Quantification of Monthly Mean Regional-Scale Albedo of Marine Stratiform Clouds in Satellite Observations and GCMs

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ABSTRACT

Planetary albedo—the reflectivity for solar radiation—is of singular importance in determining the amount of solar energy taken in by the Earth-atmosphere system. Modeling albedo, and specifically cloud albedo, correctly is crucial for realistic climate simulations. A method is presented herein by which regional cloud albedo can be quantified from the relation between total albedo and cloud fraction, which in observations is found to be approximately linear on a monthly mean scale. This analysis is based primarily on the combination of cloud fraction data from the Moderate Resolution Imaging Spectroradiometer (MODIS) and albedo data from the Clouds and the Earth’s Radiant Energy System (CERES), but the results presented are also supported by the combination of cloud fraction and proxy albedo data from satelliteborne lidar [Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO)]. These data are measured and derived completely independently from the CERES-MODIS data. Applied to low-level marine stratiform clouds in three regions (off the coasts of South America, Africa, and North America), the analysis reveals regionally uniform monthly mean cloud albedos, indicating that the variation in cloud shortwave radiative properties is small on this scale. A coherent picture of low “effective” cloud albedo emerges, in the range from 0.35 to 0.42, on the basis of data from CERES and MODIS. In its simplicity, the method presented appears to be useful as a diagnostic tool and as a constraint on climate models. To demonstrate this, the same method is applied to cloud fraction and albedo output from several current-generation climate models [from the Coupled Model Intercomparison Project, phase 3 (CMIP3), archive]. Although the multimodel mean cloud albedo estimates agree to within 20% with the satellite-based estimates for the three focus regions, model-based estimates of cloud albedo are found to display much larger variability than do the observations, within individual models as well as between models.

1. Introduction

Albedo plays an important role in Earth’s energy balance. Denoting the fraction of incoming solar radiation that is reflected back to space by Earth and atmosphere, it determines the amount of energy taken in by the system. The albedo is a complex function of the surface and atmospheric state of the Earth system and depends on several factors, such as surface properties, cloud amount, cloud physical properties, and aerosol amount. Albedo also provides the context for direct and indirect aerosol forcing of the climate.

The albedo of a partly cloudy scene can be described as the area-weighted sum of the albedos of clear and cloudy
areas (Cess 1976), presupposing knowledge of the fractional area covered by clouds (cloud fraction) and the albedos of the clear areas and of the clouds themselves. Yet, cloud albedo has proven difficult to estimate and is hence not well known. In situ measurements from aircraft can give instantaneous cloud albedo estimates, but are limited in space, time, and viewing angle. Such estimates vary widely within as well as between experiments, ranging from 0.1 to 0.9, with an average around 0.4–0.5 (see, e.g., Robinson 1958; Griggs 1968; Salomonson and Marlatt 1968; Hayasaka et al. 1994; Peng et al. 2002).

To first order, cloud albedo is what maintains the close relation between cloud fraction and planetary albedo [or reflected shortwave (SW) radiation at the top of the atmosphere (TOA)] that has been demonstrated in various observational records (Norris 2005; Loeb et al. 2007). We study the nature of this relationship in more detail and make use of its properties to test the variability of the cloud albedo within a given region, and to obtain averaged cloud albedos on regional scales, using simultaneous satellite observations of regional cloud fraction and albedo.

Hereby, we present a diagnostic tool that can be used as a test and constraint on climate models, and we illustrate this by comparing the observations with global climate model output from the World Climate Research Programme’s (WCRP) Coupled Model Intercomparison Project, phase 3, (CMIP3) multimodel dataset (Meehl et al. 2007).

Focus is placed on regions of marine stratiform clouds at low altitudes, a category of clouds that is a dominant contributor to the planetary albedo (Slingo 1990; Hartmann et al. 1992) and also remains a major contributor to uncertainties in model-predicted cloud feedback (Bony and Dufresne 2005).

2. Method

Albedo ($\alpha$) has conventionally been expressed as a function of cloud fraction ($f$):

$$\alpha = \alpha_{\text{cloud}} f + \alpha_{\text{clear}} (1 - f),$$  

with $\alpha_{\text{cloud}}$ denoting the albedo where the sky is cloudy and $\alpha_{\text{clear}}$ the albedo where the sky is clear (Cess 1976). We note that this expression is based on the assumption that a distinction between clear and cloudy skies can be made, which is not unproblematic, due to the continuum of optical depths between clear-sky aerosol and cloud (e.g., Charlson et al. 2007).

Equation (1) can be rewritten as

$$\alpha = (\alpha_{\text{cloud}} - \alpha_{\text{clear}}) f + \alpha_{\text{clear}},$$  

which illustrates that if $\alpha_{\text{cloud}}$ and $\alpha_{\text{clear}}$ are both constant on some spatial and temporal scale, then $\alpha$ is a linear function of $f$. A scatterplot of $\alpha$ versus $f$ for that space and time scale should then display a linear relation with a slope determined by the values of $\alpha_{\text{cloud}}$ and $\alpha_{\text{clear}}$, and an intercept determined by $\alpha_{\text{clear}}$. The properties of a best-fit straight line to such data can then be used to estimate $\alpha_{\text{cloud}}$. Simply as the sum of the values of the slope and the intercept of the regression line or, equivalently, as the extrapolated value where the cloud fraction is 100%.

Alternatively, we can take the derivative of Eq. (1) with respect to $f$. If $\alpha_{\text{cloud}}$ and $\alpha_{\text{clear}}$ are both independent of cloud fraction $f$, this yields

$$d\alpha/df = \alpha_{\text{cloud}} - \alpha_{\text{clear}};$$  

in agreement with the linear model of Eq. (2). In other words, the degree of linearity in the relation between cloud fraction and albedo can be considered to be a test of the independence of cloud albedo on cloud fraction.

We use this graphical-empirical method to compare cloud albedo estimates from satellite data and model output. To limit the influence of surface reflectivity, which may alter the total above-cloud albedo through multiple reflections, our application of this method for determining $\alpha_{\text{cloud}}$ is here restricted to ocean areas, where the surface albedo remains low (Jin et al. 2005). We do not make any assumptions about the surface albedo, but rather let the apparent value of the clear-sky (surface and atmosphere) albedo be determined by the intercept of the regression line. (See section 4a.)

To avoid contributions from different cloud types to $\alpha_{\text{cloud}}$ we also restrict the application to regions of homogenous cloud type. (See further discussion in section 4d.)

The regions of persistent marine stratocumulus clouds along the east side of the subtropical ocean basins meet these restricting criteria. They have also been identified as highly important to Earth’s radiative budget and as major contributors to the spread among climate models in future climate projections (Bony and Dufresne 2005). We focus here on three such regions, off the west coasts of South America (Peruvian region), Africa (Namibian region), and North America (Californian region). The choice of regions follows Klein and Hartmann (1993), and the geographical areas covered are $10^\circ$–$20^\circ$S, $80^\circ$–$90^\circ$W; $10^\circ$–$20^\circ$S, $0^\circ$–$10^\circ$E; and $20^\circ$–$30^\circ$N, $120^\circ$–$130^\circ$W, respectively.

Scatterplots of albedo versus cloud fraction, similar to those discussed here, have previously been presented by Webb et al. (2001), using global ocean daily mean data. In that study a linear relation was not evident, for two...
main reasons. First, the use of global data introduces nonlinearities due to the presence of different cloud types of varying reflectivity, which can be avoided by restricting the analysis to regions of expected homogeneous cloud cover. (See also section 4d.) Second, daily data display large variability that disrupts the linearity, even for regions dominated by one type of clouds. However, this variability is found to be reduced over increasingly long averaging times. Although we recognize an expansion of the analysis to different time scales as instructive, a thorough investigation of the averaging time is beyond the scope of the present study. We find that the use of monthly mean data yields far less spread than does the use of daily data and indicates a linearity supporting the application of the regression method as described above, and therefore we focus on the monthly time scale. In this way, rather than an instantaneous cloud albedo estimate, a climatologically significant, “effective” regional cloud albedo estimate can be obtained.

3. Data

a. CERES and MODIS

The primary satellite data sources for the presented analysis are SW TOA flux estimates from Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996; Smith et al. 2004) and cloud property estimates from the Moderate Resolution Imaging Spectroradiometer (MODIS; King et al. 2003; Platnick et al. 2003). We make use of observations by CERES and MODIS from the Terra and Aqua satellites for the period July 2002–August 2007, that is, five full years.

Terra and Aqua are both in sun-synchronous near-polar orbits, but differ in equator passing time and direction of motion. Terra crosses the equator at 1030 local time (descending), and Aqua crosses at 1330 local time (ascending).

TOA radiative fluxes are derived from instantaneous radiance observations that are converted to 24-h-average fluxes by the application of diurnal albedo models, assuming that the scene at the time of the satellite overpass remains constant throughout the day, and averaged onto a 2.5° × 2.5° grid on a daily basis. Because the CERES products used are the Earth Radiation Budget Experiment (ERBE)-like Monthly Geographical Averages (ES4), the algorithms used for conversion to daily fluxes are similar to those used in ERBE processing.

The MODIS cloud fraction estimates used are based on the retrieval of cloud optical properties (optical thickness, particle size, water path), as described in Platnick et al. (2003). These are determined from daytime scenes for a sampled set of retrieval grid points (at 5 km × 5 km resolution), and the cloud fraction is determined as the fraction of successful retrievals that are assigned either liquid water, ice, or undetermined phase in each 1° × 1° grid box. A feature of this cloud fraction estimate of possible relevance to our application is the “clear-sky restoral” algorithm setting pixels identified as cloudy to clear if they touch a pixel identified as clear. This may cause an underestimation of cloud fraction especially in areas of broken or scattered cloudiness.

The data are presented as deseasonalized monthly mean values (i.e., with a mean climatological seasonal cycle removed), to avoid influence from seasonal variability in cloud and radiative properties. The monthly mean values are the mean of the daily means. The MODIS data are linearly interpolated onto a 2.5° × 2.5° grid to match the CERES data.

b. CALIPSO

To complement the CERES-MODIS data, we use cloud fraction and proxy “lidar albedo” (Charlson et al. 2007) from the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO; Winker et al. 2009) for October 2006–September 2008 (i.e., two full years). The CALIPSO satellite is part of the A-Train and flies in formation with other satellites, including Aqua, with an afternoon equator crossing time. Full-resolution (330 m) integrated attenuated lidar backscatter (IABS) data, here binned into 5° × 5° grid boxes, are used as a proxy for albedo data. This proxy has been found to correlate well with albedo determined by near-coincident reflectance observations by CERES, as described by Charlson et al. (2007).

The CALIPSO cloud-aerosol detection algorithm, identifying clouds with a feature finder tool, is described by Vaughan et al. (2009). A daily value of cloud fraction is calculated based on the number of lidar profiles identified as cloudy, divided by the total number of lidar profiles in the grid box, excluding the number of profiles with high cloud. Profiles with cirrus are excluded from the analysis because relationships observed between IABS and low cloud become distorted when the lidar signal has interacted with nonspherical ice crystals. Only clouds with cloud-top heights lower than 3 km are included in the analysis and the data are binned into 5° × 5° grid boxes.

c. CMIP3 models

The CMIP3 general circulation model (GCM) output used is derived from simulations of the twentieth-century climate (ending in 1999) that include forcing from anthropogenic greenhouse gases, atmospheric aerosols, solar variability, and so on according to available historical data. Model documentation and references can be found online (http://www-pcmdi.llnl.gov/ipcc/model_documentation/
We use an output period of the same length as the satellite data record, from January 1995 to December 1999. All model output is linearly interpolated onto a common 2.5° × 2.5° grid to be comparable with the CERES–MODIS data.

For albedo calculation from model output, we use deseasonalized monthly means of incoming and reflected SW radiative fluxes at the TOA. For cloud fraction estimates from the model output, the model parameter “total cloud fraction” is used in the analysis. This value is a result of different cloud overlapping schemes in different models, but regardless of the cloud overlap algorithm the total cloud fraction parameter represents the radiatively effective cloud fraction, or the cloud fraction as seen from above, in each model.

d. Cloud fraction in models and observations

Defining and determining cloud fraction accurately is problematic and its value is to some degree method dependent. Yet, \( f \) plays a significant role in the earth’s climate system through its application in Eq. (1). This equation and the concept of \( f \) have been adapted to numerous surface and satellite observation schemes and to radiation and climate models (e.g., Hahn et al. 2001; Collins 2001; Martins et al. 2002; Chepfer et al. 2008), and accordingly it is also used as the basis for the present analysis.

The various satellite retrievals and models used have cloud detection schemes that are objective and internally consistent, but they do not have exactly the same definition of cloud fraction, and the threshold values for cloud formation and identification are not the same across models and observations. Satellite retrievals of global cloud fraction are based on the occurrence of cloudy conditions in individual satellite pixels, or lidar profiles in the case of CALIPSO, resulting in a gridded fractional cloud cover. For GCMs, on the other hand, the fractional cloud cover of each grid box is calculated based on the overlapping of cloud at different levels in the atmospheric column. Despite these differences, across the different datasets a near-linear relation between albedo and cloud fraction is generally present, which we see as adding credibility to our approach.

4. Results

a. CERES and MODIS satellite observations

As seen in Fig. 1, monthly mean satellite observations of albedo from CERES and cloud fraction from MODIS on the Terra and Aqua satellites for the selected areas do indeed display closely linear relations for the sampled ranges of \( f \).

In other words, on a monthly mean time scale, Eqs. (1)–(3) seem to hold, and the slope and intercept of the regression line shown in the plot can be used to estimate \( a_{\text{cloud}} \). The values of the correlation coefficient for the relations, as well as estimated cloud albedo values and related variability ranges, are summarized in Table 1.

The lack of data points for low values of \( f \) (below approximately 0.3) prevents investigation of the linearity in this range, and the physical meaning of the intercept of the
linear fit is therefore not certain. Accounting for the possibility that the linear approximation may not be appropriate for low cloud fractions, we do not lay emphasis on the $a_{\text{clear}}$ estimate. We note that the variability in the value of the intercept, corresponding to the combined effect of the surface albedo and the Rayleigh scattering by air molecules and aerosols, is an order of magnitude smaller than that of the cloud albedo estimates between the studied regions.

Figure 1 and Table 1 illustrate that the method for determining cloud albedo can even capture an apparent diurnal variation in cloud albedo in these stratocumulus cloud regions. Data from the CERES and MODIS instruments on the Aqua satellite, which has an afternoon overpass (approximately 3 h later than the Terra satellite), show flatter slopes, indicating lower estimated cloud albedos than the Terra data. This is consistent with previous findings regarding the diurnal cycle of marine stratocumulus clouds (Cahalan et al. 1994; Rozendaal et al. 1995; Wood et al. 2002).

b. CMIP3 models

The relation between cloud fraction and albedo given by Eqs. (1)–(3) and supported by the satellite data may also reasonably be expected to appear in GCMs. Figure 2 displays the results for two different fully coupled GCMs from the CMIP3 multimodel data archive: the Hadley Centre Global Environmental Model, version 1, (HadGEM1) and the Community Climate System Model, version 3 (CCSM3). These two models both show distinct

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**Table 1.** Table indicating correlation coefficient and estimated cloud albedo for three different marine stratocumulus cloud regions using satellite observations and model output. The variability ranges presented for cloud albedos are the ranges given by a 95% confidence interval for the coefficients of the regression line fit to the data. The number of data points is 992 for CERES–MODIS and 960 for HadGEM1 and CCSM3.

<table>
<thead>
<tr>
<th></th>
<th>Peruvian</th>
<th>Namibian</th>
<th>Californian</th>
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<tr>
<td>CERES–MODIS (Terra)</td>
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<tr>
<td>$a_{\text{cloud}}$</td>
<td>0.42 ± 0.016</td>
<td>0.39 ± 0.016</td>
<td>0.40 ± 0.015</td>
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<tr>
<td>CERES–MODIS (Aqua)</td>
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<tr>
<td>$a_{\text{cloud}}$</td>
<td>0.37 ± 0.011</td>
<td>0.35 ± 0.013</td>
<td>0.37 ± 0.012</td>
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<tr>
<td>HadGEM1</td>
<td></td>
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<tr>
<td>$a_{\text{cloud}}$</td>
<td>0.40 ± 0.019</td>
<td>0.42 ± 0.023</td>
<td>0.39 ± 0.025</td>
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<tr>
<td>CCSM3</td>
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<tr>
<td>$a_{\text{cloud}}$</td>
<td>0.30 ± 0.050</td>
<td>0.32 ± 0.046</td>
<td>0.42 ± 0.031</td>
</tr>
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**Fig. 2.** Scatterplots showing the relations between monthly mean albedo and monthly mean cloud fraction in two CMIP3 GCMs: (top) HadGEM1 and (bottom) CCSM3, during 1995–99 in (left to right) three marine stratocumulus cloud regions.
climatological maxima in cloudiness corresponding to the marine stratocumulus decks of interest, which is not the case for all models. For the Californian region, both models show agreement with the data with regard to the relation between albedo and cloud fraction. For the Peruvian and Namibian regions, HadGEM1 remains in agreement with the satellite data, whereas CCSM3 deviates from linearity. The correlation between albedo and cloud fraction increases somewhat when the geographical regions are slightly shifted to better capture the cloud maxima in CCSM3, but the correlation coefficients remain below 0.6, and linear relations do not become apparent. CCSM3 differs from the rest of the CMIP3 models in that it employs a parameterization where low cloud fraction increases with lower-tropospheric stability, based on the empirical findings of Klein and Hartmann (1993), for several stratocumulus cloud regions. This results in a simulated low-cloud fraction in better agreement with observational data than many other CMIP3 models (Karlsson et al. 2008), but still the radiative characteristics of the cloud cover do not agree as well with observations. HadGEM1 has an interactive parameterization of cloudiness as a function of local variability in humidity, and a boundary layer scheme with explicit stratocumulus entrainment. This model has previously been identified as the only CMIP3 model that captures the relationships between northeast Pacific Ocean low-level cloudiness and related meteorological parameters (Clement et al. 2009). In our analysis this model cannot be singled out, but is one of several models that shows fairly good agreement with the data. (See Fig. 3.) As pointed out by Broccoli and Klein (2010), the Geophysical Fluid Dynamics Laboratory Climate Model, version 2.1, (GFDLCM2.1) shares a common formulation of boundary layer mixing with HadGEM1, which is also true for GFDLCM2.0. These models also belong to those that show good agreement with the data for all three regions.

Results for 19 CMIP3 models are summarized in Fig. 3, showing the large spread in correlations and cloud albedo estimates within and between models.

In general, poor correlations and nonlinearity between albedo and cloud fraction may be effects of the nonhomogeneity of the model cloud cover in the studied regions. For instance, the presence of overlying high clouds, or cumulus in the middle of a stratus deck, could contaminate the linear relation. Yet another reason for larger spread may be the diurnal variation in cloud albedo, made apparent in Fig. 1. Whereas CERES assumes a constant cloud scene, the cloud scene in the models will vary throughout the day.

A majority of the 19 CMIP3 models considered have correlation coefficients between albedo and cloud fraction of 0.8 or higher for the studied regions, and the multimodel mean cloud albedo estimates agree within 20% with the CERES–MODIS satellite-based estimates. In most cases, however, the model estimates fall beyond the ranges indicated by satellite observations, as seen in Fig. 3.

The variabilities for the CMIP3 model estimates are also consistently larger than those for the CERES–MODIS
estimates, as seen in Fig. 3 and Table 1. As expected from the lower correlation, the estimated low-level cloud albedo variability is larger for CCSM3 than for HadGEM1, with variabilities of up to nearly 620%. (See Table 1.)

For the CERES–MODIS-estimated cloud albedo, the variabilities are within 65%.

c. CALIPSO satellite observations

Further support for this method for quantifying cloud albedo is given from its application to CALIPSO satellite-borne lidar data. Figure 4 displays the relation between cloud fraction and so-called lidar albedo (i.e., IABS; Charlson et al. 2007) for the three marine stratocumulus cloud regions.

Similar to the CERES–MODIS data, they show a high correlation and near-linear relation. This demonstrates that the results are not related to the use of measurements of reflected radiances of sunlight for both albedo and cloud cover, as is done by CERES and MODIS. The properties of the best-fit line to the CALIPSO data are indicated in Fig. 4. As they reflect the relation between cloud fraction and lidar albedo rather than actual albedo, they are not directly quantitatively comparable to the CERES–MODIS and model results. But translating the proxy data to broadband albedo, using the relation suggested by Charlson et al. (2007), yields values of cloud albedo broadly consistent with those derived from CERES–MODIS.

For the Namibian region, August and September data are excluded. Data for these months, which coincide with the peak biomass burning season in southern Africa (Eck et al. 2003), cause large spread at high cloud fractions because of attenuation of the lidar backscatter signal by aerosol layers overlying the clouds (see Chand et al. 2008). The integrated attenuated backscatter from the CALIPSO lidar, penetrating deep into the atmosphere within a narrow field of view, is more sensitive to such a signal than the passive CERES–MODIS sensors.

d. Other geographical regions

The focus of this paper is on regions of persistent stratiform cloudiness, but we find that similar results (i.e., near-linear relations between albedo and cloud fraction on monthly mean scales) hold for other ocean regions as well and are applicable not only to this cloud type. For instance, the same method can be applied to the North Pacific (30°–50°N, 160°E–160°W), with similar linearity, but higher values of estimated cloud albedo, or to the western Pacific (15°–0°S, 120°E–180°E), with similar values of cloud albedo (see Fig. 5). For these regions, Terra and Aqua estimates coincide, and there is no indication of diurnal variability in cloud albedo. For the western Pacific the cloud fraction range reaches to 0 with remaining linearity, indicating that at least for this instance the intercept of the best-fit line with the y axis can be physically interpreted. Figure 5 also illustrates that the simultaneous presence of several cloud types can disrupt the linearity, as stated in section 1. Looking at the much larger area of the entire Pacific (65°S–65°N, 120°E–120°W), a simple linear relation between cloud fraction and albedo cannot be seen.

Fig. 4. Scatterplot showing the relation between monthly mean lidar albedo (column-integrated attenuated backscatter at 532 nm) and monthly mean low cloud fraction from CALIPSO, during October 2006–September 2008, in (left to right) three marine stratocumulus cloud regions. The number of data points is 96 for the Peruvian and Californian regions and 80 for the Namibian region because of the exclusion of August and September data. According to the relation between albedo proxy and broadband CERES albedo displayed in Fig. 2 of Charlson et al. (2007), the estimated cloud albedo proxies (given by the sum of the slope and the intercept of the best-fit straight line) correspond to cloud albedos of 0.38, 0.36, and 0.39, respectively.
Independent satellite observations, from CERES–MODIS as well as from CALIPSO, over stratiform marine clouds show a close, near-linear relationship between monthly mean cloud fraction and albedo on a regional scale. We show how the properties of the relationship can provide information on the magnitude and variability of a regional monthly mean “effective” cloud albedo. Hereby, we present a method by which consistent estimates of cloud albedo in a mixed clear–cloudy scene can be made.

Applicable to models as well as observational data, the method of analysis constitutes a useful diagnostic tool. The degree of linearity as well as the magnitude of the linear slope in the relation between albedo and cloud fraction can be used as a test of the ability of models to reproduce the observational cloud radiative characteristics. Among present-generation climate models in the CMIP3 data archive there are examples of good, as well as poor, agreement with data in this regard.

The fact that satellite data indicate a near-linear relationship between albedo and cloud fraction supports the use of Eqs. (1)–(3) as a model, meaning that on the temporal and spatial scales studied, clouds and clear skies can be adequately separated, and cloud albedo and clear albedo can be considered constant, independent of cloud fraction. In other words, for the range of cloud fractions in each of the regions studied, a single value of cloud albedo on a monthly mean scale emerges, demonstrating that the radiative impacts of the clouds are primarily determined by the variation in cloud fraction, rather than by cloud albedo variation.

The cloud albedo indicated for marine stratocumulus cloud regions, which are the focus of the present study, is low, only about 20%–40% higher than the global mean total albedo. This can be compared with local and instantaneous cloud albedo estimates from aircraft, which show large diversity but are in many cases higher. While aircraft measurements of albedo are made of small numbers of individual clouds, a satellite or model grid square will contain a mixture of cloudy and clear skies, and the “effective” cloud albedo estimated from the relation between average cloud fraction and albedo in these grid squares is hence a statistical representation of the radiative properties of clouds on a regional scale.

Regionalizing the analysis is not only found to be useful as a way of separating different cloud types, but is also necessary in order to avoid nonlinearities due to the presence and averaging of clouds with different optical properties. We focus on marine stratocumulus clouds and show examples of only a few regions, but the analysis could be expanded to include additional regions, comparing their properties.

The method presented clearly allows for separation of the influence of cloud fraction from cloud albedo in the determination of the albedo of a whole scene or region. This makes it further applicable to detection and quantification of microphysical effects of cloud droplet size and population on cloud albedo, and of possible indirect aerosol effects on clouds. With simultaneous aerosol observations, the cloud and albedo data can be segregated by aerosol properties, and changes in estimated cloud albedo, if large enough, can reveal possible aerosol effects on the cloud reflectivity.

![FIG. 5. Scatterplots showing the relations between monthly mean albedo from CERES and monthly mean cloud fraction from MODIS, during 2002–07, in three regions: (left) North Pacific, (middle) western Pacific, and (right) all of the Pacific Ocean. Observations are from the Terra satellite.](image-url)
The present analysis leaves open several questions with regard to the generality of the linear relation between albedo and cloud fraction: the spatial and temporal time scales on which it is applicable, the cloud-type scenes it may be applied to, and the validity of its extrapolation to low values of cloud fraction and its subsequent use for determining clear-sky albedo. Also, the reasons for model agreement and disagreement with data, and possible use of the relation to constrain models, require further investigation, with the presented approach as a framework.

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