Final Report: Summary of Research

Coupling Satellite and Ground-Based Instruments to Map Climate Forcing by Anthropogenic Aerosols

by Professor Robert J. Charlson (P.I.) and Theodore L. Anderson
University of Washington
Seattle, WA 98195

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Abstract

Climate forcing by anthropogenic aerosols is a significant but highly uncertain factor in global climate change. Only satellites can offer the global coverage essential to reducing this uncertainty; however, satellite measurements must be coupled with correlative, in situ measurements both to constrain the aerosol optical properties required in satellite retrieval algorithms and to provide chemical identification of aerosol sources. This grant funded the third year of a three-year project which seeks to develop methodologies for combining spaceborne lidar with in-situ aerosol data sets to improve estimates of direct aerosol climate forcing. Progress under this one-year grant consisted in analysis and publication of field studies using a new in-situ capability for measuring aerosol 180° backscatter and the extinction-to-backscatter ratio. This new measurement capacity allows definitive lidar/in-situ comparisons and improves our ability to interpret lidar data in terms of climatically relevant quantities such as the extinction coefficient and optical depth. Analyzed data consisted of measurements made along the coast of Washington State, in Central Illinois, over the Indian Ocean, and in the Central Pacific. Thus, this research, combined with previous measurements by others, is rapidly building toward a global data set of extinction-to-backscatter ratio for key aerosol types. Such information will be critical to interpreting lidar data from the upcoming PICASSO-CENA[1], or P-C, satellite mission. Another aspect of this project is to investigate innovative ways to couple the lidar-satellite signal

[1] Pathfinder Instruments for Cloud and Aerosol Spaceborne Observations - Climatologie Etendue des Nuages et des Aerosols

with targeted in-situ measurements toward a direct determination of aerosol forcing. This aspect is progressing in collaboration with NASA Langley's P-C lidar simulator.

Background and Rationale

Progress on the problem of climate forcing by anthropogenic aerosols will require global-scale observations, which, in turn, can only be provided by satellite-borne instruments. With this in mind, a joint U.S./French research satellite project, PICASSO-CENA, or P-C, is being developed for a 3-year mission beginning in 2003. Following the recommendation of the U. S. National Research Council (NRC, 1996), this project will feature Light Detection and Ranging (lidar) instrumentation for obtaining quantitative, vertically resolved aerosol information both day and night, over both continents and oceans. The capabilities of spaceborne lidar have been studied previously via deployment on the U. S. Space Shuttle (Winker et al., 1996).

Lidar has become a central technology in current strategies for tropospheric aerosol research because of its demonstrated ability to map aerosol variations throughout the atmospheric column. Its use is complicated, however, by the fact that the lidar signal contains a convolution of two basic optical properties of the aerosol particles: the 180° backscatter coefficient, β_p (m^{-1} sr^{-1}) and the extinction coefficient, σ_{ep} (m^{-1}). A quantitative retrieval of either property requires knowledge their relationship along the laser path. A central goal of this project is to review and extend current knowledge of this relationship, which we express in the usual form of an extinction-to-backscatter ratio, $S = \sigma_{ep} / \beta_p$, with units of sr. (To make this definition unambiguous, the value of S in the Rayleigh limit is $8\pi/3$.)

Like other aerosol optical properties, S can be calculated from Mie Theory. Mie calculations are essential for instrument validation experiments (e.g. Fig. 1) and provide guidance to field measurements by suggesting the range of values to be expected and what the controlling factors ought to be. A study by Ackermann (1998) illustrates the latter point. He used assumptions for aerosol size distribution and refractive index to show that S should vary from about 15 to 75 sr for tropospheric aerosols, with lower values associated with weakly absorbing coarse-mode particles (i.e. seasalt and mineral dust) and higher values associated with small and/or highly absorbing accumulation-mode particles. These Mie calculations were combined with models of aerosol hydration to develop functional relationships between S and ambient relative humidity for several aerosol types.

While valuable, such calculations are inadequate in themselves for two reasons. First, Mie calculations are limited to simple, frequently unrealistic particle morphologies (homogeneous spheres, concentric spheres, and a few variants). Second, this "theoretical" approach to determining S can never be more accurate than the *empirical* knowledge of particle

size and refractive index (the latter from knowledge of particle chemistry) upon which the calculation is based. For these reasons, direct empirical determinations of S are required.

Several methods have been developed to make this measurement; however, as shown in Table 1, below, they have been deployed to date in only a handful of investigations of tropospheric aerosols. Most regions of the Earth and many major aerosol types have no published data at all and a statistically significant database exists for only one location - Tucson, Arizona. This project seeks to expand that data base in preparation for the upcoming PICASSO-CENA satellite mission. The last four entries in Table 1 show the progress that has been made so far.

Progress

This project was originally conceived in terms of three tasks. Progress is therefore grouped under those original headings, although the direction of research has clearly changed in light of the initial findings.

Task 1 Determining lidar ratios from existing in situ optical measurements

This task was superceded by the progress under Task 2. Results from direct, calibrated measurements showed that existing, nephelometric data sets combined with Mie Theory do not offer an accurate means of predicting the lidar ratio. Therefore, our focus has shifted to expanding the data base of direct measurements.

Task 2 Design of a new instrument for directly measuring 180-backscatter

Under this task we designed, built, calibrated, and deployed a new, nephelometer-type instrument for performing direct, calibrated measurements of 180° backscatter and the extinction-to-backscatter ratio, S. Results have now been reported in two published articles (Doherty et al., 1999 and Anderson et al., 2000). A summary of those and plus more recent results in included in this section.

Figure 1 shows the calibration of the device with laboratory particles of known size and refractive index (latex spheres). The continuous lines represent Mie Theory prediction. Measured points are all seen to fall within experimental uncertainties, indicated by horizontal and vertical error bars. Based on this well-calibrated methodology, measurements were made at four sites in different regions of the globe. Results are summarized in Table 1, which serves to document as well the paucity of previous measurements of S.

Task 3. Analyze 180-backscatter optical depth as an index of aerosol climate forcing

This task has evolved into a collaborative project involving the lidar simulator model developed at NASA Langley as well as radiative transfer modeling at the University of Lille, France. Progress to date has involved developing a set of atmospheric aerosol "scenes" to be jointly studied by the lidar simulator and the radiative transfer models, analyzing preliminary results from simple tests, and designing future simulation experiments.

A critical underpinning to these experiments is accurate knowledge of P-C noise uncertainty and how this effects the spatial resolution of aerosol backscatter data. The remainder of this section details progress made in that area.

In simplified form, the lidar equation is,

$$P(r) = \frac{\beta(r)}{C} e^{-2\tau(r)} + P_{bkg} \tag{1}$$

where P(r) is the total power detected by the receiver at a time corresponding to distance r, C is the net normalization constant that accounts for laser energy, system calibration, and the r^2 reduction of power with distance, $\beta(r)$ is the volume 180° backscatter coefficient at distance r, τ is the atmospheric optical depth along the laser path, and P_{bkg} is the background power due to reflected sunlight (day) or moonlight (night). In the absence of clouds, aerosol backscatter, $\beta_a(r)$, is the difference between atmospheric backscatter, $\beta(r)$, and molecular backscatter, $\beta_m(r)$, where the latter is known to the accuracy of molecular density. Similarly, in the absence of clouds overhead, $\tau(r)$ is the sum of molecular and aerosol components - $\tau_m(r)$ and $\tau_a(r)$, respectively. Thus, aerosol backscatter depends on measured power, background power (which is measured between pulses), system calibration, molecular backscatter, and attenuation along the laser path as follows:

$$\beta_a = \frac{P - P_{bkg}}{e^{-2(\tau_m + \tau_a)}} C - \beta_m \tag{2}$$

where the dependence on distance from the laser, r, is ignored for clarity. Given knowledge of β_a, the aerosol volume extinction coefficient, σ_a, is,

$$\sigma_a = \beta_a S_a \tag{3}$$

which is simply a restatement of the definition of the aerosol extinction-to-backscatter ratio, S_a. In general, σ_a and β_a are interdependent, since τ_a in Equation (2) is the path integral of σ_a. An important exception occurs at the top of the first aerosol layer, however, where τ_a=0. In this case, uncertainty in β_a depends only on knowledge of C, P_{bkg}, τ_m, and β_m, while uncertainty in σ_a has an additional component arising from uncertainty in S_a.

Differentiation of equation (2) permits a convenient method of estimating uncertainties. In the equations below, the symbol δ before a quantity is used to indicate uncertainty in that quantity.

$$\delta\beta_{a,Noise} = \delta(P - P_{bkg})e^{2\tau}C \tag{4a}$$

$$\delta\beta_{a,Calib} = (\delta C / C)\beta \tag{4b}$$

$$\delta\beta_{a,Density} = (\delta\beta_m / \beta_m)\beta_m \tag{4c}$$

$$\delta\beta_{a,Atten.} = 2(\delta\tau_m + \delta\tau_a)\beta \tag{4d}$$

Total uncertainty in backscatter is the quadratic sum of the four components listed above.

The quantity $\delta(P\text{-}P_{bkg})C$ in Eq. (4a) is the noise level of the unattenuated signal, which, for Poisson distributed noise will scale according to:

$$\delta(P - P_{bkg})C = \sqrt{N_0^2 \frac{\beta}{\beta_0} + N_{bkg}^2}\sqrt{\frac{1}{n_{bins}}} \tag{5}$$

where N_0 is the noise level associated with backscattering of magnitude β_0 and with a single lidar data bin at maximum resolution (corresponding to 30 m vertical and 333 m horizontal resolution), N_{bkg} is the noise associated with P_{bkg} at this same single-bin resolution, and n_{bins} is the number of bins over which the lidar signal has been averaged.

For attenuation, we assume that the molecular optical depth is known to the same relative accuracy as the local molecular backscatter,

$$\frac{\delta\tau_m}{\tau_m} = \frac{\delta\beta_m}{\beta_m} = \text{cons tan t} \tag{6}$$

Uncertainty in aerosol optical depth, $\delta\tau_a$, is assumed to be entirely due to uncertainty in the extinction-to-backscatter ratio, δS_a, since this is undoubtedly the dominant source of uncertainty. Previous studies with spaceborne lidar data (see Table 1 of Doherty et al., 1999), have shown that $\delta\tau_a/\tau_a$ is proportional to $\delta S_a/S_a$ at small values of τ_a but increases rapidly as τ_a itself increases. These results can be approximated as,

$$\frac{\delta\tau_a}{\tau_a} = \frac{\delta S_a}{S_a}e^{2\tau_a} \tag{7}$$

Finally, uncertainty in aerosol extinction, $\delta\sigma_a$, according to Eq. (3) has components from both β_a and S_a,

$$\delta\sigma_a = \sigma_a\sqrt{\left(\frac{\delta\beta_a}{\beta_a}\right)^2 + \left(\frac{\delta S_a}{S_a}\right)^2} \tag{8}$$

Equations (4)-(8) provide an efficient method of estimating the uncertainties in aerosol backscatter and extinction derived from spaceborne lidar measurements. There are 6 constants to be set. Table 1 lists the values used herein. The parameter n_{bins}, indicating the degree of spatial averaging, is left as a variable. Other adjustable parameters are the profiles of σ_a (or, equivalently, β_a) and S_a. A Standard US Atmosphere is assumed for the molecular profile and only cloud-free conditions are considered.

Case	N_0	β_0	N_{bkg}	$\delta C/C$	$\delta\beta_m/\beta_m$	$\delta S_a/S_a$
	$km^{-1} sr^{-1}$	$km^{-1} sr^{-1}$	$km^{-1} sr^{-1}$			
Day	1.6e-3	1.6e-3	2.4e-3	0.1	0.02	0.4
Night	1.6e-3	1.6e-3	0	0.1	0.02	0.4

Figure 2 shows the calculated uncertainty for backscatter for daytime conditions. The y-axis indicates aerosol amount, expressed as either the ratio of aerosol backscatter to molecular backscatter (on the left) or as the aerosol extinction coefficient for an assumed lidar ratio of 60 sr (on the right). The x-axis indicates the relative uncertainty in knowledge of the lidar ratio, S_a. Three sets of contour lines are given for different amounts of aerosol optical depth above the aerosol layer in question. These optical depth values are: 0.0 (blue contours), 0.1 (red contours), and 0.2 (black contours). The contour lines indicate relative uncertainty in aerosol backscatter coefficient. For zero optical depth aloft, uncertainty in S_a has no effect such that the blue contour lines are flat. However, the contour lines become steeply curved for the cases of significant optical depth aloft. This plot shows that even the backscatter coefficient is well known at the top of the first aerosol layer, but that its uncertainty once the laser has passed through significant aerosol attenuation becomes highly dependent on knowledge of the lidar ratio.

This improved knowledge of noise and total uncertainty in the P-C data products is helping us design, in collaboration with NASA Langley, a correlative in-situ measurement strategy for the P-C mission. We expect to submit this strategy for publication within the next few months.

Refereed publications

Research efforts supported by this grant (that is, the entire three-year effort) are reported in the following articles, published or submitted for publication to refereed journals. Members of the Charlson research group are indicated in bold type.

Anderson, T. L. and Ogren, J. A. (1998), Determining aerosol radiative properties using the TSI 3563 integrating nephelometer, Aerosol Sci. Technol. 29, 57-69.

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Tables

Table 1: Measurements of S for tropospheric aerosols

Method	Location	N[a]	S (sr)
slant-path lidar (b)	mixed layer, west U.S. (Tucson, AZ)	81	8-75
horizontal lidar (c)	Netherlands	10	10-50
horizontal lidar (d)	marine surface layer, Australian coast	10	40-80
multi-wavelength lidar (e)	marine boundary layer (tropical Atlantic)	1	<30
	mixed layer over rainforest (S. Amer.)	1	43-60
	Saharan dust aloft (tropical Atlantic)	2	15-62
backscatter-sonde (f)	rural, arid SW U.S.	12	42
	rural western U.S. lower troposphere	4	15-30
	rural western U.S. upper troposphere	4	15-60
space-borne lidar (g)	smoke layers in SH upper troposphere	7	50-90
Raman lidar (h)	polluted lower tropo. over Northern Germany	2	20-40
Raman lidar (i)	polluted boundary layer over Leipzig, Germany	1	55-95
Raman lidar (j)	rural, lower tropo. over Oklahoma	25	20-80
this work (k)	Northwest Coastal U.S.	3	20-70
this work (l)	Indian Ocean	50	40-90
this work (m)	polluted, Central U.S.	70	30-71
this work (n)	Central Pacific	50	28-32

a Approximate number of independent samples.

b Determines effective S over mixed layer at 694 nm. (Spinhirne et al., 1980; Reagan et al., 1984; 1988)

c 1064 nm laser. No information on location of instrument or type of aerosol investigated. (de Leeuw et al., 1986)

d 532 nm laser beam aimed 2 m over ocean surface from coastal site. Effect of waves not assessed. (Young et al., 1993).

e S at 600 nm is constrained by Mie calculations based on the measured wavelength variation of β_p. (Sasano and Browell, 1989)

f S reported at 690 nm. Method requires significant wavelength and angular adjustments, based on assumed size distributions and Mie calculations. (Rosen et al., 1997a, b)

g S at 532 nm is constrained to values yielding physically plausible lidar retrievals under the assumption of constant S throughout upper troposphere. (Kent et al., 1998)

h S at 308 nm determined from independent extinction measurements using nitrogen-Raman; one profile. (Ansmann et al., 1992)

i S at 532 nm using nitrogen-Raman as above. (Muller et al., 1998)

j S at 351 nm using nitrogen-Raman as above; profiles in lower troposphere over 8 nights (Ferrare et al., 1998)

k S at 532 nm using the nephelometric technique discussed herein. S was ca. 20 during marine flow and 60-70 during continental flow. (Doherty et al., 1999)

l As in k except for airborne measurements during the Indian Ocean Experiment.

m As in k except for surface measurements in Central Illinois.

n As in k except for surface measurements on the windward coast of Oahu.

Figures

Figure 1. Laboratory calibration of in-situ measurement of extinction-to-backscatter ratio, S.

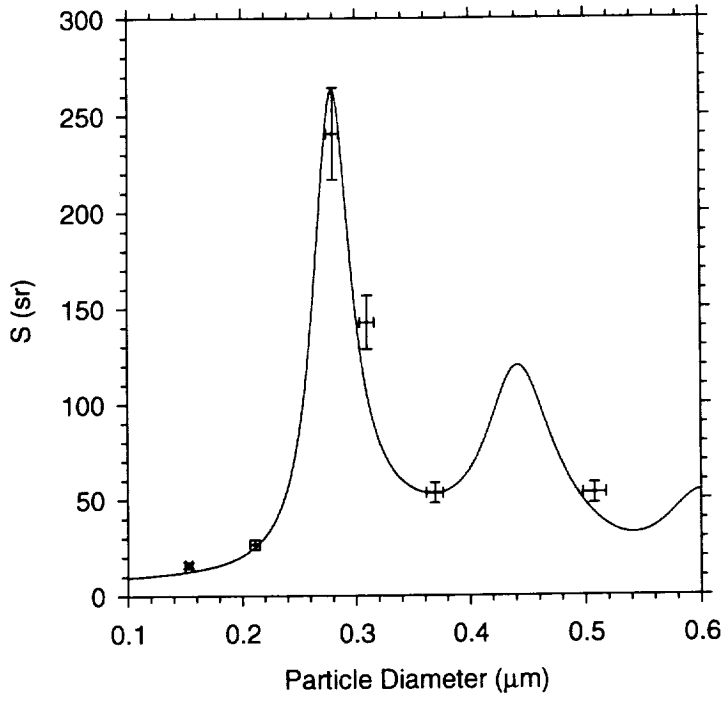

Figure 2: Relative Uncertainty in Aerosol Backscatter: $\delta\beta_a/\beta_a$