Verification of aerosol models for satellite ocean color remote sensing

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Abstract. Direct atmospheric transmittance and sky radiance were measured spectrally at the coastal site of Scripps Institution of Oceanography pier in La Jolla, California, during the winters of 1993 and 1994. Direct atmospheric transmittance was also measured aboard R/V Wecoma and on Catalina Island during the 1994 California Cooperative Oceanic Fisheries Investigation winter cruise. The data were analyzed to (1) verify whether the aerosol models selected by Gordon and Wang [1994] for the sea-viewing wide field of view sensor are adequate for ocean color remote sensing from space and (2) identify what type of in situ atmospheric optics measurements should be performed to verify atmospheric correction algorithms. Aerosol optical thickness at 870 nm was generally low at La Jolla, with most values below 0.1 after correction for stratospheric aerosols. Values were even lower offshore (R/V Wecoma, Catalina Island), and no systematic correlation was found between aerosol characteristics and meteorological conditions. Therefore a mean background model, specified regionally, may be sufficient for ocean color remote sensing from space. For optical thicknesses above 0.1, two modes of variability characterized by Ångström coefficients of 1.2 and 0.5 and corresponding to tropospheric and maritime models, respectively, were identified in the measurements. The aerosol models selected by Gordon and Wang [1994] allow one to fit, within measurement inaccuracies, the derived values of Ångström coefficient and "pseudo" phase function (the product of single-scattering albedo and phase function), key atmospheric correction parameters. Importantly, the pseudo phase function can be derived from measurements of the Ångström coefficient. This means that shipborne Sun photometer measurements made at the time of satellite overpass are usually sufficient to verify the atmospheric correction of ocean color.

1. Introduction

Remote sensing of ocean color from space provides the only means to monitor phytoplankton pigment concentration and primary production on a global scale. Large-scale, long-term information on these parameters is necessary to improve our understanding of biogeochemical (carbon, nutrient) cycles [e.g., Joint Global Ocean Flux Study (JGOFS), 1987; Abbott et al., 1994]. However, deriving marine spectral reflectances from top of atmosphere (TOA) radiances in the visible and near-infrared is not easy because of the interference of atmospheric scattering and absorption. Typically, 90% of the TOA signal is composed of photons that have not interacted with the water body [e.g., Viollier et al., 1980] and, thus, do not contain information on water composition.

The difficulty of the atmospheric correction problem has been identified early with the coastal zone color scanner (CZCS), the first space experiment dedicated to ocean color [Hovis et al., 1980; Gordon et al., 1980]. CZCS provided data from 1978 through 1986. The next generation of instruments, such as the sea-viewing wide field of view sensor (SeaWiFS) [Hooker et al., 1993; Hooker and Esaias, 1993], will improve the accuracy of the marine reflectance estimates. SeaWiFS has upgraded radiometric performance and additional wavelengths in the near-infrared for more efficient atmospheric correction [Gordon and Wang, 1994]. The launch of the ocean color and temperature scanner (OCTS) and the polarization and directionality of Earth reflectances (POLDER) radiometer [Deschamps et al., 1994] on ADEOS in 1996 and of SeaWiFS on SeaStar [Hooker et al., 1993] in 1997 will be followed by the launch of the moderate resolution imaging spectrometer (MODIS) [Ardanuy et al., 1991] on the Earth Observing System (EOS) AM-1 platform in 1998, the medium resolution imaging spectrometer (MERIS) [Rast and Bézy, 1995] on ENVISAT in 1999, etc.

Atmospheric correction algorithms attempt to retrieve the actual value of the marine reflectance from TOA measurements. Some atmospheric effects are straightforward to correct, either because they can be computed using auxiliary data or because they are weak. Molecular scattering is important at shorter wavelengths, but can be computed almost exactly knowing the surface pressure and the value of the solar constant [Gordon et al., 1988; André et Morel, 1989]. Ozone transmission can be computed accurately using an oxygen total ozone content. Water vapor transmission is usually negligible.
in ocean color spectral bands. The concentration and composition of atmospheric aerosols, on the other hand, are highly variable in space and time, and the correction of their effects is a specific problem.

Because of the space and time variability of aerosols, a priori or climatological knowledge of their optical properties is not sufficient to compute accurately their effects on the TOA signal. Aerosol parameters must be known at the time of the ocean color observations, and ideally should be derived from measurements by the same instrument. It was noted early that aerosol scattering behaved in first approximation like the product of optical thickness, single-scattering albedo, and scattering phase function [Tanré et al., 1979], resulting in a smooth spectral dependence of the scattered radiance in the visible and near-infrared for the most common aerosol models. The general concept of the aerosol correction algorithm, therefore, is to observe the aerosol-scattered radiance in the near-infrared where the marine reflectance is negligible, and to extrapolate the signal to the wavelengths in the visible [Viollier et al., 1980; Gordon and Clark, 1980].

Application of this type of algorithm to CZCS data has been limited because the only band usable for atmospheric corrections, close to the near-infrared, was centered at 670 nm in the red. One had to (1) assume a given aerosol type and spectral dependence of the aerosol scattered radiance and (2) restrict the processing to ocean cases where the marine reflectance at 670 nm could be neglected [Gordon et al., 1983]. At least two measurements in the near-infrared at wavelengths above 700 nm are necessary to determine correctly the spectral dependence of the aerosol radiance and the aerosol type. From these measurements the correction for aerosol scattering can be extrapolated to shorter wavelengths, either analytically [Deschamps et al., 1983] or with the help of aerosol models, as proposed by Gordon and Wang [1994].

The choice of aerosol models to guide the spectral extrapolation of aerosol radiance raises two main questions, both of which require in situ measurements: (1) Are the models representative of aerosol conditions encountered ocean-wide? (2) What type of in situ measurements should be made to verify the atmospheric correction algorithm, that is, to evaluate its performance?

In situ atmospheric optics measurements are not easy to perform at sea, from a ship or any platform. A few Sun photometer measurements have been reported [Tomasi and Prodi, 1982; Hoppel et al., 1990; Korotaev et al., 1993; Villevalde, 1994], and they have provided information on aerosol optical thickness and its spectral variations in the various oceans. More extensive measurements have been made from land surface sites to characterize tropospheric and stratospheric aerosol components [Bruegge, 1992; Herber et al., 1993; O'Neill et al., 1993; Smirnov et al., 1994; Dutton et al., 1994; Kent et al., 1994], including coastal sites [Obleitner, 1992]. Only a few studies address the problem of observing sky radiance, and thus the path radiance in the backward direction, at a scattering angle corresponding to space observations [Kaufman, 1993; Kaufman et al., 1994].

To provide answers to the above questions (representation of the models, type of in situ measurements), one needs to acquire atmospheric optics data under sufficiently varied aerosol conditions. The data should be collected prior to the launch of the ocean color instruments, so that recommendations and conclusions can be acted upon during the operational phase of the missions. It was our main motivation to establish, before the launch of SeaWiFS, the next ocean color sensor, a permanent atmospheric optics station on a coastal site in southern California. The measurements were performed systematically during the winters of 1993 and 1994. The measured or directly derived physical variables included spectral aerosol optical thickness and sky radiance at wavelengths used in ocean color remote sensing, that is, in the range 0.4–1 μm.

2. Background

2.1. Atmospheric Correction Algorithms

The ocean color sensor measures the solar radiation reflected by the ocean-atmosphere system, namely the TOA radiance \( L \), in spectral bands centered at wavelength \( \lambda \). For convenience, we will work with the spectral reflectance, \( \rho \), that is, the spectral radiance \( L / \rho \) normalized by the solar spectral irradiance outside the atmosphere \( E_\alpha \), and corrected for Earth-Sun distance variations:

\[ \rho = \pi L / \mu E_\alpha \]  

where \( \mu_r \) is the cosine of the solar zenith angle \( \theta_s \). In the remainder of the paper, we will omit the wavelength index \( i \) whenever possible. The TOA reflectance can be modeled as the sum of several contributions:

\[ \rho = \rho_m + \rho_a + \rho_f + \rho_p + \text{coupling terms} \]  

where \( \rho_m \) is the contribution of scattering by the water body (the signal to be retrieved), \( \rho_a \) is the contribution of foam at the surface, \( \rho_p \) is the contribution of Sun glint, and \( \rho_f \) and \( \rho_a \) are the contributions of scattering by air molecules and aerosols, respectively. Coupling terms result from multiple interactions between two or more of these individual contributions; they can be either neglected or computed knowing the aerosol model. We will assume that the measurements are made outside the directions contaminated by Sun glint and, therefore, we will neglect \( \rho_p \). We will also neglect \( \rho_f \), though \( \rho_f \) cannot be neglected at wind speeds above 10 m s\(^{-1}\) [Estep and Arnone, 1994; Frouin et al., 1996]. The atmospheric correction algorithm works on \( \rho_m \) and \( \rho_a \). The molecular scattering reflectance \( \rho_m \) is easy to compute accurately, knowing the molecular thickness of the atmosphere, a function of the atmospheric pressure at sea level. The aerosol scattering reflectance \( \rho_a \) on the other hand, depends on the aerosol amount and type, parameters that are eminently variable in time and space, and \( \rho_a \) has to be estimated from the measurements themselves.

The first generation of atmospheric correction algorithms, applied to CZCS [Gordon et al., 1980], assumed that the aerosol reflectance \( \rho_a(\lambda) \) varied with wavelength according to the following power law:

\[ \rho_a(\lambda) \sim \lambda^{-\alpha} \]  

This assumption was supported by the fact that using the single-scattering approximation, \( \rho_a \) can be expressed as

\[ \rho_a(\lambda) \sim \omega_a(\lambda) \tau_a(\lambda) \rho_s(\theta_n, \varphi, \lambda) / 4 \mu_n \mu_a \]  

where the subscript \( a \) refers to scattering by aerosols, \( \omega_a \) is the single scattering albedo, \( \tau_a \) is the scattering phase function, \( \mu_a \) is the cosine of the viewing zenith angles \( \theta_v \), and \( \varphi \) is the relative azimuth angle between solar and viewing directions. The scattering phase function and the single-scattering albedo vary weakly with wavelength. The wavelength dependence of the aerosol reflectance \( \rho_a(\lambda) \),
where the exponent \( \alpha \) is the Ångström coefficient [Ångström, 1961, 1964], close to the exponent \( n \) in (3).

The CZCS had six spectral bands, five in the visible and near-infrared (centered at 443, 520, 550, 670, and 750 nm) and one in the thermal infrared (10.5–12.5 \( \mu \)m). Only the first four channels had a good radiometric sensitivity and could be used for observation of ocean color. The first CZCS algorithm used the 670 nm channel to estimate \( \rho_a(670) \) assuming a “black” ocean, \( \rho_a(670) \approx 0 \). The value of \( \rho_a(670) \) was then extrapolated to the other wavelengths, with \( n \) corresponding to a mean aerosol model. In order to adapt the algorithm to different aerosol types, the clear water radiance concept was developed and applied to CZCS data in the open ocean [Gordon and Clark, 1981]. At low chlorophyll concentrations, \( \rho_a(520) \) and \( \rho_a(550) \) are known with enough accuracy so that not only \( \rho_a(670) \), but also \( \rho_a(520) \) and \( \rho_a(550) \), can be derived, leading to a self-determination of the wavelength dependence, hence the exponent \( n \). Thus in the clear ocean case the CZCS algorithm could accommodate for variability of both aerosol loading and type, but the solution was limited and not optimum.

A major improvement of the atmospheric correction resides in the capability of the next generation of instruments like SeaWiFS to observe in the near infrared and determine more independently the aerosol reflectance at 765 and 865 nm, with no reference to the clear water concept. On the larger SeaWiFS spectral range, 412–865 nm, however, the approximation of (3) is now not accurate enough [Gordon and Wang, 1994], primarily because its generality is affected by multiple-scattering and coupling terms. A new generation of atmospheric correction algorithms, based on aerosol models, has therefore been proposed by Gordon and Wang [1994] (see also Gordon [1997]).

Schematically, Gordon and Wang [1994] proposed first to compute exactly the radiative transfer in the ocean-atmosphere system for a set of reference aerosol models. Then, the wavelength dependence of the aerosol scattering is derived using the measurements in the near-infrared at 765 and 865 nm, and the (two) closest aerosol models are determined. Eventually, the aerosol correction at ocean color wavelengths is estimated by interpolation between the exact solutions for the two aerosol models and applied to the measurements. Multiple scattering is fully accounted for in this algorithm, but the algorithm relies strongly on the choice of the aerosol models. Gordon and Wang [1994] used nine reference aerosol models, namely, the Shettle and Fenn [1979] maritime and tropospheric models with a humidity variation of the aerosol optical properties and a coastal aerosol model, actually a mixing of the maritime and tropospheric models. The humidity is set at 70%, 90%, and 98%. These models may or may not be realistic. Shettle and Fenn [1979] developed their models using samples of aerosols for which they derived the optical characteristics. In atmospheric corrections we are more interested in the optical behavior of the aerosols through the entire atmosphere.

### 2.2. Aerosol Models

Various sets of aerosol models have been proposed to compute the radiative transfer in the Earth-atmosphere system, under varied geophysical conditions, for climate as well as remote sensing studies. They include the “haze-type” models of Diermeidjan [1969], the models of Shettle and Fenn [1979], and the models of the World Meteorological Organization (WMO) [1986] (see description below). Particular applications have required other developments, such as the global average model developed by Toon and Pollack [1976], to be used as input in global average radiative transfer calculations. The basic components and size distribution of this model vary with altitude and are supposed to represent an average of the most common aerosol conditions over the world. D'Almeida et al. [1991] established a global climatology of aerosols. They used the models of Shettle and Fenn [1979] and of WMO [1986] after some revisions and additions, namely, a polar aerosol model, a desert aerosol model, and a “mixed aerosols” model. Nilsson [1979] developed a model with variable size distribution and refractive index depending on relative humidity.

The models of Shettle and Fenn [1979], hereafter referred to as the SF models, and of WMO [1986] are based on the work of Shettle and Fenn [1976, 1979], who used ground-based as well as aircraft measurements of concentration, size distribution, and optical properties of tropospheric aerosols. Both types of models are mixtures of basic components referred to as “water soluble” and “dust-like,” which are representative of midlatitude soils, a “soot-like” component, which results from combustion and industrial sources, and an “oceanic” component, which is composed of seawater injected in the atmosphere by the wind. The main differences between the SF and WMO models are in the relative amount and the way the components are mixed. In the case of the WMO models, Mie theory is applied to each component to compute the optical parameters. The size distribution is a lognormal law. These optical parameters are mixed to obtain the final aerosol models. In the case of the SF models, the size distribution is a lognormal law with one or two modes applied to the mixtures of the basic components described above. The size distribution and the refractive index of the basic components both are altered by relative humidity. The mode radius of the size distribution, not its width, is changed with relative humidity according to the work of Hänel [1976]. As a result, Shettle and Fenn [1979] give an equivalent refractive index for specific relative humidities of 50%, 70%, 80%, 90%, 95%, 98%, and 99%, as well as tables of scattering and absorption coefficients obtained from Mie scattering computations in the spectral range 0.2–40 \( \mu \)m.

As indicated in the previous section, the models selected by Gordon and Wang [1994] for the atmospheric correction of SeaWiFS data are the SF tropospheric and maritime models for relative humidities of 70%, 90%, and 98%. The maritime models are actually the tropospheric models with the addition of sea-salt aerosols. Gordon and Wang [1994] add three additional coastal models that contain half the fraction of sea-salt aerosols of the maritime models.

The phase functions at 870 nm for specific scattering angles of 60°, 90°, and 120° are given in Table 1, along with the single-scattering albedo and the Ångström coefficient characterizing the wavelength dependence of the aerosol optical thickness between 400 and 1000 nm, for the WMO models and the nine SF models selected by Gordon and Wang [1994] (tropospheric, maritime, and coastal with humidities of 70%, 90%, and 98%, referred to as T70, T90, T98, M70, M90, M98, C70, C90, and C98, respectively). For the WMO models, the Ångström coefficients vary from 0.22 (maritime model) to 1.32.
Table 1. Optical Properties of the Aerosol Models of Shettle and Fenn [1979] and WMO [1986]

<table>
<thead>
<tr>
<th>Model</th>
<th>Ångström Coefficient 870 nm</th>
<th>Ångström Coefficient 870 nm</th>
<th>Ångström Coefficient 870 nm</th>
<th>( \sigma_{\text{aw}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>T70</td>
<td>1.46</td>
<td>0.991</td>
<td>0.328</td>
<td>0.197</td>
</tr>
<tr>
<td>T90</td>
<td>1.35</td>
<td>0.811</td>
<td>0.239</td>
<td>0.141</td>
</tr>
<tr>
<td>T98</td>
<td>1.22</td>
<td>0.718</td>
<td>0.201</td>
<td>0.118</td>
</tr>
<tr>
<td>C70</td>
<td>0.63</td>
<td>0.676</td>
<td>0.208</td>
<td>0.125</td>
</tr>
<tr>
<td>C90</td>
<td>0.40</td>
<td>0.505</td>
<td>0.134</td>
<td>0.091</td>
</tr>
<tr>
<td>C98</td>
<td>0.27</td>
<td>0.438</td>
<td>0.105</td>
<td>0.077</td>
</tr>
<tr>
<td>M70</td>
<td>0.40</td>
<td>0.611</td>
<td>0.184</td>
<td>0.111</td>
</tr>
<tr>
<td>M90</td>
<td>0.21</td>
<td>0.458</td>
<td>0.118</td>
<td>0.084</td>
</tr>
<tr>
<td>M98</td>
<td>0.12</td>
<td>0.405</td>
<td>0.093</td>
<td>0.072</td>
</tr>
<tr>
<td>WMO maritime</td>
<td>0.22</td>
<td>0.527</td>
<td>0.150</td>
<td>0.096</td>
</tr>
<tr>
<td>WMO continental</td>
<td>1.18</td>
<td>0.906</td>
<td>0.310</td>
<td>0.191</td>
</tr>
<tr>
<td>WMO urban</td>
<td>1.32</td>
<td>0.979</td>
<td>0.356</td>
<td>0.234</td>
</tr>
</tbody>
</table>

For the maritime, coastal, and tropospheric SF models, the variations are similar, from 0.10 to 1.46. Figure 1 gives the scattering phase functions at 870 nm of the continental, urban, and maritime WMO models, and four of the SF models selected by Gordon and Wang [1994], namely, tropospheric and maritime with humidity of 70% and 98%, which represent extreme cases. Examining the SF models, we see a general tendency, as we go from the small and dry particles of the T70 model to the large water particles of the M98 model, of decreasing values of the Ångström coefficient and the scattering phase function around 90°. Minima of the scattering phase functions are observed at angles 90°–120° and are well marked below 0.1 for the aerosols of maritime type. The WMO maritime model is very similar to the SF M90 model. The WMO continental model is close to the SF T70 model, but with comparatively smaller Ångström coefficient (\( \alpha = 1.18 \)) and larger absorption (\( \sigma_{\text{aw}} = 0.842 \)). The WMO urban model is an extreme model characterized by small particles, large scattering phase function, and large absorption (\( \sigma_{\text{aw}} = 0.591 \)). If we exclude this last model, the range of models selected by Gordon and Wang [1994] fits the range of the WMO models.

The physical quantity observed from space is the radiance at a scattering angle between 45° and 180°. To interpolate the aerosol signal from one wavelength \( \lambda_i \) to another wavelength \( \lambda_j \), the useful parameter is the ratio of the aerosol reflectances \( e(\lambda_i, \lambda_j) \) at these two wavelengths:

\[
e(\lambda_i, \lambda_j) = \frac{p_a(\lambda_i)}{p_a(\lambda_j)}
\]

To compute the aerosol reflectances, we can use as a first approximation the single-scattering approximation given by (4). This gives for \( e(\lambda_i, \lambda_j) \):

\[
e(\lambda_i, \lambda_j) = \frac{\sigma_{\text{aw}}(\lambda_i)}{\sigma_{\text{aw}}(\lambda_j)} p_a(\theta_i, \theta_o, \varphi_i, \lambda_i) / p_a(\theta_i, \theta_o, \varphi_i, \lambda_j)
\]

Note that Gordon and Wang [1994] included a surface reflection component in their definition of \( e \). Using (7) is appropriate for the purpose of the study, since we are analyzing ground-based measurements (no surface effects). Figure 2 displays \( e(870, 1020) \) as a function of the scattering angle for five SF models (T70, T98, M70, M98, and C70) and three WMO models (continental, maritime, and urban) at 870 nm. We chose the wavelengths of 870 and 1020 nm because our measurements of aerosol reflectance are accurate at these wavelengths (see section 3.2). The mean value of \( e(870, 1020) \) roughly corresponds to the ratio of the optical thickness at 870 and 1020 nm, not shown in Figure 2. The curves show little variation with geometry, except in the backward direction for the maritime models.

2.3. Previous Atmospheric Optics Measurements

2.3.1. Sun intensity. A radiometer with a small field of view for viewing the Sun measures the intensity of direct solar radiation and thus the direct atmospheric transmittance. The aerosol optical thickness is derived after correcting the measurements for molecular scattering and absorption. Historically, the measurements have been made in broad spectral bands. More accurate measurements of the aerosol optical thickness are now performed in narrow spectral bands corresponding to atmospheric windows. The spectral measurements give the aerosol optical thickness and its spectral dependence, allowing one to derive the classical turbidity factors, namely, the optical thickness normalized at the wavelength of 1 \( \mu \text{m} \), \( \beta \), which is directly related to the aerosol loading, and the Ång-
from solar aureole observations, that is, measurements of the measurements have been widely performed to characterize the infrared and shortwave infrared, allowing more complex re-

mades initially in the visible and later extended to the near-

infrared and shortwave infrared, allowing more complex re-

trieval of the aerosol type. Ground-based Sun photometer measurements have been widely performed to characterize the tropospheric and stratospheric aerosol components (see refer-

ences and thus the aerosol type. The measurements were

sensitive to large particles (>1 μm), while the 3nm coefficient is more

sensitive to small particles (<1 μm). The inversion of the two

types of measurements, aureole and optical thickness, gives at least two modes of the aerosol size distribution [Nakajima et

al., 1983, 1986; Tanr et al., 1988]. It is difficult, however, to

retrieve the refractive index from aureole measurements. So far, there has been no published report of solar aureole measurements at sea, although they have been attempted (T. Nakajima, personal communication, 1997). The main limitation is due to the severe requirements on platform stability, which must be maintained to better than 0.1" in three directions.

2.3.2. Solar aureole. More information can be obtained from solar aureole observations, that is, measurements of the sky radiance at forward scattering angles, typically less than 30°. Schematically, the aureole technique is sensitive to large particles (>1 μm), while the Ångström coefficient is more sensitive to small particles (<1 μm). The inversion of the two types of measurements, aureole and optical thickness, gives at least two modes of the aerosol size distribution [Nakajima et

al., 1983, 1986; Tanr et al., 1988]. It is difficult, however, to

retrieve the refractive index from aureole measurements. So far, there has been no published report of solar aureole measurements at sea, although they have been attempted (T. Nakajima, personal communication, 1997). The main limitation is due to the severe requirements on platform stability, which must be maintained to better than 0.1" in three directions.

2.3.3. Sky radiance. Because of its remote sensing signif-

icance, the scattering phase function in the backward direction at angles between 90° to 180° has been measured directly. This approach is better, at least in principle, than computing the phase function from the derived aerosol model, but accurate measurements from the ground are difficult to make at large scattering angles. Low elevations for the Sun and the direction of observation are required, which restricts the measurements possible during the day. More importantly, multiple scattering increases rapidly at low elevation, limiting the accuracy of the scattering phase function retrieval. As a result, it is generally inaccurate to observe at angles above 150°. Also, the atmosphere may not be homogeneous between solar and viewing directions. Systematic observations at various time and locations (over land) have only been reported at the angle of 120° [Kaufman, 1993; Kaufman et al., 1994], an angle close to the minimum of the scattering phase function where it is very sensitive to the aerosol model (see models above). Like solar aureole, sky radiance is difficult to measure at sea, and no measurements over the ocean have been reported in the literature.

3. Material and Methods

Extensive time series of atmospheric optics data were acquired during the winters of 1993 and 1994 (January through early March) at a coastal site, the Scripps Institution of Oceanography (SIO) pier in La Jolla, California. The measurements included direct atmospheric transmittance in order to derive the aerosol optical thickness, and sky radiance to be interpreted in relation to the aerosol scattering phase function, and they were performed in the spectral range 440 to 1020 nm. To investigate the spatial variability of the atmospheric optical properties, measurements of direct atmospheric transmittance were also performed at other sites during a few days in January 1994. The other sites were located on Catalina Island off Long Beach, California, and along the R/V Wecoma ship track during the 1994 California Cooperative Fisheries Investigation (CalCOFI) winter cruise in the Southern California Bight.

3.1. Instrumentation

3.1.1. Description. The instrument used to investigate aerosol optical properties was a Sun photometer–sky radiometer system developed by CIMEL Electronique [Holben et al., 1994]. This instrument measures the direct atmospheric transmittance at 1020, 870, 670, and 440 nm, and in additional bands located in the 940 nm water vapor absorption region. These additional bands were not used in this study. Sky radiance is also measured at 1020, 870, 670, and 440 nm. All the bands are 10 nm wide, and the field of view is 1.2°. The CIMEL instrument, when taking sky radiance measurements, was used in a semiautomatic mode, that is, the sky was scanned automatically at 40 different scattering angles. The angles of interest for our study were mainly 60°, 90°, and 120°. The Sun photometer–sky radiometer was installed at the end of the SIO pier, in order to be as much as possible under marine conditions and to minimize coastal effects (ground reflectance influence). The

Figure 2. Spectral dependency of the aerosol reflectance between 870 and 1020 nm as a function of scattering angle. The aerosol reflectance is computed using the single-scattering approximation: (a) for maritime, urban, and continental models of WMO and (b) for five models of Shettle and Fenn: T70, T98, M70, M98, and C70.
measurements (sky radiance, atmospheric transmission) were performed under clear sky conditions, generally from sunrise to sunset. The instruments used during the CalCOFI cruise and on Catalina Island were portable Sun photometers, also designed by CIMEL Electronique, but to perform optical thickness measurements at 1020, 870, 670, and 440 nm only.

Meteorological data were collected to correct the optical measurements and help with the analysis. They included wind speed, wind direction, and surface atmospheric pressure, measured every 5 min on the SIO pier, and air and dew point temperatures have been measured hourly at several meteorological stations located in the San Diego area. The SIO pier data and other meteorological data were made available by SIO staff and the National Center for Atmospheric Research, respectively.

3.1.2. Radiometric calibration. The automatic CIMEL instrument operating in sky radiance mode was calibrated in the field using a total irradiance plaque of known reflectance properties. The measurements were performed in very clear atmospheric conditions at Stevenson Peak (altitude, 1896 m), Laguna Mountains, California, on January 16, 1994. The sky radiometer was viewing vertically the calibration plaque placed horizontally. This method was also performed on the SIO pier during very clear days, for instance on January 29, 1994, and the results were in good agreement with those obtained in the Laguna Mountains. Calibration of the CIMEL instrument operating in Sun mode and of the hand-held CIMEL instruments was also performed on January 16, 1994, at Stevenson Peak, but using the Bouguer-Langle Che method.

For the radiance calibration the downward irradiance on the plaque, including the contribution of photons that had interacted with the surface and the atmosphere, was computed using the successive orders of scattering code of Deuzé et al. [1989] and measurements of the aerosol optical thickness at the calibration site. Some assumptions were made about the type of aerosol and ground reflectance. Knowledge of these parameters is not critical, however, considering the low optical thickness of the aerosols and the fact that only a small fraction of photons reflected by the ground reached the scattering plaque.

Two additional radiance calibrations were performed in the laboratory on July 1993 and April 1994 using an integrating sphere of well-known leaving radiance. The results of these two types of calibration, as well as the results of the Stevenson Peak and SIO pier calibrations, are presented in Table 2 for the two different gains used in the observation of sky radiance, “S” (sky) for scattering angles of more than 6° and “A” (aureole) for scattering angles of less than 6°.

<table>
<thead>
<tr>
<th>Channel, nm</th>
<th>Gain</th>
<th>K From Plaque on Jan. 16, 1994, at Stevenson Peak</th>
<th>K From Plaque on Jan. 29, 1994 at SIO Pier</th>
<th>K From Sphere in July 1993</th>
<th>K From Sphere in April 1994</th>
<th>Error Maximum</th>
<th>Plaque/Sphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>1020</td>
<td>S</td>
<td>8.45 × 10^{-5}</td>
<td>8.36 × 10^{-5}</td>
<td>9.02 × 10^{-5}</td>
<td>9.02 × 10^{-5}</td>
<td>0.08</td>
<td>0.07</td>
</tr>
<tr>
<td></td>
<td>A</td>
<td>7.82 × 10^{-5}</td>
<td>7.85 × 10^{-5}</td>
<td>8.86 × 10^{-5}</td>
<td>8.37 × 10^{-5}</td>
<td>0.10</td>
<td>0.10</td>
</tr>
<tr>
<td>870</td>
<td>S</td>
<td>8.20 × 10^{-5}</td>
<td>8.35 × 10^{-5}</td>
<td>9.15 × 10^{-5}</td>
<td>9.42 × 10^{-5}</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td></td>
<td>A</td>
<td>1.05 × 10^{-3}</td>
<td>1.13 × 10^{-3}</td>
<td>8.67 × 10^{-5}</td>
<td>8.10 × 10^{-3}</td>
<td>0.12</td>
<td>0.06</td>
</tr>
<tr>
<td>670</td>
<td>S</td>
<td>6.18 × 10^{-5}</td>
<td>6.24 × 10^{-5}</td>
<td>6.63 × 10^{-5}</td>
<td>6.25 × 10^{-5}</td>
<td>0.08</td>
<td>0.02</td>
</tr>
<tr>
<td></td>
<td>A</td>
<td>6.69 × 10^{-4}</td>
<td>6.68 × 10^{-4}</td>
<td>6.61 × 10^{-4}</td>
<td>0.02</td>
<td>-0.02</td>
<td>-0.02</td>
</tr>
</tbody>
</table>

Values of K are in W m^{-2} sr^{-1} nm^{-1} per numerical count. The two rightmost columns give the maximum relative error, (K_{max} - K_{min})/K_{sphere} and the relative error between the plaque field calibration on January 16, 1994, and the sphere laboratory calibration on April 1994. S (sky) is for scattering angles of more than 6° and A (aureole) is for scattering angles of less than 6°.
tation of satellite signal in the solar spectrum (5S) code [Tanré et al., 1990].

At 1020 nm the sensitivity of the detector is strongly affected by ambient temperature. For each measurement of sky radiance or direct transmission, the temperature of the detector was also measured. A correction was applied in order to take into account detector temperature variation between the day of calibration and the days of measurements.

The Sun-Earth distance factor \( D_0 \) is computed according to Paltridge and Platt [1977]:

\[
D_0(J) = 1 - e \cos \left(0.017201(J - 4)\right)
\]

where \( J \) is the Julian day and \( e = 0.01673 \) is the eccentricity factor. The next step is to write

\[
T = \exp \left(-m \tau\right)
\]

where \( m \) is the air mass and \( \tau \) is the total (i.e., aerosols and molecules) atmospheric optical thickness. The air mass \( m \) has been computed according to Kasten and Young [1989]:

\[
m(\theta_s) = 1/\left[\cos \theta_s + a(\theta_s + b)^{-e} \right]
\]

with \( a = 0.50572 \) and \( b = 6.07995 \) for \( \theta_s \) expressed in degrees, and \( e = 1.6364 \). The aerosol optical thickness \( \tau_a(\lambda) \) is derived eventually by subtracting the Rayleigh optical thickness \( \tau_r(\lambda) \) from the total optical thickness \( \tau(\lambda) \). The Rayleigh optical thickness is taken following Hansen and Travis [1974] and Gordon et al. [1988], but for a depolarization factor of 0.0279: \( \tau_r(\lambda) = (0.008524 \lambda^{-4} + 0.0000963 \lambda^{-6}) + 0.00000111 \lambda^{-7} / P/\rho_0 \) (12)

where \( P \) is the surface atmospheric pressure, actually measured at SIO pier, and \( \rho_0 = 1013.25 \) hPa is the standard atmospheric pressure.

The Ångström coefficient \( \alpha \) is determined as the slope of a linear regression, on a log-log scale, between \( \tau_a(\lambda) \) and \( \lambda \), for the four wavelengths of the measurements, 440, 670, 870, and 1020 nm. Experimental results of such a regression are given in Figure 3, for different values of the aerosol optical thickness. In Figure 3, one notices that the determination of \( \alpha \) becomes more difficult at low aerosol optical thickness, simply because the errors on the derived \( \tau_a(\lambda) \) are rather constant in absolute value, and then become increasingly important in relative value.

3.2.2. Aerosol scattering phase function retrieval. The aerosol phase function is derived from the measured sky radiances at given angles. First, for convenience, the sky radiance is converted in sky reflectance \( \rho \) according to (1). Second, the Rayleigh reflectance \( \rho_r \) is computed exactly for a pure molecular atmosphere using the successive orders of scattering code of Deuzé et al. [1989]. In the computations the Rayleigh optical thickness is given by (12), and the Rayleigh phase function is taken with a depolarization factor of 0.0279 according to Young [1981]. Third, the Rayleigh reflectance is subtracted from the sky reflectance to derive an aerosol reflectance \( \rho_a \) using

\[
\rho_a = \rho - \rho_r
\]

Finally the phase function \( \rho_a(\theta) \) at the scattering angle \( \theta \) of the observation is derived from the aerosol reflectance, using the classical single-scattering approximation [Tanré et al., 1979; Gordon et al., 1983]:

\[
\rho_a = \omega_a \tau_a P_a(\theta) / 4 \mu_a \mu_r
\]

Actually, the single-scattering albedo \( \omega_a \) remains unknown, and it is the product \( \omega_a P_a \) which is derived and hereafter referred to as the "pseudo" phase function.

Equation (14) is the first order of the development at low optical thickness of the exact single-scattering formulation, given by, at the top of the layer, for observations from space

\[
\rho_a = \omega_a P_a(\theta) \left[ 1 - \exp \left(-\tau_a(1/\mu_r + 1/\mu_a)\right) \right]/(\mu_r + \mu_a)
\]

or, at the bottom of the layer, for ground-based observations

\[
\rho_a = \omega_a P_a(\theta) \left[ 1 - \exp \left(-\tau_a(1/\mu_r + 1/\mu_a)\right) \right] \cdot \exp \left(-\tau_a/\mu_a\right) / (\mu_r + \mu_a)
\]

Equation (14) overestimates the single-scattering effect and may be a good estimation of the multiple-scattering effect [Tanré et al., 1990]. It is rather accurate at 870 and 1020 nm, where the coupling term between aerosols and molecules is negligeable. Because of the attenuation term exp \( (-\tau_a/\mu_a) \) in (16), we tested an alternative approximation for \( \rho_a \) at the bottom of the layer, with attenuation through the layer:

\[
\rho_a = \omega_a \tau_a P_a(\theta) \exp \left(-\tau_a/\mu_a\right) / (\mu_r + \mu_a)
\]

Radiative transfer simulations were made to test the validity of (14) and (17) for ground-based observations. The simulated atmosphere was a mixture of molecules and aerosols, with respective scale heights of 8 and 2 km, and a flat water surface at the bottom. The simulations were made at 870 nm for two of the above described models (tropospheric with a relative humidity of 70%, T70, and maritime with a relative humidity of 98%, M98), two aerosol optical thicknesses of 0.1 and 0.2, and two solar zenith angles, 39.9° and 80.7°. An aerosol "pseudo" phase function was then derived from the simulation, using (14) or (17), after subtracting the Rayleigh scattering.

The actual and derived "pseudo" phase functions of aerosol models T70 and M98, are plotted in Figures 4a and 4b, for \( \theta_a = 0.1 \) and two solar zenith angles, \( \theta_s = 39.9° \) and 80.7°, respectively. Equation (17) underestimates \( \rho_a \) because it includes a too large exponential attenuation and leads to a systematic overestimation of the derived phase function in the backward directions. Equation (14) gives far better results,
particularly at large scattering angles; but its accuracy is rapidly degraded at viewing zenith angles above 75° for both solar zenith angles. For θs < 75°, the worst case is observed for the maritime model M98, which gives a relative error of 10% and 20% in the range of scattering angles 100°–120°. The accuracy of (14) is limited by its ability to represent multiple scattering for such a phase function presenting a well-marked minimum.

Tables 3a and 3b summarize for the above described conditions the exact and deduced "pseudo" aerosol phase function at a scattering angle of 60° as well as the relative error (deduced — exact)/exact) induced by the approximations at 870 and 1020 nm, respectively. The relative error varies with the model and geometry, and increases roughly like the aerosol optical thickness representing the ratio of multiple (second order) to single scattering. At the other scattering angles, 90° and 120° (not shown here), aerosol scattering is lower and multiple scattering is dominated by aerosol-molecule scattering interactions. In this range of scattering angles, the ratio of multiple to single scattering depends on the molecular optical thickness but not on the aerosol optical thickness. As a result, the relative error on the retrieval of the phase function is rather constant with optical thickness at large scattering angles and varies only with solar zenith angle and aerosol type.

4. Results

4.1. Aerosol Optical Thickness and Angström Coefficient Frequencies

The aerosol optical thickness at 870 nm and Angström coefficient have been derived from the four spectral measurements of the direct atmospheric transmission made by the CIMEL instrument in bands centered at 440, 670, 870, and 1020 nm. The measurements were performed systematically when the sky could be considered as clear, actually when the cloud cover was estimated to be one tenth or less, allowing for sky radiances measurements in the principal plane. When several measurements were made within 15 min, the derived aerosol optical thicknesses at 870 nm and Angström coefficients were averaged. Twenty-five and thirty-four days of data were acquired during the periods January 23 to March 11, 1993, and January 7 to March 9, 1994, respectively.

Figures 5a and 5b present the frequency histograms of the

---

**Table 3a.** Comparison Between the Actual and Retrieved (by Approximate Equations (14) and (17)) Values of the "Pseudo" Phase Function \( \alpha_{\lambda \varphi}(60°) \) at 870 nm for Aerosol Models T70 and M98 and for a Scattering Angle of 60°

<table>
<thead>
<tr>
<th>( \tau_a )</th>
<th>Aerosol Model</th>
<th>Actual</th>
<th>Retrieved by (14)</th>
<th>Relative Error by (14)</th>
<th>Retrieved by (17)</th>
<th>Relative Error by (17)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>T70</td>
<td>0.928</td>
<td>0.903</td>
<td>-0.03</td>
<td>1.004</td>
<td>0.09</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.402</td>
<td>-0.01</td>
<td>0.448</td>
<td>0.11</td>
</tr>
<tr>
<td>62.17</td>
<td>T70</td>
<td>0.928</td>
<td>0.989</td>
<td>-0.05</td>
<td>0.982</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.402</td>
<td>-0.01</td>
<td>0.444</td>
<td>0.10</td>
</tr>
<tr>
<td>80.72</td>
<td>T70</td>
<td>0.928</td>
<td>0.738</td>
<td>-0.21</td>
<td>0.822</td>
<td>-0.12</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.363</td>
<td>-0.11</td>
<td>0.404</td>
<td>0</td>
</tr>
<tr>
<td>0.2</td>
<td>T70</td>
<td>0.928</td>
<td>0.879</td>
<td>-0.06</td>
<td>1.088</td>
<td>0.18</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.402</td>
<td>-0.01</td>
<td>0.497</td>
<td>0.24</td>
</tr>
<tr>
<td>62.17</td>
<td>T70</td>
<td>0.928</td>
<td>0.850</td>
<td>-0.09</td>
<td>1.039</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.400</td>
<td>-0.01</td>
<td>0.488</td>
<td>0.21</td>
</tr>
<tr>
<td>80.72</td>
<td>T70</td>
<td>0.928</td>
<td>0.617</td>
<td>-0.34</td>
<td>0.765</td>
<td>-0.18</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.336</td>
<td>-0.17</td>
<td>0.416</td>
<td>0.03</td>
</tr>
</tbody>
</table>

**Table 3b.** Comparison Between the Actual and Retrieved (by Approximate Equations (14) and (17)) Values of the "Pseudo" Phase Function \( \alpha_{\lambda \varphi}(60°) \) at 1020 nm for Aerosol Models T70 and M98 and for a Scattering Angle of 60°

<table>
<thead>
<tr>
<th>( \tau_a )</th>
<th>Aerosol Model</th>
<th>Actual</th>
<th>Retrieved by (14)</th>
<th>Relative Error by (14)</th>
<th>Retrieved by (17)</th>
<th>Relative Error by (17)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>T70</td>
<td>0.941</td>
<td>0.920</td>
<td>-0.03</td>
<td>0.997</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.403</td>
<td>-0.01</td>
<td>0.447</td>
<td>0.11</td>
</tr>
<tr>
<td>62.17</td>
<td>T70</td>
<td>0.941</td>
<td>0.911</td>
<td>-0.04</td>
<td>0.982</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.403</td>
<td>-0.01</td>
<td>0.445</td>
<td>0.10</td>
</tr>
<tr>
<td>80.72</td>
<td>T70</td>
<td>0.941</td>
<td>0.799</td>
<td>-0.16</td>
<td>0.866</td>
<td>-0.08</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.372</td>
<td>-0.08</td>
<td>0.414</td>
<td>0.03</td>
</tr>
<tr>
<td>0.2</td>
<td>T70</td>
<td>0.941</td>
<td>0.901</td>
<td>-0.05</td>
<td>1.058</td>
<td>0.13</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.403</td>
<td>-0.01</td>
<td>0.497</td>
<td>0.24</td>
</tr>
<tr>
<td>62.17</td>
<td>T70</td>
<td>0.941</td>
<td>0.880</td>
<td>-0.07</td>
<td>1.023</td>
<td>0.09</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.402</td>
<td>-0.01</td>
<td>0.490</td>
<td>0.22</td>
</tr>
<tr>
<td>80.72</td>
<td>T70</td>
<td>0.941</td>
<td>0.693</td>
<td>-0.27</td>
<td>0.815</td>
<td>-0.14</td>
</tr>
<tr>
<td></td>
<td>M98</td>
<td>0.404</td>
<td>0.346</td>
<td>-0.15</td>
<td>0.427</td>
<td>0.06</td>
</tr>
</tbody>
</table>
aerosol optical thickness at 870 nm, $\tau_a(870)$, during 1993 and 1994. More than 50% of the observed optical thicknesses are about 0.1 or less. The values of $\tau_a(870)$ in 1993 are slightly higher than in 1994. This simply illustrates the influence of stratospheric aerosols. According to the Stratospheric Aerosol and Gas Experiment 2 (SAGE 2) observations during the period following the eruption of the Pinatubo volcano, the stratospheric optical thickness at 870 nm, $\tau_{a,\text{strato}}(870)$, had a mean value of 0.04 and 0.02 during the 1993 and 1994 winters, respectively (see Table 4). After subtraction of the stratospheric component, one may deduce the tropospheric component of the aerosol optical thickness, $\tau_{a,\text{tropo}}(870)$, which is presented in Figure 5c. Most of the values of $\tau_{a,\text{tropo}}(870)$ are below 0.1.

As detailed below (section 5), the accuracy of the aerosol optical thickness retrieved from the measurements is a few $10^{-2}$. The stratospheric aerosol contribution derived from SAGE data is known with an accuracy of better than 0.005, which represents the dispersion of the measurements acquired in the vicinity of La Jolla (Table 4).

The aerosol variability must be analyzed in terms of not only its load, represented by its optical thickness, but also its type, represented by the Ångström coefficient. In Figures 7a and 7b are displayed scatterplots of Ångström coefficient versus aerosol optical thickness for the measurements made in 1993 and 1994. They confirm the previous analysis. At low optical thickness the measurements tend to a limit which should be the sum of stratospheric and background tropospheric aerosol components. This limit appears better defined in 1994, with $\tau_a = 0.03$ and $\alpha$ around 1.2.

The Ångström coefficient of the stratospheric component is also well known from the SAGE experiment (see Table 4). Values of $\alpha_{\text{strato}}$ are lower in 1993 than in 1994. This is explained by the post-Pinatubo conditions in 1993, with large particles injected by the eruption into the stratosphere. The estimated stratospheric conditions are indicated by a solid circle in Figure 6.

The inaccuracy of the derived Ångström coefficient, also discussed below (section 5), increases when $\tau_a$ decreases, roughly like $1/\tau_a$. Typical errors are 0.2 when $\tau_a = 0.1$, 0.4 when $\tau_a = 0.05$, and 1 when $\tau_a = 0.02$. This illustrates the difficulties in estimating accurately the Ångström coefficient at low optical thickness. Accordingly, at the lowest optical thickness, that is, at $\tau_a < 0.1$, the values of the derived Ångström coefficient are widely spread all over the geophysical range of possible values, 0 to 1.5. Thus the discussion about aerosol types is limited at low optical thickness. It is only at aerosol optical thickness above 0.1 that some trends in the Ångström coefficients and, thus, aerosol type are detectable. For both years, when $\tau_a$ increases, one identifies two modes in Figure 6: one with $\alpha$ around 1.2, which would correspond to a tropospheric model, and the other with $\alpha$ around 0.2, which would correspond to the maritime model.

The relationship between Ångström coefficient and meteorological variables (wind speed and direction, relative humidity) has been examined, using measurements performed during the most turbid days (aerosol optical thickness above 0.1 at 870 nm). A correlation was found between Ångström coefficient and cast-west component of the wind, but only in the presence of established or sustained winds. In these conditions, low Ångström coefficients (<0.8) are associated with westerly flow and high coefficients (>0.8) with easterly flow. When wind speed is low (<2 m s$^{-1}$), however, the Ångström coefficient is variable, and the correlation with wind direction is not significant. Also, the data did not show a clear relationship between Ångström coefficient and relative humidity, a factor of aerosol growth and, therefore, an indicator of particle size. Even though relationships between aerosol optical variables and me-
Table 4. Stratospheric Aerosol Optical Thicknesses, From SAGE 2 Observations

<table>
<thead>
<tr>
<th>Number of SAGE Observations</th>
<th>Observed $\tau_{a,\text{strato}}(\lambda)$ at 1 $\mu$m</th>
<th>Observed $\tau_{a,\text{strato}}(\lambda)$ at 525 nm</th>
<th>Derived $\tau_{a,\text{strato}}(\lambda)$ at 870 nm</th>
<th>Derived $\alpha_{\text{strato}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1993: $z &gt; 13.5$ km</td>
<td>8</td>
<td>0.0432 ± 0.0056</td>
<td>0.0472 ± 0.0064</td>
<td>0.0442</td>
</tr>
<tr>
<td>1994: $z &gt; 9.5$ km</td>
<td>7</td>
<td>0.0178 ± 0.0045</td>
<td>0.0275 ± 0.0072</td>
<td>0.0199</td>
</tr>
<tr>
<td>1994: $z &gt; 13.5$ km</td>
<td>10</td>
<td>0.0114 ± 0.0036</td>
<td>0.0184 ± 0.0044</td>
<td>0.0129</td>
</tr>
</tbody>
</table>

Courtesy of C. Brogniez (unpublished data, 1995).

teorological variables may be found, they are not general, presumably because atmospheric changes (vertical mixing, diurnal breeze effects) may result in varied combinations of maritime, continental, and urban aerosols at the SIO pier.

Regarding spatial variability, aerosol optical measurements made concomitantly at the SIO pier, on Catalina Island, and in the Southern California Bight during a 2-week period starting in late January 1994 (CalCOFI winter cruise) have been analyzed. R/V Wecoma was at least 200 km offshore during clear days, except on January 28 (80 km). Optical thickness was always low ($\tau_a < 0.07$ at 870 nm) on Catalina Island, located about 40 km off the mainland, and most of the time lower than optical thickness at the SIO pier. The highest values were measured on January 20, 1994, the day of highest aerosol load ($\tau_a$ of 0.15–0.2 at 870 nm) at the SIO pier. The highest value measured on board R/V Wecoma was 0.055, and it was obtained on January 29. The Ångström coefficient often was very low (<0.2) at the R/V Wecoma location, which is consistent with the predominance of maritime aerosols, and was generally lower on Catalina Island than at the SIO pier. On January 29, however, Ångström coefficients of about 1.2 were observed about 240 km from the coast, indicating the presence of aerosols of continental origin. Thus on some occasions the influence of the continent and coastal urban areas may be felt far offshore.

4.2. Sky Radiances

4.2.1. Phase function and Ångström coefficient. Phase functions derived from the sky radiance measurements at 870 nm were computed using the procedure described in section 3. This procedure implies some simplifications: A single-scattering approximation for the expression of the aerosol radiance is used, the ground effect is neglected, and the result is actually the product of the phase function and the single-scattering albedo ("pseudo" phase function). The analysis has been restricted to the morning data, before local noon, to minimize the ground effect: At scattering angles of 90° and 120°, the atmospheric path is above the ocean surface, and, therefore, the reflection from the land is reduced (but not negligible). Because the retrieval of both the "pseudo" phase function and the Ångström coefficient are increasingly inaccurate as the aerosol optical thickness decreases, the analysis has only been applied to observations involving an aerosol optical thickness above 0.075 at 870 nm.

As shown by the models of SF, as well as those of WMO, there is a tendency for the phase function to decrease around its minimum at about 100°, when the Ångström coefficient decreases (see Table 1 and Figures 1a and 1b). This is the classical result of increasing the size of the particles whatever the refractive index, that one can check against the observations.

"Pseudo" phase functions at scattering angles of 60°, 90°, and 120° are plotted versus the Ångström coefficient in Figure 7a–7c for the SF models and the observations. The values for the models are taken from Table 1. The relationship between Ångström coefficient and "pseudo" phase functions is analyzed by linear regression. The correlation coefficient for the models, $R^2$, varies from 0.74 at 60° to 0.79 at 120°, and the corresponding "pseudo" phase functions of the selected models can be

Figure 6. Ångström coefficient versus aerosol optical thickness at 870 nm: measurements (pluses) collected in (a) 1993 and (b) 1994. The stratospheric component, given by Stratospheric Aerosol and Gas Experiment (SAGE) data, is identified by a solid circle.
retrieved from the Ångström coefficient with an accuracy of 0.081, 0.030, and 0.016, respectively. Thus the model "pseudo" phase functions, which influence directly the sky radiance, can be predicted rather accurately by the Ångström coefficient.

This is confirmed by the observations also given in Figures 7a–7c for the same angles as the models. The linear regressions performed on the data show very similar intercepts and slopes at 90° and 120° and coefficients smaller by 20% at 60° when compared to the models. The data may be classified in three groups (see Table 5): a large group of 45 measurements around α = 1.3, a smaller group of 14 measurements around α = 0.5, and three measurements only around α = 0. The first group fits the SF T90 model; the second typifies the SF coastal model. It is not clear whether the three measurements in the last group correspond to very maritime conditions or to undetected thin clouds. Table 5 gives the mean and standard values of the observed phase functions for each data set. The standard deviation of the "pseudo" phase functions is about 0.02 at 90° and 120° for the two main significant data sets. This value of 0.02 can be retained as indicative of the accuracy of the phase function retrieval from the Ångström coefficient.

Previous investigators [Kaufman, 1993; Kaufman et al., 1994] have also attempted to retrieve the phase function at the scattering angle of 120° from sky radiance measurements. Table 1 of Kaufman [1993] gives the measurements of aerosol optical thickness, Ångström coefficient, and sky radiance at 613 nm, for several continental locations. The optical thickness range is 0.03–1.53, and the Ångström coefficient varies between 0.09 and 1.94. After correcting the data for molecular scattering, we have not been able to establish a relationship between α and φ(a, θ) at 613 nm, and we have only been able to retrieve a mean value of 0.2 for φ(a, θ) at 613 nm. This value is slightly larger than that predicted by most models, see Table 1. Kaufman et al. [1994] in their Table 1 provide other results of a more thorough investigation at five locations. They also give values of the phase function at 870 nm for a scattering angle of 120°. Their derived phase functions are higher than those of most models, but some relationship with the Ångström coefficient appears to exist, except for one of the sites, located in Brazil. Our study gives results at several scattering angles, which better agree with models than previously reported elsewhere.

In summary, two main points emerge from the analysis: (1) within the accuracy of the measurements, the SF models do allow one to describe the observed Ångström exponents and phase functions; and (2) the phase function can be predicted from the Ångström coefficient with a typical accuracy of 0.02 at scattering angles in the range 90°-120°.

4.2.2. Wavelength dependence of the aerosol radiance.

Computed values for the SF models show a relationship between the spectral dependence of the aerosol sky radiances, ε_a(870, 1020), and the Ångström coefficient α (Figures 8a–8c). The analysis of the actual measurements, however, has been rather disappointing on that point. The data set was carefully restricted to aerosol optical thicknesses around 0.1 (from 0.08 to 0.12) in order to avoid an effect of the optical thickness. Aerosol radiances were derived at 670, 870, and 1020 nm, after subtracting the computed molecular radiances, for scattering angles θ = 60°, 90°, and 120°. Then, the ratio of two aerosol radiances at two wavelengths, ε_a(670, 870, θ) or ε_a(870, 1020, θ), was calculated.

The only significant correlation between α and ε_a is obtained with ε_a(870, 1020, θ) for the scattering angle of 60° ($R^2 = 0.40$), and to a lesser degree of 120° ($R^2 = 0.14$), as shown in Figures 8a–8c. The agreement between measurements and models is rather good again for the scattering angle of 60°. In all other cases the inaccuracy of the derived coeffi-
5. Accuracy of the Measurements

Accuracy of the derived parameters, that is, aerosol optical thickness, Ångström coefficient, phase function, and spectral dependency of the aerosol reflectance, has been estimated. The various types of error have been regrouped into two major categories: (1) errors due to instrumentation and methods such as calibration, stability of the detectors, and accuracy of the formulas and of the exogen parameters used in the computations and (2) errors due to geophysical conditions such as the presence of clouds, stratospheric aerosols, or interaction between ground and atmosphere.

In the following, errors in the computation of solar zenith angle, air mass, and Sun-Earth distance factor have been neglected. When no sign precedes the error, the error can be positive or negative.

5.1. Errors Due to Instrumentation and Methods

5.1.1. Aerosol optical thickness. The accuracy on the calibration coefficient $\Delta \ln(I_0)m$ has been estimated using the five most regular Langley plots of all the 1994 campaign obtained with the automatic CIMEL instrument. The accuracy on the Rayleigh optical thickness $\Delta \tau_r$ has been computed assuming a 1% relative error on the computation of Rayleigh optical thickness at standard pressure and a $\pm 10$ hPa error on pressure. Error on the gaseous transmission $\Delta (\tau_g)/\tau_g$ has been neglected, except at 670 nm where it has been computed as the result of a 1 mm error on wavelength and of a 20% relative error on ozone content. Finally, a 5°C error on the temperature of the detector has been assumed, resulting in a relative error of $\Delta S/S = 0.01$ on the sensitivity of the detector at 1020 nm.

The total error is the quadratic sum of the above parameters. The final results are displayed in Table 6a. The retrieval of the Ångström coefficient is affected by the errors on the optical thicknesses, increasingly when the aerosol optical thickness decreases. The absolute error on $\tau_a$ does not depend on the actual value of $\tau_a$. Thus the relative error on $\tau_a$ varies like $1/\tau_a$ and becomes very large for low values of the optical thickness (e.g., 60% at 440 nm when $\tau_a = 0.05$).

5.1.2. Aerosol reflectance. The study has been performed at 870 and 1020 nm only. The error $\Delta \rho_a$ on the aerosol reflectance is the sum of the error $\Delta \rho$ on the measured reflectance, $\Delta \rho = \rho K/K$ where $K$ is the calibration coefficient, all other error sources being negligible, and the error $\Delta \rho_a$ on the molecular reflectance. The relative error on the calibration coefficient is about 2%. The relative error $\Delta \rho_a/\rho_a$ has been simulated for a scattering angle of 90° which shows weak variations with solar zenith angle for solar zenith angles smaller than 80°. At 870 nm, as well as at 1020 nm, the inaccuracy on $\rho_a$ is less than 2%. The relative error $\Delta \rho_a/\rho_a$ has been simulated using the SF model C70 for an aerosol optical thickness of 0.03, 0.05, 0.1, and 0.2. The results are displayed in Table 6b. The inaccuracy decreases with increasing aerosol optical thickness, from 9% to 3% at 870 nm and 6% to 3% at 1020 nm.

5.1.3. Spectral dependency. The relative error on the spectral dependency $e$ of the aerosol reflectance is the sum of the relative error on the aerosol reflectance at the two wavelengths used to compute $e$, that is, 870 and 1020 nm in our case:

$$\Delta e_a(870, 1020)/e_a(870, 1020) = \Delta \rho_a(870)/\rho_a(870) + \Delta \rho_a(1020)/\rho_a(1020)$$

(18)

The results are dependent upon aerosol optical thickness, indeed (see Table 6b). As for the aerosol reflectance, the error on $e_a$ decreases with increasing optical thickness, from 16% at $\tau_a = 0.03$ to 6% at $\tau_a = 0.2$.

5.1.4. Aerosol “pseudo” phase function. The relative error on the retrieved aerosol “pseudo” phase function is the sum of the relative error on the aerosol optical thickness and on the aerosol reflectance. An additional error results from the approximation of the aerosol reflectance (e.g., single-scattering approximation) and has been simulated for the model C70 and a scattering angle of 90°. Both errors are represented in Table 6c. Inaccuracy is large for small aerosol amount (up to 50% for $\tau_a = 0.03$) and decreases with increasing aerosol optical thickness.

5.2. Errors Due to Geophysical Conditions

Errors due to geophysical conditions are more difficult to estimate. The presence of stratospheric aerosols perturbs the retrieval of tropospheric aerosol parameters. We have seen (section 4.1) that a significant relative amount of stratospheric aerosol exists. Another perturbation may be the presence of clouds, contaminating either Sun measurements, sky measurements, or both. We concentrated our measurements during clear days, but the presence of invisible thin cirrus is always
possible. A third type of error, which can occur during retrieval of phase function from ground-based measurements, is due to interaction between ground and atmosphere [Yang et al., 1995]. This interaction does not exist or is negligible for observations from a boat or a small island, but is crucial for ground-based measurements performed in coastal (or continental) regions.

Since the coastline is approximately in the north-south direction at the SIO pier, the Sun is above the land surface in the morning, and the sky radiance measurements of $\rho(60^\circ)$ to $\rho(120^\circ)$ are made with the instrument viewing above the ocean surface. In this configuration the ground effect is negligible. In the afternoon, $\rho(60^\circ)$ to $\rho(120^\circ)$ are measured over the land surface, and interaction between surface reflection and atmospheric scattering may not be negligible any more.

Simulations were performed in order to study the land surface effect. Sky radiances, as seen from the ground, were simulated at 870 nm for the models T70 and M98 with two aerosol amounts ($\tau_a = 0.1$ and $\tau_a = 0.2$ at 870 nm) for a solar zenith angle of 39.9°, 62.1°, and 80.7°. Several values of the ground reflectance were used, namely 0, 0.2, 0.5, and 1 (academic case). Table 6d summarizes the results and gives the normalized radiance of the sky $R_n$ and the relative error, [deduced - exact]/exact, due to neglecting the coupling term. Neglecting the interaction between land surface reflection and atmospheric scattering overestimates the derived aerosol phase function. We can also notice that in Table 6d the coupling between ground and atmosphere is more important for a small solar zenith angle.

Interaction between glitter and atmosphere was also studied. At our coastal site, where land is to the east, during the morning the glitter can only influence the directions at a lower elevation from the direction of specular reflection, but these directions are not used for sky radiance measurement. During the afternoon, however, the glitter influence is effective. The normalized radiance of the sky for models T70 and M98 has been computed, for a scattering angle of 90°, a black surface and no glitter (a flat water surface), and a water surface agitated by a wind speed of 10 m s$^{-1}$. For the relation between wind speed and wave slope we used the theory developed by Cox and Munk [1954]. We found (see Table 6e) that the effect of interaction between glitter and atmosphere is small compared to that of interaction between ground and atmosphere, except at large Sun zenith angles.

6. Discussion

6.1. Aerosol Loading

Measurements at SIO pier during the 1993 and 1994 winters indicate a high frequency of aerosol optical thicknesses below 0.1 with values often around 0.05 at 870 nm. This is well below what is considered as a tropospheric background in most of the comprehensive models [Jura, 1985; WMO, 1986].

Such low values have been reported in the literature for ground-based measurements in the Antarctic [Obleitner, 1992; Herber et al., 1993], or satellite surveys using NOAA advanced very high resolution radiometer (AVHRR) over the Pacific [Stowe, 1991]. Some ground-based measurements, however, did not reveal low tropospheric backgrounds. In the work by Kaufman [1993], only 25% of the measured aerosol optical thicknesses are below 0.12 at 613 nm (0.1 at 870 nm), with a large contribution of Australian sites, while the other sites are mainly European. The SIO pier site is under the influence of an eastward airflow and far from any main terrestrial source (Asia). Natural deposition and scavenging by rainfall occurs during the long westerly journey, and these Asian aerosols do not reach the west coast of the United States. The only aerosol sources are local, maritime when there is a strong west wind regime, and urban when the breeze regime brings offshore the
local production of San Diego and Los Angeles agglomerations. Over the ocean, Hoppel et al. [1990] show some values of $\beta$, the aerosol optical thickness at 1 $\mu$m, which are lower than 0.08 in the western section of their cruise across the North Atlantic tropical ocean, that is, the Sargasso Sea. These results were confirmed by Korotkev et al. [1993] over the North Atlantic. In the North Atlantic, Saharan dust outbreaks affect large zones offshore Morocco, Mauritania, and Senegal, increasing the aerosol loading. The earliest report on low aerosol optical thickness at sea is found in the Indian Ocean by Tomasi and Prodi [1982], where they only found occasional values of $\beta$ below 0.1. In the Pacific Ocean, Villevalde et al. [1994] report some measured optical thicknesses at 551 nm which are predominantly below 0.2, and 50% of them below 0.13 at 551 nm, with the lowest optical thickness observed for a low wind speed.

It should be noted that our measurements were made during the winter season only. One expects similar results for the other seasons, however, because the region does not experience drastic weather changes during the year. In summer, sea breezes are well developed, and they produce thick coastal fog that can extend inland. Subsidence generates strong temperature inversions, with pollutants and moisture concentrated under the base of the inversion, resulting in stratocumulus clouds. These pollution conditions, however, can hardly be sampled using Sun photometry, because of the cloud cover. In winter the subsidence and temperature inversions are weaker, and they may be destroyed by the passage of cyclones, facilitating the dispersion of pollutants. In the fall and early winter, the Pacific high occasionally extends eastward, producing a southwesterly flow across California. In the Los Angeles basin the resulting wind is calm, and the effect is more high temperatures and low relative humidities (Santa Ana conditions). We may not have sampled Santa Ana conditions, but many of our measurements were made in the presence of easterly winds.

The level of aerosol loading, via the aerosol optical thickness, does affect directly the accuracy of the retrieval of ocean (color) reflectances from space. The (normalized) radiance scattered by the aerosol $\rho_o$ varies linearly like $\tau_o$:

$$\rho_o \sim \tau_o \rho_o(\theta)/4 \mu_o$$

(19)

where $\rho_o$ is the other aerosol optical parameter of interest. From our results we would take mean values of $\tau_o \sim 0.05$ and of $\rho_o \sim 0.1$ at a scattering angle of 120°, so that $\rho_o \sim 0.002$, which is far less than currently admitted. At this level of $\rho_o$, the atmospheric correction for aerosol scattering is no longer a challenge. Usually, the spectral behavior of $\rho_o$ is determined in the near-infrared and extrapolated at visible wavelengths [Gordon and Wang, 1994]. A relative 10% inaccuracy on this extrapolation would give an error of 0.0002 on $\rho_o$ that meets the requirement for an ocean color mission. Simulations of the atmospheric correction scheme have been done by Gordon and Wang [1994] for a typical aerosol optical thickness of 0.2 at 870 nm. This value appears to be several times higher than standard conditions at sea; thus the errors on the marine reflectance retrievals which are assessed in the same study are also overestimated by the same factor. Results given by Gordon and Wang [1994] are more typical of marginal conditions when the background aerosol is increased by a strong local source, likely terrestrial.

### 6.2. Aerosol Models for Atmospheric Correction of Ocean Color

For atmospheric correction of satellite ocean color radiances, Gordon and Wang [1994] propose to use models derived from Shettle and Fenn [1979], namely, two basic components, maritime and terrestrial models, and a mixture of these two components, a coastal model. In each of these models, physical and optical properties are parameterized as a function of humidity. We have examined whether the selected models are representative of realistic conditions and whether they should be complemented by other models, such as WMO or any other models.

Some of our findings are illustrated in Figure 7, which gives the "pseudo" phase function versus the Ångström coefficient as obtained from models and derived from measurements. These two parameters are key atmospheric correction parameters: The product of the "pseudo" phase function and aerosol optical thickness gives the amplitude of the perturbation, and the Ångström coefficient determines roughly the spectral variation of the perturbation. Thus Figure 7, which shows "pseudo" phase function as a function of Ångström coefficient, does allow one to visualize the main parameters of the atmospheric correction at a specific scattering angle.

Based on our analysis of data and models, we can also make

### Table 6a. Accuracy on the Retrieval of Aerosol Optical Thickness $\tau_o$

<table>
<thead>
<tr>
<th>Wavelength, nm</th>
<th>$\Delta(\ln I_o)/m$</th>
<th>$\Delta\tau_o$</th>
<th>$\Delta(S)/mS$</th>
<th>$\Delta\tau_o$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1020</td>
<td>0.0079</td>
<td>0.00016</td>
<td>0.0078</td>
<td>0.0111</td>
</tr>
<tr>
<td>870</td>
<td>0.0094</td>
<td>0.0030</td>
<td>...</td>
<td>0.0094</td>
</tr>
<tr>
<td>670</td>
<td>0.0164</td>
<td>0.00088</td>
<td>0.02...</td>
<td>0.0239</td>
</tr>
<tr>
<td>440</td>
<td>0.0306</td>
<td>0.0048</td>
<td>...</td>
<td>0.0310</td>
</tr>
</tbody>
</table>

The total error $\Delta\tau_o$ is the quadratic sum of error resulting from the determination of the calibration coefficient $I_o$, the Rayleigh optical thickness $\tau_r$, the ozone transmission $t_o$, and the temperature correction in the 1020 nm band. $\Delta\tau_o$ does not depend on the actual value of the optical thickness $\tau_o$.

### Table 6b. Relative Accuracy on the Retrieval of the Aerosol Reflectance $\rho_o$ at 870 and 1020 nm, and on Its Wavelength Dependence $e(\lambda, 870)$

<table>
<thead>
<tr>
<th>$\lambda$</th>
<th>$\tau_o = 0.03$</th>
<th>$\tau_o = 0.05$</th>
<th>$\tau_o = 0.1$</th>
<th>$\tau_o = 0.2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>870</td>
<td>0.43</td>
<td>+0.11</td>
<td>0.27</td>
<td>+0.12</td>
</tr>
<tr>
<td>1020</td>
<td>0.47</td>
<td>−0.04</td>
<td>0.29</td>
<td>−0.03</td>
</tr>
</tbody>
</table>

$\Delta\rho_o/\rho_o(1020)$

### Table 6c. Relative Accuracy on the Retrieval of the Aerosol "Pseudo" Phase Function for the Model C70 for the Scattering Angle of 90°

<table>
<thead>
<tr>
<th>$\lambda$, nm</th>
<th>$\tau_o = 0.03$</th>
<th>$\tau_o = 0.05$</th>
<th>$\tau_o = 0.1$</th>
<th>$\tau_o = 0.2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>870</td>
<td>0.43</td>
<td>+0.11</td>
<td>0.27</td>
<td>+0.12</td>
</tr>
<tr>
<td>1020</td>
<td>0.47</td>
<td>−0.04</td>
<td>0.29</td>
<td>−0.03</td>
</tr>
</tbody>
</table>

Columns denoted by A give the relative error due to errors on aerosol optical thickness and aerosol reflectance. Columns denoted by B give the relative error due to the single-scattering approximation.
the following remarks: (1) additional WMO models bring very little variability to the SF models, (2) measurements and SF model results do agree significantly within the accuracy of the retrieved parameters, (3) different aerosol models can only be identified from the measurements under marginal conditions, that is, when the aerosol optical thickness is large, (4) under normal conditions, the aerosol optical thickness is low and the measurements suggest a single background aerosol model, and (5) the background model is strongly influenced by the stratospheric component, which in 1994 was different from that in 1993.

These findings and remarks might help design the atmospheric correction scheme. At low optical thickness, under stable stratospheric conditions, it would be appropriate to measure the sky radiance measurements before making those measurements at sea. Our results suggest that Ångström coefficient measurements are sufficient to verify the determination of the aerosol type better than by only using the Ångström coefficient. In most cases, at low aerosol optical thicknesses, the sky radiance measurements are simply not accurate enough to provide useful information. More effort is needed to improve the accuracy of ground-based sky radiance measurements before making those measurements at sea. Our results suggest that Ångström coefficient measurements are sufficient to verify the determination of the aerosol models by the algorithm proposed by Gordon and Wang [1994] and, hence, the amplitude of the atmospheric perturbation on satellite ocean color radiances.

### Table 6e. Computed Normalized Sky Radiance $R_n$ for Different Ground Reflectances, $\rho_s = 0$, 0.2, 0.5, and 1, and a Scattering Angle of $90^\circ$ 

<table>
<thead>
<tr>
<th>$\tau_a$</th>
<th>$\theta_s$</th>
<th>Aerosol Model</th>
<th>$R_n$ for $\rho_s = 0$</th>
<th>$R_n$ for $\rho_s = 0$</th>
<th>$R_n$ for $\rho_s = 0$</th>
<th>$R_n$ for $\rho_s = 0$</th>
<th>$R_n$ for $\rho_s = 0$</th>
<th>$R_n$ for $\rho_s = 0$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>0.0176</td>
<td>T70</td>
<td>0.0186</td>
<td>0.06</td>
<td>T70</td>
<td>0.0186</td>
<td>0.06</td>
<td>T70</td>
</tr>
<tr>
<td>0.0089</td>
<td>M99</td>
<td>0.0097</td>
<td>0.09</td>
<td>0.09</td>
<td>M99</td>
<td>0.0097</td>
<td>0.09</td>
<td>M99</td>
</tr>
<tr>
<td>0.0112</td>
<td>T70</td>
<td>0.0070</td>
<td>0.20</td>
<td>0.20</td>
<td>T70</td>
<td>0.0070</td>
<td>0.20</td>
<td>T70</td>
</tr>
<tr>
<td>0.0119</td>
<td>M99</td>
<td>0.0116</td>
<td>0.23</td>
<td>0.23</td>
<td>M99</td>
<td>0.0116</td>
<td>0.23</td>
<td>M99</td>
</tr>
</tbody>
</table>

### 7. Summary and Conclusions

Direct atmospheric transmittance and sky radiance were measured at 440, 670, 870, and 1020 nm on the Scripps Institution of Oceanography pier during the winters of 1993 and 1994. Additional measurements of direct solar transmittance have been made in 1994 on Catalina Island (about 40 km offshore) and during the CalCOFI 1994 winter cruise. The total and aerosol optical thicknesses, and the spectral dependency of the aerosol optical thickness (Ångström coefficient), were derived from the measurements of the direct atmospheric transmittance. The aerosol scattering "pseudo" phase functions and their spectral dependency between 870 and 1020 nm were derived from the measurements of the sky radiances at the scattering angles of 60°, 90°, and 120°. The optical measurements were performed under varied atmospheric conditions and, therefore, in the presence of diversified aerosol types, including maritime, terrestrial, and urban.

This comprehensive data set on aerosol scattering properties was gathered with respect to two specific objectives. The first objective was to validate the selection of aerosol models used in the algorithms for the atmospheric correction of space-based ocean color observations, such as those of the SeaWiFS mission [Gordon and Wang, 1994]. The second objective was to identify what type of atmospheric optics measurements should be performed to verify atmospheric correction algorithms.

The analysis of data and aerosol models has provided the following main findings and conclusions:
1. Low aerosol optical thicknesses have been observed in the study area, consistent with previous measurements over the oceans. The $\tau_a$ values are influenced by the stratospheric aerosol content, higher in 1993 than in 1994. After correction of the stratospheric content, most of the $\tau_a$ values, that is, $\tau_a_{\text{strat}}$ (870) are below 0.1. In such conditions, it is impossible to identify the variability in the aerosol scattering properties, and one has to rely on a mean background model including the stratospheric component in the atmospheric correction algorithms.

2. Some variability of the aerosol scattering properties has been identified for high optical thickness, that is, above 0.1 at 870 nm. Two modes can be clearly identified. The first mode has an Angström coefficient $\alpha$ around 1.2, corresponding to a tropospheric model. The second mode has $\alpha$ below 0.5 and tending to 0.2, corresponding to a maritime model.

3. Measurements confirm the relationship between $\sin^2\Psi_a$ and $\alpha$ predicted by the models. The dispersion, however, is larger in the measurements than in the model results. The agreement is not good for the relationship between $\varepsilon$ and $\alpha$, partly because of the uncertainties in the derived $\varepsilon$ values. These findings indicate that within the accuracy of the measurements, the SF model allows one to describe the observations of Angström coefficient and phase function. Additional models bring little more. Furthermore, and importantly from a practical point of view, the phase function can be predicted from measurements of $\alpha$. Therefore shipborne measurements of $\alpha$ appear sufficient to validate the aerosol model used in the algorithms. Such measurements were performed during the CalCOFI cruise and they proved reliable.

4. There is no definite relationship between aerosol characteristics and meteorological conditions, but in some instances, small $\alpha$ values are associated with onshore airflow from the ocean and high $\alpha$ values with offshore flow from the land. Spatial variations of $\tau_a$ and $\alpha$ have been observed. In general, $\tau_a$ and $\alpha$ are lower offshore than along the coast, suggesting that the atmospheric correction offshore is much simplified.

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References


Nilsson, Meteorological influence on aerosol extinction in the 0.2-0.4 µm wavelength range, Appl. Opt., 18, 3457-3473, 1979.


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