SCIENCE OF SUN PHOTOMETRY

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Abstract
Typically, the total amount of gases and particles in a column of atmosphere cannot be determined from measurements just at Earth's surface, by a single measurement essentially at the bottom of the atmosphere column. Balloons, airplanes, and rockets are all used to perform direct measurements in the atmosphere at altitudes up to and beyond the stratosphere. Satellite-based instruments provide global views, but it is difficult to infer surface and column distributions from space-based measurements, so such measurements must still be supplemented by ground-based measurements. Sun photometry is an important way of probing the atmosphere from the ground to measure the effects of the atmosphere on Sun radiation crossing through the atmosphere to Earth's surface. These indirect technique provide information about the entire atmosphere above the observer, not just the atmosphere that can be sampled directly close to Earth's surface.

Keywords: insolation at Earth's surface, ultraviolet radiation, photosynthetically active radiation, water vapor, aerosols, Sun photometer.

1. INSOLATOIN AT EARTH'SURFACE

Insolation at Earth's surface is a fundamental measurement for understanding Earth and its atmosphere as an interconnected dynamic system. There are also obvious engineering applications of such a measurement, for designing and siting solar power arrays. From a science perspective, long-term solar monitoring helps scientists understand persistent changes in the atmosphere. Figure 1 shows modeled 4-day mean insolation over North America in late December, 2003, based on analysis of GOES visible images (Diak, Bland, and Mecikalski, 1996).
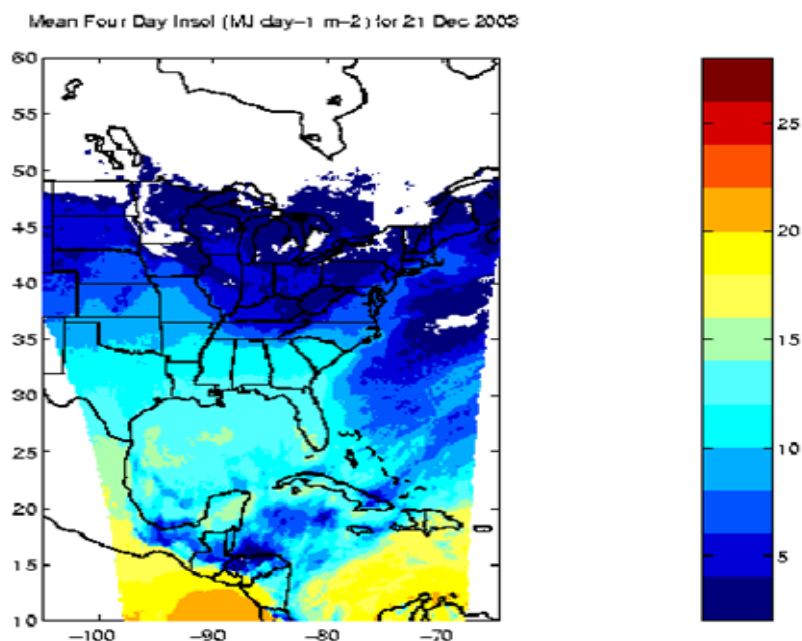
*Figure 1. 4-day mean insolation over North America, late December, 2003,based on GOES visible images.
(Diak, Bland and Mecikalski, 1996)*

2. ULTRAVIOLET RADIATION

Ultraviolet solar radiation (UV solar radiation) has wavelengths between 100 and 400. Because of its intimate connection with the concentration of stratospheric ozone, UV radiation is of great scientific interest. Although UV radiation is of special concern because overexposure is a serious public health problem. Also it is important to understand that the interactions of UV radiation with life on Earth are complex. Life on Earth has evolved within a particular radiation environment that includes some UV radiation. Disruptions to this radiation environment, causing some forms of radiation to increase or decrease, can have serious and unforeseen consequences: this is the source of concern about "ozone holes" in the stratosphere (Liou, 2002).

3. PHOTOSYNTHETICALLY ACTIVE RADIATION

Photosynthetically active radiation (PAR) refers, as its name implies, to that subset of total solar radiation that plants use for photosynthesis. Typically, this is a full-sky measurement. The spectral definition of PAR is not precise, because different kinds of plants respond to different parts of the solar spectrum. However, PAR is generally considered to include just the visible part of the solar spectrum between 400 and 700 nm. Thus, a PAR detector is similar to a pyranometer, but with its spectral response limited to just visible wavelengths.

The output from a full-sky PAR detector is somewhat correlated with total solar radiation, but the correlation is imperfect enough so that scientists require a separate measurement for PAR radiation. This is extremely important in agriculture and botany, for studying plant behavior under different lighting conditions - under crop or tree canopies, for example.

Whereas insolation is reported in units of energy density (W/m^2), PAR radiation is reported in terms of the total number of photons from the visible range, corresponding to wavelengths interval between 400 and 700 nm. This makes sense because photosynthesis involves the interaction between individual photons and molecules. Specifically, PAR is reported as the number of moles of photons per unit area, per second, in the visible spectral range, where there are 6.023×10^{23} (Avogadro's number) photons per mole. Figure 2 shows PAR from the USDA UV-B Monitoring Network at Beltsville, Maryland, on 1 September, 2006.
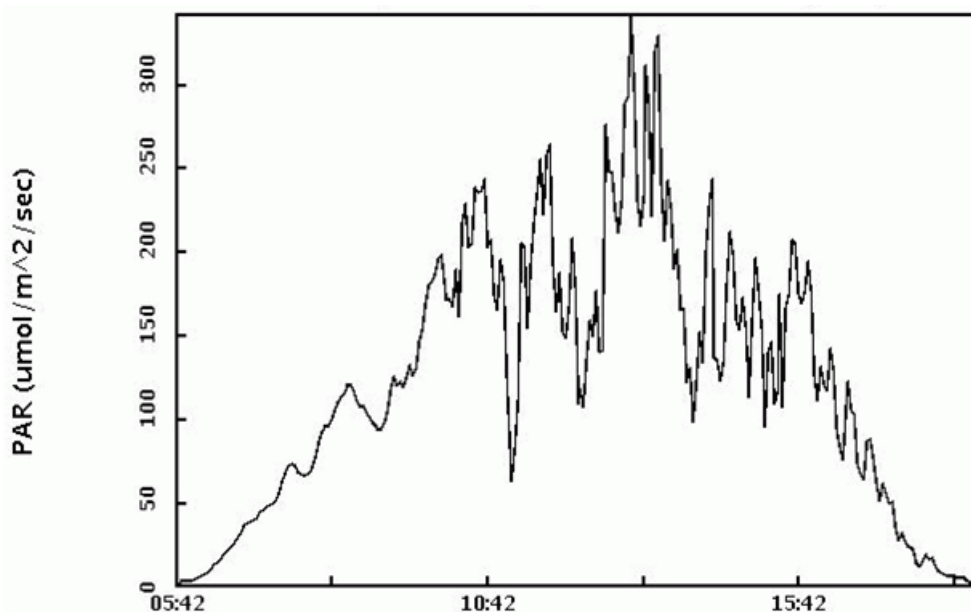

Figure 2. Photosynthetically active radiation (PAR) at Beltsille, Maryland, 1 September, 2006.
(Mims, 2006)

4. WATER VAPOR

Total atmospheric water vapor, also called *total column water vapor* or *total precipitable water vapor* (PW), is defined as the thickness of a layer of water obtained by condensing all the water vapor in a column above the observer and bringing it down to the observer's elevation. Typically, PW is a few centimeters (Reitan, 1963).

PW is distributed unevenly around the globe. Figure 3 shows a view of the Western hemisphere on 22 December, 2003, based on infrared (6750 nm) data from the GOES-12 satellite. In general, there is more water vapor over warm water and equatorial forests because of evaporation and transpiration. Not surprisingly, there is much less water vapor over deserts and at high elevations. Although it is commonly believed that there must be a lot of precipitation at the poles, because they are covered with snow and ice, the air over Earth's polar regions is usually very dry.
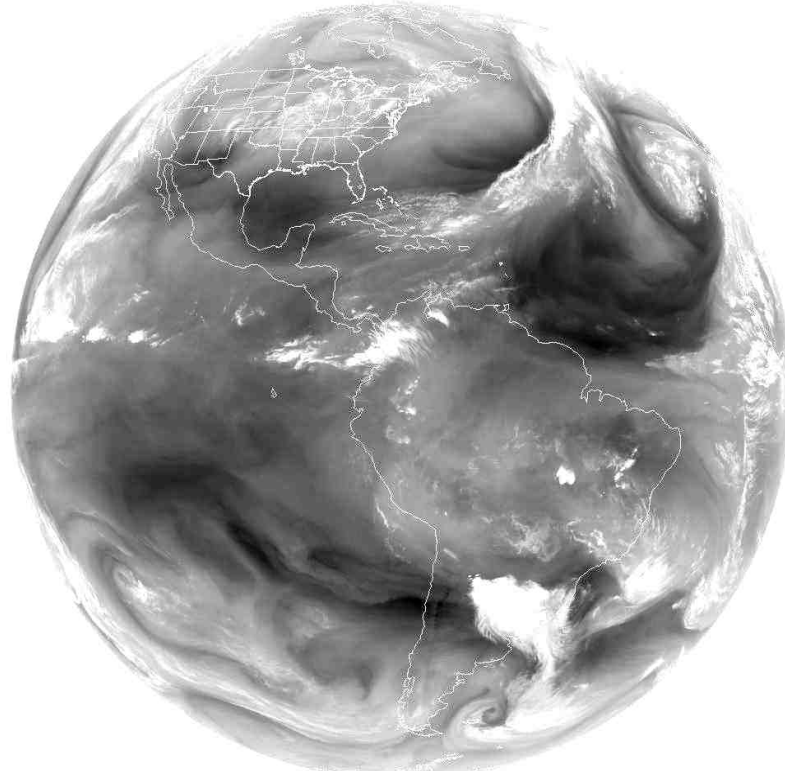
Figure 3. GOES-12 6750 nm water vapor image for the Western hemisphere, 11:45 UT, 22 December, 2003.
(Diak, Bland and Mecikalski, 1996)

5. ATMOSPHERIC AEROSOLS

Let us consider the sunlight lights up a horizontal surface situated at distance x from the Sun. If sunlight has intensity I_0 when the Sun is directly above a horizontal surface, meaning the zenith angle is 0°, then the intensity I at some other zenith angle v is given by the following formula:

$$I(x) = I_0 \cos v \qquad (1)$$

As noted above, molecules and liquid or solid particles suspended in the atmosphere (aerosols) scatter and absorb sunlight. The effect of molecular scattering on direct sunlight can be calculated theoretically, as a function of wavelength, as can absorption by gases. The remaining reduction at a particular wavelength is due to scattering and absorption by aerosols in the atmosphere. Reductions in transmission are described by a quantity known as optical thickness, or optical depth. The more scattering and absorption reduce transmitted Sunlight as viewed along a direct path, the larger the

optical thickness. The basic equation governing the transmission T of radiation through an intervening medium is given by the Lambert – Bouguer low, also knew as the Lambert – Beer low:

$$T = \frac{I_\lambda}{I_{0\lambda}} = \exp(-\tau_\lambda) \quad (2)$$

where $I_{o\lambda}$ is the original source intensity, I_λ is the intensity after radiation passes through a medium of thickness m, and τ_λ is the total atmospheric optical thickness, all at wavelength λ (Petty, 2006). The total atmospheric optical depth τ_λ consists of three basic parts: atmospheric absorption by gases such as ozone, molecular (Rayleigh) scattering by the atmosphere and scattering by aerosols (aerosol optical thickness or AOT):

$$\tau_\lambda = \tau_{abs\lambda} + \tau_{R\lambda} + \tau_{scat\lambda} \quad (3)$$

The strongly wavelength-dependent contribution of Rayleigh scattering can be calculated using theoretical models of the atmosphere, such as the following parameterized model (Bucholtz, 1995):

$$\tau_\lambda = \left(\frac{p}{p_0}\right) \cdot A \cdot \lambda \cdot \left(-B - C \cdot \lambda - \frac{D}{\lambda}\right) \quad (4)$$

where (p/p_o) is the ratio of actual barometric pressure at an observing site to standard pressure at sea level (1013.25 mbar), λ is the wavelength in units of micrometers and the coefficients A, B, C, D given in Table 1 are empirically derived values that provide a mathematical best fit to theoretical calculations:

Table 1: Coefficients for Rayleigh optical thicknes.
(Bucholtz, 1995)

Coefficient	$\lambda \leq 0.500$ μ (500 nm)	$\lambda > 0.500$ μ (500 nm)
A	6.50362×10^{-3}	8.64627×10^{-3}
B	3.55212	3.99668
C	1.35579	1.10298×10^{-2}
D	0.11563	2.71393×10^{-2}

Sources of aerosols include volcanic activity, dust from deserts and agricultural activity, sea spray, and air pollution. The amount of aerosols varies widely around the globe, and there are strong seasonal effects. In clean skies, AOT at visible wavelengths will be less than 0.1. Figure 4 shows an 8-day mean of global aerosol optical thickness from early November, 2003, as calculated from data recorded by the Moderate Resolution Imaging Spectroradiometer (MODIS) instrument on NASA's EOS/Terra and Aqua spacecraft. Cloudy weather and/or snow cover over the northern hemisphere prevent AOT calculations from being done. Bright land surfaces, such as the Sahara Desert and the Saudi Arabian Peninsula, and permanently snow/ice-covered polar regions, are inaccessible to MODIS aerosol retrievals. High values of AOT resulting from biomass burning activity are clearly evident in sub-Saharan Africa and the Amazon Basin in South America. Even though aerosols directly over the Sahara are not visible to MODIS, a plume of Saharan dust is visible over the Atlantic Ocean, blowing westward. It is common for Saharan dust to find its way to the Western hemisphere. More recently, it has also become clear that large amounts of dust and other pollution

are often blown eastward from Asia to the west coast of the United States.
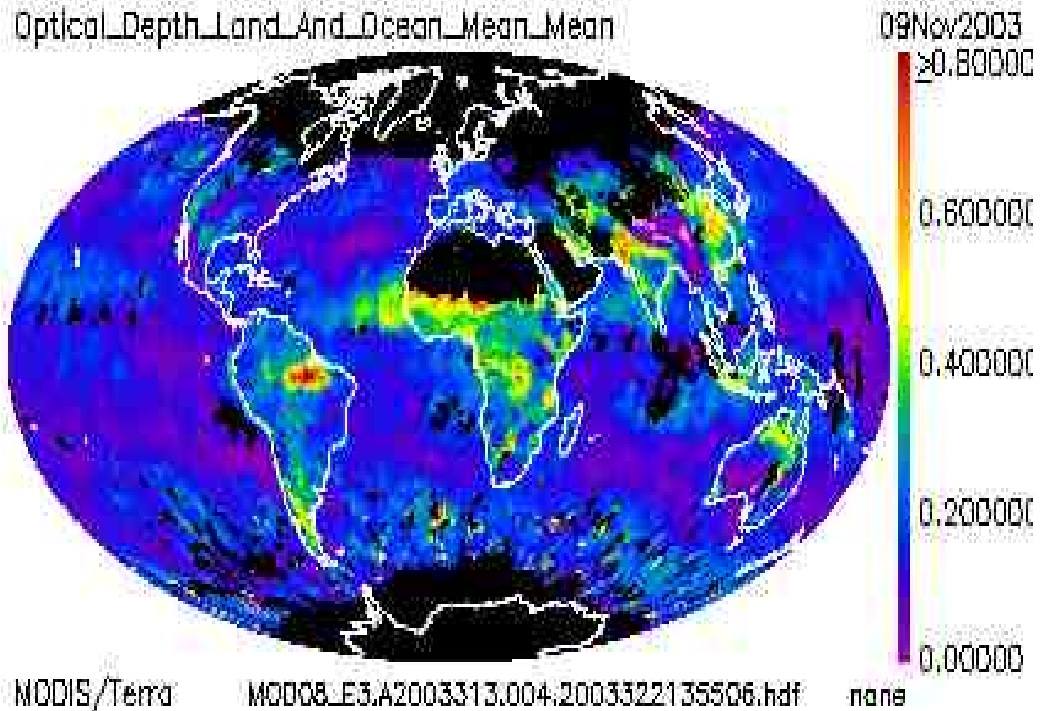

Figure 4. 8-day mean aerosol optical thickness from MODIS/Terra, early November, 2003.
(Brooks et. al., 2003)

The predominant gas absorber over the wavelength range of interest is ozone. Figure 5 shows the mean optical thickness for ozone across visible and near IR wavelengths. It is small (<0.04), but can be significant for clean air with small values of aerosol optical thickness such in the case of polar atmosphere.
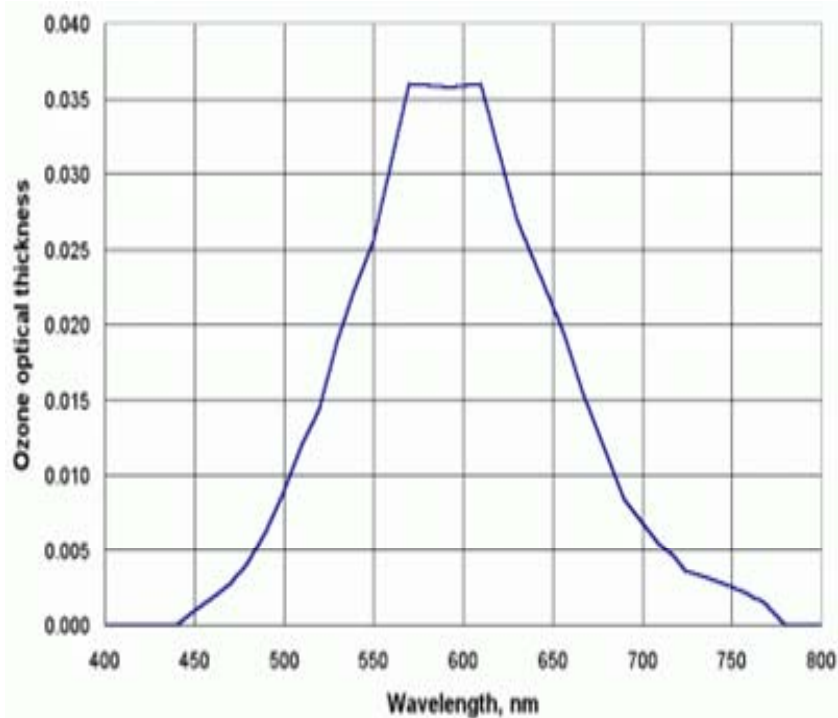

Figure 5. Ozone optical thickness at visible and near-IR wavelengths.
*(Bird and Riordan, 1994)*

6. SUN PHOTOMETRY

A Sun photometer for measuring the physical characteristics of atmospheric aerosols is a straightforward application of the Lambert - Bouguer law (Lambert – Beer low). Simple handheld Sun photometers based on interference filters were first described by Frederick Voltz [1974]. Sun photometers using LEDs as detectors were first described by Mims [1992]. In the basic circuit, shown schematically in Figure 6, an LED detector is incorporated into a transimpedance amplifier configuration that converts a small current from the detector to a voltage and amplifies it. Typically, the amplifier gain is adjusted to give an output on the order of 1-2 V in full sunlight.
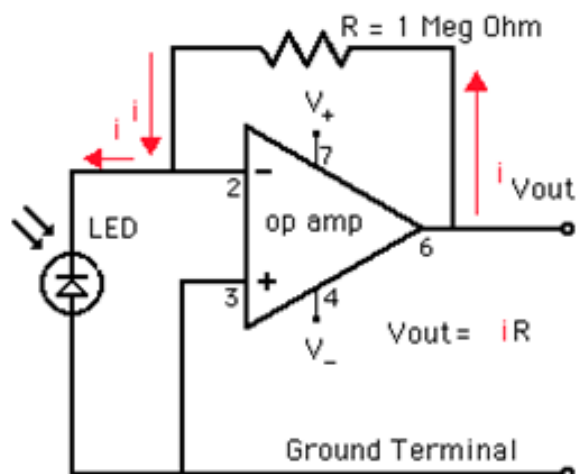
Figure 6. Sun photometer circuit. From OpAmp theory terminals 2 and 3 are at ground potential and no current flows through these pins. The current I produced by the LED when light shines on it is simply related by Ohm's law as $V_{out} = I*R$, so 1 microampere current produces 1 Volt voltage when a voltmeter is placed between V_{out} and the Ground Terminal. (Mims, 1992)

The Sun photometer originally developed for GLOBE has two channels in the visible part of the spectrum. The normalized response of these detectors is shown in Figure 7. This figure illustrates one of the problems noted earlier - that the spectral response of LED detectors is often wider than desirable for Sun photometry. The response of these LEDs extends over several tens of nanometers. Brooks and Mims [2001] have discussed how to interpret measurements made with detectors such as those shown in Figure 7. Basically, an "effective" wavelength response is calculated by summing over the response wavelengths for each detector.
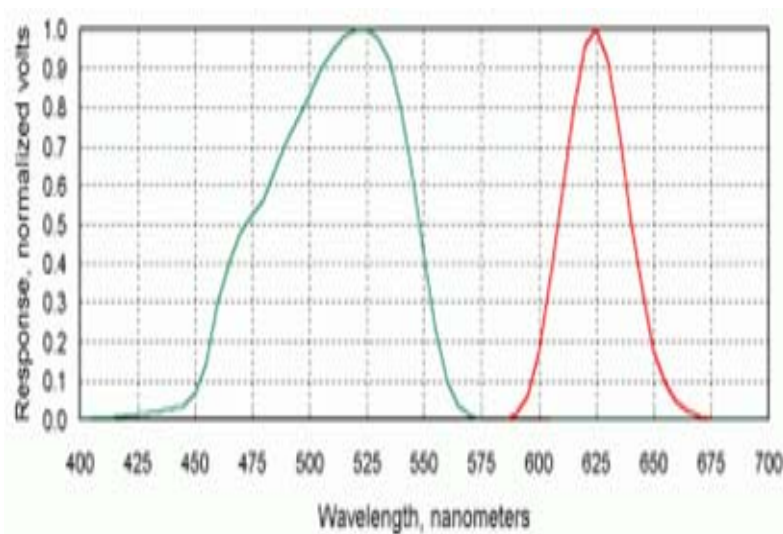
Figure 7. Normalized spectral response of green and red LED detectors used in the two-channel GLOBE Sun photometer. (Brooks and Mims, 2001)

The output voltage from the transimpedance amplifier is proportional to the amount of Sunlight reaching the detector:

$$V = V_0 \left(\frac{R_0}{R} \right)^2 * \exp\left\{ m\left[\tau_{abs} + \tau_R \left(\frac{p}{p_0} \right) \right] \right\} \quad (5)$$

Solving for aerosol optical thickness τ_{abs}, one obtains:

$$\tau_{abs} = \frac{1}{m} * \left\{ \ln\left[V_0 \left(\frac{R_0}{R} \right)^2 \right] - \ln(V) \right\} - \tau_R \left(\frac{p}{p_0} \right) \quad (6),$$
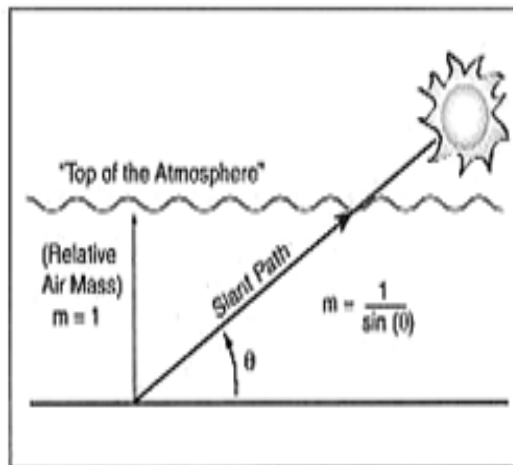
where:
R is the Earth - Sun distance at the time of a measurement,
R_o is the average Earth - Sun distance (1 astronomical unit),
p/p_o is the ratio of actual barometric pressure (station pressure) at the observing site to standard sea level barometric pressure (1013.25 mbar),
m is the airmass factor, definite by the formula $m = \dfrac{1}{\sin \theta}$ (7), explained in the Figure 8:

Figure 8. Airmass factor m corresponding to an elevating angle θ of the Sun over the horizon (Young, 1994)

That is, the equation separates the total atmospheric optical thickness into two components:
1) *molecular Rayleigh scattering*, which is a quantity that depends only on barometric pressure at the observing site: the calculated Rayleigh coefficient τ_R is multiplied by the ratio p/p_o (Bucholtz Anthony, 1995).
2) *molecular absorption and Mie scattering by aerosols*, which is a quantity that depends only on the mass factor *m*, which varies with the moment of the day when the measurement take place.
The contributions due to ozone (and perhaps other absorbing gases under some circumstances) and aerosols can be separated after the fact, either by using climatologically and latitude - dependent average ozone values, for example, or by using actual total column measurements for the time and place of the data collection. Satellite-based instruments such as the Total Ozone Mapping Spectrometer (TOMS) are sources of such data. For the GLOBE two-channel Sun photometer, a typical ozone contribution to the non-Rayleigh optical thickness is about 0.01 for the green channel and 0.03 for the red channel (Brooks and Mims, 2001), (Brooks, Forrest, Levine, Hinton, 2003).

The quantity V_o is the calibration constant for the instrument. It is the voltage output the instrument would produce if there were no atmosphere between the detector and the Sun. Obviouslly, it is not practical actually to measure the voltage outside the atmosphere. However, it is possible to infer what this voltage would be. If a sun photometer views the Sun through various values of relative air mass, and the total atmospheric optical thickness does not change, then the logarithm of the instrument voltage is proportional to relative air mass. By fitting a straight line through the data (a linear regression), the intersection of that line at the y-axis where m = 0 is the logarithm of the voltage the instrument would produce if there were no atmosphere. This is called a *Langley plot calibration*, named after Samuel P. Langley, the scientist who developed this technique in the early 20th century. (Abbott and Fowle, 1908).

It is not easy to do these calibrations, because it is not easy to find observing sites at which total atmospheric optical thickness remains constant for several hours, as is required to collect data over a range of relative air mass values. High elevation sites such as Mauna Loa Observatory in Hawaii are favored for such work because of the clean skies and stable meteorology at this site. Figure 8 shows a Langley plot calibration performed there on a GLOBE Sun photometer on June 21, 2000.
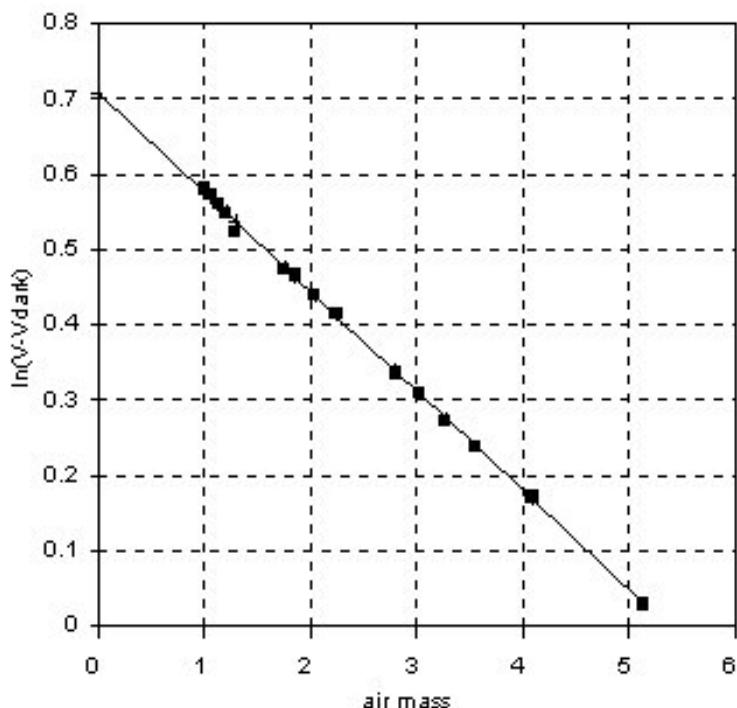
Figure 9. Langley plot Sun photometer calibration at Mauna Loa Observatory in Hawaii on June 21, 2000 (Brooks et. al. , 2003)

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