

**IMPROVED PROCESSING AND DEVELOPMENT OF THE MULTI-FILTER
ROTATING SHADOW-BAND RADIOMETER (MFRSR) NETWORK**

By

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ABSTRACT

Improved Processing and Development of the Multi-Filter Rotating Shadow-Band Radiometer (MFRSR) Network

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Information about global distributions of aerosol optical thickness (AOT) and size categorization is necessary to quantify the aerosol radiative forcing as well as a method to monitor air-quality such as fine particulate matter (PM). Efforts to provide such aerosol optical properties on a global scale clearly require satellite retrievals and therefore validation of satellite products is a clear priority. The development of suitable radiometer networks is central to this effort. Modern robotic solar instruments for retrievals of aerosols optical depth and other microphysical parameter retrievals, such as the CIMEL Sky Scanning Radiometer [Holben, 1998] as part of the Aeronet Network, are crucial to this effort but are unfortunately quite sparse and expensive and efforts to develop and utilize simpler portable instruments are highly desirable. One attractive possibility is the deployment of Multi-Filter Rotating Shadow-band Radiometer (MFRSR) in a network [Alexandrov, 2002] both for validation efforts as well as monitoring extended megacities such as the NYC area.

Furthermore, the developments of these portable networks in urban areas are particularly crucial for the following reasons:

1. The retrieval of aerosol properties over land areas and particularly urban surfaces is far more challenging than over oceans due to the bright and complex surfaces in urban areas. Cutting Edge algorithms such as the aerosol retrieval from the Advanced Polari-metric Scanner (APS) [Waquet, 2009] from current and future platforms are in need for validation and clearly would benefit from instruments that can be moved to strategic locations under the satellite track. In addition, for climate studies, separation of aerosols into fine / coarse mode constituents [Mischenko, 2004] as well as identifying and quantifying absorbing aerosols is critical. These outputs are in fact a major focus of APS (or combined APS-MODIS) and pulling these parameters out from the MFRSR processing is critical in maximizing the value.
2. In addition, urban areas have more spatial diversity in aerosols and would greatly benefit from closely spaced measurements that can probe local structure and anomalies.

In this thesis, our goal is to test and implement a newly developed retrieval algorithm developed at NASA GISS [Alexandrov et al 2006] for processing MFRSR data. In particular, we show that this algorithm significantly improves optical depth time series measurements in comparison to the most currently used Langley regression method calibration. The main reason for the improved performance of the NASA-GISS algorithm is that the 870nm calibration is based on the ratio between the direct and diffuse radiance in which only the color ratios during the calibration procedure are required to be stable. We explore the performance of the NASA-GISS algorithm

and show that the algorithm results in more stable calibration coefficients in comparison to those obtained using the conventional Langley regression. Furthermore, statistical analysis shows the stability of the aerosol color ratio as measured by the angstrom coefficient has RMSE variations three times smaller, which allows better calibrating statistics over longer periods. Further, we explore the accuracy of the MFRSR retrieval against a far more expensive coincident CIMEL and show that the MFRSR accurately retrieves both the total and the fine mode AOD but underestimates the coarse mode. This is reasonable since the MFRSR filter wavelengths only go to 870nm as compared to the 1620 nm available in the CIMEL instrument. We report our deployment of the MFRSR network over the NYC metropolitan area and describe intercomparison measurements between AOD measurements both inside and outside NYC in an effort to explore local aerosol production. Finally, the MFRSR network shall provide us with the spatial and temporal resolution needed to validate satellite data.

The thesis contents are as follows. In section 1, motivation of the importance of aerosols for both climate impacts and human health are briefly given as well as a primer on what optical quantities we are measuring to detect aerosols. In section 2, we briefly discuss the different ways to measure aerosols including satellite and ground remote sensing (and in-situ) instruments with a particular focus on establishing networks that can validate complex aerosol satellite algorithms. In section 3, we present our validation of the MFRSR algorithm of NASA GISS [Alexandrov et al 2002, 2005, 2007, 2008] against CIMEL measurements to ensure

that we are handling the processing stream properly. In section 4, we present our current and future network topology, processing, networking etc as well as present preliminary multisensory matchups for total AOD, fine and coarse mode separation and water vapor with particular focus on correlations and discrepancies. In section 5, we illustrate the value of the network through satellite validations for both MODIS and GOES aerosol retrievals. In section 6, we present our efforts to date in trying to pull out single scattering albedo and future prospects in moving the network deployment and algorithms forward.

DEDICATIONS

This dissertation is dedicated to my beloved wife Jenny, my daughter Katalina and my son Nicholas for believe in their husband and father crazy ideas and for their understanding and support for these past six years.

Special thanks to the Almighty who knew me in time of desperation and when patience was not to be found.

For my parents to whom I owe coming to this country in search of the American dream. Also thanks to my brothers and sisters for the kindness and support given at the right time.

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1 WHY AEROSOLS?

1.1 Importance to Climate

Atmospheric aerosols are organic and inorganic matter composed of fine and coarse particles that are suspended in the atmosphere and play an important role in the Earth's radiative budget [Charlson *et al.*, 1992]. The total incoming, Figure 1.1, solar radiation is about 342 W m^{-2} of which 31.3% of solar radiation is sent back to space by clouds and surface reflectance. The earth's atmosphere absorbs about 19.6% of the total incoming radiation and the earth surface absorbs about 49.1% of total solar flux.

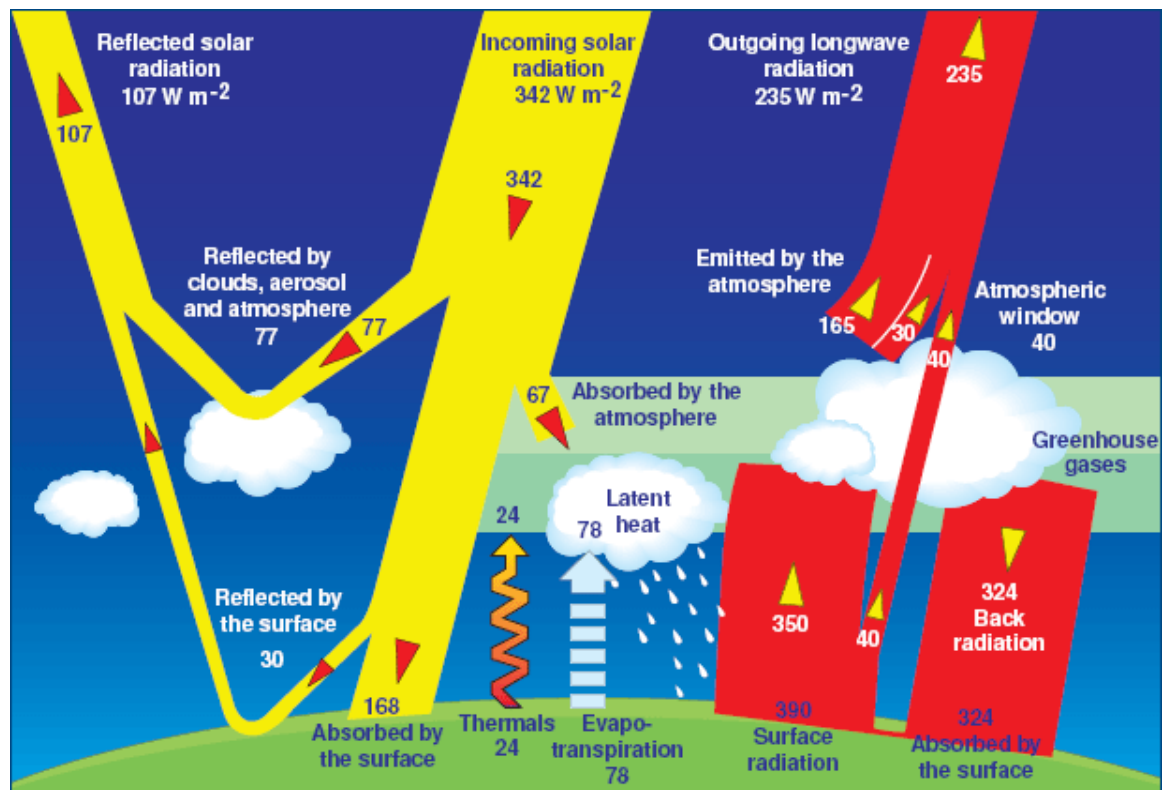


Figure 1.1 globally averaged energy budget of the atmosphere. Figure from <http://asd-www.larc.nasa.gov/ceres/brochure/> based on data from Kiehl and Trenberth 1997).

The energy absorbed by the earth's surface and atmosphere is balanced out by converting incoming solar flux into long-wave radiation and it is sent back to space as infrared radiation. The earth surface emits about 390 Wm^{-2} of which about 10.3% is directly sent into space without being extinguished by particles or the atmosphere. The atmosphere reradiate back to the earth surface about 83.1% long-wave radiation. The atmosphere and the Green house gases scatters about 67.1% infrared radiation back to space, and about 29.1% is a contribution to the atmosphere, latent heat by clouds and thermal radiation emitted by the earth surface. Aerosols also have an effect on the global climate through radiative forcing and chemical perturbations. One of the biggest natural variations of aerosols is especially due to episodic eruptions of large volcanoes which send organic matter and gases including sulfuric dioxide into the stratosphere. These perturbations are recognized as a significant climate forcing which is a factor that alters planetary radiation balance and thus tends to cause a global temperature change. Since natural aerosols particles range in radius, it is imperative for the classification of total aerosols into fine and coarse particles for the calculation of flux extinction.

1.1.1 Direct and Indirect Effect

Aerosols affect the earth radiative balance in two ways. The first way is the direct effect which involves the absorption and scattering of aerosols. In general, for scattering aerosols such as natural sea salts and clay, the sun light that is reflected back to space has a cooling effect on the climate. On the other hand, aerosols are increasingly be generated by man as a result of industrialization. These aerosols are often referred to as biomass

burning aerosols and are the residue of fuel burning processes. Unlike their natural cousin, these aerosols can also absorb radiation converting this solar energy to Infrared radiation (i.e. heat) thereby having a warming influence [www.epa.gov]. For climate applications, it is clear why we want to differentiate absorbing from scattering aerosols since their effect on climate is so different. Unfortunately, separating this from satellite measurements is very difficult and results in significant uncertainty in predicting global warming trends.

The second (indirect) mechanism is even more subtle and hard to quantify. In particular, aerosols can modify climate by modifying cloud properties and precipitation. For example, Clouds microphysics nuclei start with aerosols particles, especially clay, and grow by condensing water vapor creating many tiny droplets. High aerosol concentrations lead to an increase in small cloud droplets which reflect light more efficiency thereby increasing planetary albedo i.e. the earth have a cooling effect

Another mechanism is that the more aerosol particles there are, the more likely that the droplets remain small and these little droplets absorb most of the available water vapor. By having smaller particles attracting the water vapor, there is a tendency to reduce precipitation

1.1.2 Air Quality

Air pollution is a widespread problem in the United States, with nearly 100 million individuals exposed to levels of air pollutants that exceed one or more health-based ambient standards. The number of days not meeting these standards is particularly high in urbanized regions such as the north-east region, especially New York City metropolitan area. One of the most important air pollutants experienced in urban setting is aerosol particulate matter (PM). Studies [Wilson W.E et al 1997, Brauer M et al 2002, Jansen Nah et al 2003, Kramer U et al 2000] on a growing list of epidemiological diseases over metropolitan areas with excessive exposure to fine PM sources such as car and factory emissions, show notable links to a range of serious respiratory and cardiovascular health problems. Recent studies have shown that adverse health effects continue to be linked with exposure to particles at levels previously considered “safe” [www.epa.gov] which drive the need for stricter emission standards and better measurement of current aerosol levels.

Classification of aerosols is essential to quantify the problem and assess which aerosol features are most closely connected to health concerns. In particular, aerosols can be categorized by their size (fine or coarse) and their chemical composition (i.e. carbon – sulfates etc). Unfortunately, for prediction purposes, the concentration of aerosols varies dramatically as a function of time and location and is a complicated function of the available PM emissions sources, atmospheric physical and chemical interactions, meteorological conditions, and other variables. Fine (PM_{2.5}) and Coarse (PM₁₀) have distinct pathways in regard to their emission sources, formation processes, chemical

composition, atmospheric residence times, and transport distances. PM_{2.5} is the most important for health impacts due to fine particulates' ability to penetrate deep into the lungs. A strong EPA standard of $35\mu\text{g}/\text{m}^3$ [<http://a257.g.akamaitech.net>] has been set as a 24 hour average period.

Under the Clean Air Act, Environmental Protection Agency (EPA) is required to establish a nationally uniform air quality index for the reporting of air quality. In 1976, EPA established this index, and then called the Pollutant Standards Index, for use by state and local communities across the country. The Index provides information on pollutant concentrations for ground-level ozone, particulate matter, carbon monoxide, sulfur dioxide, and nitrogen dioxide. On July 18, 1997, EPA revised the ozone and particulate matter standards, in light of a comprehensive review of new scientific evidence including refined fine particulate matter standards. In particular, significant exposure studies have further reinforced the need to lower PM_{2.5} concentrations and currently, a notable portion of the urban North East is in non attainment of the EPA standards as seen in figure 1.2 with most of the difficulties seen in the NYC metropolitan area.

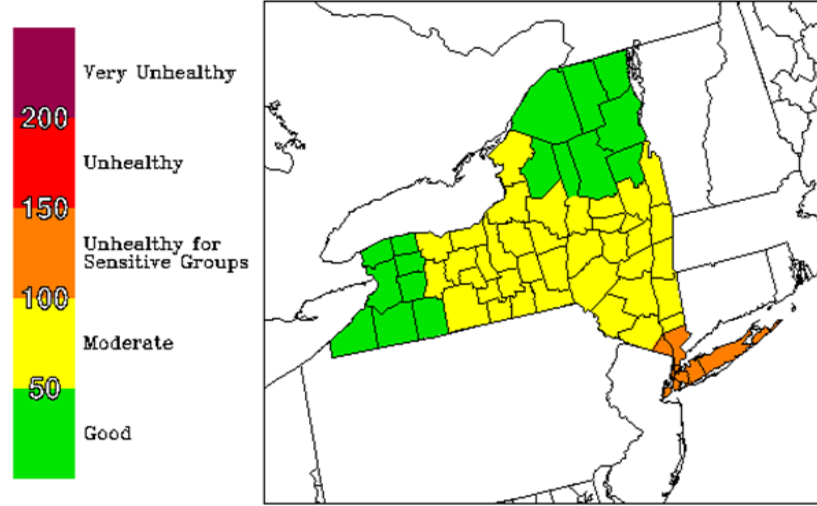


Figure 1.2: EPA PM2.5 Air Quality Map

To assess the aerosol transport problem within the metropolitan area, it is therefore important to monitor aerosol levels in a spatially distributed manner both inside and outside the metropolitan canopy. Unfortunately, direct sampling of aerosols is expensive and time consuming. Therefore, less expensive automated approaches are needed. One possibility is the use of automated sky radiometers which measure the total column (Aerosol Optical Depth) $AOD(\tau)$. Roughly, this is done by measuring the solar transmission through the atmosphere by using the relationship $T = e^{-\tau}$ [Harrison A. et al 1994]. Although it is quite apparent that a total optical depth column is only an approximate estimation for aerosol concentrations at surface level, studies tend to show that in the North East region, column AOD is a remarkably a good proxy for PM2.5. In fact, attempts to integrate satellite data into PM2.5 forecasts generally use the paradigm of the IDEA (Infusing satellite Data into Environmental Applications) product in which column AOD [www.idea.ssec.wisc.edu] is related directly to surface PM2.5 using a static universal relationship where $PM2.5 \cong 60\tau$.

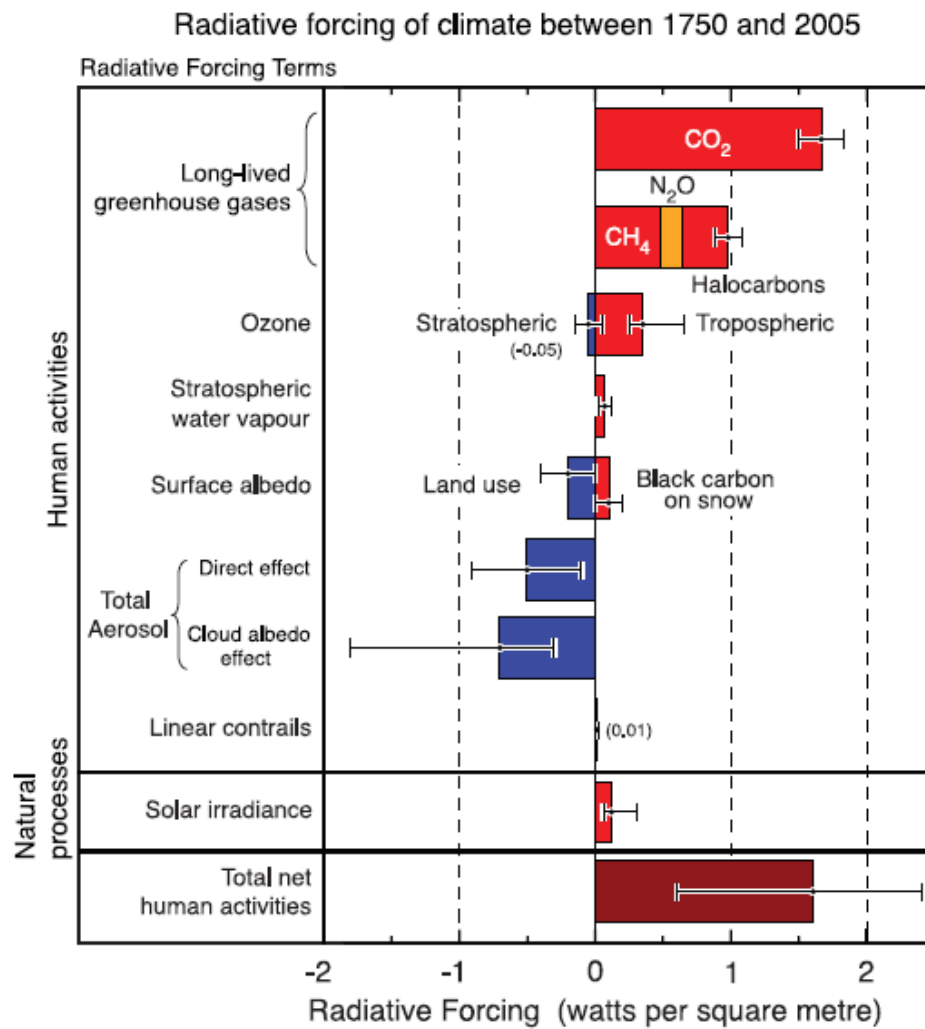


Figure 1.3 Climate Forcing due to atmospheric constituents [IPCC 2007].

1.1.3 Requirements for Climate Studies

To see the importance of aerosols on climate, we see in figure 1.3 the greatest uncertainties in the overall budget are due to aerosol direct and indirect effects. While these forcing are uncertain, they seem to imply that aerosols tend to cool the atmosphere. However, these estimates do not include possible precipitation modification and do not seriously apportion absorbing and non absorbing aerosols since. In fact, these budgets are mainly based on global AOD.

However, Aerosol Optical Depth alone is not sufficient. In fact, the following aerosol retrieval requirements for climate research over land have been suggested by Mishchenko [2004] for aerosol over land: **a)** 0.04 or (10 %) for the aerosol optical thickness; **b)** 0.03 for the single scattering albedo and **c)** 0.1 micron (or 10 %) for the effective radius. These numbers in fact are driving the need for the Advanced Polarimetric Scanner (APS) retrievals on Glory so it is extremely valuable that Ground based retrievals can be performed with these accuracy levels.

1.2 How to Measure Aerosol's?

Aerosols can be detected optically based on the fact that they scatter and absorb light. The more light scattered, the more aerosols there are in general. For ground based instruments, the quantity to measure would be the column path transmission which defines the fraction of Aerosol Optical Thickness (AOT) solar irradiance that makes it to the surface.

To quantify the transmission and relate to the type and number of aerosol particles, we need to first define local aerosol parameters that define the efficiency of scattering for a differential path element.

We first define extinction as the total removal of photons from the direct solar path. With this definition, the total extinction coefficient is the sum of absorption and scattering coefficients:

$$\alpha_{ext} = \alpha_{abs} + \alpha_{sct} \quad 1.1$$

Very important is quantifying the ratio of scattering to extinction. This ratio is called single scattering albedo denoted as scattering over extinction:

$$\tilde{\omega} = \frac{\alpha_{sct}}{\alpha_{ext}} = \frac{\alpha_{sct}}{\alpha_{abs} + \alpha_{sct}} \quad 1.2$$

The albedo ranges from pure absorbing medium ($\tilde{\omega} \approx 0$) to a solely scattering medium ($\tilde{\omega} \approx 1$). This parameter is important in differentiating absorbing aerosols such as Biomass burning (low albedo) with non absorbing aerosols (Sulfates, Sea Salts etc) and is a hard quantity to measure from space. Figure 1.4 is a cartoon that displays the direct solar beam (radiation) which reaches the earth ground through extinction-less path.

However, when the path is composed of gases and suspended particles, light photons may be scattered back into the atmosphere or absorbed by gases then reradiated like a blackbody in different directions see figure 1.1, a 83.1% back to space and 61.1% is send back to the earth surface as infrared radiation. The differential extinction is denoted $\Delta\tau = \alpha_{ext}ds$ and the full path integration is known as AOT. The optical thickness τ_λ , or extinction coefficient, is wavelength dependent and is calculated by the use of Beer-Lambert formula assuming that the extinction coefficient is constant:

$$I_\lambda = I_{0\lambda} \times e^{-\tau_\lambda \times m} \quad 1.3$$

Where I_λ is the measured spectral solar beam intensity, wavelength dependent, that reaches the earth surface without extinction, $I_{0\lambda}$, is the extraterrestrial solar beam intensity, also wavelength dependent, at the top of the atmosphere. The mass loading m is relative to the air mass unit in the zenith angle direction or solar path.

As we can above, given measurements at the ground and knowing the irradiance at the Top of Atmosphere (TOA), estimates of total optical depth can be given for each wavelength of interest. The fact that aerosols of different types have different aerosol optical depth variations with wavelength make is possible to extract some aerosol typing (particles coarse or fine mode etc).

1.2.1 Absorption and Scattering by Particles

The optical depth can also be connected to the number of scattering particles and the type of particle by rewriting the extinction coefficient as $\alpha_{ext} = N\sigma$ where N is the total particle number and σ is the extinction cross-section (per particle). The cross-section can further be connected to the particle type as $\sigma = Q_{ext}\pi a^2$ where a is particle radius and Q_{ext} is the scattering efficiency which is a complex function of particle size and refractive index and is meant to scale the scattering cross-section to that expected by geometric optics (i.e. the physical cross-section).

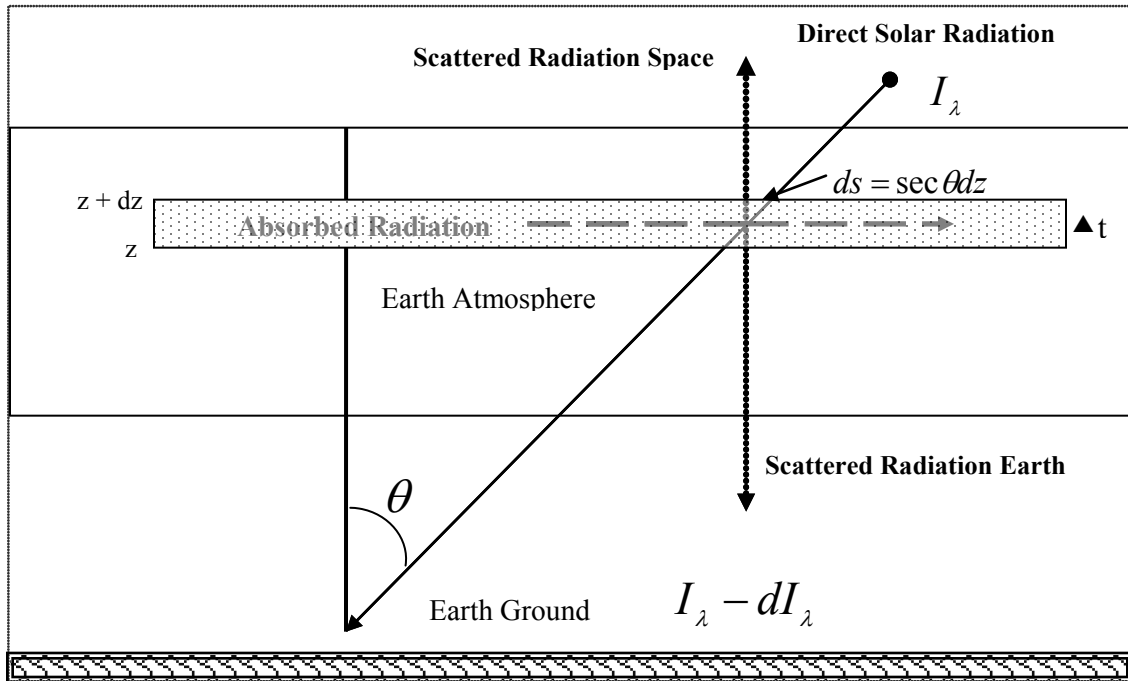


Figure 1.4 Earth Radiative budget direct sun light being scattered by $\Delta \tau$ layer of aerosols, some of the scattered or diffused radiation is send back into space and into the earth ground. A relationship between slant and vertical paths in a parallel plane.

However, aerosols do not come in one size and the optical properties must take this into account. For example, if we describe the size distribution by a sum of log-normal distributions (log normal is best since normal distributions would have a non zero probability of having negative radius which is unphysical)

$$\frac{dN}{d \ln r} = \sum_i \frac{1}{r} \frac{N_i}{w_i \sqrt{2\pi}} \exp\left(-\frac{[\ln(r/r_{g,i})]^2}{2w_i^2}\right) \quad 1.4$$

Where N_i is the total particle number concentration of i -th size mode with the mean radius $r_{g,i}$ and standard deviation w_i .

The scattering coefficients for particles with size dispersion are simply number distribution integrated extinction cross-sections:

$$\alpha_{ext} = \int \sigma_{ext}(r) \frac{dN(r)}{d \ln r} d \ln r \quad \alpha_{sca} = \int \sigma_{sca}(r) \frac{dN(r)}{d \ln r} d \ln r \quad 1.5$$

And the total column aerosol optical depth (AOD) is the path integral of the extinction coefficient from the bottom to the top of atmosphere.

$$\tau(\lambda) = \int_0^{TOA} \alpha_{ext}(\lambda, z) dz \quad 1.6$$

Furthermore, the extinction efficiency Q_{ext} coefficient based on Mie theory for spherical homogeneous particles is a function of size parameter which results in unique spectral behavior:

$$x = \frac{2\pi r}{\lambda} \quad 1.7$$

For example, particles with radius $x \ll 1$ are less effective scatters of solar radiation. In this Rayleigh scattering regime, when particles are much smaller than the incident wavelength to the negative forth power of the wavelength:

$$Q_{ext} \propto \lambda^{-4} \quad 1.8$$

This inverse 4th power behavior shows that the shortest wavelength is most efficient in scattering light and is the fundamental explanation of the blue sky color.

Since gases are well mixed, the optical depth due to the molecules in the atmosphere can be very easily quantified and depends only on the surface pressure and weakly on temperature. According to Teillet, the basic equation for Rayleigh optical depth is:

$$\tau_R = \frac{8\pi(n^2 - 1)^2 N_c}{3\lambda^4 N_s} \left(\frac{6 + 3\gamma}{6 - 7\gamma} \right) \left(\frac{P}{P_0} \right) \left(\frac{T_0}{T} \right) \quad 1.9$$

Where n is the refractive index of air, γ is the depolarization factor which describes the effect of molecular anisotropy, N_c is the columnar number density which is

$2.154 \times 10^{25} \text{ cm}^3$ for standard conditions, N_s is the molecular density for which $2.547 \times 10^{19} \text{ cm}^3$ for standard conditions, p pressure using $p_0 = 1013.25 \text{ mb}$ for standard conditions and T temperature in Kelvin using $T_0 = 288.15 \text{ K}$ for standard conditions. If we assume pressure and temperature are at standard conditions: $p = p_0$ and $T = T_0$ a useful normalized formula that can be used from Hansen and Travis:

$$\tau_R = 0.008569 \lambda^{-4} (1 + 0.0113 \lambda^{-2} + 0.00013 \lambda^{-4}) \quad 1.10$$

To compute the aerosol optical thickness from direct transmission (or measurements, the Rayleigh scattering (also known as molecular scattering) and gas absorption must be subtracted from total optical thickness. Using the total optical thickness τ_λ :

$$\tau_\lambda = \tau_{scatter} + \tau_{absorption} = \tau_{A_\lambda} + \tau_{R_\lambda} + \tau_{O_\lambda} \quad 1.10$$

The extrapolation of the extraterrestrial irradiance I_0 or zero airmass signal via Langley method [Shaw 1993] was implemented to obtain aerosol optical depths spectral measurements from the calibrated data obtained from the MFRSR (Multi filter Rotating Shadow Band) and sun photometer CIMEL data files.

2 HOW ARE AEROSOLS MEASURED

There are many agencies monitoring aerosols, including: NOAA (The National Oceanic and Atmospheric Administration). “NOAA’s Environment Satellite, Data and Information Services (NESDIS) is dedicated to providing timely access to global environmental data from satellite and other sources, to promote, protect and enhance the Nation’s economy, security, and environment and quality of life” [<http://noaa.gov>]. Their main goal is to provide operational monitoring. Until recently, their main focus has been weather related but with the increase in global warming forcing and the possibility of aerosols modifying climate and weather in meaningful ways, NOAA has become much more interested in Aerosols.

NASA (National Aeronautics and Space Administration) is another institution dedicated to the study of the atmosphere with primary focus on long term climate issues which in fact is the main focus of the NASA Goddard Institute of Space Science (GISS).

While NASA has developed multiple aerosol instruments for study including MISR (Multi-angle Imaging SpectroRadiometer) and MODIS (Moderate Resolution Imaging SpectroRadiometer), these instruments are not optimized for aerosol detection. However, GISS is soon to deploy the Advanced Polarimetric Scanner (APS) instrument on GLORY which due to its extensive angular, spectral and polarization coverage, is most likely to meet climate level aerosol retrievals

In addition, NASA Langley has pioneered the development and deployment of the most extensive and accurate network of Ground radiometers. Besides measuring optical

depth, the scanning nature of these instruments provides multiangle multiwavelength measurements of the sky radiance. Since this is a function of the aerosols, this network is particularly useful for satellite validation providing not only aerosol optical depth but fine-coarse mode partitioning, single scatter albedo and particle size distributions. Unfortunately, the network instruments are expensive and not portable by motivating the study of less accurate but more portable and inexpensive alternatives.

Environmental Protection Agency (EPA) The basic mission of the Office of Air Quality Planning and Standards within the EPA is to preserve and improve the quality of our nation's air. Using a nationwide network of monitoring sites, EPA has developed ambient air quality trends for particle pollution, also called Particulate Matter (PM) and based on epidemiological studies has set exposure guidelines for both PM₁₀ (Particles less than 10microns diameter) and PM_{2.5}.

2.1 Satellite Measurements

The Ozone Monitoring Instrument (OMI) is aboard of EOS-Aura. OMI can distinguish between aerosol types, such as smoke, dust, and sulfates, and measures cloud pressure and coverage, which provide data to derive tropospheric ozone. OMI continues the TOMS record for total ozone and other atmospheric parameters related to ozone chemistry and climate. As discussed earlier, knowledge of Ozone is important in

correcting the aerosol optical depth and OMI Ozone data products are being used as input for processing MFRSR data.

The workhorse instrument is the Moderate Resolution Imaging Spectro Radiometer (MODIS) aboard the Terra (Earth Observing System: EOS AM) and Aqua (EOS PM) satellites. These satellites orbits around the Earth in the morning from north to south across the equator, while Aqua passes south to north over the equator in the afternoon. Terra MODIS and Aqua MODIS are viewing the entire Earth's surface every 1 to 2 days, acquiring data in 36 spectral bands. Data products from MODIS give scientist an understanding of global dynamics and what is happening on the land, in the oceans, and in the lower atmosphere. MODIS plays a vital role in the development of validated, global, interactive Earth system models able to predict global change accurately enough to assist policy makers in making sound decisions concerning the protection of our environment [<http://terra.nasa.gov/>].

Aerosols are retrieved over dark surfaces only. For example, over oceans there is no problem obtaining information about aerosols, figure 2.1. And over land selection of surfaces that are dark in the IR, estimation of surface reflectivity from IR value with exclusion of clouds and glint is base on an inversion assumption of 2-modal log-normal distribution.

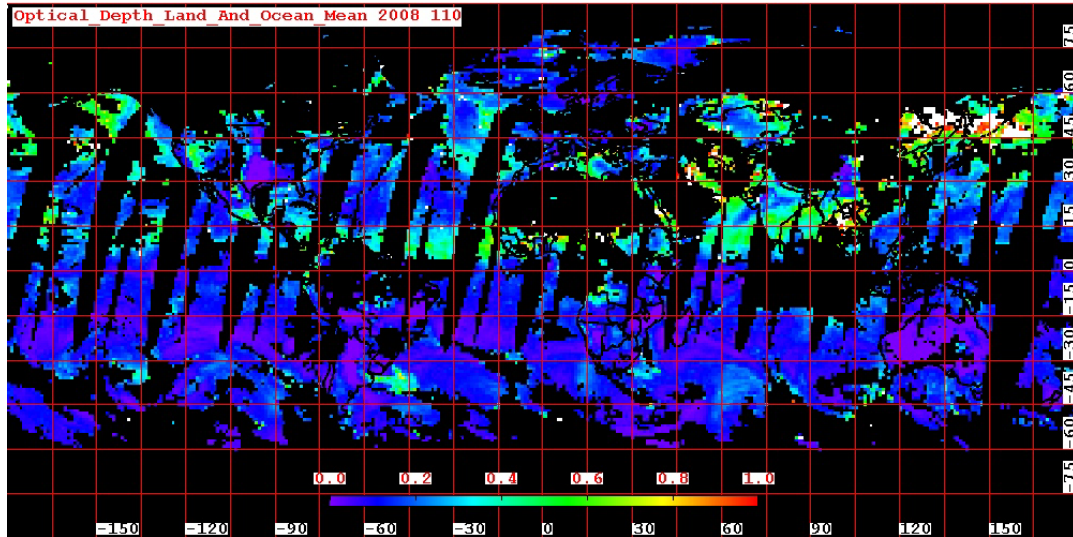


Figure 2.1 MODIS optical depth for April 19, 2008 are only defined over dark oceans

2.1.2 MODIS Strengths Weaknesses

Ground reflection is significant and hard to quantify over land. It is even complex to obtain reflection at the Top of Atmosphere (TOA) which combines reflectance from the ground and the atmosphere. MODIS AOD retrieval is described in the Collection 4 algorithm documentation was focused on using “dark” pixels such as dense vegetation as a way of isolating aerosol contribution [Kaufman, 2001]. The surface reflectance ratios that connect the visible (VIS) ground albedo to the shortwave infrared (SWIR) albedo were set to values that were suitable for vegetation. These weaknesses are not realistically met over large urban areas. Besides, the requirement of dark pixels is a prime reason for the low resolution AOD retrieval (at 10 km) by excluding bright pixels prior to processing. Therefore, any model that can reduce errors from unknown surface reflection is crucial for retrievals over land as a whole and urbanized land sections. There have been

significant efforts to improve MODIS aerosol product processing over the Collection 4 aerosol retrieval over land algorithm in Collection 5 [Levy 2004 and 2007]. New surface models are being tested using different land surfaces to obtain the required reflectance. One of these efforts was based on global AERONET models which were used to atmospherically correct MODIS TOA reflectance's to extract ground reflectance. In performing these analyses, it was necessary to obtain global aerosol models based on cluster analysis techniques.

As an example of the difficulties, we can plot the results of MODIS and MISR vs. AERONET for the CCNY location figure 2.2.

To motivate how difficult satellite retrieval of aerosols over urban areas are, we first located MODIS operational C005 L2 aerosol data over the period from 2004-2007 in a 10 km x 10 km box surrounding the AERONET station at CCNY. Similarly for MISR data, we use MISR L2 aerosol properties data in 17.6 x 17.6 km resolution box surrounding the AERONET station in the same period of time. Since we use MODIS data from Terra satellite, both MODIS and MISR data are from the same the platform.

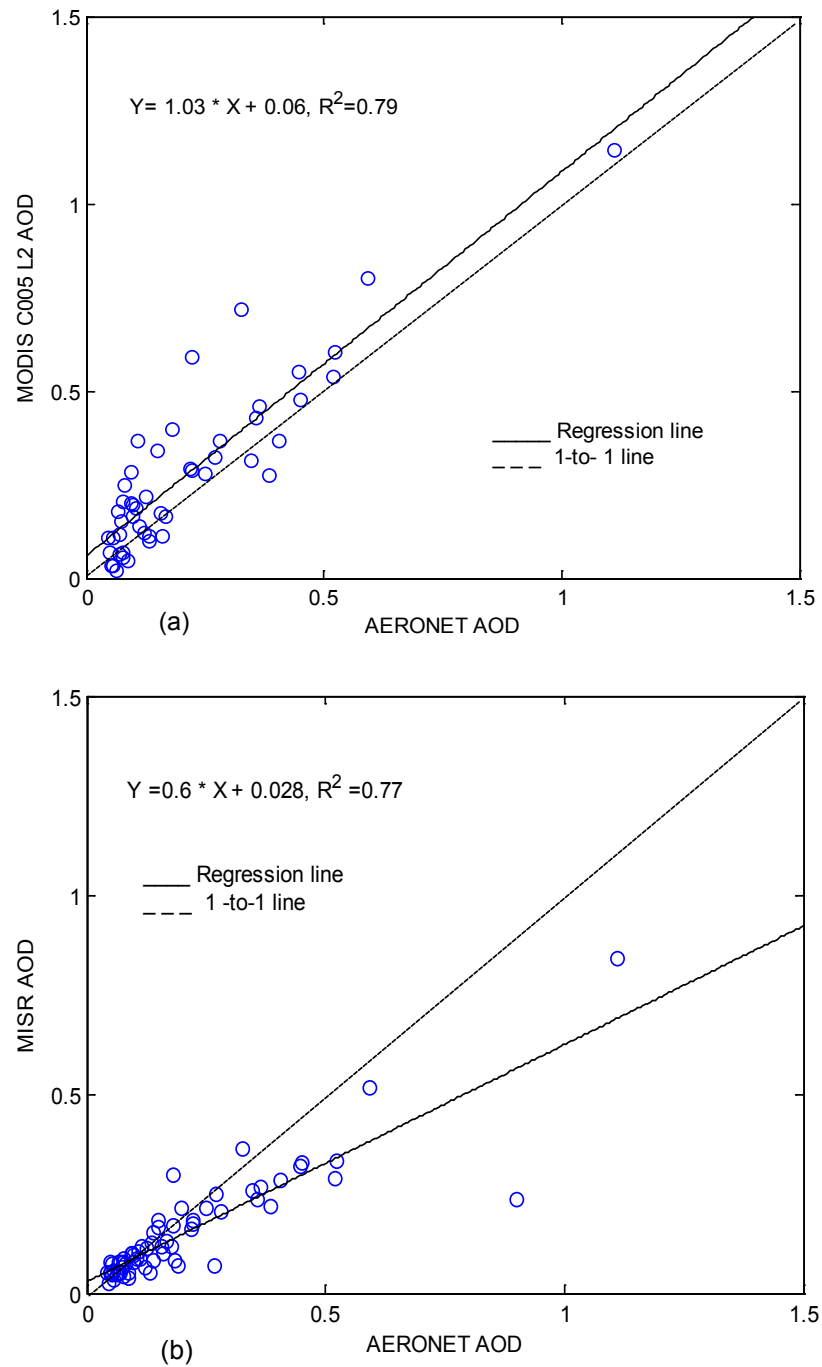


Figure 2.2 a) MODIS L2 Aerosol Optical Depth at 550nm compared with 4 hour average of AERONET aerosol optical thickness b) MISRS L2 Aerosol Optical Depth at 558nm compared with 4 hour average of AERONET aerosol optical thickness

We compare MODIS and AERONET AOD at 550 nm wavelength and MISR and AERONET at 558nm, since MODIS L2 AOD data is available at 550nm wavelength and MISR L2 AOD data is available at 558nm. For AERONET data, we interpolated AERONET AOD from 440nm, 500nm and 675nm to required wavelengths at 550nm and 558nm. The comparison of MODIS-AERONET and MISR-AERONET data are shown in figure 2.2 with the 1-1 line plotted as a dashed line and the linear fit as a solid line. The MODIS-AERONET scatter plot linear fit line shows $\tau_{\text{MODIS}} = 1.03 \tau_{\text{AERONET}} + 0.06$ and $R^2=0.79$, where number of points is $N= 67$ in figure 2.2. We recognize that MODIS AOD data is significantly overestimated in comparison with AERONET AOD data. This observation is not unique to NYC. For example, similar results were found at Mexico City urban area in MILAGRO project (Megacity Initiative: Local and Global Research Observation project). More details of MILAGRO project can be found at Castanho et al., 2007.

In figure 2.2 (b) MISR- AERONET AOD comparisons is shown, the dash line is the one to one line and solid line is the linear fit of the scattered data. The equations of the linear fit line is $\tau_{\text{MISR}} = 0.6 \tau_{\text{AERONET}} + 0.028$ and $R^2=0.77$, where number of points is $N= 67$. We find that MISR AOD data is significantly underestimated in comparison with AERONET AOD data. Similar results were found in MISR-AERONET comparison in Beijing metropolitan area. The regression analysis comparison result from metropolitan area Beijing shows that, MISR AOD as the response to AERONET, the slope is approximately 0.58 ± 0.03 and an intercept of 0.07 ± 0.02 at 558nm to Jiang et al., 2007.

In addition, Kahn et al. [2005] found similar underestimates of AOD retrievals at several urban sites and was attributed to the need to include into the aerosol models urban polluted models with lower single scattering albedo (SSA) which could raise the AOT and hence better retrieve the particle properties.

2.1.3 Regional Training of Urban Surface Models Using Ground Sensors.

However, it is possible that using MFRSR AOD measurements together with MODIS, the urban surface can be refined. The idea is to implement on a local scale what MODIS is on a global scale in retying to determine the ratios between the VIS and SWIR channels.

Following MODIS methodology, we only use atmospheres which we could verify a-priori as fine mode. In fine mode dominated aerosol conditions, the SWIR reflectance at TOA is totally due to the ground since the SWIR wavelength is transparent to the fine mode aerosol atmosphere [Kaufman et al. 1997]. Therefore we can approximate the surface reflectance at $2.12 \mu\text{m}$ as the TOA reflectance and not concern ourselves with aerosol contributions at the SWIR during surface retrieval. To identify fine mode dominant cases, we select only comparisons when the angstrom coefficient $\alpha > 1$ (as defined in equation (2.1)), and to further reduce errors due to misclassification of aerosols, we limit the AOD to $\tau_{0.55\mu\text{m}} > 0.2$.

$$\alpha = \frac{\log(\tau_{675nm} / \tau_{1020nm})}{\log(1020 / 675)} \quad 2.1$$

Moreover, to ensure horizontal homogeneity in the aerosol during the surface retrieval, we require at least 10 cloud cleared points to exist in AERONET measurement in the 4 hour interval (approximately 2 hours before and 2 hours after MODIS/Terra passing time) surrounding the MODIS observations and relative standard deviation in the AOD to be less than 20%. Using a reasonable assumption of mean wind speeds between 10 and 20 km / hr, 4 hour time stability can be converted into 40-80 km spatial homogeneity [Anderson et al 2003]. These filters help to reduce possible errors in our assumption that the aerosols are of the urban type and to ensure that regions in the vicinity of the AERONET site are described approximately by the same aerosol model.

To use the AERONET AOD data, we must be able to inject this information directly into the radiative transfer code. Assuming a Lambertian surface albedo, the surface reflectance can be written as

$$\rho_{\lambda}^{surf}(\theta_0, \theta, \Delta\phi, \tau) = \frac{\rho_{\lambda}^{TOA}(\theta_0, \theta, \Delta\phi, \tau) - \rho_{\lambda}^{path}(\theta_0, \theta, \Delta\phi, \tau)}{\bar{s}(\theta_0, \tau)(\rho_{\lambda}^{TOA}(\theta_0, \theta, \Delta\phi, \tau) - \rho_{\lambda}^{path}(\theta_0, \theta, \Delta\phi, \tau)) + T_{\lambda}^{up}(\theta_0, \theta, \tau)T_{\lambda}^{dn}(\theta_0, \tau)} \quad 2.2$$

In equation (2.2), ρ_{λ}^{TOA} is the measured top of atmosphere reflectance from MODIS L1 B cloud free data (where λ = MODIS band 3 (0.47 μ m) and MODIS band 1 (0.66 μ m)), ρ_{λ}^{path} is the atmospheric path reflectance, $\bar{s}$ is the atmospheric albedo, T_{λ}^{dn} is

the downward transmission, and T_{λ}^{up} is the upward transmission, θ_0 , θ , $\Delta\varphi$ are the solar zenith angle, sensor (satellite) zenith angle and relative solar sensor azimuth angle respectively [Remer et al., 2003, Levy et al, 2007 b]. Ingesting the AERONET AOD and MODIS urban aerosol phase function is sufficient to obtain the surface reflections for both visible channels independently. Furthermore, since the aerosols are fine mode dominated, the ground reflectance at $2.12 \mu\text{m}$ is directly obtained allowing us a measure of the VIS-SWIR surface reflectance ratio.

Using equation (2.2), the surface reflectance's can be calculated. The results of these calculations as a function of the scattering angle are shown in Figure 2.3 for 3 spatial resolutions (10 km, 3 km and 1.5 km) surrounding the City College of New York (AERONET site CCNY). We also masked the inland water area contamination to our derived surface reflectances, since CCNY is so close to Hudson River. More details of the masked can be found at Oo et al, 2010. We first note that the calculated VIS/SWIR reflectance ratios are higher than those implemented from MODIS operational algorithm defined VIS/SWIR reflectance ratios and these ratios are in good agreement with VIS/SWIR ratios result from Hyperion data for New York City urban area [Gross et al 2003]. Furthermore, we do not detect any meaningful trend behavior as a function of angle even though the data is aggregated over the entire multi-year training period. In fact, any trend observed is significantly smaller than the fluctuations around the trend line. It should also be pointed out that the approach was validated against vegetation targets returning values consistent with the C004 and C005 values.

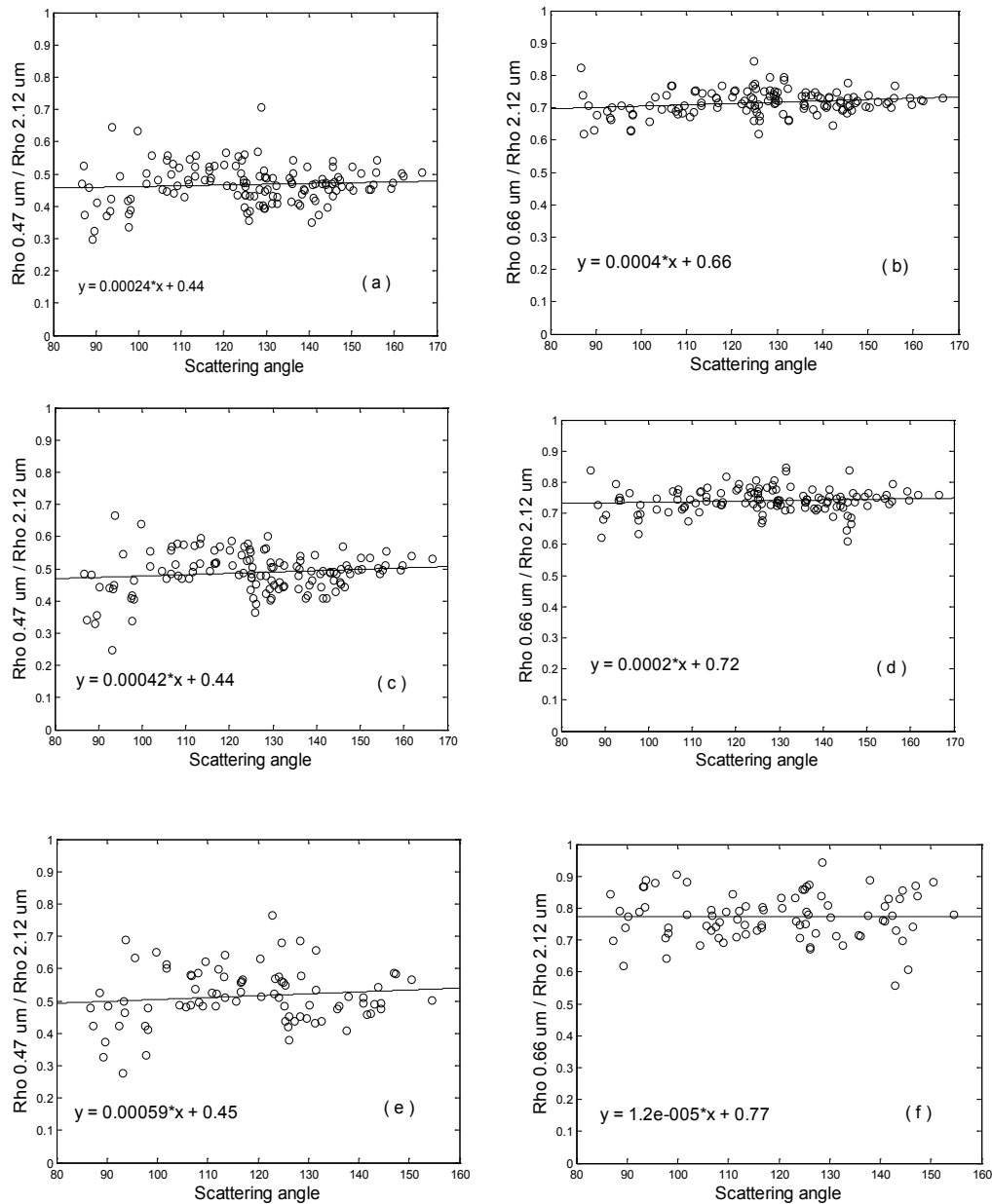


Figure 2.3 Surface reflectance ratio of 0.47 μm / 2.12 μm and 0.66 μm / 2.12 μm with respect to (x axis) scattering angle. Mean reflectance ratio for different spatial resolution: 10 km x 10 km (a-b), 3 km x 3 km (c-d) and 1.5 km x 1.5 km (e-f) surrounding to the nearest pixel to AERONET site at CCNY from 2001 to 2006 with fine mode aerosol dominant atmosphere (with inland water areas and rivers masked)

When these ratios are used in the operational AOD retrieval processing stream, a significant improvement is found as shown in Figure 2.4. In fact, the uncertainty even at higher resolution is quite small. In both these figures, the operational error bounds of $(\pm .05 \pm .15\tau)$ for AOD retrieval over land were added and show that except for very low AOD values where even small errors in reflection compensation can lead to significant AOD errors, the AOD retrievals are well within these uncertainty bounds.

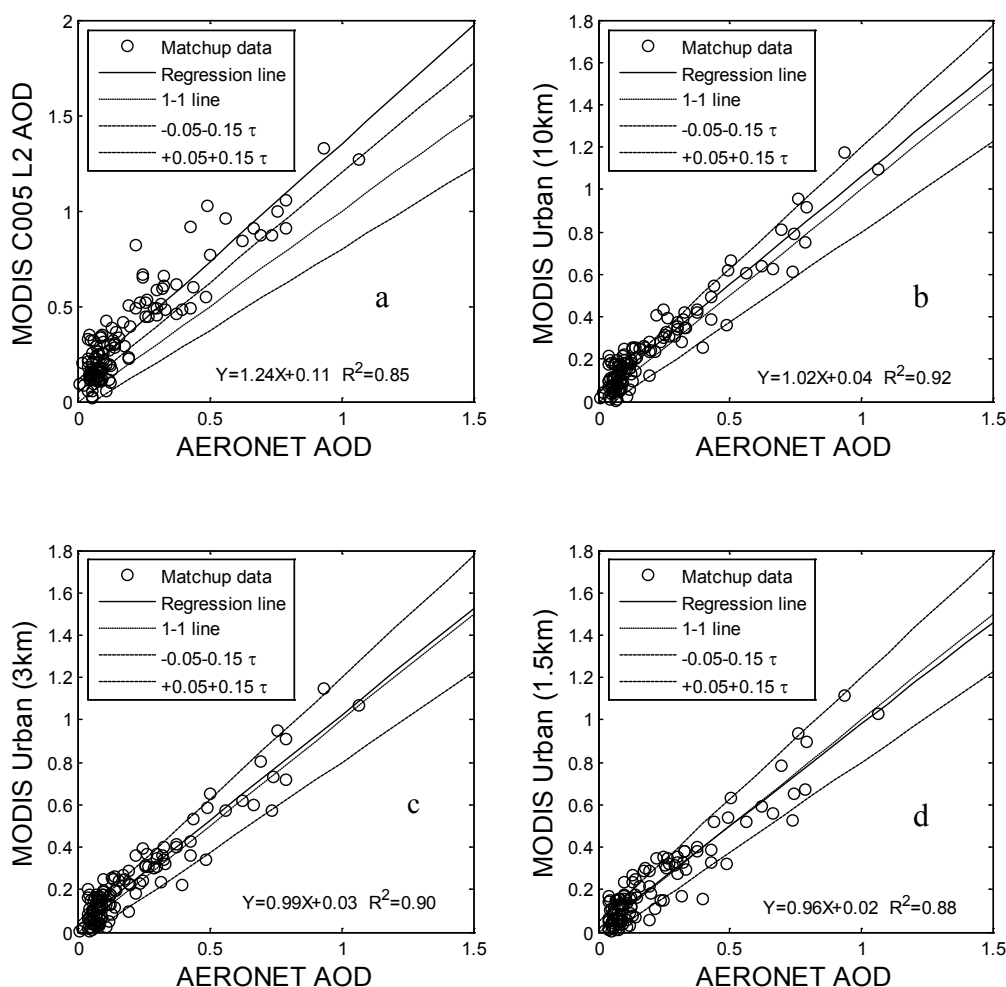


Figure 2.4 (a) MODIS L2 Aerosol Optical Depth at $0.55\ \mu\text{m}$ compared with 4 hour ($\sim 2\text{hr}$ before and 2hr after MODIS (Terra) satellite overpass time) average of AERONET aerosol optical thickness. (b), (c) and (d) are retrieved AOD with new surface reflectance

VIS/SWIR ratio plot with average of AERONET AOD, inland water areas and rivers masked.

The main points to take from this analysis are:

1) Using ancillary data from ground sensors should be able to improve the surface reflection models. Using AERONET required that the aerosol loading was very stable over the day. However, with MFRSR network measurements at critical locations, the aerosol loading can be determined over a much broader region

2) Using the regional surface ratio values, the MODIS AOD output can be obtained at much higher spatial resolution. The implementation of MFRSR instruments that are closely spaced on urban scale can ultimately attempt to confirm this.

2.2 In-situ Instruments

Ground base instruments offer a great potential for aerosol optical retrievals as well as directly measuring Aerosol Particulate Matter. For our work, we focus mainly on the sky radiometer which provides estimates of aerosol optical depth and microphysical properties, the MFRSR shadowband radiometer and the TEOM instrument from EPA (deployed over New York through NYS DEC Department of Environment Conservation) for in-situ measurements.

2.2.1 Tapered Element Oscillating Microbalance --TEOM

The Tapered Element Oscillating Microbalance (TEOM) is a low-volume air sampler that uses a mass sensor to measure airborne particle mass in real time. A TEOM contains the following parts: an inlet that can be adjusted for a particulate size, a flow splitter, a sample filter, a microbalance, a flow controller timer, and manufacture software which fully automates the instrument operation. In practice, the TEOM collects PM on a filter located on the top of a hollow, oscillating tapered tube. A small portion of the incoming air flow is drawn through the filter and through the tube. The oscillation frequency of the tapered inlet tube is inversely proportional to the mass of the sample that is deposited on the collection filter. The frequency decreases as mass accumulates on the filter, providing a direct measure of inertial mass. The typical measurement is collected over a period of ten minutes. The sample chamber is maintained above ambient temperatures (30-50C) to minimize the effect of temperature changes and thermal expansion of the tapered element that may affect the oscillation frequency, and to reduce particle-bound water [Patashnick et al., 1991]. TEOM and AIRNow measure particles simultaneously. The TEOM measure continuous total mass concentration of PM_{2.5} or PM₁₀ depending on the type of cyclone you use in the sample inlet. The instrument is easy to manipulate and also easy to download the data. The difficulties we can encounter have to do with the instruments calibration, or air leaking from some connections. But once the instrument is working properly, within the parameters, we should not have any problem.

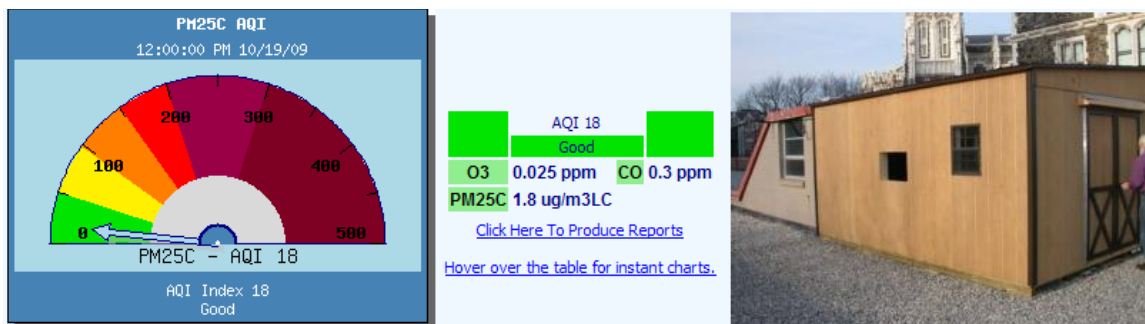


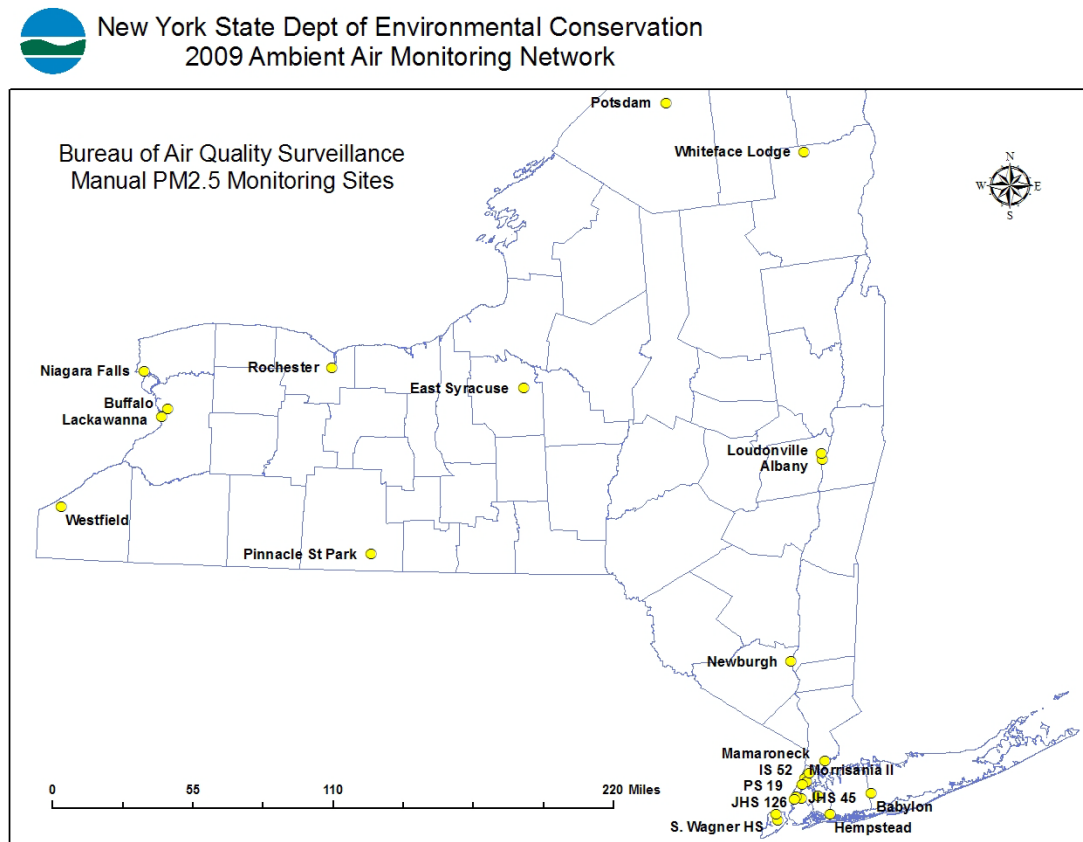
Figure 2.5 Data products from CCNY can be downloaded from:
<http://www.dec.ny.gov/airmon/stationStatus.php?stationNo=73>

The City College of New York (CCNY) contains an Aeronomy Lab, located atop the Administration Building on the CCNY campus to gather air quality data since spring 2007; figure 2.5 displays the air quality gauge reading. PM 2.5 and O₃ Data can be retrieve <http://www.dec.ny.gov/airmon/stationStatus.php?stationNo=73>. The New York State Department of Environmental Conservation operates equipment in our lab to gather and report this concentration data representing 2.5 micrometers in size or smaller aerodynamic diameter. Since airborne particles have irregular shapes, their aerodynamic behavior is expressed as the diameter of an idealized spherical particle known as aerodynamic diameter. Studies by Allen et al. 1997, Chung et al. 2001 and Cook et al. 1995 have shown that the concentration of PM10 or PM2.5 mass using the TEOM are often lower than PM measurements produced by other methods. Analysis of the constituents typically found in such cases indicates that this is caused by loss in the heated sample chamber of semi volatile PM, such as ammonium nitrate or organic compounds. While most of the volatile components are found in the fine PM fraction, discrepancies between PM10 TEOM and reference samplers have also been observed.

MEASUREMENT	INTERVAL	EQUIPMENT
PM ₁ size distribution between 0.01- and 1- mm	10-minute	TSI 3936L85 Scanning Mobility Particle Sizer (SMPS)
PM _{2.5} mass	5-minute	Rupprecht and Patashnick 1400a Tapered Element Oscillating Microbalance (TEOM) with a nafion-based Sample Equilibration System (SES) that allows for 30-degC sampling)
PM _{2.5} organic and elemental carbon mass	4-hour	Sunset Laboratory Non-dispersive Infrared (NDIR) Carbon Analyzer
PM _{2.5} organic and inorganic composition	24-hour	Fourier Transform Infrared Spectrometer analysis of PM collected on filters
Gas phase SO ₂ concentration	1-minute	API 100e UV pulsed fluorescence monitor
Gas phase NO _x concentration	1-minute	API 200e Chemiluminescence monitor
Gas phase CO concentration	1-minute	API 300e Gas-filter correlation monitor
Gas phase O ₃ concentration	1-minute	API 400e UV absorbance monitor

Table 2.1 Table products for TOEM at the CCNY.

Figure 2.6 shows the density of PM_{2.5} sites in the NYS network with a particular emphasis on the NYC area.

Figure 2.6 PM_{2.5} monitoring network for NYSDEC

2.3 Radiometric Measurements

2.3.1 CIMEL Sun Photometer CE-318A

The electronic of ground-based atmospheric aerosol measurements using sun photometry has changed substantially since Volz [1959] introduced the first handheld analog instrument. Modern digital and microcontroller units of supreme quality and robustness can collect data accurately and quickly with the onboard processing microprocessor [Schmid et al., 1997, Ehsani, 1998, Forgan, 1994, and Morys et al., 1998]. Sun photometers are narrow field of view radiometer instruments designed to measure solar irradiance. Their typical field-of-view is between 6 and 10 degrees with well defined spectral bands; each of them in the order of 10 nm FWHM (Full Width Half Maximum). Modern instruments are electronically controlled like the CE-318A made by CIMEL a French manufacture.



Figure 2.7 The CE 318 automatic sun tracking photometer has been designed and realized to be a very accurate sun photometer with all the qualities of a field instrument.

The CE318A is a sun photometer that works with solar power. It has a robotic arm managed by a microprocessor that uses a sun tracking mechanism to measure the sun and the sky irradiance. Sun photometer uses a filtered detector that measures the spectral extinction of direct beam radiation according to equation 1.3, the Beer-Lambert-Bouguer law with minor changes.

$$V_{\lambda} = V_{0\lambda} d^2 t_y \exp(\tau_{\lambda} m) \quad 2.3$$

Where V_{λ} is the digital voltage measured at specific spectral band (λ) as a function of ($V_{0\lambda}$) the extraterrestrial voltage modified by the relative (d) earth sun distance ratio. The exponent is the same as equation 1.3 with an extra factor t_y which is the gaseous absorption correction. Direct and diffuse solar measurements can be used to recover atmospheric parameters including AOD, perceptible water, sky irradiance distributions and ozone amount. Aerosol volume and size distribution are retrievable by inversion modeling from spectral aerosol optical depth [Rollin]. This instrument has a 1.2° degree field of view and two detectors to measure direct sun, aureole and sky irradiance. Figure 2.7 shows a sun photometer CE-318A, it contains two collimators that measure the aureole at about 3° degrees from the sun. The instrument is mounted on an arm robot sensor pointed at nadir when not taking data to prevent contamination of the optical collimator from suspended particles and rain. The sun/aureole collimator is protected by a quartz window allowing observation with a UV enhanced silicon detector with sufficient signal-to-noise for spectral observations between 300 and 1020 nm [Tandre, 1998]. The sun photometer makes two measurements direct sun and sky irradiance each measurement with preprogrammed sequences.

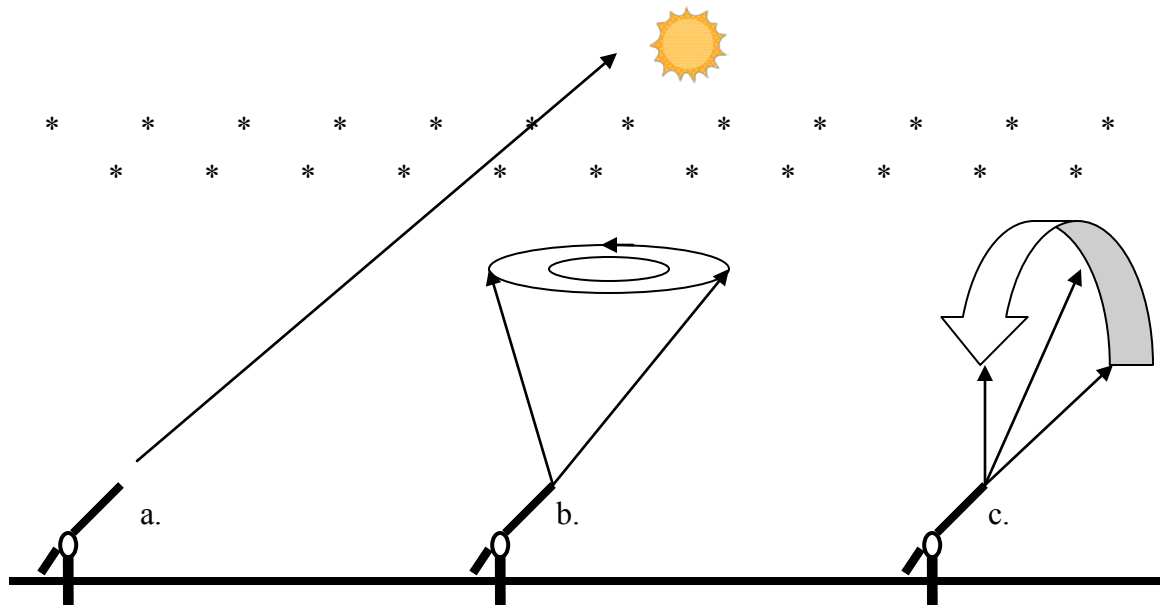


Figure 2.8 Representations of the a) Direct, b) Almucantar and c) Principal Plane measurements. Almucantar and principal plane are done only at 440, 670, 870 and 1020 nm channels. Direct measurement is done at all wave lengths: 340, 380, 440, 500, 670, 870, 940 and 1020 nm

The direct sun, figure 2.8a, measurements are made four times per hour at eight spectral bands. Using a combination of special filters it takes measurements at selected spectral bands: 340, 380, 440, 500, 675, 870, 1020 and 1640 nm that requires about ten seconds, table 2.2. The direct sun measures are repeated 3 times with 30 seconds interval to obtain a triple observation. This observation is used for cloud contamination. This triple observation is taken every 15 minutes for morning and afternoon which creates a larger temporal resolution that allows comparison between observations for cloud contamination since cloud observations are larger than aerosol measurements. The direct sun measurement gives the following products aerosols optical depth, perceptible water and Angstrom coefficient.

	Spectral Range nm	Target	No. Obs.	Obs. Inter.	Appli.
Direct Sun	340-1020	sun	1 ea. λ	~8 sec. for .8 λ	AOT, Pw, α
Triple Obs.	340-1020	sun	3 direct sun	3@30 sec. apart 1 min total	AOT, Pw, α and cloud screening
Std. Measu. Seq.	340-1020	sun	Variable: depends on day length.	Ea. 15 min m=2am to m=2pm	AOT, Pw, α
Langley	340-1020	sun	16 am & pm btw m 7 & 2	m=7-5 increm. .5m ; m=5-2 increm. .25m	AOT, Pw, α , Langley Cal.
Basic Sky	440-1020	sky	1 ea. λ	none	Sky Radiance
Langley Sky	440-1020	sky	16 btw. m 7 & 2	m=7-5, .5m m=5-2, .25m	Stability Langley plot
Almucantar	440-1020	sky	72 table 3.2	>8/day:m=4,3,2,1. 7 hrly. 9AM to 3PM	AOT, α , Size dist. $P(\theta)$
Polarization	870	sky	42 table 3.2	hrly m=3AM to m=3pm	Size dist. $P(\theta)$
Prin. Plane	440-1020	sky	42 table 3.2	hrly m=3AM to m=3pm	AOT, α , Size dist. $P(\theta)$

Table 2.2 Measurement sequences of the CIMEL sun/sky scanning spectral radiometer where m is the airmass ($\sec \theta$) [Holben 1998]

	Sun	Sky($^{\circ}$)
Almucantar Azimuth angle relative to sun	0 $^{\circ}$	6.0,5.0,4.5,4.0,3.5,3.0,2.5,2.0,-2.0,-2.5,-3.0,3.5,-4.0,-4.5,-5.0,-6.0,-8.0,-10.0,-12.0,-14.0,-16.0,-18.0,-20.0,-25.0,-30.0,-35.0,-40.0,-45.0,-50.0,-60.0,-70.0,-80.0,-90.0,-100.0,-110.0,-120.0,-130.0,-140.0,-160.0,-180.0. Duplicate above sequence for a complete counter clockwise rotation to -6
Principal Plane: Standard scattering angle from sun (negative is below sun)	0 $^{\circ}$	-6.0,-5.0,-4.5,-4.0,-3.5,-3.0,-2.5,-2.0,2.0,2.5,3.0,3.5,4.0,4.5,5.0,6.0,8.0,10.0,12.0,14.0,16.0,18.0,20.0,25.0,30.0,35.0,40.0,45.0,50.0,60.0,70.0,80.0,90.0,100.0,110.0,120.0,130.0,140.0
Principal Plane: Polarization scattering angle from sun (negative is in the anti solar direction)	-	-85.0,-80.0,-75,-70,-65.0,-60.0,-55.0,-50.0,-45.0,-40.0,-35.0,-30.0,-25.0,-20.0,-15.0,-10.0,-5.0,5.0,10.0,15.0,20.0,25.0,30.0,35.0,40.0,45.0,50.0,55.0,60.0,65.0,70.0,75.0,80.0,85.0

Table 2.3 Almucantar and Principal Plane sequences for the Standard and Polarization [Holben 1998]

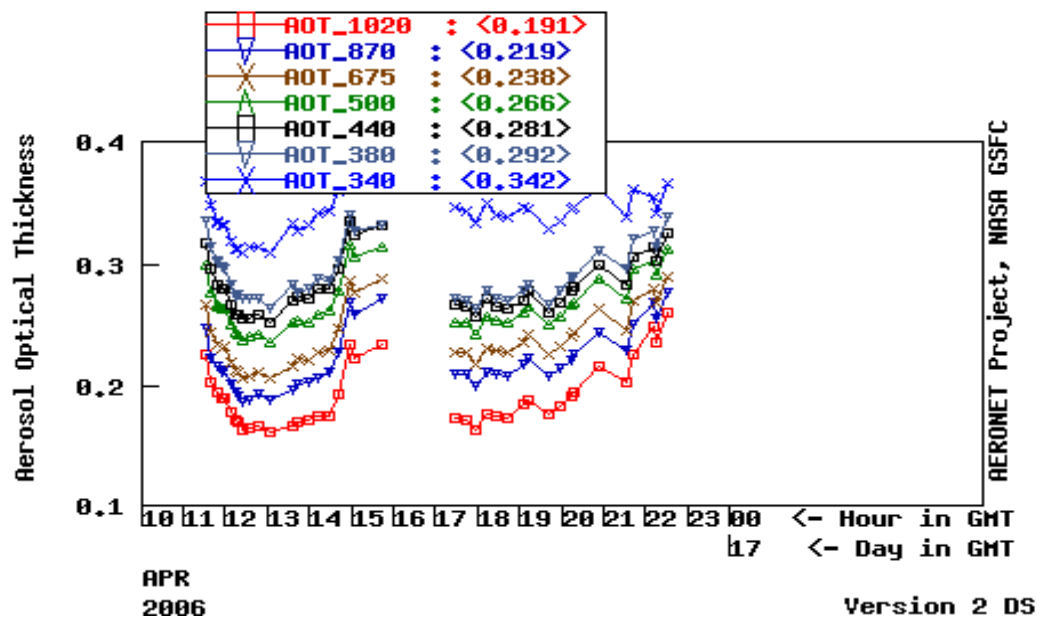
Following the triple observation, Langley calibration intervals also take place at every 15 minutes. Langley calibration observations of direct sun are made between airmass of 2 to 7 for both morning and afternoon. The relative Airmass (m) is approximately by using $\sec(\theta_z)$ of the θ_z zenith angle. The zenith angles between 60 $^{\circ}$ to 82 $^{\circ}$ degrees are essential for the correct determination of the Langley calibration slope. Diffuse or sky measurements are approximately once per hour at selected spectral bands: 440, 670, 870 and 1020 nm, table 2.2 and table 2.3. A single spectral measurement sequence, Langley sky, is made immediately after the Langley airmass direct sun measurement, 20 $^{\circ}$ degrees from the sun. This is used to assess the stability of the Langley plot analysis according to O'Neill et al., 1989. There are two diffuse sky measurements, figure 2.10 b and c, the Almucantar and the Principal Plane. The almucantar scans the horizontal plane at the

solar zenith angle θ_z , with measurements at specified azimuths relative to the sun counterclockwise through 360° of azimuth. Almucantar measurements are typically made at an airmass of 2 or less to maintain large scattering angles [Rollins E.M.]. The principal plane is a swept of the sky in a vertical direction rotating orthogonal to the azimuth plane spanning to 150° degrees. The philosophy of the almucantar and principal plane measurements is to acquire aureole and sky irradiances observations through large range of scattering angles from the sun to construct a constant aerosol profile to retrieve size distribution, phase function and AOD [Holben, 1998]. Sky-scanning channels are calibrated by the use of the NASA Goddard Space Flight Center's 2-m integrating spheres, and the direct solar scans are compared with a reference instrument calibrated with Mauna Loa Langley Observatory. Since this device measures the spectral sky radiance at known angular distances from the sun, then it requires precise aureole measurements near the solar disk and good stray light rejection. Sun photometer have expanded the aerosol knowledge through inversion of the sky radiances to derive aerosol microphysical properties such as size distribution and optical properties such as phase function [Nakajima et al. (1983, 1996), Tanré et al. (1988), Shiobara et al. (1991), and Kaufman et al., (1994)]. According to Holben, the measured uncertainty in the retrieved AOT is estimated to be approximately 0.01 - 0.02 [Holben, 1998]. Our sun photometer is identified at the NASA AERONET (AErosol RObotic NETwork) as instrument #237.

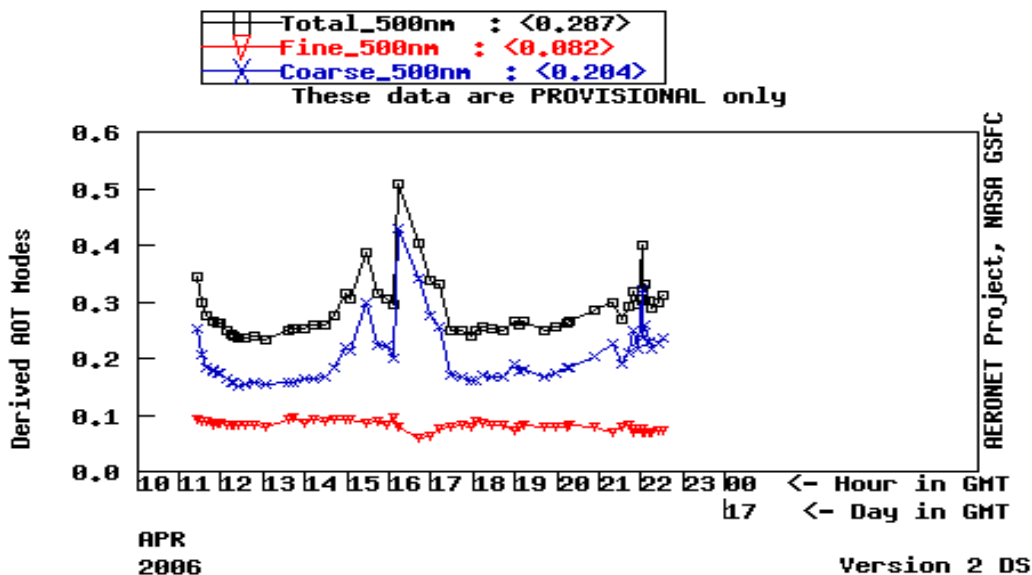
Light is absorbed and scattered by atmospheric gasses, water vapor and aerosols. The concentration of different atmospheric components can be determined by the attenuation at wavelengths which are strongly absorbed or scattered. Total Optical Depth (TOD) is the sum of the Rayleigh Optical Depth (from atmospheric gasses, e.g. Nitrogen, Oxygen, and Argon), the Ozone Optical Depth, the Mixed Gas Optical Depth (e.g. Carbon dioxide, oxides of Nitrogen), the Water Vapor Optical Depth and the Aerosol Optical Depth (AOD). AOD is of the most interest because it cannot be directly measured by other methods, and is determined by subtraction of the other quantities. Aerosol phase functions and size distribution can also be inferred. The sun photometer measurements are also important as standards for satellite-derived values of AOD. Data collected from the AERONET device# 237 are analyzed by Langley Regression using different levels to obtain the AOD. Level 1.0 raw data no calibrated and no cloud screen. Level 1.5 Cloud screened and Level 2.0 is cloud screened and data has been calibrated. The multi-filter rotating shadow-band radiometer (MFRSR) employs a different strategy. It measures spectral total and diffuse radiation to obtain the direct component from which aerosol optical thickness is computed using equation 1.3. The instrument nominally measures at one minute intervals and has been shown to be reliable over long periods of time. The measurements are networked to a common server by a modem interface and the data processed by a common analysis system [Harrison et al., 1994].

An example of the data output from AERONET for multispectral AOD is given in figure 2.11. This example is clearly of interest as one where strong coarse mode loading exists.

CCNY , N 40°49'15", W 73°56'56", Alt 100 m,
 PI : Barry_Gross, gross@ccny.cuny.edu
 Level 1.5 AOT; Data from 16 APR 2006



CCNY , N 40°49'16", W 73°56'56", Alt 100 m,
 PI : Barry_Gross, gross@ccny.cuny.edu
 Aerosol Mode from Level 1.0 AOT; 16 APR 2006



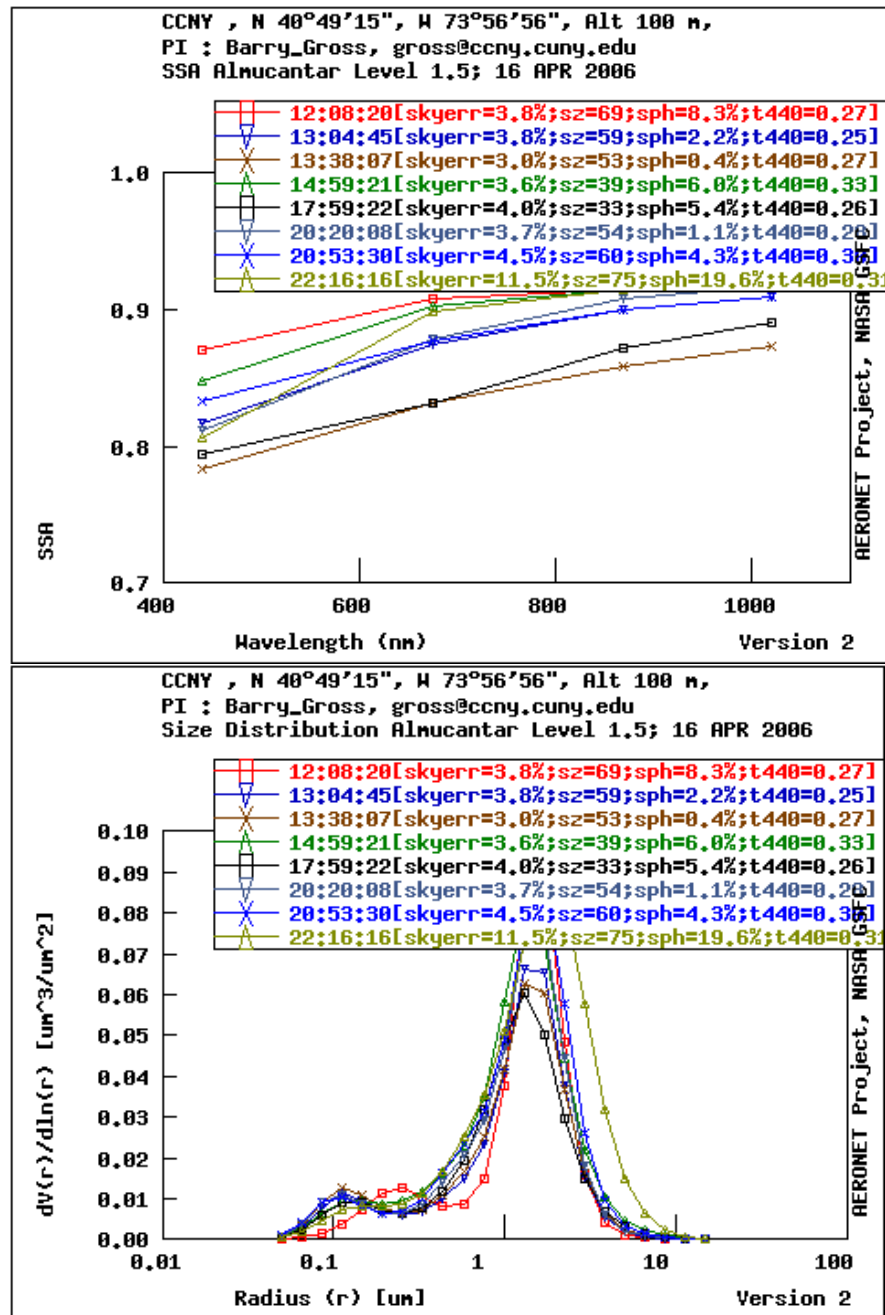


Figure 2.9 1) Data from CCNY CIMEL for April 16, 2006 obtained from the AERONET as one of the data products, AOD level 1.5. 2) Fine – Coarse Mode AOD 3) Multispectral SSA. 4) Aerosol Size Distribution

2.3.2 Multi-Filter Rotating Shadow-Band Radiometer (MFRSR)

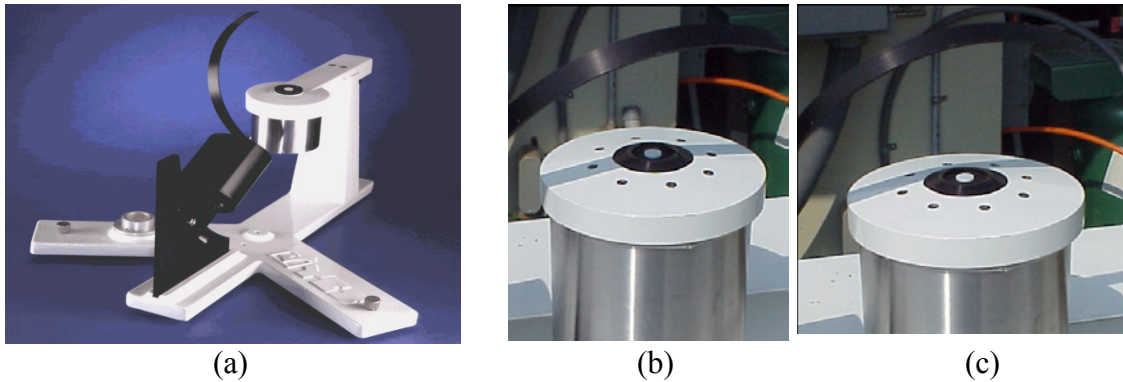


Figure 2.10 a. MFR-7 measures total, diffuse, and direct irradiance at six wavelengths (415, 500, 615, 673, 870, and 940 nm, each 10 nm FWHM) in the visible/NIR, and includes one unfiltered broad-band silicon pyranometer, manufactured by YESINC. (b) Middle measurement sun blocked. (c) 90° blocked to either side of the diffuser to get the other two measurements.

The Model MFR-7 is a microprocessor controlled data acquisition instrument designed to collect solar irradiance. The MFR was developed by a team of scientists: L. Harrison, J. Michalsky and J. Bernt in a collaborative effort between the Atmospheric Science Research Center and the State University of New York and Battelle's Pacific Northwest Laboratory. The MFRSR is a field instrument that concurrently makes instantaneous spectral measurements of horizontal solar irradiance at six nominal wavelengths 415, 500, 615, 673, 870, and 940 nm, of Global, diffuse, and direct normal components. Figure 2.11 displays the spectral band filters from the MFRSR diffuser head. In figure 2.12 displays the global (total), diffused (scattered) and direct for the 870 nm spectral channel. The direct normal solar radiation seems larger than the total due to the fact that it was divided by the airmass to get the direct-normal radiation. In addition it contains one unfiltered broadband silicon pyranometer whose spectral response is between .30 μm

to $1.10\ \mu\text{m}$. Since it uses the same detector to measure global and diffuse irradiance, the MFRSR is a versatile computer-controlled rotating shadow-band radiometer instrument. The MFR makes two physical measurements global and diffuse. There are two independent measurements used to calculate the direct-normal irradiance. Those two measurements are done by the curved plastic strip showing on figure 2.10a and b that rotates around the diffuser (hence the name shadowband). Before these measurements there is one at nadir, where the clip is at the bottom of the diffuser, which the unfiltered broadband silicon detector detects the total-horizontal irradiance. Then the band rotates and stops 9° degrees before and after the diffuser to take total sky measurements. There is a measurement that blocks the sun by shading the diffuser, figure 2.10b. The MFRSR makes its diffuse-normal by averaging the two side readings and the fully shaded to compensate for errors that the clip does not block the sun disks completely from the diffuser. It then subtracts the corrected diffused measurement from the total (global)-horizontal to obtain the direct-horizontal measurement. Finally, it divides the direct-horizontal by the cosine of the solar zenith angle (or the airmass) to obtain the direct-normal, figure 2.12.

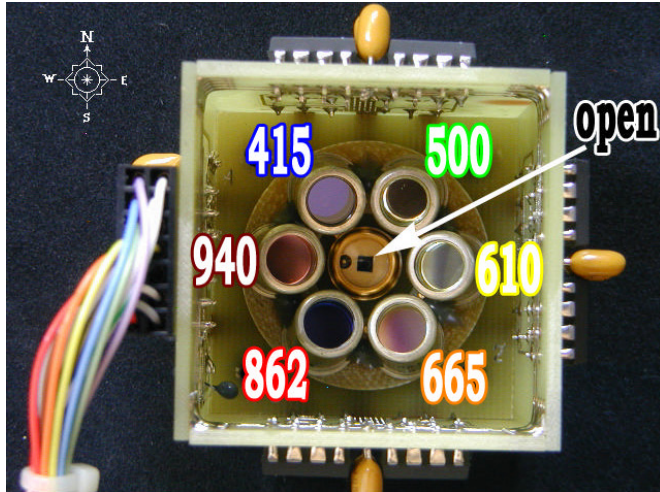


Figure 2.11 MFRSR Spectral Band filters, at the center is the silicon with a range of 415 – 940 nm.

The MFRSR Channels are classified as the following: channels 415 and 870 are mostly sensitive to aerosols, channels 500, 615, and 673 are sensitive to ozone, NO_2 and aerosols, and channel 940 is sensitive to water vapor, this channel is excluded from this dissertation. The MFRSR is an important instrument in climatologically research due to its automated operation and its robustness in constructing a network for spatial and temporal analysis of AOT, and it would be the main instrument.

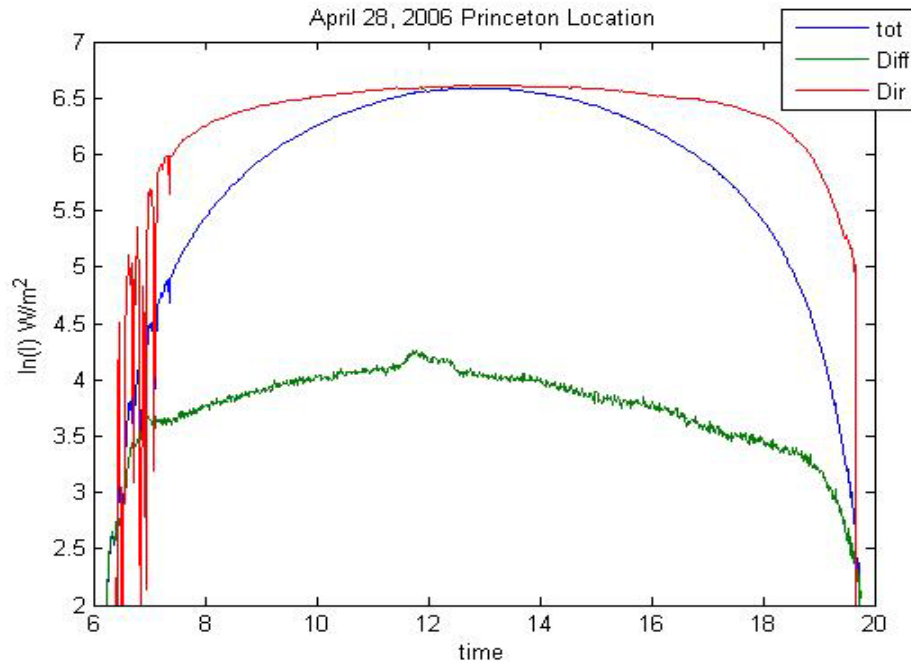


Figure 2.12 Daily solar plots from Princeton Site for April 28, 2006 using data from the 870 nm channel. Total/Global horizontal (blue curve), the diffused horizontal (Green curve) and the direct normal irradiance (red curve). The dips in the direct curve represents cloud contamination, even a simple thin cirrus cloud can create a problem.

2.4 Sun Photonmeter Retrieval Algorithms

2.4.1 Langley Regression Retrieval Algorithm

The Langley regression technique is a linear fit to the Beer-Lambert formula used to determine the solar constant and aerosol optical depth. The solar constant is the amount of incoming radiation at the top of the atmosphere over a unit area. Attenuation of the solar radiation depends linearly on the total amount of attenuators along any path, figure 2.13. The traditional way of perform Langley analysis is to assume that the optical

depth of the atmosphere does not vary significantly during the day. Such behavior is not typical for most sites, so averaging the solar constant for every spectral channel determined from Langley analysis must be performed over 20 to 40 clear days within about 3 months of data is a requirement for the most accurate solar constant [Harrison and Michalsky 1994]. The Beer-Lambert-Bouguer law, equation 1.3, for the atmosphere:

$$I_{\lambda}^g(x) = I_{\lambda}^0 e^{\tau_x m_n}, \text{ where } m_n \text{ is the air-mass which is close associated as } \frac{1}{\cos(\theta_z)}$$

, θ_z is the solar zenith angle; τ_x is the total optical depth including aerosols, ozone, and Rayleigh (molecules which must be subtracted from every channel before obtaining the aerosol optical thickness. Again, the Rayleigh contribution is obtained as:

$$\tau_R(\lambda) = 0.008569 \lambda^{-4} \left(1 + 0.0113 \lambda^{-2} + 0.00013 \lambda^{-4} \right) P P_0^{-1} \quad 2.4$$

Here, $I_{\lambda}^g(x)$ is the ground measured spectral radiance for any arbitrary path x; I_{λ}^0 is the extraterrestrial irradiance obtained by linear regression over clear cloud free once for the morning and/or once for the afternoon. The solar constant is also called the calibration constant which is an average of calibrations and is used to obtain Aerosol Optical Thickness for the rest of the data.

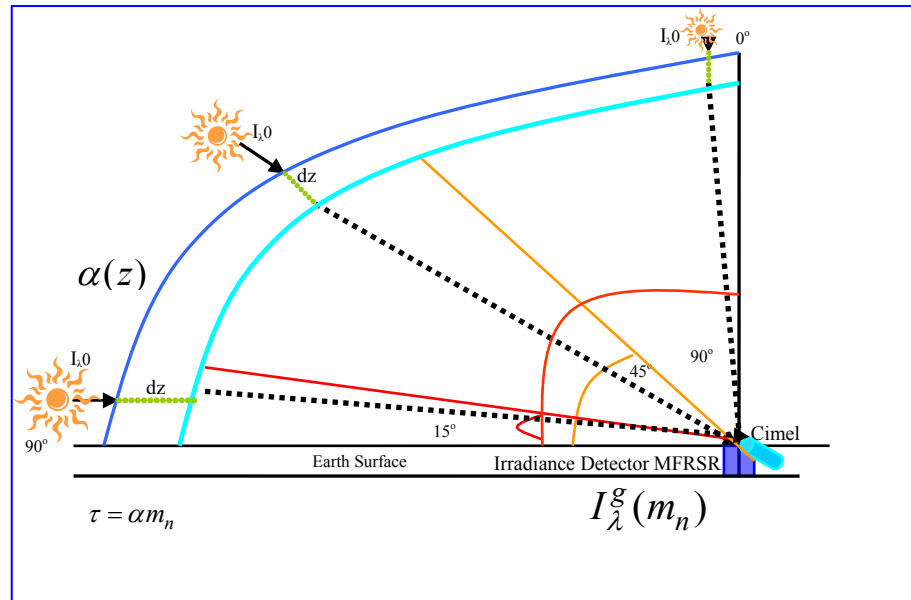


Figure 2.13 Different solar paths, at lower air-masses there are small changes in the planetary boundary layer while at larger air-masses changes are drastically. The best mass load is between 2.5 and 4.5 ($\theta_z \sim 61.5$ and 76.5 degrees).

Figure 2.13 follows the solar path and illustrates the change the solar azimuth angle. Most important is the assumption that the atmosphere does not change with respect to time during the Langley Regression.

We are able to extrapolate I_{λ}^0 from measured solar irradiance taken at several path lengths, limiting the airmass between $2.5 < m < 4.5$ for best measured results to reduce the problem to a linear regression, the logarithmic form

$\ln(I_\lambda^g) = -\tau_\lambda \times m_n + \ln(I_\lambda^0)$ is used. This representation is linear of the form:

$y = ax + b$, where the $b = \ln(I_{\lambda}^0)$ and $a = \tau_{\lambda}$ which is the negative optical depth is the slope of the line.

The best fit values for the coefficients $\ln(I_\lambda^0)$ and τ_x can be determined by applying a least-square regression under clear sky conditions. For the MFRSR data we perform this operation for morning and afternoon due to the non homogeneous atmosphere and average the solar calibration. Examples of the Langley regression are below.

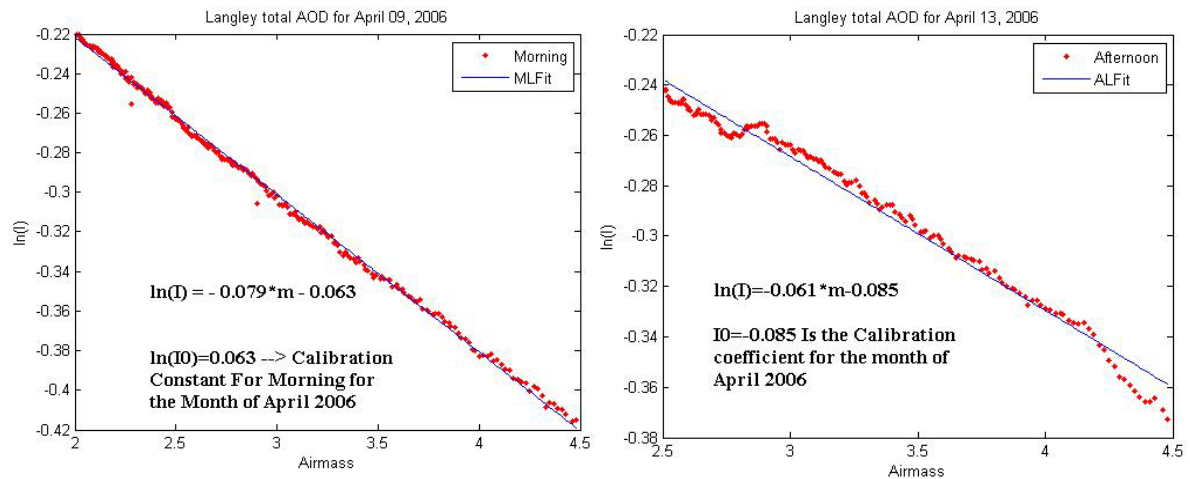


Figure 2.14 Results for the SGP MFRSR data using Langley regression yielding two calibration constants, one for the morning and the other for the afternoon. Data must be cloud cleaned and refit again to obtain best results over a longer period of time.

We note that the y intercept is not the same so an averaging of the solar constants over this period would be necessary.

2.4.2 NASA-GISS Retrieval Algorithm

The problem with the Langley regression approach is the need of constant AOD during the calibration period. The algorithm for implementing calibration coefficients using direct to diffuse ratios requires only that the “type of aerosol” be fixed during the

calibration period. Following Alexandrov, the atmospheric column extinction optical depth corresponding to the i^{th} channel can be expressed as:

$$\tau_i = -\ln\left(\frac{I_i}{I_i^o}\right) \times \mu - c_i \mu \quad 2.5a$$

$$\tilde{\tau}_i = \tau_i + c_i \mu \quad 2.5b$$

where μ is the inverse airmass (which is associated to $\cos(\text{zenith Angle})$), $\frac{I_i}{I_i^o}$ are solar radiation intensities, I_i is the ground measured irradiance, I_0 is the corresponding to the top of the atmosphere, $c_i = -\ln(C_i)$ are the Calibration coefficients and $\tilde{\tau}_i$ is for the un-calibrated optical depth [Alexandrov, et.al 2002].

The general idea in the new scheme is that instead of treating the optical depth as a constant to be determined through regression, an independent optical depth measurement based on the ratio of diffuse/direct transmission at the bottom of the atmosphere can be obtained which does not require calibration.

$$\Phi(\tau) \equiv \frac{L_{diff}(\tau)}{L_{dir}(\tau)} = \Phi_{meas} \rightarrow \tau_{diff} \quad 2.6$$

In later papers, an effort to include the surface albedo was also done but since the affects of the surface albedo seem to result in small and constant shifts in the optical depth, they can be absorbed into the so called opacity factor (D) which describes the observations that the diffuse/direct derived optical depth is different than the true optical

depth by some constant which seems to be stable over the duration of the regression analysis.

Lumping these terms together into the opacity factor

$$\tau_{dir} = \tau_{diff} + D \quad 2.7$$

$$\tau_{870}^{diff} \approx -\mu[\ln(I_g) + C_{870}] + D$$

Combining equations 2.5 and 2.7 we find D and C_{870} by regression

$$\tau_{dir} = \tau_{diff} + c \cdot \mu + D \quad 2.8$$

The direct to diffuse approach can be underlined by the following steps:

1. First, assume a reasonable urban atmosphere phase function and generate from the MFRSR diffuse/direct measurements the diffuse optical depth as function of time (solar angle). Assume no surface albedo.
2. Regression calculates D and C_{870} which can then be fed back into Eq. 2.7 to get the time dependant optical depth τ_{870}^{dir} .
- 3.

$$\tilde{\tau}_1 = q_1 \tau_a + \beta_1 x_{NO2} + c_1 \mu \quad 2.9a$$

$$\tilde{\tau}_2 = q_2 \tau_a + \beta_2 x_{NO2} + \gamma_2 x_{O3} + c_2 \mu \quad 2.9b$$

$$\tilde{\tau}_3 = q_3 \tau_a + \beta_3 x_{NO2} + \gamma_3 x_{O3} + c_3 \mu \quad 2.9c$$

$$\tilde{\tau}_4 = q_4 \tau_a + \beta_4 x_{NO2} + \gamma_4 x_{O3} + c_4 \mu \quad 2.9d$$

$$\tilde{\tau}_5 = \tau_a + c_5 \mu \quad 2.10$$

$\tilde{\tau}_i$ is the measured optical depth in the i^{th} channel $q_i = Q_{\text{ext}}^i / Q_{\text{ext}}^5$ is the Mie scattering extinction coefficients ratios normalized to the 870 channel corresponding β_i and γ_i are the effective spectral absorption coefficients of NO₂ and O₃ weighted by the Solar flux and filter response function of each respective amount on NO₂ and O₃ [Alexandrov, 2002, 2005]. The two unknown gas concentrations can actually be eliminated and equation 2.10 can be used (since we solved it earlier) to obtain the reduced set.

$$x = \frac{\tilde{\tau}_5}{\mu}; b_{ij} = \frac{\beta_i}{\beta_j}; g_{ij} = \frac{\gamma_i}{\gamma_j}$$

$$F_i = B_i(x - c_5) + A_i \quad i = 3, 4 \text{ where}$$

$$F_i = \mu^{-1} [\tilde{\tau}_i - b_{i1}\tilde{\tau}_1 - g_{i2}(\tilde{\tau}_2 - b_{21}\tilde{\tau}_1)] \quad 2.11a$$

$$B_i = q_i - b_{i1}q_1 - g_{i2}(q_2 - b_{21}q_1) \quad 2.11b$$

$$A_i = c_i - b_{i1}c_1 - g_{i2}(c_2 - b_{21}c_1) \quad 2.11c$$

Using $i = 3, 4$ and $b_{ij} = \beta_i / \beta_j$ and $g_{ij} = \gamma_i / \gamma_j$ are the weighted spectral coefficient ratios.

3. Given c_5 and using the fact that F_i can be calculated from the raw uncalibrated data for each time, we can obtain the value of B_i as the regressed extrapolation $F_i(x - c_5)$ for $i=3$ and 4.

4. Since B_3 and B_4 are now known, we have two relations which are used to fix two size distribution parameters in the aerosol distribution. In the 2002 Alexandrov paper, the aerosol model was monomodal and the 2 parameters were the $\mathbf{r}_{\text{eff}}$ and $\mathbf{v}_{\text{eff}}$ but more recently, a bimodal distribution was used as the core model and the 2 parameters which

are to be retrieved are the fine mode effective radius and the ratio of coarse to fine mode volume.

5. Once the parameters are retrieved (in a LSQ sense), the spectral extinction coefficients q_i $i = 1, 2$ for the short wavelength channels can be obtained directly from Mie theory and thus using equations 2.9, all calibrations can also be determined.

The outputs of this algorithm therefore provide a time series of aerosol optical depth and effective particle size as well as calibration coefficients of the instrument. The benefits of this approach are:

1. This method allows us to have independent calibration errors; one calibration error does not affect the accuracy of the other channels.
2. This approach is more robust than the traditional Langley method and it is applicable to wider range of atmospheric conditions. In particular, Langley approach relies on optical depth stability [Alexandrov, 2004, 2005] while the NASA GISS algorithm depends on fixed values of aerosol type (not amount).

3 Validations of Measurements

3.1 Validation of Measurements (SGP Site)

We repeat for our own understanding the validation procedures based on simultaneous comparisons of MFRSR retrievals. Comparisons were taken for January 2006 to Dec 15, 2006 at the Atmospheric Radiation Measurement (ARM) Program's Southern Great Plains Site (SGP; <http://www.arm.gov/docs/sites/sgp>) [Ackerman, T. P., and G. Stokes, 2003].

The ARM Site contains rich well calibrated data sets for various devices including the Multi-Filter Rotating Shadow-Band Radiometer and at this particular site there is a corresponding CIMEL device to correlate data. The site is located at the South Great Plains SGP Central Facility, Lamont, OK 36° 36' 18.0" N, 97° 29' 6.0" W with an altitude of 320 meters of sea level, it was the first site to collect ground data for the DOE's.

Two MFRSR sites data sets for 2008 (C1 and E13) as identified in figure 3.1 were collected as well as CIMEL data at the C1 site. A good example of the MFRSR AOD retrieval in cases where the AOD can vary significantly is illustrated in figure 3.2. These results on the 500 channel indicate that the optical depth can be accurately obtained in comparison to the CIMEL instrument while increasing the number of data points.

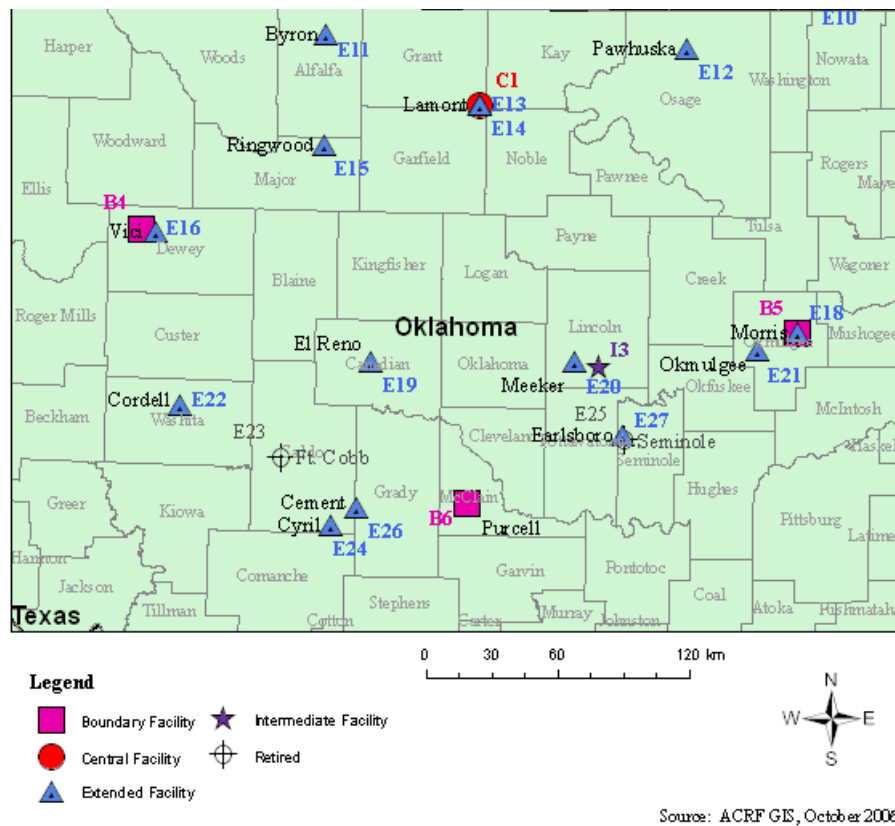


Figure 3.1 South Great Plains SGP Facility includes Central (C1) and Extended Facility (E13)

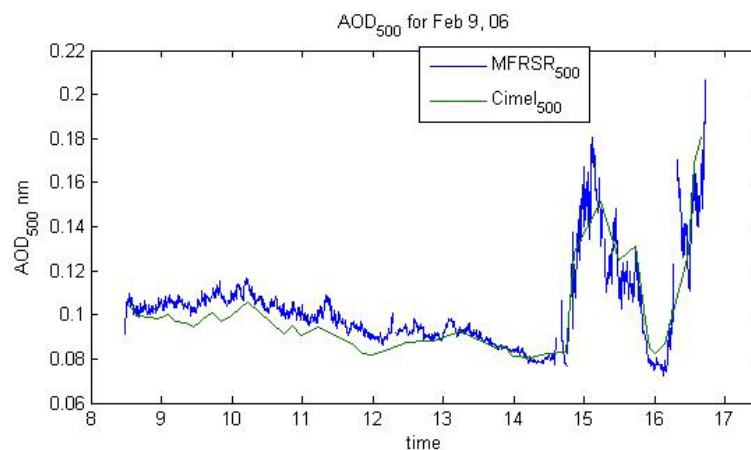


Figure 3.2 CIMEL vs MFRSR AOD retrieval (500nm) at C1.

In section 2, we emphasized the need for stable optical depth using the direct to diffuse ratios (processed aerosols from the direct to diffuse ration) which are critical to the processing of the aerosols. The level of improvement can be seen from statistical analysis of the ARM CIMEL data which shows that the normalized optical depth variability figure 3.3 is approximately three times larger than the optical depth ratio variability.

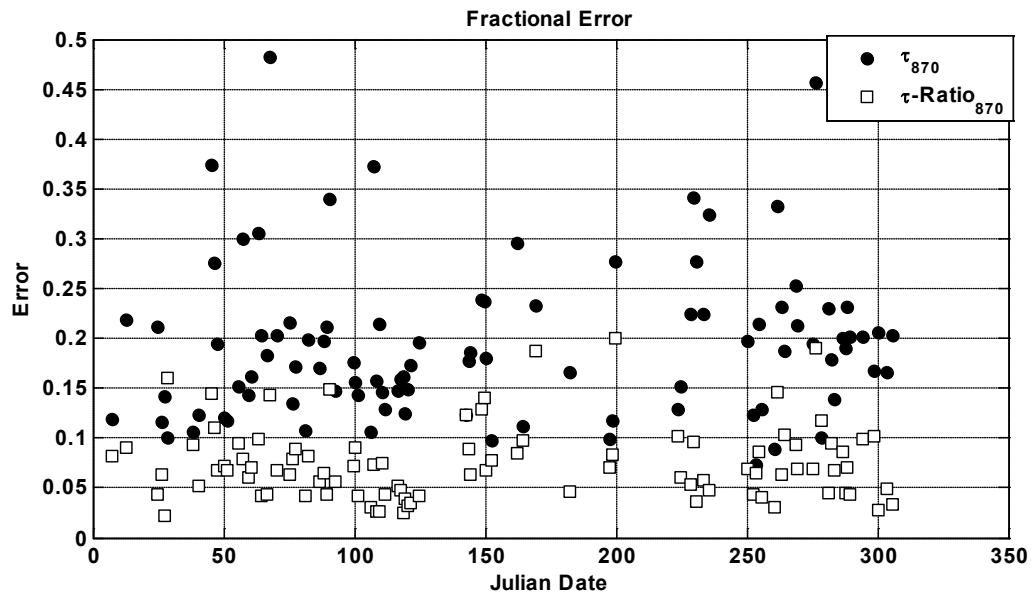


Figure 3.3 calculating the fractional error between AOD at 870 nm channel vs. the Diffuse to Direct clearly shows an improvement of AOD.

While we have seen that the MFRSR retrieval is often in good agreement with the AERONET retrievals, this is not always the case. In figure 3.4, the NASA-GISS algorithm comparison with CIMEL shows a strong correlation of 93.9% for April 18. However, on April 9, significant deviations were seen in the matchup. This may be attributed to small AOD loading. In addition, plotting the angstrom coefficient from CIMEL in figure 3.5, we see a significant variability in the aerosol properties which implies that the assumptions used in calibrating the MFRSR AOD were not well met.

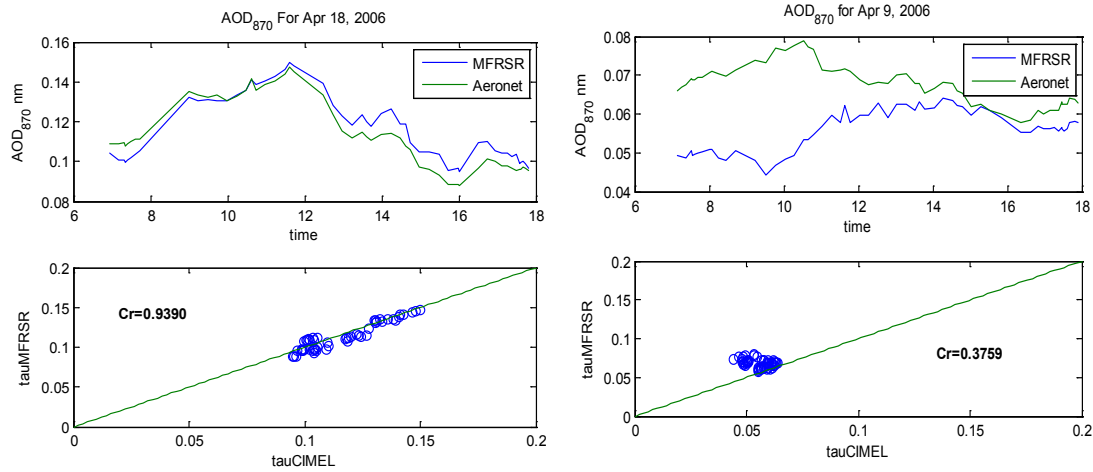


Figure 3.4 Comparison of the NASA algorithm with the CIMEL for 2 days.

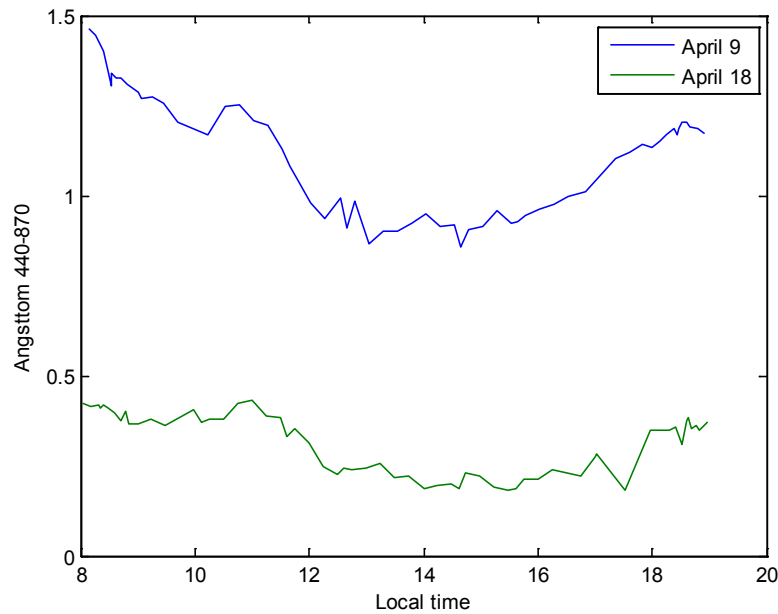


Figure 3.5 Angstrom Coefficients 440-870 for 2 cases illustrating metric defining aerosol type variability

In addition, we also checked the stability of the calibration coefficients. In figure 3.6, we plot the time series for both calibration methodologies and note the GISS calibration is

more stable throughout the year than the Langley analysis. In particular, we see a significant decrease in the calibration curve from May until October in comparison to that obtained from Langley regression.

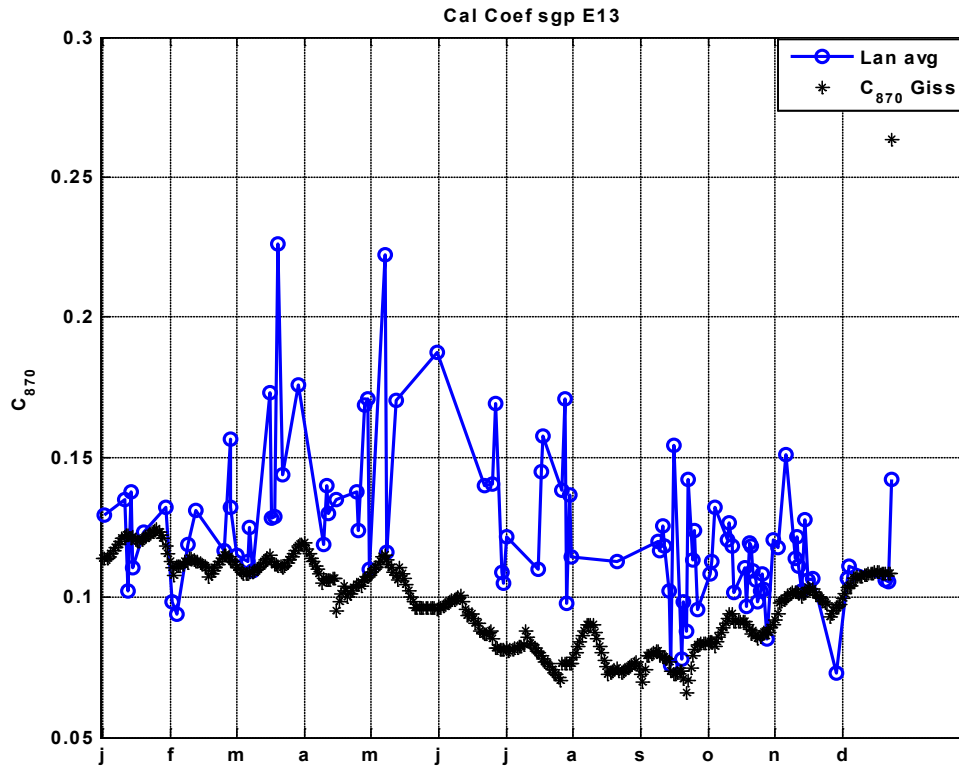


Figure 3.6 Calibration coefficients at 870 between GISS algorithm and Langley Regression for the year 2008 for the E13 SGP Cart Site Data set.

3.1.1 Validation of MFRSR Products (Fine – Coarse Mode)

On the other hand, the lack of spectral coverage of the MFRSR data means that the MFRSR should be less capable in pulling out coarse mode aerosols. Therefore, it is useful to compare the fine and coarse retrievals. Because of the lack of information channels, MFRSR calculations are based on fixed models of fine and coarse modes. In

figure 3.7, we find Aerosol optical thickness comparison (AOT) of the fine and coarse mode. Whereas the fine mode is much more accurately retrieved by MFRSR processing than there are significant errors in the coarse mode retrieval. In particular, the error seems to occur when the fine / coarse ratio undergoes significant dynamic change. This is reasonable since the original processing relies on a homogeneous aerosol model type.

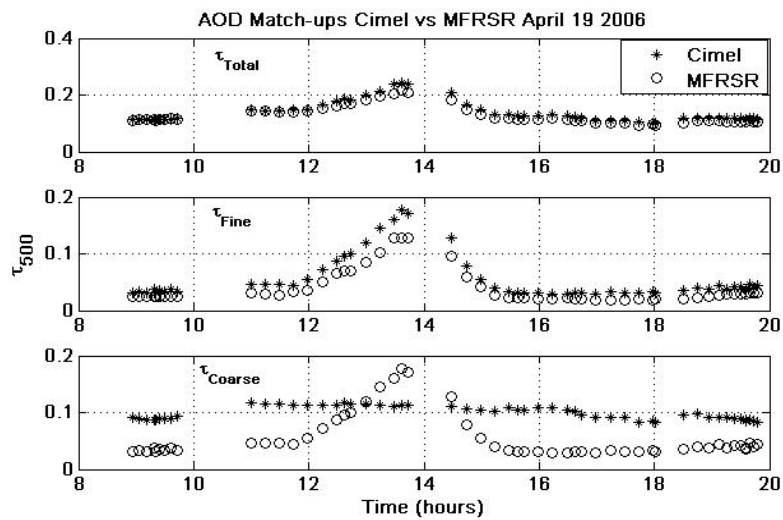


Figure 3.7 Time series comparison of AOT's Aeronet between CIMEL and MFRSR for Total top graph, Fine middle graph and coarse bottom graph.

3.1.2 Statistical Match-Ups

In figure 3.8, the regression plots are given for total, fine and coarse mode AOD retrieval. Errors are compatible with the numbers given by Alexandrov [2008] with mean absolute errors on the order of .012. Note in particular the coarse mode has the largest relative error as can be expected from the MFRSR system.

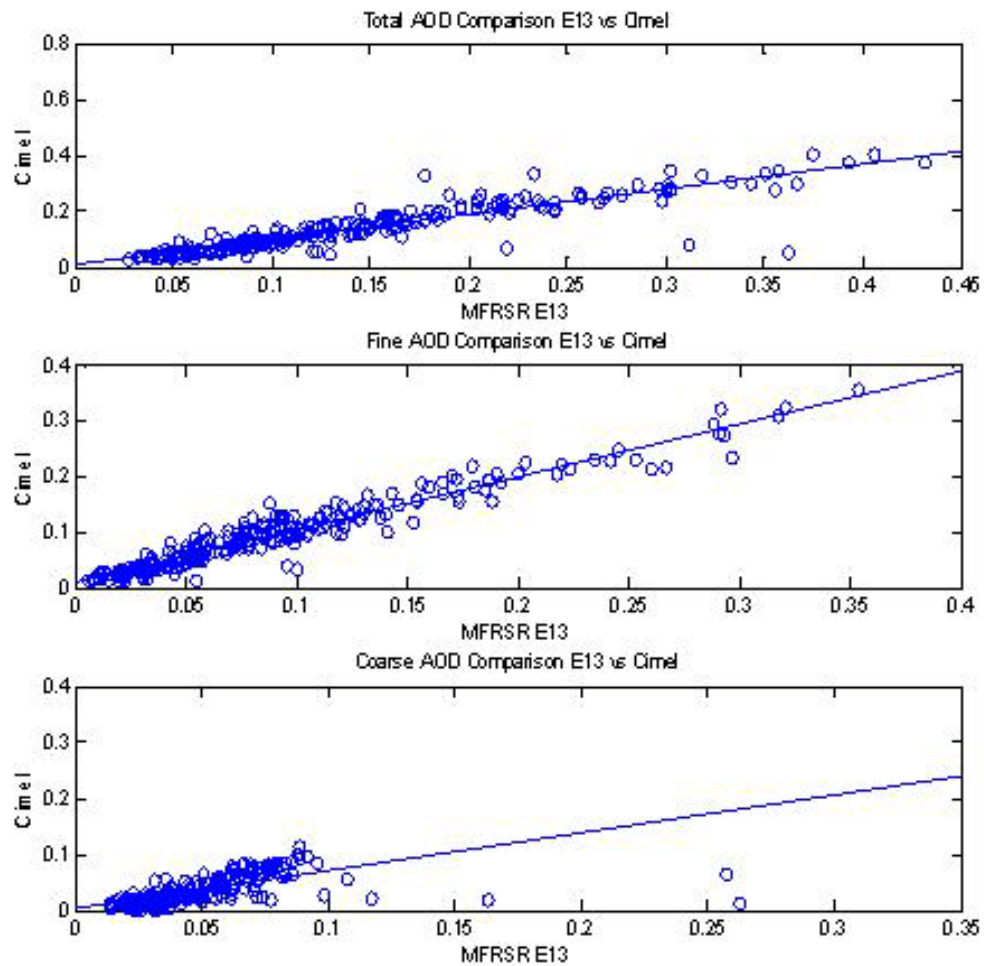


Figure 3.8 Annual Regression plots for a) Total, b) Fine and c) Coarse Mode.

3.1.3 Match-Up GISS Algorithm and Langley

To see if the Langley processing can match the NASA-GISS algorithm, we plot several days where the difference between the morning and afternoon Langley calibration coefficients were less than 2%. Under such conditions, we expect the two approaches to be quite consistent as illustrated in figure 3.9.

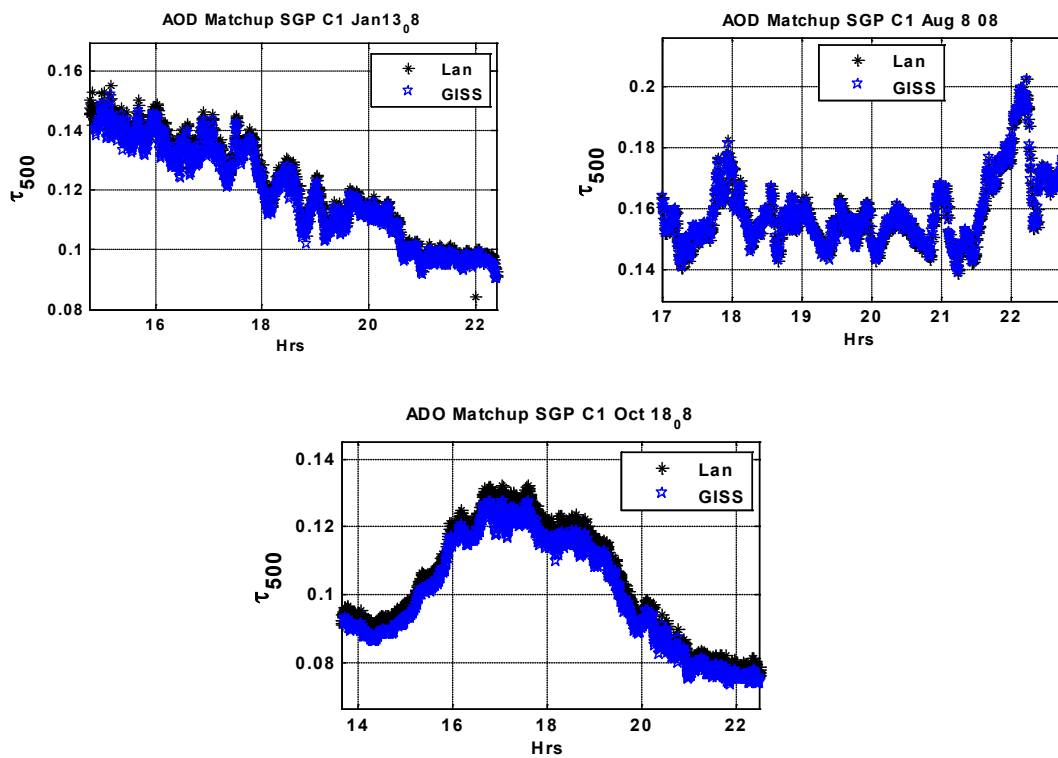


Figure 3.9 Results for the few of the best days that Langley and GISS were coincident in evaluating the Total AOD.

Calibrations from Langley are averaged and they do not make any big changes in the total AOD but the only noticeable difference is that the Langley Total AOD is most of the time lower than the one obtained by the GISS algorithm. GISS Algorithm is more robust

when it comes to obtain daily AOD's because its dependency on the color ratios does not make it vulnerable to changes in the inhomogeneous atmosphere.

4 MFRSR Network – North East Region US

4.1 Ground Base Network Instrumentations

4.1.1 AERONET

The AERONET program is a federation of ground-based remote sensing aerosol networks established by NASA and LOA-PHOTONS (<http://loaphotons.univ-lille1.fr>). The network has been expanded by collaborators from national agencies, institutes, universities, individual scientists, and partners (<http://aeronet.gsfc.nasa.gov>), figure 4.1a. The sun photometer network provides globally distributed near-real-time observations of aerosol spectral optical thickness and sky radiance as well as derived parameters such as particle size distributions, single-scattering albedo and complex refractive index. The standardized network procedures [Holben et al. 1998; Smirnov et al. 2000] of instrument maintenance, calibration, cloud screening, and data processing allow for quantitative comparison of the aerosol data obtained in different times and locations. The network provides a long-term, continuous and readily accessible public domain database of aerosol optical, microphysical and radiative properties for aerosol research and characterization and validation of satellite retrievals.

However, the network instruments are expensive and need to be located in very controlled environments. This makes the density of instruments quite sparse and quite impractical when exploring urban questions.

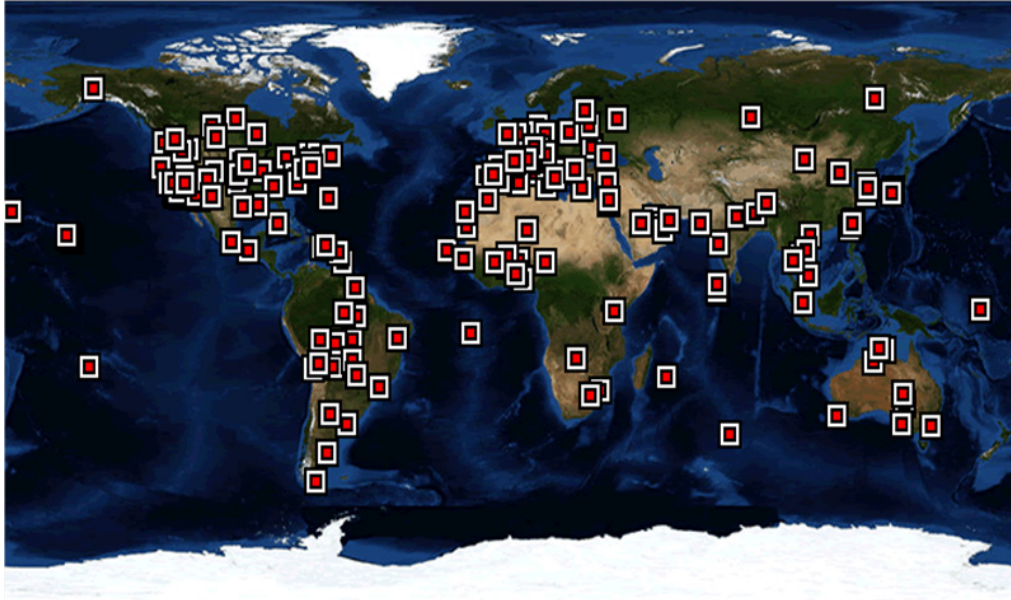


Figure 4.1a AERONET Global Network Distribution with at least 2 years of data.

In particular, when looking at the very important North East Sector, the number of sites is extremely small as seen in Figure 4.1b



Figure 4.1b AERONET Regional Distribution with at least 2 years of data.

4.1.2 Multi-Filter Rotating Shadow-Band (MFRSR) Radiometer Network

The North Eastern region of the US MFRSR network began during the fall of 2006 with the help of NASA-GISS and other collaborators as shown in table 4.1. The goal of the network is to monitor atmospheric aerosols and to create a spatial and temporal database of aerosol optical depth including fine and coarse mode. The continuation of the development of the network will improve the AOD data products for the Northeast region of the USA and provide a unique test-bed for satellite validation data that will be available as NetCDF format or raw ASCII format for further processing. Currently we are in the process of adding more devices along with contributors across this region. The full network plan will include two concentric rings. The inner ring will probe the New York Metropolitan area while the outer ring will look at transport between these two rings for a better determination of aerosol sources.

Site	Location	Latitude	Longitude
CCNY	The CCNY of NY	40.834 N.	73.946 W.
PTZ	Princeton U. NJ	40.349 N.	74.650 W.
LGCC	LIC, Queens, NY	40.444 N.	73.564 W.
LAMONT	Lamont-Doherty Observatory, NY	41.010 N.	73.910 W.
MECC	Medgar Evers C. NY	40.000 N.	73.000 W.

Table 4.1 Represents the actual MFRSR Network and data is been collecting.

OUTER RING					
Instrument	Site	Location	Distance(km)	Lat	Lon
MFRSR, LIDAR	PTZ	PRINCETON	21.05	N 40.34914	W 74.65022
MFRSR	SCSU	South Connecticut U.	3.27	N 41.330976	W 72.94681
MFRSR	BNL	Brook Heaven Labs	55.61	N 40.86274	W 72.8876
MFRSR	SNP	SUNY New Platz U	21.47	N 41.738313	W 74.09077
MFRSR, LIDAR	DEL	Delaware Gap	6.40	N 40.981649	W 75.14107
MFRSR(MECC)	PUR	Purchase College	14.69	N41.04	W73.70
SeaPRISM	PLAT	Costal Platform	12.56	N40.96	W73.34

Table 4.2 represents the future location and distance to Glory satellite tracks also instruments implementation for the year 2010 to be deployed.

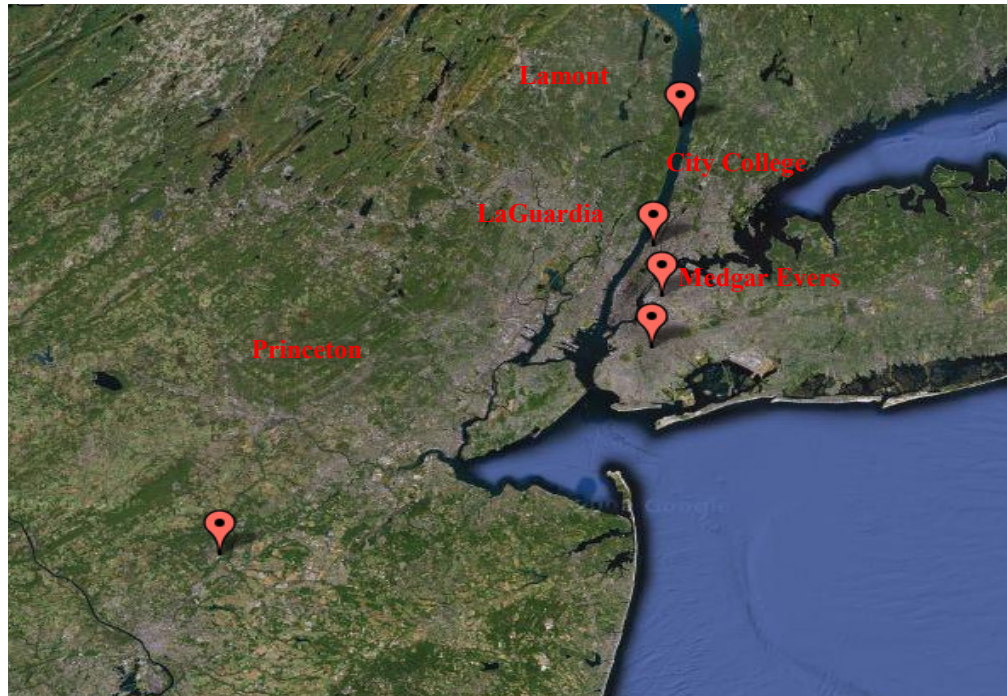


Figure 4.2 Google Earth images of current Northeast Inner Circle MFRSR network sites used for the urban studies of AOD transport.

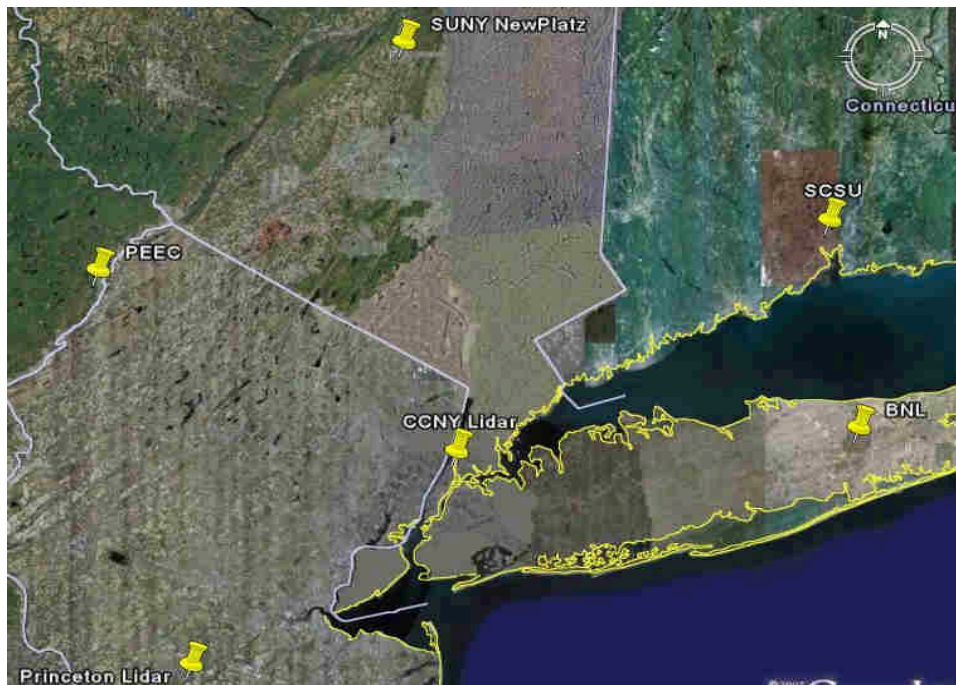


Figure 4.3 Google Earth image of current Northeast Outer Circle MFRSR network sites used for the rural studies of AOD transport it is being implemented.

Unfortunately, progress has been slowed by difficulties in instrument repair.

Historically, the MFRSR radiometer has proven to be a reliable portable and relatively inexpensive solution. However, until recently, the MFRSR was in general used only to measure multispectral optical depth extinction from the direct horizontal solar beam, and the diffuse horizontal beam was simply used as a sky correction parameter. With the implementation of new algorithms which make use of both the direct and diffuse horizontal solar beams are developed and have increased the retrieval potential beyond simple optical depth. The MFRSR network goal is to contain these model upgrades into single robust retrieval AOD properties from the CCNY-NASA-GISS network. The goal of this network will be the automated retrieval of total, fine and coarse mode AOD, effective fine mode radius and can possible be used to retrieve Single

Scattering Albedo (SSA). Some of these upgrades are already being used with the existing MFRSR instruments with the help of our collaborator Dr. Mikhail Alexandrov (Columbia University and NASA-GISS).

Alexandrov et al [2005, 2007] has demonstrated the ability of MFRSR's AOD to separate fine-coarse mode AOD by using gamma distributions. In particular, based on matchups made at the DOE Arms site with AERONET, standard deviation uncertainties in aerosol optical depth in all cases are less than .01 for both fine and coarse mode while a systemic over bias in the MFRSR retrievals of .01 is seen in the coarse mode retrievals. Most impressive is the fact that no significant bias is seen in the fine mode retrieval. This excellent performance is clearly of value in validation of GOES, MODIS and the new APS instrument which will be launched on GLORY in 2011.

In figure 4.4, we again illustrate the performance of the GISS algorithm in retrieving total and fine mode aerosol between the CCNY CIMEL and the coincident CCNY MFRSR. In particular, we observe a good performance between the MFRSR (solid circle) and the AERONET retrievals (open circles) through a single day from the CCNY site.

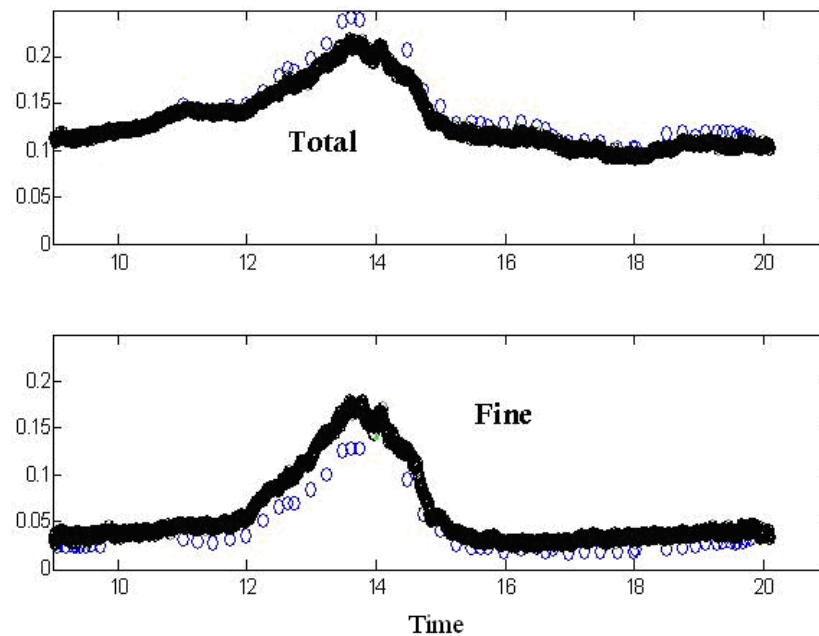


Figure 4.4 Intercomparison of MFRSR and AERONET AOD retrievals as function of time at 500nm.

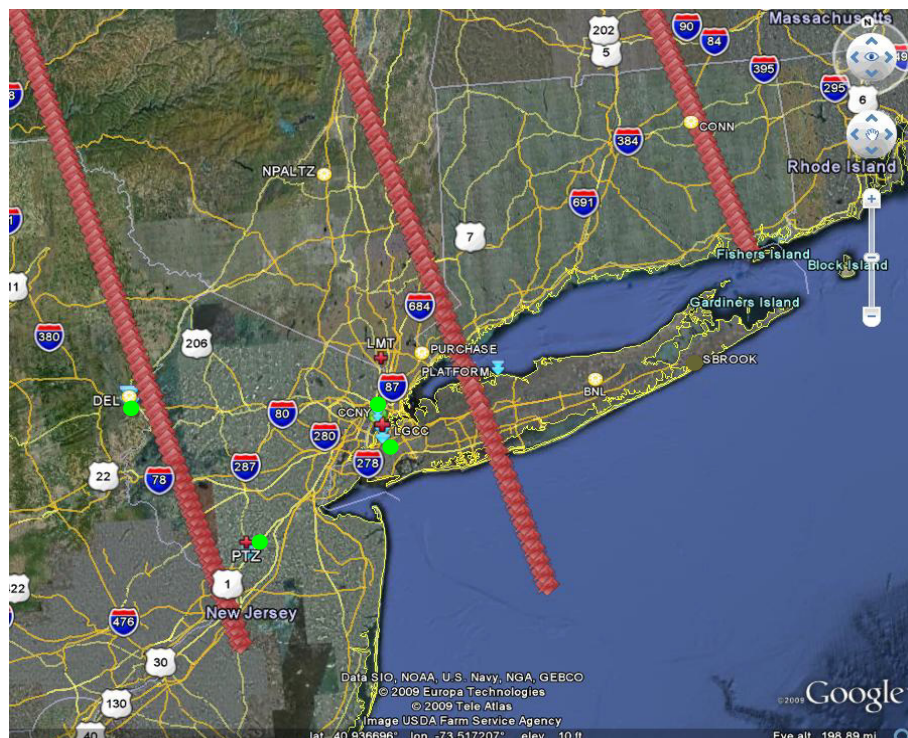


Figure 4.5 Existing (red) and proposed MFRSR deployment relative to APS ground tracks. Red crosses indicate active MFRSR sites, Yellow squares indicate MFRSR In deployment in near future, green circles are LIDARs and blue triangles are current and future CIMEL's.

When full deployment is made, the MFRSR network together with contributing Lidar assets will make the NYC region critical as a satellite validation test bed in an area where retrieval of aerosols is hard to retrieve (see section 2.1). Useful statistics on the spatial distribution of optical parameters will also be collected for all deployed MFRSR instruments as illustrated in figure 4.5 and in detail in table 4.2. Most important for APS validation are the near coincident locations of the future network implementation MFRSR at Southern Connecticut State University (SCSU) and the Delaware Water Gap (DEL) with distances $< 6\text{km}$, as well as SUNY Purchase and an AERONET (Sea-Prism) under deployment on a platform in the Long Island Sound, which will be operational as part of the coastal water observing activities. The latter two ground sites are within 15 km from GLORY ground track. Currently, five MFRSR sites are operational while the rest of the deployments will be accomplished in 2010.

4.2 Network Devices

The first steps in creating a MFRSR network is intercommunication between sites to the main branch located at the CCNY. The first MFRSR was installed at the City College of New York. It was located at 40.834 North and 73.946 West NYC, (Top of the Steinman Hall Building located at 140 Street and Convent Ave.) The MFRSR are not network ready, meaning that the only way to extract the collected data from the memory was to use a MODEM (Modulating/Demodulating telephone device) directly connected to a computer using a NULL Cable MODEM DB9 (9 pins connector device) and interlaced it with the Communication port of the computer. This was a very cumbersome way to

communicate with the MFRSR located two floors down, using approximately 100 ft of cable. To create the MFRSR network we need it to interlace the MFRSR box with a RS-232 to Ethernet Mini Serial Server see figure 4.6.

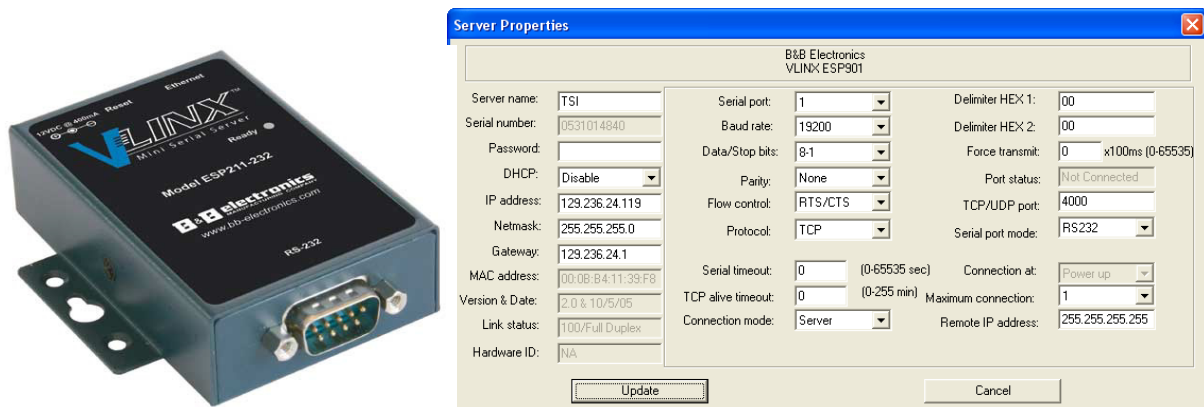


Figure 4.6 RS-232 to Ethernet mini serial server software driven simulates a virtual communication port on the computer collecting data over the network.

This RS-232 to Ethernet device was a key component to start the network data collection over the Local Area Network (LAN). The VLink connects directly over the network using TCP/IP (Transmission Control Protocol / Internet Protocol) base connectivity. The software that comes with VLink helps you to manage one or many devices connected in a similar way. There is a virtual (COM) communication port created by the VLink software, the COM port emulates the MODEM of the MFRSR in order to download the processed data.

4.2.1 Local Area Network Communication Protocol

We are using TCP/IP which is a routable protocol and can be accessed over long distance. There are two main networks components for delivering data between sites private and public addresses. Private and public IP addresses are classified into three classes see table 4.3 with different amount of devices (IP's) that can be attached. Table 4.3 displays only the private IP's that can be used within the network the rest are public IP addresses.

CLASS	SUBNET MASK	NETWORKS	DEVICES
A—10.0.0.0	255.0.0.0	126	16,777,214
B – 172.16-30.0.0	255.255.0.0	16,384	65,534
C – 192.168.0.0	255.255.255.0	2,097,152	254

Table 4.3 Contain the IP addresses for private use, the remaining are for public usage starting at 1.0.0.0 and ending at 224.255.255.254 and are owned by many institutions for the use and network connectivity.

Each IP class is made of eight bits per dot called an octet for example 00001110.0.0.0 this represents 10.0.0.0. Since each octet contains eight bits the count is from right to left using binary (2^n) for each bit position where $n = 0, 1, 2, \dots, 7$ for a maximum value of 255 per octet. The subnet mask differentiates between local and remote network connectivity. For the MFRSR network we are using class C (192.168.1.100 or 192.168.0.100) IP address to communicate between local computer and the MFRSR Box. At the sites listed on table 4.1 are implemented into the site (North East US) NEUS network. There are sites that require strong communication than others. It all depends on the college and the

network administrators to provide us with a solution. The only requirements that we entail are: the need of two IP's addresses that will not compromise their local network, in other words a demilitarized DMZ Zone IP's that we will have full control from CCNY Remote sensing Lab. In order to administer every site devices, either the computer or the data-logger instruments we need to remote control them from CCNY. We can either use Microsoft remote desktop connectivity or use third party free software like (Ultra VNC) Virtual Network Connection, which allows computer management tools from remote locations. When our requirement cannot be met by the accommodating college, we suggest VPN and FTP from CCNY to their location. For instance, LGCC only allows Virtual Private Network (VPN) connectivity and they use Cisco Web enable VPN, figure 4.10. VPN uses a tunnel across the internet to connect over networks and be allowed into their LAN's to manage our equipment with the help of VNC to control MFRSR desktop connections, figure 4.9. VNC can be accessed from any network to an IP that is visible on public network.

Configuration of a File Transfer Protocol FTP is crucial for the transport of data from a remote site into the main site at CCNY. We had installed FTP in each machine and created several script files (Files with commands executed line by line) to extract the following files: daily collection of the (.mtm) morning to morning file, and at the end of the day transfer the daily collected data that is in the .mtm or in a .txt file. We use Scheduler software to execute the scripts and allow us to automate the daily collection of data. Collected data is then published on the web <http://sky.ccny.cuny.edu/wc/Shadowband/index.php#s2> where data can be safely

downloaded refer to figure 4.7. Figure 4.8 is the YESDAS inc processing of .mtm files and can be seen daily as data is being collected the site can be accessed by: <http://134.74.16.39>.

WEATHER CENTER RAW SHADOWBAN DATA ARCHIVE

The City College of New York

138th Street & Convent Avenue, New York, NY 10031

These data sets were obtained from the MFR-7 Shadowband instruments (Yankee Environmental Systems, Inc.) located at several remote sites. Calibration corrections have been applied and differ from the same day's data that can be downloaded through the 'Current Shadowband Data' link, on the CCNY-YESDAS Web server, only in the file format. The format presented here contain the following ASCII format, data is angle corrected and calibrated further processing of data is required. Data files contains the following structure {col1 time, col2 sec (theta), Total-horizontal [nominally 415, 500, 615, 673, 870, and 940 nm], Diffuse-horizontal [415, 500, 615, 673, 870, 940], Direct-normal [415, 500, 615, 673, 870, 940] only daytime otherwise *4 for night data is assigned. These data are presented here for your ease of access and use. Please note that the filters (sol and cal files) for each instrument have slightly different spectral characteristics unique to each site as detailed in the top of each download file.

Click on the desired link shown below to view the data, or alternatively, right-click and select 'save as' to save the file to your local machine. The files are in ASCII tab delimited format.

Directory Contents for The City College of New York (CCNY)

<u>File</u>	<u>Data Date</u>	<u>Size</u>
CCNYDay20020926.txt	Sep 26, 2002	4kb
CCNYDay20020927.txt	Sep 27, 2002	4kb

4.7 Raw data that can be downloaded from the above stated hyperlink.

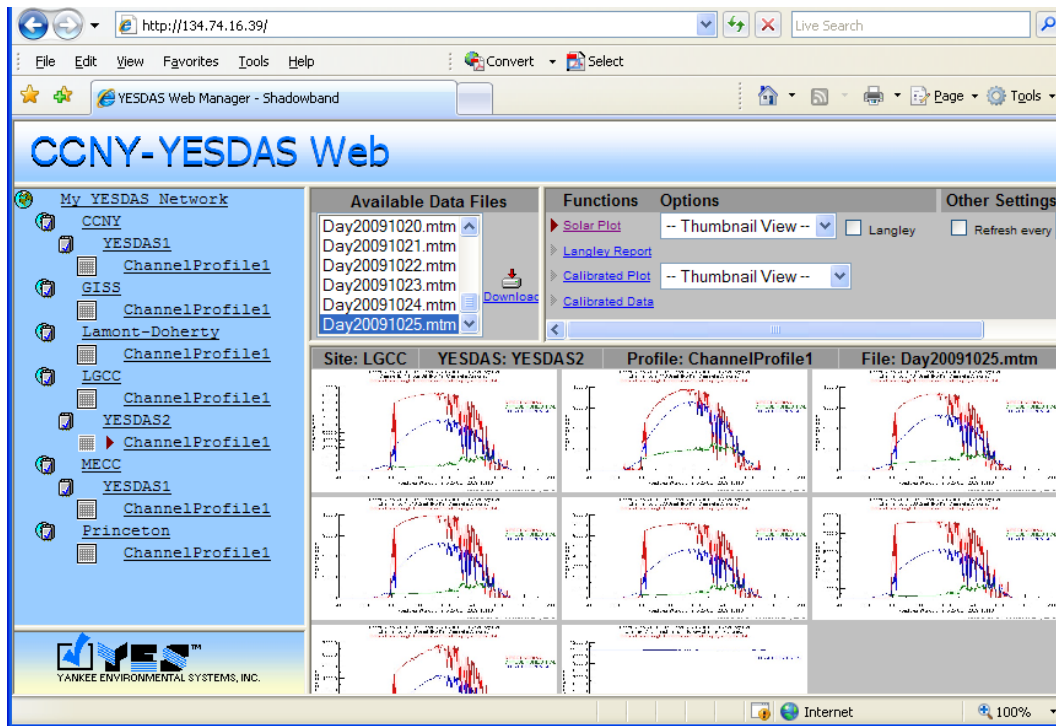


Figure 4.8 this is a separate real-time web browser to safely download data from each site. Data can also be reviewed before been downloaded. This site will be among other site with AOD products obtain from the NASA-GISS algorithm.

Types of configuration and remote access refer to figures 4.9 and 4.10.

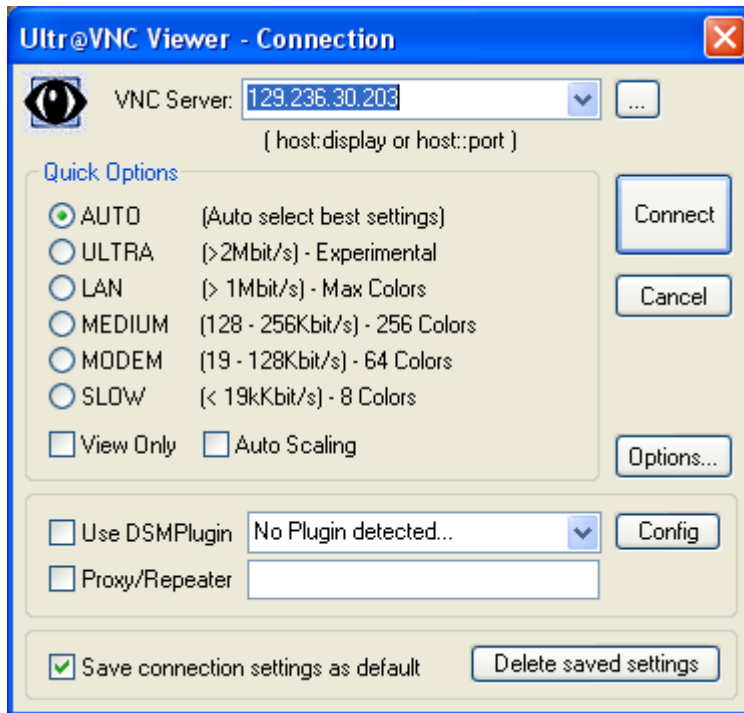


Figure 4.9 Network administrator uses VNC connectivity to any remote site for troubleshooting, maintenance and data transport.

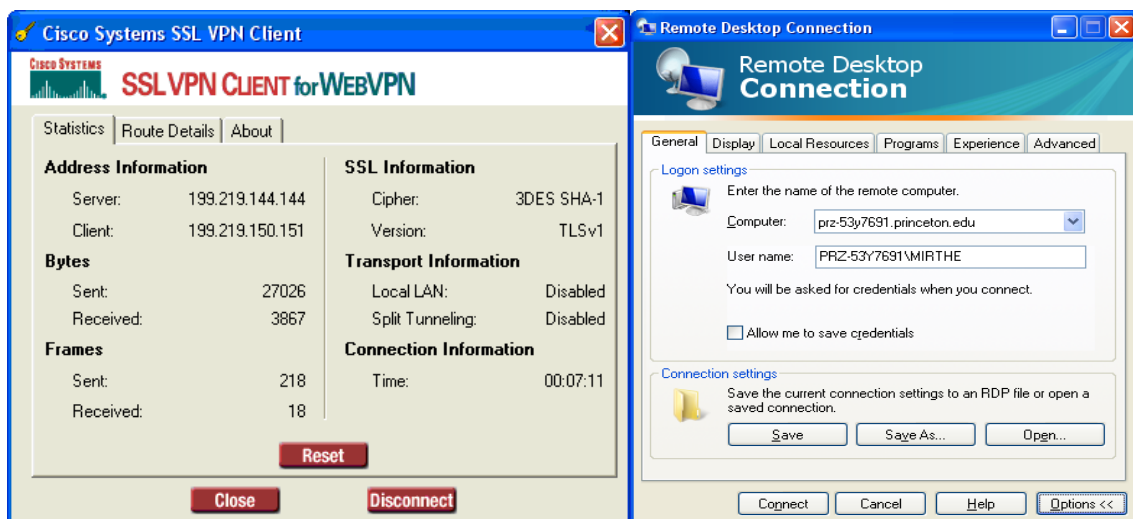


Figure 4.10 Ways to connect to remote site using VPN or remote desktop communication protocols.

4.3 Data Storage

Data is being collected locally and remotely for each site. At the end of the day a script will separate data files into a folder where data is sorted accordingly to file extension and later be collected by the FTP script to transfer all fully completed files to be archived at CCNY. However, fractions of the .mtm (morning to morning used by the data logger) file is being collected every 15 minutes to be displayed on the raw data website <http://134.74.16.36> and also at the main CCNY site located at: <http://sky.ccny.cuny.edu/wc/Shadowband/index.php#s2>. At the CCNY site data is also backed-up weekly into a backup storage device. Data is available as MFRSR Raw ASCII as well as NETCDF files (.ndf) These Raw or processed files using the NASA-GISS algorithm can be downloaded from: http://sky.ccny.cuny.edu/wc/map_Shadowband.html by clicking the appropriate link of choice see figure 4.11. We are currently working in a new interactive web browser to display AOD products processed by NASA-GISS algorithm. The first interactive web browser should be ready by June of 2010 and also pull down operational MODIS and GOES aerosol products.

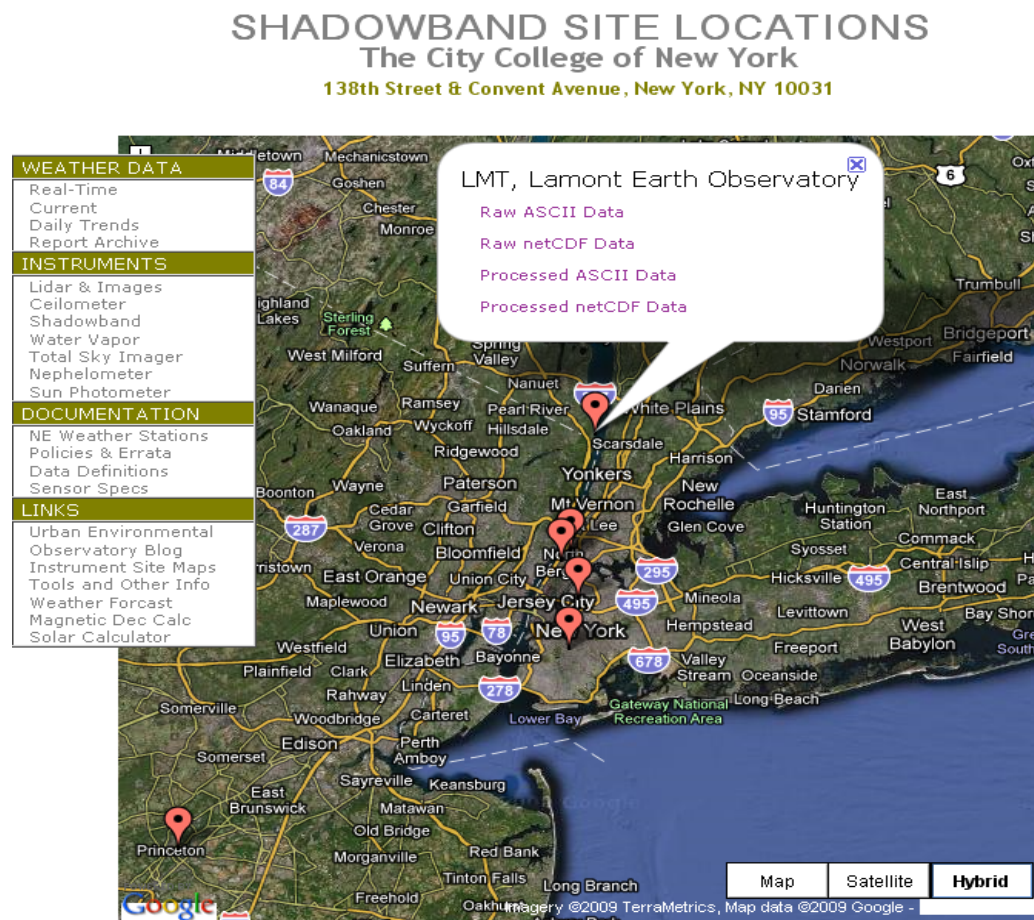


Figure 4.11 NEUS Sites to select type of data to be downloaded. User has a choice of data to select from Raw ASCII and netCDF or Processed data according to NASA-GISS algorithm Processed ASCII or/and netCDF files.

4.4 Applications of Network

We first look at Lamont and Princeton as two sites outside the urban canopy.

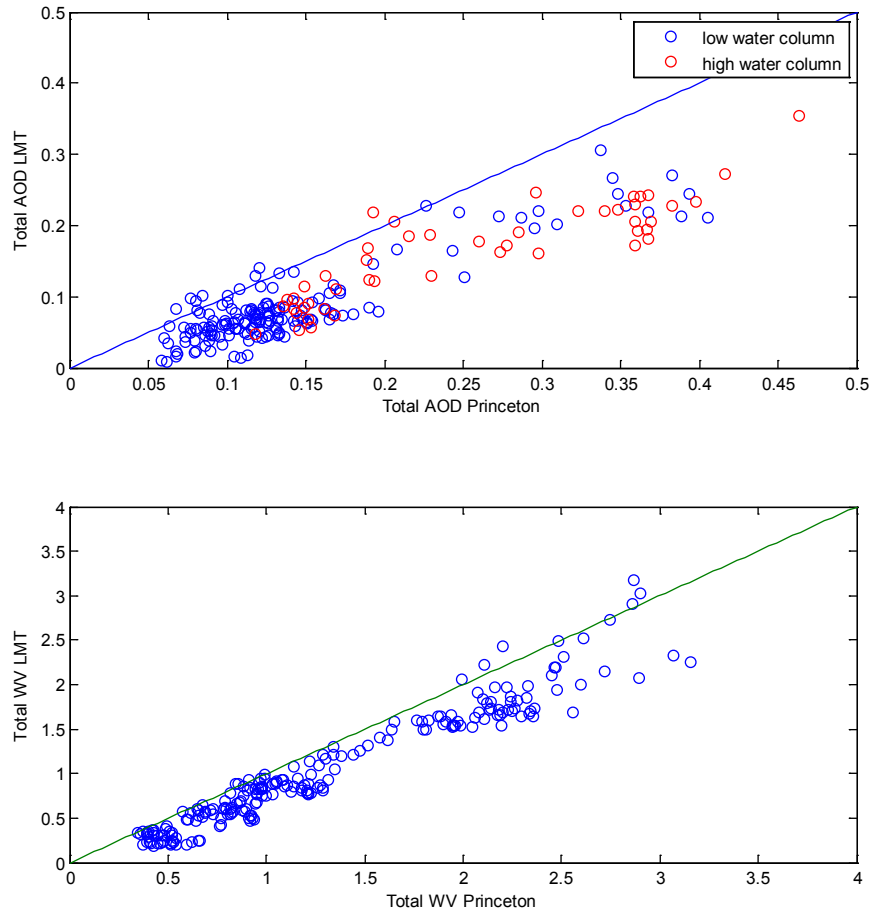


Figure 4.12 Comparisons of AOD and water vapor between suburban sites.

The main result we see from this comparison is the increase in statistical AOD that seems to be highly correlated to an increase in total water vapor. In addition, Princeton is also upwind from a number of local emission sources so it is not surprising that AOD is larger

at Princeton than Lamont. The situation becomes more interesting when considering an urban location such as Medgar Evers in comparison to Princeton.

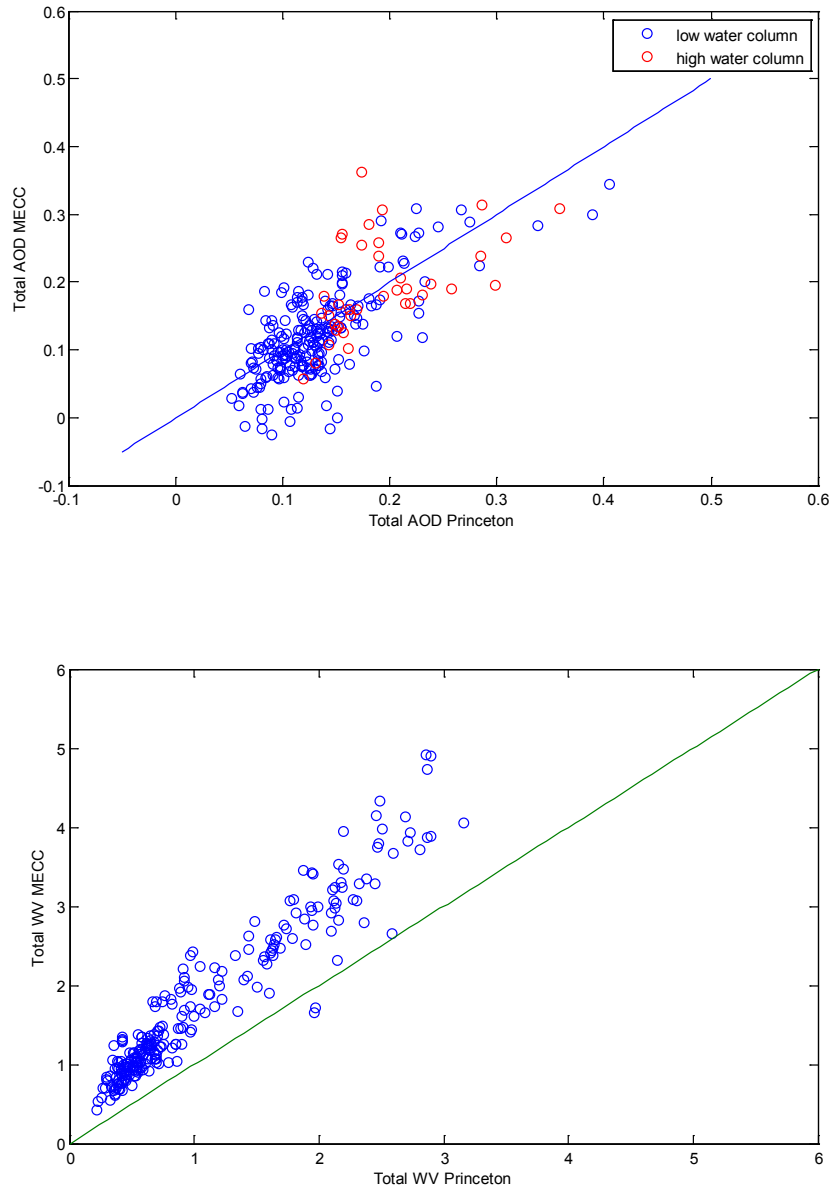


Fig. 4.13 Comparisons of AOD and water vapor between suburban site (PRN) and urban site MECC.

While the correlation in AOD is significant, it is less as expected due to the greater distances. On the other hand, a much enhanced water vapor column is seen which is

complex to explain for now it needs to be researched more. Therefore, we use the CCNY CIMEL to assess the Medgar Evers data.

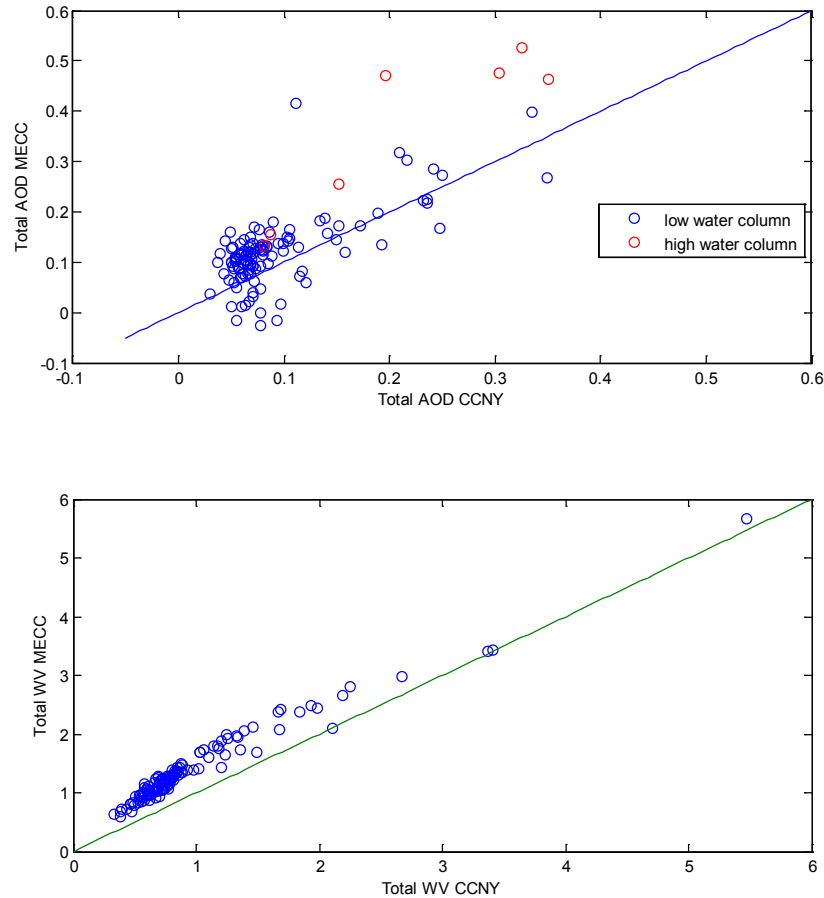


Fig. 4.14 Comparisons of MECC and CCNY to assess strange overestimation of Water Vapor.

We note that MECC water vapor seems to be running very high even relative to CCNY site. For low WV, the bias is there although surprisingly for higher WV, the bias seems to be significantly reduced. It may possibly be some issue with the MECC Water vapor channel.

Based on the above results, there is a clear cleaning out of the aerosols by the time they reach Lamont. Local Emissions raise the AOD at Princeton to urban levels.

Unfortunately, there seem to be quite significant anomalies in the total column water vapor although all sites are highly correlated.

More information can be gained from CCNY CIMEL vs. MFRSR comparison taken from July – December 2004. We first note, in general, high correlations but MFRSR shows an increased bias for the WV. In addition, MFRSR retrievals of AOD seem to have some significant cloud contamination cases.

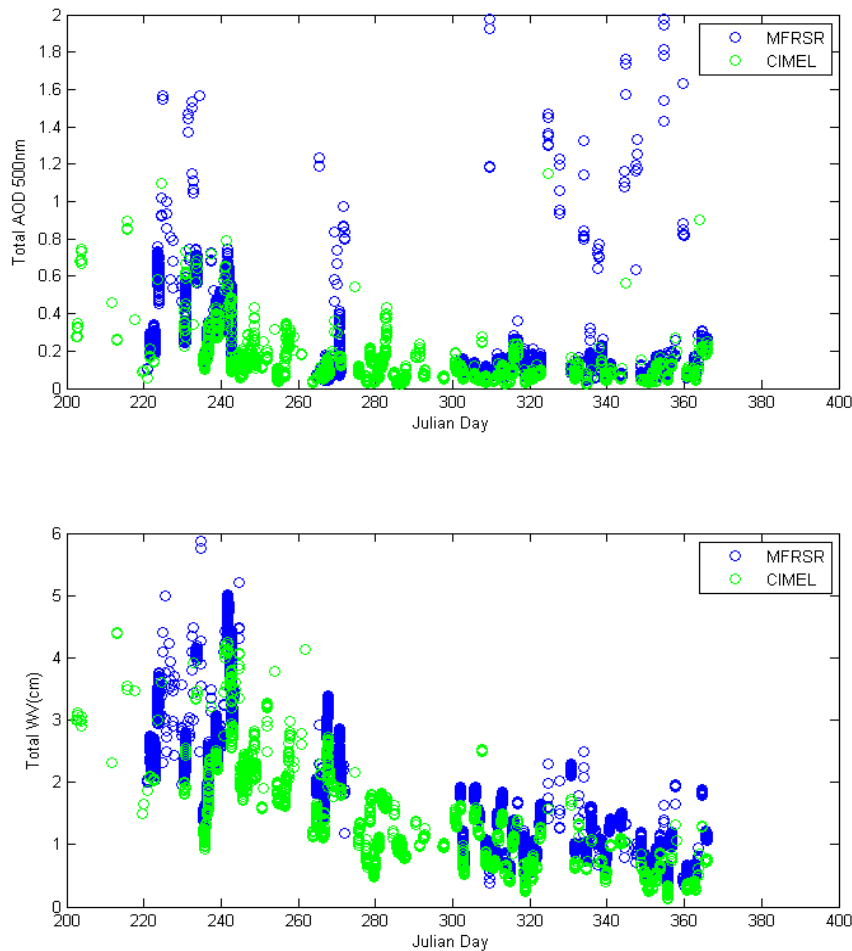


Fig. 4.15 Time series comparisons of total AOD and WV at CCNY (CIMEL vs MFRSR)

In figure 4.16, daily averages of simultaneous measurements are plotted.

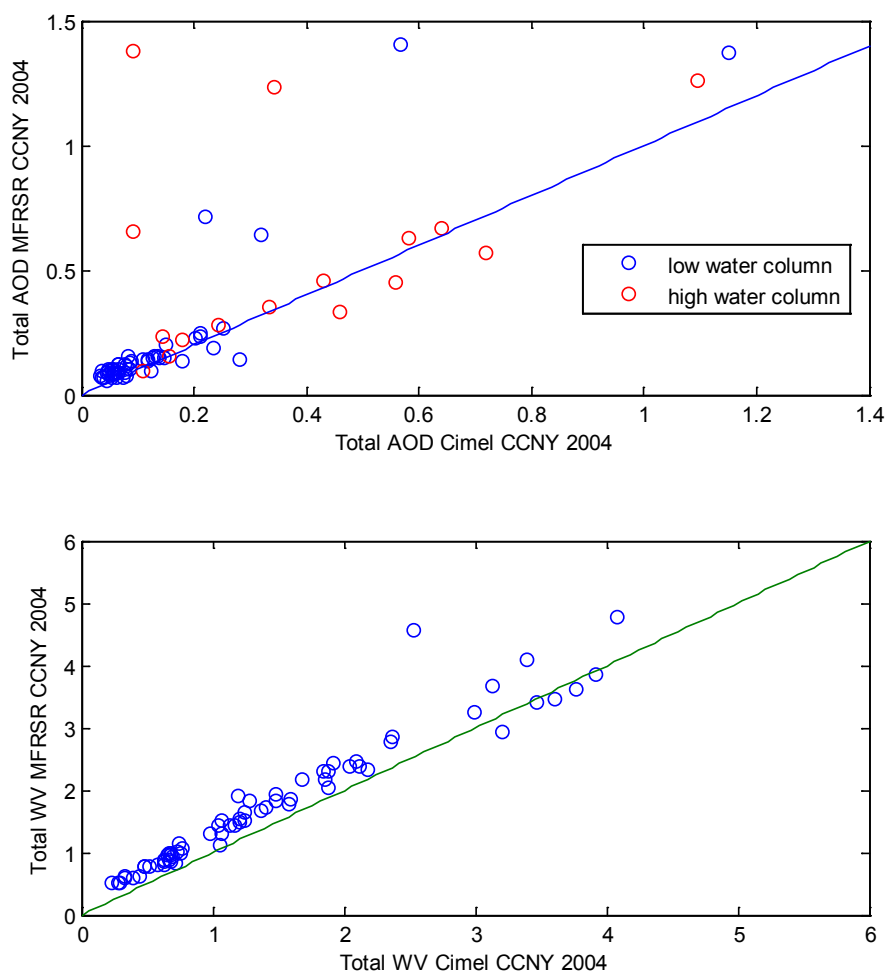


Figure 4.16 Daily averaged regressions between CIMEL and MFRSR (CCNY)

Very high correlations are seen but a clear high bias is observed in the MFRSR data for the WV. Therefore, our comparisons of WV between MECC MFRSR and CCNY CIMEL where MECC had anomalously high WV can be at least partially explained by the high MFRSR WV bias.

However, to see the true potential for the MFRSR, we also reduce the time averaging to $\frac{1}{2}$ hour in figure 4.17. Note in particular the removal of the anomalous

MFRSR AOD. This is a signature that on this time scale, the MFRSR results have no comparable CIMEL results which we take to be a manifestation of the CIMEL Level 1.5 cloud clearing.

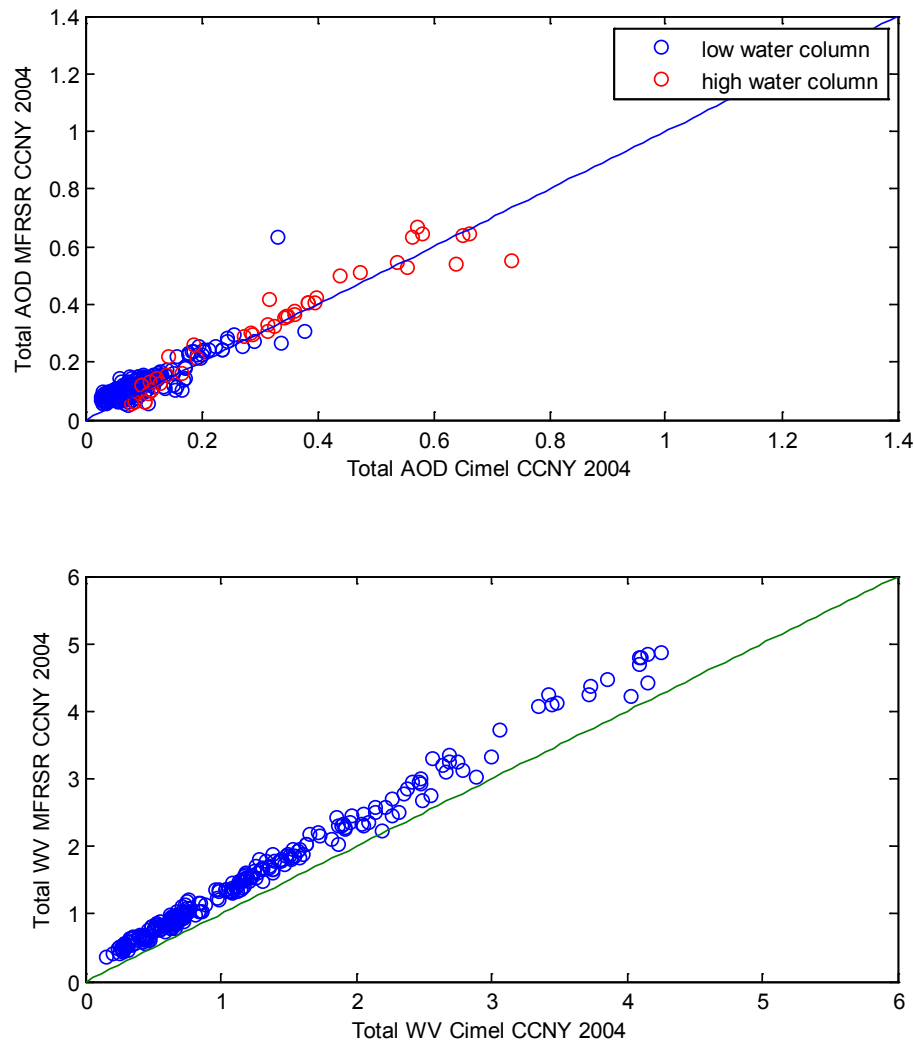


Figure 4.17 30 minute regressions between CIMEL and MFRSR (CCNY)
The water vapor bias however is still clearly observed and is not modified in amplitude.

Interestingly, we also note a bias in the AOD particularly for low water vapor cases where the aerosols phase function is perhaps not hygroscopically mixed.

Finally, to illustrate that the CIMEL water vapor retrieval is not the problem, we compare CIMEL against GPS retrievals of water vapor in Figure 4.18. We note no bias of CIMEL retrievals. The solid lines are the 10% error lines and the RMS percentage error < 5%. Therefore, the MFRSR retrievals of water vapor have a positive bias (at least for low to moderate WV loadings) that must be investigated further.

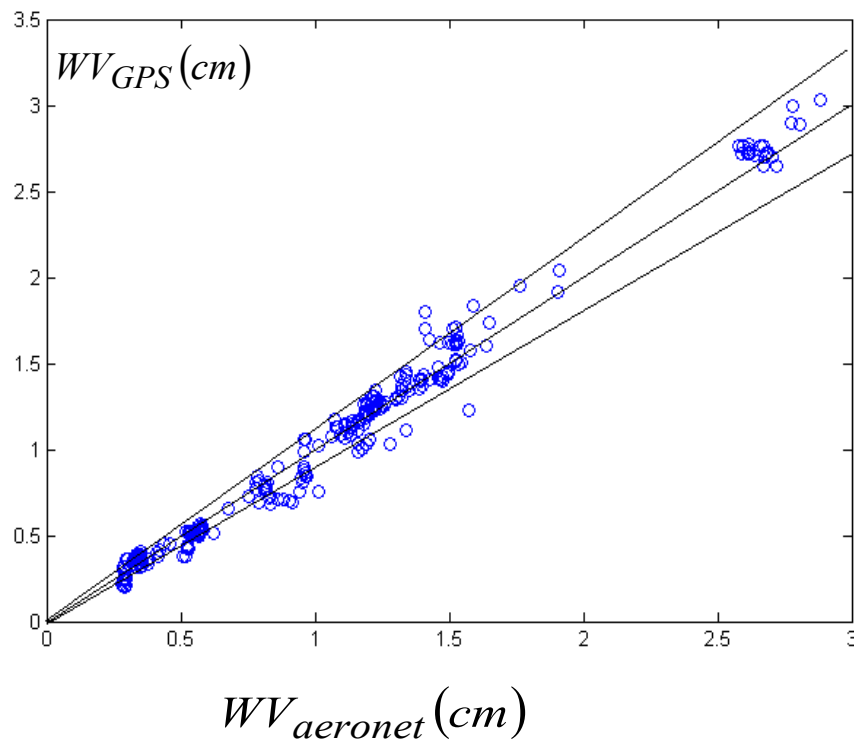


Figure 4.18 GPS-Aeronet Intercomparison Water Vapor comparison with 10 percent error bars.

Regarding the aerosol retrievals of figure 4.17, we note a small positive bias in the total AOD at 500nm. To get more insight, we also look at the Fine and Coarse Mode Retrievals. Note in particular, the high MFRSR AOD values are primarily significant overestimates of the coarse mode which reinforces that the retrieval is picking up clouds and misclassifying as coarse mode dominated.

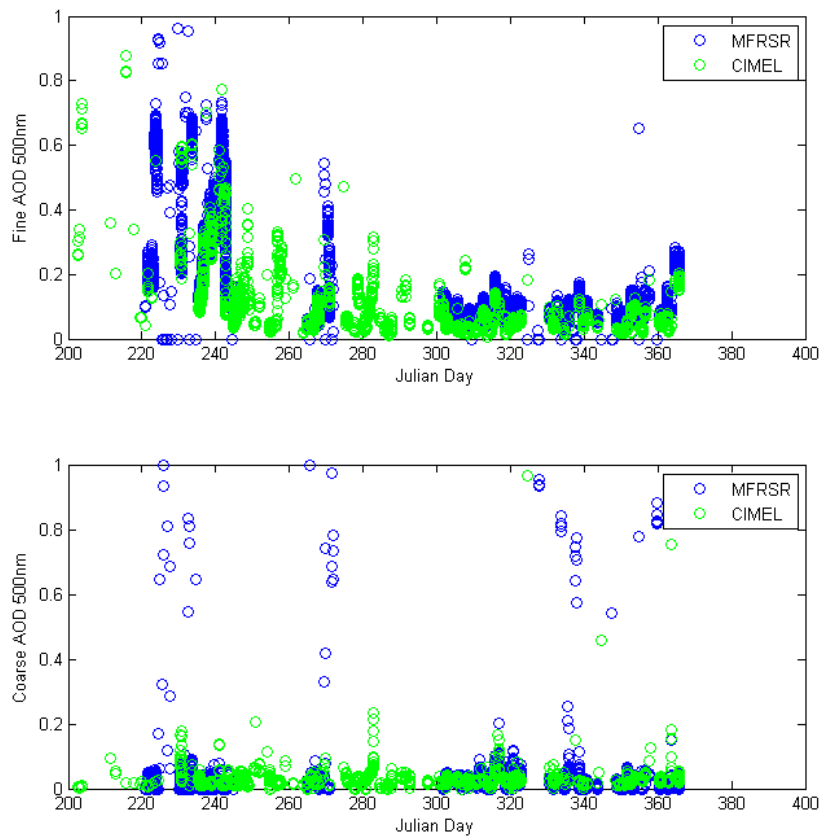


Figure 4.19 Time series comparisons of Fine and Coarse AOD at CCNY (CIMEL vs MFRSR)

The regression plot (30 minute averaging) in Figure 4.20 shows a small but significant over bias in the fine mode AOD and the coarse mode retrieval are in general much less accurate as expected.

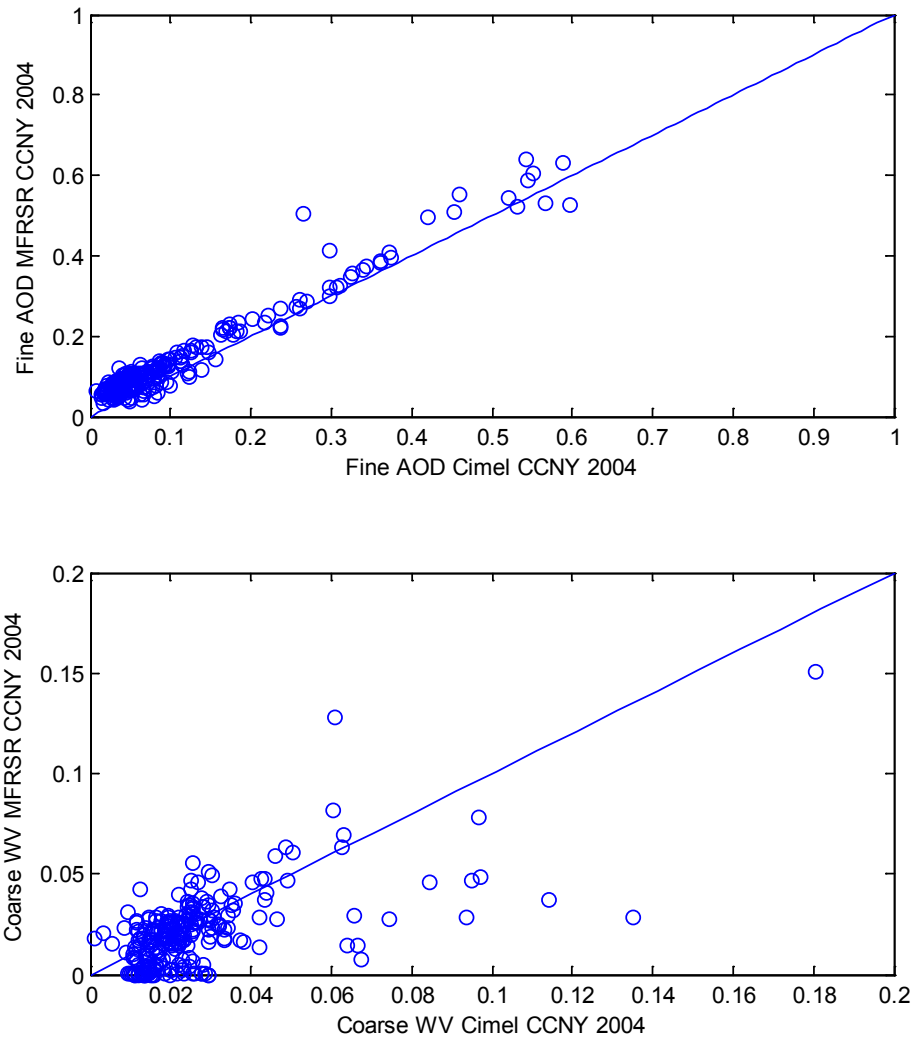


Figure 4.20 30 minute regressions Fine and Coarse Mode.

When the regressions for the daily averages are compared, we see an improvement in coarse mode but no removal of bias in the fine mode.

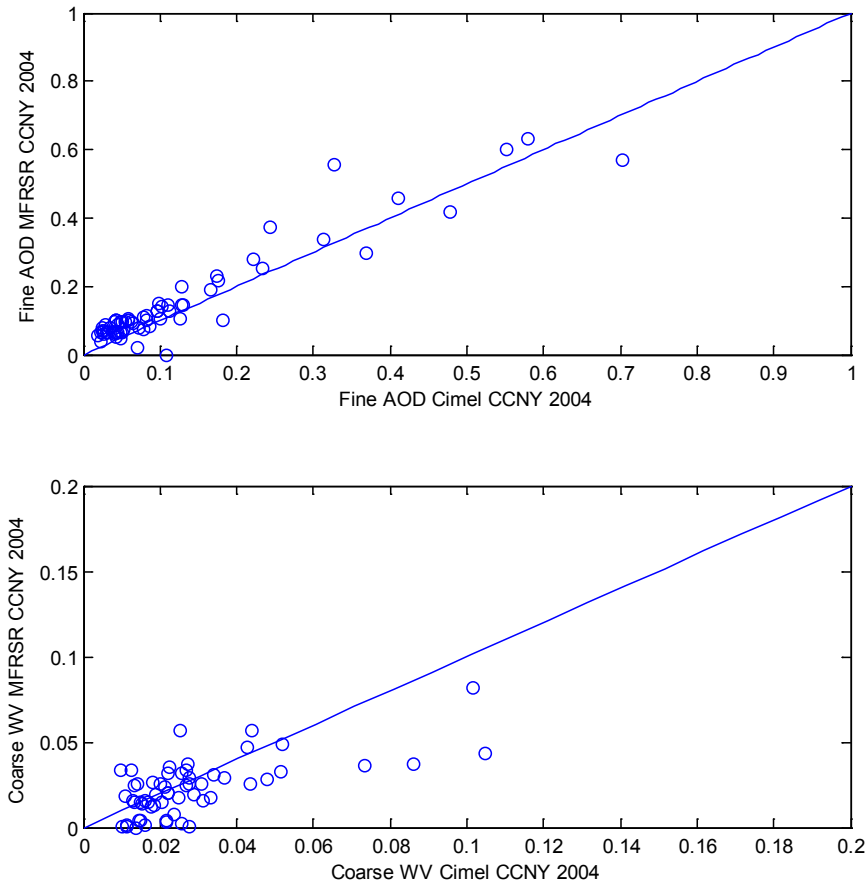


Figure 4.21 Daily regressions Fine and Coarse Mode.

The bias we see in NYC was not observed in the ARM site and might require a modification in the choice of aerosol phase function as well as possible ozone retrieval errors affecting the fine mode retrieval.

4.5 Spatial Resolution Networks

One of the most important outputs for an MFRSR networks is the capability of observing spatial variability in AOD products which might provide a better clue for pollutant sources and effects in urban areas. Unlike meteorological variables such as water vapor fields, aerosol variability can be quite significant with even larger variability expected for urban areas. To try to assess this issue, we first begin by making time series measurements with the AERONET CIMEL radiometer and look for the time auto-correlation function between measurements. In other words, we look to calculate the variations of AOD as a function of the time difference between measurements as defined below:

$$\Delta \tau(T_d) = \frac{1}{N(T_d)} \sum_{|T_i - T_j| = T_d} |\tau_i - \tau_j| \quad 4.1$$

Of course, it is not possible to have the time delay vary continuously so instead we look to separate the time into time cells $T_{D,k} \in I_k$ $I_k \equiv [(k-1) * \Delta T_D, k * \Delta T_D]$ so that the differential AOD for the k^{th} time differential is the average of all individual differentials for the k^{th} interval.

$$\Delta \tau_k = \frac{1}{N_k} \sum_{|T_i - T_j| \in I_k} |\tau_i - \tau_j| \quad 4.2$$

Once this autocorrelation is calculated, we can of course look at other statistics as well besides the mean value. For example, we can also look to calculate the frequency of

occurrence that the differential between AOD measurements is above a certain measurement threshold.

$$f_k = \frac{N(\tau_{ij}^k \leq .02)}{N(\tau_{ij}^k)} \quad 4.3$$

Where τ_{ij}^k symbolizes all optical depth differentials in interval k and N represents the number function (For Example it can represent how many cases).

In looking at the above statistics, it is useful to separate out the cases where significant water column is or is not present since it is well known that significant amounts of atmospheric water vapor lead to significant modifications of aerosol optical depth through hygroscopic interaction. The choice of separation is made as 3 cm which accounts for approximately 35 % of the cases that occur in the urban NYC area as seen in figure 4.22.

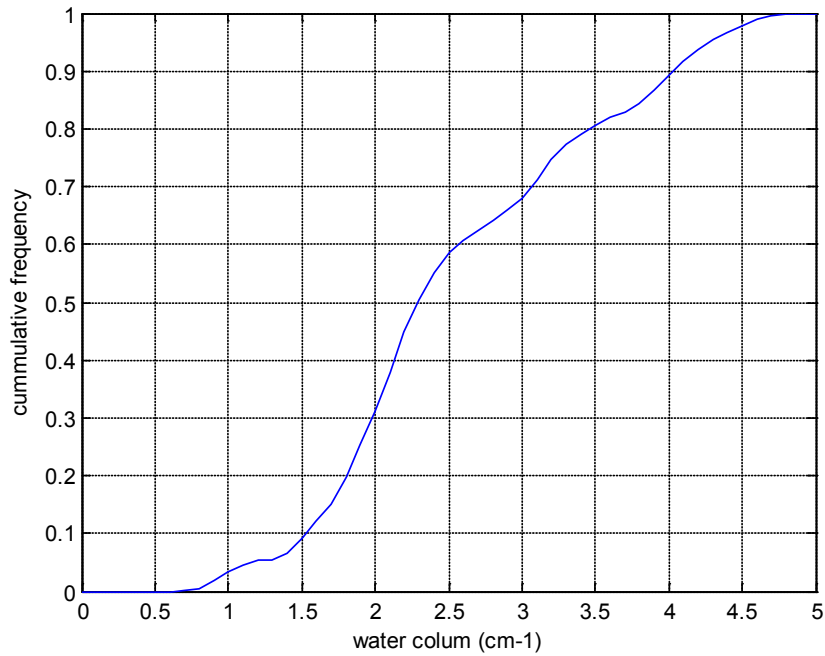


Fig. 4.22 Cumulative distribution of water column cm-1)

The results of the statistical analysis for the year of 2007 are plotted below in figure 4.23 and 4.25. To simulate the minimum thresholds of CIMEL, we only keep AOD values > 0.06 . We note for example that the mean autocorrelation time needed to resolve above the MFRSR retrieval limit of .02 is > 1 hr for low column water cases but is on the order of 10 minutes for the high water column. This is also sharply put into focus with the frequency histogram of figure 4.24 Here we note that for high water column cases $> 50\%$ show significant variations within the first hour and therefore it is important that these cases be observed.

However, it is important to realize that the time autocorrelation is not useful to our assessment of the density of MFRSR instruments that are needed to observe the natural variability of aerosols. However, we could get a reasonable estimate if we can measure the wind speed which would allow us as a first estimate to convert time

autocorrelations to spatial autocorrelations. Of course, once these autocorrelations are known ahead of time, we can then use multiple instruments directly to see how well these pseudo spatial correlations compare with true spatial correlations at different instrument distances.

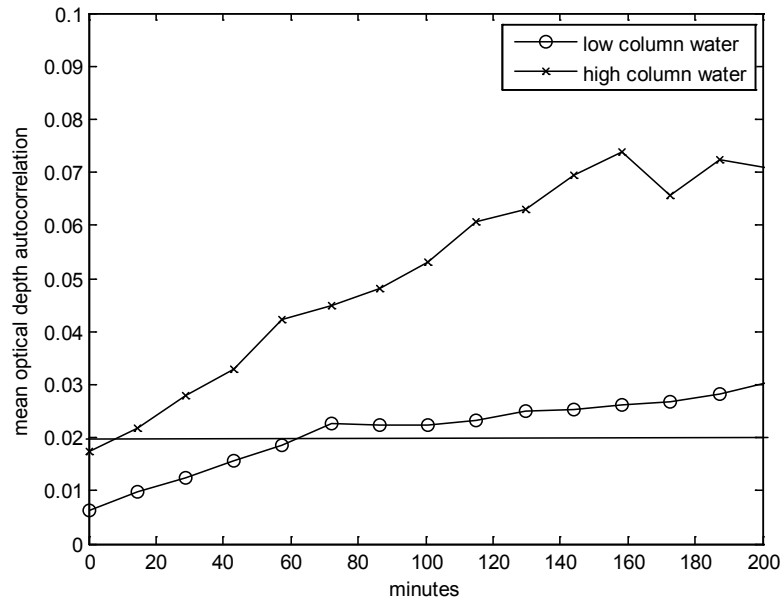


Figure 4.23 Mean Autocorrelation for both low water column and high water column cases. Threshold WV is set at 3 cm

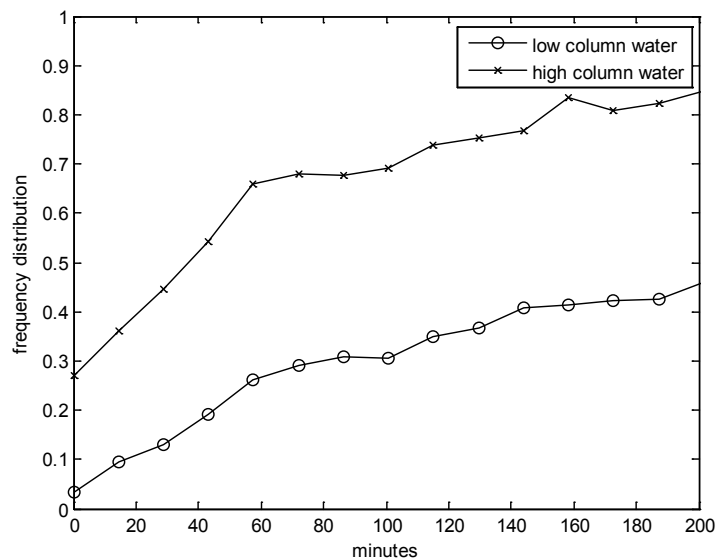


Figure 4.24 Percentage of cases where optical depth differential > .02

One issue that has to be looked at carefully however is that surface winds are not sufficient to characterize the aerosol column since the aerosols is distributed vertically. However, if we assume that the aerosols lie within the planetary boundary Layer and the average PBL height is on the order of 2km, we can get a better estimate of the true effective wind speed. In figure 4.25, we plot the regression plot between the surface wind and 2km averaged winds as obtained from the 40km NAM Meteorological Model used in the NOAA HYSPLIT Back trajectory engine.

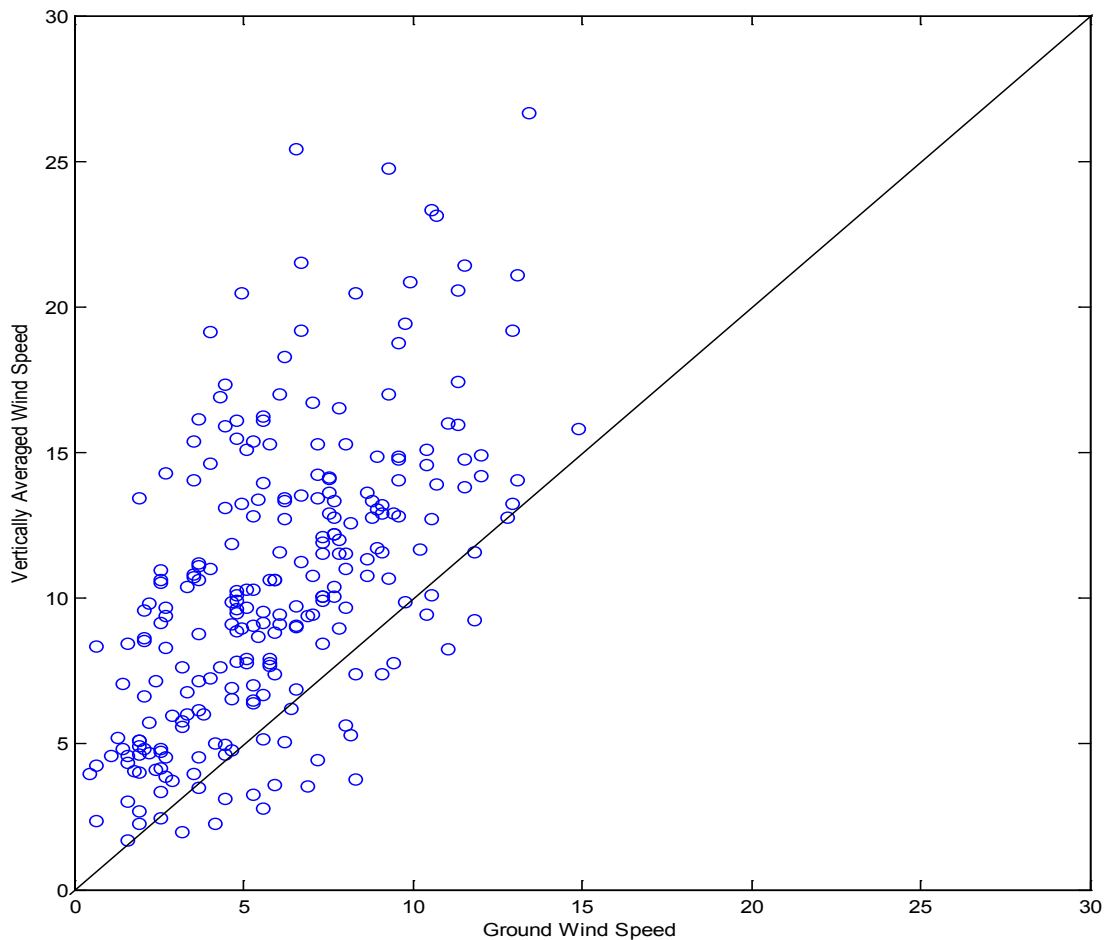


Figure 4.25 Scatter plot of ground wind vs. PBL averaged winds.

We note in particular, although the winds are correlated to some level, the magnitude of the wind speed needed is nearly twice as large as that based on the ground measurements. Therefore, using the surface winds as a measure of converting time delay to space delay would significantly underestimate the spatial correlation scales. Incorporating the PBL averaged winds; we simply multiplied each time differential by the wind speed where the

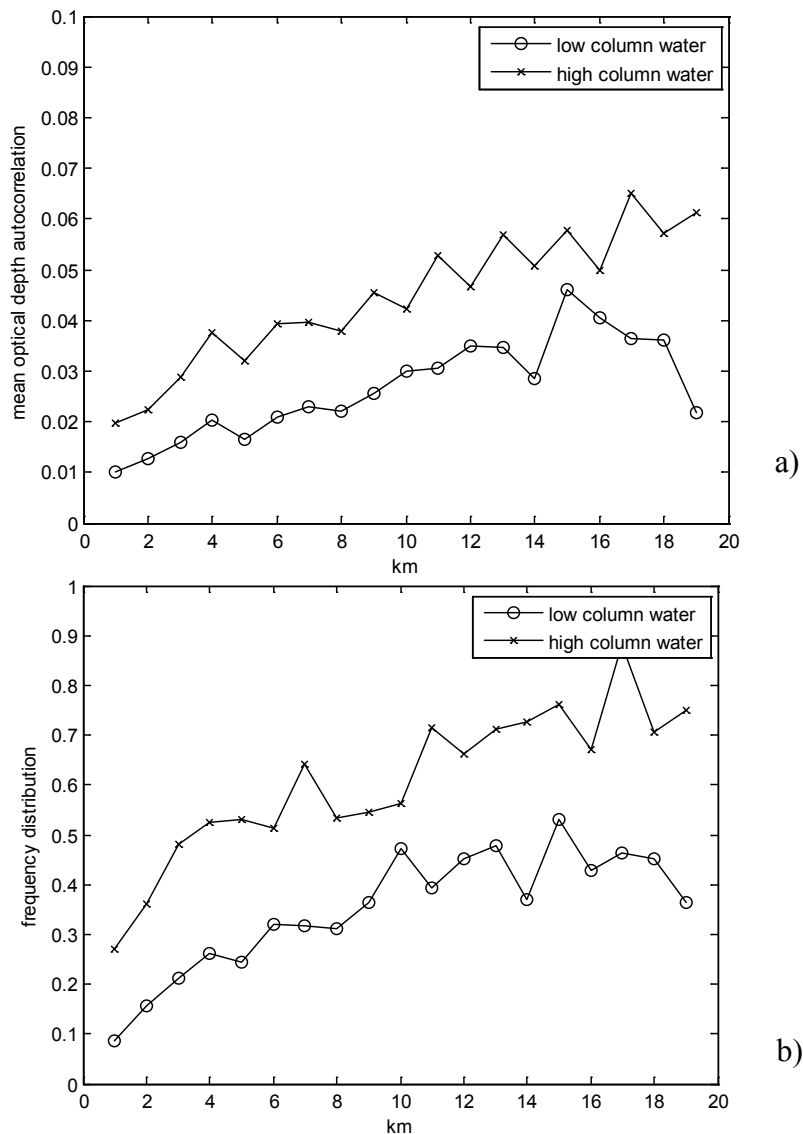


Figure 4.26 a) Mean Autocorrelation for both low water column and high water column cases. b) Frequency distribution.

Wind was evaluated at the center of the time differential. $d_{ij} = \bar{w}_{ij}T_{ij}$. This does not cause any issues since the periods are very small for the most interesting short term dynamics (< 5 km). The scaling results are given in figure 4.26 after accounting for a local wind speed scaling. Therefore, we see that spatial differentials as short as 4km should provide meaningful results that can be seen over the error levels from the ARM data.

However, we do encounter higher errors in the CCNY data with biases in the total (and fine mode) AOD $\sim .02$. This might make it more difficult to obtain spatial patterns unless the source of these biases can be ascertained.

4.6 Identification of AOD source from Network instruments.

In an attempt to locate the most important sources of pollution, we have looked at mapping the statistics of the AOD retrievals by the wind direction to see which trajectories are most likely to have enhanced AOD. The wind speed calculations were done for the last 4 months from Sept 1 to Dec 31 for 2008 where the best MFRSR data is available at the moment. Unfortunately this misses some of the most important aspects of the heavy pollutant mechanisms but more extensive comparisons are to be carried out.

To account for the entire AOD within the PBL, we use average wind directions averaged from 0-2km. since surface wind directions are not appropriate when considering total PBL column AOD. The mean AOD is plotted on polar plots overlaid on a Regional MapQuest Map. The data were averaged into 20 degree by width angular bins.

Our preliminary observations are as follows.

1. One significant AOD signatures common to both sites comes from the W-SW (~210 degrees) which seems to be consistent with the I-95 pollution corridor.
2. There is also a common spike near 120 degrees which may be more due to transport from the Ohio-Valley perhaps. This direction in fact seems to be most frequent wind path taken during this period.
3. The Princeton site seems to have an interesting spike from the S-SE which may come from local pollution sources. More study is needed over entire annual cycle.

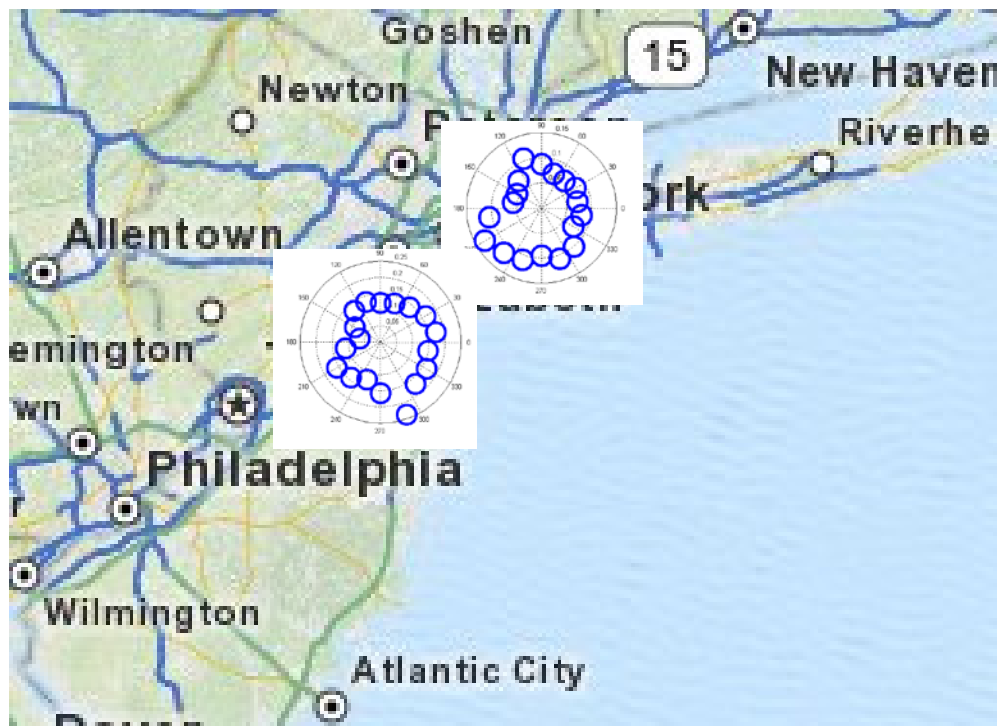


Figure 4.27 displays the statistics of the AOD retrievals by the wind direction for two sites Princeton and CCNY.

4.7 Future Network Work

Finally, it is important to appreciate the fact that the MFRSR network is expected to grow on a yearly basis. For instance, we are currently adding a new device to the network at Stevens Institute of Technology located (40.74267, -74.027349) in Hoboken NJ. This device should be in the network list as soon data collection and transfer mechanisms are error free. Every instrument should be coordinated with other instruments to provide suitable validation capabilities for current and future sensors. A general architecture of what is planned by end of 2010 is given in table 4.2 and also on figure 4.5 with the new Glory Satellite path. This architecture plus the ability of the MFRSR to not only retrieve AOD but extract fine/coarse mode fraction, effective fine radii and possibly Single Scatter albedo is crucial for validating sensors such as GOES, GOES-R, MODIS and APS on GLORY.

At the same time, we will continue to develop the web based tools that will allow us to pull operational products including GOES-GASP and MODIS AOD and provide users with matchup data over both spatially for a given overpass as well as time series record.

5 Match-ups Against Satellite Data, MODIS AND GOES

5.1 Satellite Retrieval Validation

As we discussed in 2.1.2, operational detection of aerosol over urban land was limited to single observations from polar orbiters and difficulties from land surfaces were mostly difficult to overcome. Therefore, these platforms were relegated to aerosol detection over dark surfaces. On the other hand, efforts to improve retrieval over brighter surfaces have been made. For example, MODIS uses the 2.1 micron channel to measure the ground albedo and then makes use of quasi-empirical relations connecting the ground albedo in the blue and red channels to the 2.1 micron channels. Although we have seen the difficulties in applying this over urban surfaces, regional studies that use collocated aerosol measurements to obtain regional surface Spectral Regression Coefficients (SRC) seem to be very promising and offer the possibility that more robust models for the SRC's for different land types can be obtained. Unfortunately, observations from polar orbiters are not frequent enough to address many air quality measurement and prediction. Specifically, there is a strong need for near real time aerosol information with high latency. This hole is somewhat filled with ground based remote sensor networks such as AERONET and AIRNow who's in-situ instrumentation provides accurate aerosol information at high temporal resolution but unfortunately their coverage is extremely limited and hampers our ability to monitor and forecast particulate matter aerosol properties on a global scale.

Therefore, it is not surprising that efforts to retrieve aerosols from Geostationary Platforms are significant. In particular, based on the general idea that ground albedo does not change dramatically over a time scale much longer than the aerosol fluctuations, it is possible to isolate the ground albedo contributions simply by making observations over sufficiently long periods of time and expecting that in such a long time period, “clean:” atmospheres in which aerosol loading is negligible can be observed. Then assuming the same surface albedo holds over the entire period, the aerosol contribution can now be determined for all other cloud cleared observations. This very intuitive idea is the underlying basis for Geostationary Aerosol and Smoke Product (GASP). This algorithm has been described in references [Prados et al., 2007; Knapp et al., 2002, 2005] has now been made operational and provides data through the NOAA – CLASS Data system. However, it must be pointed out that since GOES has a single VIS channel, strong assumptions must be made to process the data. Just like for AVHRR, the aerosol type must be assumed so that fixed aerosol LUT’s based on a single aerosol class may be constructed. In particular, these one parameter look up tables are based on the continental aerosol model used in 6S. This model is basically a background continental model and has a single scattering albedo is 0.89. In addition, the aerosol optical depth of this background continental model is assumed to be .02. Full details can be found in [Vermote et al., 1997]. In addition, the TOA LUT’s include Rayleigh scattering as well as ozone and water vapor absorption. Rayleigh scattering depends on surface pressure and is calculated using Digital Elevation Maps and gases are treated simply using the U.S. standard atmosphere.

As mentioned above, the most critical aspect is the development of the GASP algorithm is the generation of realistic surface estimates. This is done by first creating a composite clear-sky background image for each time of day which is equivalent to a fixed solar zenith angle for a given pixel. This composite image is generated by collecting images over the previous 28 days including the current image, and the selecting the second darkest pixel at each location over that time period. The choice of using the second darkest pixel instead of the darkest is to avoid or reduce cases where cloud shadow contamination might arise [Knapp et al., 2005]. Although the choice of 28 days is fixed statically operationally, it should be pointed out that such rules may not be optimal and can vary seasonally and geographically seasonally. In particular, the choice is made as a way of balancing between a longer time period which increases the chances of finding an image with clear sky conditions and a shorter time period to increase the probability that the underlying surface reflectivity is stable.

5.2 GASP PERFORMANCE

The performance of GASP product taken from Prados 2007 using the AERONET station at City College of New York (CCNY) illustrates the strong potential for GASP for urban areas. However, while no bulk bias is seen in general and the correlations are fairly uniform over the diurnal cycle, figure 5.1 shows significant biases as defined by their RMS errors rise dramatically before 1215 UTC or after 2115 UTC and vary between 0.11 and 0.15 but increase considerably before and after that time period. Reasons for this include larger solar zenith angles enhancing imprecise atmospheric assumptions, a greater tendency for surface angular affects to be removed as well as a greater likelihood

of cloud shadows affecting the surface composite map. Therefore, optimized performance should be time limited between 1215 and 2115 UTC.

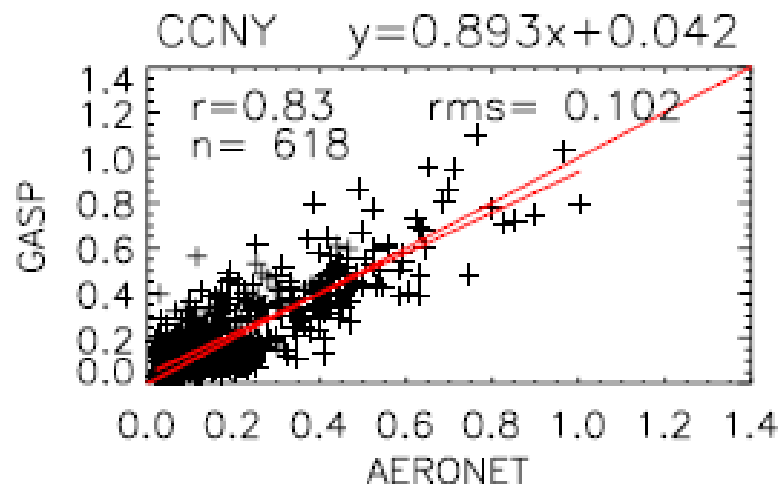


Figure 5.1 GASP AOD vs AERONET AOD at CCNY. Note although it has a high correlation, it lack of significant bias and contain a strong AOD RMS deviations.

5.3 SATELLITE MFRSR MATCHUPS

However, the MFRSR network makes it possible to improve validation statistics by increasing the likely number of matchups. For example, over a 4 week period with 3 sites reporting (Princeton, MECC and Lamont) a comparison of GASP and MFRSR total column AOD was made and displayed in figure 5.2. We note in particular, the reasonable unbiased agreement for AOD above the threshold of 0.1. We also note a small percentage of negative GASP AOD retrievals consistent with the statistics reported by the GASP algorithm group.

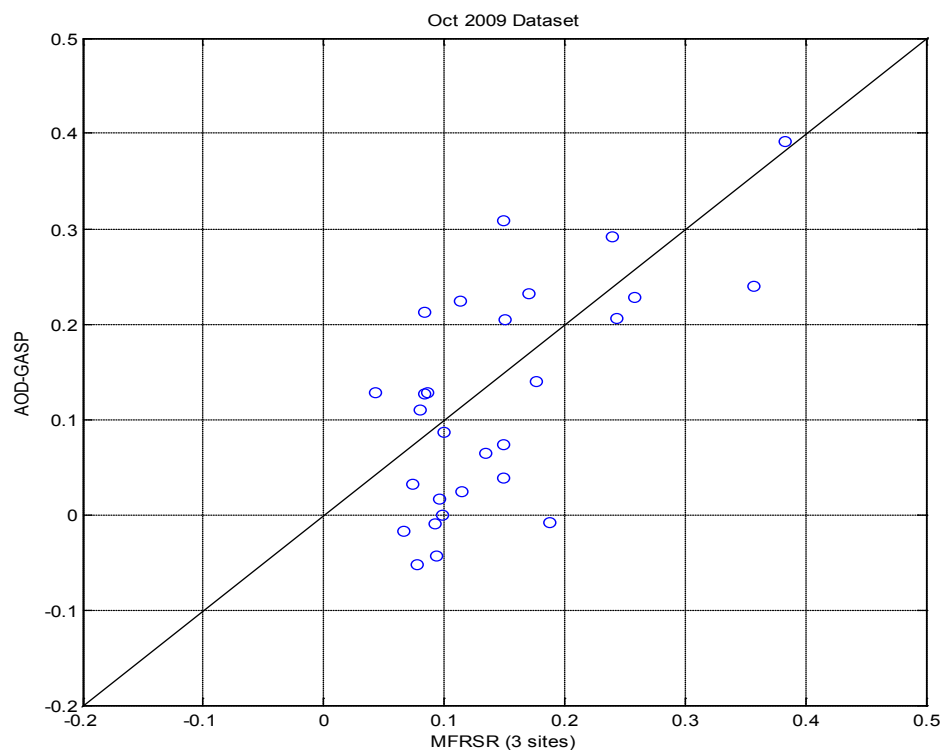


Figure 5.2 GASP MFRSR AOD matchup for Oct 1-15 2009.

On the other hand, the MODIS performance over urban areas shows significant overestimation due to the underestimation of the surface albedo in the VIS channels as a consequence of the underestimation of the VIS – SWIR surface correlation factor. This bias is shown in figure 5.3 where we find that MODIS retrieval slightly under-estimates AOD for Princeton (rural) but severely over-estimates the AOD for urban sites like NYC and MECC Brooklyn. However, if we reprocess the AOD using our regional values for the VIS-SWIR coefficients, the retrieval should be much improved. This is demonstrated in Figure 5.4 which plots the differences in C005 algorithm using both the operational surface and CCNY regionally trained surface model at MECC.

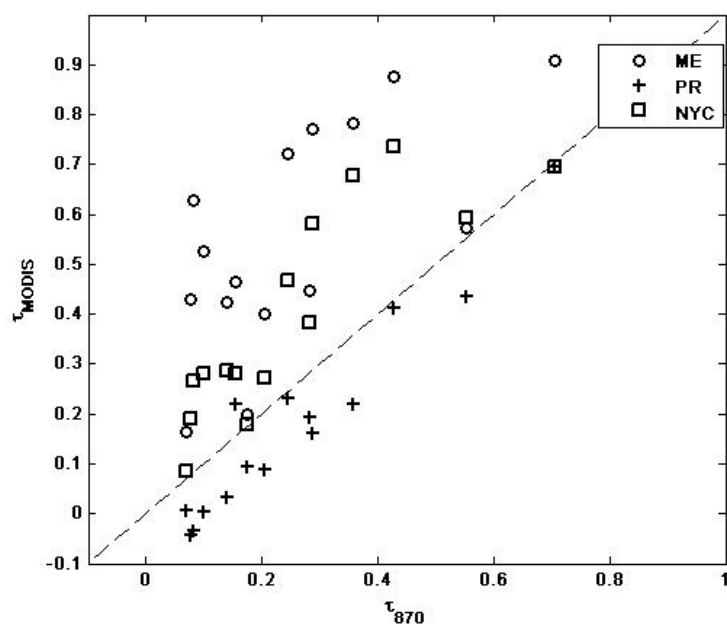


Figure 5.3 Comparison of MFRSR AOD vs. MODIS AOD between Sites: Princeton, Medgar Evers and CCNY for Sep-Oct 2007 period. Princeton is under estimated compared to CCNY and MECC.

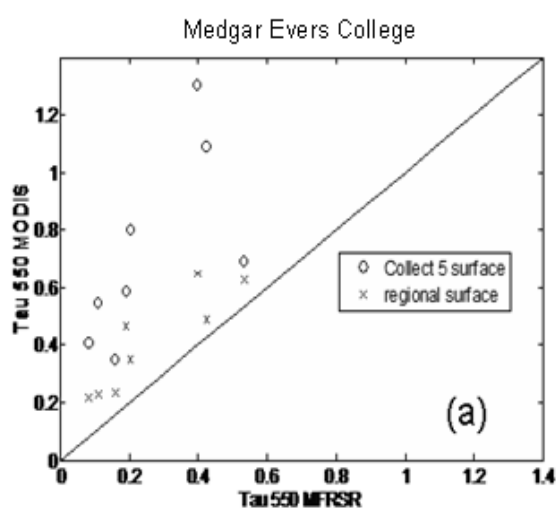


Figure 5.4 a) Comparison of MODIS derived AOD, 'x' represent using surface reflectance ratios obtained from CCNY and 'o' represent using MODIS operational result, with MFRSR AOD results b) Geo-location of Comparison

It is clear that the MFRSR AOD retrievals are significantly improved when using the regional model. It is also useful to note that the training was done at CCNY and not at MECC illustrating that spatial homogeneity filters we employ in training the surface are quite useful. However, we would again like to emphasize that with a spatially distributed MFRSR network, regional training of the surface would be far improved since we would not require spatial homogeneity.

All MFRSR retrieved data will be made available through the MFRSR page at <http://sky.ccny.cuny.edu/wc/Shadowband/Pascii/index.php#s3>

6 Improvement of MFRSR to Measure Single Scatter Albedo

The need to measure SSA is based on the fact that the difference between scattering and absorbing aerosols makes a substantial difference on energy budget since scattering aerosols are a cooling mechanism while absorbing aerosols provide a warming mechanism.

Unfortunately, retrieval of SSA (or absorbing aerosol) from space is very difficult and simple multiwavelength multiangular irradiance measurements are quite limited. However, the Ozone Monitoring Imager (OMI) is able to measure absorbing aerosols particularly if they aloft plumes [Torres 1998] while the soon to be launched GLORY Advanced Polarimetric Scanner (APS) with polarization capabilities and large spectral coverage are optimized to retrieve aerosol SSA [Waquet 2009]. Clearly, ground retrievals to match against these satellite retrievals would provide added value to the MFRSR network

6.1 Current capabilities

Indirect methods for the determination of the aerosol properties from solar irradiance measurements have been proposed and tested in several studies, using either global irradiance spectra [Kylling *et al.*, 1998; Kazantzidis *et al.*, 2001], or the direct to diffuse irradiance ratio at select wavelengths [King, 1979; Slusser *et al.*, 2001] or radiances measured by satellites [Herman *et al.*, 1999; King *et al.*, 2003]. Ground-based Sun photometry is a vital source of information for characterization of atmospheric aerosols,

and provides a validation tool for global satellite aerosol retrievals. Apart from CIMEL Sun/sky radiometers, the multi-filter rotating shadowband radiometer (MFRSR) is also widely used for measurements of visible range spectral aerosol optical depth (AOD or τ_λ). (According to AMS Glossary of Meteorology, optical thickness gives the line integral of extinction along any line of sight (e.g., Sun photometer to the Sun), while optical depth is optical thickness projected onto a vertical path). Usually, the principal component measured by these SpectroRadiometer is the global irradiance incident on a horizontal surface along with the direct irradiance, which in turn can be used to derive the optical depth of the aerosols [Huber *et al.*, 1995; Marenco *et al.*, 1997; Kylling *et al.*, 1998] in the operational spectral region. The diffuse component of the solar radiation is can in principle be derived as the difference between the global and the direct irradiance.

The spectral AOD and Angstrom exponent are the only significant retrievals available from the major networks of MFRSR data, unlike the advanced data products availability from AERONET. For the past several years, Alexandrov *et al.* [2002a; 2005] worked on designing, improving and testing retrieval algorithms for MFRSR data to provide aerosol size distribution parameters. Alexandrov *et al.* [2002a] used a simplified monomodal aerosol size model, which tends to produce large NO₂ columns that are strongly correlated with AOD [Alexandrov *et al.*, 2002b]. Alexandrov *et al.* [2008] used a more realistic bimodal aerosol model with the fine mode effective radius to be smaller than 0.2 μm , which still showed strong correlation with AOD. The method of getting the optical depth from MFRSR is suitable in getting the effective variance and effective radius but since these measurements are not sensitive to albedo, we needed to obtain the

diffuse to direct irradiance ratios. In the present study, we focus on investigating the effect of atmospheric turbidity, surface albedo and SZA on the diffuse-to-direct irradiance ratio and thereby the effect of single scattering albedo.

It is hoped that this study will provide us with the tools to account for the errors that make the MFRSR measurement of the diffuse / direct irradiance ratio needed to make SSA retrieval feasible.

6.2 Simulation Studies

6.2.1 Aerosol Model

The retrieval technique consists of three steps. The first step is the current MFRSR NASA-GISS processing provides the aerosol size distribution. Once the full microphysical description is retrieved, the aerosol optical parameters using Mie Theory from the microphysical parameters are calculated and then ingested in the Radiative Transfer model for obtaining both the diffuse and direct irradiance measurements at ground level.

6.2.1.1 Bimodal Size Distribution

As explained by (Alexandrov 2008), a more realistic bimodal aerosol size distribution was adopted with two distinct classes: fine and coarse particles, each having a distinct mode in the aerosol volume distribution function. We assume that the aerosol

size distribution can be described by a combination of two lognormal distributions that represent fine (f) and coarse modes [Dubovik *et al.*, 2002]:

$$n(r) = \frac{dV(r)}{d \ln r} = \frac{N_f}{\sqrt{2\pi}\sigma_f} \exp\left(-\frac{[\ln r - \ln R_f]^2}{2\sigma_f^2}\right) + \frac{N_c}{\sqrt{2\pi}\sigma_c} \exp\left(-\frac{[\ln r - \ln R_c]^2}{2\sigma_c^2}\right) \quad 6.1$$

where r (in μm) is particle radius, N (in cm^{-3}), R (in μm) and σ^2 are the volume concentration, volume median particle radius and variance, respectively. The particle number size distribution is linked with volume size distribution as $n(r) r (4\pi r^3/3) = dV(r)/d \ln r$. The shape of the aerosol size distribution is assumed to be independent of altitude so that the vertically integrated number density N is related by a scale height factor to the surface number density.

The variables physically independent of particle size (if we do not consider processes like coagulation) are the total number of particles and the particle number fraction:

$$\hat{v}_f = \frac{N_f}{N_f + N_c} \quad 6.2$$

where N_f and N_c are column numbers of fine and coarse mode particles, and $\hat{v}_f$ is a parameter of the normalized total aerosol size distribution:

$$n(r) = \hat{v}_f n_f(r) + (1 - \hat{v}_f) n_c(r) \quad 6.3$$

Where $n_f(r)$ and $n_c(r)$ are (normalized) size distribution functions for respective fine and coarse mode aerosol. Table 6.1 summarizes the assumed parameters of an aerosol bimodal size distribution in our present study.

Size distribution parameter	Fine mode	Coarse mode
Mean particle radius (r_m)	$r_f = 0.0085 \mu m$	$r_c = 0.035 \mu m$
Standard deviation (σ)	$\sigma_f = 2.34$	$\sigma_c = 1.88$
Fine mode fraction ($\hat{v}_f$)	0.81	

Table 6.1 The assumed parameters of an aerosol bimodal size distribution.

6.2.2 Computation of Aerosol Optical Parameters

Light extinction by small spherical particles can be described by Mie theory [Bohren and Huffman, 1983; Mischchenko, 2002], with key parameter as the complex refractive index of the particle relative to the surrounding air: $m_\lambda = n_\lambda + ik_\lambda$. We can relate (through the Lorentz-Mie theory) the aerosol optical depth τ_a with size distribution and complex refractive index as equation 6.4.

$$\tau_\lambda = \int_{r_{\min}}^{r_{\max}} n(r) Q_{\text{ext},\lambda}(r, m_\lambda) dr \quad \tau_\lambda = \int_{r_{\min}}^{r_{\max}} n(r) Q_{\text{ext},\lambda}(r, m_\lambda) dr \quad 6.4$$

where $n(r)$ is particle number size distribution, and $Q_{\text{ext},\lambda}$ is extinction cross-sections.

Complex refractive index characterizes scattering (n_λ) and absorbing (k_λ) properties of the particles. For simplicity, the absorption and scattering effects of a soot-based nucleation mode were neglected (as were relative humidity induced variations of the real part of the refractive index) in favor of a fixed purely scattering refractive index

(n) of 1.5 for fine and coarse modes. The Mie derived parameters for lognormal distribution include extinction coefficient (β_{ext}), scattering coefficient (β_{sca}), single scattering albedo (ω_0), asymmetry parameter (g) and the phase functions $p(\theta)$ for the scattering angles (θ) between 0 and π . These parameters are obtained for fine and coarse modes distinctly at 555 nm wavelength.

As per our assumption of the existence of distinct fine and coarse particle modes, the aerosol scattering phase function can be written as the weighted mean:

$$P(\theta) = \frac{\beta_{\text{ext}}^f \omega_0^f P^f(\theta) + \beta_{\text{ext}}^c \omega_0^c P^c(\theta)}{\beta_{\text{ext}}^f \omega_0^f + \beta_{\text{ext}}^c \omega_0^c} \quad 6.5$$

Similarly, the weighted mean of the single scattering albedo (ω_0) is given by:

$$\omega_0 = \frac{\beta_{\text{ext}}^f \omega_0^f + \beta_{\text{ext}}^c \omega_0^c}{\beta_{\text{ext}}^f + \beta_{\text{ext}}^c} \quad 6.6$$

The superscripts f and c on the corresponding symbols indicate the fine and coarse mode components respectively. The total aerosol optical depth can be expressed as a linear sum of the optical depth of the fine (f) and coarse I modes:

$$\tau_a = \tau_f + \tau_c$$

It is standard fare to expand the phase functions in terms of Legendre polynomials given by: $P_l(\xi)$

$$P_l(\xi) = \sum_{n=0}^N a_n p_n(\xi) \quad 6.7$$

Where $P_l(\xi)$ is the phase function, ξ is the cosine of the scattering angle θ , a_n are the Legendre coefficients and p_n are Legendre polynomials. Note that n represents the

polynomial order of $P_n(\xi)$ and hence, the equation approximates the phase function as a polynomial of order N .

The Legendre polynomials are computed using the orthogonality of Legendre polynomials [Liou, 1992]:

$$a_n = \frac{(2n+1)}{2} \int_{-1}^1 P_l(\xi) P_n(\xi) d\xi \quad 6.8$$

where the first term in the expansion gives the single scattering albedo [Liou, 1992]:

$$a_0 = \frac{1}{4\pi} \int_0^{2\pi} \int_{-1}^1 P_l(\xi) d\xi d\phi \quad 6.9$$

For pure scattering ($\omega_0 = 1$), this satisfies the normalization condition: $a_0 = \omega_0 = 1$. It is customary to limit the number of Legendre polynomials in radiative transfer calculations to the number of computational polar angles, or streams, with the rationale being that little can be gained from phase functions that are more precise than the model that employs them [Thomas and Stamnes, 1999; p194]. Since the computation time is proportional to the cube of the number of streams [Stamnes et al., 2000], it is important to limit the number of streams (and hence the number of Legendre moments). This poses a problem for calculations involving the highly anisotropic phase functions associated with aerosols, as low-order Legendre polynomials are not capable of capturing the strong forward scattering peak. Modern radiative transfer codes (such as DISORT) account for strong forward scattering by approximating the forward peak as a Dirac δ -function weighted according to the number of streams [Thomas and Stamnes, 1999, p 191; Wiscombe, 1977; Stamnes et al., 2000]. The realistic phase functions are used for the present study. When realistic phase functions are not available, it is usual practice to derive using the Henyey

Greenstein approximation computed by the asymmetry parameter (g). Note that HG phase functions are not accurate in backward scattering direction. The remainder of the phase function (i.e., polar angles outside of the forward peak) is scaled and the number of moments is limited to the number of streams. Accurate representation of the Legendre coefficients is necessary to avoid a nonsensical phase function expansion.

6.2.3 Radiative Transfer Calculations Using SBDART Model

Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) is a software tool that computes the plane parallel radiative transfer in clear and cloudy conditions within the Earth's atmosphere and at the surface [Ricchiazzi *et al.*, 1998]. In SBDART calculations, the user has freedom to set various parameters for the input to the model. The only subjectively selected parameters for the input are atmospheric profile (US 62), solar zenith angle and wavelength of observation, surface albedo and aerosol properties which include aerosol optical depth, single scattering albedo and the computed Legendre moments of the phase functions. The zero'th moment is assumed by SBDART for the Legendre moments and should not be specified and the number of streams are taken as eight (8) for each wavelength. The SBDART automatically combines aerosol properties with the gas properties while making atmospheric radiative transfer computations when aerosol presence is considered.

The computation of input parameters by SBDART yields both the upwelling and down-welling irradiance (in $Wm^{-2}nm^{-1}$) at the top of the atmosphere (TOA) and at the surface levels for the computed solar zenith angles. The computed down-welling diffuse

and direct irradiance at the surface are used to generate the look-up-table (Table 2) of down-welling diffuse to direct ratio in terms of total aerosol optical depth (AOD or τ_a) and single scattering albedo (ω_0).

	Single Scattering Albedo						
	1.000	0.992	0.977	0.962	0.926	0.863	0.710
AOD							
0.1	0.263	0.261	0.258	0.255	0.248	0.235	0.206
0.2	0.426	0.422	0.415	0.408	0.392	0.364	0.301
0.3	0.605	0.599	0.586	0.575	0.547	0.500	0.400
0.4	0.801	0.792	0.773	0.755	0.714	0.645	0.501
0.5	1.018	1.004	0.977	0.952	0.895	0.799	0.606
0.6	1.256	1.237	1.201	1.167	1.090	0.963	0.715
0.7	1.518	1.493	1.445	1.401	1.300	1.138	0.829
0.8	1.807	1.774	1.713	1.656	1.528	1.325	0.947
0.9	2.125	2.084	2.006	1.934	1.774	1.524	1.070
1	2.477	2.425	2.327	2.237	2.041	1.736	1.198

Table 2: Look-Up-Table of down-welling diffuse to direct irradiance ratio in terms of τ_a and ω_0 at 555 nm.

6.2.4 Results and Discussion

6.2.4.1 Effect of Single Scattering Albedo on Down-Welling Diffuse Irradiance

The diffuse-to-direct irradiance ratio, ($F = L_{\text{diff}}/L_{\text{dir}}$), which is an indicator of the atmospheric transmittance [King, 1979], is investigated as a function of the aerosol optical depth, τ_a , and single scattering albedo, ω_0 . The single scattering albedo influences directly the diffuse irradiance, since it describes the proportion of the scattered photons in a beam to the total attenuation (scattering plus absorption). The changes in diffuse-to-direct irradiance at 555 nm due to changes in ω_0 from 1.0 (purely scattering), for a range of aerosol optical depth from 0.1 to 1.0 in increments of 0.1, for each τ_a is shown in Figure 6.1.

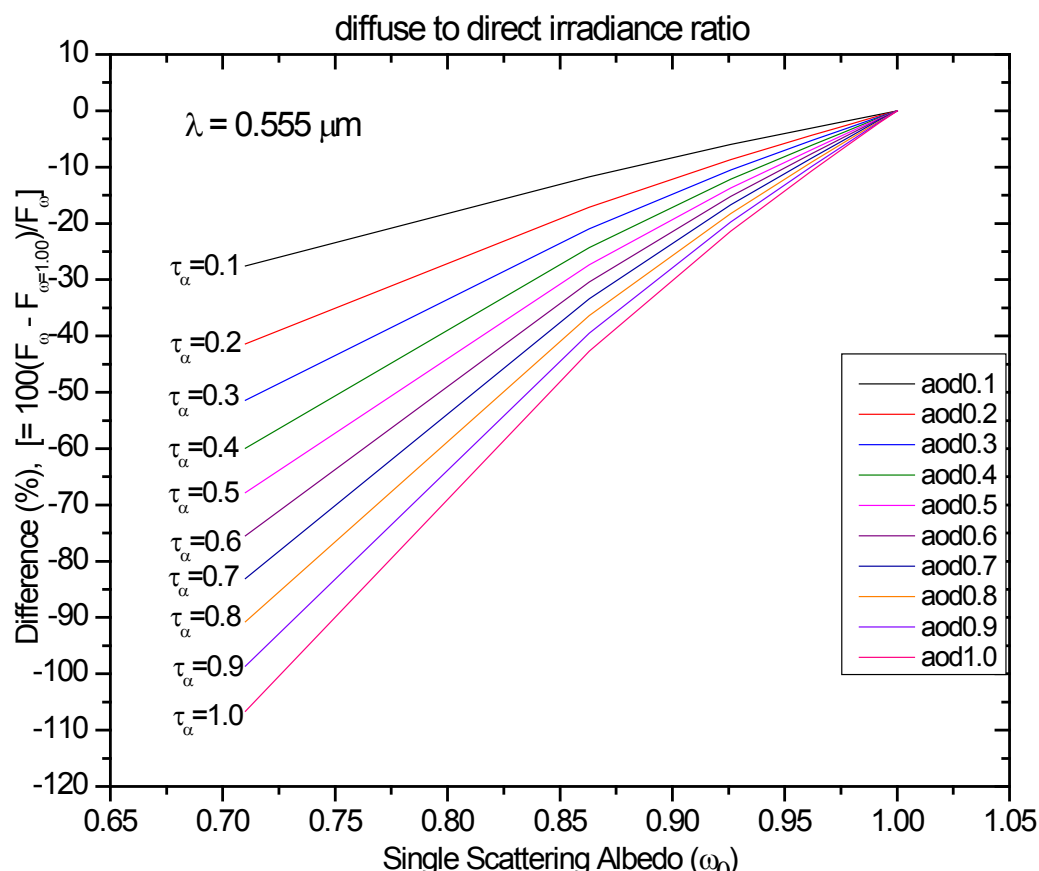


Fig 6.1: Changes in diffuse-to-direct irradiance at 555 nm due to changes in ω_0 from 1.0, for a range of τ_a from 0.1 to 1.0. Differences in radiation quantities (F) are expressed in percent.

Differences in radiation quantities (F) are expressed in percent ($100[F_\omega - F_{\omega=1.00}]/F_\omega$). The diffuse and direct components obtained at each of the solar zenith angles $0, \pi/6, \pi/4$ and $\pi/3$ are mean weighted. The direct irradiance obtained at each τ_a for different single scattering albedos are constant at different wavelengths indicating that the direct irradiance is not sensitive to single scattering albedo (ω_0) variation but sensitive to total aerosol optical depth (τ_a) values. In contrast, the diffuse irradiance is sensitive to both τ_a and ω_0 . The accuracy of the retrieved ω_0 depends on the sensitivity of the radiation quantities to changes in ω_0 and is limited by the measurements' accuracy and the detection limit of the instruments at low-intensity conditions (i.e., large SZAs, small

wavelengths, high aerosol optical depth). Also, the inversion of the multi-spectral optical depth provided particle size information that allows us to limit the degrees of freedom so we are only inverting ratio versus albedo (assuming AOD and other microphysical parameters are locked by the previous step).

6.2.4.2 MFRSR Measurement Details Relevant to Our Study

The diffuse to direct ratio is clearly sensitive to the SSA as long as the microphysical retrievals are obtained earlier. However, Single scattering albedo ω_0 retrievals from MFRSR data are not easy, since a significant underestimation of ω_0 values can be caused by the error in separation between direct and diffuse irradiances by MFRSR shadow-banding procedure. As explained in detail in Alexandrov 2006, when the band covers the sun which is the orientation meant to estimate the diffuse beam alone, it is clear that the part of the sky radiance (including the solar aureole) behind the band is blocked out. This clearly results in a reduction of the diffuse component. However, the band attempts to make some adjustment by estimating the aureole with band measurements (plus / minus) 9 degrees. However, these measurements do not completely account for the aureole so that the diffuse radiation is still underestimated.

In the present work, we make an attempt to quantify the effects of the errors in the diffuse and direct radiance using empirical ΔL corrections. In other words, we assume the total irradiance is correct but that the underestimation of the diffuse radiance by ΔL is offset by an overestimate of the direct irradiance by the same amount. Of course, the loading of the atmosphere would obviously modify the ΔL factor so some form of iterative scheme which recalculates the correction should be implemented. Furthermore,

the corrections should be at some level of approximation proportional to the diffuse radiance so we quantify the mismatch as fixed percentages in the spirit of Alexandrov [2006].

By estimating the error in the diffuse and direct beam by an empirical factor, we can look at the error we can expect for the SSA retrieval.

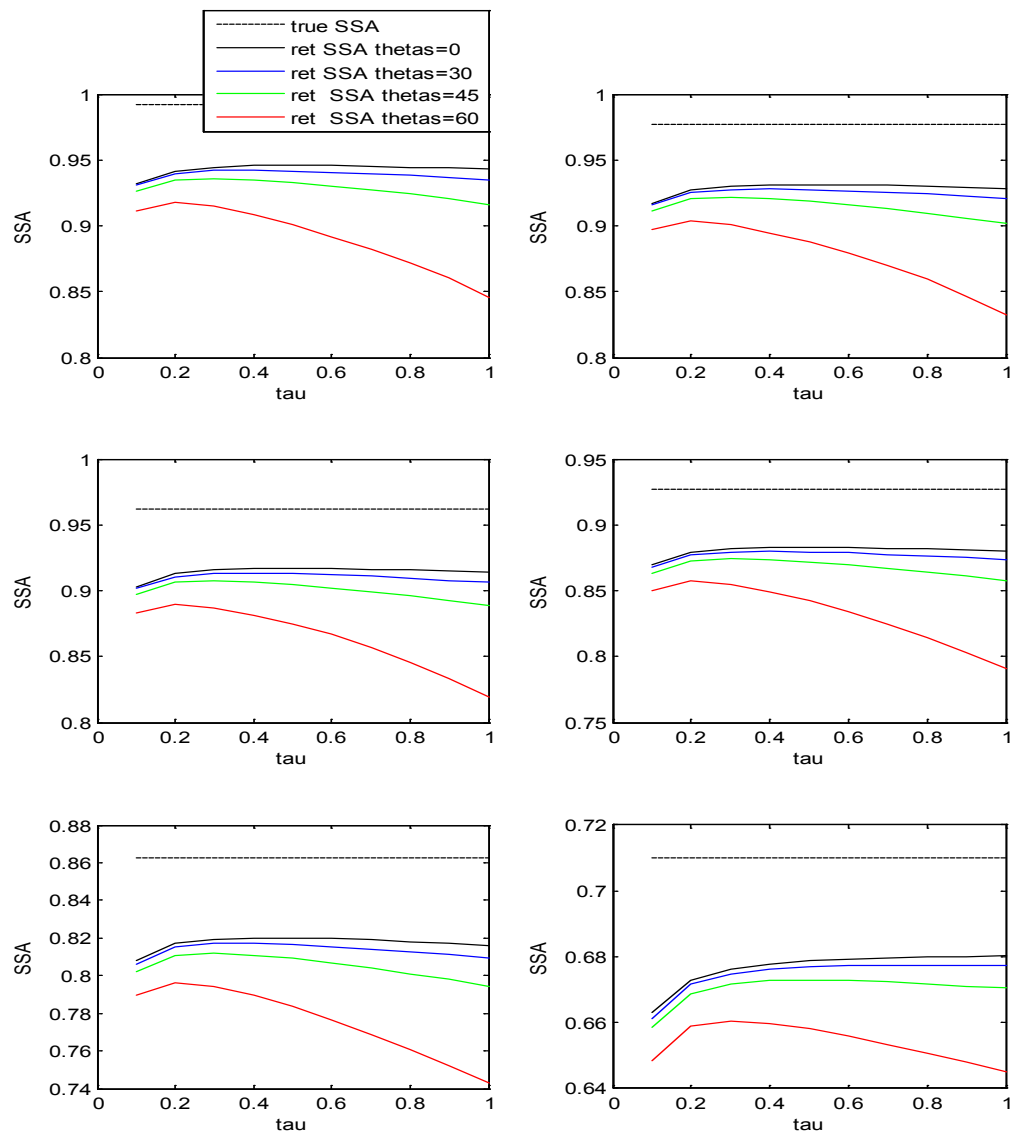


Figure 6.2 a) Sensitivity of SSA retrieval as function of optical depth and solar zenith angle. Assuming fine particles so correction factor $\sim .04$

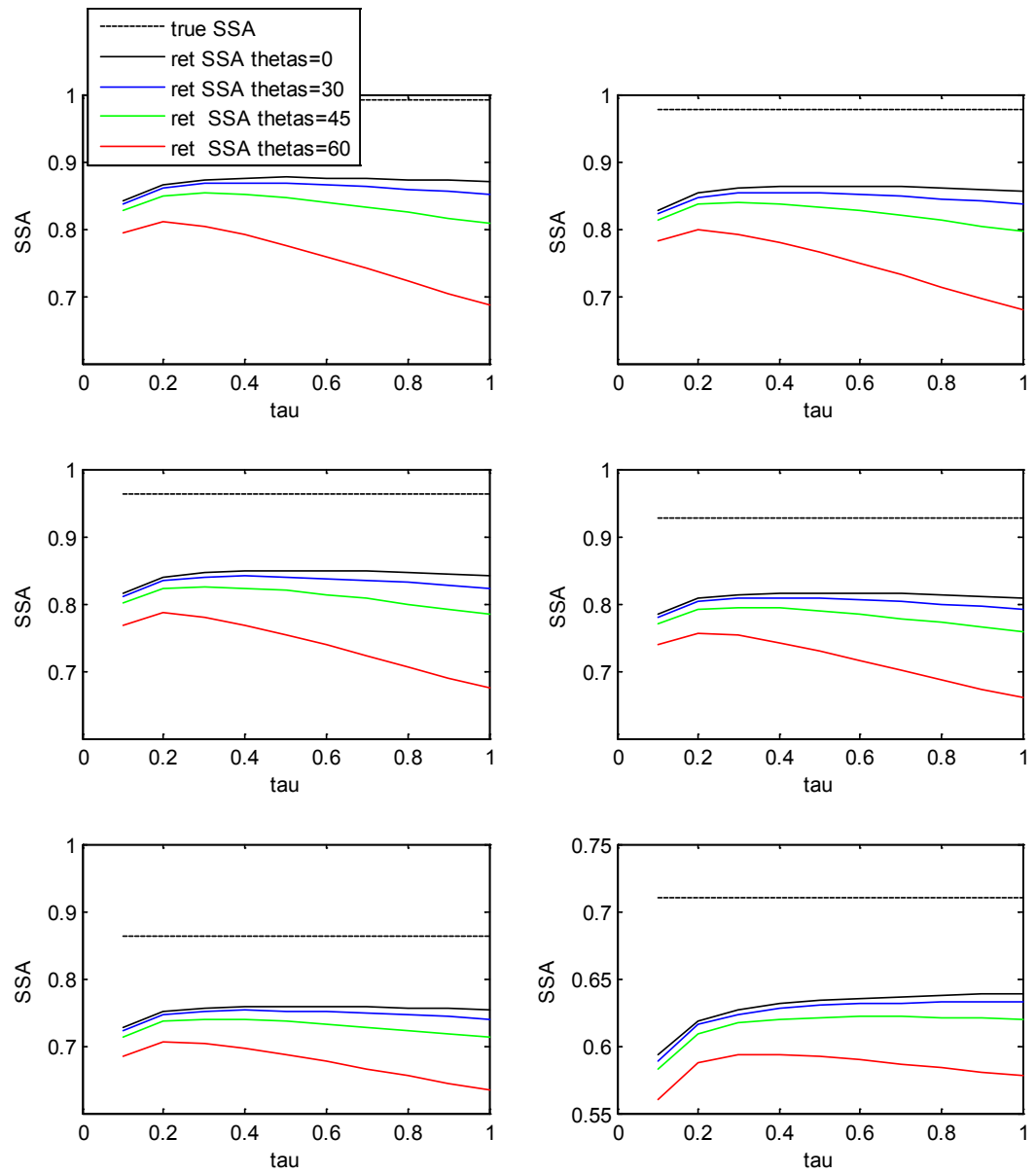


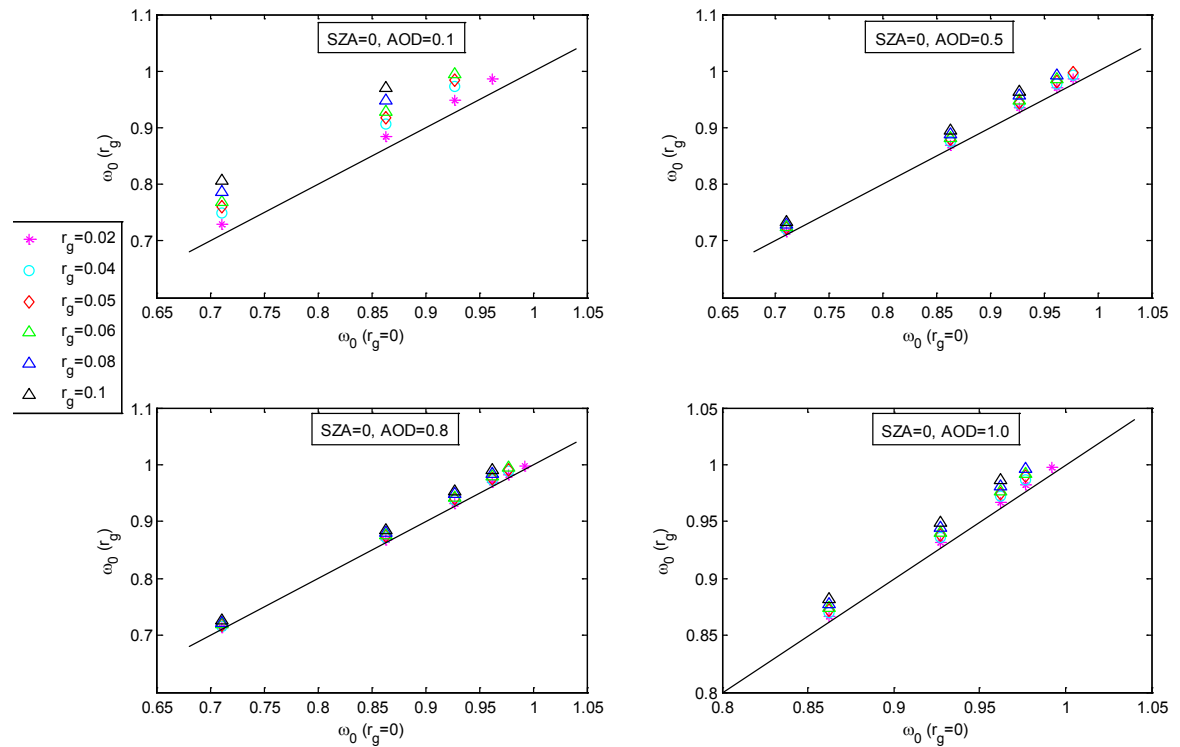
Figure 6.2 b) Sensitivity of SSA retrieval as function of optical depth and solar zenith angle. Assuming coarse particles so correction factor $\sim .10$

The amount that the diffuse / direct ratio was perturbed is based on Alexandrov (2006), the direct beam would increase by $L_{dir} = L_{dir} + \delta L_{diff} / \cos \theta$ and the diffuse would be decreased by $L_{diff} = L_{diff} - \delta L_{diff}$. This overall decrease in the diffuse / direct ratio leads to a decrease in the SSA when using the correct diffuse / direct ratios. Of course, if the above model is correct, this would be correctable by simply changing the sign of the perturbation when redefining the ratio. In the analysis of figure 6.2a, using $\delta = .04$ which would correspond to an aerosol classification in which $\tau_{870}^f / \tau_{870}^{tot} = .8$ which is a reasonable number for urban aerosols.

Based on the analysis, for fine mode dominated aerosols, the retrieval error for SSA is approximately 10% which should be manageable. The biggest errors occur for small AOD as well as large solar zenith angles but even unsophisticated corrections should allow the errors to be manageable. Furthermore, we do not see too much change in the % error as a function of the original SSA. These results seem contrary with the extreme errors that occur when the SSA retrieval was tested against AERONET at SGP (Alexandrov 2006). However, since it might be reasonable that particles at the SGP site may be more coarse mode dominated, we would get much larger errors. The needs of further tests are clearly needed. Therefore, we consider $\tau_{870}^f / \tau_{870}^{tot} = .5$ to be more reasonable resulting in $\delta = .1$. The results in this case are shown in figure 6.2b and show errors can be ~ 20-30% especially at high zenith angle and high SSA.

Of course, we are interested in exploring how well the empirical corrections fit with first principle calculations from radiative transfer (either single scattering or full multiple scattering treatment).

In addition, we also were interested in exploring the errors that result in not accounting for the surface albedo. Therefore, we ran SBDART for different Lambertian surface reflectance varying from vegetation (.02 to urban (~ 0.1) in the VIS (550nm) channels. At low AOD (.1), the effects are most significant with errors ~15-20% but become less significant at high AOD. For the Lambertian surface, the solar angle is not very important.



a)

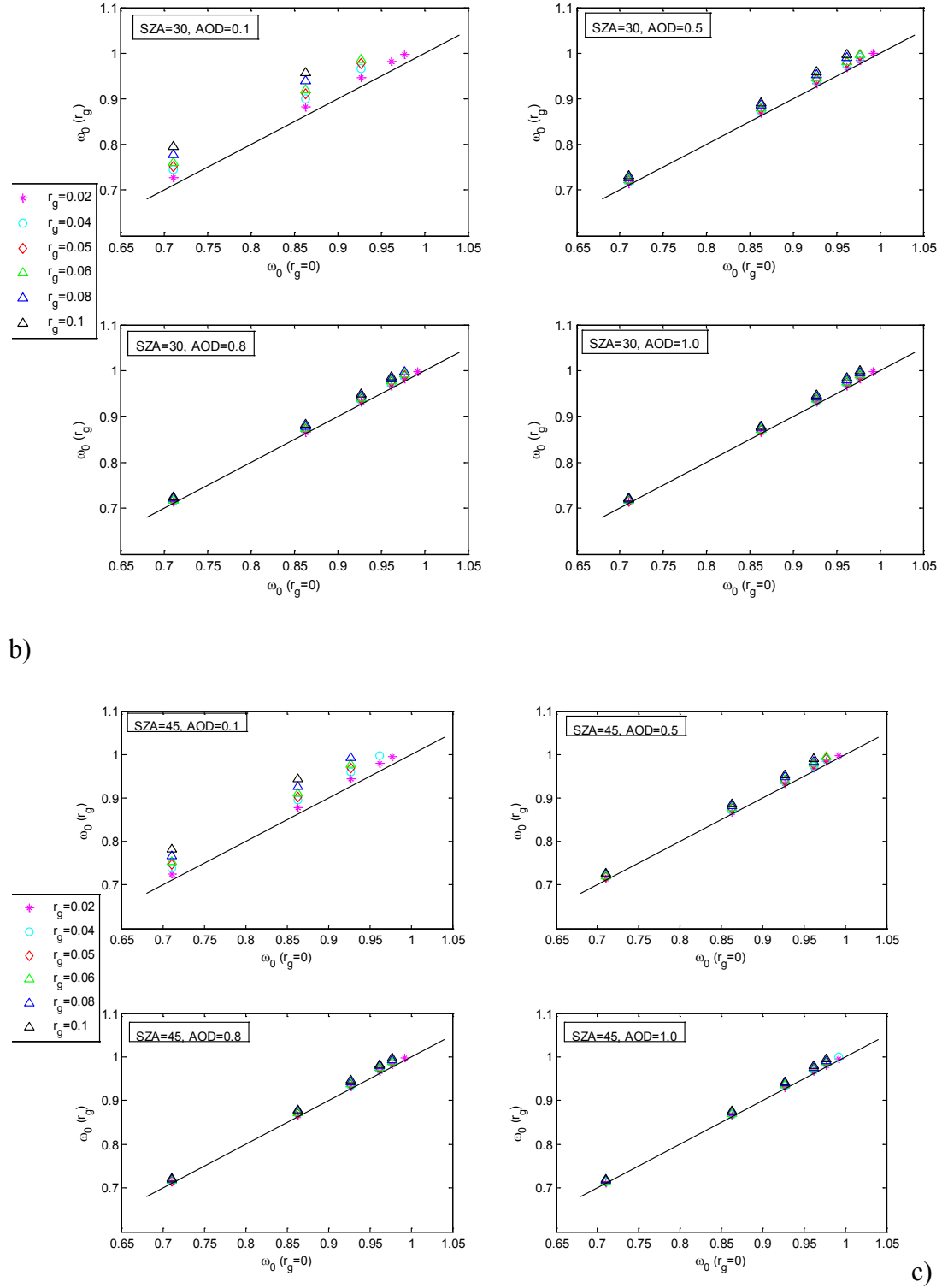


Figure 6.3 Scatter plots of $\omega_0(r_g=0)$ and $\omega_0(r_g)$, where $r_g = 0.02, 0.04, 0.05, 0.06, 0.08$ and 0.1 obtained for different total aerosol optical depths $0.1, 0.5, 0.8$ and 1.0 at 555 nm wavelength at different solar zenith angles SZA = (a) 0 , (b) 30 , (c) 45

6.3 Future work

We have seen that if we have some estimate of errors in the diffuse to direct ratio, we can estimate the correction needed for improving the single scattering albedo ω_0 . However, it would be useful to try to incorporate the aerosol properties and the shadowband geometry directly into the calculation. In this way, we can imagine a scenario that the LUT for the direct / diffuse ratio is recalculated taking into account the correction a-priori.

Further, the work in this direction will be:

- Based on the single scattering approximation, we can estimate the phase function (i.e., the diffuse radiance as a function of angle) assuming that the retrieval of the R_f and v_f parameters are sufficiently accurate and therefore determine an estimate of ΔL based on geometry of the MFRSR shadowing directly.
- We can also see if the full Radiative Transfer (RT) solution accounting for multiple scattering will significantly alter the ΔL correction, which in turn can improve the ω_0 estimate directly. In particular,
- To compare the estimate of ΔL against the empirical correction approach of *Alexandrov et al.*, [2006].
- Also, we observe that for $\tau_a \sim 0.1$, ground reflectance (r_g) ~ 0.1 can be sufficient to contaminate the ω_0 retrieval. Therefore, the estimates of the VIS – SWIR ratios obtained from simultaneous CIMEL and MODIS measurements will be incorporated. In other words, incorporating local MODIS measurements on the

SWIR (i.e., 2130 nm band) and the fairly stable spectral ratios, we can estimate the surface albedo for the MFRSR channels and compensate for another source of error.

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- ❖ MCP Windows NT 4.0
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