. 2020). The resulting PW values are provided in centimetres.

GNSS-derived PW is obtained from zenith tropospheric delays of the GNSS signal for the period 1999 to 2022. The *zenith total delay* (ZTD) can be expressed as the sum of a hydrostatic component, the *zenith hydrostatic delay* (ZHD), and a wet component, the *zenith wet delay* (ZWD).

The ZHD is computed using the Saastamoinen model (Askne and Nordius 1987), which requires knowledge of surface pressure and station altitude. Since in situ pressure measurements were not available for the entire GNSS period, mean sea level pressure from the ERA5 reanalysis (CCCS 2018) was scaled to the station altitude using the following barometric formula:

$$P_{site} = P_{MSL} \left(1 - \frac{\Gamma}{T_0} \cdot z_{site} \right)^a$$

where $\Gamma = 6.5 \times 10^{-3} \text{ K m}^{-1}$ is the mean tropospheric lapse rate, $T_0 = 288.15 \text{ K}$ is the reference surface temperature, z_{site} is the mean sea level station altitude in m, and $a = 5.25588$.

Subtracting ZHD from the observed ZTD, it is possible to derive ZWD, that is finally converted into PW using a proportionality factor Π , so that $PW = \Pi \cdot ZWD$. The factor Π depends on the weighted mean atmospheric temperature T_m (Bevis et al. 1992; Bevis et al. 1994). In this study, Π is computed from a long-term radiosonde climatology of temperature and relative humidity, following the method of Basili et al. (2001), which provides reduced uncertainty compared to regressions based solely on surface temperature (Basili et al. 2001).

Lastly, radiosonde profiles can be used to retrieve PW values. The dataset of radiosonde and ozonsonde profiles collected at LAO between 1994 to 2023 has been used. This dataset is maintained by CETEMPS, and available upon request. First, temperature (T) and relative humidity (RH) profiles are filtered to remove erroneous measurements, and smoothed using a 60s moving median filter to remove high-frequency noise. The water vapor density ρ_w is computed as:

$$\rho_w = \frac{RH}{100} \cdot \frac{e_s(T)}{R_w \cdot T} \quad (4.1)$$

where $e_s(T)$ is the saturation vapor pressure over water, computed following Hyland and Wexler as reported in Sommer et al. (2023), R_w is the specific gas constant for water vapor, and T is air temperature in K. Then, PW is obtained by vertically integrating ρ_w profile from Equation (4.1), and dividing by the density of liquid water (assumed equal to 1000 kg m^{-3}). Unless otherwise stated, the PW from radiosonde is used as reference value.

4.2.2 Data processing and statistical analysis

For most analyses, PW observations are aggregated into daily means, requiring at least three measurements per day for sun-photometer and GNSS data. Monthly means are subsequently derived from them, for long-term variability and trend analyses.

Instrumental consistency is assessed using temporally co-located observations. Radiosonde launches were paired with the closest GNSS or sun-photometer measurement within a maximum time window of one hour. Statistical agreement was quantified using Pearson’s correlation coefficient (r), linear regression, mean bias error (MBE), and mean absolute error (MAE).

Long-term trends are evaluated on de-seasonalized monthly PW anomalies using the modified Mann-Kendall test accounting for autocorrelation (Hamed and Ramachandra Rao 1998). Trend magnitudes are estimated using the Theil-Sen slope estimator (Sen 1968; Theil 1992), as implemented in Hussain and Mahmud (2019).

The annual cycle is characterized using climatological monthly means. With high-frequency instruments, like sun-photometer and GNSS, the diurnal cycle is analysed using hourly PW anomalies, defined as deviations from the respective daily mean. Data are grouped by climatological season (DJF, MAM, JJA, SON) to highlight seasonal differences in intra-day variability.

4.3 Results

4.3.1 Intercomparison of PW measurements

The comparison of co-located observations shows a strong and statistically significant agreement among all instruments, with Pearson correlation coefficients ranging between $r = 0.94$ and $r = 0.99$ (Table 4.2). Scatter plots of paired measurements and corresponding linear regression fits are shown in Figure 4.1. The best agreement is obtained between GNSS and radiosonde, which also exhibits the lowest bias and absolute error.

Sun-photometer PW values systematically underestimate those from GNSS and radiosondes, with mean negative biases of approximately 10–15 %, slightly larger than typically reported in the literature (Fragkos et al. 2019; Halthore et al. 1997; Pérez-Ramírez et al. 2014). As shown in Figure 4.1, the magnitude of the sun-photometer bias increases with increasing PW, indicating a dependency on atmospheric moisture conditions. This behavior has been previously reported for inland and mountainous sites and is partly attributed to the lack of seasonal variability in temperature and humidity profiles assumed by the retrieval algorithm (Campanelli et al. 2018; Vaquero-Martínez et al. 2022).

4.3.2 Long-term trend and annual cycle

Daily PW statistics for the three instruments are summarized in Table 4.3. Mean PW values at the L’Aquila site range between approximately 1.3 cm to 1.6 cm, depending on the observing technique, confirming the semi-arid character of the study area. Monthly mean PW time series are shown in Figure 4.2.

After removing the mean seasonal cycle, the de-seasonalized monthly anomalies are analyzed for the presence of monotonic trends. All datasets exhibit weak positive tendencies, with Theil-Sen slopes of the order of 10^{-3} cm month $^{-1}$. These trends correspond to

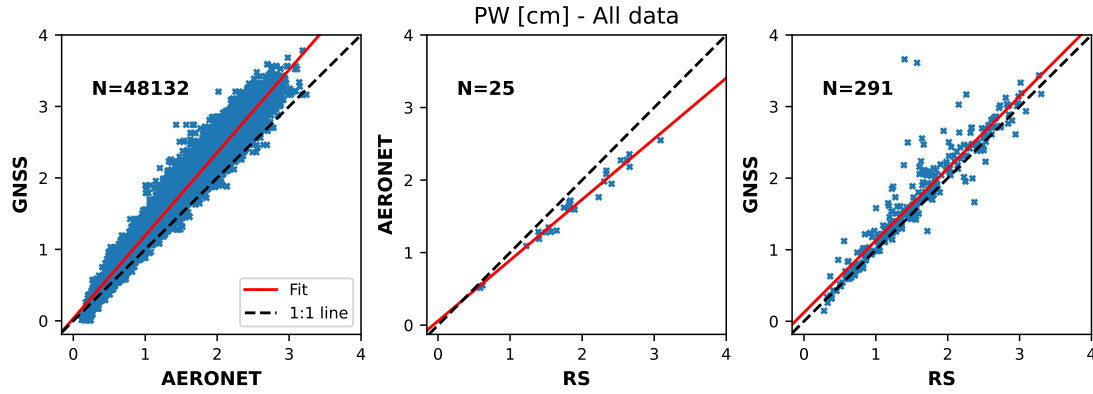


Figure 4.1: Comparison of PW between co-located AERONET, GNSS and radiosounding (RS) observations. The black dashed line represents the line of perfect agreement between observations, the red solid line represents the linear regression line. All values are in centimeters.

Table 4.2: Statistics of linear regression fit using co-located observations, grouped by instruments pair (ANET: AERONET, RS: radiosounding). Intercept, MBE and MAE in centimeters. r : Pearson's correlation coefficient, $Int.$: intercept, MBE : mean bias error, MAE : mean absolute error.

Pair (Y - X)	Count	r	Int. (Slope)	MBE		MAE	
				cm	%	cm	%
GNSS-ANET	48 132	0.987	0.03 (1.16)	0.262	18.59	0.264	19.09
ANET-RS	25	0.987	0.06 (0.84)	-0.252	-12.96	0.252	12.96
GNSS-RS	291	0.936	0.12 (1.01)	0.132	9.28	0.170	12.06

Table 4.3: PW statistics by instrument from monthly climatological observations. All values in centimeters except *count* value. SE : standard error of the mean, SD : standard deviation, $P.05$: 5th percentile, $P.95$: 95th percentile.

PW [cm]	Count	Mean $\pm$ SE	SD	Median	P.05	P.95	Min	Max
AERONET	1765	1.27 $\pm$ 0.15	0.53	1.18	0.69	2.03	0.68	2.04
GNSS	6138	1.61 $\pm$ 0.17	0.58	1.61	0.90	2.39	0.88	2.43
Radiosonde	501	1.50 $\pm$ 0.17	0.58	1.45	0.76	2.33	0.73	2.42

less than 0.02 cm yr^{-1} and are statistically significant only for the GNSS and radiosonde series. Despite their limited magnitude, the consistency between independent observing systems supports the robustness of the detected signal.

The climatological annual cycle of PW, derived from daily observations and shown

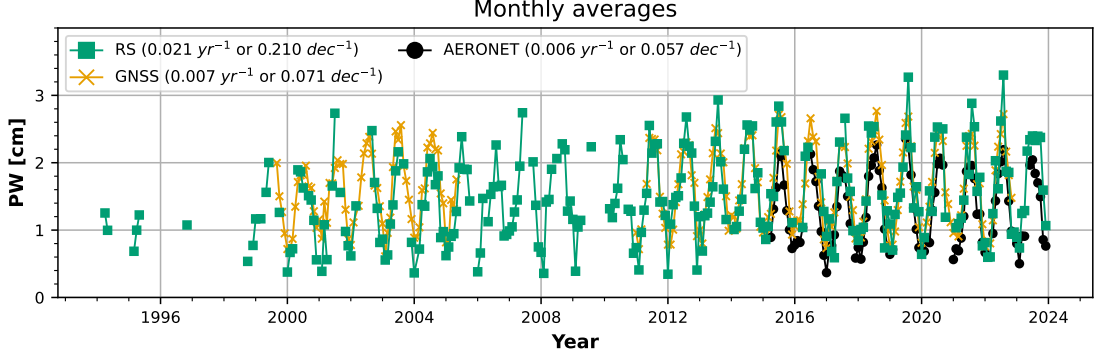


Figure 4.2: Time series of monthly PW for AERONET, GNSS and radiosounding (RS) observations. For a data gap larger than 60 days, the time series line is interrupted. In the legend is reported also the trend from Theil-Sen slope estimator both in units of centimeters per year, and centimeters per decade.

in Figure 4.3, exhibits a clear seasonal modulation. Minimum PW values occur during winter, with typical values between 0.7 cm to 0.9 cm, while maximum values are reached in summer, when PW can exceed 2 cm (Table 4.4). This pattern reflects enhanced evapotranspiration and convective activity during the warm season.

The intra-monthly variability, quantified by the standard deviation, is largest between late spring and early autumn and smallest during winter. This behavior is consistent with previous studies conducted over continental and mountainous regions in the Northern Hemisphere (Kalinnikov and Khutorova 2017). When cloud-screened GNSS and radiosonde data are considered, the bias relative to sun-photometer observations is reduced, particularly outside the summer months (Figure 4.3, right plot).

4.3.3 Diurnal cycle

The diurnal evolution of PW anomalies is illustrated in Figure 4.4, where hourly deviations from the daily mean are shown for each climatological season. GNSS-derived PW exhibits a systematic increase from morning to late afternoon, with the largest peak-to-peak amplitude during summer and the smallest during winter. This seasonal modulation closely follows the availability of solar radiation. In the same figure, a strong correspondence is observed between PW anomalies and near-surface temperature anomalies, suggesting that local radiative forcing and evapotranspiration processes dominate the diurnal PW cycle, rather than horizontal moisture advection. Similar patterns have been reported for other inland and orographically sheltered sites (Dai et al. 2002; Ortiz de Galisteo et al. 2011; Ziskin Ziv et al. 2021). The temperature anomaly is calculated in the same way as PW anomalies, so as hourly temperatures minus the relative daily mean.

An anomalous local maximum during early morning hours is observed in sun-photometer PW anomalies in Figure 4.4, a feature not reproduced by GNSS measurements.

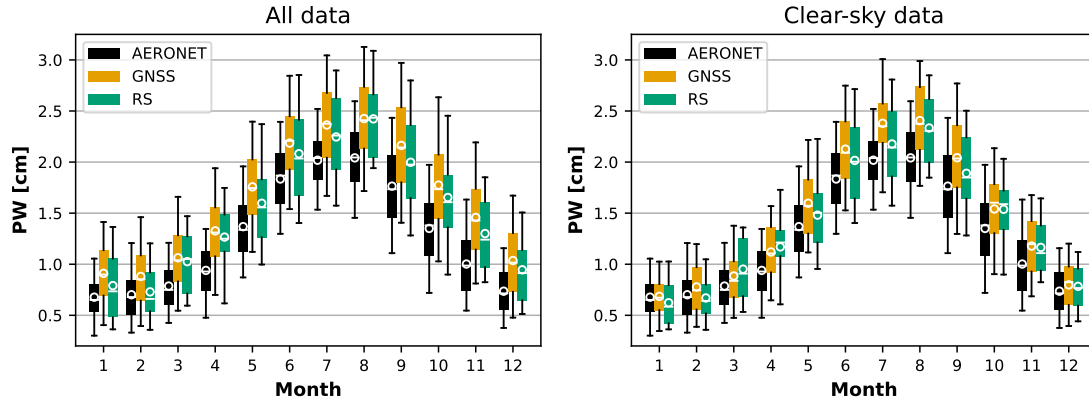


Figure 4.3: Monthly Boxplots of PW from AERONET, GNSS and radiosounding (RS) daily observations. The horizontal solid line inside the box is the median, the circle is the mean. The box extends from 25-th to 75-th percentile, and the whiskers extend to 1.5 times the interquartile range below/above these percentiles. The data falling outside the whiskers are not shown. All values are in centimeters.

Table 4.4: Monthly climatological PW average by instrument (RS: radiosounding) from daily observations. All values in centimeters except *Count* values. *SD*: standard deviation.

	AERONET			GNSS			RS		
	Count	PW	SD	Count	PW	SD	Count	PW	SD
Jan	103	0.68	0.24	550	0.91	0.31	40	0.79	0.35
Feb	97	0.70	0.25	459	0.88	0.32	37	0.73	0.29
Mar	96	0.79	0.25	487	1.06	0.34	43	1.03	0.34
Apr	141	0.94	0.28	473	1.33	0.37	50	1.27	0.35
May	135	1.37	0.35	535	1.76	0.39	51	1.60	0.46
Jun	160	1.83	0.35	503	2.19	0.40	52	2.08	0.47
Jul	197	2.02	0.32	522	2.36	0.44	31	2.25	0.46
Aug	244	2.04	0.35	521	2.43	0.43	28	2.42	0.44
Sep	195	1.76	0.42	505	2.16	0.50	41	2.00	0.49
Oct	171	1.35	0.37	540	1.77	0.48	50	1.65	0.48
Nov	92	1.00	0.33	529	1.46	0.43	36	1.30	0.37
Dec	134	0.74	0.24	514	1.04	0.38	42	0.95	0.35

Since a physical increase in column-integrated water vapor should be consistently detected by both instruments, this behavior is likely related to observational or geometric effects affecting sun-photometer retrievals at low solar elevation angles. This feature is not observed in other similar works (e.g., Ortiz de Galisteo et al. 2011; Torres et al. 2010). Further investigation is required to fully explain this behavior and assess its potential

impact on diurnal PW studies.

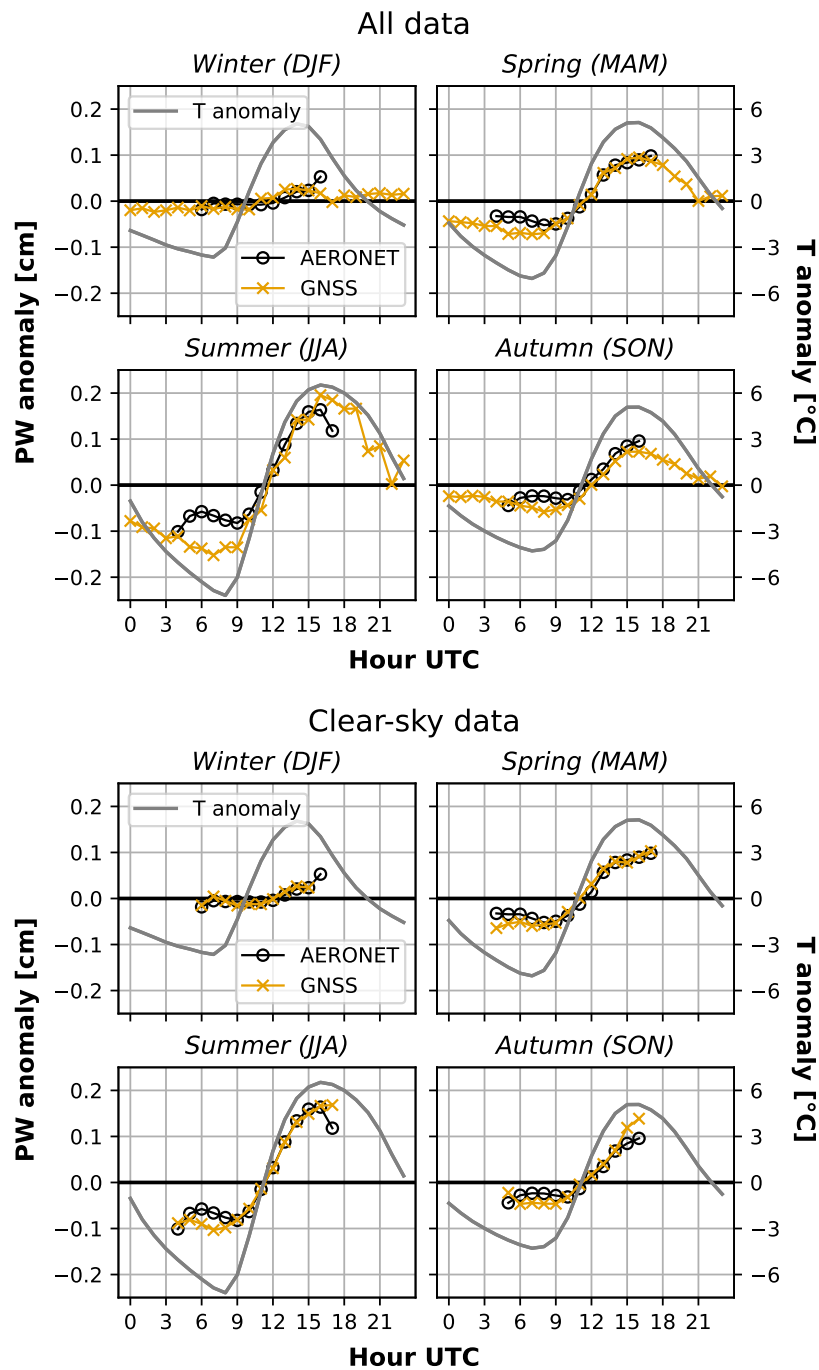


Figure 4.4: Hourly PW anomaly for AERONET and GNSS data, grouped by season. Upper plot with all the GNSS data available, lower plot with cloud-filtered GNSS data. The PW anomaly is in units of centimeters, the temperature anomaly is in units of degrees Celsius.

Chapter 5

Lunar photometry at LAO

5.1 Introduction

The photometry measurements from photometer like the Cimel CE318 described in Section 3.2.4, were restricted to unpolarized Sun’s radiation until the last two decades. Since then, various successful attempts to increase the information content retrieved from the ground-based observations, allowed to better characterize the aerosol properties and to increase the temporal continuity of measurements. In this chapter will be explored advancements in ground-based photometry related to the addition of polarized and nighttime observations. Despite the photometer installed at LAO in 2024 is capable of measuring polarized radiation, during the doctoral work has not been possible to integrate this kind of data in the workflow for lack of time. In Section 5.2 will be provided only a theoretical introduction about polarized photometry and some of the most relevant results from literature, with the analysis of polarized data at LAO to be referred to further studies. Concerning the nighttime photometry, in Section 5.3 is provided an overview of the theoretical background necessary to obtain AOD measurements from lunar photometry. The methodology shown is very similar to the one used by AERONET, then it is possible to compare the results. Indeed, in Section 5.3.1 it is shown the

5.2 Polarized observations

Polarimetric measurements have become a fundamental component in modern aerosol remote sensing using sun–sky radiometers. Unlike traditional photometers that record only the total radiance intensity, polarimetric instruments capture additional information on the state of polarization of scattered light, typically the Stokes parameters Q and U (see Section 1.1), the degree of linear polarization (DoLP), and the angle of polarization (AoP). This expanded set of measurements provides access to portions of the scattering matrix that are highly sensitive to aerosol optical and micro-physical properties, such particle size, refractive index, shape, and absorption, thus enhancing retrieval accuracy and reducing ambiguities inherent to intensity-only observations (Mishchenko and Travis

1997).

The interaction of sunlight with atmospheric aerosol modifies both the intensity and polarization state of the scattered radiation. Modern instruments such as polarized CIMEL sun/sky radiometers combine multi-spectral and multi-angular measurements with polarization channels, significantly enriching the information content available for inversion procedures (Z. Li et al. 2006; Z. Li et al. 2009). The CIMEL polarized sun-sky radiometers can provide the Stokes vector components: I, Q and U, typically ignoring V (L. Li et al. 2014).

Accurate polarimetric retrievals require rigorous calibration of the detectors and polarizing elements. Various studies have proposed methods to calibrate DoLP and Stokes parameters using solar radiation or dedicated laboratory systems. Solar-based calibration schemes ensure the correction of instrumental biases and long-term drifts (Z. Li et al. 2010), while more advanced setups such as the POLBOX system improve the precision of polarization measurements and enable operational field calibration of radiometers (Z. Li et al. 2018). Additional work has refined the computation of Stokes parameters and AoP from polarized CIMEL observations to ensure consistency and traceability of the measurements (L. Li et al. 2014).

The main advantage of using polarimetric measurements is observed in the inversion problem. The inversion seeks aerosol parameters that minimize the mismatch between measured and modeled radiances/polarization. The use of polarimetric measurements greatly reduce the degeneracy between aerosol size distribution, refractive index, and shape that often limits intensity-only retrievals, improving inversion stability and limiting the multiplicity of mathematically admissible solutions (dos Santos Oliveira et al. 2023; Z. Li et al. 2009). The degree of linear polarization provides strong constraints on the effective radius, fine-to-coarse mode partitioning, and the width of the particle size distribution, enabling more reliable microphysical solutions (Z. Li et al. 2006; Z. Li et al. 2009).

This is particularly important for complex aerosol mixtures or conditions where absorption and size parameters are strongly coupled. This additional information content has direct implication in climate change studies helping in evaluate the atmosphere heating by aerosol absorption (C. Chen et al. 2022). Studies using AERONET polarized observations confirm that polarization adds independent information to the inversion problem, allowing improved estimation of both the real and imaginary parts of the refractive index (Fedarenka et al. 2016). The sensitivity of polarization to particle asphericity is especially relevant for dust and other non-spherical aerosol types. Whereas radiance measurements alone often fail to discriminate spherical from irregular particles, polarimetric signals encode characteristic features of non-spherical scattering patterns, thus improving retrievals of shape-related properties (Fedarenka et al. 2016).

While the AOD is primarily derived from direct-sun extinction measurements and does not require polarization, polarimetric information contributes indirectly to its accuracy. Precise calibration of polarized channels reduces systematic biases in sky radiances, which in turn enhances the reliability of inversion-based estimates of single-scattering albedo, phase functions, and other optical properties linked to AOD partitioning across modes (Z. Li et al. 2010; Z. Li et al. 2018).

In summary, incorporating polarization into photometric measurements represents a significant leap forward in aerosol remote sensing capability. By providing additional observables that are highly sensitive to particle size, composition, and morphology, polarimetry enhances the precision and robustness of inversion algorithms, minimizes retrieval ambiguities, and enables a more complete exploitation of radiative transfer physics. Despite this, during the doctoral work has not been possible to analyze the information from polarimetric measurements, although the polarimetric data are measured since the inclusion of the new lunar photometer in AERONET, both for direct and sky observations, and available on the AERONET database.

5.3 Nighttime observations

While solar photometry is well-established for daytime aerosol characterization, nighttime retrievals remain challenging due to the absence of a strong and direct source of light, as the Sun in daytime. The last decade developments in lunar photometry (e.g., Á. Barreto et al. 2016), allow for aerosol measurements using the Moon as a source of light. Indeed, in contrast to the stellar photometry that requires very specialized instruments (e.g., Leiterer et al. 1995; Perez-Ramirez et al. 2008), the lunar photometry shares the same instrument design used in sun photometry, except for an increased sensitivity of detectors. This advantage is compensated by strong limitations due to the Moon itself. Indeed, the magnitude of the lunar irradiance depends strongly on the phase angle, then, also increasing the sensitivity of the instruments, the lunar photometry is a technique applicable only in the brighter phase of the lunar cycle (i.e., between first quarter and last quarter).

The Moon is not an emitter of radiation, as the Sun or stars, but only a reflector of Sun's radiation. The radiation coming from the Moon depends on many factors: Sun-Moon-Earth geometry, lunar libration, lunar phase angle and lunar surface's inhomogeneities. This produces a continuous change in the Moon's reflected radiance reaching the Earth, also along the same night. Despite that, Moon's properties as a reflector are stable in time to the order of 10^{-8} yr^{-1} (Kieffer 1997) so that the Moon results in an optimal calibration target.

As can be seen in the left plot on Figure 5.1, the lunar surface reflectance depends strongly on the phase angle, and the combined effect of the factors listed above produce also an asymmetry between positive and negative phases. Thus, a model of Moon's irradiance is required to know the quantity of reflected radiation reaching the top of the atmosphere. The U.S. Geological Survey (USGS) acquired an extensive dataset of ground-based lunar disk-integrated irradiance observations over 32 channels, corrected for atmospheric extinction and calibrated against stars, through the RObotic Lunar Observatory (ROLO) (Kieffer and Stone 2005). From these observations was possible to obtain an empirical model of the spectral lunar disk reflectance (A_λ) based on geometric

variable:

$$\begin{aligned} \ln A_\lambda = & \sum_{i=0}^3 a_{i,\lambda} g^i + \sum_{j=1}^3 b_{j,\lambda} \Phi^{2j-1} + c_1 \phi + c_2 \theta + c_3 \Phi \phi + c_4 \Phi \theta \\ & + d_{1,\lambda} \exp^{-g/p_1} + d_{2,\lambda} \exp^{-g/p_2} + d_{3,\lambda} \cos \left(\frac{g - p_3}{p_4} \right) \end{aligned} \quad (5.1)$$

where g is the Moon's phase angle, ϕ and θ are the selenographic sub-observer longitude and latitude respectively, and Φ is the sub-solar longitude. The coefficients a , b , c , d and p are obtained by means of fitting the model over the observational dataset, and are provided in Kieffer and Stone (2005).

This reflectance model requires an additional fitting over the true reflectance values from laboratory spectra of soil samples collected by Apollo-16 mission, to correct some numerical spikes due to the empirical nature of the model, as shown in the right plot on Figure 5.1. The lunar reflectance model residuals have a relative uncertainty of about 1 %, so that the model is able to track accurately the changes of irradiance (Stone et al. 2020). Combining all the error sources, the lunar irradiance from the ROLO model shows an absolute uncertainty of about 5 % to 8 %, mainly due to the uncertainty in the atmospheric extinction correction (Stone et al. 2020).

Because of the uncertainty in the ROLO model, many efforts are in place to minimize its absolute uncertainty to less than 1 %. Some of them using airborne or space-based observations like for the NASA projects Airborne Lunar Spectral Irradiance (air-LUSI) (Woodward et al. 2022) or ARCSTONE (Swanson et al. 2020). Other using empirical corrections like the ROLO Implementation for Moon photometry Observation (RIMO) (A. Barreto et al. 2019) with additional correction from the RIMO Correction Factor (Román et al. 2020), or using a completely new set of ground-based observations as for the Lunar Irradiance Model of ESA (European Space Agency) (LIME) (Toledano et al. 2024).

The exoatmospheric spectral irradiance of the Moon ($E_{0,\lambda}^M$) can be written as:

$$E_{0,\lambda}^M = \frac{\Omega_{Moon} \cdot A_\lambda \cdot E_{0,\lambda}^S}{\pi \cdot \left(\frac{d_{SM}}{1 \text{ au}} \right)^2 \cdot \left(\frac{d_{ME}}{384\,400 \text{ km}} \right)^2} \quad (5.2)$$

where $E_{0,\lambda}^S$ is the exoatmospheric spectral irradiance of the Sun, $\Omega_{Moon} = 6.417\,75 \times 10^{-5}$ sr is the solid angle of the Moon at the standard Moon-observer distance of 384 400 km, d_{SM} is the actual Sun-Moon distance in au, d_{ME} is the actual Moon-observer distance in km, and A_λ is the (adimensional) disk-integrated lunar reflectance provided by the ROLO model in Equation (5.1).

Recalling Equation (3.4) used in sun-photometry, also for lunar observations is required the knowledge of the calibration constant. In this case, different ways exist to solve the problem. The most naive way to obtain the lunar calibration constant is to perform a Langley plot. Similar to the classic daytime Langley plot, it requires clean, cloud-free and stable atmospheric conditions, in order to span air masses between 2 and 5 at moonrise

and moonset (A. Barreto et al. 2013). Because the irradiance of the Moon changes along the night, the $V_{0,\lambda}$ retrieved at different air mass values changes as well. Then, the V_0^N calibration constant, once corrected for the Moon's illumination changes along the night, can be used only for that specific night, simply because is not truly constant (A. Barreto et al. 2019). One solution is to normalize $V_{0,\lambda}^N$ in respect to the lunar irradiance, so that:

$$\frac{V_{0,\lambda}^N}{E_{0,\lambda}^M} = \kappa_\lambda^N \quad (5.3)$$

where $E_{0,\lambda}^M$ is the exoatmospheric spectral lunar irradiance from Equation (5.2), and κ_λ^N will be the true spectral calibration constant for the lunar photometer. Again, the κ_λ is obtained as the intercept of the fitting line of $\ln(V_{0,\lambda}^N/E_{0,\lambda}^M)$ at different air masses as in the usual Langley plot technique, and it strictly accounts for the instrument's photometric responsivity, but also for any residual systematic offset difference between modeled and actual exoatmospheric lunar irradiance (A. Barreto et al. 2013). This method is called Lunar-Langley plot, and can be applied to all the lunar photometers available, but is highly dependent on the uncertainty of the lunar irradiance model being used. As a matter of fact, some systematic bias observed in the ROLO model and its derivative, is the cause of a small variability up to 5% in the κ_λ^N .

Another way is to use the so-called Sun-Moon gain factor method (Á. Barreto et al. 2016). This technique can be applied only to instruments that share the same setup for daytime and nighttime observations, except for internal electronic gain of the detectors, as is the case of Cimel photometers. Thus, it is possible to retrieve the lunar calibration constant (κ_λ^N) from daytime calibration constant (κ_λ^D) as:

$$\kappa_\lambda^N = \kappa_\lambda^D \cdot G \quad (5.4)$$

where G is the gain used in lunar observation, corresponding to a nominal value of 4096. The value of κ_λ^D can be obtained during daytime Langley plot with Equation (5.3), using the solar irradiance value ($E_{0,\lambda}^S$) instead of the lunar irradiance. The Sun-Moon gain factor has the advantage to be independent by the lunar irradiance model.

The experimental determination of the gain G using a common target (e.g., a calibration integrating sphere), with both the Sun and Moon gains, is obtained as the ratio of the two observed signals, so that:

$$\frac{DC^N}{DC^D} = G \quad (5.5)$$

where DC is the observed signal, in terms of digital counts, using the relative gain (superscript N for Moon gain, D for Sun gain). The experimental values of G differ less than 0.3% from the nominal value (Román et al. 2020), and this uncertainty expected to introduce an absolute error about 0.011 in the AOD computation (Á. Barreto et al. 2016).

In literature, can be found many examples of an AOD dependency from the Moon's phase angle (e.g., Geogdzhayev and Marshak 2018; Lachérade et al. 2014; Román et al. 2020; Uchiyama et al. 2019). This effect can be related to a bias in the ROLO lunar model, and it affects also its derivative RIMO model. Whenever the formula to retrieve

the AOD involves the model reflectance or irradiance, it is necessary to correct for such bias to obtain the correct lunar optical depth.

In the AERONET network, where the ROLO model is used, this bias has a linear correction (AERONET 2019), so that the lunar calibration constant can be written as:

$$\kappa_{\lambda}^N = \kappa_{\lambda}^S \cdot G(1 + \delta) \quad (5.6)$$

with $\delta = c_1 + (c_2 \cdot g)$ where c_1, c_2 are coefficients depending on the photometer channel, and g is the Moon's phase angle. Equation (5.6) reflects the current correction applied in AERONET lunar processing (AERONET 2019, and Dr. África Barreto personal communication). It is important to highlight that the calibration coefficient C in AERONET (2019) is the inverse of the calibration constant κ used in this work.

Otherwise, when using the RIMO model and applying the RCF correction (Román et al. 2020), the bias has a quadratic correction respect the Moon's phase, so that the lunar calibration constant can be written as:

$$\kappa_{\lambda}^N = \kappa_{\lambda}^S \cdot G(c_1 + c_2 \cdot g + c_3 \cdot g^2) \quad (5.7)$$

where c_1, c_2, c_3 are coefficients depending on the photometer channel, and g is the Moon's phase angle. For Prede photometers the correction is still quadratic but without the linear term (i.e., $c_2 = 0$) (Uchiyama et al. 2019).

Finally, the formula to retrieve the lunar AOD follow from (3.6), using a suitable value of $V_{0,\lambda}$ obtained by means of one of the calibration methods shown above:

$$\tau_{a,\lambda} = \frac{1}{m_a} \cdot \left[\ln \left(\frac{V_{0,\lambda}^N}{V_{\lambda}} \right) - m_{atm} \cdot \tau_{atm} \right] \quad (5.8)$$

The combined uncertainty of τ_a depends on the method used to obtain the calibration coefficient $V_{0,\lambda}^N$, and it ranges from 0.011 to 0.019, with 0.001 changes due to lunar phase as well (Á. Barreto et al. 2016).

The geometrical parameters, as well as the Moon's zenith angle, used to compute the lunar reflectance, Equation (5.1), and irradiance, Equation (5.2), can be obtained from the Spacecraft, Planet, Instrument, C-matrix, Events (SPICE) observation geometry information system¹ (Acton 1996; Acton et al. 2018) developed by NASA's Navigation and Ancillary Information Facility (NAIF). Using a set of kernels providing the information about celestial bodies' ephemeris, bodies frame and reference system, and time parameters, SPICE is able to provide the required distances and angles at every instant.

5.3.1 Detection of nighttime aerosol at L'Aquila site

The sun-sky-lunar Cimel CE318-TP9 photometer was installed in April 2024, but became the official instrument into AERONET network only since May 2025. After the installation,

¹<https://naif.jpl.nasa.gov/naif/index.html>, last check on November 28, 2025

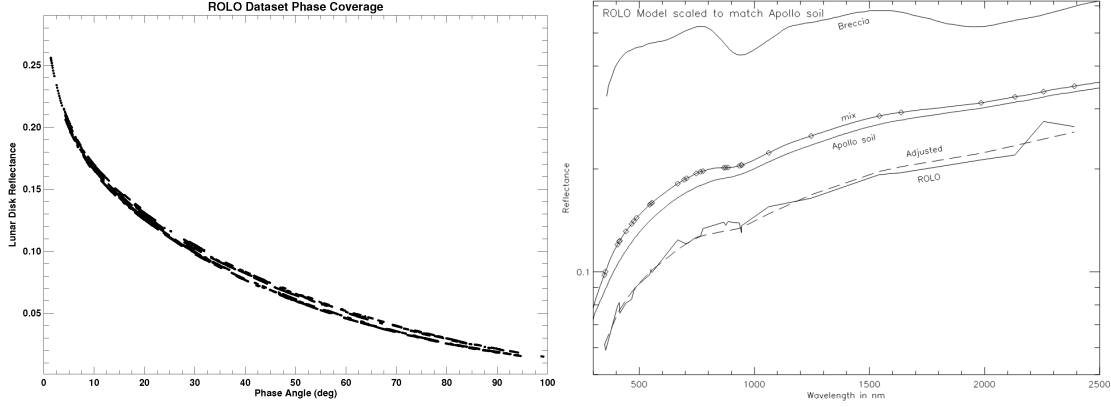


Figure 5.1: *Left:* dependency of the lunar reflectance from the phase angle and asymmetry between waxing and waning phases, as seen from the observations in the ROLO dataset; adapted from Stone et al. (2020). *Right:* scaling (dashed line) of the noisy ROLO reflectance with the smooth reflectance from lunar samples; adapted from Kieffer and Stone (2005).

the code for lunar AOD retrieval has been developed based on works from A. Barreto et al. (2019), Á. Barreto et al. (2016), and Román et al. (2020). The raw data are extracted through the instrument’s software as text ASCII file, one for each available scenario. Files with extension *.NSU* contain daytime AOD, and *.NLU* files contain nighttime AOD. Both scenarios collect triplets of direct irradiance observations. In addition, auxiliary information about pressure are necessary, which are provided in *.STA* files. In contrast to AERONET, the raw data processing does not include temperature correction because it was not possible to obtain the temperature response function from the calibration center.

5.3.1.1 Methodology

The formula to retrieve the lunar AOD follows from Equation (5.8) using the Sun-Moon gain factor (Á. Barreto et al. 2016) and RIMO correction factor (Román et al. 2020):

$$\tau_{a,\lambda} = \frac{1}{m_a} \cdot \left[\ln \left(\frac{E_{0,\lambda}^M \cdot \kappa_{\lambda}^N}{V_{\lambda}} \right) - m_{atm} \cdot \tau_{atm} \right] \quad (5.9)$$

where $E_{0,\lambda}^M$ is the one in Equation (5.2), and κ_{λ}^N is the one on Equation (5.7). Because the daytime calibration constant is nothing but $\kappa_{\lambda}^D = V_{0,\lambda}^D / E_{0,\lambda}^S$, the solar irradiance terms cancel out, and Equation (5.9) depends only on the knowledge of the daytime calibration constant $V_{0,\lambda}^D$ by means of one of methods explained in Section 3.2, and transmissivity of other gases contributing to extinction. The spectral $V_{0,\lambda}^D$ constants used for raw data processing are obtained from the last calibration of lunar photometer performed on February 2025 by PHOTONS/Laboratoire d’Optique Atmosphérique (LOA) at the University of Lille in France.

The air mass terms are computed from the Sun/Moon zenith angle (θ , in degrees) using the formula of Kasten and Young (1989) for molecular (i.e., Rayleigh), NO_2 and aerosol components (Giles et al. 2019):

$$m_a = \cos(\theta) + 0.50572 \cdot (96.07995 - \theta)^{-1.6364} \quad (5.10)$$

and using the formula of Komhyr et al. (1989) for gases component:

$$m_g = \sec(\theta) - 0.0018167 \cdot [\sec(\theta) - 1] - 0.0028750 \cdot [\sec(\theta) - 1]^2 - 0.0008083 \cdot [\sec(\theta) - 1]^3 \quad (5.11)$$

Both the formulae require the use of the apparent zenith angle (Kasten and Young 1989). If the air mass computation produces any negative or complex value, that is unphysical, the related observation is removed.

The spectral molecular optical depth ($\tau_{R,\lambda}$) is computed from Bodhaine et al. (1999) as:

$$\tau_{R,\lambda} = \frac{p}{p_0} \cdot 0.002152 \left(\frac{1.045996 - 341.29061 \cdot \lambda^{-2} - 0.90230850 \cdot \lambda^2}{1 + 0.0027059889 \cdot \lambda^{-2} - 85.968563 \cdot \lambda^2} \right)$$

where p is the site actual pressure in hPa, $p_0 = 1013.25$ hPa is the sea-level reference pressure, and λ is the wavelength in μm .

For CO_2 and CH_4 the optical depth is obtained just scaling the absorption constants to site's pressure (Giles et al. 2019), and it is constant along the year, so that:

$$\tau_{CO_2} = 0.0087 \frac{p}{p_0}$$

$$\tau_{CH_4} = 0.0047 \frac{p}{p_0}$$

For NO_2 and O_3 the optical depth is obtained multiplying the monthly columnar abundance of the species (ρ_i) by its spectral absorption cross-section ($\sigma_{i,\lambda}$) as:

$$\tau_{i,\lambda} = \rho_i \cdot \sigma_{i,\lambda}$$

In contrast to AERONET, where the monthly columnar abundance is retrieved from satellite observations, in this work it is obtained as the monthly climatological value from the Copernicus Atmosphere Monitoring Service (CAMS) reanalysis (Inness et al. 2019). In particular, the dataset *CAMS global reanalysis (EAC4) monthly averaged fields* (CAMS 2020), provides monthly columnar mass values in kg m^{-2} for O_3 and NO_2 with a $0.75^\circ \times 0.75^\circ$ grid resolution, since 2003. The use of reanalysis instead of satellite data, ensures a continuous dataset of gases columnar abundance that is constantly updated and homogenized along its whole time range. The monthly climatology has been created averaging, for each month, all data from 2003 to 2023, extracting the pixel corresponding to

the LAO site, and converting to columnar abundance in mol m^{-2} through the appropriate molar weight value. This methodology allows using the climatology for computing a near-real time AOD, leaving room to recompute a-posteriori the AOD with the effective monthly columnar abundance if available, that would reflect a more realistic value in respect to the climatology.

The spectral absorption areal cross-section in $\text{cm}^2 \text{mol}^{-1}$ is obtained from GOME FM data (Burrows et al. 1998; Burrows et al. 1999) for both species, assuming 273 K and vacuum wavelength condition. These cross-sections are available on the site of the Institute for Environmental Physics (IUP) of the University of Bremen ² (courtesy of Dr. Carlo Arosio, IUP).

The astronomical distances and the Moon’s zenith angle, are obtained using the *SPICE Toolkit* version N0067 ³ through its Python implementation in *Spiceypy* library (Annex et al. 2020). To run SPICE, some kernels are required providing information about celestial bodies position, orientation and reference system. In this work the DE440 version of the kernels has been used, consisting in the planetary and lunar ephemeris (`de440s.bsp`) (Park et al. 2021), the planetary orientation (`moon_pa_de440_200625.bpc` and `earth_1962_240827_2124_combined.bpc`) and frame (`moon_de440_220930.tf` and `earth_assoc_itrf93.tf`), for the Moon and the Earth, respectively (Park et al. 2021; Seidelmann et al. 2002). In addition, some general-purpose kernels provide other planetary (`pck00011.tpc`) and time parameters (`naif0012.tls`). Finally, the NAIF *Pinpoint* tool is used to generate custom kernels with the position of the LAO site in respect to the Earth reference system (`lao_station.bsp` and `lao_station.tf`).

5.3.1.2 Quality control

Raw DNs are quality controlled in a manner similar to AERONET, that consist in removing observed triplets if:

1. radiometric sensitivity of 870 and 1020 (silicon) channels is less than 100 DN.
2. abnormal temperature is detected, so that $T < -40^\circ\text{C}$ or $T > 100^\circ\text{C}$.
3. air mass more than 7 is found.
4. variance of DN triplet, in terms of root mean squared deviation divided by triplet’s mean, is more than 0.16.
5. signal-to-noise ratio is low, so that $DN < \frac{V_{0,\lambda}}{1500}$, meaning too much diffuse light in respect to the direct beam.

The first three will remove the whole triplet if conditions are met, instead the last two are wavelength-wise, so that only the wavelength that fulfill such conditions is removed. In

²<https://www.iup.uni-bremen.de/gruppen/molspec/databases/index.html>, last check on March 15, 2026

³<https://naif.jpl.nasa.gov/naif/toolkit.html>, last check on March 15, 2026

Giles et al. (2019) more criteria are additionally applied for a fine-grained pre-screening of data.

The above criteria are able to remove most observations affected by strong attenuation due to stratus or cumulus clouds, but a proper cloud-screening algorithm is still needed. In this work, the screening algorithm used by AERONET Version 3 that is based on AOD and AE values has been partially implemented. This algorithm is able to detect also the influence from thin cirrus cloud and is able to overcome some over-screening affecting the previous algorithm version shown in Smirnov et al. (2000). This algorithm is based on the fact that aerosol optical depth is expected to have a temporal variability much slower in respect to cloud optical depth (Giles et al. 2019; Smirnov et al. 2000). The observations are removed if:

1. value of AOD is less than -0.01 . This value account for uncertainties in calibration, temperature correction and gases absorption (Smirnov et al. 2000).
2. AOD triplet variability is higher than the maximum between 0.01 or $0.015 \cdot \tau_\lambda$, for 675 nm, 870 nm and 1020 nm wavelengths simultaneously.
3. $AE_{440-870}$ is less than -1 or greater than 3 .

Again, the first two are wavelength-wise. In respect to the AERONET algorithm, the time series smoothing check and cirrus detection has not been implemented because of lack of time.

Because of lack of knowledge about temperature dependency coefficients, and also of coefficients for water vapor optical depth computation, the longer wavelengths (i.e., 1020 nm and 1640 nm) as well as the 937 nm channel have been discarded. Also, the UV channels (i.e., 340 nm and 380 nm) are not accounted because they are not measured at nighttime. Then the following results will be restricted to 440 nm, 500 nm, 675 nm and 870 nm channels.

5.3.1.3 Results

The previous procedure was developed after the installation of the lunar photometer in April 2024, but it was possible to compare it against AERONET lunar products only since May 2025, when the lunar photometer become the official instrument for LAO site in the AERONET network. First, a qualitative visual analysis of the results for some selected cases is done. In particular, the selection focused on those days when observations from late afternoon until next morning were available, in order to compare the lunar AOD against the previous and next daytime AODs.

An example of such visual continuity check can be seen in Figure 5.2. The days between December 15, 2024 to December 17, 2024 were characterized by a very low aerosol content, with $AOD_{500} < 0.1$. The upper panel shows a very well reasonable continuity between daytime and nighttime, both for AOD and AE. The lower panel shows a clear-sky day, as recorder by the ceilometer, except for some clouds on afternoon of December 16, where AOD jumps outside of plot limits and AE drop toward 0.

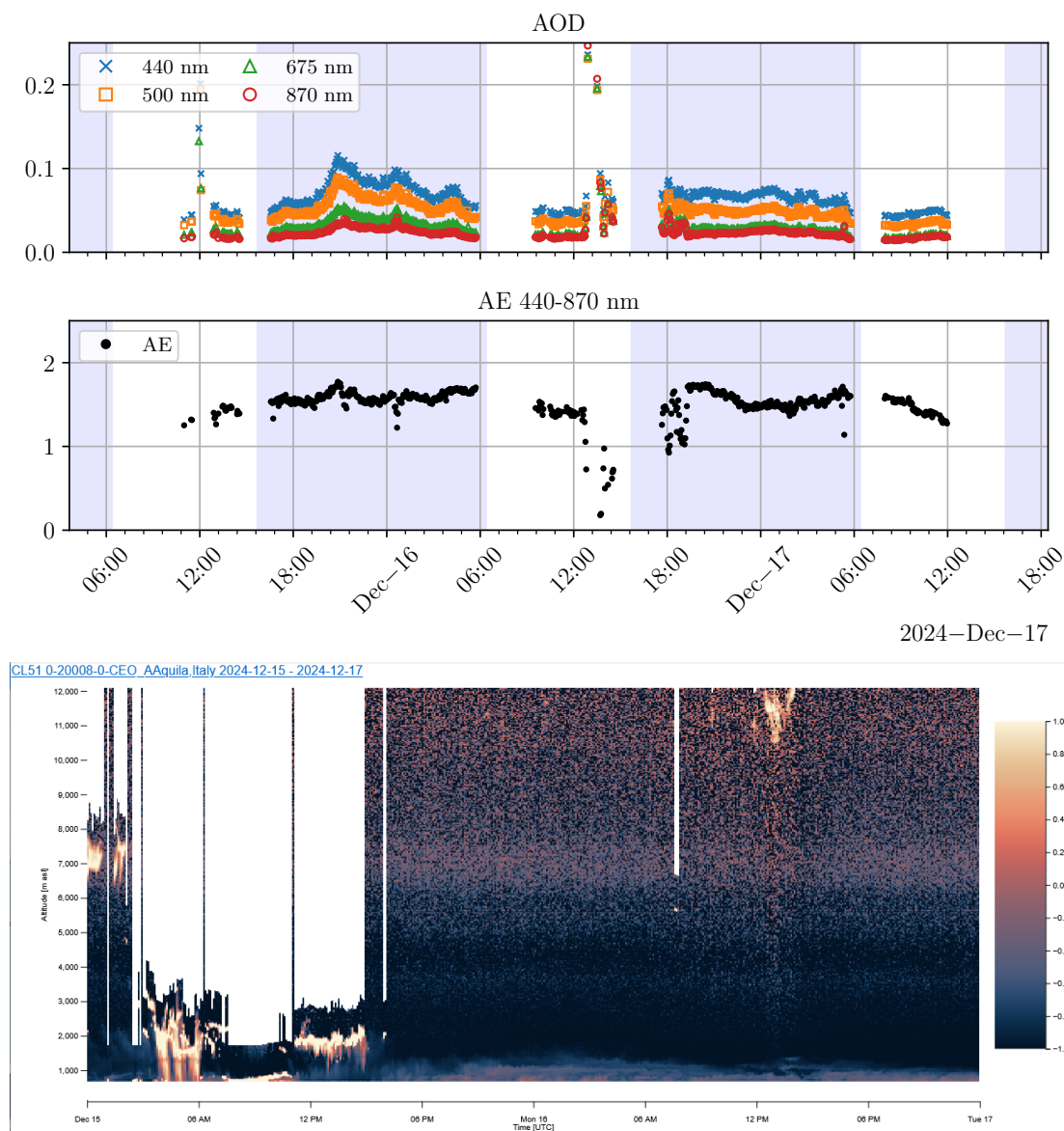


Figure 5.2: Spectral AOD (upper panel), AE (middle panel), and ceilometer RCS profile (lower panel) between December 15, 2024 and December 17, 2024. The RCS ranges up to December 17, 2024 midnight only (adapted from E-PROFILE website).

Table 5.1: Comparison of RIMO-like against AERONET over coincident daytime and nighttime observations in the period between June 1, 2025 to October 31, 2025. For each different spectral aerosol optical depth (AOD) and Angstrom exponent (AE) are reported the number of coincident observations, the Pearson correlation coefficient (r), the absolute and relative mean bias of residuals (MBE and % MBE), the standard deviation of the residuals (SD), and the slope and intercept for the linear regression of RIMO-like data.

	Case	Number	r	MBE (% MBE)	SD	slope (intercept)
Day	AOD 440	9286	0.999	-0.005 (-0.042)	0.004	0.994 (-0.004)
	AOD 500	9284	0.999	-0.003 (-0.034)	0.003	0.996 (-0.003)
	AOD 675	9288	0.999	0.000 (-0.002)	0.002	0.998 (0.000)
	AOD 870	9289	0.999	-0.001 (-0.027)	0.001	0.999 (-0.001)
	AE ₄₄₀₋₈₇₀	9222	0.997	-0.011 (-0.014)	0.026	1.026 (-0.046)
Night	AOD 440	1578	0.999	-0.003 (-0.024)	0.002	0.994 (-0.002)
	AOD 500	1914	0.999	-0.002 (-0.023)	0.002	0.997 (-0.002)
	AOD 675	2190	0.999	-0.001 (-0.011)	0.001	0.999 (-0.001)
	AOD 870	2230	0.999	-0.001 (-0.013)	0.001	0.999 (-0.001)
	AE ₄₄₀₋₈₇₀	1575	0.978	-0.007 (-0.003)	0.068	0.965 (0.042)

In contrast to very clean atmospheric conditions, the LAO site is located at a latitude where episodes of Sahara dust intrusion can happen. On June 18, 2024 there was one severe episode, with dust aerosol lasting for several days and reducing the atmospheric visibility. At that time the lunar photometer was already active, and it was possible to observe the development of dust layer both at day and night. In Figure 5.3 is possible to see the growth of dust layer caught by the ceilometer (lower left panel) and photometer (upper left panel). The AOD at 500 nm increases from values below 0.1 up to values above 0.6 once the dust layer is well-developed. The presence of large particles (i.e., like mineral dust) is confirmed by the Angstrom exponent values near 0. To assess the origin of such large particles is possible to use the NOAA HYSPLIT (Rolph et al. 2017; Stein et al. 2015) back-trajectories simulations (right panel), which confirm the elevated air mass containing those large aerosols comes from the North Africa desert region. Again, the visual inspection confirms a plausible behavior in the daytime-nighttime continuity.

Because the AOD retrievals use the RIMO approach coupled with simplified quality-control and cloud-screening procedures, a quality evaluation of the results become necessary. Because of its long history and huge amount of studies, AERONET results are assumed as ground truth data. For a long-term comparison all the dataset available both from AERONET and raw local data, from June 1, 2025 to October 31, 2025, is collected. Figure 5.4 shows the full dataset of locally collected data at LAO with the lunar photometer. Before the lunar photometer become the official instrument in AERONET (i.e, dashed line), there are gaps related to the calibration (February 2025), and to the fact that data download was done manually at that time, so some data went lost. The figure

shows spectral AOD (upper panel) and AE (lower panel) computed with the simplified processing described previously.

The result of RIMO against AERONET comparison is reported in Table 5.1, with statistics about the linear correlation for spectral AOD at the four wavelengths analyzed, and AE values. The correlation is very good and statistically significant for all cases. The Angstrom exponent shows the larger bias and standard deviation of residuals. The AOD bias is well below the instrumental uncertainty of 0.01 for daytime. The linear regression shows that RIMO algorithm tends to slightly underestimate in respect to AERONET. In Figure 5.5 an example of coincident observations from RIMO and AERONET on the time range between October 3, 2025 and October 4, 2025. The AOD at 500nm and AE agree very well both day and night as reported by Table 5.1 statistics.

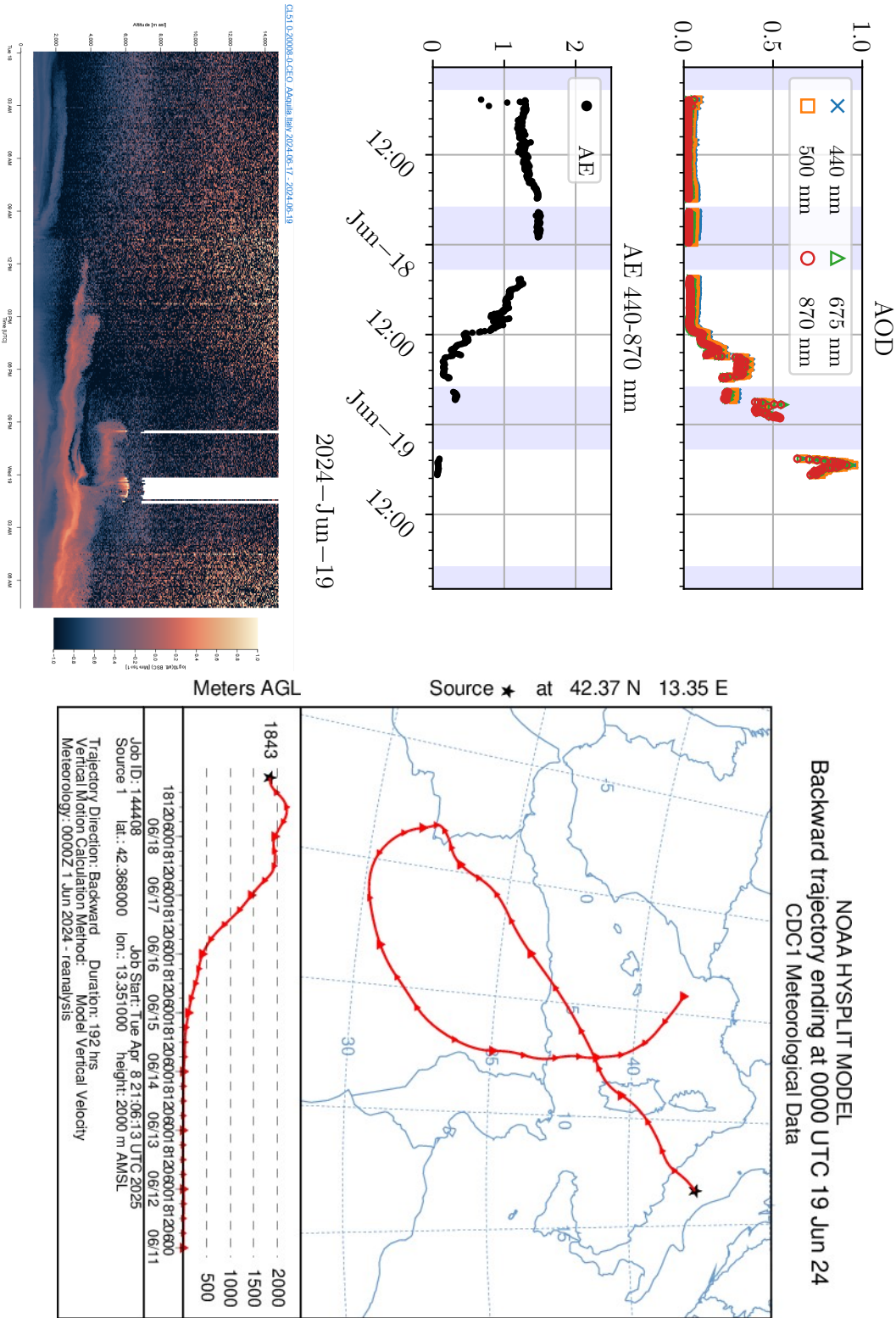


Figure 5.3: *Upper left:* photometer AOD and AE for June 17-19, 2024. *Lower left:* ceilometer backscattering profile during same period (adapted from E-PROFILE website). *Right:* HYSPLIT back-trajectory of the elevated dust air mass over LOA on morning of June 19, 2024.

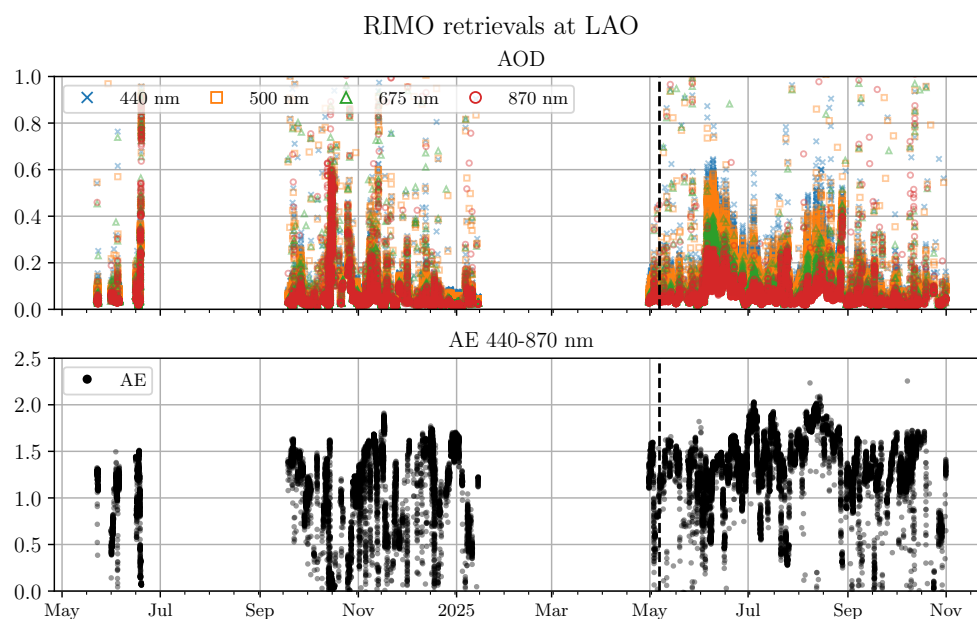


Figure 5.4: Time series of photometric observations of spectral AOD (upper panel) and AE (lower panel) collected with the lunar CE318-TP9 photometer at LAO. Nighttime observations are processed using RIMO algorithm. The black dashed line indicates the date since AERONET lunar products become available.

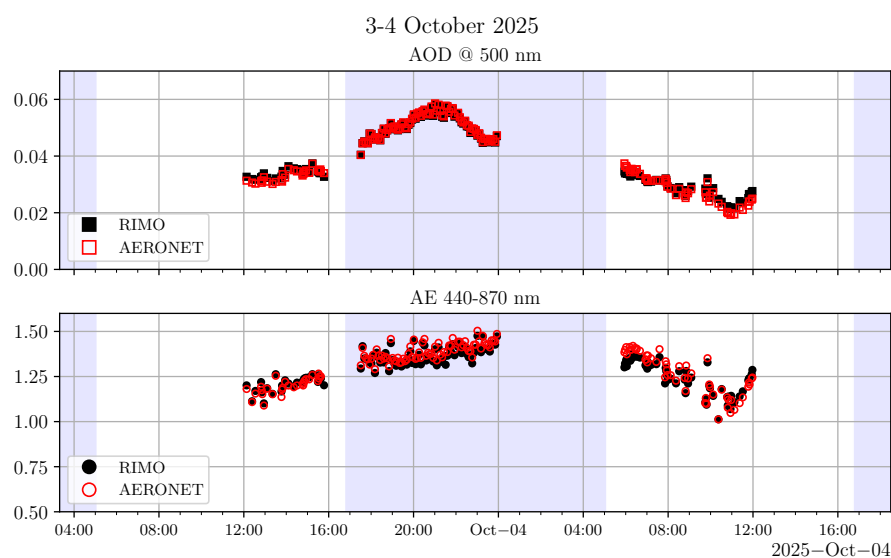


Figure 5.5: Comparison of RIMO and AERONET observations in daytime (white background) and nighttime (light blue background) for a night in October 2025. AOD at 500 nm (upper panel) and AE (lower panel) are shown. Good continuity between day and night behavior is observed.

Chapter 6

Aerosol retrieval with GRASP

6.1 Introduction

The *Generalized Retrieval of Atmosphere and Surface Properties* (GRASP) algorithm is an advanced inversion framework developed to retrieve atmospheric and surface properties from remote sensing observations in a statistically consistent and physically constrained manner. GRASP was originally conceived to generalize statistically optimized inversion techniques for aerosol retrieval from multi-angular, multi-spectral, and polarimetric measurements, and has subsequently evolved into a versatile platform capable of integrating heterogeneous observations from ground-based, airborne, and satellite sensors (Dubovik et al. 2011; Dubovik et al. 2014; Dubovik et al. 2021).

A defining feature of GRASP is its flexibility: the algorithm is designed to assimilate diverse measurement types while explicitly accounting for their uncertainties and for multiple sources of a priori information. This makes GRASP particularly suitable for complex and ill-posed inverse problems commonly encountered in atmospheric remote sensing.

GRASP algorithm is available both in open and proprietary versions, the latter developed for specific research and industrial projects. In this doctoral work, the free version of GRASP, also called *GRASP Open*, has been used, which is freely available online upon free registration.

6.2 GRASP forward and inverse modelling

The forward model in GRASP starts from a state vector $\mathbf{x}$, which contains atmospheric and surface parameters representing the optical, micro-physical and thermodynamic state of the system, to compute the observations through a model $\mathbf{F}(\mathbf{x})$, which accounts for radiative transfer processes including multiple scattering and polarization effects (Dubovik et al. 2011). Such forward model represents a well-posed problem because, once the physical processes regulating the radiative transfer are known, is it possible to reproduce observations related to the state of system.

On the contrary, the inverse problem consists in estimating a state vector $\mathbf{x}$ from a set of observations $\mathbf{y}$, so that:

$$\mathbf{y} = \mathbf{F}(\mathbf{x}) + \epsilon$$

where ϵ represent the noise associated to the observations. Less trivially, ϵ is the mismatch between the theoretical model and the true measurement.

Such inverse problems are typically ill-posed, as the number of unknowns may exceed the information content of the measurements, and the observations are affected by instrumental noise and modeling uncertainties. GRASP addresses these issues through a statistically optimized inversion approach based on generalized least squares with multiple a priori constraints.

6.2.1 Multi-Term Least Squares Method

The theoretical foundation of GRASP inversion is the *Multi-Term Least Squares Method* (MT-LSM), comprehensively described by Dubovik et al. (2021). The inversion is formulated as the minimization of a global cost function that combines observational constraints and multiple a priori terms. The cost function $\Phi(\mathbf{x})$ can be written as:

$$\begin{aligned} \Phi(\mathbf{x}) = & \sum_{i=1}^{N_d} [\mathbf{y}_i - \mathbf{F}_i(\mathbf{x})]^T \mathbf{S}_{\epsilon,i}^{-1} [\mathbf{y}_i - \mathbf{F}_i(\mathbf{x})] \\ & + \sum_{j=1}^{N_a} (\mathbf{x} - \mathbf{x}_{a,j})^T \mathbf{S}_{a,j}^{-1} (\mathbf{x} - \mathbf{x}_{a,j}) \end{aligned} \quad (6.1)$$

The first term on the right represents the noise of the observations weighted by the noise covariance matrix of observations ($\mathbf{S}_{\epsilon}$), which represents the expected observations noise. The second term represents the noise of the state vector in terms of residuals respect the most likely state defined by the a priori. In this case the covariance matrix of the a priori ($\mathbf{S}_a$) contains all physical limits and previous knowledge about parameters, both retrieved and not retrieved, in the state vector that helps in limiting the information provided by observations when they have not enough information content.

The summation in these two terms, comes from the fact that each observation is statistically independent. Indeed, GRASP assumes that each instrument, or better each kind of observations, is independent to the others. Then, AOD and radiance, despite being both measured by a photometer, are independent, and GRASP accounts the internal noise correlation at different spectral channels (and angles in case of radiance) for each of them.

Additional regularization terms can be introduced to enforce smoothness and reduce numerical oscillations, to ensure physical consistency and stabilize the solution:

$$\Phi_{\text{reg}}(\mathbf{x}) = \sum_k (\mathbf{H}_k \mathbf{x})^T \mathbf{S}_k^{-1} (\mathbf{H}_k \mathbf{x}) \quad (6.2)$$

where the operator $\mathbf{H}$ is a derivative to be applied on the state vector. The strength of the regularization term is set by (1) the order m of derivatives ($m = 0$, constant;

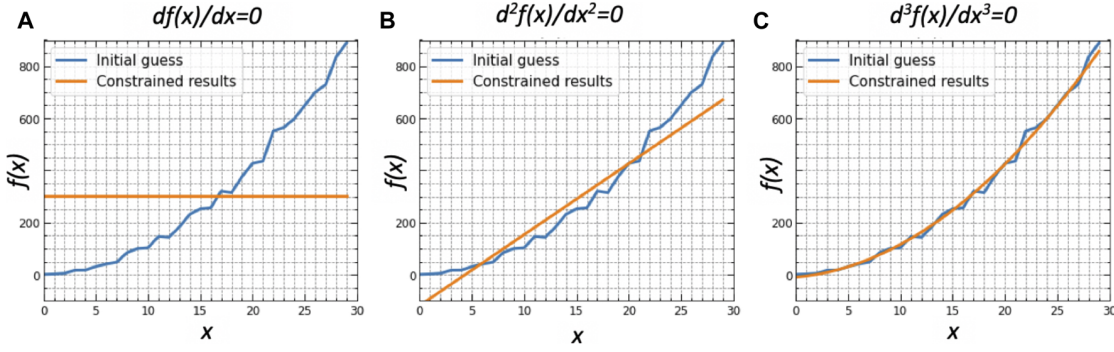


Figure 6.1: Effect of regularization constraints with derivatives of different order. The iterative retrieval, using the same initial guess, is limited by first (A), second (B) and third (C) derivatives. From Figure 2 in Dubovik et al. (2021).

$m = 1$, straight line; $m = 2$, parabola, etc.), and (2) the weight of regularization, called *Lagrange parameter* (Λ) (Dubovik et al. 2021). An example of different order regularization derivatives is shown in Figure 6.1. The regularization term is often used for retrieval of continuous functions, such as size distribution or spectrally-dependent parameters (e.g., complex refractive index), with stronger weights on parameters as complex refractive index and size distribution, or for parameters that may show spatial autocorrelation, as is the case of lidar measurements (Dubovik et al. 2021; Lopatin et al. 2021).

The Λ parameter exists both for a priori (Λ_{ap}) and regularization (Λ_{reg}) terms, and is a critical point to obtain meaningful inversion. The Λ_{ap} means how much the a priori is reliable in respect to the solution, with value of $\Lambda_{ap} \approx 1$ telling the algorithm to rely on the a priori only. The Λ_{reg} means how much will be the strength of smoothing applied to the solution. Values of $\Lambda_{reg} \gtrsim 1$ tend to produce very smooth solution, but it may mask high frequency signals. In general, no universal values of these Lagrange parameters exists, and should be tuned depending on the reliability of observations and a priori, and the autocorrelation of the parameter (e.g., spectral or vertical range).

Finally, the total cost function is obtained by summing all contributions from $\Phi(\mathbf{x})$ and $\Phi_{reg}(\mathbf{x})$, allowing for a statistically consistent balance between measurements and prior knowledge. From a practical point of view, GRASP performs computation in logarithmic space which allows defining a log-normal probability distribution function for positively-defined variables without the risk of generating negative values in the distribution (Dubovik et al. 2021).

6.2.2 Error estimation

An important advantage of the GRASP framework is the explicit characterization of retrieval uncertainties through the propagation of measurement and a priori errors into the retrieved state vector. Under the assumption of local linearity of the forward model around the solution $\hat{\mathbf{x}}$, the posterior covariance matrix of the retrieved parameters can be

approximated as (Dubovik et al. 2011; Dubovik et al. 2021):

$$\mathbf{S}_{\hat{x}} = \left(\sum_i \mathbf{K}_i^T \mathbf{S}_{\epsilon,i}^{-1} \mathbf{K}_i + \sum_j \mathbf{S}_{a,j}^{-1} + \sum_k \mathbf{H}_k^T \mathbf{S}_k^{-1} \mathbf{H}_k \right)^{-1} \quad (6.3)$$

where $\mathbf{K}_i = \partial \mathbf{F}_i / \partial \mathbf{x}$ denotes the Jacobian matrix of the forward model with respect to the state vector, and then the sensitivity of observations to the parameters in the state vector. Because of the covariance matrix of observations, the first term accounts also for internal correlations in each kind of independent observable. Finally, the Equation (6.3) highlights how observational information, a priori constraints, and regularization terms jointly determine the uncertainty of the retrieved parameters.

The posterior covariance matrix provides not only variances of individual parameters but also their cross-correlations, offering insight into parameter dependencies and information content. In GRASP applications, this information is essential for assessing retrieval reliability, performing sensitivity analyses, and ensuring a rigorous interpretation of the results, particularly in multi-sensor and synergistic retrieval configurations.

Because GRASP works in logarithmic space, also the calculated uncertainty is obtained in that space. Then the standard deviation (σ) from GRASP can be returned into the parameter space as (Dubovik et al. 2024):

$$\begin{aligned} x_{max} &= x \cdot \exp(\sigma_x) \\ x_{min} &= x \cdot \exp(-\sigma_x) \end{aligned} \quad (6.4)$$

where x is a parameter in the state vector ($\mathbf{x}$) retrieved by GRASP. For a more detailed assessment of the uncertainty, the algorithm is able to provide both the random ($\hat{\sigma}$) and systematic (σ_{bias}) components of the error so that the total standard deviation in GRASP space is (Dubovik et al. 2024):

$$\sigma = \sqrt{\hat{\sigma}^2 + \sigma_{bias}^2}$$

The effect of these uncertainty components on the retrieved parameters is well assessed in Herrera et al. (2022).

6.2.3 Aerosol retrieval

Within GRASP, the retrieved quantities are components of a multidimensional state vector whose structure is defined according to the physical model and the available observations (Dubovik et al. 2011). Concerning aerosols, the typical retrieved parameters include aerosol size distributions, complex refractive indices, particle sphericity, aerosol layer height and thickness, and optical properties such as aerosol optical depth, single-scattering albedo, and phase function. Surface reflectance properties can also be retrieved simultaneously, ensuring consistency between atmospheric and surface contributions.

Trivially, the number of retrieved parameters should be balanced by enough observations and a priori information in order to allow the MT-LSM to converge to a meaningful solution.

6.2.3.1 Inversion using photometer observations

One of the most established applications of GRASP is the inversion of observations from ground-based photometers. In this context, GRASP generalizes earlier inversion schemes by enabling a flexible definition of the state vector and by incorporating multiple a priori constraints in a rigorous statistical framework (Dubovik et al. 2011). In this case spectral AOD and angular radiance information are used. The inclusion of polarimetric observations improves the sensitivity of retrieved size distribution and real refractive index for fine mode aerosols, as well as of the fraction of spherical particles (Fedarenka et al. 2016; Z. Li et al. 2009). On the other hand, is it possible to perform GRASP inversion using spectral AOD only, but it is necessary to constrain the aerosol size distribution to a log-normal distribution to allow the inversion to converge (Torres et al. 2017). This latter method provides aerosol inversions also at nighttime, when sky radiance observations are not performed.

6.2.3.2 Inversion using photometer and lidar observations

A major extension of GRASP is its capability to exploit the synergy between column-integrated photometric observations and vertically resolved lidar measurements. The Generalized Aerosol Retrieval from Radiometer and Lidar Combined data *GARRLiC* algorithm represents a specific configuration of GRASP designed for this purpose (Lopatin et al. 2013). Lidar observations are introduced as additional terms in the cost function using dedicated forward models, providing strong constraints on aerosol vertical distribution. Practically, *GARRLiC* algorithm allows transferring micro-physical information to the vertical, and vertical information into the micro-physics. In GRASP, photometry-only retrievals rely more heavily on a priori information to constrain aerosol micro-physical properties, while lidar-only retrievals are primarily stabilized by smoothing constrain acting on the vertical structure. In synergistic photometer–lidar inversions, the need for informative a priori is reduced, whereas regularization remains essential, allowing the retrieval to be more data-driven while preserving physical consistency.

The GRASP framework has been further extended to integrate a wide range of ground-based and in situ observations, including multi-wavelength lidar, radiometers, radiosondes, and in situ aerosol measurements (Lopatin et al. 2021). Recent advancements combine polarimetric photometry with other active or passive sensors to further refine aerosol characterization. For instance, the joint use of lidar backscatter profiles and polarimetric radiances improves the retrieval of vertical distributions and micro-physical parameters, demonstrating higher accuracy compared to using either technique alone (dos Santos Oliveira et al. 2023). This integration highlights the versatility and robustness of polarization-enhanced datasets for multi-instrument aerosol remote sensing.

6.3 GRASP aerosol retrieval at LAO

The synergistic use of different instruments within GRASP is an approach that has become increasingly used in scientific research over the past few decades (e.g., Benavent-Oltra et al. 2021; Lopatin et al. 2021) and is being actively developed at networks level in ACTRIS (Miladi et al. 2019; Miladi et al. 2025). A number of studies exists on the combination of sun-photometer and high-power multi-wavelength lidar because of big vertical information content that is possible to extract from this kind of lidar systems (e.g., Benavent-Oltra et al. 2021; dos Santos Oliveira et al. 2023; Lopatin et al. 2013). The importance of accounting the vertical information is especially relevant at high aerosol loading, where ignoring vertical layering for smoke and dust aerosol types, can introduce biases in aerosol property retrievals when AOD above 0.3 and 0.5 for the polarimetric and radiometric observations respectively (Kalashnikova et al. 2011). On the other hand, multi-wavelength lidar systems are generally expensive and can not operate continuously 24/7, often requiring operators to start and to monitor the measurement (Wiegner et al. 2014). Despite the effort to coordinates the operation of such systems as for the European Aerosol Research Lidar Network (EARLINET) (Pappalardo et al. 2014), the uneven and limited spread of such systems around the globe is still an issue (Wiegner et al. 2014).

As already explained in Section 3.3, the ALCs are single-wavelength low-power lidars, originally designed to provide the cloud base altitude. Despite their lower SNR respect high-power lidars, they are sensitive enough to detect aerosol layers in the lower troposphere. Moreover, they are cheaper in respect to multi-wavelength lidars, and can operate continuously in an automatic way, resulting in a more widespread around the globe (e.g., Román et al. 2018; Wiegner et al. 2014).

Given such considerations, it is worthwhile to evaluate the synergy of photometers and ceilometers as a convenient approach for retrieving aerosol properties at a higher spatial density than that achievable with the combination of photometers and research-grade lidars. This synergy has been previously assessed in the work of Román et al. (2018). Starting from it, the GRASP inversion at L'Aquila site has been implemented using the combination of Cimel CE318-TP9 sun-sky-lunar photometer and Vaisala CL51 ceilometer, also including lunar photometric measurements as a way to extend further the methodology to nighttime observations.

6.3.1 Case study

During doctoral work, several specific episodes, ranging from desert dust intrusions to aerosol-free condition, have been analyzed to set up the GRASP algorithm. In this chapter, a single example case is analyzed, that encompasses most of the aspects and limitations related to GRASP retrievals using the synergy of photometer and ceilometer.

The selected period ranges from June 5, 2025 to June 6, 2025, including the night between these dates. In Figure 6.2 is reported the AOD and the AE from the pre-processed AERONET data. The aerosol column shows a build-up of aerosol optical depth in the morning of June 5, 2025 following the arrival of small aerosols which increase the AE

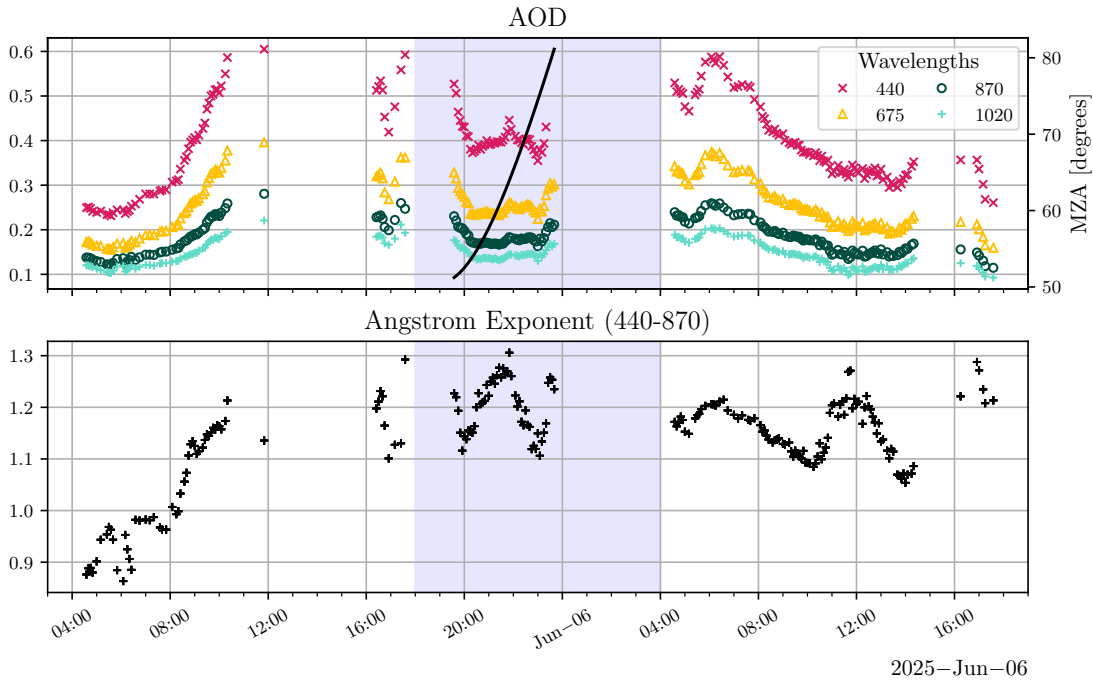


Figure 6.2: Spectral AOD (upper panel) and AE (lower panel) time series for the selected case study. The shaded area represents nighttime period. The black solid line in upper panel is the Moon's zenith angle (MZA) in degrees. All the wavelengths are reported in units of nanometers. Time is in UTC.

values from below 1 up to 1.2. The afternoon is characterized by cloud coverage that prevents observations with photometer as can be seen in left plot of Figure 6.5. During nighttime the AOD decreases again but AE showing a predominance of small aerosols reaching up to 1.3. The nighttime observations stop after midnight because the Moon goes below the horizon, as suggested by the Moon zenith angles growing up to 80° , thus limiting the availability of lunar observations also when the Moon's illumination is favorable. On June 6, 2025, the AOD increases a bit in early morning, then decreases along the day, while the AE still oscillates between 1.1 to 1.2.

The case described above is likely to be linked to air masses from North Africa, bringing a little of dust until the morning of June 5, 2025, followed by smoke from wildfires. In Figure 6.3 the back-trajectories at various heights for the June 6, 2025 morning. On June 3, 2025, the lowest trajectory collected smoke from wildfires lying in bottom atmospheric layers over the North Africa. Such air mass arrived over L'Aquila site the June 5, 2025 night (trajectory not shown) and continued to advect smoke until the next day.

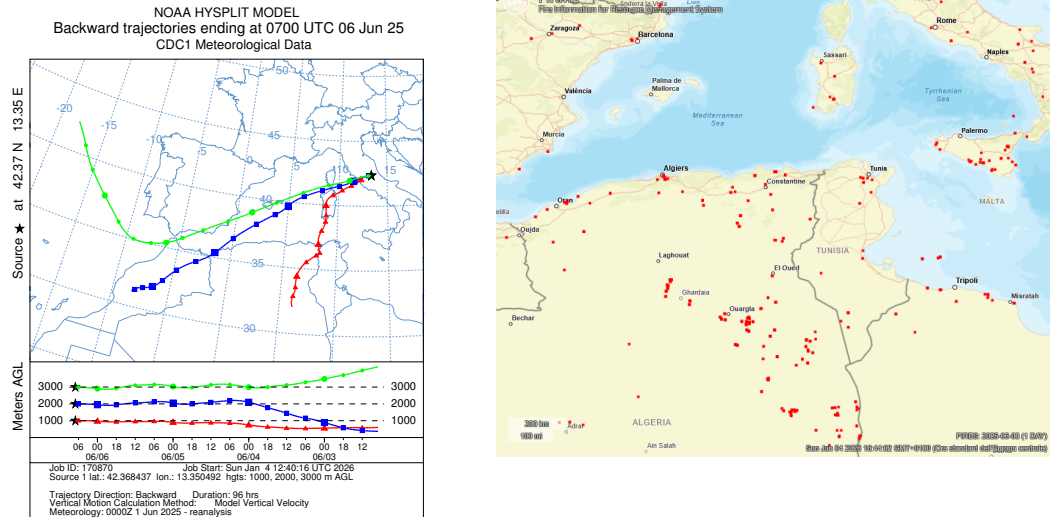


Figure 6.3: *Left:* back-trajectory for June 6, 2025 at 7 UTC. *Right:* Active fires on June 3, 2025 (NASA/LANCE/FIRMS). On that day the lowest HYSPLIT trajectory (red line) passes over active fire's area (red points) along the border between Algeria and Tunisia in the central part of the map.

6.3.2 Methods

As mentioned above, to extend the work of Román et al. (2018) to include nighttime observations, it requires some changes and assumptions. Indeed, with the current technology in photometers, it is not possible to measure sky radiance at nighttime, but AOD only. This seriously limits the retrieval of aerosol optical properties as they are strongly influenced by the phase function (Dubovik et al. 2006). Strategies have been proposed in literature (e.g., Benavent-Oltra et al. 2019), mostly relying on the assumption of constant optical properties during nighttime, or interpolating optical properties between last observation before night and first observation after night. Avoid completely the use of other input data and relying on spectral AOD only, it has been proved to be a fruitful approach, but it requires using a simpler log-normal size distribution to constrain enough the inversion (Torres et al. 2017). This latter approach is less general as it requires assuming in advance the number of modes in the size distribution.

Following Román et al. (2018), wavelengths at 440 nm, 675 nm, 870 nm and 1020 nm are used for photometric sky radiance and AOD, and at 910 nm for ceilometer's backscatter

Table 6.1: Parameters that can be retrieved from photometer-ceilometer synergy in GRASP.

Integrated properties	Vertically-resolved properties
<i>Micro-physical</i>	<i>Micro-physical</i>
Size distribution	Aerosol profile
Sphere fraction	<i>Optical (extensive)</i>
Aerosol concentration	Extinction coefficient
<i>Optical (intensive)</i>	Backscatter coefficient
Real refractive index	Absorption coefficient
Imaginary refractive index	Scattering coefficient
Aerosol optical depth	
Absorption AOD	
Angstrom exponent	
Single-scattering albedo	
Lidar ratio	

signal. Since the ceilometer has only one wavelength, it is not possible to discriminate between different aerosol modes along the vertical because this is accomplished through a different spectral response between small and large particles. Then, it is necessary to assume a single type of aerosol in the whole column, with unique micro-physical and optical intensive properties, and eventually with an uneven concentration along the profile.

Starting from considerations above, the following work assesses a mixed approach. In daytime the approach described in Román et al. (2018) is used. In nighttime, the approach described in Torres et al. (2017) is used, but with a more general binned size distribution to retain the skill to catch multi-modal distributions, and the addition of ceilometer profile data. Table 6.1 shows the aerosol properties that can be retrieved with the Román et al. (2018) method.

For this work the version v1.1.6 of GRASP ¹ is used. The GRASP algorithm is available as source code, and it needs to be compiled and installed on Linux-based systems.

6.3.3 Datasets and preprocessing

6.3.3.1 Sky radiance

Sky radiance is available only during daytime, and it represents the primary data to check for coincidence with the other observation types. For this study, raw sky radiance from AERONET in the almucantar geometry has been used.

This dataset requires a number of checks, as described in Román et al. (2018), in order to be used as input for GRASP. First, the almucantar geometry is composed by left and right branches, then it is necessary to check that difference between any angle's pair

¹<https://code.grasp-open.com/open/grasp>, last check on April 1, 2026.

is less than 20 %. Second, the scattering angles are grouped into bins to check there is at least one observation in each bin, and no less than 10 observation in total. The angle bins (right open interval) are 3.2° - 6° , 6° - 30° , 30° - 78° , and 78° - 181° . Finally, the radiance must be normalized in respect to the extraterrestrial irradiance value. The latter given by the reference quiet solar spectrum version v02r01 in (Coddington et al. 2016) and available on the NOAA Climate Data Record (Coddington et al. 2017).

6.3.3.2 AOD

The AOD is obtained from AERONET L1.5 dataset Version 3 algorithm (Giles et al. 2019). The same dataset is used also for nighttime AOD observations to ensure consistency in the data processing, although in Section 5.3.1 it is shown a very good agreement between AOD from AERONET and RIMO methods, so the two dataset are interchangeable.

The AOD data is then resampled into 5 min bins using average value. This allows to reduce further the number of points and smooth the AOD variability, although L1.5 data is already cloud-screened.

The selection of coincident observations depends on the part of the day. In daytime, the 5-min AOD nearest to sky radiance observations is selected in a time window of ± 16 min. For nighttime case the 5-min AOD become the primary data to check for coincident ceilometer profiles.

6.3.3.3 RCS

The ceilometer signal is provided in terms of RCS as in Equation (3.13). The RCS is provided in the L1 NetCDF produced by the instrument software, with a time resolution of 30 s. Because the NetCDF contains a flag about cloud-contaminated profiles, such profiles have been removed to retain only cloud-free signals.

The RCS preprocessing mainly consists in the normal processing of lidar signals. First, the RCS is time-averaged into 30 min blocks, considering only the blocks with at least 5 profiles. Second, the RCS is range-averaged with a moving mean filter and using a 210 m window. This processing helps in removing the high-frequency noise in the backscatter signal.

To reduce, the computational load in the GRASP algorithm the RCS is further normalized into 60 logarithmic bins (Lopatin et al. 2013). This logarithmic space is commonly defined between the overlap height and an altitude that can be considered aerosol-free, that is between 250 m to 7000 m AGL usually (e.g., Lopatin et al. 2013; Román et al. 2018). The average of signals in each bin is normalized in respect the signal's integral between the minimum and maximum heights of the logarithmic space. As explained in Román et al. (2018), it can happen that the normalization provides negative values that are unphysical. In such case, the normalization is done again setting a lower maximum height, until the normalized signals are all positive. With the Vaisala CL51, negative values are encountered more frequently because of the signal sensitivity to temperature shown in Section 3.3.1, that very often allow normalization in daytime only decreasing maximum height down to 3000-4000 m. For easy of coding and computation,

such iterative procedure has been skipped, so that in case of negative normalized values the profile is discarded, and all profiles with successful normalization are up to 7000 m. On the contrary, lowering the maximum height produces normalized profiles with different vertical grids, and making disadvantageous to save such profiles into file format like NetCDF, which take advantage from variables sharing the same grid.

Finally, the RCS nearest to sky radiance (daytime) or AOD (nighttime) is selected in a time window of ± 16 min. In the latter case, since the AOD are grouped into 5 min bins, multiple AOD observations can be associated to the same RCS averaged profile, which spans a 30 min time range instead.

6.3.3.4 Surface reflectance

Surface reflectance in terms of *bidirectional reflectance distribution function* (BRDF) is required to constrain the inversion when sky radiance are used. The spectral BRDF, in units of sr^{-1} can be described by a variety of more or less empirical models (Strahler et al. 1999). In the GRASP context, the BRDF is described by the RossThick-LiSparse-R model that is expressed as (Strahler et al. 1999):

$$\begin{aligned} \text{BRDF}_\lambda = & f_{iso}(\lambda) \\ & + f_{vol}(\lambda) \cdot K_{vol}(\theta_S, \theta, \phi) \\ & + f_{geo}(\lambda) \cdot K_{geo}(\theta_S, \theta, \phi) \end{aligned} \quad (6.5)$$

where terms f_i are parameters for respective kernels K_i . The kernels depend on illumination zenith angle θ_S , and on viewing zenith θ and azimuth ϕ angles. The parameters are obtained by means of inverting observations against kernels. The model in Equation (6.5) is a superposition accounting for isotropic surface scattering (*iso*), volumetric scattering from a dense surface of small scatterers (*vol*), and geometric scattering from sparse scatterers accounting for their shadows (*geo*) (Strahler et al. 1999).

In GRASP, the spectral parameters f_i are required to compute the surface reflectance. These parameters are obtained from satellite MODIS MCD43A1 Version 6.1 Bidirectional Reflectance Distribution Function and Albedo Model Parameters (Schaaf and Wang 2021) data. This dataset provides daily values of spectral f_i at 500 m spatial resolution. These spectral f_i values are then used to compute a monthly climatology .

Finally, because MODIS sensors's wavelengths are not the same as the photometer, climatological spectral f_i values are linearly interpolated to the wavelengths used for this study.

Although not used in the current work, in case polarimetric observations are used, the surface reflectance data should provide also information about polarization. One possible dataset of polarized surface observation is provided by multi-angle polarimetric satellite's sensors, like POLDER-3 onboard of PARASOL satellite that collected data between 2005 to 2013. The polarimetric reflectance is described by means of *bidirectional polarized distribution function* (BPDF). Such function can be represented by several empirical models or machine learning techniques, see Liu et al. (2020) for more details and references.

Table 6.2: GRASP input parameters provided by each observation type.

	AOD	Radiance	RCS
Quantity			
Wavelengths	4	4	1
Angles	-	27	-
Heights	-	-	60
Total	4	108	60

6.3.4 GRASP input data

The preprocessed data shown before are provided to GRASP in a text file with SDATA format. All the observations spatially and temporally co-located form a single block of data allocated to a specific location and time instant. The SDATA file may contain multiple blocks of observations done at different times and/or locations (Dubovik et al. 2024).

A block of data contains values for each type of observation, each wavelength, each scattering angle/altitude (when applicable). In Table 6.2 is shown the number of values existing for each type of observations. Then, each block contains 172 values in daytime, but only 64 values at nighttime because sky radiance is not measured.

The multi-pixel approach can be used to account for information provided by adjacent pixels, both in space and in time, helping in increasing the aerosol characterization (Dubovik et al. 2021). The data collected for this case study come from co-located instruments for which could be suitable to apply the temporal multi-pixel approach, but, for sake of simplicity, a tight temporal co-location has been used during pre-processing stage in order to use a simpler single-pixel approach.

6.3.5 GRASP setting file

The GRASP setting with options for the forward model, the inversion, the a priori information, and the regularization constrain is provided in a YAML configuration file Dubovik et al. (2024). As said before, sky radiance is not available at nighttime, so it is necessary to use different configuration files for day and night. Moreover, the lack of angular scattering properties requires a decrease in the number of parameters that GRASP can retrieve. Following Benavent-Oltra et al. (2019) and Torres et al. (2017), the complex refractive index and the sphere fraction are fixed to values found in the nearest daytime retrieval, that in this case study correspond to values retrieved on June 6, 2025 morning.

First, it is necessary to provide the observation's uncertainty. The AOD has an absolute uncertainty of 0.01, and the sky radiance a relative uncertainty of 5 % (Holben et al. 1998). The RCS uncertainty is not known but can be assumed from statistical consideration. The ceilometer does not provide the signal variability, in terms of standard

deviation, for the 30 s sampling bins, but it is possible to obtain the standard deviation for 30 min temporal windows that are used to average the signal. This provides, in average, a relative uncertainty of about 25 % at 1-sigma. As described in Section 3.3.1, the water vapor content in the atmosphere can affect the backscatter coefficient by several percents (Wiegner and Gasteiger 2015). Despite the correction for water vapor absorption was not applied in this work, given the PW analysis performed in Chapter 4, it is possible to roughly estimate the error on the signal, since the backscatter coefficient is proportional to RCS, through Equations (3.11) and (3.13). Given the error values found by Wiegner and Gasteiger (2015) for mid-latitudes, it seems reasonable to assume a maximum 15 % of error given the low amount of water vapor column at LAO (see Chapter 4). The sum of these contributions provides a conservative relative uncertainty of 40 % for RCS.

Second, the particle size distribution is defined as a binned distribution with 22 logarithmic bins between $0.5\text{ }\mu\text{m}$ to $15\text{ }\mu\text{m}$. Such setting for the PSD is the most general, as it can catch also multi-modal aerosol distributions, and covering most of the dimensional range for common atmospheric aerosols. Moreover, this PSD setup is the same used in AERONET, so enabling to do a direct comparison of the two PSD.

Finally, it is necessary to set up the a priori and the regularization constraints to allow the convergence of the algorithm to the proper solution. Tuning the Lagrange parameters for these constraints took considerable efforts and the goal was achieved only for the PSD as will be shown next. Moreover, the a priori means providing a first-guess, along with the minimum and maximum values, for each parameter.

In Table 6.3 are reported the a priori (except the BRDF for sake of space) and the regularization constraints for the single-pixel solution. The a priori boundaries are fixed because reflects (reasonable) physical limits of the parameters. The first-guess roles is basically to speed up the convergence of the solution, so it is fixed for each parameter except for the size distribution, which is usually assumed equal to $\text{AOD}_{440}/10$ (Román et al. 2018). For this case study, given an average AOD_{440} of 0.39 (daytime) and 0.41 (nighttime), a value of 0.04 is used for the PSD first-guess in all retrievals. A total of 94 and 83 parameters must be retrieved by the algorithm in daytime and nighttime, respectively. The lower number in nighttime is due to the fixed values of refractive index and sphere fraction.

In case the information from observations are not enough to retrieve parameters, may be necessary to introduce assumptions to reduce the unknowns. As example, moving from triangular size distribution to log-normal distribution reduces the parameters to retrieve from 22 to 2, because only mean value and standard deviation are needed to construct the latter distribution (Torres et al. 2017). Similar considerations can be done using wavelength-constant refractive indices instead of spectral ones.

In this work several scripts has been developed in order to automatically preprocess the data, create the SDATA file, and update the setting file with spectral BRDF coefficients interpolated on-the-fly on the relevant wavelengths extracted from the SDATA file. The latter step requires a template for the setting file that must be created by the user in advance. This processing chain (Figure 6.4) reduces at minimum the manual efforts for the user, and has been developed taking into account the near real time retrieval of

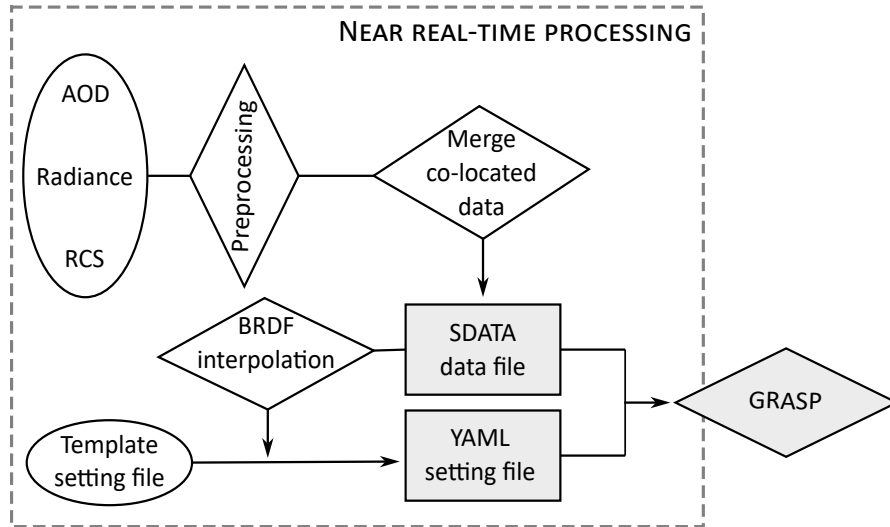


Figure 6.4: Schematic workflow of the automatized procedure for photometer-ceilometer synergy in GRASP.

aerosol, once properly configured.

The vertically-resolved parameters in Table 6.1 but not listed in Table 6.3, can be simply obtained by means of some calculations starting from the normalized aerosol profile (AP). As example, see Equation (6.6) with formula to obtain extinction and backscatter profiles.

6.3.6 Results

The RCS signal does not undergo to any adjustment in terms of background correction, both instrumental and atmospheric, in this work. As a consequence, the negative values in the signal prevents the normalization up to 7000 m most of the daytime. As example, the morning of June 5, 2025 (Figure 3.3) shows that already at 8am UTC the profile are affected by negative values.

In Figure 6.5 is shown the RCS after pre-processing (left) and after normalization (right). During daytime available normalized profiles drastically reduce in respect to profiles before normalization. Then, accounting also for co-located AOD/radiance observations, in this study case there are only 4 and 30 inversions available in daytime and nighttime, respectively.

As a matter of facts, the correction of background and pre-processing are fundamental steps for the correct usage of lidar signals into inversion algorithm, and they will be implemented in a next stage after this doctoral work.

Table 6.3: GRASP inversion setting for both day and night. For each parameter is shown the number of element to retrieve, the a priori first-guess (FG) value, the a priori (Λ_{ap}) and regularization (Λ_{reg}) Lagrange parameter, and the order of regularization derivative. The dash in night setting means that the same value of daytime is used. PSD: particle size distribution, RRI: real refractive index, IRI: imaginary refractive index, SF: sphere fraction, AC: aerosol concentration, AP: normalized aerosol profile.

	Nr.	Retr.	Day			Night			Order
			FG	Λ_{ap}	Λ_{reg}	FG	Λ_{ap}	Λ_{reg}	
PSD	22	Yes	0.04	10^{-5}	1	–	–	–	3
RRI	5	Yes [†]	1.45	10^{-2}	1	1.52	–	–	1
IRI	5	Yes [†]	5×10^{-3}	10^{-3}	0.1	6.3×10^{-3}	–	–	1
SF	1	Yes [†]	0.5	10^{-3}	0	0.1	–	–	0
AC	1	Yes	0.05	10^{-3}	0	–	–	–	0
AP	60	Yes	0.01	0	0.1	–	–	10^{-5}	3
BRDF	15	No	*	10^{-3}	0				0

[†] Parameters not retrieved at night.

* The spectral FG values for BRDF are not reported for sake of simplicity.

6.3.6.1 Fitting quality

After running inversions, a first check is the quality of fitting in terms of bias between the true observations and ones obtained by means of GRASP forward model using the best set of parameters retrieved. In general, in a good inversion the fitting bias should be well below the uncertainty associated to the observation.

For what concern the AOD, the fitting at each wavelength is well below the uncertainty of 0.01 for both day and night cases (Table 6.4).

In the case of RCS, some results are shown in Figure 6.6. In daytime, despite the pre-processing and log-normalization the observed signal clearly shows a noisy behavior above 3500 m (upper left plot). Considering the nominal $\pm 25\%$ 1-sigma relative noise of RCS signals, a very good agreement is found below such threshold height. On the upper right plot is reported the normalized residual by heights, averaged over all cases available in daytime. The normalized residual is defined as $(y_{fit} - y_{obs})/\sigma_{obs}$. Again, the absolute values of normalized residual, and then the agreement of fitting, is below 1 up to 3000 m. Upper heights have an oscillating behavior likely due to the noise in observations. The nighttime case (lower plots) highlights a difficulty of the fitting in reproducing the observations. Such behavior affects mainly the ranges between 3500 m to 5500 m. The fitted profile (lower left plot) seems to be constrained, and this is likely due to a too strong smoothing polynomial. Since the polynomial order and weight are the same for day and night inversions, and radiance information should be linked only to optical properties, the cause of such behavior is still not clear.

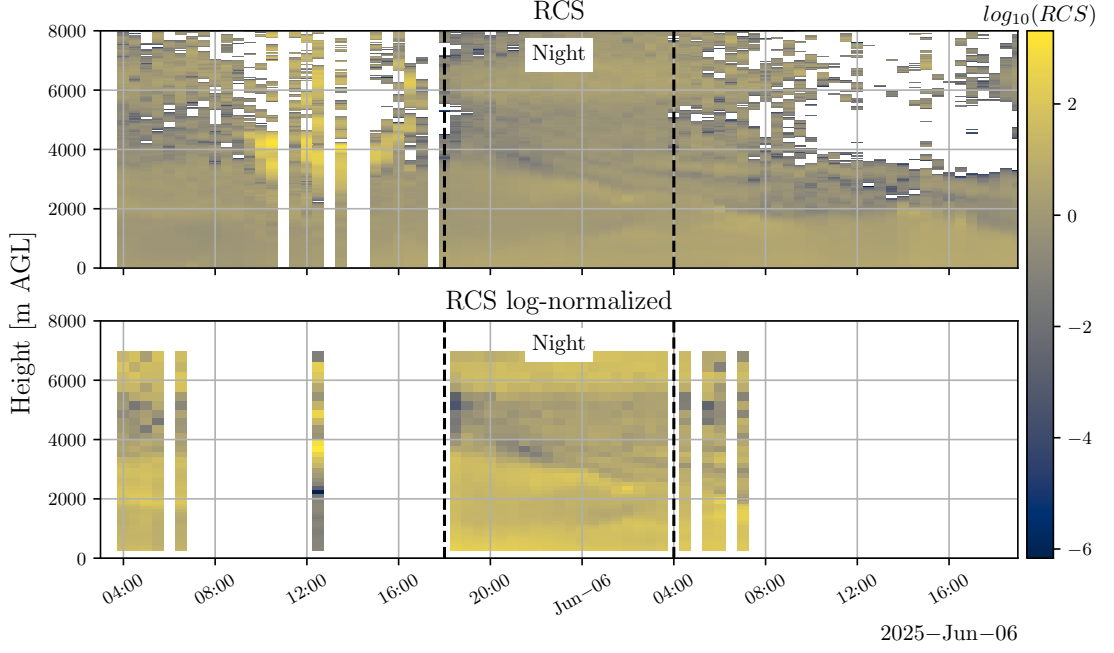


Figure 6.5: Noise effect on normalization of ceilometer signal. The pre-processed signal (upper plot) covers most of the time range analyzed except for central part of day on June 6, 2025 that is affected by clouds. In the normalized signals (lower plot), the presence of negative values below 7000 m drastically reduces availability of usable observations during daytime. Time is in UTC.

Finally, the radiance is measured only in daytime. Contrary to previous variables, the radiance is measured at scattering angles which values depend on the illuminating source’s zenith angle. Given the time required to measure all angles at one wavelength, the next wavelength measures radiance at different scattering angles because the zenith angle of the source changes slightly in the meanwhile. As a consequence, it is not possible to average the fitting statistics over time on a common scattering angle scale, as it was done above for the RCS on height scale. In Figure 6.7 are shown the angular distribution of normalized radiance at 440 nm (left plot), and the spectral distribution of normalized residuals computed in respect to the nominal $\pm 5\%$ 1-sigma relative noise of radiance observations (right plot). This latter plot highlights the very good fitting for radiance at each wavelength, as no residuals fall behind the threshold of 1-sigma dashed lines.

6.3.6.2 Retrieved parameters

The parameters directly retrieved by GRASP are shown in Table 6.3, and correspond to characteristics in the setting file for which is enabled the retrieval. Moreover, some additional optical parameters shown in Table 6.1, as for example SSA, are derived from them. For all of retrieved and derived parameters, GRASP can provide a standard

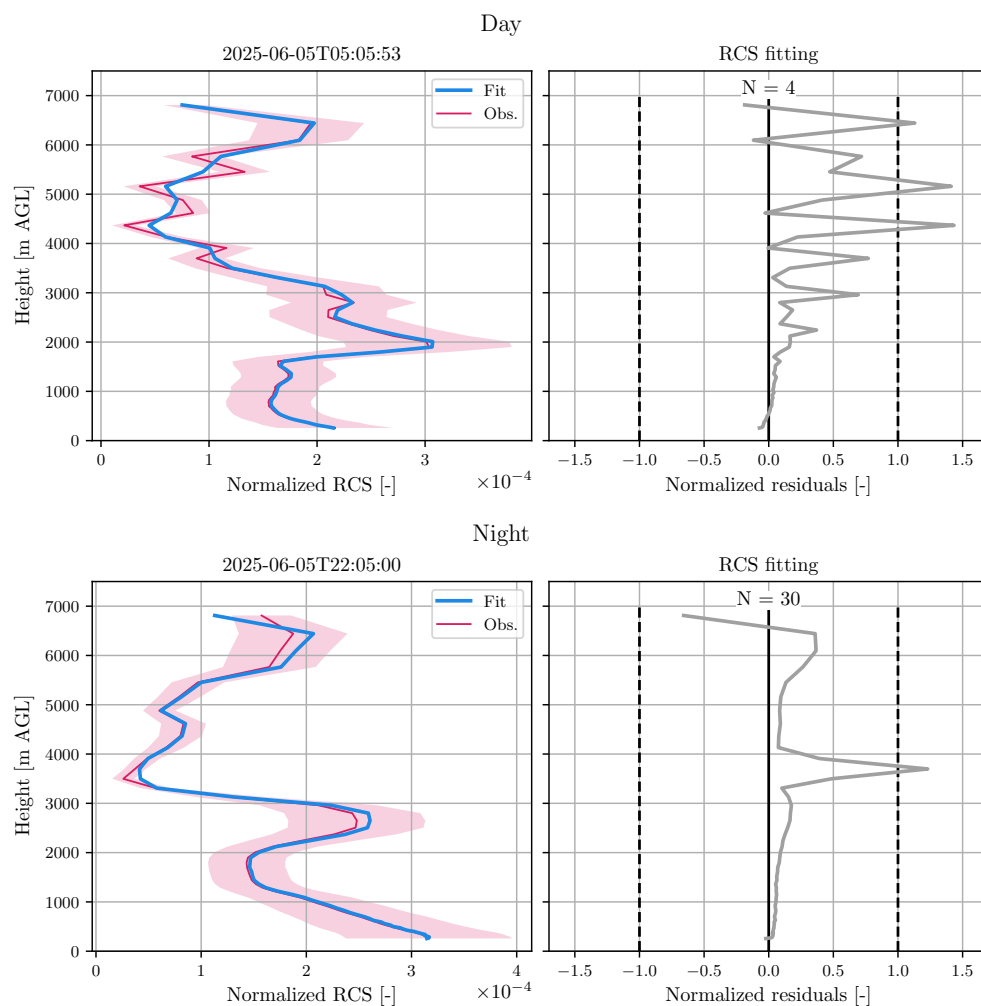


Figure 6.6: *Left:* single example of best-fit from retrieval against the observation for normalized RCS; the shaded area represents the instrumental $\pm 25\%$ 1-sigma relative noise associated to RCS observation. *Right:* average normalized residuals of fitting by height for all cases available (N); the dashed black lines are ± 1 -sigma uncertainty for RCS observations. Upper (lower) panels for daytime (nighttime) observations.

Table 6.4: Average bias (i.e., fitting minus observation) for spectral AOD, for day and night. N: number of observations.

Wavelength	Day		Night	
	N	Bias	N	Bias
440	4	-3.95×10^{-4}	30	0.22×10^{-4}
675	4	-0.27×10^{-4}	30	1.36×10^{-4}
870	4	6.40×10^{-4}	30	-6.28×10^{-4}
1020	4	4.92×10^{-4}	30	6.86×10^{-4}

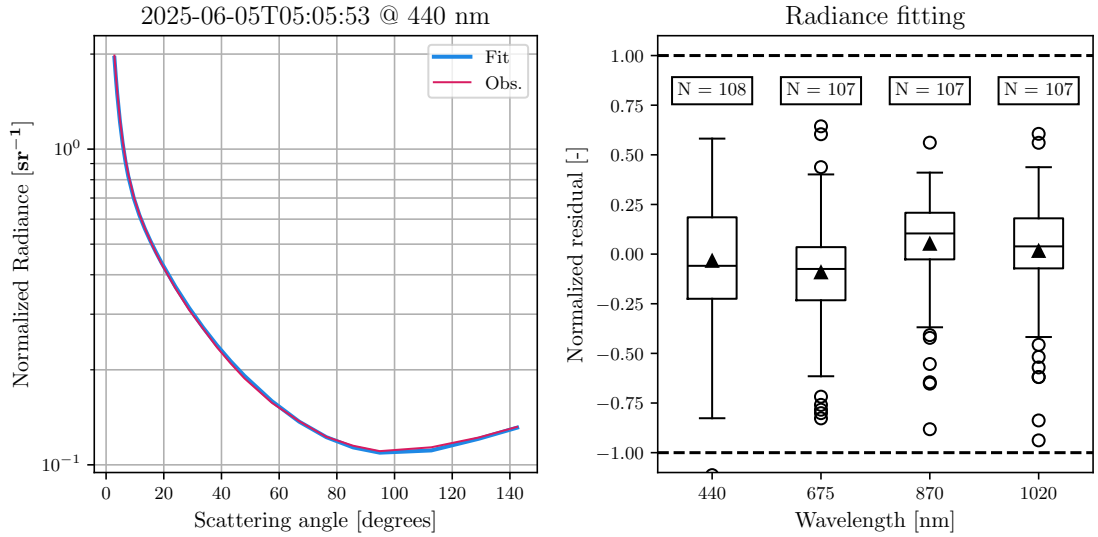


Figure 6.7: *Left:* single example of best-fit from retrieval against the observation, for angular distribution of normalized radiance at 440 nm. *Right:* spectral boxplots for normalized residuals of radiance fitting, accounting for all scattering angles and observations; the dashed black lines are ± 1 -sigma uncertainty for radiance observations, the horizontal solid line inside the box is the median, the triangle is the mean. The box extends from 25-th to 75-th percentile, and the whiskers extend to 1.5 times the interquartile range from these percentiles. The circles are outliers. N: number of observations.

deviation, from which it is possible to compute the parameter's uncertainty limits as shown in Equation (6.4). Despite the good fitting performance, the retrieval's uncertainty for this case study returned abnormal high values for all parameters, that may indicate some numerical instability due to the low information content of this synergy setup. Trying different values of a priori boundaries, Λ_{ap} and Λ_{reg} , it was possible to obtain acceptable uncertainty for PSD only. Except for the PSD, all the other parameters will be shown without the uncertainty. Because the error analysis was done only at the end of the doctoral work, no proper correction of that numerical instability has been possible, thus requiring additional study in the future. Despite this issue precludes to trust completely results of the retrieval, it is still possible to do an exploratory comparison of the retrieved parameters relying on simple statistical consideration only.

Most of retrieved parameters cannot be validated in this work because missing independent observations, especially at night, but, as reported in Román et al. (2018), there is a high correlation between the retrieved aerosol volume concentration and airborne and in-situ measurements. The only parameters that is possible to compare against a reference are the size distribution and refractive indices during daytime, for which exist retrievals from AERONET network. At nighttime AERONET does not provide any inversion. Given the availability of inversions (i.e., 4 in daytime, 30 in nighttime), only 2 daytime and 1 nighttime cases will be shown next, in order to evaluate the aerosol evolution in time, specifically at 5:05am and 9pm on June 5, 2025, and at 7:12am on June 6, 2025.

The particle size distribution in Figure 6.8 shows a good visual agreement against AERONET in daytime, and a reasonable evolution at nighttime. Indeed, the PSD evolves from a predominant coarse mode toward a predominant fine mode moving from one day to the other, while the nighttime PSD represents an intermediate case between them. Such evolution confirms the preliminary analysis done in Section 6.3.1 looking at AERONET data with the aid of back-trajectory simulations. Daytime cases can be analyzed on a statistical basis looking at the overall relative difference and root mean squared relative error (RMSRE) of GRASP retrievals in respect to AERONET. The median (mean) value is about -5% (-7%) and 4% (10%), for June 5, 2025 and June 6, 2025 cases, respectively. The RMSRE is about 33% and 56% , respectively. The big difference among median and mean on June 6, 2025 is explained by the large bias between GRASP and AERONET at the edges of the distribution, especially for the coarse mode, with the GRASP retrieval being higher in respect to AERONET. The difference at the edges may be due to different constraints used in GRASP and AERONET to force the edges to values near zero (i.e., in this work the maximum allowed values at the edge for GRASP is 0.0001). In general, more uncertainty at edges is expected because particles in such parts of the distribution contribute less to the information content retrieved by the photometer in respect to particles in the middle of the distribution (Herrera et al. 2022). On the other hand, since the AERONET PSD seems to stay within the uncertainty of retrieved PSD except for the largest particles in the distribution, such difference in the very edge of the coarse mode may come from the increased information content provided by the ceilometer. The nighttime PSD (Figure 6.8, central panel) has been computed using fixed

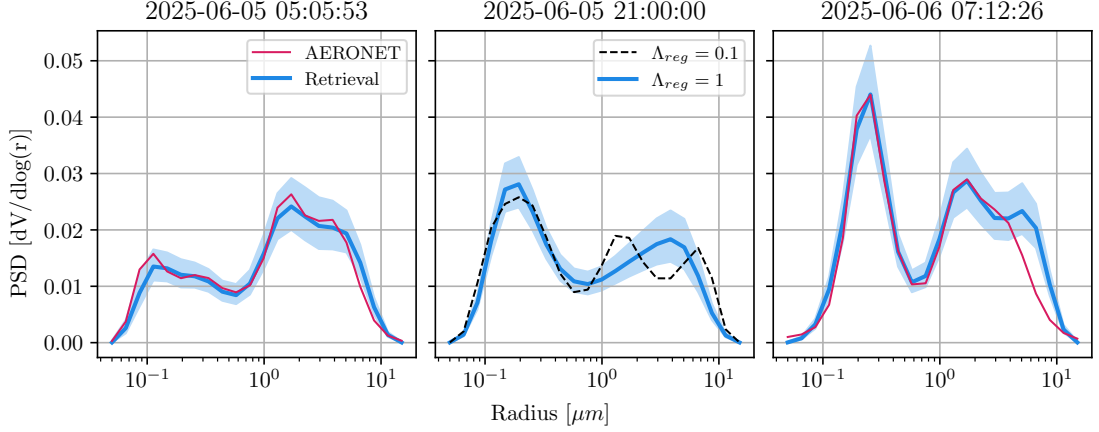


Figure 6.8: Evolution of the particle size distribution (PSD) for three selected times of the case study. The radius is in logarithmic scale. The shaded area is the $1\text{-}\sigma$ uncertainty computed by GRASP. The nighttime case (central panel) shows PSD with different smoothing Lagrange parameter (Λ_{reg}). $\Lambda_{reg} = 1$ is used also for daytime retrievals.

value of refractive indices and fraction of spherical particles from the nearest daytime inversion, that is spectrally averaging the retrieval on June 6, 2025 at 7:12am, giving values of $\text{RRI} = 1.53$ and $\text{IRI} = 0.0063$, as well as a percent of spherical particles of 36 %. The same panel shows the effect of regularization Lagrange parameters. Nighttime PSD has been smoothed with the same $\Lambda_{reg} = 1$ as for daytime, and with a less aggressive value equal to $\Lambda_{reg} = 0.1$. Is possible to see that using the same value of daytime with a reduced information content, as during nighttime, it may mask some features like the double peak in coarse mode. Further studies are required to understand if it may be worth to use different Λ_{reg} during day and night.

The refractive indices are compared only at the four wavelengths used by the photometer, but GRASP retrieves them at all wavelengths, also including the ceilometer 910 nm channel. Figure 6.9 shows a good agreement for what concern the real part, which follow the behavior of AERONET spectral real refractive index. The imaginary part shows more differences in respect to AERONET values, in agreement with the fact that the retrieval of imaginary part has a higher uncertainty in respect to the real part (Herrera et al. 2022). The spectral average of residuals for the real part gives a relative bias of 0.34 % and a RMSRE of 0.88 % on June 5, 2025 at 5:05am, and relative bias of -1.37% and $\text{RMSRE} = 1.38\%$ on June 6, 2025 at 7:12am. For the imaginary, the same statistics gives a relative bias of 1.28 % and 10.66 %, and a RMSRE of 39.22 % and 24.6 %, for the same days respectively. These results confirm the better skill in retrieving real refractive index in respect to the imaginary refractive index. In Figure 6.9, the nighttime refractive indices are equal to the spectral average of the one retrieved by nearest daytime inversion, because the information from AOD and ceilometer is not enough to correctly retrieve the refractive index (e.g., Benavent-Oltra et al. 2019; Torres

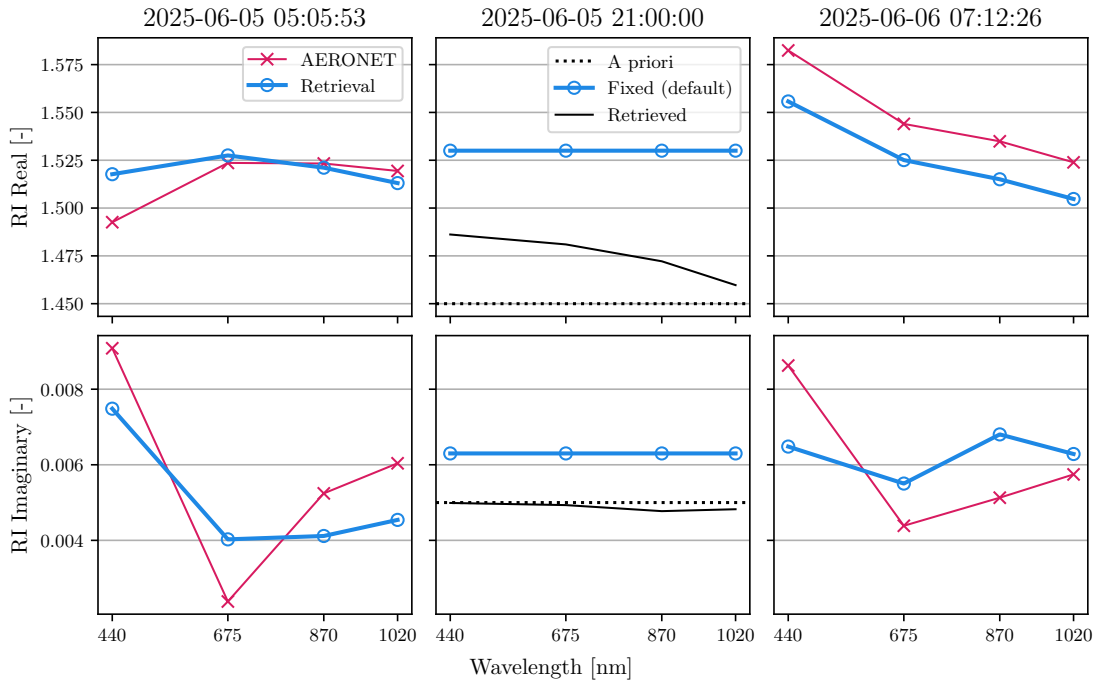


Figure 6.9: Spectral refractive index for three selected times of the case study. Upper panels for RRI, lower panels for IRI. At nighttime (central panel) is shown both the default method with fixed refractive indices from daytime retrieval on June 6, 2025 at 7:12am, and the refractive indices obtained from night retrieval using only AOD and RCS data (solid black). The a priori (dotted black) is the same for all cases except for the default method.

et al. 2017). Anyway, GRASP retrieval of refractive indices at nighttime has been tested as well. In such case, the daytime first-guess of Table 6.3 has been used, in order to compare against daytime retrievals. Refractive indices are retrieved but with unmeaningful results (Figure 6.9, central panel). Indeed, the magnitude of refractive indices is about the same value as the first-guess in the a priori, as confirmed by a relative bias of 1.71 % and -2.41 %, and a RMSRE of 1.84 % and 2.95 %, for RRI and IRI respectively. These results suggest that the lack of enough information does not allow GRASP to effectively converge toward a meaningful solution, although no strong constraint in a priori is used (i.e., large Lagrange parameter).

Instead of looking at the real and imaginary part of the refractive index separately, the overall optical behavior of an aerosol is better described by the SSA, as radiative transfer properties are sensitive to absorption via sensitivity to the SSA rather than the absolute value of absorption (Dubovik et al. 2002). The agreement of SSA retrieved by GRASP against the one retrieved by AERONET is very good, as shown in Figure 6.10. The average relative bias is 0.44 % and -0.75 %, and the RMSRE is 1.46 % and 1.31 %, for the two daytime cases on June 5, 2025 and June 6, 2025, respectively.

The spectral behavior of SSA, together with the PSD can be used for a first guess

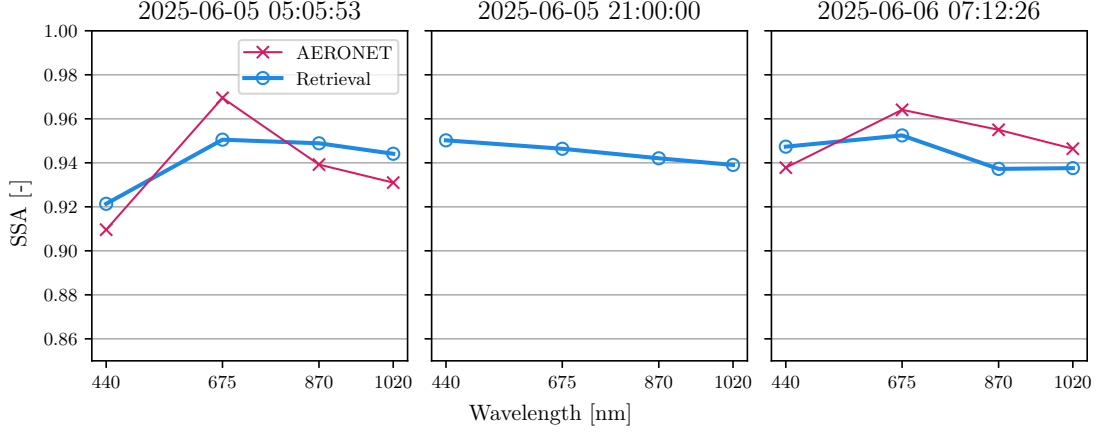


Figure 6.10: Evolution of the spectral SSA for three selected times of the case study.

about aerosol typing, using as reference values the properties shown in Figure 1.5. On June 5, 2025, SSA shows a very slightly spectral dependency with more absorption at shorter wavelengths, and the PSD shows a prevailing coarse mode. Such properties are likely representing a predominant load of desert dust in the aerosol mixture. On the other hand, on June 6, 2025, SSA is spectrally almost flat, and PSD shows a prevailing fine mode. These characteristics are likely representing a predominant load of urban aerosol, but with a notable amount of dust still there. This appears to be in contrast with the back-trajectories seen in Figure 6.3 which suggests clearly an advection of biomass burning-like aerosol. It is important to highlight how this kind of aerosol typing is very rough, because of photometer and ceilometer synergy which does not allow retrieving the optical properties for fine and coarse mode separately. Indeed, any optical properties (i.e., RRI, IRI, SSA) is associated to the mixture of aerosol, so it may be not possible to recognize the typical aerosol type behavior shown in Figure 1.5, as likely happens on June 6, 2025 morning case.

Finally, the profile information provided by the ceilometer allows retrieving some vertically-resolved aerosol characteristics by means of combination of the aerosol profile (AP) with other retrieved quantities. As example the aerosol volume concentration profile (VCP) is obtained multiplying AP with the scalar total aerosol volume concentration (VC), so that $VCP(z) = AP(z) \cdot VC$ at each height. The VCP is important to account the effect of aerosol layers in energy budget simulations. In Figure 6.11 the VCP is shown for illustrative purposes only, because of a sub-optimal setup of the inversion's parameters. Indeed, the nighttime case has relative large uncertainty, and on June 6, 2025 the profile is too noisy. Because the noise in daytime RCS is a relevant issue for ceilometers, again there might be a need for different inversion settings in a priori/regularization depending on daytime or nighttime retrieval. Currently, the $\Lambda_{ap} = 0$ as reported in Lopatin et al. (2013), gives the maximum weight to RCS observations, but on the other hand, as for the PSD, it is extremely difficult to define an informative a priori for such variable parameter.

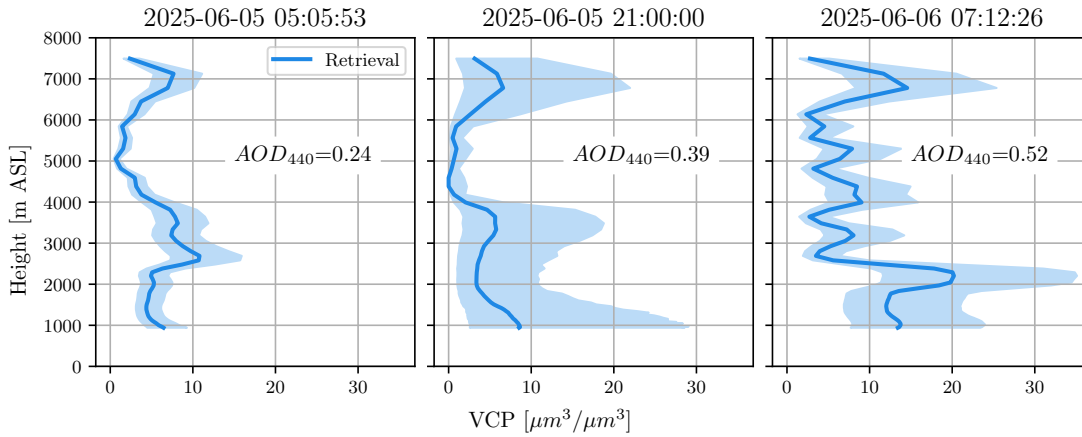


Figure 6.11: Evolution of the aerosol volume concentration profile for three selected times of the case study. The shaded area is the $1\text{-}\sigma$ uncertainty computed by GRASP.

However, the algorithm is able to detect the increase in AOD and allow identifying the lowering of main aerosol layer from about 3 km on June 5, 2025, to 2 km on June 6, 2025.

Another interesting parameters are the extinction (α) and backscatter (β) coefficient profiles, obtained as:

$$\begin{aligned}\alpha(\lambda, z) &= AOD(\lambda) \cdot AP(z) \\ \beta(\lambda, z) &= AOD(\lambda) \cdot \frac{AP(z)}{LR(\lambda)}\end{aligned}\tag{6.6}$$

where LR is the spectral lidar ratio provided by the inversion. As shown in Section 3.3, the extinction (backscatter) coefficient can be obtained by means of Klett inversion with independent lidar system or by models so that is possible to do a validation of the α (β) from GRASP inversion. Although initially planned, a comparison of β from GRASP against the one obtained from ECMWF-CAMS dataset has not been possible for a lack of time, but will be exploited in further study once the optimization of inversion parameters will be satisfying. Another way to validate the vertically-resolved extinction coefficient is shown in Herreras et al. (2019).

Chapter 7

Conclusions

This doctoral thesis investigated ground-based remote sensing techniques for the characterization of atmospheric aerosols and precipitable water vapor, with a particular focus on observations performed at the L'Aquila Atmospheric Observatory (LAO). The overarching objective of this work was to assess and extend the information content of sun-sky-lunar photometer measurements, both at daytime and nighttime, and to evaluate their integration with ceilometer observations through the GRASP inversion framework. The observational context of the study was then introduced, describing the LAO site and its role within major international monitoring networks. These networks highlight the importance of long-term, well-calibrated, and traceable measurements for climate-relevant applications, and part of the doctoral work consisted of carrying out maintenance and installation work at the LAO site in order to make it ready to participate in ACTRIS networks.

A significant component of this work was dedicated to the climatological analysis of precipitable water vapor at the L'Aquila site. PW estimates derived from sun-photometry were systematically compared with independent GNSS and radiosonde measurements over multiple temporal scales. The results demonstrate a generally good agreement among the different techniques, with residual biases largely attributable to cloud contamination and methodological differences. The analysis revealed a pronounced annual cycle and a clear diurnal modulation of PW, closely linked to near-surface temperature variations, suggesting that local radiative forcing and evapotranspiration processes dominate the short-term PW variability at the site. Some peculiar behavior in diurnal modulation was observed with sun-photometry during morning, but no clear explanation was found and still under investigation.

The thesis further explored last decade advancements in ground-based photometry, with particular emphasis on nighttime observations. A dedicated processing chain for retrieving aerosol optical depth from lunar photometry was implemented. The resulting AOD and AE were validated against AERONET nighttime observations, showing high correlation and low bias across multiple wavelengths. These results confirm the robustness of the lunar algorithm applied, and demonstrate its capability to extend aerosol monitoring into the nighttime, thereby improving the temporal continuity of aerosol observations.

But, it is necessary to account also the limits of lunar photometry. The uncertainty associated to the lunar algorithm is still higher in respect to solar photometry, and the geometry of the Moon in the sky remains a strong limitation to the availability of lunar observations. Next steps on lunar photometry will involve the implementation of remaining channel's corrections, like temperature sensitivity.

A central element of this work is the use of GRASP algorithm as a unified and statistically consistent framework for the inversion of heterogeneous remote sensing observations. This framework is used to retrieve aerosol micro-physical and optical properties, but provides also a rigorous quantification of retrieval errors and parameter's correlation, which is essential for the interpretation of complex multi-sensor inversions.

Building upon this foundation, a retrieval of aerosol properties from the synergy of sun-sky-lunar photometer and ceilometer observations was implemented in GRASP starting from previous works found in literature. The current doctoral work extends the article of Román et al. (2018) by explicitly accounting for nighttime measurements. A representative case study demonstrated that this synergy allows the retrieval of vertically resolved aerosol optical properties, despite intrinsic limitations due to a reduced spectral information from the ceilometer and missing sky radiance measurements at night from photometer. Despite various trials, was not possible to set up the algorithm for a satisfying inversion, resulting in large uncertainty and huge sensitivity to the constraints. Such case study highlighted the importance of a complete pre-processing of input data, especially for the ceilometer, and accurate tuning of GRASP through synthetic scenarios. Anyway, the preliminary results indicate that such a configuration represents a viable and cost-effective alternative to research-grade lidar systems for a simple aerosol characterization in the lower troposphere. Future studies will be aimed at improving the pre-processing of ceilometer data, and finding a better tuning of the algorithm starting from synthetic scenarios.

Overall, this doctoral work demonstrates that sun-sky-lunar photometry, when combined with complementary observations from ceilometer and a flexible inversion framework such as GRASP, can substantially enhance the characterization of atmospheric aerosols at the local scale. The methodologies developed and assessed in this work contribute to ongoing efforts to improve ground-based aerosol monitoring at L'Aquila site, including extended long-term analyses and integration within wider observational networks as ACTRIS, and provide a solid basis for future developments of real-time products that can be timely ingested in data-assimilation schemes of weather and climatic models.

Bibliography

- Acton, C.H. (1996). “Ancillary data services of NASA’s Navigation and Ancillary Information Facility”. In: *Planetary and Space Science* 44.1, pp. 65–70. ISSN: 0032-0633. DOI: 10.1016/0032-0633(95)00107-7.
- Acton, C.H. et al. (2018). “A look towards the future in the handling of space science mission geometry”. In: *Planetary and Space Science* 150, pp. 9–12. ISSN: 0032-0633. DOI: 10.1016/j.pss.2017.02.013.
- AERONET (2019). “Retrieval of nocturnal AOD on AERONET instruments 2015-2019. Transfer of Solar irradiance calibration to Lunar irradiance calibration using Mauna Loa and Izana observations”.
- Almansa, A.F. et al. (2020). “Column Integrated Water Vapor and Aerosol Load Characterization with the New ZEN-R52 Radiometer”. In: *Remote Sensing* 12.9, p. 1424. ISSN: 2072-4292. DOI: 10.3390/rs12091424.
- American Meteorological Society (2025). *Precipitable water*. Ed. by Glossary of Meteorology. URL: https://glossary.ametsoc.org/wiki/Precipitable_water.
- Ångström, A. (1929). “On the Atmospheric Transmission of Sun Radiation and on Dust in the Air”. In: *Geografiska Annaler* 11, p. 156. ISSN: 1651-3215. DOI: 10.2307/519399.
- Annex, A. et al. (2020). “SpiceyPy: a Pythonic Wrapper for the SPICE Toolkit”. In: *Journal of Open Source Software* 5.46, p. 2050. ISSN: 2475-9066. DOI: 10.21105/joss.02050.
- Askne, J. and Nordius, H. (1987). “Estimation of tropospheric delay for microwaves from surface weather data”. In: *Radio Science* 22.3, pp. 379–386. ISSN: 1944-799X. DOI: 10.1029/rs022i003p00379.
- Barreto, A. et al. (2019). “Evaluation of night-time aerosols measurements and lunar irradiance models in the frame of the first multi-instrument nocturnal intercomparison campaign”. In: *Atmospheric Environment* 202, pp. 190–211. ISSN: 1352-2310. DOI: 10.1016/j.atmosenv.2019.01.006.
- Barreto, A. et al. (2013). “A new method for nocturnal aerosol measurements with a lunar photometer prototype”. In: *Atmospheric Measurement Techniques* 6.3, pp. 585–598. ISSN: 1867-8548. DOI: 10.5194/amt-6-585-2013.
- Barreto, Á. et al. (2016). “The new sun-sky-lunar Cimel CE318-T multiband photometer – a comprehensive performance evaluation”. In: *Atmospheric Measurement Techniques* 9.2, pp. 631–654. ISSN: 1867-8548. DOI: 10.5194/amt-9-631-2016.
- Basili, P. et al. (2001). “Atmospheric water vapor retrieval by means of both a GPS network and a microwave radiometer during an experimental campaign in Cagliari,

- Italy, in 1999". In: *IEEE Transactions on Geoscience and Remote Sensing* 39.11, pp. 2436–2443. ISSN: 0196-2892. DOI: 10.1109/36.964980.
- Bedoya-Velázquez, A. et al. (2021). "Ceilometer inversion method using water-vapor correction from co-located microwave radiometer for aerosol retrievals". In: *Atmospheric Research* 250, p. 105379. ISSN: 0169-8095. DOI: 10.1016/j.atmosres.2020.105379.
- Benavent-Oltra, J.A. et al. (2019). "Different strategies to retrieve aerosol properties at night-time with the GRASP algorithm". In: *Atmospheric Chemistry and Physics* 19.22, pp. 14149–14171. ISSN: 1680-7324. DOI: 10.5194/acp-19-14149-2019.
- Benavent-Oltra, J.A. et al. (2021). "Overview of the SLOPE I and II campaigns: aerosol properties retrieved with lidar and sun-sky photometer measurements". In: *Atmospheric Chemistry and Physics* 21.12, pp. 9269–9287. ISSN: 1680-7324. DOI: 10.5194/acp-21-9269-2021.
- Berkoff, T.A. et al. (2011). "Nocturnal Aerosol Optical Depth Measurements with a Small-Aperture Automated Photometer Using the Moon as a Light Source". In: *Journal of Atmospheric and Oceanic Technology* 28.10, pp. 1297–1306. ISSN: 1520-0426. DOI: 10.1175/jtech-d-10-05036.1.
- Bevis, M. et al. (1992). "GPS meteorology: Remote sensing of atmospheric water vapor using the global positioning system". In: *Journal of Geophysical Research: Atmospheres* 97.D14, pp. 15787–15801. ISSN: 0148-0227. DOI: 10.1029/92jd01517.
- Bevis, M. et al. (1994). "GPS Meteorology: Mapping Zenith Wet Delays onto Precipitable Water". In: *Journal of Applied Meteorology* 33.3, pp. 379–386. ISSN: 1520-0450. DOI: 10.1175/1520-0450(1994)033<0379:gmmzwd>2.0.co;2.
- Bodhaine, B.A. et al. (1999). "On Rayleigh Optical Depth Calculations". In: *Journal of Atmospheric and Oceanic Technology* 16.11, pp. 1854–1861. ISSN: 1520-0426. DOI: 10.1175/1520-0426(1999)016<1854:orodc>2.0.co;2.
- Brasseur, G.P. and Jacob, D.J. (2017). *Modeling of Atmospheric Chemistry*. Cambridge University Press. ISBN: 9781107146969. DOI: 10.1017/9781107146969.
- Burrows, J.P. et al. (1998). "Atmospheric Remote-Sensing Reference Data from GOME: Part 1. Temperature-Dependent Absorption Cross-Sections of NO₂ in the 231–794 nm Range". In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 60.6, pp. 1025–1031. ISSN: 0022-4073. DOI: 10.1016/s0022-4073(97)00197-0.
- Burrows, J.P. et al. (1999). "Atmospheric Remote-Sensing Reference Data From GOME - 2. Temperature-Dependent Absorption Cross Sections Of O₃ In The 231–794nm Range". In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 61.4, pp. 509–517. ISSN: 0022-4073. DOI: 10.1016/s0022-4073(98)00037-5.
- Campanelli, M., Nakajima, T., and Olivieri, B. (2004). "Determination of the solar calibration constant for a sun-sky radiometer: proposal of an in-situ procedure". In: *Applied Optics* 43.3, p. 651. ISSN: 1539-4522. DOI: 10.1364/ao.43.000651.
- Campanelli, M. et al. (2018). "Precipitable water vapour content from ESR/SKYNET sun-sky radiometers: validation against GNSS/GPS and AERONET over three different sites in Europe". In: *Atmospheric Measurement Techniques* 11.1, pp. 81–94. ISSN: 1867-8548. DOI: 10.5194/amt-11-81-2018.

- Chen, C. et al. (2022). “Multi-angular polarimetric remote sensing to pinpoint global aerosol absorption and direct radiative forcing”. In: *Nature Communications* 13.1. ISSN: 2041-1723. DOI: 10.1038/s41467-022-35147-y.
- Chen, J. et al. (2025). “Water Vapor Correction in Measurements of Aerosol Backscatter Coefficients Using a 910 nm Vaisala CL51 Ceilometer”. In: *Remote Sensing* 17.12, p. 2013. ISSN: 2072-4292. DOI: 10.3390/rs17122013.
- Chen, S. et al. (2021). “Calibration of Automatic Sun Photometer with Temperature Correction in Field Environment”. In: *Remote Sensing* 14.1, p. 66. ISSN: 2072-4292. DOI: 10.3390/rs14010066.
- Coddington, O. et al. (2016). “A Solar Irradiance Climate Data Record”. In: *Bulletin of the American Meteorological Society* 97.7, pp. 1265–1282. ISSN: 1520-0477. DOI: 10.1175/bams-d-14-00265.1.
- Coddington, O. et al. (2017). *NOAA Climate Data Record (CDR) of Solar Spectral Irradiance (SSI), NRLSSI Version 2.1*. DOI: 10.7289/V53776SW.
- Copernicus Atmosphere Monitoring Service (2020). *CAMS global reanalysis (EAC4) monthly averaged fields*. DOI: 10.24381/FD75FFF2.
- Copernicus Climate Change Service (2018). *ERA5 hourly data on single levels from 1940 to present*. Copernicus Climate Change Service (C3S) Climate Data Store (CDS). DOI: 10.24381/CDS.ADBB2D47.
- (2019). *ERA5 explorer*. URL: <https://cds.climate.copernicus.eu/cdsapp#!/software/app-era5-explorer?tab=overview>.
- Curci, G. et al. (2021). “Building a local climate reference dataset: Application to the Abruzzo region (Central Italy), 1930–2019”. In: *International Journal of Climatology* 41.8, pp. 4414–4436. ISSN: 1097-0088. DOI: 10.1002/joc.7081.
- Dai, A. et al. (2002). “Diurnal variation in water vapor over North America and its implications for sampling errors in radiosonde humidity”. In: *Journal of Geophysical Research: Atmospheres* 107.D10. ISSN: 0148-0227. DOI: 10.1029/2001jd000642.
- Dos Santos Oliveira, D.C.F. et al. (2023). “Evaluation of the Accuracy of the Aerosol Optical and Microphysical Retrievals by the GRASP Algorithm from Combined Measurements of a Polarized Sun-Sky-Lunar Photometer and a Three-Wavelength Elastic Lidar”. In: *Remote Sensing* 15.20, p. 5010. ISSN: 2072-4292. DOI: 10.3390/rs15205010.
- Dubovik, O. et al. (2011). “Statistically optimized inversion algorithm for enhanced retrieval of aerosol properties from spectral multi-angle polarimetric satellite observations”. In: *Atmospheric Measurement Techniques* 4.5, pp. 975–1018. ISSN: 1867-8548. DOI: 10.5194/amt-4-975-2011.
- Dubovik, O., Ducos, F., and Fuertes, D. (2024). *The GRASP Package. An overview. Generalized Retrieval of Atmosphere and Surface Properties*. URL: <http://www.grasp-open.com/doc>.
- Dubovik, O. and King, M.D. (2000). “A flexible inversion algorithm for retrieval of aerosol optical properties from Sun and sky radiance measurements”. In: *Journal of Geophysical Research: Atmospheres* 105.D16, pp. 20673–20696. ISSN: 0148-0227. DOI: 10.1029/2000jd900282.

- Dubovik, O. et al. (2002). “Variability of Absorption and Optical Properties of Key Aerosol Types Observed in Worldwide Locations”. In: *Journal of the Atmospheric Sciences* 59.3, pp. 590–608. ISSN: 1520-0469. DOI: 10.1175/1520-0469(2002)059<0590:voaaop>2.0.co;2.
- Dubovik, O. et al. (2006). “Application of spheroid models to account for aerosol particle nonsphericity in remote sensing of desert dust”. In: *Journal of Geophysical Research: Atmospheres* 111.D11. ISSN: 0148-0227. DOI: 10.1029/2005jd006619.
- Dubovik, O. et al. (2014). “GRASP: a versatile algorithm for characterizing the atmosphere”. In: *SPIE Newsroom*. ISSN: 1818-2259. DOI: 10.1117/2.1201408.005558.
- Dubovik, O. et al. (2021). “A Comprehensive Description of Multi-Term LSM for Applying Multiple a Priori Constraints in Problems of Atmospheric Remote Sensing: GRASP Algorithm, Concept, and Applications”. In: *Frontiers in Remote Sensing* 2. ISSN: 2673-6187. DOI: 10.3389/frsen.2021.706851.
- Eck, T.F. et al. (1999). “Wavelength dependence of the optical depth of biomass burning, urban, and desert dust aerosols”. In: *Journal of Geophysical Research: Atmospheres* 104.D24, pp. 31333–31349. ISSN: 0148-0227. DOI: 10.1029/1999jd900923.
- Fedarenka, A. et al. (2016). “Utilization of AERONET polarimetric measurements for improving retrieval of aerosol microphysics: GSFC, Beijing and Dakar data analysis”. In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 179, pp. 72–97. ISSN: 0022-4073. DOI: 10.1016/j.jqsrt.2016.03.021.
- Fernald, F.G. (1984). “Analysis of atmospheric lidar observations: some comments”. In: *Applied Optics* 23.5, p. 652. ISSN: 1539-4522. DOI: 10.1364/ao.23.000652.
- Fragkos, K. et al. (2019). “Assessment of the total precipitable water from a sun photometer, microwave radiometer and radiosondes at a continental site in southeastern Europe”. In: *Atmospheric Measurement Techniques* 12.3, pp. 1979–1997. ISSN: 1867-8548. DOI: 10.5194/amt-12-1979-2019.
- Frey, L., Bender, F.A.-M., and Svensson, G. (2021). “Processes controlling the vertical aerosol distribution in marine stratocumulus regions – a sensitivity study using the climate model NorESM1-M”. In: *Atmospheric Chemistry and Physics* 21.1, pp. 577–595. ISSN: 1680-7324. DOI: 10.5194/acp-21-577-2021.
- Gaetani, M. et al. (2012). “A synoptic characterization of the dust transport and associated thermal anomalies in the Mediterranean basin”. In: *International Journal of Climatology* 36.7, pp. 2779–2791. ISSN: 1097-0088. DOI: 10.1002/joc.3615.
- Gassó, S. and Knobelspiesse, K.D. (2022). “Circular polarization in atmospheric aerosols”. In: *Atmospheric Chemistry and Physics* 22.20, pp. 13581–13605. ISSN: 1680-7324. DOI: 10.5194/acp-22-13581-2022.
- Geogdzhayev, I.V. and Marshak, A. (2018). “Calibration of the DSCOVR EPIC visible and NIR channels using MODIS Terra and Aqua data and EPIC lunar observations”. In: *Atmospheric Measurement Techniques* 11.1, pp. 359–368. ISSN: 1867-8548. DOI: 10.5194/amt-11-359-2018.
- Giles, D.M. et al. (2019). “Advancements in the Aerosol Robotic Network (AERONET) Version 3 database – automated near-real-time quality control algorithm with improved cloud screening for Sun photometer aerosol optical depth (AOD) measurements”.

- In: *Atmospheric Measurement Techniques* 12.1, pp. 169–209. ISSN: 1867-8548. DOI: 10.5194/amt-12-169-2019.
- Grainger, K.N. (2025). *Some useful formulae for particle size distributions and op-tical properties*. URL: <https://eodg.atm.ox.ac.uk/user/grainger/research/aerosols.pdf>.
- Halothore, R.N. et al. (1997). “Sun photometric measurements of atmospheric water vapor column abundance in the 940-nm band”. In: *Journal of Geophysical Research: Atmospheres* 102.D4, pp. 4343–4352. ISSN: 0148-0227. DOI: 10.1029/96jd03247.
- Hamed, K.H. and Ramachandra Rao, A. (1998). “A modified Mann-Kendall trend test for autocorrelated data”. In: *Journal of Hydrology* 204.1–4, pp. 182–196. ISSN: 0022-1694. DOI: 10.1016/S0022-1694(97)00125-X.
- Hashimoto, M. et al. (2012). “Development of a new data-processing method for SKYNET sky radiometer observations”. In: *Atmospheric Measurement Techniques* 5.11, pp. 2723–2737. ISSN: 1867-8548. DOI: 10.5194/amt-5-2723-2012.
- Heese, B. et al. (2010). “Ceilometer lidar comparison: backscatter coefficient retrieval and signal-to-noise ratio determination”. In: *Atmospheric Measurement Techniques* 3.6, pp. 1763–1770. ISSN: 1867-8548. DOI: 10.5194/amt-3-1763-2010.
- Herrera, M.E. et al. (2022). “Estimates of remote sensing retrieval errors by the GRASP algorithm: application to ground-based observations, concept and validation”. In: *Atmospheric Measurement Techniques* 15.20, pp. 6075–6126. ISSN: 1867-8548. DOI: 10.5194/amt-15-6075-2022.
- Herreras, M. et al. (2019). “Evaluation of retrieved aerosol extinction profiles using as reference the aerosol optical depth differences between various heights”. In: *Atmospheric Research* 230, p. 104625. ISSN: 0169-8095. DOI: 10.1016/j.atmosres.2019.104625.
- Holben, B.N. et al. (2006). “Aeronet’s Version 2.0 quality assurance criteria”. In: *Remote Sensing of the Atmosphere and Clouds*. Ed. by S.-C. Tsay et al. SPIE. DOI: 10.1117/12.706524.
- Holben, B.N. et al. (1998). “AERONET - A Federated Instrument Network and Data Archive for Aerosol Characterization”. In: *Remote Sensing of Environment* 66.1, pp. 1–16. ISSN: 0034-4257. DOI: 10.1016/S0034-4257(98)00031-5.
- Hopkin, E. et al. (2019). “A robust automated technique for operational calibration of ceilometers using the integrated backscatter from totally attenuating liquid clouds”. In: *Atmospheric Measurement Techniques* 12.7, pp. 4131–4147. ISSN: 1867-8548. DOI: 10.5194/amt-12-4131-2019.
- Hussain, M. and Mahmud, I. (2019). “pyMannKendall: a python package for non parametric Mann Kendall family of trend tests.” In: *Journal of Open Source Software* 4.39, p. 1556. ISSN: 2475-9066. DOI: 10.21105/joss.01556.
- Illingworth, A.J. et al. (2019). “How Can Existing Ground-Based Profiling Instruments Improve European Weather Forecasts?” In: *Bulletin of the American Meteorological Society* 100.4, pp. 605–619. ISSN: 1520-0477. DOI: 10.1175/bams-d-17-0231.1.
- Inness, A. et al. (2019). “The CAMS reanalysis of atmospheric composition”. In: *Atmospheric Chemistry and Physics* 19.6, pp. 3515–3556. ISSN: 1680-7324. DOI: 10.5194/acp-19-3515-2019.

- Intergovernmental Panel on Climate Change (2023). *Climate Change 2021 – The Physical Science Basis: Working Group I Contribution to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press. ISBN: 9781009157896. DOI: 10.1017/9781009157896.
- Kalashnikova, O.V. et al. (2011). “Sensitivity of multi-angle photo-polarimetry to vertical layering and mixing of absorbing aerosols: Quantifying measurement uncertainties”. In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 112.13, pp. 2149–2163. ISSN: 0022-4073. DOI: 10.1016/j.jqsrt.2011.05.010.
- Kalinnikov, V.V. and Khutorova, O.G. (2017). “Diurnal variations in integrated water vapor derived from a GPS ground network in the Volga–Ural region of Russia”. In: *Annales Geophysicae* 35.3, pp. 453–464. ISSN: 1432-0576. DOI: 10.5194/angeo-35-453-2017.
- Kasten, F. and Young, A.T. (1989). “Revised optical air mass tables and approximation formula”. In: *Applied Optics* 28.22, p. 4735. ISSN: 1539-4522. DOI: 10.1364/ao.28.004735.
- Kazadzis, S. et al. (2018). “The World Optical Depth Research and Calibration Center (WORCC) quality assurance and quality control of GAW-PFR AOD measurements”. In: *Geoscientific Instrumentation, Methods and Data Systems* 7.1, pp. 39–53. ISSN: 2193-0864. DOI: 10.5194/gi-7-39-2018.
- Khatri, P. and Takamura, T. (2009). “An Algorithm to Screen Cloud-Affected Data for Sky Radiometer Data Analysis”. In: *Journal of the Meteorological Society of Japan. Ser. II* 87.1, pp. 189–204. ISSN: 2186-9057. DOI: 10.2151/jmsj.87.189.
- Kieffer, H.H. (1997). “Photometric Stability of the Lunar Surface”. In: *Icarus* 130.2, pp. 323–327. ISSN: 0019-1035. DOI: 10.1006/icar.1997.5822.
- Kieffer, H.H. and Stone, T.C. (2005). “The Spectral Irradiance of the Moon”. In: *The Astronomical Journal* 129.6, pp. 2887–2901. ISSN: 1538-3881. DOI: 10.1086/430185.
- Klett, J.D. (1981). “Stable analytical inversion solution for processing lidar returns”. In: *Applied Optics* 20.2, p. 211. ISSN: 1539-4522. DOI: 10.1364/ao.20.000211.
- Komhyr, W.D., Grass, R.D., and Leonard, R.K. (1989). “Dobson spectrophotometer 83: A standard for total ozone measurements, 1962–1987”. In: *Journal of Geophysical Research: Atmospheres* 94.D7, pp. 9847–9861. ISSN: 0148-0227. DOI: 10.1029/jd094id07p09847.
- Kotthaus, S. et al. (2016). “Recommendations for processing atmospheric attenuated backscatter profiles from Vaisala CL31 ceilometers”. In: *Atmospheric Measurement Techniques* 9.8, pp. 3769–3791. ISSN: 1867-8548. DOI: 10.5194/amt-9-3769-2016.
- Kovalev, V.A. and Eichinger, W.E. (2004). *Elastic Lidar: Theory, Practice, and Analysis Methods*. Wiley. ISBN: 9780471643173. DOI: 10.1002/0471643173.
- Kudo, R. et al. (2021). “Optimal use of the Prede POM sky radiometer for aerosol, water vapor, and ozone retrievals”. In: *Atmospheric Measurement Techniques* 14.5, pp. 3395–3426. ISSN: 1867-8548. DOI: 10.5194/amt-14-3395-2021.
- Lach  rade, S. et al. (2014). “POLO: a unique dataset to derive the phase angle dependence of the Moon irradiance”. In: *Sensors, Systems, and Next-Generation Satellites XVIII*.

- Ed. by R. Meynart, S.P. Neeck, and H. Shimoda. Vol. 9241. SPIE, p. 924112. DOI: 10.1117/12.2067283.
- Leiterer, U. et al. (1995). “A new star photometer developed for spectral aerosol optical thickness measurements in Lindenberg”. In: *Contributions to Atmospheric Physics* 68.2. Cited by: 29, pp. 133–141. URL: <https://www.scopus.com/inward/record.uri?eid=2-s2.0-0029474873&partnerID=40&md5=cfd08c878d3d7b0a78884b37ffca613>.
- Li, L. et al. (2014). “A method to calculate Stokes parameters and angle of polarization of skylight from polarized CIMEL sun/sky radiometers”. In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 149, pp. 334–346. ISSN: 0022-4073. DOI: 10.1016/j.jqsrt.2014.09.003.
- Li, Z. et al. (2006). “Retrieval of aerosol optical and physical properties from ground-based spectral, multi-angular, and polarized sun-photometer measurements”. In: *Remote Sensing of Environment* 101.4, pp. 519–533. ISSN: 0034-4257. DOI: 10.1016/j.rse.2006.01.012.
- Li, Z. et al. (2009). “Improvements for ground-based remote sensing of atmospheric aerosol properties by additional polarimetric measurements”. In: *Journal of Quantitative Spectroscopy and Radiative Transfer* 110.17, pp. 1954–1961. ISSN: 0022-4073. DOI: 10.1016/j.jqsrt.2009.04.009.
- Li, Z. et al. (2010). “Calibration of the degree of linear polarization measurement of polarized radiometer using solar light”. In: *Applied Optics* 49.8, p. 1249. ISSN: 1539-4522. DOI: 10.1364/ao.49.001249.
- Li, Z. et al. (2018). “Calibration of the degree of linear polarization measurements of the polarized Sun-sky radiometer based on the POLBOX system”. In: *Applied Optics* 57.5, p. 1011. ISSN: 2155-3165. DOI: 10.1364/ao.57.001011.
- Liou, K.-N. (2002). *An introduction to atmospheric radiation*. 2nd ed. International geophysics series v. 84. Includes bibliographical references (p. 543-555) and index. - Description based on print version record. Amsterdam: Academic Press. 583 pp. ISBN: 0080491677.
- Liu, S. et al. (2020). “Modeling Bidirectional Polarization Distribution Function of Land Surfaces Using Machine Learning Techniques”. In: *Remote Sensing* 12.23, p. 3891. ISSN: 2072-4292. DOI: 10.3390/rs12233891.
- Lopatin, A. et al. (2013). “Enhancement of aerosol characterization using synergy of lidar and sun-photometer coincident observations: the GARRLiC algorithm”. In: *Atmospheric Measurement Techniques* 6.8, pp. 2065–2088. ISSN: 1867-8548. DOI: 10.5194/amt-6-2065-2013.
- Lopatin, A. et al. (2021). “Synergy processing of diverse ground-based remote sensing and in situ data using the GRASP algorithm: applications to radiometer, lidar and radiosonde observations”. In: *Atmospheric Measurement Techniques* 14.3, pp. 2575–2614. ISSN: 1867-8548. DOI: 10.5194/amt-14-2575-2021.
- Madonna, F. et al. (2018). “Intercomparison of aerosol measurements performed with multi-wavelength Raman lidars, automatic lidars and ceilometers in the framework of INTERACT-II campaign”. In: *Atmospheric Measurement Techniques* 11.4, pp. 2459–2475. ISSN: 1867-8548. DOI: 10.5194/amt-11-2459-2018.

- Mattis, I., Geiß, A., and Wagner, F. (2024). *Quality assurance of automatic low power lidars and ceilometers (ALCs)*. Presentation at the ACTRIS Week 2024, Matera, Italy.
- Miladi, L. et al. (2019). *Final description document of GARRLiC products*. Tech. rep. D10.15. Version 1.01. ACTRIS. URL: https://actris.eu/sites/default/files/Documents/ACTRIS-2/Deliverables/WP10_D10.15_M42.pdf (visited on 12/20/2025).
- Miladi, L. et al. (2025). *Generalized Aerosol Retrieval from Radiometer and Lidar-Combined data / Generalized Retrieval of Aerosol and Surface Properties (GARRLiC/GRASP)*. Tech. rep. Version 2.02. ACTRIS. URL: https://www.earlinet.org/index.php?eID=tx_securedownloads&p=125&u=0&g=0&t=1766333132&hash=21152df06179e9a71be6e4e2dc962e1de1c03520&file=fileadmin/user_upload/EARLINET_Database/GARRLiC_documentation_20250207.pdf (visited on 12/20/2025).
- Mishchenko, M.I. and Travis, L.D. (1997). “Satellite retrieval of aerosol properties over the ocean using polarization as well as intensity of reflected sunlight”. In: *Journal of Geophysical Research: Atmospheres* 102.D14, pp. 16989–17013. ISSN: 0148-0227. DOI: 10.1029/96jd02425.
- Momoi, M. et al. (2020). “Development of on-site self-calibration and retrieval methods for sky-radiometer observations of precipitable water vapor”. In: *Atmospheric Measurement Techniques* 13.5, pp. 2635–2658. ISSN: 1867-8548. DOI: 10.5194/amt-13-2635-2020.
- Mona, L. et al. (2006). “Saharan dust intrusions in the Mediterranean area: Three years of Raman lidar measurements”. In: *Journal of Geophysical Research: Atmospheres* 111.D16. ISSN: 0148-0227. DOI: 10.1029/2005jd006569.
- Nakajima, T. et al. (1996). “Use of sky brightness measurements from ground for remote sensing of particulate polydispersions”. In: *Applied Optics* 35.15, p. 2672. ISSN: 1539-4522. DOI: 10.1364/ao.35.002672.
- Nakajima, T. et al. (2020). “An overview of and issues with sky radiometer technology and SKYNET”. In: *Atmospheric Measurement Techniques* 13.8, pp. 4195–4218. ISSN: 1867-8548. DOI: 10.5194/amt-13-4195-2020.
- O’Connor, E.J., Illingworth, A.J., and Hogan, R.J. (2004). “A Technique for Autocalibration of Cloud Lidar”. In: *Journal of Atmospheric and Oceanic Technology* 21.5, pp. 777–786. ISSN: 1520-0426. DOI: 10.1175/1520-0426(2004)021<0777:atfaoc>2.0.co;2.
- Ortiz de Galisteo, J.P. et al. (2011). “Diurnal cycle of precipitable water vapor over Spain”. In: *Quarterly Journal of the Royal Meteorological Society* 137.657, pp. 948–958. ISSN: 1477-870X. DOI: 10.1002/qj.811.
- Pappalardo, G. et al. (2014). “EARLINET: towards an advanced sustainable European aerosol lidar network”. In: *Atmospheric Measurement Techniques* 7.8, pp. 2389–2409. ISSN: 1867-8548. DOI: 10.5194/amt-7-2389-2014.
- Park, R.S. et al. (2021). “The JPL Planetary and Lunar Ephemerides DE440 and DE441”. In: *The Astronomical Journal* 161.3, p. 105. ISSN: 1538-3881. DOI: 10.3847/1538-3881/abd414.
- Perez-Ramirez, D. et al. (2008). “Development and calibration of a star photometer to measure the aerosol optical depth: Smoke observations at a high mountain site”. In: *Atmospheric Environment* 42.11, pp. 2733–2738. ISSN: 1352-2310. DOI: 10.1016/j.atmosenv.2007.06.009.

- Pérez-Ramírez, D. et al. (2014). "Evaluation of AERONET precipitable water vapor versus microwave radiometry, GPS, and radiosondes at ARM sites". In: *Journal of Geophysical Research: Atmospheres* 119.15, pp. 9596–9613. ISSN: 2169-8996. DOI: 10.1002/2014jd021730.
- Rolph, G., Stein, A., and Stunder, B. (2017). "Real-time Environmental Applications and Display sYstem: READY". In: *Environmental Modelling and Software* 95, pp. 210–228. ISSN: 1364-8152. DOI: 10.1016/j.envsoft.2017.06.025.
- Román, R. et al. (2018). "Retrieval of aerosol profiles combining sunphotometer and ceilometer measurements in GRASP code". In: *Atmospheric Research* 204, pp. 161–177. ISSN: 0169-8095. DOI: 10.1016/j.atmosres.2018.01.021.
- Román, R. et al. (2020). "Correction of a lunar-irradiance model for aerosol optical depth retrieval and comparison with a star photometer". In: *Atmospheric Measurement Techniques* 13.11, pp. 6293–6310. ISSN: 1867-8548. DOI: 10.5194/amt-13-6293-2020.
- Sasano, Y., Browell, E.V., and Ismail, S. (1985). "Error caused by using a constant extinction/backscattering ratio in the lidar solution". In: *Applied Optics* 24.22, p. 3929. ISSN: 1539-4522. DOI: 10.1364/ao.24.003929.
- Schaaf, C. and Wang, Z. (2021). *MODIS/Terra+Aqua BRDF/Albedo Model Parameters Daily L3 Global - 500m V061*. DOI: 10.5067/MODIS/MCD43A1.061.
- Scorzini, A.R. and Leopardi, M. (2018). "Precipitation and temperature trends over central Italy (Abruzzo Region): 1951–2012". In: *Theoretical and Applied Climatology* 135.3–4, pp. 959–977. ISSN: 1434-4483. DOI: 10.1007/s00704-018-2427-3.
- Seidelmann, P.K. et al. (2002). "Report of the IAU/IAG Working Group on Cartographic Coordinates and Rotational Elements of the Planets and Satellites: 2000". In: *Celestial Mechanics and Dynamical Astronomy* 82.1, pp. 83–111. ISSN: 0923-2958. DOI: 10.1023/a:1013939327465.
- Seinfeld, J.H. and Pandis, S.N. (1998). *Atmospheric chemistry and physics. From air pollution to climate change*. A Wiley Interscience publication. "A Wiley-Interscience publication." - Includes bibliographical references and index. New York: Wiley. 1326 pp. ISBN: 0471178160.
- Sen, P.K. (1968). "Estimates of the Regression Coefficient Based on Kendall's Tau". In: *Journal of the American Statistical Association* 63.324, pp. 1379–1389. ISSN: 1537-274X. DOI: 10.1080/01621459.1968.10480934.
- Shaw, G.E. (1983). "Sun Photometry". In: *Bulletin of the American Meteorological Society* 64.1, pp. 4–10. ISSN: 0003-0007. DOI: 10.1175/1520-0477(1983)064<0004:sp>2.0.co;2.
- Shaw, G.E., Reagan, J.A., and Herman, B.M. (1973). "Investigations of Atmospheric Extinction Using Direct Solar Radiation Measurements Made with a Multiple Wavelength Radiometer". In: *Journal of Applied Meteorology* 12.2, pp. 374–380. ISSN: 0021-8952. DOI: 10.1175/1520-0450(1973)012<0374:ioaeud>2.0.co;2.
- Sinyuk, A. et al. (2012). "Assessment of error in aerosol optical depth measured by AERONET due to aerosol forward scattering". In: *Geophysical Research Letters* 39.23. ISSN: 1944-8007. DOI: 10.1029/2012gl053894.

- Sinyuk, A. et al. (2020). “The AERONET Version 3 aerosol retrieval algorithm, associated uncertainties and comparisons to Version 2”. In: *Atmospheric Measurement Techniques* 13.6, pp. 3375–3411. ISSN: 1867-8548. DOI: 10.5194/amt-13-3375-2020.
- Smirnov, A. et al. (2000). “Cloud-Screening and Quality Control Algorithms for the AERONET Database”. In: *Remote Sensing of Environment* 73.3, pp. 337–349. ISSN: 0034-4257. DOI: 10.1016/S0034-4257(00)00109-7.
- Smirnov, A. et al. (2009). “Maritime Aerosol Network as a component of Aerosol Robotic Network”. In: *Journal of Geophysical Research: Atmospheres* 114.D6. ISSN: 0148-0227. DOI: 10.1029/2008jd011257.
- Sommer, M. et al. (2023). *GRUAN characterisation and data processing of the Vaisala RS41 radiosonde*. Tech. rep. GRUAN-TD-8. Version 1.0.0. GRUAN Lead Centre, DWD. 206 pp. URL: <https://www.gruan.org/documentation/gruan/td/gruan-td-8>.
- Stein, A.F. et al. (2015). “NOAA’s HYSPLIT Atmospheric Transport and Dispersion Modeling System”. In: *Bulletin of the American Meteorological Society* 96.12, pp. 2059–2077. ISSN: 1520-0477. DOI: 10.1175/bams-d-14-00110.1.
- Stier, P. et al. (2007). “Aerosol absorption and radiative forcing”. In: *Atmospheric Chemistry and Physics* 7.19, pp. 5237–5261. ISSN: 1680-7324. DOI: 10.5194/acp-7-5237-2007.
- Stone, T.C. et al. (2020). “The Moon as a Climate-Quality Radiometric Calibration Reference”. In: *Remote Sensing* 12.11, p. 1837. ISSN: 2072-4292. DOI: 10.3390/rs12111837.
- Strahler, A.H. et al. (1999). “MODIS BRDF/albedo product: algorithm theoretical basis document version 5.0”. In: *MODIS documentation* 23.4, pp. 42–47. URL: https://modis.gsfc.nasa.gov/data/atbd/atbd_mod09.pdf (visited on 12/22/2025).
- Swanson, R. et al. (2020). “The ARCSTONE Project to Calibrate Lunar Reflectance”. In: *2020 IEEE Aerospace Conference*. IEEE. DOI: 10.1109/aero47225.2020.9172629.
- Takamura, T., Nakajima, T., and SKYNET community group (2004). “Overview of SKYNET and its activities”. In: *Optica Pura Y Aplicada* 37.3, pp. 3303–3308. URL: <https://opa.sedoptica.es/en/article/342>.
- Theil, H. (1992). “A Rank-Invariant Method of Linear and Polynomial Regression Analysis”. In: *Henri Theil’s Contributions to Economics and Econometrics*. Springer Netherlands, pp. 345–381. ISBN: 9789401125468. DOI: 10.1007/978-94-011-2546-8_20.
- Toledano, C. et al. (2018). “Assessment of Sun photometer Langley calibration at the high-elevation sites Mauna Loa and Izaña”. In: *Atmospheric Chemistry and Physics* 18.19, pp. 14555–14567. ISSN: 1680-7324. DOI: 10.5194/acp-18-14555-2018.
- Toledano, C. et al. (2024). “LIME: Lunar Irradiance Model of ESA, a new tool for absolute radiometric calibration using the Moon”. In: *Atmospheric Chemistry and Physics* 24.6, pp. 3649–3671. ISSN: 1680-7324. DOI: 10.5194/acp-24-3649-2024.
- Torres, B. et al. (2010). “Precipitable water vapor characterization in the Gulf of Cadiz region (southwestern Spain) based on Sun photometer, GPS, and radiosonde data”. In: *Journal of Geophysical Research: Atmospheres* 115.D18. ISSN: 0148-0227. DOI: 10.1029/2009jd012724.

- Torres, B. et al. (2014). “Sensitivity of aerosol retrieval to geometrical configuration of ground-based sun/sky radiometer observations”. In: *Atmospheric Chemistry and Physics* 14.2, pp. 847–875. ISSN: 1680-7324. DOI: 10.5194/acp-14-847-2014.
- Torres, B. et al. (2017). “Advanced characterisation of aerosol size properties from measurements of spectral optical depth using the GRASP algorithm”. In: *Atmospheric Measurement Techniques* 10.10, pp. 3743–3781. ISSN: 1867-8548. DOI: 10.5194/amt-10-3743-2017.
- Uchiyama, A. et al. (2019). “Nocturnal aerosol optical depth measurements with modified sky radiometer POM-02 using the moon as a light source”. In: *Atmospheric Measurement Techniques* 12.12, pp. 6465–6488. ISSN: 1867-8548. DOI: 10.5194/amt-12-6465-2019.
- Van Malderen, R. et al. (2014). “A multi-site intercomparison of integrated water vapour observations for climate change analysis”. In: *Atmospheric Measurement Techniques* 7.8, pp. 2487–2512. ISSN: 1867-8548. DOI: 10.5194/amt-7-2487-2014.
- Vaquero-Martínez, J. et al. (2022). “Comparison of CIMEL sun-photometer and ground-based GNSS integrated water vapor over south-western European sites”. In: *Atmospheric Research* 275, p. 106217. ISSN: 0169-8095. DOI: 10.1016/j.atmosres.2022.106217.
- Wagner, T. et al. (2006). “Global trends (1996–2003) of total column precipitable water observed by Global Ozone Monitoring Experiment (GOME) on ERS-2 and their relation to near-surface temperature”. In: *Journal of Geophysical Research: Atmospheres* 111.D12. ISSN: 0148-0227. DOI: 10.1029/2005jd006523.
- Wallace, J.M. and Hoobs, P.V. (2006). *Atmospheric Science: An introductory survey*. Second edition. International Geophysics 92. Academic Press. 506 pp. ISBN: 0080499538.
- Wiegner, M. and Gasteiger, J. (2015). “Correction of water vapor absorption for aerosol remote sensing with ceilometers”. In: *Atmospheric Measurement Techniques* 8.9, pp. 3971–3984. ISSN: 1867-8548. DOI: 10.5194/amt-8-3971-2015.
- Wiegner, M. et al. (2014). “What is the benefit of ceilometers for aerosol remote sensing? An answer from EARLINET”. In: *Atmospheric Measurement Techniques* 7.7, pp. 1979–1997. ISSN: 1867-8548. DOI: 10.5194/amt-7-1979-2014.
- Wiegner, M. et al. (2019). “Aerosol backscatter profiles from ceilometers: validation of water vapor correction in the framework of CeiLinEx2015”. In: *Atmospheric Measurement Techniques* 12.1, pp. 471–490. ISSN: 1867-8548. DOI: 10.5194/amt-12-471-2019.
- Woodward, J.T. et al. (2022). “Measurements of absolute, SI-traceable lunar irradiance with the airborne lunar spectral irradiance (air-LUSI) instrument”. In: *Metrologia* 59.3, p. 034001. ISSN: 1681-7575. DOI: 10.1088/1681-7575/ac64dc.
- World Meteorological Organization (2005). *WMO/GAW Experts Workshop on a Global Surface-Based Network for Long Term Observations of Column Aerosol Optical Properties*. Tech. rep. WMO/TD-No. 1287. 148 pp. URL: <https://library.wmo.int/idurl/4/41246>.
- Zieger, P. et al. (2013). “Effects of relative humidity on aerosol light scattering: results from different European sites”. In: *Atmospheric Chemistry and Physics* 13.21, pp. 10609–10631. ISSN: 1680-7324. DOI: 10.5194/acp-13-10609-2013.

- Ziskin Ziv, S. et al. (2021). “The diurnal variability of precipitable water vapor derived from GPS tropospheric path delays over the Eastern Mediterranean”. In: *Atmospheric Research* 249, p. 105307. ISSN: 0169-8095. DOI: 10.1016/j.atmosres.2020.105307.