# 4 Direct observation of the sun for aerosol retrieval

Colette Brogniez, Jacqueline Lenoble, Glenn Shaw

## 4.1 Introduction

The simplest remote sensing methods rely on the observation of the extinction of radiation, as it is defined in Chapter 2, in Eqs 2.4 and 2.5 (Beer exponential extinction law), which we recall here as Eqs 4.1 and 4.2:

$$I(s_2) = I(s_1)\exp(-\tau_e), \tag{4.1}$$

where

$$\tau_e = \int_{s_1}^{s_2} \sigma_e(s)ds. \tag{4.2}$$

If the source radiance $I(s_1)$ is known, the simple measure of $I(s_2)$ at a distance $(s_2-s_1)$ from the source directly gives access to the total extinction optical thickness of the atmospheric layer between the points $s_1$ and $s_2$ along the radiation path length.

This procedure has been used with artificial sources, especially at the Earth's surface, in order to observe fog or pollution episodes. Among natural sources (passive remote sensing), the moon and stars can be used, but of course the sun is the most important and widely-used source.

Ground-based observation of the direct solar beam, known as sunphotometry, has been used for decades. Sunphotometers are often associated with instruments measuring the sky radiance (and eventually the sky polarization) in selected directions and at selected

J. Lenoble et al. (eds.), *Aerosol Remote Sensing*, Springer Praxis Books,
DOI 10.1007/978-3-642-17725-5_4, © Springer-Verlag Berlin Heidelberg 2013

wavelengths; this allows a more complete retrieval of the aerosol characteristics. A history of ground-based aerosol remote sensing and a description of present-day networks that use sunphotometry with sky observations is presented in Chapter 6. In this chapter, we will address the basic principles of sunphotometry and the problems linked to calibration.

From space, it seems less obvious as to how we might observe the sun directly attenuated by atmospheric layers. However, this is achieved by the so-called "occultation methods", allowing an efficient sounding of the stratosphere and higher layers of the troposphere. Section 4.3 presents the basic principles of satellite occultation methods as used in the Stratospheric Aerosol and Gas Experiment (SAGE), the Polar Ozone and Aerosol Measurement (POAM) instruments, and some others.

Section 4.4. presents a few conclusions.

## 4.2 Ground-based sunphotometry

At the Earth's surface, the transmitted solar radiation can easily be observed by a sunphotometer pointing to the sun. The monochromatic irradiance measured on a plane perpendicular to the solar beam is

$$E_\lambda = E_{0\lambda} \exp(-M\tau_\lambda), \tag{4.3}$$

where $E_{0\lambda}$ is the extraterrestrial solar irradiance, $\tau_\lambda$ the vertical total atmospheric optical depth defined in Chapter 2, and $M$ the air mass depending on the solar zenith angle. For the plane parallel approximation, $M$ is the secant of the solar zenith angle. Note that atmospheric air mass deviates from the plane parallel approximation for low solar elevation angles. Calculations of $M$ for spherical atmospheres can and have been made and must be used for solar elevations less than about 10 degrees. Useful tables and convenient fitted closed form mathematical approximations for the air mass function are given in Kasten (1965) and Young (1994). Eq. (4.3) allows the determination of $\tau_\lambda$, if the extraterrestrial irradiance $E_{0\lambda}$ is known. Subtracting the gaseous optical depth (gas absorption and molecular scattering) from the total optical depth provides the aerosol optical depth.

Translating a ground-based measurement of the solar radiation into aerosol optical depth requires knowing the $E_{0\lambda}$ and the optical depths of water vapor, ozone and Rayleigh scattering for the bandpass of every wavelength of the instrument. Spectral transmission over the solar spectrum for the non-aerosol components is well known. All that is needed is the total column loading of these constituents. Ozone amounts can be estimated from climatology or, for more precise measurements, from open data sources. The magnitude of the Rayleigh scattering is a function of atmospheric pressure, which can be approximated by altitude above sea level or auxilliary information. Total column water vapor amounts can be obtained by auxilliary measurements, or derived from the sunphotometer measurements themselves using observations in highly absorbing bands. The best option in terms of correcting for water vapor absorption in the aerosol measurement is to choose spectral bands for the instrument that measure in the atmospheric "windows", avoiding water vapor absorption altogether. Assuming that the corrections for gas absorption and molecular scattering are known precisely, then the accuracy of the aerosol optical depth measurement is dependent on the accuracy of determining $E_{0\lambda}$.

Because the solution of Eq. 4.3 for $\tau_\lambda$ involves the ratio $E_\lambda / E_{0\lambda}$ the units of the value of the extraterrestrial irradiance are immaterial and only need to match the units measured by the sunphotometer. Often values are left in instrument-measured "voltage". The point is that the extraterrestrial irradiance needed in Eq. 4.3 to solve for $\tau_\lambda$ is the irradiance that would be measured by the individual instrument in question if that instrument were to be transported to the top of the atmosphere. Calibrating a sunphotometer involves determining $E_{0\lambda}$ for a particular instrument and for every wavelength band in that instrument. Calibration is a continuing necessity during the life span of the instrument, as band pass filters can degrade and other instrument characteristics can change over time (Ichoku et al., 2002a).

The typical approach to calibration is the so-called Langley (or Bouguer-Langley) method (Shaw et al., 1973), measuring the irradiance for a large range of solar zenith angles, and plotting $\ln(E_\lambda)$ versus $M$. The extrapolation of the plotted line to $M$=0 provides $E_{0\lambda}$, and the slope is $-\tau_\lambda$ (see Figure 4.1). This method was used extensively at the Smithsonian Institution between 1920 and 1950, with the objective of obtaining the solar constant, i.e.

$$E_0 = \int_0^\infty E_{0\lambda}\, d\lambda \tag{4.4}$$

The Langley plots are now used mainly for measuring the aerosol optical depth.

Langley plot calibration of sunphotometers, though simple in principle, has a number of subtle pitfalls. One must exercise great care in the Langley calibration because it is here that the largest and most difficult-to-define systematic errors occur. The Langley method assumes that, during the measurements at different elevation angles, the atmosphere is temporally invariant and horizontally homogeneous (within about 50 km of the observer). The latter problem may be reduced by choosing the observation location to ensure that power plant plumes or other interfering aerosol plumes are not passing through the field of view, but the problem of temporal stability of the atmosphere is much more difficult to satisfy. Almost all calibrations of sunphotometers conducted at continental locations have the possibility (and high probability) of being seriously marred because of time-changing drifts in atmospheric transmission. The only exceptions are those measurements made from high-altitude mountain observatories, but even those locations often have problems because of complex mountain meteorology and upslope thermally-driven currents. For example, though the air at the Mauna Loa Observatory (MLO) can be exceedingly transparent, it is affected in late morning and afternoon hours by marine aerosol that reaches the observatory as the marine inversion layer breaks down under solar heating. Langley plots from Mauna Loa give calibration constants varying slightly from day to day, by about one percent when data are taken all day long. However, when data acquired only before about 10 or 11 am local time are analyzed (before up-flowing air had reached the observatory) the calibration constants repeat to a few parts in a thousand day after day.

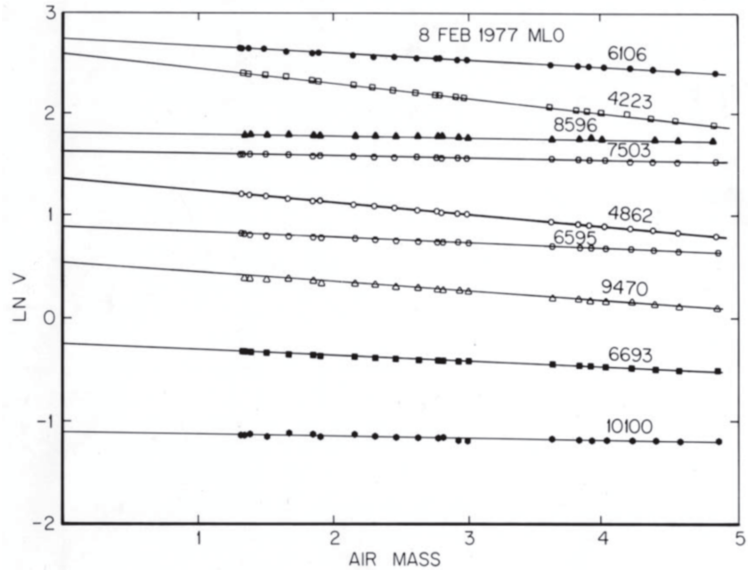



**Figure 4.1** Example of a Langley-Bouguer plot for multiwavelength sun photometer data taken at Mauna Loa. The ordinate is instrument voltage and the abscissa is the atmospheric air mass (secant of the solar zenith angle for plane parallel approximation). Theory suggests a straight line or linear relationship, but such a relationship can also occur when some of the assumptions in the theory formulation break down (see text). The individual line fits are for data taken at different narrow (10 nm) wavelength bands through well-blocked interference filters. Numbers indicate wavelengths in Ångströms.

At locations not as favorably endowed as Mauna Loa, the problem of time-varying atmospheric transmission, as it relates to the accuracy of calibration by the Langley method, is more severe and challenging to handle. What can be particularly annoying is that time variations in transparencies often have systematic trends of a type that provide a nearly linear Langley plot. If such time trends occur and if they persist day after day because of systematic diurnal trends in local meteorological conditions, then inaccurate calibration constants for the sunphotometer will be derived (Shaw, 1976) and – if cross-comparisons of instruments are made – propagate errors through an entire sunphotometer network.

To illustrate the insidious nature of the problem of local time drifts in atmospheric transparency, a modelled Langley plot is shown in Figure 4.2, for a situation when the atmospheric optical depth varies parabolically about the noon point; a case that actually occurs, to some approximation, quite frequently in polluted areas due to trapping and build up of aerosols under a temperature inversion and break up and dispersal of the aerosol in the afternoon hours. Photometers calibrated by the Langley method at Tucson, Arizona, for example, consistently show lower calibration voltage by 1 to 3 percent when compared to calibration conducted during pristine conditions in the morning hours at Mauna Loa because of this reason. Even though the Langley plots are consistently providing low values of extrapolated zero air mass intercepts at Tucson, the squared sum of the residuals about the least square fit to the exponential Bouguer law is often very low, because of the semi-parabolic variation of optical depth around solar noon. Notice that the modelled Langley plots are very close to straight lines, even though the optical depth is strongly varying (Shaw, 1976).

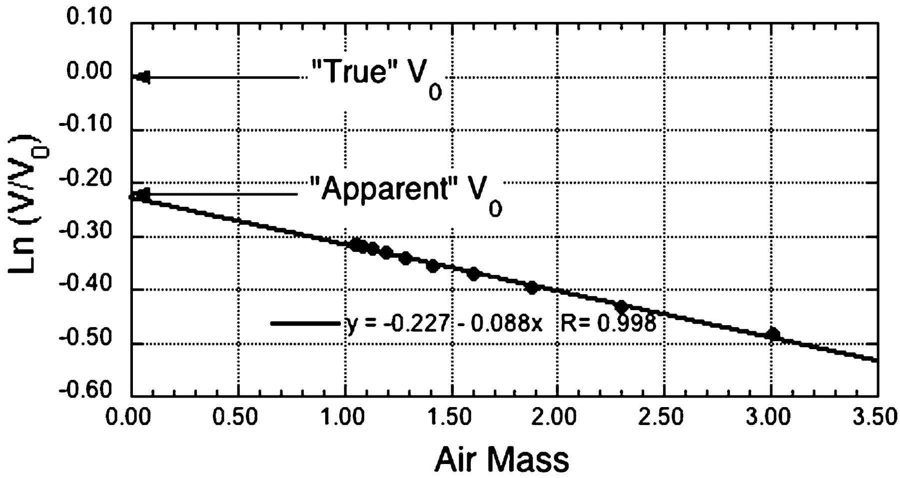
The only practical way to guard against this sort of calibration systematic errors is to conduct Langley plot calibrations from high-altitude observatories during stable conditions in the mornings and, simultaneously, constantly measure the aerosol particle concentration to ensure both that aerosols below the station remain below it during the calibration period and that the calibration repeats on several days. Even this is insufficient, because it may be possible that new particles are produced by sunlight, a kind of natural photochemical smog, which could conceivably maximize around the solar noon point. If this occurs, as Figure 4.2 illustrates, the calibration constants could still be in error. This error, however, would be very low if the aerosol optical depth is small, as indeed it typically is from an excellent station such as Mauna Loa.

Sunphotometers typically are manufactured with baffling and collimators to maintain a relatively small field of view (FOV) (several degrees) to minimize the scattered sky radiation. Though a correction for this forward-scattered radiation is performed, the photometers have to be accurately pointed at the sun. With handheld instruments, "operator error", the inability of an operator to keep the instrument steadily pointed at the solar disk, is a common source of inaccuracy.

Care must also be taken when performing accurate determinations of optical depth to ensure that the temperature of the detector/filter combination remains within fairly narrow bounds. It is common for optical detectors to have a rather large and usually positive temperature coefficient at the redder wavelengths (perhaps 1% per 10 degrees C), and a smaller and sometimes negative coefficient at bluer wavelengths, for example in the 350 to 450 nm region.

Sunphotometers are calibrated frequently either by referencing to a standard lamp or by comparing with a "master" instrument which has undergone accurate and extensive Langley plot calibration from a high-altitude mountain station with known excellent optical conditions, such as the Mauna Loa Observatory. It is well to use a great deal of caution when performing such intercomparisons, for a number of reasons. First, the spectral distribution of sunlight may be quite different in a pristine mountain environment compared with that in a polluted environment, and thus small amounts of signal due to out-of-band light leakage can be rather different for the two situations. Second, to be reliable, the pass bands and detector sensitivities in the two instruments (master and instrument undergoing calibration) must be either known or carefully included in the comparison, or must be identical, which is rarely the case. Even filters ordered in batch quantities often have spurious slight differences in pass band characteristics. It is quite possible that one of the two pass bands might contain a slight gaseous absorption feature not detected in the other, for instance. There must be other, perhaps not well understood, systematic error sources.

(a)




(b)

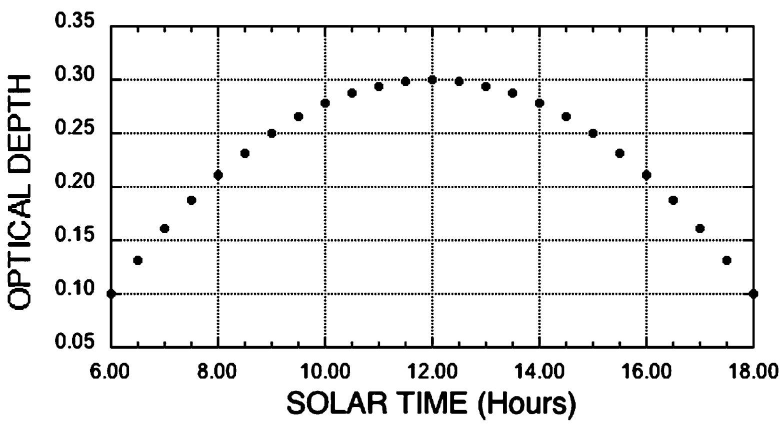



**Figure 4.2** (a) Modelled Langley-Bouguer plot for a hypothetical situation where the optical depth varies parabolically around solar noon. Note the goodness of the linear fit, in spite of massive breakdown in the assumption of optical depth remaining constant during the observation period. The "correct" zero air mass intercept in this modelled exercise is 1.0. Note the significant error in extrapolation.

(b) Assumed parabolically shaped variation of optical depth with time, varying around the solar noon point for the Langley model illustrated in Figure 4.2(a).

One of us (GS) frequently has seen unexpected differences of several percentage points in the extrapolated calibration constant when comparing two instruments, both referenced against master instruments that had been calibrated at Mauna Loa using the direct Langley method.

Extrapolated zero air mass intercepts can be performed to an accuracy of about one half percent, or possibly slightly better, if calibrations are very carefully performed under only ideal conditions, rejecting data when there is interfering aerosol plume as detected by a scattering nephelometer, or very thin cirrus or other visible sky contamination. A data set taken over a period of one year at MLO (Shaw, 1982) provided extrapolated values of zero air mass intercepts identical to one another within a half percent for most wavelength bands. Figure 4.3 shows a histogram for Langley plot intercepts taken during clean conditions at Mauna Loa. This, incidentally, indicates that extraterrestrial solar spectral irradiance was constant to at least half of one percent over a one-year period!

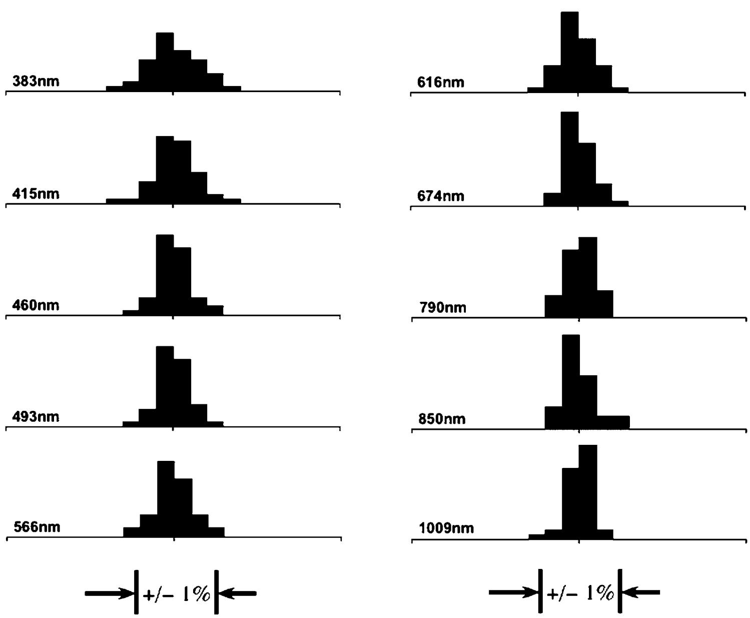



**Figure 4.3** Histograms of the frequency of occurrence for about 200 especially clean days of the zero air mass intercept of the extraterrestrial solar spectral irradiance from Langley plots taken at Mauna Loa Observatory, for several wavelengths indicated in nm. The rms variation in day-to-day extrapolated values is approximately 0.3 percent for most wavelengths.

Sometimes investigators calibrate sunphotometers by viewing a standard lamp; for example, type FEL 1000-watt lamps are frequently used in conjunction with stable electric current sources. One of us (GS) has had quite a lot of experience with intercomparisons of instruments calibrated in this manner, but with somewhat disappointing results. For example, a sunphotometer calibrated with a standard lamp, referencing an instrument accurately calibrated at MLO, might, when the secondary instrument is taken to MLO, provide zero air mass intercepts disagreeing by a few percentage points. The reasons probably have to do with the very different spectral distribution of the lamp's light (approximating a black body source at 1600 degrees C) with that of the sun (approximating a black body source at 5500 degrees C).

In summary, one should exercise a great deal of scepticism in calibrating sunphotometers. It must always be kept firmly in mind that knowing the zero air mass calibration value to only 2 percent error, a small calibration error for most instrumental calibration purposes, can bring inadmissible errors into sun photometry, especially in situations where the optical depths are quite small. For performing successful science or when assessing slow changes in atmospheric turbidity, one must keep an open mind and always stay alert to possible subtle errors that can and do creep into sun photometry. Calibration errors are often very subtle and difficult to detect. It is recommended that instruments at mountaintop observatories are calibrated regularly.

Another technique for measuring atmospheric turbidity, simpler, but less accurate, consists of measuring alternately global and diffuse irradiances on a horizontal plane with a movable shading disc or shading ring that occasionally cuts off the direct solar component. The direct sun signal is deduced by subtracting the "diffuse" sky radiation signal from the "global" (diffuse sky plus direct sun). This can be done using spectrometers (Lenoble et al., 2008, Brogniez et al., 2008), or filter radiometers (Bigelow et al., 1998). Such instruments need to be carefully calibrated and a good stability of the atmosphere during the two measures is necessary; this last condition is especially compelling for instruments that need quite a long time (a few minutes) for recording a large spectrum. In addition, a bias is due to the size of the shadower being larger than the solar disc; this bias is approximately corrected by computing the contribution of the circumsolar sky radiance, which needs an assumption about the atmospheric scattering properties (including aerosols).

A few sunphotometers, instead of being ground-based, operate from airborne platforms (Russell et al., 2005), enabling vertical profiles of aerosol optical thickness to be retrieved. A paragraph on airborne sunphotometers with ample references for further review can be found in Chapter 6, Section 6.7.

## 4.3 Occultation methods

From space, observing the direct solar radiation is performed in occultation experiments. The first observations of aerosols by solar occultation were performed by the Stratospheric Aerosol Measurement (SAM) instrument flown on the Apollo-Soyuz mission launched in 1975 by NASA (Pepin and McCormick, 1976). Though the mission lasted only few days it proved the capability of this technique of observation for stratospheric studies and opened the way to other missions. History of space-borne instruments using the occultation technique and the long-term data record of stratospheric aerosols obtained by this technique is presented in Chapter 7.

As the satellite orbits the Earth, the solar occultation instrument points towards the sun and measures its irradiance. The instrument observes sunsets as the spacecraft moves from the sunlit toward the dark side of the Earth. Before each sunset starts, the line of sight (LOS) between the instrument and the sun is well above the atmosphere so that the sun's irradiance as measured by the instrument is unattenuated. As the spacecraft moves toward the dark side, the LOS passes through a portion of the atmosphere, and the sun's irradiance is attenuated due to scattering and absorption by atmospheric constituents. A set of measurements for a series of LOS constitutes an event. During sunrise events, the measurement sequence is just the reverse of that during sunset. As the spacecraft moves from the dark towards the sunlit side of Earth, the sun is first viewed through the atmosphere, and then along an unobstructed path when the spacecraft moves toward the sunlit side. During its rise and set the sun is thus observed through the atmospheric layers and each LOS corresponds to a tangent altitude $Z_t$, which is the point on the LOS closest to the Earth's surface (cf. Figure 4.4). The precise determination of the tangent altitude is a difficult step, critical for deriving vertical profiles of all species, gas and spectral aerosol extinctions and is achieved differently for different instruments (for example, Chu et al.,1989, Glaccum et al.,1996). It should be noted that spacecraft sunsets do not correspond systematically to astronomical ones, it depends on spacecraft orbital parameters.

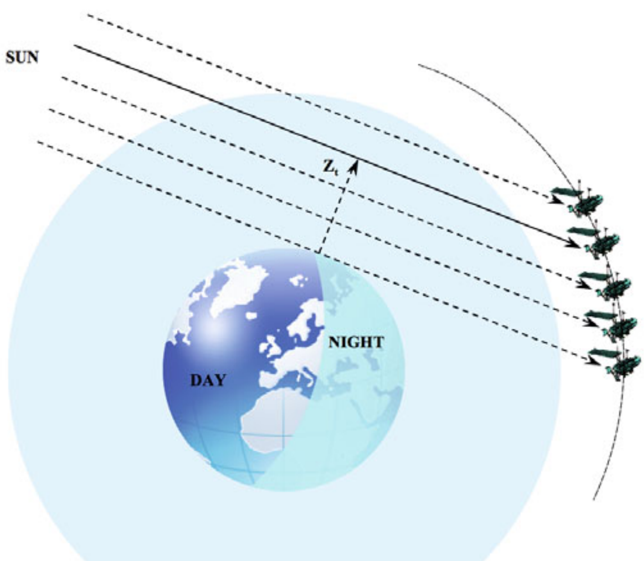



**Figure 4.4** Solar occultation measurement geometry (without refraction). $Z_t$ denotes the tangent altitude.

The sequence of LOS irradiance measurements during an event enables the characterization of the composition of the atmosphere. In the visible and near-infrared, the emission term in the source function of the transfer equation (Eqs 3.1 to 3.4) is negligible compared to the directly transmitted radiance. Moreover, provided that the FOV is small, the contribution of the scattering source term for single and multi-scattered radiations is negligible as well. The slant path transmission along the LOS at a tangent altitude $Z_t$ is obtained by dividing the corresponding irradiance measurement by the unattenuated irradiance measured when the LOS does not intersect the atmosphere, at $Z_t^{\infty}$ tangent altitude

$$T(Z_t) = E(Z_t)/E(Z_t^{\infty}) \tag{4.5}$$

This is analogous to solving for $\tau_\lambda$ in Eq. 4.3 for ground-based sunphotometry. In the occultation technique there is no need for on-board calibration facilities because, due to the unattenuated light measurement, the instrument is recalibrated just after or just before an event so that there is no sensitivity to any drift in the instrument performances. This provides a more accurate calibration than can be attained by ground-based sunphotometers because irradiance at the top of atmosphere is measured directly rather than being extrapolated from a set of measurements in a Langley plot. Therefore this technique is well suited for long-term monitoring of atmospheric species.

The attenuation is caused by the combination of molecular scattering, gaseous absorption and aerosol extinction. Thus, to enable characterization of each contributor, more information is needed than can be provided by a single wavelength, and measurements are generally performed at several wavelengths, as are made for ground-based sunphotometers. In the UV-Visible-NIR wavelength domain the absorbing gases are ozone, nitrogen dioxide, water vapor, oxygen and carbon dioxide. In the far IR several other gases such as methane, nitrous oxide, CO, HCl, contribute also to the absorption.

The total slant path optical depth is derived from the transmission $T(Z_t, \lambda)$ at each tangent altitude $Z_t$ and at wavelength $\lambda$ as

$$\tau_{total}^{SP}\left(Z_t, \lambda\right) = -\ln\left(T\left(Z_t, \lambda\right)\right) \tag{4.6}$$

Vertical profiles of gas concentration and of spectral aerosol extinction coefficients can be retrieved combining these multi-wavelength measurements obtained at all tangent altitudes. Thus two steps are needed in the retrieval processing: a spectral inversion and a spatial inversion (details are given in Chapter 8). Below we summarize some important features of the retrieval.

For the spatial inversion of one event the atmosphere is divided into thin spherical layers and assumption of spherical symmetry is made. This hypothesis is not always fulfilled since it would require horizontal homogeneity over a distance of several hundreds of kilometers. For example, along a LOS at a 10 km tangent altitude, a 1-km thickness layer located at 40 km should be homogeneous over a distance of about 1200 km (Hamill et al., 2006). Heterogeneity along the LOS can lead to significant errors in the retrieved products (Swartz et al., 2006, Berthet et al., 2007).

(a)

(b)

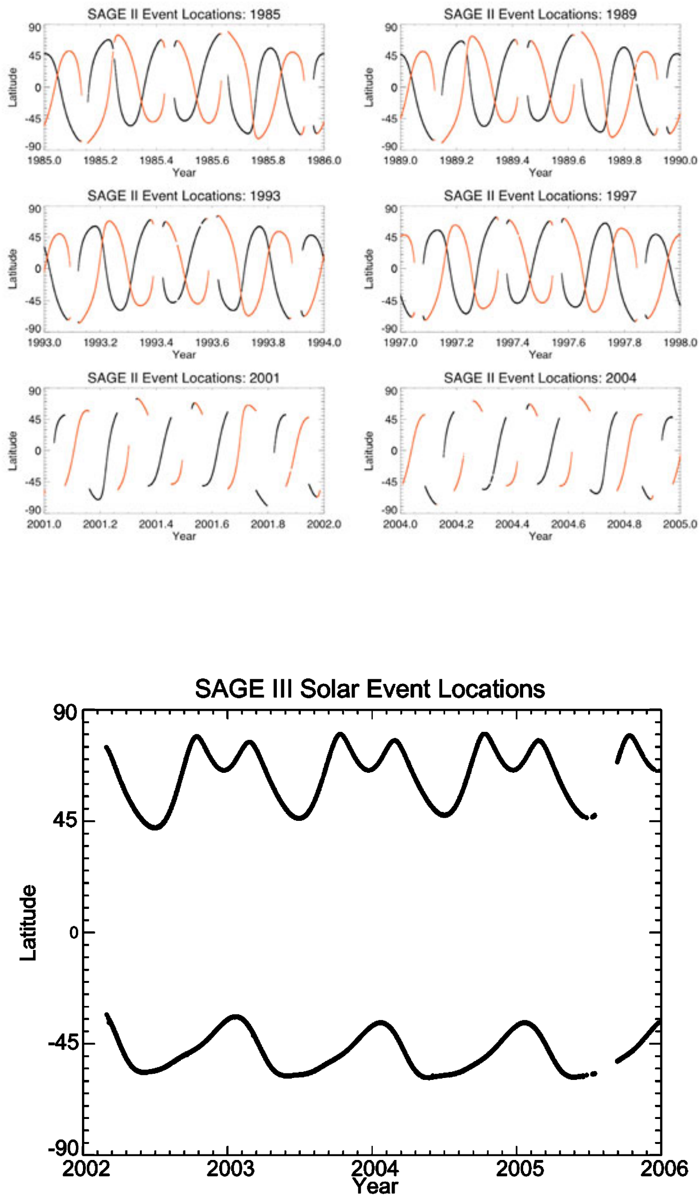



**Figure 4.5** Time-latitude measurement coverage for (a) SAGE II/ERBS for few years between 1984 and 2005. Sunrises, as seen from the spacecraft, are in black; sunsets in red. (b) SAGE III/Meteor-3M for all years between 2002–2005. Sunrises, as seen from the spacecraft, occur in the Southern hemisphere, sunsets in the Northern hemisphere. (Courtesy of L. Thomason, NASA LaRC).

Due to the long path through the atmosphere, the occultation technique is able to detect minor species in the stratosphere or in the troposphere that shorter path lengths, for example from a nadir-viewing satellite, would be lost in the noise. The only caveat is that the optical depth cannot be large enough to overwhelm the signal. Indeed, for retrieval of any species using occultation methods the optical thickness cannot be so large as to obscure the source (the sun). In general this is the case outside strong absorption bands and outside the wavelength domain where the molecular scattering is large, i.e. in the UV range. Of course, the presence of clouds limits the altitude coverage.

An advantage of the space occultation technique compared with ground-based sunphotometry, is that it enables retrieval of information on the vertical distribution of aerosols while ground-based measurements only give access to an integrated column. Compared with airborne sunphotometry, which is limited to altitudes lower than about 10 km, space-based occultation provides information on higher altitudes.

The same method can also be applied to sources other than the sun, such as the moon (for example, with the SAGE III lunar mode) or another star (for example, with Global Ozone Monitoring by Occultation of Stars (GOMOS)).

Space occultation measurements provide better vertical resolution (though it depends on the field-of-view) than passive space nadir measurements, but the geographical and temporal coverage at any location are worse, depending on the spacecraft orbit. Typically, the spacecraft orbits the Earth in approximately 90 minutes or 16 times per day, depending on orbital parameters. Since each orbit provides two measurement opportunities (one sunrise and one sunset), the instrument can acquire 32 separate solar occultation measurements during a 24-hour period at different geographical locations over the Earth depending on the spacecraft orbit. The geographical and temporal coverage of the measurements depend on the satellite orbital parameters. For example, Figure 4.5 shows the time-latitude measurement coverage of SAGE II (inclined orbit) and SAGE III (polar orbit). SAGE II was able to observe sunrises and sunsets in both hemispheres, between 90°S and 90°N, whereas SAGE III observed sunsets in the Northern hemisphere between about 45°N and 80°N, and sunrises in the Southern hemisphere between about 30°S and 60°S. For SAGE II, one can notice the slight drift over the duration of the experiment. During the last years of SAGE II, coverage was not full due to instrumental problems. Similarly, SAGE III encountered a failure in mid-2005.

The number of measurement opportunities and the geographical coverage can be increased if measurements are made using both lunar and solar occultation events, or if stellar occultation events are used.

A few instruments have operated from balloon-borne platforms rather than from space-borne ones (Renard et al., 1996, 2000), and a few others have pointed at the Earth's limb rather than at the sun (SAGE III, Optical Spectograph and InfraRed Imaging System – OSIRIS, on the Odin satellite). In this latter case, the aerosol characterization is much more difficult to achieve but the geographical coverage is extended. More details on the retrievals are available for SAGE III in Rault and Loughman (2007) and for OSIRIS in Boussara et al. (2007).

## 4.4 Conclusion

The direct observation of the sun with ground-based sunphotometers allows retrieval of the total atmospheric optical depth, at one or several wavelengths. Although very simple in principle, the method needs great care in the instrument calibration and utilization for obtaining accurate results. Subtracting from the total optical depth the other components (molecular scattering and gas absorption) leads to the aerosol optical depth. The molecular scattering contribution estimate is provided by ancillary measurements of pressure and temperature. Concerning gas absorption, a correction is often needed even if measurements generally avoid strong gas absorption bands. Spectroscopic data banks are thus necessary, as well as gas concentration data coming, for example from climatology, as is the case to correct for the ozone contribution. These aerosol optical depth data enable the estimation of the columnar aerosol content. An estimate of the aerosol size representative of the column atmosphere can be achieved via spectral variations of the aerosol optical depth (see Chapter 5), the simultaneity of the spectral observations is thus critical. Some instruments employ a filter wheel to measure different wavelengths introducing a time delay between wavelengths. Some others perform measurements simultaneously in several channels, for example, Microtops instruments (Ichoku et al., 2002a). Nevertheless, the time delay is often short (a few seconds for Cimel-AERONET instruments), therefore the precision is similar for both techniques. For additonal aerosol characterization, such as detailed size distribution and information on real and imaginary refractive index, additional measurements are required. Ground-based radiometers linked to sunphotometry that measure sky radiance distribution and/or polarization can produce these additional data, and are described in Chapter 6.

Solar occultation measurements with space-borne instruments also use the extinction of the direct solar beam through the atmosphere during sunset or sunrise, and allow the retrieval of the spectral slant optical depth. Generally, one aims at retrieving not only aerosol, but also a few important gas components, such as ozone. Species separation leads to spectral aerosol slant optical depths and is followed by retrieval of vertical profiles of spectral aerosol extinction coefficients. Species separation and spatial inversion are explained in Chapter 8.