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Aerosol direct radiative effect at the top of the atmosphere over cloud free ocean derived from four years of MODIS data

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Abstract. A four year record of MODIS spaceborne data provides a new measurement tool to assess the aerosol direct radiative effect at the top of the atmosphere. MODIS derives the aerosol optical thickness and microphysical properties from the scattered sunlight at 0.55–2.1 μm. The monthly MODIS data used here are accumulated measurements across a wide range of view and scattering angles and represent the aerosol's spectrally resolved angular properties. We use these data consistently to compute with estimated accuracy of $\pm 0.6\,\mathrm{W\,m^{-2}}$ the reflected sunlight by the aerosol over global oceans in cloud free conditions. The MODIS high spatial resolution (0.5 km) allows observation of the aerosol impact between clouds that can be missed by other sensors with larger footprints. We found that over the clear-sky global ocean the aerosol reflected $5.3\pm0.6\,\mathrm{Wm^{-2}}$ with an average radiative efficiency of $-49\pm2\,\mathrm{Wm^{-2}}$ per unit optical thickness. The seasonal and regional distribution of the aerosol radiative effects are discussed. The analysis adds a new measurement perspective to a climate change problem dominated so far by models.

1 Introduction

Traditionally, chemical transport and general circulation models enjoyed a monopoly on estimating the role of aerosols in the Earth's climate. Model results form the basis of almost every previous estimate of the aerosol effect on climate (IPCC, 2001). Observations of aerosols from ground-based, airborne or satellite instruments are used only to validate these models. The prevailing strategy dictates that measurements improve models, and then models, not measurements, answer climate questions. However, there is a wide range of discrepancy in model results because of the many inherent assumptions involved in modeling the aerosol effect on climate. Models must properly estimate the source terms of the many aerosol species, properly model the aerosol sink terms, and simulate the transport. Even if the model properly simulates the global distribution of aerosol concentration, assumptions have to be made of the aerosol optical properties in order to convert mass concentrations to the radiative fluxes. Because of the complexity of the problem, it is no wonder that the uncertainties in estimating aerosol effects on climate are growing, rather than shrinking.

To narrow the uncertainties associated with estimating aerosol effects on climate, the time has come to include measurement-based estimates of aerosol radiative effects and forcing. With the launch of EOS-Terra carrying the Moderate resolution Imaging Spectroradiometer (MODIS), Multi-angle Imaging (MISR) and Clouds and Radiant Energy System (CERES), we are suddenly "data rich". These instruments, along with subsequent instruments on EOS-Aqua, EOS-Aura, ICESat, and Parasol, are designed specifically to observe aerosols and the Earth's radiation budget. They provide global information in a way that previous ground-based or airborne instruments could not, and they provide quantitative information about aerosol that is not only more accurate than our heritage instruments, but also more complete in terms of aerosol characterization. With these increased capabilities, aerosol observations from satellite can provide an independent measure of some key climate parameters in parallel with model predictions.

One key measurement that satellites are able to provide is the direct shortwave radiative effect of aerosols at the top of the atmosphere. By aerosol direct shortwave radiative *effect* we mean the difference in shortwave radiative flux between having aerosols present and having no aerosols at all. This is different from aerosol shortwave direct radiative *forcing*, which is the radiative effect of anthropogenic aerosols only. Analysis suggests that by characterizing aerosol particle size from space, there is information available to the satellites to

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classify aerosol into natural and anthropogenic and therefore to determine the anthropogenic portion of the aerosol loading and subsequently determine aerosol forcing from the aerosol effect (Kaufman et al., 2002, 2005a). However, the focus of the present study is the straightforward estimate of aerosol total direct radiative effect.

In this study, we make global and regional estimates of the clear-sky aerosol shortwave radiative effect over the oceans using an internally consistent set of parameters from the MODIS aerosol retrieval. We first put the present study in context with other measurement-based estimates of aerosol effect. We then describe the MODIS aerosol retrieval over ocean and the information available. The paper then describes the radiative transfer model, how we adapt the MODIS data to be used as inputs to the model, how we calculate the regional and global instantaneous and 24 h daily averages of the aerosol direct radiative effect. The results include estimates of monthly mean direct aerosol radiative effect over the oceans, globally and in 13 regional sections, for both the Terra and Aqua satellites.

2 Background

There have been various approaches to using satellite data as the basis for determining aerosol direct radiative effect. One approach is to combine the satellite data with chemical transport model information (Yu et al., 2004). This method allows apportionment of radiative effects to chemical species, but requires assumption of aerosol optical properties. Another approach is to use MODIS to measure aerosol loading in the form of aerosol optical thickness and to use simultaneous observations of the radiation field by CERES (Christopher and Zhang, 2002; Zhang et al., 2005b). Using CERES eliminates the need to assume aerosol optical properties, but does require aerosol dependent angular distribution models (Loeb et al., 2003a, b; Zhang et al., 2005a). Furthermore, the large CERES footprint (20 km at nadir) biases results of clear sky direct radiative effects to situations dominated by large high pressure systems. Loeb and Manalo-Smith (2005) reduce this cloud-free sky bias by basing their estimate on the finer resolution MODIS observations. They first determine the relationship between MODIS narrowband radiances and CERES broadband ones, and use the relationship to make a narrowband to broadband conversion.

In this study we present an alternative method using MODIS data alone to estimate direct aerosol radiative effect over the oceans. Unlike the CERES studies, above, we use an offline radiative transfer model (Chou et al., 1992) to make the conversion between MODIS-measured narrowband angular radiances and broadband hemispheric fluxes in one step. In this way we avoid the empirical model that translates CERES angular measurements to hemispheric flux. Unlike the other studies that use models we do not have to go looking for outside sources for information to use as input to the model. The MODIS aerosol retrieval provides a model of aerosol optical properties that match the spectral radiance at the top of atmosphere to within 3%. A similar method maintaining consistency between retrieval and flux calculations was done using POLDER data (Boucher and Tanré, 2000).

Radiance is a better predictor of reflected flux at top of the atmosphere than any single retrieved parameter (ie. aerosol optical thickness). In Fig. 1 we plot the results from the MODIS aerosol LookUp tables. These include both top of atmosphere spectral radiances and fluxes calculated using the full radiative transfer code of Ahmad and Fraser (1982) for a variety of geometries, aerosol optical thicknesses (τ_a) and aerosol optical models. In the first panel we show flux as a function of aerosol optical thickness. We can predict flux from τ_a, but there is scatter due to uncertainties in the other aerosol optical properties. In the second plot we show flux as a function of radiance for several specific geometries. For any individual observation, the uncertainty in predicting flux from radiance is much smaller. Using the retrieved parameters as a consistent set is closer to the original radiance, and thus a better predictor of the flux. However, other uncertainties affect our results that do not appear in the simulated atmospheres used to produce Fig. 1. Some of these other uncertainties can be quantified, such as assumptions of ocean surface albedo. These will be addressed quantitatively in Sect. 6 below. Other assumptions such as a bimodal aerosol model, particle sphericity, or unexpected chemistry affecting the UV cannot be easily quantified at this time, but these effects are expected to be small.

3 The MODIS aerosol retrieval over ocean

The MODIS satellite sensor has been observing and reporting on aerosol characteristics since the beginning of the Terra satellite mission in 2000 (Ichoku et al., 2002; Chu et al., 2002; Remer et al. 2002). MODIS measures radiance ($\mathrm{W\,m^{-2}\,sr^{-1}}$), denoted as L, in 36 channels. Reflectance is calculated from these measurements according to the definition $\rho=\pi\,\mathrm{L}/(\mu_o \mathrm{E}_o)$ where μ_o is the cosine of the solar zenith angle and E_o is the extraterrestrial solar flux ($\mathrm{W\,m^{-2}}$) in the given spectral band. Of the 36 MODIS channels 6 channels (0.55–2.13 μm) are directly used to retrieve aerosol information from scenes over ocean (Tanré et al., 1997; Remer et al. 2005). While MODIS spatial resolution ranges from 250 m to 1000 m depending on wavelength, the 6 channels used in the aerosol algorithm are all at resolution of 250 or 500 m. The 250 m bands are degraded to 500 m, and thus the basic resolution of the MODIS aerosol retrieval input is uniformly 500 m. This broad spectral range, coupled with the 500 m spatial resolution in these bands, permits a unique view of aerosols that cannot be duplicated with any other sensor. Because of the fine spatial resolution and specialized cloud mask (Martins et al., 2002; Gao et al., 2002; Brennan et al., 2005), MODIS retrieves aerosol properties closer to clouds than other satellites such as AVHRR with its 1 km resolution or especially CERES with its 20 km footprint. On the other hand, close proximity to clouds may introduce cloud contamination into the aerosol optical thickness retrieval. Recent studies estimate the proportion of the retrieved aerosol optical thickness attributed to cloud effects including side-scattered light and cloud shadows (Kaufman et al., 2005b; Zhang et al., 2005c; Coakley et al., 2005). Kaufman et al. (2005b) concluded that undetected cirrus represents10% of the τ over the oceans. Comparison to AERONET as a function of cloud cover indicates additional uncertainty of 5% in the τ due to clouds.

The MODIS aerosol retrieval makes use of a LookUp Table (LUT) consisting of calculated upwelling radiances (or when normalized as above, solar reflectances) at top of atmosphere for each of the six wavelengths for a rough ocean surface, a variety of geometries, aerosol amounts and aerosol models (Remer et al., 2005). There are 9 aerosol models in the LUT. Four of the models represent submicron (fine) mode aerosol particles, and five of the models represent supermicron (coarse) mode particles. Each of the nine models consists of a monomodal lognormal size distribution, and real and imaginary refractive indices. Thus, a unique spectral dependence of extinction, single scattering albedo (ω_o) and assymetry parameter (g) is defined for each model.

In the retrieval process, the algorithm is looking for a combination of fine and coarse mode models to accurately represent the spectral reflectances measured by MODIS at the top of atmosphere. The modes from the LUT are combined using η as the weighting parameter,

$$\rho_\lambda^{\mathrm{LUT}}(\tau_a) = \eta\rho_\lambda^f(\tau_a) + [1-\eta]\rho_\lambda^c(\tau_a) \tag{1}$$

The inversion finds the pair of fine and coarse modes and the τ_a and η that minimizes the error (ε) defined as

$$\varepsilon = \sqrt{\frac{\sum_{\lambda=1}^{6} N_\lambda \left(\frac{\rho_\lambda^m - \rho_\lambda^{\mathrm{LUT}}}{\rho_\lambda^m + 0.01}\right)^2}{\sum_{\lambda=1}^{6} N_\lambda}} \tag{2}$$

where N_λ is the number of pixels at wavelength λ, ρ_λ^m is the measured MODIS reflectance at the wavelength λ and $\rho_\lambda^{\mathrm{LUT}}$ is calculated from the combination of modes in the Look Up Table, defined by Eq. (1). The 0.01 prevents a division by zero for the longer wavelengths under clean conditions. Typically solutions are found with $\varepsilon<3\%$ (Remer et al., 2005).

The solution represents the best fit of the LUT reflectances to the actual reflectances that MODIS measures. The combination of the two chosen modes, τ_a and η represent a derived aerosol model from which a variety of parameters including ω_o and g can be inferred. The combination of τ_a, ω_o and g represent the aerosol optical properties that best fit the spectral reflectances at top of atmosphere given the assumptions embedded in the LUT calculations such as bulk ocean reflectivity and the ozone, water vapor and aerosol profiles. We refer to this hereafter as an internally consistent set of aerosol optical parameters. This is not saying that the MODIS algorithm is retrieving ω_o or g with any accuracy. There could be and are compensating errors associated with the retrieval of any one of the parameters. For this reason

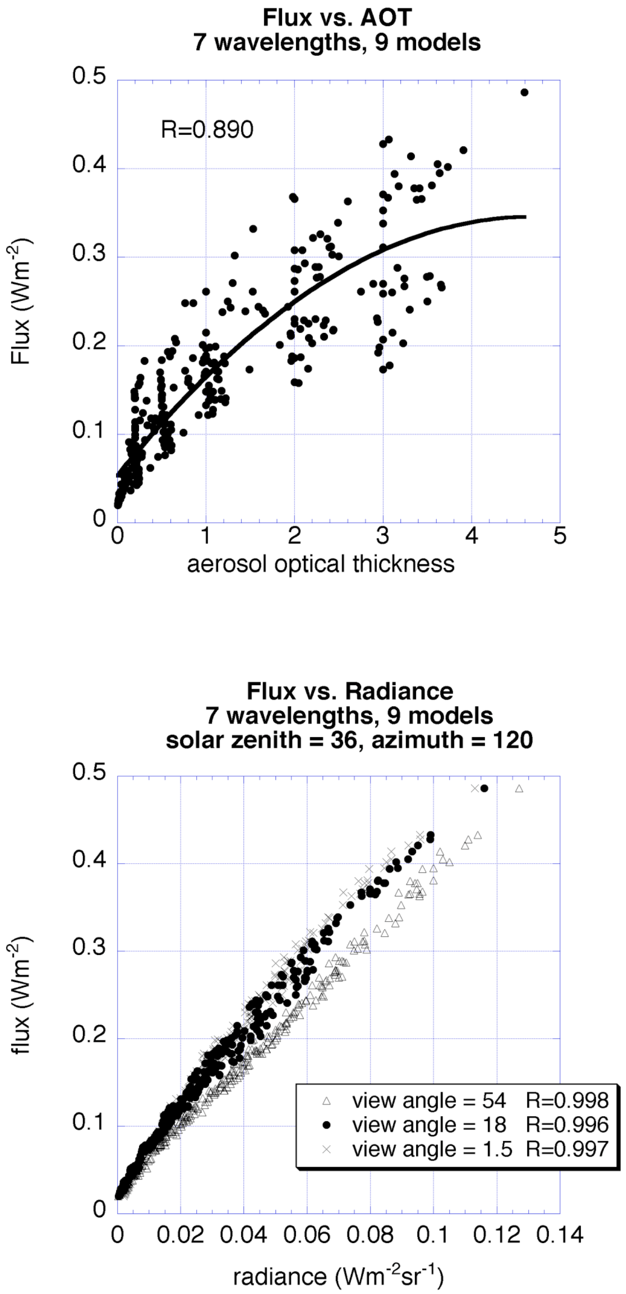
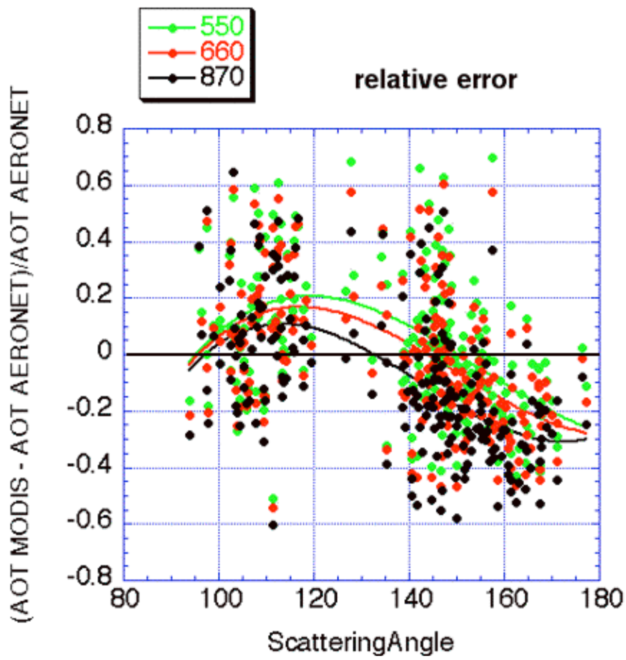


Fig. 1. Top of atmosphere reflected flux from the MODIS Look Up Tables, plotted as a function of aerosol optical thickness (top) for all 9 models and 7 wavelengths, and as a function of top of atmosphere radiance (bottom) for the same mix of models and wavelengths, and 3 selected geometries.



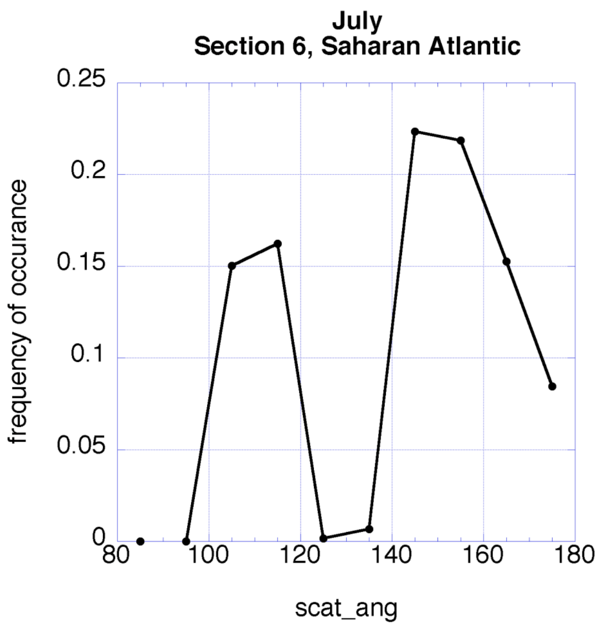


Fig. 2. (Top) Difference between MODIS aerosol optical thickness retrieval at three wavelengths and corresponding AERONET measurements for situations identified as dominated by Saharan dust, plotted as a function of scattering angle. (Bottom) Frequency histogram of scattering angle of MODIS measurements in Section 6 during July. Section 6, the tropical north Atlantic is a region heavily influenced by transported Saharan dust.

we do not make an attempt to estimate radiative effects at the surface, which are particularly sensitive to the value of ω_o. However, the combination of MODIS retrieved τ_a, ω_o and g, when used consistently has to produce the best fit to the spectral reflectances at top of atmosphere.

4 Estimating aerosol radiative effect at top of atmosphere

4.1 The MODIS aerosol data

We will use the results of the MODIS aerosol retrieval as an internally consistent set of aerosol optical properties: τ_a, ω_o and g, that will be input into a column radiative transfer climate model (Chou et al., 1992; Chou and Suarez, 1999) to calculate the upwelling hemispheric broadband fluxes at the top of atmosphere. The MODIS data we use are the Level 3 monthly mean aerosol optical thickness by model, reported at 0.55 μm on a 1 degree grid over oceans (King et al., 2003). This product gives us the monthly statistics based on the original 500 m resolution data. The data from the Terra satellite form a time series from September 2001 to October 2002, and additionally from June 2003 to October 2004. The 7 months of data in 2002–2003 are missing due to a reprocessing of the data occurring during the time of this analysis. The data from the Aqua satellite form a continuous time series from October 2002 to November 2004.

Because we are not constructing fluxes from an angular dependence model (ADM) like CERES does, we can estimate flux from a single geometry. However, because the retrieval is not perfect there could be systematic biases that are correlated to scattering angle. For example, in dust regimes (Fig. 2a), above 140 degrees the optical depth retrieval is biased low, while at lower scattering angles it is biased high. Over the course of a month, MODIS views the same 1 degree square with a wide variety of angles (Fig. 2b). If we divide the error of Fig. 2a at 660 nm into the same scattering angle bins of Fig. 2b, the average magnitude of the error in any bin can reach 0.25 for some scattering angles. However, weighting the error by the frequency of the observations in the month and summing over all scattering angles, the magnitude of the monthly mean error in this case is less than 0.02. This is a particularly spectacular example of the reduction of error due to monthly averaging. In general, by following a similar method of analysis in other cases we expect a reduction of error by approximately a factor of 3.

The MODIS-derived aerosol optical thickness product has been compared extensively with AERONET observations (Holben et al., 1998). Comparisons are made both in terms of individual observations collocated in space and time (Ichoku et al., 2005; Remer et al., 2005) and also comparisons of independently derived monthly mean values (Remer et al., 2005; Kleidman et al., 2005). These evaluations suggest that the MODIS aerosol optical thickness retrieval over oceans agrees with AERONET to within $\pm 0.03 \pm 0.05\tau_a$. Even where the scatter from individual retrievals exceeds expectations, the scatter is random, suggesting that long-term statistics may be even more accurate (Remer et al., 2005).

When MODIS data are collocated in time with AERONET data, MODIS benefits partially from AERONET's more aggressive cloud clearing algorithm. Thus, uncertainty may be larger and biases may exist in MODIS retrievals of aerosol optical thickness that have not been previously reported in the validation studies. For example, MODIS may incorrectly make an observation and report an optical thickness for a scene with cloud contamination. AERONET would not make an observation in those conditions. Therefore, that contaminated MODIS retrieval would never make it to the validation scatter plots because there would be no corresponding AERONET point. Because of these missing points, the reported uncertainty of $\pm 0.03 \pm 0.05\tau_a$ may be overly optimistic, and MODIS retrievals could be biased high at all levels and scales. Recently this potential problem has been addressed and quantitatively estimated. We know that the cloud fraction in the validation data sets used to collocate MODIS and AERONET is 50% lower than the global cloud fraction. Thus, the probability of cloud contamination in the MODIS retrievals of the validation data set is lower than in the overall global data set. Also, recent analysis of MODIS-derived thin cirrus reflectances and aerosol optical thickness retrievals suggests that roughly 0.01–0.02 of the MODIS aerosol optical thickness at 0.55 μm may be attributed to thin cirrus contamination and not aerosol at all (Kaufman et al., 2005b).

4.2 The radiative transfer model

We use the radiative transfer model CLIRAD-SW (Chou et al., 1992; Chou and Suarez, 1999) to calculate the hemispherical flux at the top of the atmosphere. CLIRAD-SW includes the absorption and/or scattering due to water vapor, various gases, aerosols clouds and the surface. Fluxes are integrated over the full solar spectrum, from 0.175 μm to 10 μm. The reflection and transmission of clouds and aerosol layers are calculated from the δ-Eddington approximation and the fluxes calculated using the two-stream adding approximation. Note that we use the model only in cloud free conditions.

CLIRAD-SW requires input of aerosol optical properties in 11 spectral bands, 7 in the ultraviolet, 1 in the 0.40–0.70 μm visible range, 1 in the near-infrared (0.70–1.22 μm), and 2 in the mid-infrared (1.22–10.0 μm). MODIS reports aerosol optical properties in 7 bands (0.47–2.13 μm), none in the ultraviolet. We translate the MODIS values to the wavelengths needed by the model by finding the wavelength of the solar-weighted MODIS extinction in each of CLIRAD-SW's bands,

$$\overline{\beta \mathrm{ex}}(\bar{\lambda}, \mathrm{mode}) = \frac{\int_{\lambda 1}^{\lambda 2} S(\lambda)\beta \mathrm{ex}(\lambda, \mathrm{mode})d\lambda}{\int_{\lambda 1}^{\lambda 2} S(\lambda)d\lambda} \tag{3}$$

with S(λ) the solar spectrum (Neckel and Labs, 1981), βex(λ, mode) the spectral extinction for each of the MODIS

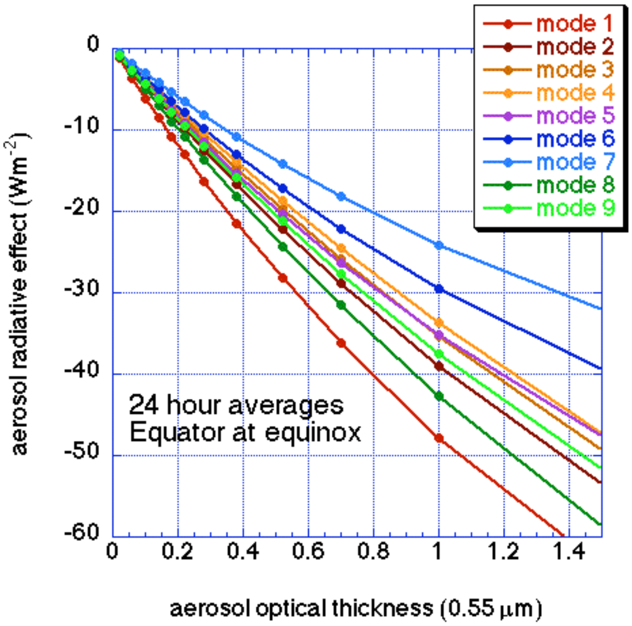


Fig. 3. Daily averaged aerosol radiative effect for a 12 h day with the solar zenith angle equal to 0 at noon, a variety of aerosol optical thicknesses and the nine modes of the MODIS aerosol retrieval over ocean.

modes, and $\overline{\beta \mathrm{ex}}(\bar{\lambda}$, mode) the weighted value used for the CLIRAD-SW input for the band defined between λ1 and λ2. The representative wavelength is $\bar{\lambda}$, and the MODIS optical properties (τ_a, ω_o and g) are interpolated or extrapolated to this value for each of the nine MODIS modes and each CLIRAD-SW band.

The interpolation/extrapolation of MODIS values to CLIRAD-SW bands introduces uncertainty in the final derivation of radiative effect. However, Ichoku et al. (2003) discuss that the final results of radiative effect calculations, especially at top of the atmosphere are mostly insensitive to the extrapolation to the UV or mid-IR bands. The main sensitivity of translating input from the MODIS observations to the CLIRAD-SW bands is to the interpolation in the only visible band, λ=0.40 μm to λ=0.70 μm, corresponding closely to the MODIS primary channel (0.555 μm), and making the interpolation more certain. The uncertainty in the final results from many sources of error is fully discussed in Sect. 6.

We use the midlatitude profiles for temperature and humidity for all model runs. The sensitivity tests in Ichoku et al. (2003) show that the results at top of atmosphere are insensitive to choice of atmospheric profile. Sensitivity to total column amounts of water vapor and ozone are described in Sect. 6.

In all model runs we set sea surface albedo to a constant value of 0.07. Sea surface albedo is a function of the ocean condition (foam, chlorophyll, sediments) and also a strong function of solar zenith angle. Jin et al. (2002) use modeling

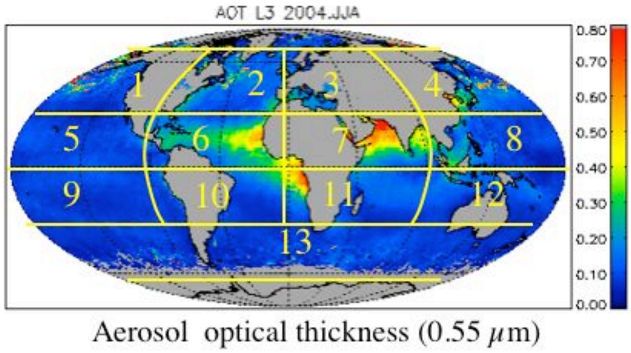


Fig. 4. Terra-MODIS observed seasonal mean aerosol optical thickness over oceans at 0.55 μm for the months June–July–August 2004. The 13 regional sections are also identified.

supported by observations to show that variability in ocean condition contributes to variability in sea surface albedo of 0.01 or less. However, the sea surface albedo can range from 0.09 to 0.04, over the solar zenith angles encountered in our data set. Our constant value of 0.07 corresponds to a solar zenith angle of approximately 55° (Jin et al., 2002), which turns out to be 5° higher than the global mean value of our data set. A 5° difference in mean solar zenith angle results in a 0.012 too high estimate of ocean surface albedo. To determine a correction factor for this offset, we run the model for one fine mode (model 3) and one coarse mode (model 7) with a solar zenith angle of 50 degrees and a constant atmosphere, but change the sea surface albedo from 0.07 to 0.058. We weight the results of the two modes by the global fraction of fine and coarse modes in our data set (50% fine and 50% coarse). The resulting uncertainties are a function of aerosol optical thickness. Therefore, we calculate the global mean uncertainty by weighting by the global mean frequency histogram of aerosol optical thickness, and adjust this instantaneous value to represent the 24 h average using the procedure that will be described in Sect. 4.4. We find that a sea surface albedo that is 0.012 too high will produce an approximately 0.4 Wm^{-2} too low estimate of aerosol effect. The final global mean results reported in this paper will automatically include an adjustment to better match the sea surface albedo of our data set. No corrections are performed on regional or monthly results. Thus, the uncertainty associated with a range of sea surface albedos of 0.04 to 0.09 results in an uncertainty in regional values of approximately 1 Wm^{-2}.

We run CLIRAD-SW separately for each of the 9 sets of aerosol optical properties corresponding to the 9 MODIS modes, for a range of aerosol optical thickness values and for 9 solar zenith angles. From the model output we subtract the net radiative flux at top of the atmosphere for no aerosol optical thickness (τ_a=0) from the values calculated at each of the other values of aerosol optical thickness. This becomes a Look Up Table (LUT) of aerosol effect at the top of the atmosphere. An example of such results are displayed in Fig. 3 averaged over the 24 h period for a location at the equator at the equinox so that we are simulating a 12 h day with the solar zenith angle equal to 0 at noon. We see that for a specific τ_a, even for a moderate value such as 0.20, the effect at top of the atmosphere can vary by approximately $-5\,\mathrm{W\,m}^{-2}$, depending on the type of aerosol present.

4.3 The distribution of aerosol type

The MODIS Level 3 monthly mean statistics include the product, Optical_Depth_By_Models_Ocean, that provides the optical depth at wavelength 0.55 μm attributed to each of the 9 modes in the MODIS algorithm. This product provides the basis for determining the distribution of aerosol properties over the world's oceans. As an illustration we divide the global oceans into 13 sections defined in Fig. 4, and calculate the mean optical thickness attributed to each of the MODIS modes for every month. Examples of the distribution of τ_a among the different modes observed from the Terra satellite for three such sections and one section from the Aqua satellite are shown in Fig. 5.

Section 9 is the cleanest of the 13 sections in terms of aerosol loading with an annual average τ_a=0.09. In this southern tropical Pacific section the primary mode chosen by MODIS is mode=7, and to a lesser extent mode=6, both corresponding to coarse marine sea salt aerosols. Fine modes 1 and 4 also make a contribution, especially in the non-summer months. The fine mode may represent dimethyl sulfide (DMS). There is almost no contribution from fine modes 2 and 3, or coarse modes 5, 8 and 9. This is how Terra-MODIS interprets the background marine aerosol, and Aqua-MODIS (not shown) is similar but with less coarse mode 6, slightly more in modes 1 and 9.

Section 6, off the coast of West Africa contains both transported Saharan dust and biomass burning smoke with an annual average τ_a=0.20. In contrast to Section 9, we see that in Terra Section 6 modes 8 and 9 make a contribution to the total aerosol optical thickness. These two modes correspond to mineral dust. In addition, mode 4 is much stronger than in the purely background aerosol of Section 9. The broad size distribution of mineral dust includes long tails into the submicron region that the MODIS retrieval interprets as optical thickness in the largest fine mode. The winter months tend to have a different distribution of modes than the rest of the year, possibly due to a greater contribution by biomass burning aerosol during that season. The Aqua Section 6 distribution (not shown) is similar to Terra, but with less contribution by mode 6, and more in the dust modes 8 and 9.

Section 4 is the region down stream from north and central Asia with an annual mean τ_a=0.20. In Terra-MODIS we see a broad distribution of aerosol modes, with the summer months exhibiting large increases in fine modes 2 and 3. MODIS interprets smoke and pollution particles mostly as an increase in modes 2 and 3. Although dust is prevalent in this region in the Spring months only a slight elevation in

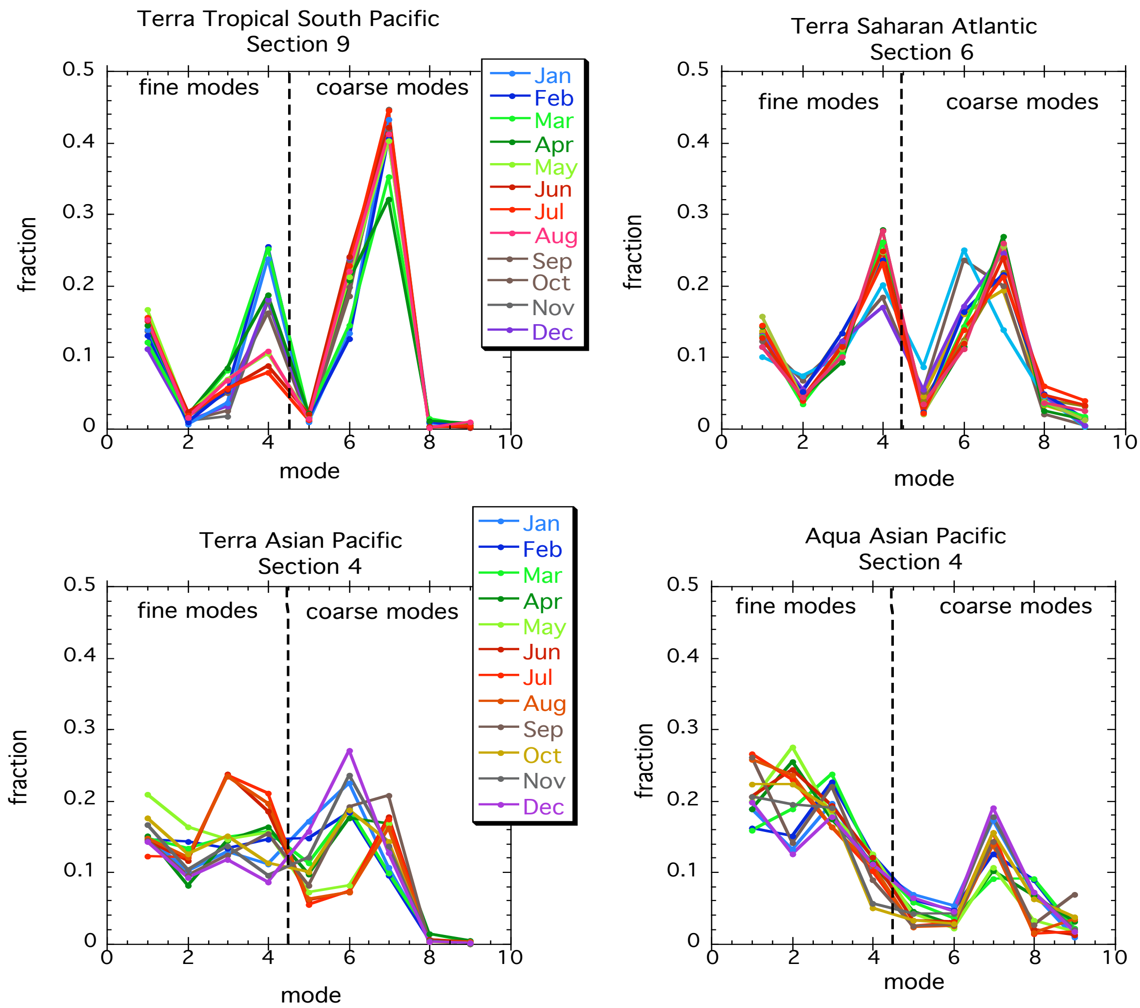


Fig. 5. Fraction of aerosol optical thickness attributed to each of the 9 MODIS modes for four example sections of Fig. 4 as functions of month. Months are composites of all available years of data. Three of the panels show distribution of mode optical thickness observed from the Terra satellite and the last panel (bottom right) shows observations from the Aqua satellite.

mode 8 is noted. The Aqua-MODIS representation in this section is quite different, showing very little optical thickness due to mode 6, much more optical thickness in the dust modes of 8 and 9, and very different distributions amongst the fine modes. Annual mean fine mode fraction from Terra for Section 4 is 0.60, while for Aqua it is 0.70. Note that unlike annual mean values of fine mode fraction published in other studies these mean values were not weighted by τ_a and are used only to compare Terra and Aqua here. Differences between Terra and Aqua arise from a combination of basic calibration differences in the two instruments and also small changes to the MODIS aerosol retrieval algorithms that may be implemented at different times in the separate processing for Terra and Aqua. The MODIS retrieval of aerosol size and choice of aerosol model are especially sensitive to instrument calibration (Chu et al., 2005).

The examples in Fig. 5 demonstrate two points. The first is that the global distribution of aerosol optical properties is more complex than simply the distribution of aerosol optical thickness, or even the distribution of fine mode fraction. The second point is that differences between Terra and Aqua demonstrate the sensitivity of the retrieval algorithm to small perturbations in instrument calibration and software.

4.4 Deriving regional and global daily average aerosol radiative effect

To calculate the aerosol radiative effect we combine the distribution of aerosol modes from the MODIS retrieval (Fig. 5) with the calculated radiative effect as a function of mode (Fig. 3). The MODIS-measured aerosol optical thickness in each mode, τ_a (mode, lat, lon) and the solar zenith angle are used as indices in the radiative effect look-up table, F[τ_a

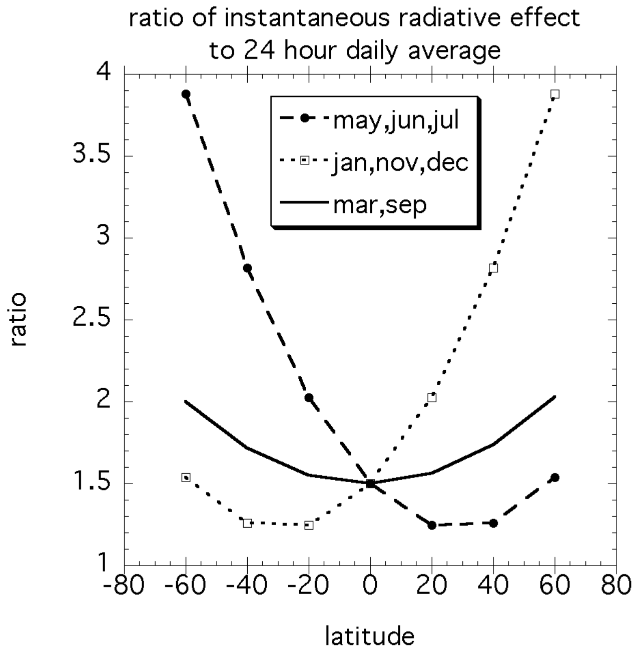


Fig. 6. Ratio of instantaneous radiative effect ($F^{calc}I$) to 24 h daily average radiative effect ($F^{calc}24$) as a function of latitude and month. Shown are selected months. The same ratio applies for Terra and Aqua.

(mode, lat, lon),θ_o]. Then we sum the results over all nine modes.

$$F(\text{lat, lon}) = \sum_{\text{mode}=1}^{9} F[\tau_a(\text{mode, lat, lon}), \theta_0] \tag{4}$$

This is the monthly mean aerosol effect at top of atmosphere for a particular 1 degree grid square, instantaneously at the time of satellite overpass.

We estimate the 24 h daily average radiative effect from the instantaneous values calculated from the MODIS observations. To do so, we return to the CLIRAD-SW model and simulate the diurnal cycle in hourly increments of the aerosol effect for 7 latitudes and 12 months, assuming that the aerosol AOT and properties do not vary systematically through the day. We combine the results of the nine MODIS modes based on the annual mean global aerosol optical thickness and distribution over the nine modes. From this modeling effort we are able to calculate the daily average and the ratio of the instantaneous at the time of satellite overpass to the daily average. The Terra overpass is considered to be 10:30 a.m., and the Aqua over pass 1:30 p.m. An example of these ratios is shown in Fig. 6. Thus for any particular month,

$$F24(\text{lat, lon}) = F(\text{lat, lon}) \frac{F^{calc}24(\text{lat, month})}{F^{calc}I(\text{lat, month})} \tag{5}$$

with F24(lat,lon) the 24 h daily average radiative effect for the grid square based on the MODIS observations, F(lat,lon) the MODIS-derived instantaneous radiative effect from Eq. (4), $F^{calc}24$(lat,month) the model-derived daily average for month and latitude and $F^{calc}I$(lat,month) the model-derived value at the instantaneous time of overpass.

The ratios of $F^{calc}24$(lat,month)/$F^{calc}I$(lat,month) are dependent on aerosol optical thickness and type. On a global mean basis there is a 2% uncertainty in F24(lat,lon) introduced by the ratios due to uncertainty in aerosol type, based on the uncertainty in fine mode fraction of ± 0.25. There is an additional 3% uncertainty introduced by uncertainties in the global mean aerosol optical thickness. Individual regions and months will have larger uncertainty. Because of the symmetry around solar noon of the Terra and Aqua over pass times, the ratios are the same for both satellites.

The Level 3 monthly mean MODIS data that we use will report a monthly mean value in any grid square that has at least one retrieval in that square during the month. Because the basic resolution of the MODIS aerosol retrieval is 10 km, a grid square may have as many as 3000 retrievals in a 30 day month. Clouds, glint, geometry and orbital considerations reduce that number considerably. However, there does remain a significant difference between a grid square with just one 10 km retrieval in the entire month and another square with several hundred retrievals. This difference would be minimal had we used daily data instead of monthly. In order to reconstruct the statistics realized from daily data as we calculate regional and global means, we simply weight each monthly value by the number of MODIS aerosol observations for that month and grid square, Nobs(lat,lon). We also weight by cosine of the latitude to account for the decreasing surface area and corresponding decreasing contribution to the total global or regional radiative effect toward the poles.

$$F24(\text{sect}) = \sum_{\text{lat}} \sum_{\text{lon}} F24(\text{lat, lon}) \text{Nobs}(\text{lat, lon}) \cos(\text{lat}) \tag{6}$$

$$F24\_\text{global} = \sum_{\text{lat}} \sum_{\text{lon}} F24(\text{lat, lon}) \text{Nobs}(\text{lat, lon}) \cos(\text{lat}) \tag{7}$$

where F24(sect) is the daily mean radiative effect at top of atmosphere for one of the 13 sections defined in Fig. 4 and F24_global is the global value. F24(sect) and F24_global are calculated for every month of available data.

5 Results

Figure 7 shows the 24 h MODIS-derived aerosol radiative effect from the Terra satellite at top of the atmosphere for four seasons, and Fig. 8 gives the numerical values for both the aerosol optical thickness and the radiative effect. The locations noted for high aerosol loading unsurprisingly also show prominent radiative effect from these aerosols. Such locations as the Atlantic coast of Africa (Swap et al., 2003; Tanré et al., 2003), the coasts of Asia (Huebert et al., 2003) and the northern midlatitudes in spring (Chin et al., 2004) all report

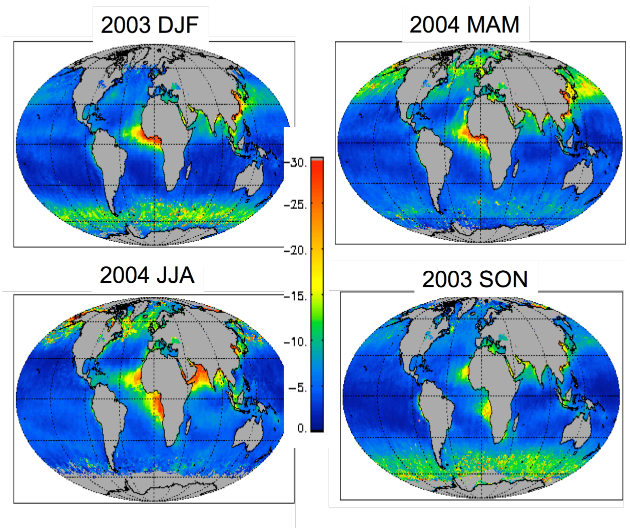


Fig. 7. Global distribution of MODIS-observed aerosol radiative effect at top of atmosphere from the Terra satellite for four seasons: Northern Winter 2003–2004 (upper left), Spring 2004 (upper right), Summer 2004 (lower left) and Fall 2003 (lower right). Units are in Wm^{-2}.

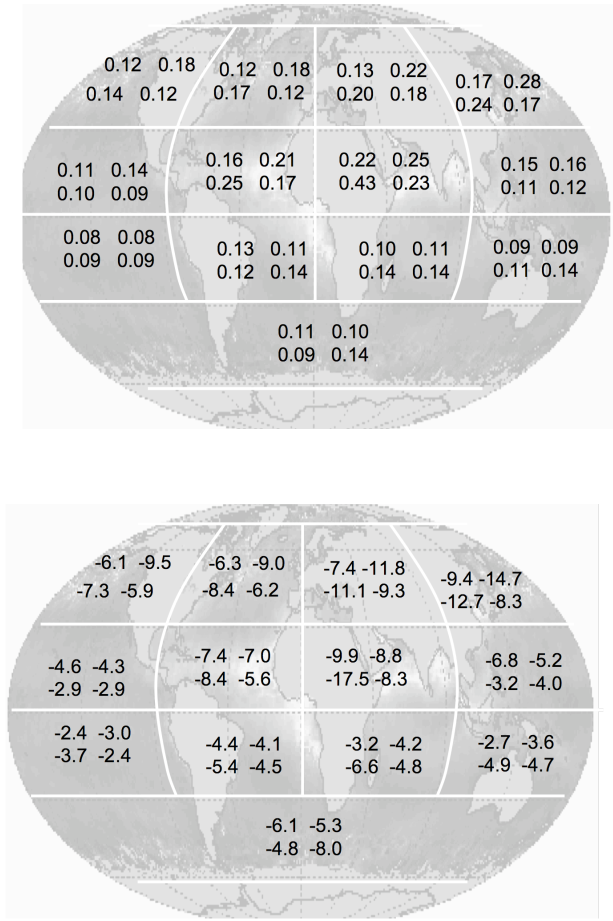


Fig. 8. Seasonal values of aerosol optical thickness (top) and aerosol radiative effect at the top of the atmosphere (bottom) from the Terra satellite. The four numbers in each latitude-longitude section represents a seasonal mean for that section from all available monthly data. Starting from the upper left corner and reading from left to right, the seasons are Northern Winter, Spring, Summer and Fall, respectively. Radiative effect values of the bottom panel are fluxes in units of $\mathrm{W\,m^{-2}}$.

radiative effect in excess of $-15\,\mathrm{W\,m^{-2}}$. More surprising is the band of strong effect that occurs in the southern midlatitudes during Northern Fall and Winter.

Figure 9 shows time series of Terra-MODIS monthly mean aerosol optical thickness, τ_a, for each section and also the global value for both Terra and Aqua satellites. These τ_a are weighted by the number of retrievals in each grid box, analogous to Eqs. (6) and (7) for F(lat,lon). These weighted τ_a are biased low when compared to unweighted values, but better represent the clear-sky direct radiative effect, which is the subject of the present study. Annual mean values of the weighted τ_a over the global oceans for Terra-MODIS is 0.13, the unweighted value is ∼0.14. For Aqua-MODIS the weighted and unweighted values are 0.12 and 0.13, respectively. The time series plots show a great amount of variation in optical thickness among sections, hemispheres and seasons. However, the global mean value remains remarkably constant. The sections of highest aerosol optical thickness include the Asian outflow (section 4), the Saharan outflow (section 6) and the Arabian Sea (section 7). Note that the cleanest region is the south tropical Pacific, but that the midlatitude southern ocean also has relatively little aerosol loading, despite the strong radiative effect seen in Fig. 7.

The center row of Fig. 9 shows a time series of monthly mean aerosol radiative effect from Terra-MODIS for each section, F24(sect), and also F24_global for both Terra and Aqua. The same regional and seasonal variations are seen in the radiative effect as in the optical thickness. The bottom row of Fig. 9 shows a time series for radiative efficiency in units of $\mathrm{W\,m^{-2}}$ per unit τ_a, again from Terra-MODIS. Radiative efficiency is defined as the slope of the linear regression equation calculated from the relationship of F24 and τ_a. In this work it is not a simple ratio of F24/τ_a. There is much more variability in the radiative efficiency than in either τ_a or F24, not only regionally, but globally as well. The higher the latitude the larger the solar zenith angle and the greater the radiative efficiency. Section 13, the midlatitude southern ocean, has a strong radiative efficiency, explaining the apparent contradiction between low aerosol optical thickness and relatively high F24.

Table 1 gives the annual mean global values of τ_a, F24_global and the radiative efficiency for 5 complete calendar years, 2 from Terra and 3 from Aqua. Note that these values include the automatic adjustment to match the global mean sea surface albedo for our data set ($0.4\,\mathrm{Wm^{-2}}$).

Fig. 9. Time series of monthly mean aerosol optical thickness (top row), radiative effect (center row), and radiative efficiency (bottom row) from Terra-MODIS for each of the 13 sections defined in Fig. 4 (dots). Also shown are the global mean values from both Terra (black line) and Aqua (blue line). The left panels show the northern midlatitudes, the center panels the northern tropics and the right panels the southern hemisphere. Terra is missing 7 months of data (2002–2003) due to data unavailability during reprocessing.

The global mean value of F24 for Terra is approximately $-6.0\pm0.7\,\mathrm{W\,m^{-2}}$ and $-6.3\pm0.7\,\mathrm{W\,m^{-2}}$ for Aqua. The global mean value of aerosol efficiency is approximately $-46\,\mathrm{W\,m^{-2}}\tau_a^{-1}$ for Terra and $-51\,\mathrm{W\,m^{-2}}\tau_a^{-1}$ for Aqua.

The year to year variation of either platform is remarkably small. However, even though the two platforms agree to within the given error bars, Aqua does report higher values. This is not due to a global diurnal variation of observed τ_a, because Aqua 's value of τ_a is actually smaller

Table 1. Annual global mean aerosol optical thickness (τ_a), radiative effect at top of atmosphere (F24_global) and radiative efficiency (F24/τ_a) observed from Terra- and Aqua-MODIS during various calendar years.

year	τ_a	F24_global (W m^{-2})	F24/τ_a (W m$^{-2}\tau_a^{-1}$)	F24 corrected for clouds
Terra Sep'01 to Aug'02	0.130	−5.9±0.6	−45.0	−5.0 to −5.2
Terra Sep'03 to Aug'04	0.129	−6.0±0.6	−46.5	−5.1 to −5.3
Aqua Sep'03 to Aug'04	0.122	−6.2±0.6	−50.5	−5.2 to −5.4
Aqua Dec'02 to Nov'03	0.123	−6.3±0.6	−51.4	−5.3 to −5.5
Aqua Dec'03 to Nov'04	0.123	−6.3±0.6	−51.0	−5.3 to −5.5

F24 corrected for clouds is an approximation based on estimates of cloud contamination in the aerosol optical thickness product of 0.015 to 0.020 on a global basis, over the oceans. Discussion in Sect. 6.

than Terra's in this data set. The two platforms do report different distributions of aerosol over the 9 modes (Fig. 5), suggesting either different aerosol types at the two overpass times, or more likely, diurnal differences of cloud contamination in the aerosol retrievals or uncertainties in the two sensors' calibrations or properties that result in retrievals of different aerosol modes. For example, the 1.6 μm channel on Aqua is not functioning well and the aerosol retrieval is sometimes reduced to 5 channels of input. The partitioning of the aerosol optical thickness into different modes will be much more sensitive to subtle changes in instrument calibration and characterization than the derivation of total aerosol optical thickness (Tanré et al., 1997; Chu et al., 2005).

A more detailed comparison between Terra and Aqua is shown in Fig. 10. Here monthly sectional means derived from the two sensors are plotted against each other in scatter plots. Northern and southern hemispheres are plotted separately, with midlatitude separated from tropical sections by symbol. We use different scales on the axes in the two hemispheres. Aqua aerosol optical thickness (τ_a) is systematically lower than Terra's for all sections and seasons, north and south of the equator, both midlatitudes and tropics. However, Aqua's radiative effect (F24) is similar to Terra's in the midlatitudes, while systematically more negative in the tropics. The reason is the stronger efficiency (F24/τ_a) observed by Aqua in all regions and seasons. The stronger efficiency compensates for the lower τ_a in the midlatitudes, but overcompensates in the tropics, causing the Aqua tropical F24 values to be more negative than Terra's. For these matching monthly-sectional mean values, Aqua τ_a are lower than Terra's by 8% in the midlatitudes and 3% in the tropics. The Aqua efficiencies are stronger by 6% in the midlatitudes and 15% in the tropics, while the Aqua radiative effect (F24) is 2% less negative than Terra's in the midlatitudes but 12% more negative in the tropics.

All estimates of radiative effect reported above describe the radiative effect per unit of clear-sky area. This is the quantity commonly reported by other studies (Boucher and Tanré, 2000; Christopher and Zhang, 2002; Loeb and Manalo-Smith, 2005). This quantity only represents the amount of energy reflected to space by the aerosol if the region is completely cloud free. In fact, the regions are not cloud free, and some exhibit annual mean cloud fractions exceeding 0.75. Thus, the true effect that clear-sky aerosols have on the Earth's radiative balance is much less than reported above, or reported in other studies. When we weight the above calculated radiative effect by the MODIS-derived cloud-free fraction the global annual mean effect for the Terra satellite is −2.2 W m^{-2}, less than half of the value assuming 100% cloud free area. In weighting by cloud-free area we cannot separate thin clouds from thicker clouds. Aerosol under a thin cloud also affects the Earth's radiative balance. Thus, the −2.2 Wm^{-2} is an underestimate of the aerosol effect on the planet and the −5 to −6 W m^{-2} from Table 1 is an over estimate, although the latter value is an unambiguous estimate of the radiative effect per unit of clear-sky area.

6 Estimating uncertainty

6.1 Unbiased uncertainty

The uncertainties appearing in Table 1 are based on the following sources of unbiased uncertainty. The first source of error is the calibration uncertainty of the MODIS radiances themselves, ∼2%, which will generate a larger error in the aerosol radiative effect, ∼4%. The second source of error are the initial MODIS retrievals of the sets of parameters, τ_a, ω_o and g, which match the observed spectral radiances to within 3% (Eq. 2), and thus over an ensemble of measurements of various view angles encountered during a month of MODIS observations should also represent the aerosol effect at top of atmosphere to within the same uncertainty.

The third source of error arrives from choosing input parameters for the CLIRAD-SW model. We estimate the uncertainties on the annual global aerosol effect by perturbing our assumed values one at a time and then running the model for a representative fine mode (mode 3) and a representative coarse mode (mode 7). We then combine the uncertainties from the two modes using the global mean fine mode

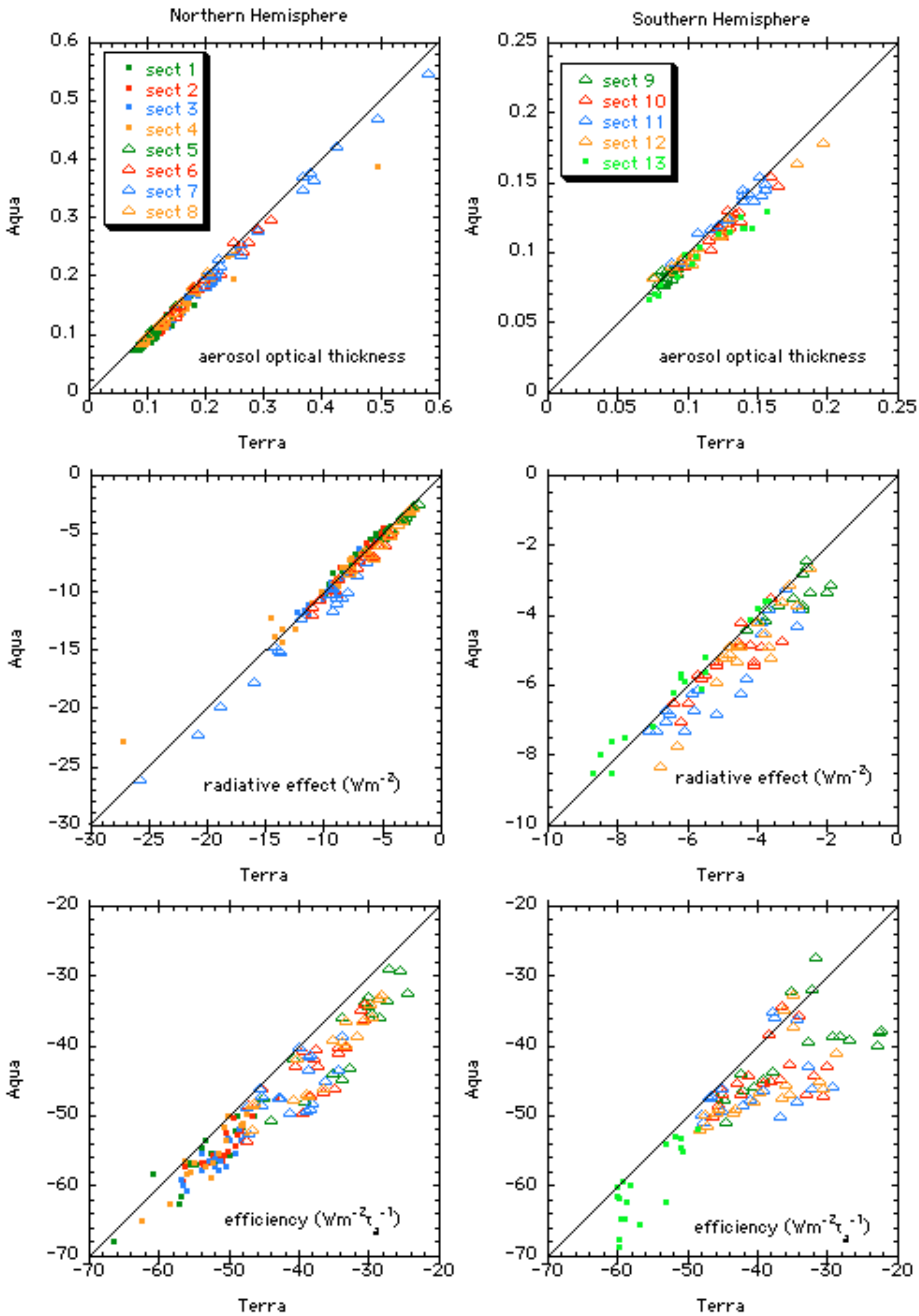


Fig. 10. Scatter plots of quantities derived from Aqua data plotted against those derived from Terra data. Each point is a monthly-sectional mean from each of the 13 sections whenever both satellites reported values. The quantities shown are aerosol optical thickness-τ_a (top), radiative effect – F24 (center) and radiative efficiency – F24/τ_a (bottom). The left column is for the northern hemisphere and the right column shows southern hemisphere results. Midlatitudes in both hemispheres are denoted by dots. Tropical sections in both hemispheres are denoted by open triangles.

Table 2. Five types of unbiased uncertainty originating from (1) the inherent calibration uncertainty of the measured radiances from the MODIS instrument, (2) the ability of the retrieval to match reflectances at TOA with τ, ω_o and g, (3) initializing the radiative transfer model, (4) calculating F24 from the instantaneous satellite observation, and (5) estimating the magnitude of the cloud contamination correction.

Source of error	Parameter	perturbation	% change in aerosol effect
(1) Instrument calibration	MODIS radiances		4
(2) Retrieval	Matching TOA radiances		3
(3) Input parameters for the RT model	Extrapolate SSA to UV	0.035	1
	Extrapolate SSA to MidIR	0.05	1
	Extrapolate AOT to UV	25%	1
	τ confined to layer 870–561 hPa		4
	τ confined to layer surface – 799 hPa		2
	Total column water	25%	2
	Total ozone	25%	1
	Ocean albedo	0.01	7
(4) Calculating 24 h average Flux	Aerosol type	0.25 in fine mode fraction	2
	Aerosol amount	0.015	3
(5) Cloud contamination correction	Uncertainty in estimating magnitude of correction		3
Total unbiased uncertainty			11

fraction, which is roughly 0.5. Some of the resulting uncertainties are a function of aerosol optical thickness. Therefore, we calculate the global mean uncertainty by weighting by the global mean frequency histogram of aerosol optical thickness. The resulting percent change in aerosol effect due to the given perturbation is listed in Table 2. The perturbations represent departures from annual, global mean conditions. Regional and monthly uncertainties are larger. In particular the perturbation in sea surface albedo represents the 0.01 uncertainty due to foam, chlorophyll, sediments etc. (Jin et al., 2002) and not the systematic relationship between sea surface albedo and solar zenith angle that we correct for in the global values of Table 1 and characterize as a 1 Wm^{-2} uncertainty in the regional results.

Another source of error arises from converting instantaneous radiative effect to 24 h daily averaged values. In making the conversion we model the diurnal cycle of radiative effect based on assuming global mean aerosol optical thickness and global mean distribution of aerosol type over the 9 MODIS modes. We determine uncertainty to these assumptions of aerosol properties from sensitivity studies that deviated aerosol type and amount based on the uncertainty of the MODIS aerosol retrievals for global mean fine mode fraction ($\pm$0.25) and aerosol optical thickness ($\pm$0.02). The uncertainty to the conversion due to aerosol type adds a 2% error, while the uncertainty due to aerosol amounts introduces a 3% error. We take these errors originating in the conversion to 24 h averages to be unbiased, although there could be systematic biases if assumptions underlying the original aerosol optical models are not realistic.

The last source of unbiased error arises from uncertainties associated in correcting for cloud contamination. Cloud contamination itself is a biased error, and we discuss the correction below. However, correcting for this offset introduces unbiased uncertainty in the final numbers. We estimate the uncertainty in the correction based on the uncertainty in global estimates of cirrus contamination in the aerosol optical thickness product ($\sim$0.005). The resulting uncertainty in the aerosol radiative effect is approximately 3%. Combining all these sources of uncertainty in a root square error sense results in an overall *unbiased* uncertainty of 11% in the cloud-corrected estimated aerosol radiative effect. Uncertainty is higher for monthly and regional values.

6.2 Residual cloud contamination

The above error analysis assumes all uncertainties are unbiased. Another source of uncertainty concerns the issue of residual cloud contamination in the retrievals, which introduce a biased error into the estimation of aerosol radiative effect. Cloud contamination will always increase aerosol optical thickness and therefore systematically introduce a high bias to our estimates of radiative effect. As discussed above in Sect. 4.1, we estimate the potential increase of optical thickness due to contamination may be as high as 0.015 to 0.020 optical thickness on a global basis (Kaufman et al., 2005b; Zhang et al., 2005c). Clouds will also modify the aerosol retrieval of the other two parameters of the solution set (ω_o and g), creating their own signature in the calculated fluxes and estimates of radiative effect. It is unclear at this point, exactly how to interpret the effect of cloud contamination on the final results. While clouds consist of large particles and cloud contamination will shift aerosol retrievals to the coarse modes, the coarse modes (modes 5 to 9 in Fig. 3) do not have separable efficiencies from the fine modes (modes 1 to 4). We do not know how cloud contamination affects the efficiencies. However, if we assume that the global efficiencies in Table 1 remain the same with only the global mean aerosol optical thickness affected then as an approximation we can calculate a "cloud corrected" F24 by multiplying the Table 1 efficiencies by their respective global values of (τ_a–$\Delta\tau_a$), where $\Delta\tau_a$ is the amount of optical thickness attributed to cloud contamination (0.015 to 0.020). For the first row of Table 1 (τ_a–$\Delta\tau_a$) is 0.11 to 0.115, which when multiplied by $-45\,\mathrm{Wm^{-2}}$ per τ_a gives us a range of corrected F24 to be -5.0 to $-5.2\,\mathrm{W\,m^{-2}}$. Applying the same calculation to the other years and satellites listed in Table 1 suggests that the Terra $-6.0\,\mathrm{Wm^{-2}}$ and the Aqua $-6.3\,\mathrm{Wm^{-2}}$ listed in the table should be taken as an upper bound of the estimate, and a cloud free number may be closer to -5.0 to $-5.5\,\mathrm{W\,m^{-2}}$.

6.3 Precision

Another way of evaluating the usefulness of the method is to estimate the method's precision. We can do this by comparing Terra and Aqua results. Differences between the two platforms may be due to physical differences in the aerosol between the two overpass times, but this is unlikely. Thus, if we assume that the aerosol properties remain constant between overpass times, then the estimated aerosol radiative effect, F24, should be the same. In Fig. 10, we show that the two instruments agree to within 2% in midlatitudes and to within 12% in the tropics. While the reasons for the regional difference are unclear, diurnal differences in cloudiness and cloud contamination of the aerosol optical thickness and chosen modes may contribute. Overall, we find that the method's precision for global estimates is 5%.

6.4 Other sources of uncertainty

While we have attempted to quantify the major sources of uncertainty and the precision of the method, there are other sources of uncertainty having to do with the assumptions in the MODIS retrieval such as particle shape. However, these other parameters are expected to introduce only small additional uncertainty. For example, we know that particle nonsphericity only affects dust aerosol, and then only increases uncertainty in τ by $\sim$7% for monthly mean values. Effects on flux retrievals will be less (Fig. 1), and a global annual mean over all types of aerosol will decrease the uncertainty further.

Likewise, if the true ocean surface properties differed from the assumptions used in the original retrieval a bias will be introduced to the retrieved aerosol characteristics. The retrieval cannot decouple aerosol characteristics from errors in surface reflectance assumptions. The bias inherent in the aerosol retrieval from the surface will be carried through to the calculations of outgoing radiative flux. The difference between the calculated aerosol-laden flux and the calculated clean case will include both the aerosol effect and biases introduced from erroneous surface assumptions. Thus, the values we calculate in this work and attribute solely to the aerosol may contain artifacts originating from our original assumptions of surface reflectance in the MODIS retrieval. This differs from the uncertainties introduced when choosing input to the radiative transfer model for calculations of aerosol flux at top of atmosphere and quantified in Table 2.

If there were a global bias in the aerosol retrievals, then it should show up as a bias when we compare MODIS retrievals to AERONET observations. Such comparisons suggest a negligible bias of 0.005 in optical thickness at 0.55 μm (Remer et al., 2005). There is some concern that the MODIS-AERONET comparisons are limited to island and coastal waters, and may not reveal a general bias over the open ocean. We will explore this possibility.

The MODIS retrieval assumes that the water leaving reflectance at 0.55 μm is 0.005 and at longer wavelengths it is zero. These values were chosen from remote sensing experience that began with AVHRR. Recent analysis of more than 1000 spectra of water leaving reflectance measurements taken from ocean going cruises (Maritorena et al., 2002) shows that 88% of the observations report water leaving reflectance at 0.55 μm within $\pm$0.001 of 0.002, and 75% of reflectances at 0.67 μm are less than 0.0003. It does appear that open ocean values at 0.55 μm are 0.003 less than what are assumed by the MODIS algorithm, but the longer wavelengths, at least at 0.67 μm, are very close to zero, as assumed. Because the MODIS algorithm inverts six wavelengths to retrieve the aerosol characteristics, an over prediction of 0.003 in surface reflectance in one channel does not necessarily result in a 0.03 under prediction of optical thickness in that channel, as would be expected for a single channel inversion. The inconsistency with the 0.55 μm channel's assumed surface reflectance will more likely affect the choice of models, the spectral signature of the optical thickness and the retrieved size parameters.

To estimate the effect on our results we turn to the sensitivity studies of Tanré et al. (1997). These were performed for perturbations 3 times larger than the bias expected from the Maritorena et al. (2002) in situ data. Scaling the Tanré et al. (1997) results to match the observed perturbations results in a bias of 0.08 in retrieved fine mode fraction and 0.006 in optical thickness. From Fig. 3 we see that there is no systematic trend in radiative effect as fine modes progress to coarse modes. A 0.08 bias applied to the average fine mode and average coarse mode at optical thickness near the global mean value (0.13) results in an uncertainty in aerosol effect of $\pm0.07\,\mathrm{Wm^{-2}}$. We can think of situations where choice of models can increase this uncertainty, but also situations where the uncertainty can be less.

We conclude by stating that errors in the original MODIS aerosol retrieval from improper assumptions can contribute to errors in the estimates of radiative effect that are not included in the estimate of uncertainty in Table 2. Simply, there is uncertainty in the estimate of uncertainty. However, because of the overall good agreement between MODIS retrievals and AERONET observations, even with some bias due to the locations of the AERONET stations, the additional uncertainty is well-within the stated bounds of the global estimates.

7 Conclusions

We have estimated the global value of total clear-sky aerosol shortwave radiative effect over the oceans in cloud free conditions to be $-6.0\pm0.7\,\mathrm{Wm^{-2}}$ to $-6.3\pm0.7\,\mathrm{Wm^{-2}}$ using an internally consistent set of aerosol optical parameters. Correcting for estimated cloud contamination, these numbers become $-5.0\pm0.6\,\mathrm{Wm^{-2}}$ to $-5.5\pm0.6\,\mathrm{Wm^{-2}}$. The global values of aerosol optical thickness and radiative effect are remarkably consistent from season to season and year to year.

These values are essentially the same as those found using different satellites and methods. Yu et al. (2005) present a comprehensive review and comparison of different observationally-based estimates of aerosol radiative effects. Studies that use MODIS aerosol optical thickness in conjunction with CERES observations of radiative fluxes to determine the global annual radiative effect over the oceans report an annual value of $-5.3\,\mathrm{Wm^{-2}}$ (Zhang et al., 2005b) and $-5.5\,\mathrm{Wm^{-2}}$ (Loeb and Manalo-Smith, 2005). Using POLDER data consistently in a method similar to the one employed here gives $-5.7\,\mathrm{Wm^{-2}}$. The results also resemble those of Yu et al. (2004) who combine MODIS aerosol optical thickness retrievals with results of a chemical transport model. Their value for annual aerosol radiative effect over the oceans is $-5.1\,\mathrm{Wm^{-2}}$.

Individual regions show greater variability, spatially, seasonally and annually. For the most part, aerosol shortwave radiative effect is directly proportional to aerosol optical thickness, with the regions and seasons experiencing the highest optical thickness also experiencing the greatest radiative effect. However, because of the increased solar zenith angle at higher latitudes, the midlatitude and polar regions have higher radiative efficiency and greater radiative effect for the same optical thickness found in the tropics. There are also differences in radiative efficiency due to different optical properties of aerosol in different regions.

The numbers above represent the aerosol effect per unit of clear-sky area, the quantity typically quoted in previous work. The actual effect on the Earth's radiative balance will be substantially less due to cloudiness and clear-sky fraction less than 1.0. Assuming the aerosol has no effect on the radiative balance for the portion of the globe that MODIS identifies as cloudy, we calculate global clear-sky aerosol effect to be $-2.2\,\mathrm{Wm^{-2}}$ for the Terra satellite. However, this number is an underestimate due to aerosol acting beneath thin clouds. The actual effect on the Earth's radiative balance must fall between the $\sim-5.3\,\mathrm{Wm^{-2}}$ that assumes 100% clear sky and the $-2.2\,\mathrm{Wm^{-2}}$ that underestimates the effect beneath thin clouds.

There is a systematic bias between the results from the Terra and Aqua satellites with Terra showing 5% less effect and 11% weaker radiative efficiency than Aqua, despite its consistently higher values of optical thickness. Most of the differences between Terra and Aqua occur in the tropics. Note that the 5% difference is slightly smaller, not larger and in opposite direction than the difference in the AOT between the two satellites. This is the result of the compensation effects between errors made in the derivation of the AOT and in calculations of the aerosol radiative effect. If the difference between Terra and Aqua is taken as an objective measure of the overall precision in estimating aerosol radiative effects by this method, then the precision of estimating global values is 5%, or $\pm0.27\,\mathrm{W\,m^{-2}}$ for a mean value of $-5.3\,\mathrm{W\,m^{-2}}$. Thus the precision is about half of the estimated uncertainty in the method.

The MODIS analysis of the aerosol effect on the radiative fluxes adds a new measurement perspective to a climate change problem dominated so far by models. In fact the results of this study used in conjunction with estimates of the anthropogenic fraction of the aerosol optical thickness (Kaufman et al., 2005a) show excellent agreement between the MODIS-derived estimates of anthropogenic aerosol radiative forcing and the same quantity calculated by models.

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