Influence of Absorbing Aerosols on the Inference of Solar Surface Radiation Budget and Cloud Absorption

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ABSTRACT

This study addresses the impact of absorbing aerosols on the retrieval of the solar surface radiation budget (SSRB) and on the inference of cloud absorption using multiple global datasets. The data pertain to the radiation budgets at the top of the atmosphere (TOA), at the surface, and to precipitation and tropical biomass burning. Satellite-based SSRB data were derived from the Earth Radiation Budget Experiment and the International Satellite Cloud Climatology Program using different inversion algorithms. A manifestation of the aerosol effect emerges from a zonal comparison between satellite-based and surface-observed SSRB, which shows good agreement in most regions except over the tropical continents active in biomass burning. Another indication arises from the variation of the ratio of cloud radiative forcing at the TOA and at the surface, which was used in many recent studies addressing the cloud absorption problem. The author's studies showed that the ratio is around unity under most circumstances except when there is heavy urban/industrial pollution or fires. These exceptions register discrepancy between observed and modeled SSRB. The discrepancy is found to increase with decreasing cloudiness, implying that it has more to do with the treatment of aerosols than clouds, although minor influences by other factors may also exist. The largest discrepancy is observed in the month of minimal cloud cover and maximal aerosol loading. The corresponding maximum monthly mean aerosol optical thickness is estimated to be around 1.0 by a parameterization developed in this study. After the effects of aerosols on SSRB are accounted for using biomass burning and precipitation data, disagreements no longer exist between the theory and observation with regard to the transfer of solar radiation. It should be pointed out that the tropical data employed in this study are limited to a small number of continental sites.

1. Introduction

Clouds and aerosols are the two most important factors modulating the solar energy reaching the ground and trapped in the atmosphere. They thereby play key roles in the earth's climate (Charlock and Sellers 1980; Coakley et al. 1987; Hobbs 1993). Understanding their roles is an immense and painstaking endeavor as it entails knowledge of their spatial distribution, temporal evolution, and the processes controlling their changes and interactions with other components of the earth's climate system (Hansen et al. 1995). While clouds have been monitored for over a century at ground and more than a decade from space, many critical cloud properties are still badly wanting such as cloud vertical structure, phase, and microphysics (Wielicki et al. 1995). Apart from the complex interactions and feedbacks between clouds, radiation, and climate, the basic knowledge of clouds in absorbing solar energy is still being fervently debated (Cess et al. 1995; Ramanathan et al. 1995; Pilewski and Valero 1995; Chou et al. 1995; Li et al. 1995a; Li and Moreau 1996; Wiscombe 1995; Charlock et al. 1995; Arking et al. 1996; Arking 1996; Ackerman and Toon 1996; Stephens 1996; Cess and Zhang 1996; Pilewski and Valero 1996, etc.). It is thus not surprising that the treatment of clouds and the associated feedbacks in general circulation models (GCMs) constitutes the largest uncertainty in climate modeling (Cess et al. 1989; IPCC 1995).

The importance of aerosols was underlined by the finding that adding anthropogenic aerosols (e.g., Langner and Rodhe 1991) to GCMs can greatly ameliorate changes due to the buildup of greenhouse gases (Charlson et al. 1992; Kiehl and Briegleb 1993; Mitchell et al. 1995). However, the aerosol climatology used so far is extremely crude and deals mostly with sulfate particles. We have very poor knowledge of many other types of aerosols that are of climatic significance such as those produced from biomass burning (Penner et al. 1992). Smoke particles can significantly modify the earth's radiation budget and boundary layer meteorology by reflecting sunlight to space and absorbing solar radiation in the atmosphere (Coakley et al. 1983; Christopher et al. 1996). While numerous projects have been in place for monitoring and studying different types of aerosols (NASA 1996), routine global observation for

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even such a basic aerosol attribute as optical thickness will not be feasible until the Earth Observation System era (King et al. 1992), despite limited success achieved over oceans (Rao et al. 1989; Ignatov et al. 1995). A complete characterization of aerosol optical properties is notoriously difficult as it requires information about, among others, particle size distribution and chemical composition, both of which are extremely difficult to obtain on large scales and on a routine basis. Therefore, before extensive and credible aerosol data become available, regional and global studies on both the direct and indirect forcing by aerosols (Schwartz et al. 1995) need to resort to inference, in addition to direct observation.

This study, for example, infers the influence of strong absorbing aerosols on the solar surface radiation budget (SSRB). The work has a close tie with the author's recent investigation on the effect of clouds on atmospheric absorption of solar radiation (Li et al. 1995; Li and Moreau 1996). By virtue of the ratio (R) of shortwave cloud radiative forcing (CRF), the difference in solar net radiative fluxes between all-sky and clear-sky conditions, at the top of the atmosphere (TOA) to that at the surface (SFC), that is,

$$R = CRF_{SFC}/CRF_{TOA}, \quad (1)$$

several studies (Cess et al. 1995; Ramanathan et al. 1995; Pilewski and Valero 1995) suggested an enormous enhancement of solar absorption by clouds. The claim was based on their finding that the observed value of R is around 1.5, whereas modeled R is slightly larger than unity ($\sim$1.0–1.1). The discrepancy represents a difference of 25 W m^{-2} in global and annual mean SSRB due to an inadequate treatment of cloud absorption alone (Cess et al. 1995). However, our observational studies (Li et al. 1995a; Li and Moreau 1996) suggest that the ratio is more variable but agrees with model calculation in most circumstances except for a few cases in the midlatitudes and a lot of cases in the Tropics. Given that these exceptional cases correspond to either heavy industrial pollution in western Europe or active biomass burning in the Tropics, a premise was made that the large values of R stem from the effects of strong absorbing aerosols. The premise, albeit plausible, was questioned due to the lack of scientific insight (Wiscombe 1995). Moreover, a critical question remains open as to whether the large R signifies a genuine enhancement of cloud absorption due to the presence of absorbing aerosols embedded in clouds (Stephens and Tsay 1990), or simply an artifact resulting from an incorrect determination of R due to potential overestimation of the clear-sky SSRB by the algorithm of Li et al. (1993a), which does not account for strong absorbing aerosols. An attempt is made here to unravel the issue and to remedy the erroneous estimates of R that were obtained before. It is admitted that the present investigation is not rigorous per se, but it does provide more convincing evidence to support our proposition.

The following section describes briefly the datasets employed. The scientific issues that are dealt with are further elaborated in section 3. Section 4 analyzes the impact of absorbing aerosols on the inference of SSRB and on the determination of cloud radiative forcing ratio. Section 5 presents a simple method of estimating the approximate values of aerosol optical thickness with known TOA and surface radiation budgets. The study is concluded in section 6.

2. Datasets

Five observational datasets are employed in this study, namely, the Earth Radiation Budget Experiment (ERBE), the retrieved SSRB from ERBE and the International Satellite Cloud Climatology Program (ISCCP), the Global Energy Balance Archive (GEBA), the Global Precipitation Climatology Project (GPCP), and the Database for Tropical Biomass Burning (DTBB). While the original resolutions of these datasets vary, they were all converted into the equal-area cells of 280 km $\times$ 280 km as adopted in ISCCP C1 grid system. Since detailed descriptions for all of the datasets can be found elsewhere, a brief discussion on each dataset is given here.

ERBE was a multiple-satellite observation system that provided calibrated broadband shortwave and longwave radiation budgets at the TOA (Barkstrom et al. 1989). Regional and monthly mean fluxes under all-sky and clear-sky conditions for the period of 1985–89 are available, from which the CRF_{TOA} can be computed (Ramanathan et al. 1989). In addition, Li and Leighton (1993) derived a global dataset of SSRB from the ERBE TOA measurements using the inversion algorithm of Li et al. (1993a). While the algorithm contains several sets of coefficients for different sky and cloud conditions, the clear-sky set was employed for retrieving all-sky monthly mean SSRB following a validation study (Li et al. 1993b). Examination of the uncertainties in the resulting SSRB estimates with respect to clouds is thus instrumental in revealing if there exists any cloud absorption anomaly. In addition to the ERBE-based SSRB data, two other independent satellite-based SSRB products were also employed. They were derived from the ISCCP using the algorithm of Pinker and Laszlo (1992) and that of Zhang and Rossow (1995) and Rossow and Zhang (1995).

GEBA is an assembly of global monthly mean surface heat balance observations including SSRB prepared by the Swiss Federal Institute of Technology in Zurich, Switzerland (Ohmura and Gilgen 1991). GEBA radiation data encompass primarily surface downwelling irradiance (insolation) observed with pyranometers deployed in many countries around the world. The GEBA data employed here were extracted from the World Climate Research Project (WCRP) SRB dataset (Whitlock et al. 1995) in which the original site-specific surface irradiance data were averaged over the ISCCP C1 cells. Only monthly mean fluxes are available from GEBA that originate from the World Radiation Data Center (WRDC) located in St. Peterburg, Russia. Rigorous quality controls were applied to the WRDC data (Ohmura and Gilgen 1991).

To compare the ERBE-based estimates of SSRB with GEBA surface insolation, surface albedo is needed, since the algorithm of Li et al. (1993) retrieves directly the surface net solar flux. Surface broadband albedos have been inferred from clear-sky ERBE satellite measurements (Staylor and Wilber 1990; Li and Garand 1994) and ISCCP (Pinker and Laszlo 1992). Notwithstanding different inversion algorithms, the two ERBE-based products are in good agreement. All-sky surface albedos were obtained by correcting the clear-sky ERBE-based albedo to account for the effect of clouds (Darnell et al. 1992). Note that clouds can alter surface albedo by modifying the spectral distribution of incident irradiance that interacts with the spectral dependence of surface albedo. The resulting all-sky surface albedos were validated favorably in several regions (Whitlock et al. 1995) that were thus employed to convert SSRB into downwelling fluxes. Overall, the resulting surface insolation estimated from satellites, contains little bias errors and moderate random uncertainties with respect to GEBA observations (Li et al. 1995b).

Since satellite and ground data differ in spatial and temporal coverage, a comparison between them suffers from uncertainties resulting from mismatch. The uncertainty due to spatial mismatch depends primarily on the density of surface observation stations, ranging from 24 to 6 W m^{-2} as the number of surface stations increase from 1 to 10 within the area of an ISCCP cell (Li et al. 1995b). Errors due to temporal mismatch depend on the frequency of diurnal sampling. The satellite sampling rate for shortwave radiation is, on average, twice a day from two ERBE satellites. After a correction for diurnal variability that takes advantage of the variable overpass time by one of the ERBE satellites (ERBS) (Brooks et al. 1986), the overall temporal sampling error for the monthly mean regional ERBE product was estimated to be about 4 W m^{-2} (Wielicki et al. 1995). The temporal sampling error of surface measurements should be even smaller than satellite data, because of more frequent surface observation. Therefore, errors due to temporal mismatch are negligible, relative to spatial mismatch. Under most circumstances, it is a sound assumption that both the spatial and temporal sampling errors are random. The analysis is thus focused more on the trend of variation than on individual numbers.

GPCP contains terrestrial and global gridded monthly precipitation analyses for about 10 years starting from 1986 (WCRP 1990), produced at the Global Precipitation Climatology Centre (GPCC) in Germany. GPCC collected and applied quality control to the worldwide rain gauge measurements of monthly precipitation, and calculated aerial mean totals from conventionally observed data over land, which were merged with precipitation analyses from numerical weather forecasting and estimates from satellite-based retrievals. Global terrestrial precipitation data from 1986 to 1989 at 2.5° latitude–longitude grids are employed here.

The fifth dataset, DTBB, contains the amount of biomass burned in tropical America, Africa, and Asia during the late 1970s (Hao and Liu 1994). Monthly and annual total amounts were given at 5° × 5° latitude–longitude grids. While the data were differentiated by the type of burning such as forest fires, savanna fires, and burning of fuel wood and agricultural residues, this application does not distinguish fire types. The annual total amount of biomass burned was obtained from various surveys, whereas the monthly amounts were inferred from surface zone concentration, which is proportional to the burned biomass amount (Hao and Liu 1994). It is conceivable that fire activity changes significantly from one year to another as does the total amount of biomass burned. In comparison, the seasonal variability of fires is expected to be more stable, especially in the Tropics where fire density and frequency are largely controlled by rainfall, which generally shows good seasonal regularity (Arino and Mellnotte 1995). Therefore, the monthly proportions of burned biomass out of the total annual amount were adopted in this investigation.

3. Issues

While all the satellite products employed here have been compared to GEBA data (Li et al. 1995b; Rossow and Zhang 1995; Whitlock et al. 1995), no zonal comparisons were reported. It is from the zonal comparison that the potential effect of biomass burning aerosol emerges. Figure 1 presents a comparison between estimated (ERBE-based) and observed (GEBA) surface insolation for various latitude zones having GEBA stations. The number of GEBA stations within a latitude zone of 2.5° is given in the top panel. The middle and bottom panels denote the mean and one standard deviation of the differences within the latitudinal zone. It should be emphasized that the comparison should not be regarded as being between the *zonal means* of the observed and inferred surface insolation, because spatial sampling is far from sufficient to represent the true zonal mean values, especially in the Tropics and polar regions (cf. Fig. 6 of Li and Moreau 1996). GEBA stations are located only over a small portion of the continents. Nevertheless, the strong dependence of the difference on latitude shown in Fig. 1 is revealing: almost zero in the midlatitudes (30°–65°) and up to 30 W m^{-2} or more in the Tropics (positive) and the polar region (negative). Statistical uncertainties of the differences are denoted by the ratios of the standard deviation over the square root of the number of samples, both of which change dramatically with latitude. In the midlatitudes, the number of samples is much larger and the standard deviation is much smaller than in other regions. Therefore, the small differences in the midlatitudes are more reliable

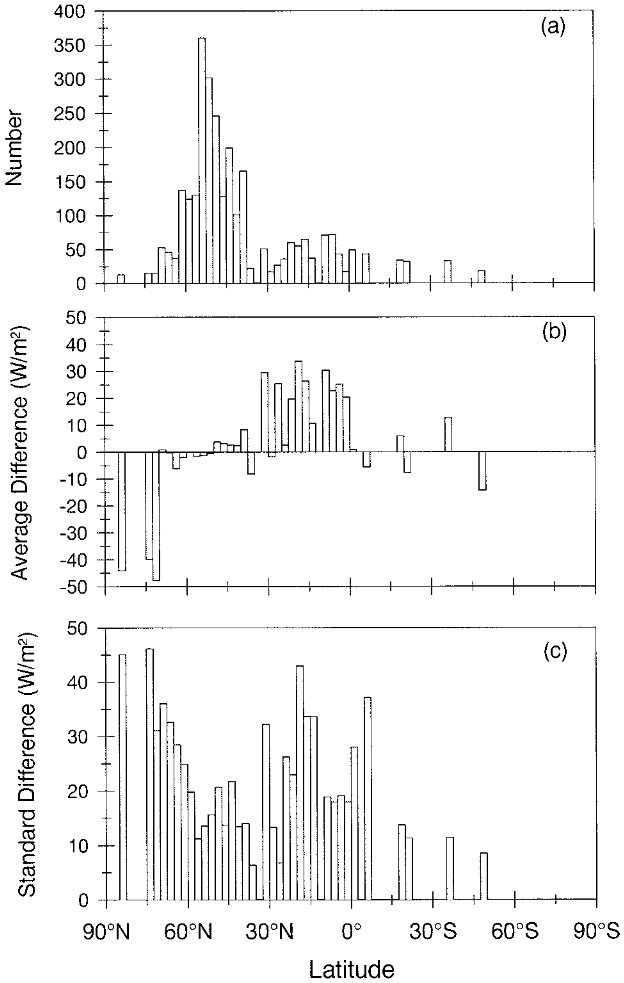

FIG. 1. Zonal comparisons of the monthly mean SSRB estimated from ERBE (Li and Leighton 1993) against ground-based measurements archived in GEBA: (a) number of data samples in each 2.5° latitude zone, (b) mean difference (satellite minus surface), and (c) standard difference.

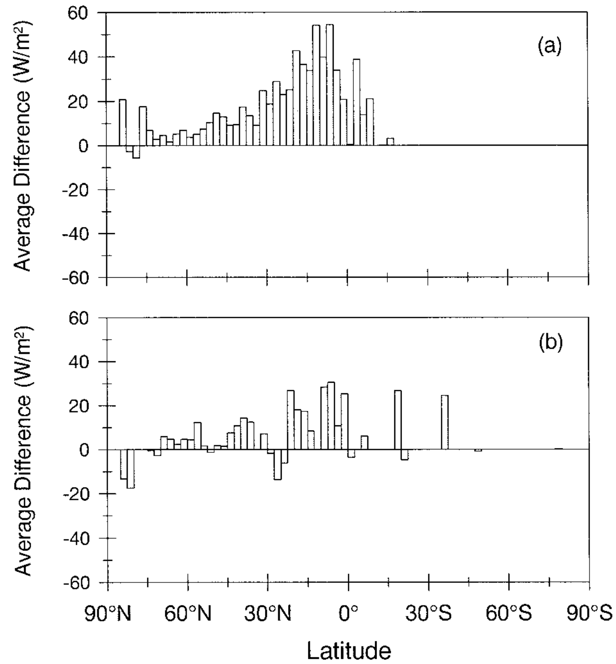

FIG. 2. Zonal comparisons of the monthly mean SSRB estimated from ISCCP using the algorithms of (a) Pinker and Laszlo (1992) and (b) Rossow and Zhang (1995) against ground-based observations from GEBA.

than the underestimation in the polar region and the overestimation in the Tropics. Interestingly, similar latitudinal trends also exist in the comparisons of the two ISCCP-based products, although their values are generally higher than observations in almost all latitude zones (Fig. 2).

The large overestimation in the Tropics, a common feature of the three comparisons, is a positive sign of the influence of biomass burning on the retrieval of SSRB, as many of the tropical GEBA sites are prone to fires during dry seasons. The algorithms and input data used for producing these satellite products differ in many respects but have a common shortcoming, no account being taken for the influence of biomass burning aerosols. Biomass burning is a frequent and widespread phenomenon in the tropical regions (Crutzen and Andreae 1990; Levine 1991; Cahoon et al. 1992), as is evident from Table 1, which shows the zonal mean amounts of total biomass burned in the tropical continents obtained by Hao and Liu (1994). Note that the correspondence between Table 1 and Figs. 1 and 2 is not very close, because of the nonuniform and sparse distribution of surface radiation stations and difference in observation period. For example, during the dry season of 1987, there were 350 000 fires in the Amazon Basin (Setzer and Pereira 1991) where there were no GEBA stations. But this does not mean that these fires had no impact on surface radiation measurements, since fire smoke can travel thousands of kilometers. Smoke particles are potentially far more absorbing than clouds and many other types of aerosols, due to the existence of graphitic (black) carbon (Ackerman and Toon 1981; Chylek and Ramaswamy 1982), a strong absorber of solar radiation in the solar spectrum (Chylek et al. 1984; Ramaswamy and Kiehl 1985). It can be shown that the algorithm of Li et al. (1993a) is not affected by con-

TABLE 1. Zonal and monthly mean amounts of biomass burned during the late 1970s (in teregrams of dry mass burned) (computed according to Hao and Liu 1994).

Northern Hemisphere						
30°–35°N	25°–30°N	20°–25°N	15°–20°N	10°–15°N	5°–10°N	0°–5°N
0.567	2.157	3.357	3.480	4.548	7.031	4.517
Southern Hemisphere						
5°S–0°	10°–5°S	15°–10°S	20°–15°S	25°–20°S	30°–25°S	35°–30°S
3.700	5.748	8.535	9.587	8.418	4.900	3.150

servative aerosols and influenced only slightly by weak absorbing aerosols. The single scattering albedo for biomass burning aerosols is highly variable but generally within 0.8 and 1.0 (Lenoble 1991) depending on many factors such as biomass type, weather condition, and the moisture content of burned material (Kaufman et al. 1994). The single scattering albedo for tropical biomass burning aerosols was estimated to be 0.90 ± 0.01 (Kaufman et al. 1992), which is similar to the model aerosol defined for rural continental conditions (WCP-112 1986). Aerosol optical thickness from biomass burning is even more variable depending on combustion efficiency, fire age, distance from a fire, and prevailing wind direction, etc. While aerosol optical thickness can be very high near a burning site, the background value is reported to be in the range of 0.5 to 1.0 in the Tropics (Holben et al. 1991; Kaufman et al. 1992).

The second analysis that may be affected by absorbing aerosols is concerned with the determination of cloud radiative forcing ratio (R) as defined in Eq. 1. The ratio indicates the *additional* effect of clouds on total absorption of the solar energy in the atmospheric column, relative to clear-sky atmospheric absorption. Here, $R > 1$ implies an enhancement of absorption by clouds with respect to the corresponding clear-sky counterpart, while $R < 1$ means that the column absorption is less for a cloudy atmosphere than for a clear one. The latter can occur for high clouds that reflect photons that would otherwise be absorbed by water vapor and aerosols in the lower atmosphere.

Li et al. (1995a) computed and analyzed the variation of R using global satellite and surface observations, in which $\mathrm{CRF}_{\mathrm{TOA}}$ was derived exclusively from ERBE while $\mathrm{CRF}_{\mathrm{SFC}}$ was determined from a combination of surface observation and satellite-based estimation. Following the definition of CRF, we have

$$\mathrm{CRF}_{\mathrm{SFC}} = \mathrm{INSO}_{\mathrm{ALL,SFC}}(1 - A_{\mathrm{ALL,SFC}}) - \mathrm{NET}_{\mathrm{CLR,SFC}}, \tag{2}$$

where $\mathrm{INSO}_{\mathrm{ALL,SFC}}$ and $A_{\mathrm{ALL,SFC}}$ denote all-sky monthly mean surface insolation and albedo respectively, and $\mathrm{NET}_{\mathrm{CLR,SFC}}$ denotes the clear-sky surface net solar flux. Among these three parameters, $\mathrm{INSO}_{\mathrm{ALL,SFC}}$ is most sensitive to cloud and has been observed routinely at the surface. Its values were thus taken from in situ measurements as archived in GEBA. The remaining two more conservative variables, $A_{\mathrm{ALL,SFC}}$ and $\mathrm{NET}_{\mathrm{CLR,SFC}}$, are not available from GEBA. They were estimated from ERBE using the algorithms described in Darnell et al. (1992) and Li et al. (1993a), respectively. While the majority of the ensuing values of R are around unity (Li et al. 1995a), its values in the Tropics are generally so large that they seem to corroborate the finding of anomalous cloud absorption (Cess et al. 1995; Ramanathan et al. 1995; Pilewski and Valero 1995). At first glance, this appears rather compelling as the smoke particles embedded in a cloud layer can enhance its ability to absorb solar radiation (Chylek et al. 1984). However, as demonstrated by Li and Moreau (1996), the ratio (R) is not an absolute measure of cloud absorption but a relative index indicating whether cloudy atmospheric absorption is larger than its clear-sky counterpart. Apparently, absorbing aerosols can also increase clear-sky atmospheric absorption by an amount that could be even larger than that under cloudy conditions. This is because the predominant scattering effect by cloud droplets diminishes the chance for photons to be captured by the smoke particles below clouds. While it is possible that some fire smokes can reach above the top of a cloud layer, the bulk of smoke is usually situated below the cloud layer. Therefore, we were skeptical about the values of R obtained in this region (Li et al. 1995a). It is likely that the high values of R originate from the erroneous estimation of A_{SFC} or $\mathrm{NET}_{\mathrm{SFC}}$ due to the neglect of the effect of absorbing aerosols.

4. Analysis of the impact of absorbing aerosols

The unavailability of ground-truth measurements on $A_{\mathrm{ALL,SFC}}$ and $\mathrm{NET}_{\mathrm{CLR,SFC}}$ does not allow an investigation into which parameter suffers greater uncertainty as a result of the inexplicit treatment of aerosols. A sensitivity test is conducted here to help determine the likelihood of the errors in the two estimated quantities. The test was done with a plane-parallel radiative transfer model that is described in detail by Masuda et al. (1995). The model is a doubling–adding code with 120 spectral bands applied to a vertically inhomogeneous atmosphere of eight layers. Radiative transfer calculations were carried out for the tropical model atmosphere of McClachey et al. (1972), over different types of surfaces (ocean, vegetated land, desert, and snow/ice) with the continental model aerosol (CON-I) of varying loadings (WCP-112). CON-I model aerosol was chosen, as its single scattering albedo is close to the median value of biomass burning aerosols from natural fires (Radke et al. 1988; Lenoble 1991) and, in particular, similar to those observed in the Tropics (Kaufman et al. 1992). Surface types are differentiated by their distinct spectral and angular reflectances.

Figure 3 shows the simulated relationships between the TOA and surface albedos and between surface net solar flux and TOA reflected flux for the solar zenith angle (SZA) of 60°. For a given SZA and aerosol loading, the relationship is highly linear but the intercept and slope is modified by aerosol optical thickness. The influence of aerosols on the retrieval of surface albedo from TOA albedo is minimal for terrestrial surfaces, and maximal for dark oceans and bright snow/ice-covered surfaces, as the absorbing aerosols tend to increase and decrease TOA albedo over dark and bright surfaces, respectively. This is conceivable from the well-known criterion proposed by Chylek and Coakley (1974) and Coakley and Chylek (1975) that determines the perturbation of TOA albedo by aerosols as a function of surface albedo, aerosol single-scattering albedo, and back-

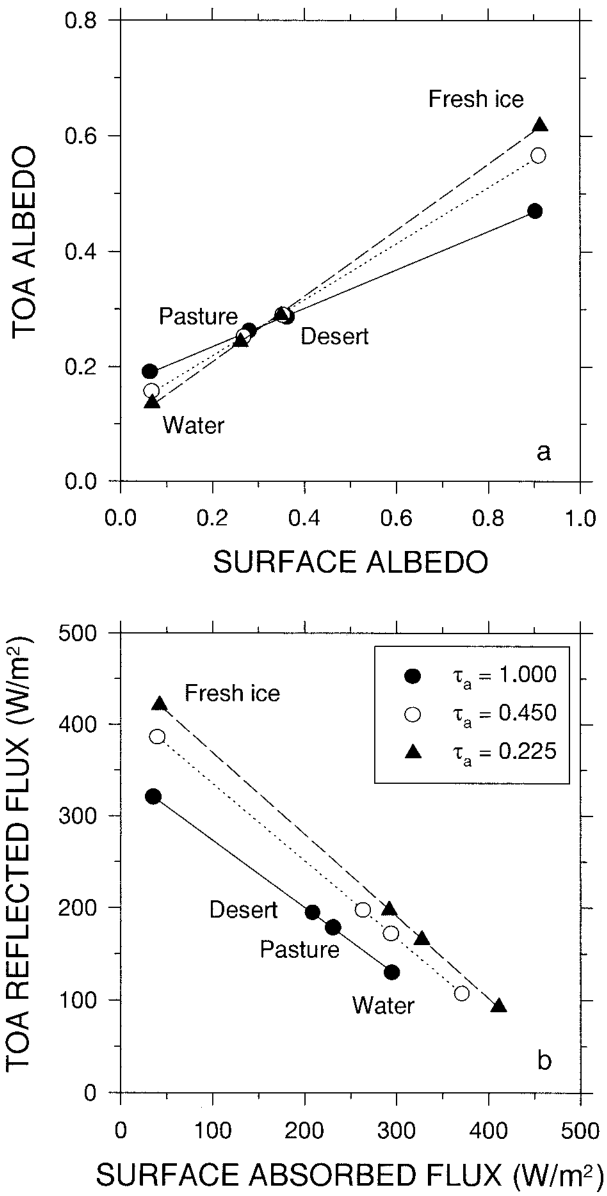

FIG. 3. Sensitivity tests of the relationship (a) between TOA and surface albedos and (b) between TOA reflected flux and surface absorbed flux for varying aerosol optical thickness. The data were simulated using a radiative transfer model with the continental aerosol under clear skies over different types of surfaces.

scattering fraction. In comparison, aerosols can diminish substantially the accuracy of the estimate of NET_{SFC} over any type of surface if variation in aerosol amount is not considered. It is thus envisioned that errors incurred in the determination of surface cloud radiative forcing by Eq. 2 are potentially tainted more by the estimation of $NET_{CLR,SFC}$ than $A_{ALL,SFC}$ for such an absorbing aerosol.

Since there are no separate records on monthly mean clear-sky radiation measurements, it is impossible to estimate directly the errors in the estimates of

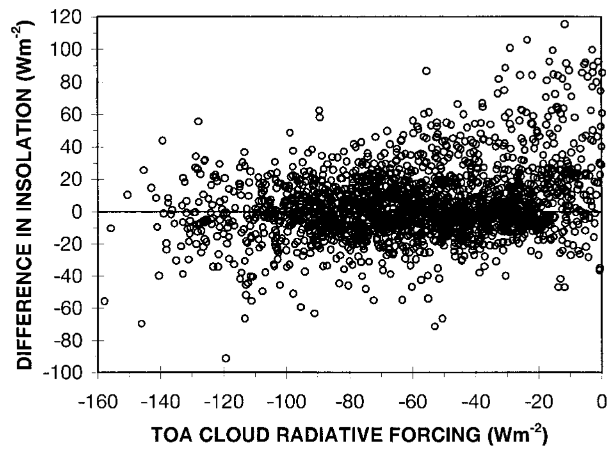

FIG. 4. Variation of the differences in surface insolation between satellite estimation and surface observation with TOA CRF.

$NET_{CLR,SFC}$. An indirect inference is thereby invoked based on the variation of the difference between estimated and observed $INSO_{SFC}$ with respect to cloud radiative forcing at the TOA, as is shown in Fig. 4. It is worth emphasizing that the estimation of surface insolation is for all-sky conditions, but the estimated quantities are not used for computing R but for inference of clear-sky estimation uncertainties. Intriguingly, Fig. 4 shows that the difference increases as the magnitude of CRF_{TOA} decreases. When $|CRF_{TOA}|$ is larger than 60 W m^{-2}, the mean difference is close to zero, suggesting that the estimates under clearer conditions are subject to larger uncertainties than under cloudy conditions. This trend is opposite of what would be expected if there were a cloud absorption anomaly. If clouds enhanced atmospheric absorption substantially relative to clear-sky absorption, the algorithm of Li et al. (1993a), which does not account for this effect, would overestimate SSRB under cloudy conditions and the overestimation should increase with the magnitude of CRF_{TOA}. Therefore, the trend revealed in Fig. 4 itself negates the claim of a cloud absorption anomaly but rather corroborates the argument that the systematic errors in SSRB from satellite estimation, or GCM simulation alike, are caused more by the calculation of radiative transfer under clear-sky conditions than under cloudy conditions (Barker and Li 1995; Arking 1996; Li et al. 1997).

The trend shown in Fig. 4 implies that the estimates of clear-sky surface net flux and the ensuing values of cloud-forcing ratio may be questionable in some cases. Li et al. (1995a) surmised that their estimates of large R are not as reliable as those of low R. This premise is confirmed by Fig. 5, which differentiates the results shown in Fig. 4 into $R > 1.2$ and $R < 1.2$. The value 1.2 represents approximately the typical upper limit of R obtained from a conventional radiative transfer model (Li and Moreau 1996). Remarkably, the increasing trend displayed in Fig. 4 corresponds almost exclusively to $R > 1.2$, confirming that these large values of R are

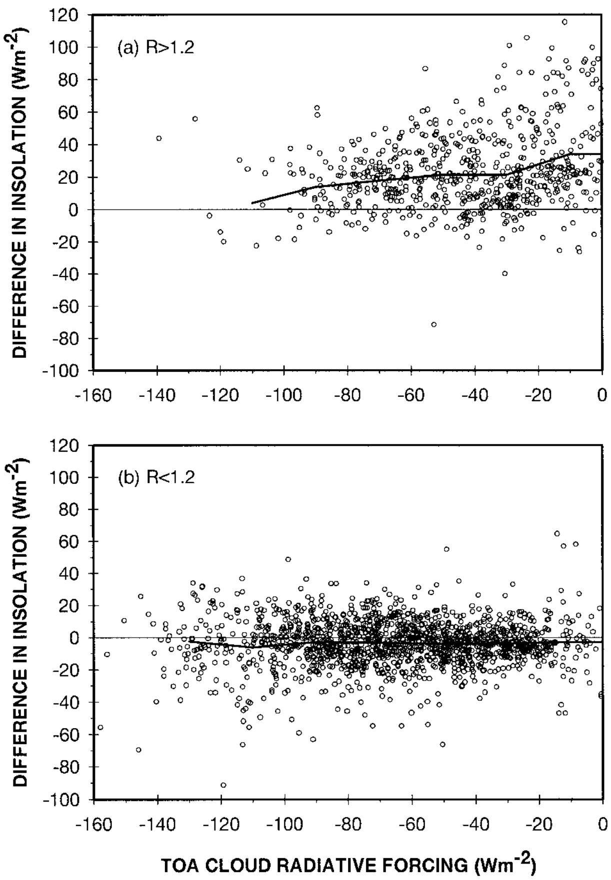

FIG. 5. Same as Fig. 4 but the data are grouped according the ratio of cloud radiative forcing (R) (a) greater and (b) less than 1.2. Also plotted are the curves of the mean differences in insolation computed for each CRF_{TOA} bin of 20 W m^{-2}.

spurious resulting from overestimation of $NET_{CLR,SFC}$. In contrast, the large majority of the cases with $R < 1.2$ exhibit little dependence on CRF_{TOA}, as the mean difference is close to zero for any value of CRF_{TOA} denoting variable amounts of cloud. Thus, these relatively low values of R are more accurate than the high ones. In addition, Fig. 5b suggests that clouds, overall, have relatively little impact on atmospheric absorption, in accord with conventional wisdom. However, this conclusion must be understood within the context of the uncertainty displayed in the figure.

As mentioned earlier, the differences between the estimated and observed surface fluxes encompass both true estimation error and artifacts resulting from the mismatch between satellite and surface measurements in time and space. The spread of the data points in Fig. 5b is comparable to the standard deviation of the matchup error (Li et al. 1995b). The matchup error appears to be random and independent of cloud condition, according to Fig. 5b. On the other hand, from the first glance at Fig. 5a, one could draw a conclusion that the trend is caused by the matchup error that is contingent upon cloud, greater under clear conditions than cloudy ones. While the data used here do not convey information to unravel the paradox unambiguously, the second explanation seems unlikely for the following reasons. First, a more detailed regional analysis (cf. Fig. 6) reveals that significant trends as shown in Fig. 5a are observed primarily in the tropical regions during the fire season. The matchup error due to temporal sampling in the Tropics should be even less than other regions as a result of transient observation time by ERBS. Second, the magnitude of potential temporal matchup is much smaller than the discrepancy shown in Fig. 5a, as discussed earlier. Third, there is no compelling reason to believe that the matchup errors due to spatial or temporal mismatch increase systematically as cloud amount decreases. Conversely, it is more plausible that the trend is induced by the absorbing aerosols whose effects are stronger under clear-sky conditions than cloudy conditions. The trend is thus more likely a physical nuisance, rather than a statistical artifact.

To gain more insight into, and to correct for, the nuisance, data presented in Fig. 5a are further scrutinized by analyzing the trend for each individual site. Such an analysis eliminates the differences in atmospheric and surface conditions among various sites. Besides, the two flux variables are normalized by the incident irradiance at the TOA so that data from different months are compatible as the seasonal cycle driven by the solar zenith angle and daylight duration is removed. Figure 6 presents the results for all the cases with $R > 1.2$ over four geographic regions abundant in absorbing aerosols: South America, Africa, Asia (near the equator), and western Europe. The proportions of the cases with $R > 1.2$ out of the total cases are equal to 0.75, 0.87, 0.67, and 0.14 for the above four regions, respectively. The problem is thus far more common in the Tropics than in midlatitudes, so far as the GEBA data are concerned. It follows from Fig. 6 that the difference in atmospheric transmittance varies approximately linearly with normalized CRF_{TOA}. The linearity is better defined in the three tropical regions, especially South America and Africa where fire activity is more frequent and widespread than other parts of the world. The high correlation coefficients in these regions may be explained by a dual dependence of the amount of biomass burned on precipitation that is further linked to CRF_{TOA}. First, precipitation has a scavenging effect, removing any aerosol particles from the atmosphere. Second, biomass burning activity in the Tropics is dictated primarily by precipitation. In comparison, precipitation has only a washout effect on industrial aerosol that is more plentiful in western Europe (Langner and Rodhe 1991). Low values of the correlation coefficient usually correspond to small ranges of variation in CRF_{TOA}, which are thereby subject to large statistical uncertainties.

The linear trend renders simple determination of the estimation error in clear-sky atmospheric transmittance. The intercept of the linear relationship between the difference in transmittance and normalized CRF_{TOA} denotes the maximum error in the estimated transmittance cor-

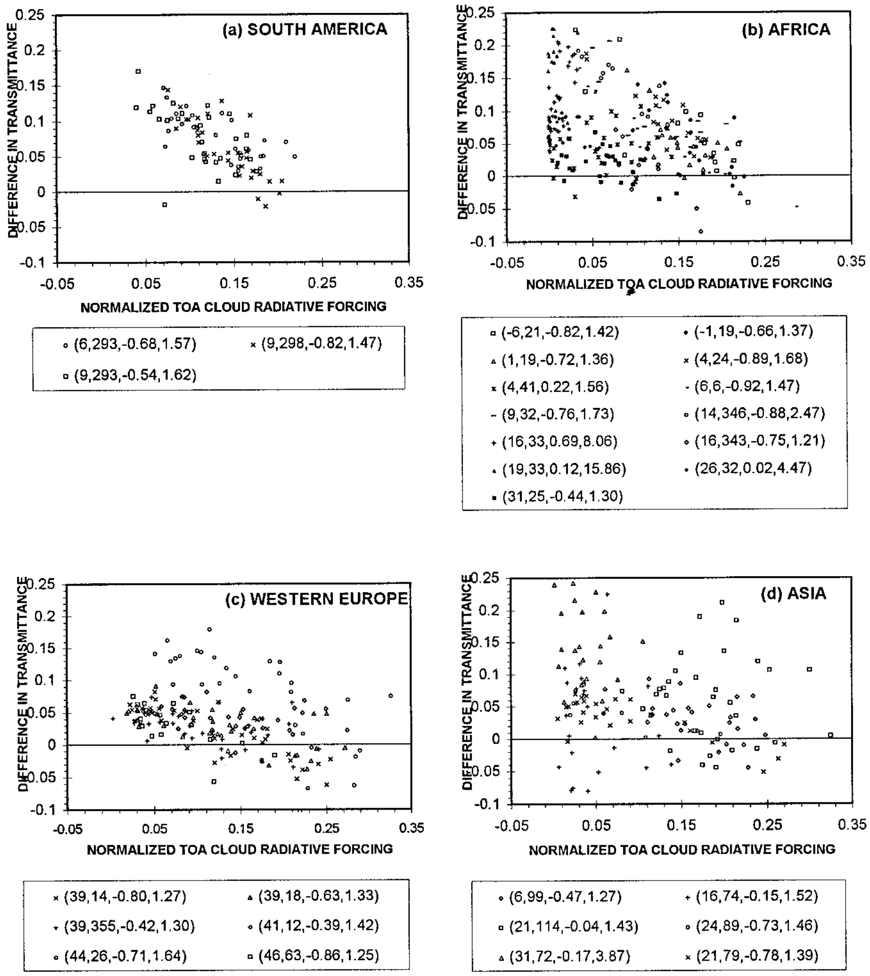

FIG. 6. Variation of the difference in atmospheric transmittance between satellite estimation and surface observation with normalized TOA shortwave cloud radiative forcing for four regions with $R > 1.2$ including (a) South America, (b) Africa, (c) Western Europe, and (d) Asia. Linear regression was conducted for each site. The four numbers beside the legends denote (from left to right) the rounded numbers of latitude (negative for SH) and longitude at the centers of ISCCP C1 cells containing GEBA sites, linear correlation coefficients, and the original value of cloud radiative forcing ratio.

responding to $\mathrm{CRF_{TOA}} = 0$. It is seen from Fig. 6 that the error ranges from 0 to 0.25, or 0 to 100 W m^{-2} in terms of insolation. This maximum error occurs during the haziest month with minimum cloud cover or precipitation. In other months of more cloud cover, errors in the estimates of clear-sky atmospheric transmittance or surface insolation should be smaller. This is, in particular, the case for tropical biomass burning aerosols whose loading is subject to strong seasonal variation as is its impact on the estimation of clear-sky fluxes. However, without clear-sky data in each month, errors for all months other than the haziest one cannot be determined directly. By necessity, they are approximated by the fractions of the maximal errors, and the reduction factor, f, is assumed to be proportional to an aerosol index. Therefore, the estimation errors in clear-sky atmospheric transmittance in all months are given by

$$\delta t' = f\delta t, \tag{3}$$

where δt is the intercept of the regression equation described above. The reduction factor f is equal to unity for the month(s) of maximum aerosol loading. In the Tropics, the amount of biomass burned estimated by Hao and Liu (1994) serves as the indicator, and f is simply taken as the fraction of individual monthly amount over the maximum monthly amount of biomass burned. In midlatitudes, monthly mean precipitation serves as the indicator and f is computed by

$$f = (p_{\max} - p)/(p_{\max} - p_{\min}), \tag{4}$$

where p, $p_{\max}$, and $p_{\min}$ denote the precipitation in the month of interest, and the maximum and minimum monthly mean precipitation during the year. The correction rests on an assumption that maximum error oc-

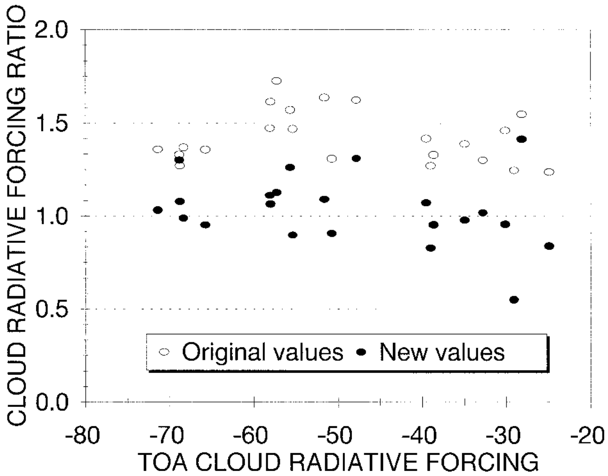

FIG. 7. Comparison of the original and revised values of cloud radiative forcing ratio for all the cases with the original value of R > 1.2 except those with TOA cloud forcing larger than -25 W m^{-2}.

curs in the month of maximum biomass burning in the Tropics or minimum precipitation in midlatitudes; no error exists for the month of minimum biomass burning or maximum precipitation; errors in the remaining months vary linearly with the aerosol indicator. While the assumption is somewhat ad-hoc, qualitatively it should be better than the two alternative assumptions, that is, that there are no errors in the clear-sky estimates as was implied in the previous studies (Li et al. 1995a; Li and Moreau 1996) and that the errors remain invariant throughout seasons as determined from the regression.

From $\delta t'$, the original estimates of clear-sky surface net solar flux ($NET_{CLR,SFC}$) and the surface cloud radiative forcing ratio R are modified. Revision was done by retaining the original values of all variables in Eq. (2) except for the estimates of $NET_{CLR,SFC}$ whose values were adjusted by amounts determined by $\delta t'$ and surface albedo. Figure 7 compares the old and new values of R for all cases with the original R larger than 1.2. The comparison is confined to TOA cloud radiative forcing less than -25 W m^{-2} in view of the matchup errors between satellite and surface measurements (Li et al. 1995b). Small signal-to-noise ratio for $CRF_{TOA} > -25$ W m^{-2} leads to a dramatic fluctuation in R. The new magnitudes of R are considerably smaller than the old ones, the mean values of the former and the latter being equal to 1.05 and 1.43, respectively. The revised values of R in the Tropics are not significantly at variance with model results (Li and Moreau 1996). Due to the crude correction, though physically sound, it cannot be claimed that these new numbers are absolutely correct. It is fair to say, however, that they are more credible than the old ones. These modified values are still slightly larger than model results that are generally less than unity for high-reaching tropical clouds (Li and Moreau 1996). The difference is, however, well within the estimation uncertainty due to poor knowledge on the vertical distribution of aerosols. After all, the study abates the possibility that absorbing aerosols lead to a substantial enhancement of solar absorption in a cloudy atmospheric column.

Note that this study has no bearing on other studies reporting anomalously high cloud absorption in the Tropics (e.g., Ramanathan et al. 1995; Pilewskie and Valero 1995). Their study regions are remote tropical oceans, far from the sources of absorbing aerosols as encountered in this study. If their findings are true and if absorbing aerosols play an essential role, a possible scenario for the enhanced absorption would be that the absorbing aerosols are blown away from the continental sources and overlay upon boundary marine clouds. In this case, a considerable increase in atmospheric absorption is possible, since the clouds serve as a multiple reflector rather than a shelter.

5. Estimation of aerosol optical thickness and its impact on the inference of SSRB

The above analyses show the effects of absorbing aerosols on the inference of SSRB. Such an impact can be accounted for if the optical properties of aerosols are known. Conversely, one may infer certain bulk aerosol properties if coincident and collocated TOA and surface radiation measurements are available. This section is intended more for demonstrating this proposition than presenting rigorous results.

As mentioned in the introduction, a complete characterization of the radiative properties of an aerosol requires several parameters. At present and in the near future, only aerosol optical thickness can be acquired with sound accuracy on a large-scale and routine basis from both ground-based and space-borne sensors. This parameter is fortunately the most significant variable affecting the transfer of solar radiation. Without observational information on other parameters, model aerosol types are often used. Based on the continental type of aerosol, for instance, Masuda et al. (1995) proposed a parameterization correcting the effect of variable aerosol optical thickness. The parameterization relates the difference in surface net solar flux normalized by the TOA incoming solar radiation (Δs) to aerosol optical thickness (τ) for a given TOA albedo (A_{TOA}) and cosine of solar zenith (μ) by

$$\Delta s = [0.00521 - 0.00246\mu - (0.09058 + 0.28465A_{TOA})\tau]. \qquad (5)$$

If surface observations of SSRB are available together with coincident and collocated A_{TOA}, one is able to estimate τ approximately by assuming a continental type of aerosol. As a result, the ensuing value of τ does not necessarily represent the true optical thickness but an equivalent one. For the biomass burning aerosol, the difference may not be very large since its optical properties are somewhat similar to those of CON-I. It should

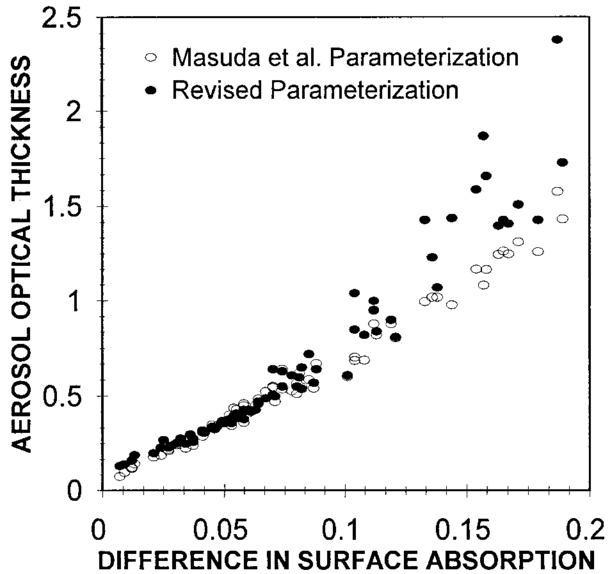

FIG. 8. Comparison of the aerosol optical thickness as a function of the difference in normalized surface absorbed flux derived from the two parameterization schemes using all data under study.

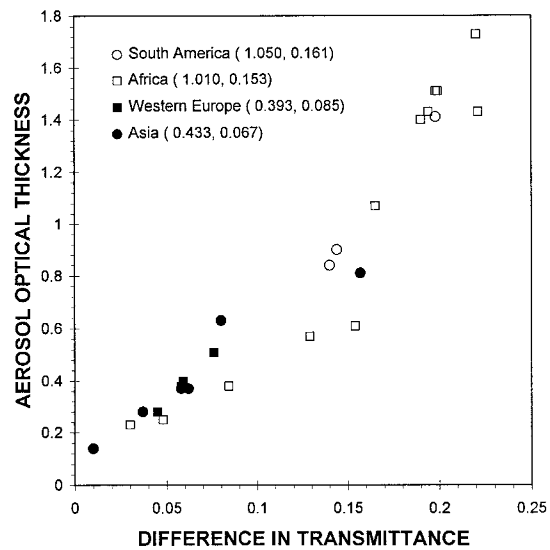

FIG. 9. Estimated aerosol optical thickness as a function of the atmospheric transmittance derived from the new parameterization for heavy smoky or polluted conditions as identified in Fig. 6 (original $R > 1.2$) over four regions.

be noted, however, that this parameterization was designed primarily for low and moderate aerosol loadings. It is clear from Fig. 3b that the relationship between SSRB and aerosol optical thickness is nonlinear. The linearity of the relationship (5) can lead to large errors for heavy aerosol conditions. Parameterization was thereby redone using the results of more extensive radiative transfer calculations with the model of Masuda et al. (1995) for seven different values of τ ranging from 0 to 3. Following the framework of Li et al. (1993a), the new parameterization is given by

$$\Delta s = \Delta\alpha + \Delta\beta A_{TOA} \tag{6}$$

$$\Delta\alpha = \alpha_1\Delta\tau + \alpha_2\Delta\tau^2 + \alpha_3\Delta\tau^3 \tag{7}$$

$$\Delta\beta = \beta_1\Delta\tau + \beta_2\Delta\tau^2 \tag{8}$$

$$\Delta\tau = \tau - 0.1 \tag{9}$$

$$\alpha_1 = -0.84038 \exp(-1.386\mu) - 0.4982 \log_{10}(\mu) + 0.1272 \tag{10}$$

$$\alpha_2 = 0.59792 \exp(-2.0419\mu) + 0.30306 \log_{10}(\mu) - 0.06751 \tag{11}$$

$$\alpha_3 = -0.10557 \exp(-2.2949\mu) - 0.05475 \log_{10}(\mu) + 0.1147 \tag{12}$$

$$\beta_1 = -0.2877 + 0.1597\mu - 0.2474\mu^2 + 0.1012\mu^3 \tag{13}$$

$$\beta_2 = -0.03, \tag{14}$$

where $\Delta\tau$ denotes the difference between actual aerosol optical thickness and the background one employed in the development of the inversion algorithm (Li et al. 1993a; Masuda et al. 1995). The new parameterization produces very close results to the detailed model calculations for τ up to 3, and similar results to Eq. (5) for τ up to 0.6, after which the agreement deteriorates quickly as τ increases. This is clearly seen from Fig. 8, which shows the relationship between Δs and τ where τ is inverted with both the original and new parameterizations using the observational values Δs, A_{TOA}, and μ for all the sites. Note that the inferred τ determined from the intercept of the regression discussed earlier corresponds to the haziest month at each site. They are thereby somewhat larger than the climatological values (d'Almeida et al. 1991). In addition, there was an inherent assumption that the discrepancy between surface-observed and satellite-estimated SSRB is due exclusively to aerosol. Errors caused by such factors as mismatch were factored into aerosol effect. This may explain the scattering around the general trend exhibited in Fig. 8. The majority of the moderate and large values of τ correspond to the cases of either heavy biomass burning or industrial pollution as identified in Figs. 6 and 7 (cf. Fig. 9). Aerosol loading is rather high in two active fire regions: South America and Africa, with a mean aerosol optical thickness of 1.05 and 1.01, respectively. These numbers compare well with the ground-based measurements made in the Tropics during the burning season (Holben et al. 1991). In comparison, aerosol loading is relatively low in two other regions with the mean optical thickness of 0.39 and 0.43 for western Europe and Asia, respectively.

While such heavy aerosol loadings can drastically influence the inference of regional SSRB as indicated by Fig. 9, their impact on the estimates of zonal and global mean SSRB is much smaller. If the explanation of biomass burning holds, the *real* zonal mean bias errors in the satellite-based estimates of SSRB should be considerably smaller than the differences shown in Figs. 1 and 2, since biomass burning occurs primarily over land during the dry season only, about three months. Tropical landmass encompasses about one-fourth the area of the Tropics (30°S to 30°N), which occupies half the area of the earth. Assuming that the effect of biomass burning extends across the entire tropical continent for half a year, the upper limit of the bias errors in the estimates of global and annual mean SSRB resulting from biomass burning is estimated to be less than 2 W m^{-2}. While this may be significant for climate change studies, it is much smaller than the discrepancy found among various contemporary estimates of SSRB (Li et al. 1997).

6. Summary

This study infers the influence of absorbing aerosols, in particular those produced from tropical biomass burning, on the estimation of the solar surface radiation budget (SSRB) and on the investigations of the effects of clouds on atmospheric absorption of solar energy using the cloud radiative forcing (CRF) ratio. Multiple observational datasets were employed including the Earth Radiation Budget Experiment (ERBE), satellite-based products of SSRB retrieved from ERBE and the International Satellite Cloud Climatology Program (ISCCP), the Global Energy Balance Archive (GEBA), the Global Precipitation Climatology Project (GPCP), and a Database for Tropical Biomass Burning.

Comparisons in different latitudinal zones of the SSRB derived from ERBE and ISCCP using independent retrieving algorithms against GEBA surface measurements exhibit a similar zonal trend: better agreement in the midlatitude than in The tropics. While the statistical uncertainties of the differences are much larger in the Tropics than in midlatitudes due to fewer observation stations and shorter duration, the common trend discloses the influence of strong absorbing aerosols produced from biomass burning on SSRB. This effect was also conjectured to produce the CRF ratio much larger than model calculations in our earlier papers (Li et al. 1995a; Li and Moreau 1996). It was unclear then whether the large CRF ratio denotes the enhancement of atmospheric absorption by clouds because of the presence of graphitic carbon, or is an artifact due to potentially incorrect estimation of SSRB under clear-sky conditions.

To address this question, the differences between the SSRB estimated from ERBE and the measurements contained in GEBA are analyzed with respect to the TOA CRF (CRF_{TOA}). The trend of the variation is found to be instrumental in revealing the effect of clouds on atmospheric absorption and in detecting the effect of aerosols on estimation of SSRB. For the majority of cases under study, no significant trends are found and the mean differences are close to zero throughout the range of CRF_{TOA} denoting variable amounts of cloud. This reinforces the concept that clouds have little impact on the absorption of solar radiation in the atmospheric column, as we understand from the conventional radiative transfer theory/model. Notable trends do exist for a small number of cases that are, however, opposite to what would be expected if there was a significant cloud absorption anomaly, since the estimation error increases as cloud amount decreases. These cases occur mostly in the tropical regions abundant with biomass burning aerosols and in parts of western Europe with heavy pollution. Maximum aerosol optical thickness is estimated to be around 1.0 in parts of Africa and South America, and around 0.4 in parts of western Europe and Asia. The spatial trend corroborates the effect of absorbing aerosols on inference of SSRB that reaches a maximum under clear-sky conditions.

Under most circumstances, the relationship between the difference in atmospheric transmittance and the TOA normalized CRF is linear. This allows one to infer the errors in the satellite-based estimates of SSRB under clear-sky conditions, which are unavailable from observation. Maximum estimation error for a given site was approximated by the intercept of the linear regression. Estimation errors in other months were computed based on an ad hoc assumption that the error is proportional to the amount of burned biomass in the Tropics and inversely proportional to precipitation in other regions. After correcting these errors, the mean CRF ratio in the Tropics diminishes from 1.5 to 1.0. It is thus concluded that significant enhancement of atmospheric absorption by clouds is unlikely even in regions with the presence of strong absorbing aerosols, which is in agreement with the finding of Chylek and Wong (1995). It is admitted that this conclusion is based partially on inference and partially on observation. The attribution to absorbing aerosols constitutes a more likely explanation, but not necessarily the sole one. Other factors such as water vapor may also contribute somewhat to the discrepancy, but they seem to be minor agents.

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