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Passive remote sensing of tropospheric aerosol and atmospheric correction for the aerosol effect

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Abstract. The launch of ADEOS in August 1996 with POLDER, TOMS, and OCTS instruments on board and the future launch of EOS-AM 1 in mid-1998 with MODIS and MISR instruments on board start a new era in remote sensing of aerosol as part of a new remote sensing of the whole Earth system (see a list of the acronyms in the Notation section of the paper). These platforms will be followed by other international platforms with unique aerosol sensing capability, some still in this century (e.g., ENVISAT in 1999). These international spaceborne multispectral, multiangular, and polarization measurements, combined for the first time with international automatic, routine monitoring of aerosol from the ground, are expected to form a quantum leap in our ability to observe the highly variable global aerosol. This new capability is contrasted with present single-channel techniques for AVHRR, Meteosat, and GOES that although poorly calibrated and poorly characterized already generated important aerosol global maps and regional transport assessments. The new data will improve significantly atmospheric corrections for the aerosol effect on remote sensing of the oceans and be used to generate first real-time atmospheric corrections over the land. This special issue summarizes the science behind this change in remote sensing, and the sensitivity studies and applications of the new algorithms to data from present satellite and aircraft instruments. Background information and a summary of a critical discussion that took place in a workshop devoted to this topic is given in this introductory paper. In the discussion it was concluded that the anticipated remote sensing of aerosol simultaneously from several space platforms with different observation strategies, together with continuous validations around the world, is expected to be of significant importance to test remote sensing approaches to characterize the complex and highly variable aerosol field. So far, we have only partial understanding of the information content and accuracy of the radiative transfer inversion of aerosol information from the satellite data, due to lack of sufficient theoretical analysis and applications to proper field data. This limitation will make the anticipated new data even more interesting and challenging. A main concern is the present inadequate ability to sense aerosol absorption, from space or from the ground. Absorption is a critical parameter for climate studies and atmospheric corrections. Over oceans, main concerns are the effects of white caps and dust on the correction scheme. Future improvement in aerosol retrieval and atmospheric corrections will require better climatology of the aerosol properties and understanding of the effects of mixed composition and shape of the particles. The main ingredient missing in the planned remote sensing of aerosol are spaceborne and ground-based lidar observations of the aerosol profiles.

1. Introduction

The demand for detailed information on the aerosol spatial distribution and variation with time is increasing much faster than any foreseeable future supply of such information. Aerosols are liquid and solid particles suspended in the air from natural or man-made sources. Aerosol particles affect climate directly by interacting with solar and terrestrial radiation and indirectly by their effect on cloud microphysics, albedo, and precipitation (for review, see *Andreae* [1995], *Charlson and Heintzenberg* [1995]). Tropospheric aerosol forcing is comparable to global net cloud forcing of approximately -1 W m^{-2}. However, on a regional basis the calculated mean aerosol direct radiative forcing caused by mineral dust over the ocean amounts to about -10 W m^{-2} [*Tegen et al.*, 1996]. The relatively strong aerosol forcing is due to the far smaller compensation of solar and terrestrial radiation effects by aerosols as compared to clouds. The effect of aerosol on clouds is caused by soluble submicron particles that serve as cloud condensation nuclei and larger dust particles that are efficient ice nuclei. In both cases the cloud microphysics and properties are affected.

The effect of man-made aerosol on the planetary albedo can

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Table 1. Present Major Satellite Sensors Applicable for Remote Sensing of Aerosol

Sensor/Agency	Launch Date	Spectral Channels λ, μm	Spectral Channels $\Delta\lambda$, nm	Pixel Size, km^2	Remote Sensing Application
AVHRR/NOAA	since 1979	4 bands [0.64] [0.83] [3.75] [11.5]	150 200	1.00 × 1.00 or 4.00 × 4.00	operational remote sensing of τ_a over oceans; Ångström coefficient over ocean; τ_a over land using dense vegetation or contrast effects; ω_0 using spectral or spatial contrast
Landsat-TM/NASA	since 1982	6 bands [0.47–2.20]	20	0.03 × 0.03	τ_a over ocean; τ_a of dust over land using contrast effect
Landsat-MSS/NASA	since 1971	4 bands [0.55–0.90]	100	0.08 × 0.08	τ_a; *Ångström coefficient*
VISSR-GOES/NOAA	since 1975	1 band [0.66]	300	1.00 × 1.00	τ_a
SAGE I, II/NASA	since 1979	7 bands [0.38–1.08]	2–20	limb occultation	aerosol extinct., NO_2, H_2O and O_3 profiles
TOMS-Nimbus 7/NASA	since 1978	2 bands [0.34]–[0.38]			presence of absorbing aerosols
OCTS-ADEOS/NASDA	1996	9 bands [0.41–0.86] and 3.9	20–40 330	0.70 × 0.70	τ_a; size distribution over water
POLDER-ADEOS/ CNES-NASDA	1996	8 bands [0.44–0.91] 3 polarized bands; multiview angles	20	6.00 × 7.00	τ_a; Ångström coefficient; aerosol model
ERBE-CERES-EOS/ NASA	1984/1998	0.3–50 μm; 0.3–5 μm; 8–12 μm			aerosol radiative forcing

delay or temporarily screen the presence of greenhouse warming. In fact, it is suspected to slow the increase in the global temperature in the last century, to decrease the diurnal temperature range, and to decrease the anthropogenic warming in the northern hemisphere relative to the southern hemisphere [*Karl et al.*, 1995; *Hansen et al.*, 1997]. Natural oceanic aerosol, originating from DMS emissions by phytoplankton, may form a feedback mechanism that can reduce future increases of ocean temperature. Thus uncertainty in aerosol science is generating probably one of the largest uncertainties in predicting anthropogenic climate change.

Our understanding of the importance of aerosol to atmospheric and Earth processes is expanding beyond that of sulfate radiative forcing. It is recognized that smoke aerosol [*Liousse et al.*, 1996], black carbon from urban/industrial sources, and dust (natural and due to land use change) are just as important. Aerosols absorb solar radiation, thereby changing the temperature vertical profile. Under some conditions they serve as the surfaces that enhance and change the heterogeneous chemistry of reactive gases [*Taylor et al.*, 1983] like tropospheric ozone. Either dissolved in precipitation, deposited with snow, or directly deposited on surfaces, they form the major fertilizers for many ecosystems. Dust aerosol, originating also from land use change in Africa, and deposited into the Atlantic Ocean, is hypothesized to be a major source of iron used by the phytoplankton that may capture excess CO_2 [*Young et al.*, 1991]. It is also the source of topsoil in Atlantic islands [*Prospero and Nees*, 1986] and the Amazon Basin. The difference between the spatial and the vertical distribution of aerosol from greenhouse gases, resulting from their short lifetime, is expected to introduce even more important climatic effects, from cooling in the North Atlantic region to possible reduction of atmospheric mixing in the tropics. Recently, a comparison of the pattern of predicted global warming with temperature measurements, including sulfate aerosol, indicated for the first time the detection of "fingerprints" of the anthropogenic climate change [*Santer et al.*, 1996].

To understand these aerosol impacts, we need to know the variation of the spatial distribution of aerosol, expressed by the optical thickness or mass concentration. We also need to know the aerosol absorption, their scattering properties, vertical profiles, size distributions, compositions, and surface area. Aerosol chemistry and interaction with water vapor also plays a major role. These properties can have an important diurnal cycle. Frequent global measurements of the variation of the aerosol spatial distribution and some key properties are only achievable by Earth observations from space. The new multinational expanded series of Earth satellite systems, which is an unprecedented effort in human history, is a partial response to this growing demand for applications. It includes measurements of direct relevance to aerosol research.

We shall soon have an array of instruments representing new technology of spaceborne precise monitoring of aerosol from space. Starting with the successful launch of the ADEOS system in August 1996, with POLDER, TOMS, and OCTS on board, to the planned launch of EOS-AM 1 with MISR, MODIS, and CERES in 1998, ENVISAT and ADEOS 2 in 1999, and EOS-PM 1 in 2000, we shall have, still in this century, an array of sensors, with new capability and precise onboard calibration and registration devices, for multiplicative pathways of monitoring aerosol, their properties, and radiative forcing from space. The improved AVHRR and SeaWiFS instruments, with aerosol capability, are planned to be launched shortly. The geostationary GOES and Meteosat satellites supplement the detailed diurnal aerosol cycle. Tables 1 and 2 summarize, respectively, the present and future capability of these satellite systems for monitoring aerosol. European, Japanese, and U.S. sponsored teams of scientists from many countries are working together in designing new techniques of extracting aerosol information from these data. The satellite platforms are accompanied by the first automatic network that monitors aerosol remotely from the ground from tens of locations around the world, for validation of the satellite retrievals and supplementing them with vital information not obtainable from space [*Holben et al.*, 1997]. Plans for systematic measurements from ground-based and spaceborne lidar systems are on the way, though not in this century. Lidars are needed to observe the vertical stratification of the aerosol. New technics

Table 2. Future Major Satellite Sensors Applicable for Remote Sensing of Aerosol

Sensor/Agency	Launch Date	Spectral Channels λ, μm	Spectral Channels $\Delta\lambda$, nm	Pixel Size, km^2	Remote Sensing Application
SeaWiFS-SeaStar/NASA	this issue	8 bands [0.41–0.86]	20–40	1.00 × 1.00 every 4 km	τ_a Ångström coefficient
MODIS-EOS/NASA	1998	12 bands [0.41–2.10] and 3.96	10–20	0.25 × 0.25 1.00 × 1.00	τ_a and size distribution over water
			100		τ_a and ω_0 over land
MISR-EOS/NASA	1998	4 bands [0.47–0.86]	10–20	0.25 × 0.25 1.00 × 1.00	τ_a, size distribution and phase function over water
		9 view angles			τ_a over land
MERIS-ENVISAT/ESA	1998	15 bands [0.40–1.02]	adjustable	0.25 × 0.25 1.00 × 1.00	τ_a and size distribution
GLI-ADEOS II/NASDA	1999	12 bands [0.41–2.10] and 3.75	10–20 100	0.25 × 0.25	τ_a and size distribution

τ_a, aerosol optical thickness; ω_0, single-scattering albedo; λ, central wavelength; $\Delta\lambda$, bandwidth.

to analyze previous records of satellite data, like the UV technic applied to 15 years of TOMS data and AVHRR, are being developed.

Parallel to the effort of remote sensing of aerosol from space, an effort to remove their influence from satellite data used for remote sensing of oceanic and land biota was developed. The use of satellite imagery over land for deriving quantities such as bidirectional reflectance distribution function (BRDF), albedo, vegetation indices, leaf area index (LAI), and fraction of photosynthetically active radiation (FPAR) requires that the signal measured at the top of the atmosphere be corrected for atmospheric effects and converted to surface reflectance. Atmospheric correction of image data requires inputs that describe the variable atmospheric constituents influencing surface reflectances as measured at satellite altitudes and a correct modeling of atmospheric scattering and absorption. Aerosols are among the most variable of these atmospheric constituents. A more complex problem is atmospheric correction over oceans, where most of the signal measured at the top of the atmosphere is composed of photons that have not interacted with the water body. In the blue spectral band, where phytoplankton pigments absorb substantially, typically 90% of the satellite radiance originates from the atmosphere. In the red, where phytoplankton fluoresces, the figure becomes 99.5 to 99.9%. Therefore performing accurate atmospheric correction of satellite radiances over ocean is a formidable challenge, all the more so because the satellite sensors cannot be absolutely calibrated to better than a few percent in terms of measured radiance. It appears logically paradoxical to use the satellite data to derive the aerosol content (by assuming certain properties of the underlying surface) and then to correct the same satellite data for the aerosol effect, using the aerosol information. This paradox is circumvented by using the multidimensionality of the data stream. Depending on the satellite sensor, the dimensions include the spatial dimensions, the spectral, view angle, and optical dimensions. The optical dimension includes radiance and polarization. Several methods have been developed to use these dimensions to separate the aerosol signal from the surface signal. The methods differ from land to ocean because of the difference in their optical properties.

This unprecedented new effort, and the parallel plans for instruments to be flown in the next century (e.g., EOSP), was the reason for the 5 days, highly scientifically stimulating and socially relaxing workshop, in Washington, D. C., during a nice spring week in 1996. Some 50 international experts on remote sensing of aerosol and of atmospheric corrections, who are presently responsible for the development of algorithms for interpretation of data from the new satellite systems and ground-based instrumentation, participated in the workshop. Most of the papers presented in the workshop are given in this special issue. Summaries of the discussions that followed each session and which occurred during the lunches, dinners, and cruise are presented in this paper after a background section on the aerosol properties and their impact on remote sensing. The collection of papers in this special issue describes the different methodologies and technologies that are being used to derive aerosol information from space and to correct the space imagery for the aerosol effect on observations of the surface.

2. Background Information on Aerosol Radiative Properties

Sources and Types of Aerosols

Aerosol particles originate from sources with different properties [*d'Almeida et al.*, 1991]: sea-salt particles from the ocean, wind-blown mineral particles, including desert dust, sulfate, and nitrate aerosols resulting from gas to particle conversion, organic material, carbonaceous substances from biomass burning, and industrial combustions. These particles are generally produced at the Earth's surface and remain located in the boundary layer, or raised to higher altitudes during their transport. For example, the Saharan desert dust observed around 5–6 km above the Atlantic Ocean, or smoke from large fires emitted to 3–4 km height.

A special category of aerosols are the sulfuric acid particles produced by oxidation of sulfur dioxide in the stratosphere. In unperturbed conditions, this so-called "Junge layer" is very tenuous. However, after a strong volcanic eruption injecting a mass of SO_2, the amount of stratospheric aerosol can be increased by 2 orders of magnitude, leading to a contribution to the total optical depth similar to that of tropospheric aerosols [*Stowe et al.*, 1992]. Retrieval of aerosols from remotely sensed data relies on the choice of an "aerosol model"; dealing with two different types of aerosols (one in the troposphere and one in the stratosphere) complicates the problem. Fortunately, the stratospheric aerosol, which makes a rather homogeneous and stable layer, is monitored by spaceborne occultation experiments [*McCormick et al.*, 1979; *Yue et al.*, 1991]. A possible approach would be to introduce them as a known correction term in the remote sensing of tropospheric aerosols, this correction term being only important after volcanic eruptions.

Aerosol characteristics

The aerosol particles are characterized by their shape, their size, their chemical composition, and total amount, which in turn determines their radiative characteristics. Remote sensing relies on the impact of aerosols on backscattering and transmission of radiation by the Earth's atmosphere and therefore relies on the aerosol radiative characteristics [*Lenoble*, 1993]. Establishment of the climatology of these characteristics, for remote sensing and climate assessment is a major objective of the scientific community in the last several decades [*International Aerosol Climatology Project* (*IACP*), 1991].

These characteristics, given for a wavelength λ, include the vertical profile of the scattering σ_s and absorption σ_a coefficients and the scattering phase function. Instead of σ_s and σ_a, one can use the extinction coefficient $\sigma_e = \sigma_a + \sigma_s$ and the single scattering albedo $\omega = \sigma_s/\sigma_e$; for nonabsorbing aerosols, $\omega = 1$. The aerosol depth is defined by

$$\tau_{e,s,a} = \int_0^\infty \sigma_{e,s,a}(z)\, dz, \tag{1}$$

for extinction, scattering, and absorption. The angular distribution of the scattered photons is characterized by the phase function $p(\theta)$, where θ is the scattering angle, between the incidence and the scattering directions. As $p(\theta)$ is normalized to 4π by integration over all directions, it does not depend on the total amount of the particles. If one is interested in the polarization effect of scattering, the scalar function $p(\theta)$ has to be replaced by a 4×4 phase matrix $P(\theta)$.

The phase function (or phase matrix) and the scattering and absorption coefficients depend on the shape, composition, and size distribution of the particles and on the refractive index m ($m = m_r - i\, m_i$) of the individual chemical components; if the particles are not absorbing, the refractive index $m = m_r$ is real; if they are absorbing, m is complex. The imaginary part m_i describes the absorption of each chemical component. For a given aerosol model (shape, size, and refractive index), σ_a and σ_s are proportional to the aerosol concentration.

The liquid particles are generally spherical and submicron, whereas the solid ones have various irregular shapes and are larger. A general assumption so far has been that irregular, randomly distributed particles behave similarly to spherical particles. The principal reason for this assumption was the convenience of using Mie theory. However, with the availability of new methods as the T matrix, the scattering by nonspherical particles, with simple geometry, is the subject of active research. Three papers in this issue are devoted to this topic [*West et al.*, this issue; *Mishchenko et al.*, this issue; *Kahn et al.*, this issue]. A puzzling result is that spheroids, either oblate or prolate, scatter less than spheres in the backward direction and more between 80° and 150°, and therefore randomly oriented spheroids have a phase function different from spheres [*Mishchenko et al.*, this issue]. A similar result is observed for large hexagonal ice crystals. This finding questions the use of spherical particle models for dust-like aerosols.

The particle size impact on radiation is easier to discuss for spherical particles, but the problem is fundamentally the same for other shapes. In this case, there is only one size parameter, the radius r, and the well-known Mie theory shows that the extinction coefficient is given by

$$\sigma_e = \int_0^\infty Q_e(x, m)\pi r^2 n(r)\, dr; \tag{2}$$

Q_e is the Mie extinction efficiency factor, given in terms of the Mie size parameter $x = 2\pi r/\lambda$, and depending on the complex refractive index; $n(r) = dN/dr$ is the number size distribution of the N particles. Although $n(r)$ is generally used to define an aerosol model, it does not carry useful physical information; the surface size distribution $dS/dr = 4\pi r^2 n(r)$, related to the surface available for heterogeneous chemistry, or the volume size distribution $dV/dr = 4\pi r^3 n(r)/3$, related to the total amount of particulate matter, are more informative. Note that in (2) the impact of a particle on extinction is weighted by its geometrical section πr^2 and therefore is represented by the surface size distribution.

If some information can be gained from space observations about the aerosol size distribution, it is obviously limited to the size range which influences the radiative characteristics. From (2), one can guess that the very small particles, although numerous, do not contribute much to σ_e; the large ones contribute more, but generally $n(r)$ decreases faster than r^{-2}, and there is also a limit of detection for large r. Figure 1 shows, at 0.4 and 1.0 μm, the normalized kernels, which weight the volume size distribution in an equation similar to (2). It appears that the contribution in this wavelength domain is approximately limited to particles between 0.08 and 2 μm.

The fine structure of the size distribution has a small effect on radiation; the usual approach in remote sensing is to adopt a simple mathematical form for $n(r)$, with adjustable parameters, which are sought from observations. The most often used are the Junge distribution,

$$\begin{aligned} n(r) &= Cr^{-\nu} & r_1 < r < r_2 \\ n(r) &= 0, & r < r_1,\ r > r_2 \end{aligned} \tag{3}$$

where ν is the adjustable parameter, and C is a constant related to the total number of particles; and the lognormal distribution (LND)

$$n(r) = \frac{N}{\sqrt{2\pi} r \ln \sigma} \exp\left[-\frac{\ln^2 r/r_m}{2 \ln^2 \sigma}\right], \tag{4}$$

with two adjustable parameters r_m and σ.

More refined models use segmented Junge distributions, or bimodal LNDs. It is not easy to distinguish between these distributions from remote sensing observations. *Tanré et al.* [this issue] state that the spectral radiance of an aerosol with a bi-LND size distribution can be very well simulated by a single LND with appropriate mean radius and width. *Hansen and Hovenier* [1974] have shown that the dominant parameter for the radiative effect of aerosols is the effective radius r_{eff}, defined as an average radius weighted by $\pi r^2 n(r)$,

$$r_{\text{ef}} = \int_0^\infty r^3 n(r)\, dr \bigg/ \int_0^\infty r^2 n(r)\, dr; \tag{5}$$

however, this may not be true when the size distribution is wide or bimodal. It is an open problem to clarify how many parameters of the size distribution are needed to define the aerosol radiative characteristics and how many can actually be retrieved from satellite observations alone.

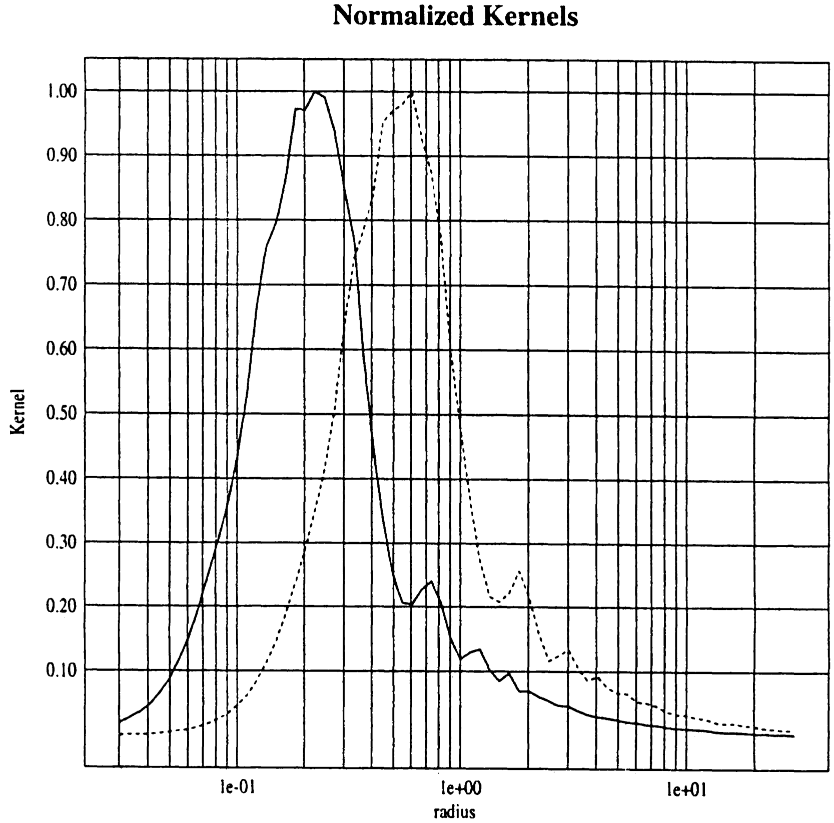

Figure 1. Normalized kernel function for the volume size distribution, at 0.4 μm (solid curve) and at 1.0 μm (dotted curve); the refractive index is 1.45.

A Brief History of Aerosol Remote Sensing

Aerosols can be observed and analyzed, either in situ by impactors or particle counters, or from a distance by active (lidars) or passive remote sensing instruments. Passive remote sensing is based on the modification of the solar radiation field (or in some cases of the terrestrial longwave radiation) induced by the aerosol particles; it has been used for decades and is still used from ground-based stations. Extending these remote sensing techniques to satellite instruments was very appealing, with the objective of obtaining a global view of the atmospheric aerosols.

The simplest ground-based observation is the extinction of the direct solar beam, and its satellite counterpart is the solar occultation method, unfortunately restricted to the stratosphere and high troposphere aerosols [*McCormick et al.*, 1979]. The sky radiance and the sky polarization [*Sekera*, 1956] also contain information on the aerosols and are useful to complement the extinction measurements.

It was early recognized that the solar radiation backscattered to space was influenced by the aerosol and could be analyzed by methods similar to those used for the analysis of sky radiance, although the surface reflectance introduces a further difficulty [*Fraser*, 1964]. The first applications of satellite remote sensing of aerosols began in the mid-1970s and concerned the detection of desert particles above the ocean [*Fraser*, 1976; *Mekler et al.*, 1977; *Norton et al.*, 1980; *Griggs*, 1979; *Carlson and Wendling*, 1977); they used Landsat, GOES, and AVHRR data. Simulations were performed by *Koepke and Quenzel* [1979, 1981] in order to define the optimum viewing geometry and the optimum wavelengths. References of papers published on the issue before 1989 can be found in the work of *Stowe et al.* [1990]. More recent work is reviewed by *Kaufman* [1995].

The correction of the aerosol influence on the remote sensing of ocean color was considered by *Gordon [1978]. A* major application of satellite observations is providing detailed images of the land surface, and removing the blurring effect of the atmosphere was the subject of several studies [*Tanré et al.*, 1979; *Otterman et al.*, 1980]; it opened the way to aerosol detection over land surfaces.

Till recently, only data from instruments not at best designed for aerosol studies were available. The present major breakthrough is the availability of new spaceborne instruments designed also for the remote sensing of aerosol. A consequence is that not only the aerosol optical depth but also a more detailed characterization of the aerosols can now be sought.

Basic Principles of Aerosol Remote Sensing

The simplest case is remote sensing above a black surface, where the observed radiance at the top of the atmosphere (TOA) L^{TOA} is only due to photons backscattered by the atmosphere. For the sake of simplicity we omit the wavelength subscript in what follows. We define the atmospheric reflectance as

$$R^{ATM}(\mu, \phi, \mu_0, \phi_0) = \pi L^{TOA}(\mu, \phi, \mu_0, \phi_0)/\mu_0 f, \quad (6)$$

where f is the extraterrestrial irradiance; (μ_0, ϕ_0) refer to the Sun direction and (μ, ϕ) to the observation direction; μ is the cosine of the zenith angle, and ϕ is an azimuth angle. The radiance $L(\tau; \mu, \phi)$ satisfies the classical equation of radiative transfer

$$\mu \frac{dL(\tau;\ \mu,\ \phi)}{d\tau} = L(\tau;\ \mu,\ \phi) - \frac{\omega(\tau)}{4\pi} \cdot \int_0^{2\pi} \int_{-1}^{+1} p(\tau;\ \mu,\ \phi;\ \mu',\ \phi') L(\tau;\ \mu',\ \phi')\, d\mu'\, d\phi'; \quad (7)$$

where τ is the optical depth, above an altitude z, due to the molecules (Rayleigh scattering) and aerosol particles; $p(\tau;\ \mu,\ \phi;\ \mu_0,\ \phi_0)$ is the phase function (molecules and aerosols) at the depth τ for the scattering angle between the directions $(\mu_0,\ \phi_0)$ and $(\mu,\ \phi)$. We have omitted in (7) the parameters of the Sun direction. The boundary conditions are given by the incoming solar radiation at the top of the atmosphere and by the black ground surface (no illumination at the bottom of the atmosphere).

The molecular composition of the atmosphere is known, thus if the aerosol radiative characteristics are also known at all levels, solving (7) gives the radiance as a function of altitude and direction, including the upward radiance at the top of the atmosphere $L^{\mathrm{TOA}} = L(\tau = 0;\ \mu > 0,\ \phi)$. This is the "direct" or "forward" problem.

The observed quantity in remote sensing is the TOA radiance, or the atmospheric reflectance, and the sought parameters are the aerosol characteristics. This is a much more complex "inverse" problem. Inverse problems are generally ill-conditioned, and a lot of work has been devoted to this particular aspect of physical mathematics [*Twomey*, 1977]. Not all the aerosol characteristics can be retrieved, and there is a hierarchy between the importance of the various parameters in the observed reflectance; the first parameter sought from limited observations is the optical depth, assuming a given aerosol model. When multiangle and/or multiwavelength data are available, possibly with polarization, one can hope to extend the retrieval to include also the size distribution, the refractive index, or even the shape of the particles.

Remote sensing of aerosol above the oceans in the red and near infrared illustrates the case of a black surface at the bottom of the atmosphere. The first maps of aerosol optical depth were produced operationally above the oceans by NOAA using the 0.63 μm channel of AVHRR. *Stowe et al.* [this issue] present the second-generation algorithm for AVHRR products and the improvement expected from a second channel in the near infrared. *Nakajima and Higurashi* [this issue] show the oil fire smoke detected by AVHRR in the Persian Gulf region in 1991; with channels 1 and 2 of NOAA 11 AVHRR, they retrieve the aerosol optical depth and the size parameter of a Junge distribution; a comparison with ground-based measurements leads to a retrieval of the absorption. *Tanré et al.* [this issue] present an algorithm for retrieving the aerosol optical depth and two aerosol size characteristics (the relative concentration of the two modes and the mean radius of the main mode) from MODIS. *Leroy et al.* [this issue] present the POLDER algorithm; they retrieve the optical depth and its spectral variation in the near-infrared channels; the polarization allows them to derive the refractive index. *Kahn et al.* [this issue] expects to retrieve information about the particle nonsphericity from MISR.

When the Earth's surface is not black, photons reflected by the surface as well as photons backscattered by the atmosphere reach the instrument. Assuming the simplest case of a uniform Lambertian surface with a reflectance R^{SURF}, the measured reflectance can be expressed as

$$R^{\mathrm{MEAS}}(\mu,\ \phi;\ \mu_0,\ \phi_0) = R^{\mathrm{ATM}}(\mu,\ \phi;\ \mu_0,\ \phi_0) + T^{\mathrm{ATM}}(\mu_0)T^{\mathrm{ATM}}(\mu)R^{\mathrm{SURF}}/(1 - R^{\mathrm{SURF}}R^{\mathrm{ATM}}), \quad (8)$$

where $T^{\mathrm{ATM}}(\mu_0)$ and $T^{\mathrm{ATM}}(\mu)$ stand for the total transmittance (direct and diffuse) of the atmosphere from the Sun to the surface and from the surface to the instrument, respectively; T^{ATM} is computed from the downward radiance at the bottom of the atmosphere obtained from the solution of (7); R^{ATM} is the atmospheric reflectance integrated over both directions. The factor $1/(1 - R^{\mathrm{SURF}}R^{\mathrm{ATM}})$ in the third term of (8) expresses the multiple reflections between the surface and the atmosphere which converge as a geometric series; it can be neglected when both R^{ATM} and R^{SURF} are small. Decoupling the two contributions of the surface and of the atmosphere is not easy, even if the surface reflectance is perfectly known, which is not generally the case. In the absence of aerosol the instrument measures the surface reflectance, and it is the difference of reflectance between R^{MEAS}, in the presence of aerosols, and R^{SURF}, which contains the required information about the aerosol loading. Because of compensating effects, this difference is very small for bright surfaces [*Fraser and Kaufman*, 1985].

If the surface is non-Lambertian, its bidirectional reflectance distribution function (BRDF) affects the interactions between the surface and the atmosphere [*Lee and Kaufman*, 1986]. If it is nonuniform, the apparent reflectance of the target pixel contains a contribution from the surrounding pixels, weighted by their distance from the target, and depending on the atmospheric scattering; this leads to a blurring effect, reducing the contrasts, which can be used for detecting the aerosols [*Tanré et al.*, 1988]. Equation (8) is no more valid in the general case of a nonuniform, non-Lambertian surface; the complete expressions are given by *Vermote et al.* [this issue].

Remote sensing over land surfaces, which are generally not black, is much less advanced than above the oceans, despite its importance. *Kaufman et al.* [this issue] use MODIS mid-infrared channels (2.13 and 3.75 μm), where the aerosol contribution is small to identify dark pixels, estimate their reflectance at 0.47 and 0.66 μm, and then derive the aerosol optical depth and an information on the aerosol type. *Herman et al.* [this issue (b)] propose to use the polarized radiance from POLDER to observe aerosols over land surfaces; they base their argument on ground-based sky measurements, which are well represented by Mie theory, and on data of the POLDER airborne version, which exhibit a rather small ground polarization. *Martonchik* [this issue] presents results of an airborne instrument simulating MISR type imagery and uses the MISR algorithm to retrieve the aerosol optical depth and the surface reflectance.

TOMS on Nimbus 7 has observed the solar backscattered radiation in several ultraviolet channels during 14 years; its main objective is the retrieval of ozone, but using its 340 and 380 nm channels, outside the ozone absorption band, it is possible to sense the presence of absorbing aerosols [*Herman*, this issue (a)]. TOMS has detected biomass burning particles and followed their large-scale transport.

The paper presented by *Ackerman* [this issue] is the only one in this special issue to consider the detection of aerosol based on their impact on the longwave terrestrial radiation using the MODIS observations at 8.5, 11, and 12 μm.

Atmospheric Corrections

Atmospheric corrections are very different for ocean and land surfaces. Above the oceans the surface reflection consists of the surface specular Fresnel reflectance, dominant in the sunlight direction; foam or whitecap reflectance; and diffuse reflection due to photons backscattered by particles suspended in the water. The last term carries information about the water content. In the red and near infrared the water strong absorption allows almost no light penetration and no backscattering; this allows, outside the glitter and the whitecap areas, the retrieval of aerosols above the ocean surface. The water diffuse reflectance is maximum in the blue or green, depending on the water turbidity and chlorophyll concentration, resulting in the "ocean color" used for marine biology studies. This water contribution to the reflectance observed at the top of the atmosphere has to be separated from the atmospheric contribution: the Rayleigh scattering and aerosol contributions. Rayleigh scattering is well known, and aerosol contribution can be obtained from the red and near-infrared channels. This should, in principle, solve the problem; unfortunately, the aerosol retrieval is neither complete nor perfect, and as the atmospheric correction is often much larger than the water reflectance, it needs to be evaluated quite accurately. The main question addressed in all the studies on atmospheric correction of ocean color is how to best extrapolate the near-infrared aerosol observations to the short wavelengths and avoid pure surface contribution (Fresnel and foam reflectance). *Gordon* [this issue] reviews the basic concepts of atmospheric correction over the oceans as they were first developed for CZCS; with the increased sensitivity of the new instruments, improved algorithms are necessary; details of the algorithms currently developed for SeaWiFS, MODIS, and MISR are presented and discussed. *Fraser et al.* [this issue] describe an ocean-atmosphere model used to build lookup tables to derive the water-leaving radiance. *Fukushima and Toratani* [this issue] analyze the impact of Asian desert dust on CZCS data.

Over the land there is no straightforward method to separate the surface and the atmospheric contributions to the TOA radiance. The algorithms proposed try to use the multiangle and multichannel data to retrieve simultaneously the atmosphere and surface parameters. *Leroy et al.* [this issue] gives a detailed description of the POLDER algorithm. *Vermote et al.* [this issue] do the same for MODIS. *Wanner et al.* [this issue] present the algorithm to be used in producing a global BRDF/albedo product from MODIS and MISR combined data.

Validations

The aerosol products derived from observations are validated by ground-based remote sensing measurements, which can be completed by particle counting and chemical analysis; profiles can be provided by lidars, or by airborne instruments. The most widely used instrument is the ground-based Sun photometer, which measures the extinction of the direct solar beam through the atmosphere and therefore the aerosol optical depth; it is generally equipped with narrowband filters and provides directly the spectral variation of the optical depth, which contains information about the aerosol size distribution. More information about the aerosol characteristics can be retrieved from radiative transfer inversions of sky radiance and polarization measurements.

Because the tropospheric aerosols are highly variable, the ground-based observations must be performed as close as possible in time to the satellite passage. The Aerosol Robotic Network (AERONET) of ground-based Sun/sky photometers [*Holben et al.*, 1997] is used on a continuous basis for POLDER [*Bréon et al.*, this issue] and MODIS validations [*Kaufman et al.*, this issue].

Two papers concern directly with the validation of sea surface observations. *Clark et al.* [this issue] uses a complete set of ship-based and buoy-based instrumentation for MODIS validation; beside Sun/sky photometers, they include measurements of the water-leaving radiance, of in situ radiance or irradiance, and of phytoplankton pigments. *Kishino et al.* [this issue] describe a moored optical buoy system for OCTS validation; it comprises in situ radiance and phytoplankton measurements. In the following we shall summarize the discussions conducted and conclusions reached in the workshop.

3. Discussion on Remote Sensing of Aerosols

Evaluation of the new aerosol remote sensing techniques use radiative transfer simulations or analysis of results from present sensors and aircraft data. However, this assessment is partially hampered by the lack of a full sensitivity study, using, for example, a principal component analysis [*Tanré et al.*, 1996] for planned multiwavelength, multiangle, polarization detection sensors for upwelling solar radiation over land or ocean. The basic, unknown optical parameters of aerosols that should be determined are (1) aerosol column concentrations, (2) optical thickness τ_a as a function of wavelength λ, (3) aerosol phase function (or phase matrix) as a function of λ, and (4) single-scattering albedo as a function of λ. Equivalently, the following physical properties may be determined: (1) size distribution, (2) refractive index or chemical composition, and (3) shape of particles. Other atmospheric ingredients, such as water vapor, may also have an effect on the retrieved parameters. Furthermore, the reflective properties of the underlying surface must be known.

Status of Algorithms

All the methods require some assumptions, to specify one or more of the following: aerosol size distribution, aerosol height profile, aerosol composition/index of refraction, and surface reflectivity properties. The required assumptions depend on what is being retrieved and what measurements are made. A single measurement retrieval is limited to retrieving a singular parameter; that is, the aerosol optical thickness and other required inputs must be assumed. Multiple measurement retrievals generally seek to obtain additional information, such as an estimate of the aerosol size distribution and/or the complex index of refraction or single-scattering albedo.

Because of the typical size range of aerosol particles, the usable range of wavelengths of light for remote sensing of aerosol particles is mostly restricted to the solar radiation above 300 nm, due to ozone absorption at shorter wavelengths. Gaseous absorption limits the choice of the spectral bands in the longer wavelengths. Four basic methodologies were identified to determine one or more of the optical properties (see Tables 1 and 2): (1) a single measurement (present AVHRR, GLI-land, GOES, Meteosat, MODIS-land, TOMS), (2) multispectral measurements (advanced AVHRR, GLI-ocean, MERIS, MISR, MODIS-ocean, OCTS, POLDER, SeaWiFS), (3) multiangle measurements (EOSP, MISR, POLDER), and (4) polarization (EOSP, POLDER). Various combinations of these basic techniques may also be used. The four measurement techniques do not yield vertical profile information. This may be obtained from the following measurement schemes: (5) active lidar, (6) limb scanning, and (7) occultation. Techniques 6 and 7 are presently operational for stratospheric aerosols, but their use is limited in the troposphere because the extinction along the path becomes very large, even at high altitudes, and the lower troposphere is generally blocked by clouds along the long slant path. Since the next generation of satellites, at least until 2001, do not include active sensors, the vertical profile in the troposphere is out of reach at present.

Sensor Group 1: Multiwavelength Near Nadir View

The retrievable aerosol parameters depend in this case more on the covered width of the entire spectral domain than on the spectral resolution of the different channels. The darker the surface the higher the accuracy of the retrieved aerosol optical thickness τ_a. Several sensors, whose specifications are mainly driven by the application of water color measurements over oceans, will soon be in space (MERIS, OCTS, SeaWiFS). While SeaWiFS and MERIS are restricted to wavelengths <1 μm, these sensors will be improvements of the CZCS for water color measurements and aerosol optical thickness estimates, with an envisaged accuracy of $\Delta\tau_a = 0.03$–0.05. In order to improve aerosol information over land, MERIS has two additional narrowband ($\Delta\lambda = 2.5$ nm) channels separated by only 10 nm in the oxygen-A absorption band (mainly used for cloud height) that will be used to estimate the fraction of aerosol optical thickness residing in the planetary boundary layer (PBL). The wavelength dependence of aerosol optical thickness in the visible and near infrared, which indicates the aerosol size distribution type, is retrievable over land by sensors like MODIS and GLI, having spectral channels up to 2.2 μm. It uses the fact that the very dark pixels over forests (at 470 and 670 nm) or lakes (at 670 and 865 nm) with spatial resolution ($\geq$250 m) contain mostly radiance backscattered by aerosols. The retrievable aerosol information is as follows: (1) total aerosol optical thickness estimates in the red (670 nm) over land by all sensors with $\Delta\tau_a \approx 0.05 \pm 20\%$ using the dark pixel approach; (2) total aerosol optical thickness estimates for $\lambda \geq 670$ nm over ocean by all sensors with $\Delta\tau_a \approx 0.03$–$0.05$; (3) wavelength dependence of extinction in the blue (470 nm) and red (670 nm) using channels in the mid-IR ($\lambda \geq 2.10$ μm) for estimating the land surface reflectance (example: MODIS); (4) aerosol size information over the oceans; (5) PBL portion of aerosol optical thickness by sensors with more than two channels in O_2 absorption bands (example: MERIS); (6) no detection of aerosol absorption is planned, though some methods were suggested [*Kaufman*, 1987; *Ferrare et al.*, 1990].

Sensor Group 2: Multiangle, Multiwavelength View

The additional information by viewing the same surface area under different angles allows, in principle, the disentangling of surface reflection characteristics and aerosol backscattering. The sensors MISR on EOS-AM 1 and POLDER (without its polarization capability) on ADEOS can be used to derive, in addition to the total aerosol optical thickness, its spectral dependence if either surface reflection characteristics are well known (e.g., over the ocean) or the viewing geometry covers large parts of the aerosol-scattering function and the bidirectional reflectance function of the surface. No thorough sensitivity study has yet been performed. As shown by the evaluation of TOMS channels in the UV-A portion of the spectrum [*Herman et al.*, this issue (a)], high-scattering optical thickness by molecules allows the detection of absorbing aerosol. The multiangle instruments, in their shortwave channels at very long slant paths (thus reaching high optical thickness), also contain information on aerosol absorption, which may be best retrievable from limb-darkening effects. The retrievable aerosol information is (1) spectral aerosol optical thickness in the visible and NIR, (2) aerosol size information, (3) aerosol absorption might be retrievable in the blue portion of the visible spectrum (especially important for mineral dust).

Sensor Group 3: Multiangle, Multiwavelength With Polarization

The capability to measure the degree of polarization of upwelling radiances facilitates the separation of surface reflection and aerosol backscattering, since over land only Rayleigh scattering and aerosol backscattering in optically thin atmospheres contribute strongly to polarization.

POLDER, in addition to its several viewing angles (up to 14) and 9 spectral channels from 443 to 910 mm, has also three channels with a polarizing filter in front. This offers, in principle, a significant enhancement over multiwavelength, multiangle scanners alone. Whether the strongly varying geometry and the dependence of the degree of polarization on the nonsphericity of the particles allows the disentangling of aerosol information even for aspherical mineral dust remains open. However, the total optical thickness of the aerosol plus an improved spectral dependence of optical thickness will enhance the basic aerosol information derivable over land surfaces. For certain angle intervals the aerosol phase function can be retrieved for near-spherical particles or small particles, thus size distribution information might exist for certain latitudes and seasons. The partial separation of surface reflection and aerosol backscattering achievable through the measurement of polarization also might offer a way to retrieve aerosol absorption estimates at longer wavelengths than for the multiangle, multiwavelength sensors. The retrievable aerosol information is (1) spectral aerosol optical thickness in the visible and NIR, (2) aerosol size information, (3) aerosol refractive index, (4) aerosol phase function for certain angle intervals depending on latitude and season, (5) aerosol absorption might be retrievable at longer wavelengths than for sensors without polarization measurements.

Issues and Concerns

Better use of existing information. An assessment of the aerosol radiative forcing over land requires a complete description of the spatial and vertical distribution of their spectral extinction and absorption coefficients and of the phase functions. We are far from getting these observations entirely from satellite sensors. However, together with already known land use characteristics and aerosol type information derived from ground-based remote sensing, we should be soon able to extract global aerosol information from satellites over land, extending our understanding beyond that already demonstrated over oceans by the AVHRR. The immediate task is therefore to conduct more comprehensive sensitivity studies for multiwavelength, multiangle sensors currently under construction, thereby including the assessment of their potential to give information on absorption.

Combined surface-aerosol retrievals. We should no longer try to separate atmospheric retrievals from corrections of surface reflection effects but rather evaluate spectral radiances for a combined surface reflection and aerosol characteristics retrieval, applying inverse modeling techniques and using the full knowledge already available for deriving these characteristics. This would soon lead to the buildup of a land reflection characteristics data set so urgently needed for atmospheric correction for the large remote sensing community deriving land use classifications.

International effort. The full information contained in the different data sets from different sensors and satellites available in the near future can only be retrieved by an international effort, led by a global change research program with coordination between space measurements and surface-based networks.

Synthesis data set. We propose to construct a synthesis data set, a time series of spectral aerosol optical thickness and other properties with global coverage. This data set should, as demonstrated for other climate variables by the GEWEX experiment, merge in situ networks like the spectral Sun photometers at GAW stations and the AERONET spectral Sun/sky radiometers with several satellite sensors. The data set initiative should be coordinated by scientists from the International Global Atmospheric Chemistry (IGAC) core project of IGBP and GEWEX of WCRP.

Addition of an active sensor. The full aerosol information needed for climate change assessments must come from a combination of radiometers and a profiling active (lidar) aerosol instrument. Knowledge of the vertical heating rate profile due to aerosols and thus the realistic forcing of climate by aerosols is accessible through a combination of passive and active aerosol sensors. There is an urgent need for the latter.

4. Discussion on Atmospheric Corrections Over Land

A major effort in atmospheric correction is underway for several of the new satellite instruments (POLDER, MISR, and MODIS) for retrieval of atmospheric aerosol parameters that will facilitate atmospheric corrections in near-real-time [*Vermote et al.*, this issue; *Kaufman and Tanré*, 1996; *Martonchik*, this issue]. The availability of aerosol information over much of the globe on a nearly daily basis also holds considerable promise for the routine atmospheric correction of other sensors that will retrieve surface reflectances but not aerosol parameters. Routine atmospheric corrections have the potential to increase significantly the mass-market uses of quantitative Earth observation products from a wide variety of sensors [*Teillet*, 1995]. Progress in atmospheric correction over land has also led to the inclusion of the atmospheric point spread function (for higher spatial resolution imagery) and the coupling of surface reflectance anisotropies and atmospheric effects.

Status of Algorithms

Choice of radiative transfer code. Although there is presently no standardization as to which atmospheric radiative transfer code should be used, most of the predominantly used codes tend to disagree significantly only for large aerosol optical thicknesses and large off-nadir angles [*Lenoble*, 1985; *Royer et al.*, 1988]. The proper use of a given atmospheric code should therefore be of greater concern than which code to use. The choice of code is a factor in the correction of imaging spectrometer data for atmospheric gas absorption [*Staenz et al.*, 1994].

Computation issues. Radiometric and atmospheric corrections typically require complex, pixel specific computations. Consequently, software algorithms usually include simplifying approximations and extensive use of lookup tables for more rapid computations. Thus decisions have to be made with respect to input/output/interpolation implications in addition to the standard trade-off between greater accuracy and CPU time. Special attention needs to be devoted to the interplay between the different data sets involved in atmospheric correction, including the atmospheric, the bidirectional reflectance distribution function (BRDF), and surface reflectance products, as well as ancillary data sets such as cloud masks, land cover, average climatologies, and digital terrain elevation models. The key issues encompass product level, data flow, timeliness, differences in spatial resolution, girding, and formatting.

Filling aerosol product gaps. The main processing issue concerns the filling of spatial and temporal gaps in the aerosol products needed for atmospheric correction over land. Interpolations of available aerosol data sets will be used to fill smaller gaps (up to 50 km), and aerosol parameters based on climatological averages will be used to drive atmospheric corrections in larger gaps. Thus there is a definite requirement for comprehensive, up-to-date, and readily accessible global climatologies for aerosols. Accuracy assessment or figure-of-merit products should also accompany both the aerosol and the surface reflectance products.

Consistency of assumptions. Atmospheric correction involves assumptions in the absence of specific knowledge about aerosol model characteristics. These assumptions are needed for the satellite-based aerosol retrievals, the ground-based aerosol retrievals used for validation, and the atmospheric computations in the image correction itself. Some level of consistency in these assumptions should be sought; for example, the same aerosol model should be used to derive the optical thickness τ_a from satellite data and for actual correction. It is recognized that a lot can be learned from the use of a diversity of methods for the different instruments.

Issues and Concerns

BRDF-atmosphere coupling. The surface of the Earth provides a lower boundary for the atmospheric radiation field. This lower boundary is characterized by the directional behavior of its reflectance, usually described in terms of the bidirectional reflectance distribution function of the surface. The BRDF function can only be derived from atmospherically corrected reflectance data, but performing that correction requires knowledge of the surface BRDF. Studies conducted by the MODIS, MISR, and POLDER science teams show that BRDF effects are large enough to require their inclusion in accurate atmospheric correction algorithms [*Leroy et al.*, this issue; *Vermote et al.*, this issue; *Wanner et al.*, this issue].

In practice, the multiangular information needed to construct the BRDF is seldom available in a timely manner. Remote sensing observations sampling different regions of the viewing and illumination hemispheres have to be accumulated over 2–4 weeks. If atmospheric correction is to proceed in near-real-time fashion, a decision has to be made as to how to supply the necessary BRDF information. In deciding upon a method, timeliness, the amount of ancillary data required, and the accuracy of the resulting coupling parameters have to be considered [*Vermote et al.*, this issue; *Wanner et al.*, this issue]. Either the current BRDF has to be approximated using a recent retrieval or the information about surface physical and optical properties needs to be available to allow reliable physical modeling of the BRDF from a limited number of samples.

The situation is partially alleviated somewhat for sensors such as POLDER and MISR that acquire strings of multiangular data almost simultaneously.

Adjacency effects. The atmospheric point spread function leads to adjacency effects, especially for satellite and aircraft imagery acquired at spatial resolutions of 250 m or finer. Performing an atmospheric correction on such high-resolution imagery with high spatial contrast (e.g., scenes containing coastlines, spotty vegetation, etc.) and ignoring adjacency effects can produce surface reflectance errors that are even larger than errors due to uncertainties in the aerosol properties [*Diner et al.*, 1989]. For resolutions of 1 km and coarser, however, uncertainties in the aerosol properties tend to be the dominant error source in the correction procedure.

Single-scattering albedo. The single most important aerosol parameter that will not be well captured by planned satellite systems is the single-scattering albedo. Since this parameter is intimately coupled to the aerosol phase function, MISR, in particular, with its wide coverage of scattering angle in the span of a few minutes, has the potential to at least discriminate between strongly absorbing and relatively nonabsorbing aerosols. These indirect determinations of the single-scattering albedo are probably adequate for most atmospheric correction applications but may not be accurate enough for other disciplines such as radiative balance and atmospheric heating studies.

Cirrus clouds. The new 1.38 μm spectral channel on MODIS [*Gao and Kaufman*, 1986] will allow the identification of thin cirrus clouds not detectable in other spectral bands. Consequently, MODIS algorithms are being developed to correct for thin cirrus and flag the relevant pixels as having undergone an experimental correction. This cirrus capability will undoubtedly be added to future Earth observation satellites whenever possible (e.g., GLI).

5. Discussion on Atmospheric Corrections Over Ocean

Satellite ocean color imagery is contaminated by atmospheric scattering/absorption and surface reflection. This contamination must be removed to retrieve water-leaving radiance, the signal that contains information on water composition, and in particular, biomass content. The process is called atmospheric correction, even though it corrects also for surface effects. Atmospheric correction is a prerequisite to any quantitative utilization of the satellite data in biogeochemistry applications and global change studies.

Status of Algorithms

Despite difficulties and inherent limitations, CZCS, the first ocean color sensor to fly aboard a satellite, demonstrated, using atmospheric corrections, the feasibility of measuring phytoplankton pigments from space on a global scale. Since the CZCS, several sensors with improved ocean color capabilities have been designed, and some are now orbiting the Earth. They include OCTS and POLDER on ADEOS, SeaWiFS on SeaStar, MODIS and MISR on EOS-AM 1, MERIS on ENVISAT, and GLI on ADEOS 2. These sensors have characteristics that will allow large-scale, long-term views of phytoplankton abundance and, for some sensors, other variables such as the concentration of colored dissolved organic material, an important carbon pool.

Compared with CZCS, the new sensors have more spectral bands, reduced radiometric noise, and for most of the sensors, improved calibration capabilities. The additional bands in the near infrared, where the ocean may be considered black, will help extrapolation of the atmospheric radiance to the blue and green channels. The directional and polarization measurements of the POLDER instrument and only directional of MISR will improve determination of the aerosol model. The new algorithms developed for these sensors include better handling of multiple scattering [*Fraser*, this issue; *Gordon*, this issue]. One expects therefore that a more accurate atmospheric correction will be possible with these new sensors.

Wang and Gordon [1994] exploit the new angular information in their proposed algorithm for MISR, using radiance at 865 nm to select the aerosol models. *Gordon* [this issue] indicates that this single-wavelength approach, is more attractive than the two-wavelength SeaWiFS or MODIS approach, since it does not require assuming a phytoplankton pigment concentration less than 0.5 mg m^{-3}. Compared with the two-wavelength approach, the single-wavelength approach appears to perform similarly, except in winter geometrical conditions, for which the two-band correction is significantly better.

For the POLDER instrument, another use of angular information in atmospheric correction is described. In the so-called "class 1" algorithm [*Deschamps*, workshop communication] the aerosol path radiance at 670, 765, and 865 nm is first computed from the satellite radiance at the respective wavelengths by subtracting molecule, glitter, and foam contributions, as well as coupling terms (foam-molecule, glitter-molecule). This requires a first guess of the aerosol optical thickness (to estimate atmospheric transmittance) and the phytoplankton pigment concentration (to estimate ocean reflectance at 670 nm). An average spectral dependence of the aerosol path radiance between 670 and 865 nm is then determined for the various viewing geometries (the same target can be viewed by the POLDER instrument in up to 14 directions) and compared to precalculated values, allowing a determination of the aerosol type. In the next step, knowing the aerosol type, an average τ_a at 865 nm is obtained for the various viewing geometries as well as a dispersion coefficient (ratio of standard deviation and average value). The dispersion coefficient is minimized by modifying the foam contribution to the satellite radiance, since foam effects are not well known [e.g., *Frouin et al.*, 1996]. Finally, from τ_a at 865 nm and the aerosol type, the aerosol path radiance at the ocean color wavelengths is easily computed. Iteration may be necessary if the retrieved τ_a at 865 nm and phytoplankton pigment concentration differ substantially from the first-guess values. *Deschamps et al.* [1996] have applied successfully the above "class 1" algorithm to POLDER images acquired in October 1996 over the Mediterranean Sea and other European waters, providing the first view of phytoplankton pigment patterns in these regions since 1986 (when the CZCS stopped acquiring data).

Issues and Concerns

Despite (1) progress made during recent years in algorithm development, most notably the inclusion of multiple-scattering effects, additional spectral channels for a better determination of the aerosol type, (2) the availability of new data on the optical properties of the ocean, surface, and aerosols, and (3) the improved characteristics of the new satellite sensors, a number of issues must be addressed before the atmospheric correction can be qualified as generally accurate over the range of atmospheric and oceanic conditions expected to be encountered. They include the effects of absorbing aerosols, the impact of aerosol vertical structure, the influence of thin clouds in the sensor's field of view, the effects of stratospheric aerosols, the validity of the aerosol models, the influence of whitecaps on the ocean surface, the sensor sensitivity to polarization, the influence of instrument stray light, the impact of angular distribution of water-leaving and glitter radiances, the separation of case 1 and case 2 waters, the effects of sensor calibration errors, and the validation of the retrieved water-leaving radiances. Most of these issues have been examined by *Gordon* [this issue] and are summarized here with recommendations based on workshop discussions.

Absorbing aerosols. In the presence of absorbing aerosols (e.g., dust or black carbon particles) the current atmospheric correction schemes do not perform adequately. The aerosol path radiance in the red and near infrared cannot be readily extrapolated to shorter wavelengths, even with the addition of specific aerosol models. *Gordon* [this issue] shows that to improve the atmospheric correction in the presence of absorbing aerosols, the candidate models must have single-scattering albedos similar to that of the actual aerosols. It is not easy, unfortunately, to detect the presence of absorbing aerosols, furthermore, optical properties of dust, which are nonspherical, are poorly known, and more observations (ground based, aircraft) are needed of the optical properties of absorbing aerosols.

Fukushima and Toratani [this issue] propose using the Ångström exponent between 550 and 670 nm to detect absorbing aerosols, but their assumption that the ocean reflectance is known at 550 nm may not hold in many cases. *Gordon* [this issue] suggests that MODIS observations at $\lambda > 1.2$ μm may be useful, since the spectral dependence of the aerosol path radiance between 865 nm and these wavelengths is much lower for dust than for nonabsorbing aerosols. One promising approach, however, might be to use ultraviolet data from TOMS. Ultraviolet wavelengths might also improve the atmospheric correction more directly, by constraining the aerosol path radiance extrapolated from red and near-infrared wavelengths. Such an approach has been suggested by *Lee et al.* [1995]. GLI will have a spectral band centered at 380 nm, which can be used for this purpose; but the water-leaving radiance may not be null at this wavelength, especially in the open ocean, complicating the retrieval of the aerosol path radiance. On the other hand, polarization and bidirectionality measurements, such as those of the POLDER, are very sensitive to aerosol type and probably offer the best way to discriminate efficiently the aerosol type.

Whitecaps. The presence of whitecaps (foam, streaks, underwater bubbles) at the ocean surface can affect considerably the satellite signal in the visible and near infrared, making it more difficult to estimate the aerosol path radiance. The optical properties of whitecaps and their fractional coverage, unfortunately, are not well known. *Gordon and Wang* [1994], assuming that whitecap reflectance is wavelength independent, found that the effects of whitecaps on the retrieval of water-leaving radiance in the blue is quite small (<0.001 in reflectance). However, recent in situ measurements [*Frouin et al.*, 1996; *Moore et al.*, 1996] suggest that whitecap reflectance decreases substantially in the near infrared, contrary to previous theoretical and laboratory studies. If confirmed, errors of an order of magnitude larger than the acceptable errors for biological applications are expected by neglecting the spectral dependence of the whitecap reflectance.

More in situ measurements of whitecaps are in order, to determine their optical properties (spectral, bidirectional, and polarization) and how these properties depend on environmental factors (wind speed, atmospheric stability, ocean stratification, and composition). Methods for retrieval of the effective whitecap reflectance (i.e., the product of fractional coverage and reflectance) should be developed for ***MODIS*** and GLI using additional wavelengths in the near infrared. Therefore large errors may occur in using satellite observations that are potentially affected by an effective whitecap reflectance ≥ 0.001, i.e., corresponding to wind speeds above 8–9 m s^{-1}.

Aerosol vertical structure. Current atmospheric correction algorithms assume that aerosols are located either in the boundary layer, below molecules, or distributed vertically according to some climatology. As long as the aerosols are not absorbing or weakly absorbing, the influence of vertical structure is negligible, but the effect cannot be neglected for aerosols located above 7–8 km [*Ding and Gordon*, 1995]. As the single-scattering albedo decreases, however, the error becomes progressively larger. In the presence of uniformly mixed urban type aerosols, for instance, the *Gordon and Wang* [1994] algorithm, which assumes a two-layered atmosphere, will overestimate the aerosol path radiance in the blue by more than 0.01 in reflectance units [*Gordon*, this issue], well beyond the acceptable error limit (0.001). Thus for absorbing aerosols, information on the vertical profile is important. The use of measurements sensitive to the atmospheric profiles, for example, lidars, TOMS, instruments with the 765 nm oxygen band (MERIS and GLI) should be explored.

Cirrus clouds. Thin cirrus clouds, transparent in most of the visible and near-infrared spectral range, may go undetected in ocean color imagery obtained from sensors such as SeaWiFS. If undetected, these clouds may be interpreted as tropospheric aerosols in atmospheric correction algorithms, possibly leading to unacceptable errors in aerosol path radiance. More study is required for a quantitative assessment of the impact of cirrus clouds on the atmospheric corrections. The use of thermal infrared channels, or the new cirrus 1.38 μm channel, may help their detection. OCTS, MODIS, and GLI have these channels (except of 1.38 μm on OCTS).

Stratospheric aerosols. Stratospheric aerosols, following volcanic eruptions, affect the atmospheric correction over the oceans, which is basically geared toward tropospheric aerosols. *Gao and Kaufman* [1996] and *Gordon* [this issue], however, suggest that radiance measured at 1.38 μm (MODIS, GLI), resulting mostly from scattering by stratospheric aerosols, can be used to estimate this contribution. Exogenous stratospheric aerosol data from occultation experiments (i.e., SAGE II and follow-ons), are probably the best alternative to correct the effect on ocean color radiances. Simulations made for the POLDER instrument [*Herman*, 1996] show that by using this approach, the expected errors on the water-leaving radiance in the blue are only a few 0.0001 in reflectance units, even at large solar and viewing zenith angles.

Validity of aerosol models. Current atmospheric correction algorithms [*Gordon and Wang*; 1994] require a set of candidate aerosol models usually taken from the work of *Shettle and Fenn* [1979]. These models, however, are essentially built from ground-based physicochemical analysis of aerosol samples and therefore must be validated in terms of phase function [*Remer et al.*, 1996], optical thickness, and single-scattering albedo, including spectral dependencies (*i.e.*, the variables that directly affect aerosol path radiances). Furthermore, the characteristics of dust are poorly known [*Sokolik and Toon*, 1996], and measurements (ground based, aircraft) are needed to develop realistic models. Errors in pigment concentrations, usually in the 15% error range, become unacceptable in the presence of dust.

Validation of the aerosol models can be accomplished by using instruments such as the AERONET Sun/sky radiometers [*Holben et al.*, 1997], which measures spectral solar transmission and sky radiance (including solar aureole) in the almucantar and principal plane. The AERONET measurements can be inverted to retrieve aerosol size distribution, phase function, and single-scattering albedo [e.g., *Nakajima et al.*, 1986; *Kaufman et al.*, 1994; *Wang and Gordon*, 1993], but they also provide, almost directly, estimates of the "pseudo" phase function (product of phase function and single-scattering albedo). These radiometers cannot be easily operated onboard ships, because they require accurate positioning, although a Japanese version of the Sun/sky radiometer has been successfully used at sea (T. Nakajima, personal communication, 1997). The AERONET international network has been put in place over land, by Holben and Tanré, and is being expanded to include oceanic stations (coastal regions, islands).

A first effort to validate the aerosol models selected by *Gordon and Wang* [1994] was made by *Schwindling* [1995], who collected Sun/sky radiometer data at the Scripps Institution of Oceanography (SIO) Pier, California, during the winters of 1993 and 1994. He found that for the low-aerosol optical thickness (0.1 at 870 nm), a unique aerosol model, representative of stratospheric and tropospheric backgrounds, might be sufficient to operate atmospheric correction algorithms. Estimates of the "pseudo" aerosol phase function and the Ångström exponent were found to fit the models of *Shettle and Fenn* [1979], which thus appeared to be representative of the SIO Pier site. *Schwindling* [1995] also derived relationships between "pseudo" phase function and Ångström exponent, suggesting that measurements of spectral optical thickness might be sufficient in many cases to identify the aerosol type.

Atmospheric transmittance and Sun glint effects. Atmospheric correction algorithms assume that photons specularly reflected by the surface are directly transmitted through the atmosphere. In a scattering atmosphere, some of the photons reflected by the ocean glint might be scattered into other directions. Thus residual glitter effects in the satellite imagery may not be accounted for properly, leading to atmospheric correction errors. Although the magnitude of the errors has yet to be quantified, it is straightforward to incorporate the glitter reflectance diagram into the algorithms [*Fraser et al.*, this issue].

Ocean BRDF. Backscattering of sunlight by the water body, on the other hand, is not isotropic, as shown by *Morel and Gentilli* [1993] and *Morel et al.* [1995]. The bidirectional characteristics of the water body reflectance affect the signal transmitted to the satellite but are ignored in current atmospheric correction algorithms. The effect on diffuse atmospheric transmittance is typically less than 5% [*Gordon*, this issue], but there is direct repercussions on water-leaving radiance. Accurate computation of the diffuse atmospheric transmittance requires a knowledge of the upwelling radiance distribution, which depends on water composition. Iterative schemes may be devised.

Gaseous absorption. In atmospheric correction algorithms, gaseous transmittance in the visible is generally assumed to be governed uniquely by ozone absorption (except for spectral bands affected by oxygen absorption). There is also water vapor absorption in the visible, not only due to absorption lines but also due to a continuum. The ocean color spectral bands generally avoid water vapor absorption lines, but they are affected by the continuum (1–2% transmittance effect in the visible). Even though water vapor is located low in the troposphere mostly under the molecular scattering, the effect may be significant for remote sensing of ocean color. The POLDER "class 1" algorithm takes into account absorption by the water vapor continuum, which requires an estimate of the water vapor amount (also provided by the POLDER instrument). More study is needed, however, to assess the impact of this type of absorption on water-leaving radiance retrievals.

Instrument polarization. Satellite ocean color sensors, especially scanners, are sensitive to the polarization of the radiance they are trying to measure. The sensitivity varies with the type of sensor. SeaWiFS, for example, has a very good polarization tolerance (a fraction of a percent), due to the presence of a polarization scrambler. MODIS, on the other hand, is trying to achieve a 2% polarization tolerance. If one considers the extreme case of a molecular atmosphere and solar and viewing angles giving a scattering angle of 90°, the incident radiance at the entrance of the sensor will be completely polarized. A 2% polarization tolerance for the sensor will translate into a maximum error of 2% on the radiance measurement, which typically corresponds to a 20% error on the water-leaving radiance in the blue. This error is unacceptable, 5% being the limit for biological applications. Thus for sensors like MODIS, polarization effects must be removed or well characterized if one wants to use the data quantitatively. A realistic atmosphere, however, contains not only molecules but also aerosols, and aerosols are less effective at polarizing incident radiance. Sensor polarization effects therefore will be smaller in actuality, but the polarization characteristics of the measured radiance are unknown in the presence of aerosols, complicating the removal procedure. *Gordon* [this issue] suggests that sensor polarization effects can be estimated with sufficient accuracy for ocean color remote sensing by assuming a pure molecular atmosphere.

Stray light/adjacency. Detectors of large field-of-view sensors are affected by stray light of diverse origins (reflection by diopters, lens edges, elements of the filter assemblies, etc.). In the vicinity of high-reflectance objects (e.g., clouds, ice sheets) the level of contamination may be unacceptable for ocean color remote sensing (the signal from the ocean is quite small). For MODIS, as much as 100 pixels adjacent to a bright target (about 100 km) may exhibit significant stray light. For POLDER any bright pixel may affect at least 40 surrounding pixels (about 250 km). If the artifacts, when mapped properly, cannot be corrected before launch, they must be removed after launch. This is generally possible (done systematically for the POLDER instrument) but extremely time consuming, since the effects depend on the location, shape, and intensity of the contaminating sources, and the task might overburden the ground segments of satellite missions.

Another type of adjacency effect, due to photons reflected by the target's environment and scattered by the atmosphere into the sensor's field of view, may add to the instrument stray light effect. According to *Deschamps et al.* [1983], in the presence of a clear atmosphere the additional effect may be felt over a distance of about 10 km. Although no studies of the consequences for ocean color remote sensing have been made, the effect can be corrected by knowing the reflectance of the environment and the characteristics of the atmosphere (aerosol properties). Still, the two types of adjacency effects, instrumental and atmospheric, may not be easily separated. This emphasizes the complexity of the atmospheric correction problem in coastal regions, where, furthermore, the water body may not be considered black in the red (the basis of current algorithms).

6. Discussion on Validation

Validation of global satellite data sets requires a two-tiered approach consisting of both (1) networks of long-term surface monitoring sites and (2) short-term, yet comprehensive, field experiments to understand the processes and assumptions used in the satellite retrievals. In the case of atmospheric aerosol applications focused on aerosol radiative forcing, the most important field experiment is a column closure experiment, in which satellite overpasses are coordinated with surface measurements of column aerosol and water vapor properties, together with a comprehensive set of in situ aircraft observations of aerosol microphysics (size distribution, single-scattering albedo), radiation (both spectral and broadband), and chemistry. These observations should characterize the vertical column between the surface and the upper troposphere, so the sampling strategy consists of both horizontal flight legs as well as vertical profiles.

Long-Term Network of Surface Sites

In order to validate satellite-derived optical thickness estimates worldwide, it is essential to have a well-calibrated, and characterized, surface network of Sun and sky radiometers. For aerosol validation purposes it is extremely important that this network of quality-controlled radiometers be located in major, and distinctive, aerosol regimes of the world. In particular, we recommend that such a surface network be located in the following major aerosol regimes: (1) urban/industrial aerosol (e.g., East Coast of the United States, Japan, and/or Europe and numerous city environments worldwide), (2) biomass burning (e.g., Brazil, central Africa, and Indonesia), (3) Asian absorbing aerosol (e.g., China or Hong Kong), (4) Arctic haze (e.g., Greenland, Spitsbergen, Barrow), (5) marine aerosol (e.g., Hawaii, Bermuda, Canary Islands), (6) mineral dust (e.g., Sahelian dust from the continent of Africa), and (7) yellow sand from China.

To maximize economic value, it is further recommended that this network be collocated with other surface networks wherever possible (e.g., ARM, BSRN, GAW) and that there be well-established quality assurance/quality control protocols and procedures established. This necessitates locations where there is a responsible site manager with good network communications and that there is adequate logistics support for visiting or shipping of instrument components.

The workshop also endorsed a two-tiered approach to long-term surface networks: (1) a basic site and (2) a super site. For the basic site the minimum set of measurements and conditions should include a Sun/sky radiometer (such as those in AERONET [*Holben et al.*, 1997]) for aerosol optical thickness and columnar size distribution and "uniform" surface conditions. The super site, which would be a subset of the larger network, should have one representative in each aerosol regime identified above, should include the basic site system and surface conditions and, in addition, include some or all of the following enhancements: (1) surface radiation budget measurements (both shortwave and longwave), (2) in situ measurements of single-scattering albedo, (3) skylight polarization measurements, (4) monostatic or scanning lidar (e.g., micropulse lidar such as that described by *Spinhirne* [1993]), (5) chemical composition of aerosol, (6) collocation with radiosonde, and (7) extended wavelength range of Sun and sky radiometer (should include measurements to 1.6 and 2.2 μm). The basic sites could likewise incorporate some of these additional measurement capabilities (e.g., polarization, which currently exists in many of the Sun/sky radiometers in AERONET, at least at 0.86 μm).

Sun and Sky Radiometer Standards

The workshop participants discussed the need to identify and verify that the surface suite of Sun/sky radiometers (e.g., AERONET) meet certain minimum standards. It was also recognized that a different set of criteria could be justifiably placed on the basic sites which would be less stringent than for the super sites.

For the basic sites (perhaps 60–80 sites) it is necessary for the uncertainty in the aerosol optical thickness $\Delta\tau_a < 0.02$ at any single wavelength (with a multiwavelength precision of 0.01). For the sky radiance measurements to be of value, they must be measured with an accuracy of 5%, requiring careful attention to periodic calibration at a high-altitude mountain location and intercomparisons with "standard" instruments. The Sun/sky radiometers at the super sites should perform even better, with $\Delta\tau_a < 0.01$ at any single wavelength (with a multiwavelength precision of 0.005) and sky radiance measurements made with a goal of 2.5% absolute accuracy.

Atmospheric Corrections

For this purpose as well as for characterizing the aerosol optical thickness retrievals, it is necessary to have a complete set of comprehensive validation measurements acquired at a few super sites if at all possible. These efforts can be supplemented by participation in intensive, focused field campaigns, such as column closure experiments or other national or international intensive field campaigns. Core validation of specific aspects of atmospheric correction should take place at numerous sites as part of core EOS validation activities. Finally, global statistics can be used to monitor the health of the atmospheric correction algorithms over time. The possibility of deploying basic instrument packages on numerous truck-mounted, airborne, or ship platforms was also discussed.

Another indispensable validation measurement for atmospheric correction is the upwelling radiance or surface reflectance for a variety of target types in different ocean regions as well as land cover and climatological zones. An important issue in this respect is that of scaling surface-based measurements at specific locations up to the size of several satellite pixels. Since this scaling usually involves either homogeneous sites and/or multialtitude aircraft measurements, validation of the surface signal is necessarily constrained and focused. The validation of higher-level products based on surface reflectance and albedo is also important but beyond the scope of this discussion.

For atmospheric correction over the ocean, the ocean science community has developed marine optical buoys for measuring downwelling irradiance and upwelling radiance, both at the ocean surface and in the subsurface waters. At the present time, there are two such marine optical buoys [*Clark et al.*, 1996; *Kishino et al.*, 1996]. One is located in case 1 waters off the coast of Lanai, Hawaii, while the second is located in case 1 and, occasionally, case 2 waters off the Sea of Japan. The global biological ocean community will rely on these two sites for ground truth and validation of water-leaving radiances derived from various ocean color sensors (e.g., SeaWiFS, MODIS, OCTS, MERIS, and GLI). As a consequence of the key role that these two validation sites play in ocean color validation, it is exceedingly important that the atmosphere also be well characterized at these locations by using Sun and sky radiometers such as those in the AERONET.

Issues and Concerns

Because of the necessity of making aerosol measurements during clear-sky (or low cloudiness) conditions, there is a distinct possibility that the satellite and surface network-derived aerosol optical thickness will be biased to clear-sky conditions (low optical thickness or low humidity conditions). This is unavoidable but important to recognize. We were also concerned as to whether we may have overlooked some valuable auxiliary measurements, not identified above. For all surface networks it is essential that the site reflectance be characterized, as the satellite radiometers will be retrieving aerosol optical thickness looking down at the site.

7. Conclusions

The general consensus of the remote sensing community represented in the workshop was that we can measure remotely the aerosol optical thickness to within 0.03 to 0.05. Some information on aerosol size distribution and single-scattering albedo may also be derived but to within uncertain accuracies. No estimates were made as to the accuracies of other retrieved parameters although the consensus was that they had significant errors.

The main sources of uncertainty in the optical thickness determinations were considered to be surface reflectivity, instrument calibration and instability for low optical thickness, and aerosol model assumptions for high optical thickness. Main sources of uncertainty for other parameters were not determined but appear to be related to uncertainties in the assumed quantities and, to some extent, to the numerical procedures used. It was generally agreed that aerosol optical thickness is the easiest parameter to determine and that all other properties are more difficult and also less accurate. Significant improvement over present capabilities is expected with the new instrumentation, which will include polarization measurements as well as angular and wavelength scanners. A new and promising approach was presented which involved near-ultraviolet measurements, utilizing the aerosol absorption. This technique promises to provide a long sequence of aerosol data over land and a measure of aerosol absorption. Quantitative application depends on knowledge of aerosol vertical distribution and optical properties in the UV.

Regarding atmospheric corrections, new algorithms different in philosophy are now being developed. Over land, the availability of aerosol information holds considerable promise for routine corrections; more sophisticated effects such as the atmospheric point spread function and the coupling of surface and atmospheric BRDFs can be corrected for. Over ocean, instead of considering the atmosphere and the surface as perturbing water-leaving radiance (the signal of interest), new algorithms view the atmosphere, surface, and water body as forming a coupled system, and they attempt to retrieve simultaneously the geophysical variables that affect the satellite radiances in a single step. The chief advantage of this approach is that the information from all the sensor wavelengths is used to perform the atmospheric correction, not only the information in the red and near infrared, where the ocean reflectance is negligibly small.

Some unanswered questions that were raised are as follows:

1. How can we measure single-scattering albedo over oceans? What could be the accuracy when the best we can do in the laboratory is ±30%?
2. What is the best second parameter to try to determine with a two-channel or multichannel instrument? Some suggestions were the effective radius, the Ångström wavelength exponent, and the ratio between the accumulation and the coarse particle modes.
3. Can ultraviolet absorption provide quantitative aerosol information?
4. How may aerosol climatology be used to help derive aerosol properties from satellite observations?
5. How can we correct for whitecap effect over the ocean?

Finally, the workshop made the following recommendations:

1. Implementation of ground-based and airborne measurements to augment satellite observations is essential.
2. Intercomparison of the different observational techniques is vital.
3. Initiation of sensitivity studies to determine information content in remote sensing and the effects of uncertainties in all the various parameters on the derived properties is important.
4. Ranking of parameters by their impact on measurements should be investigated.
5. Further study of the proposed UV method for determining aerosol presence and absorption is necessary.
6. Extending aerosol climatology should be done as soon as is practicable.
7. Development of techniques to measure aerosol absorption and single-scattering albedo and to derive realistic models for absorbing aerosols should be encouraged.
8. Studies of the usefulness of other techniques such as occultation, limb scatter, lidar, and two-band measurements for retrieving the vertical profile are urgently needed.
9. Further study of whitecap optical properties and how they vary with environmental factors is needed.

Notation

AERONET	Aerosol Robotic Network.
AVHRR	advanced very high resolution radiometer.
CERES	clouds and the Earth's radiant energy system.
CZCS	coastal zone color scanner.
ERBE	Earth Radiative Budget Experiment.
GAW	Global Atmosphere Watch.
GEWEX	Global Energy and Water Cycle Experiment.
GLI	global imager.
GOES	Geostationary Operational Environmental Satellite.
IGAC	International Global Atmospheric Chemistry.
MODIS	moderate resolution imaging spectroradiometer.
MERIS	medium resolution imaging spectrometer.
MISR	Multiangle imaging spectroradiometer.
MSS	multispectral scanner.
OCTS	ocean color and temperature scanner.
POLDER	Polarization and Directionality of the Earth's Reflectance.
SAGE	Stratospheric Aerosol and Gas Experiment.
SeaWiFS	sea-viewing wide field-of-view sensor.
TM	Thematic Mapper.

TOMS total ozone mapping spectrometer.
WCRP World Climate Research Program.

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