On Solar Energy Disposition: A Perspective from Observation and Modeling

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ABSTRACT

Solar energy disposition (SED) concerns the amount of solar radiation reflected to space, absorbed in the atmosphere, and absorbed at the surface. The state of knowledge on SED is examined by comparing eight datasets from surface and satellite observation and modeling by general circulation models. The discrepancies among these contemporary estimates of SED are so large that wisdom on conventional SED is wanting. Thanks to satellite observations, the earth's radiation budget (ERB) at the top of the atmosphere is reasonably well known. Current GCMs manage to reproduce a reasonable global and annual mean ERB, but often fail to simulate the variations in ERB associated with certain cloud regimes such as tropical convection and storm tracks. In comparison to ERB, knowledge of the surface radiation budget (SRB) and the atmospheric radiation budget (ARB) is still rather poor, owing to the inherent problems in both in situ observations and remote sensing. The major shortcoming of in situ observations lies in insufficient sampling, while the remote sensing techniques suffer from lack of information on some variables affecting the radiative transfer process, and dependence, directly or indirectly, on radiative transfer models. Nevertheless, satellite-based SRB products agree fairly well overall with ground-based observations. GCM-simulated SRBs and ARBs are not only subject to large regional uncertainties associated with clouds, but also to systematic errors of the order of 25 W m^{-2}, due possibly to the neglect of aerosol and/or inaccurate computation of water vapor absorption. Analyses of various datasets suggest that the SED based on ERBE satellite data appears to be more reliable, indicating 30% reflection to space, 24% absorption in the atmosphere, and 46% absorption at the surface.

1. Introduction

Solar energy reaching our planet is partly reflected to space, partly absorbed in the atmosphere, and partly absorbed at the earth's surface. This partitioning of the solar energy incident at the top of the atmosphere (TOA), hereafter called solar energy disposition (SED), is determined by the optical properties of the atmospheric column, which, in turn, is influenced by the SED. The key variables of the column that control SED include those associated with the amount, vertical distribution, and optical properties of clouds, moisture and aerosols, as well as surface optical properties. Feedback involving these variables and the SED is important in modeling the climate system response to external perturbations, such as changes in the concentrations of CO_2 and other greenhouse gases. At this point, cloud feedback is the principal contributor to the large uncertainty in climate system response (Cess et al. 1989; Arking 1991).

Not only does the SED play an active role in the energetics of the climate system, it is also closely linked to the hydrologic cycle via dynamic and thermodynamic processes (Randall et al. 1989; Stephens and Greenwald 1991; Wielicki et al. 1995). About half the solar energy absorbed at the surface is used to evaporate water, which eventually forms clouds. Latent heat released in cloud formation is a major source of energy driving the atmospheric circulation, especially in the Tropics, and is comparable in magnitude to the solar radiation directly absorbed by the atmosphere. A sensitivity study with a general circulation model (GCM) shows that modifying the parti-

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tioning of solar energy between the atmosphere and surface could substantially alter the modeled fields of cloud cover, temperature, precipitation, humidity, and atmospheric circulation pattern (Kiehl et al. 1995). Understanding the earth's climate and the ability to model it, therefore, require an accurate representation of the radiation energy budget at the TOA (Hartmann et al. 1986; Ramanathan 1987; Stowe 1988) and at the surface (Suttles and Ohring 1986; Wielicki et al. 1995). Together they determine how much of the solar energy is absorbed in the atmosphere.

The earth radiation budget (ERB) in the atmosphere–surface system has been monitored from space for more than two decades, while the surface radiation budget (SRB) has been observed at various sites for more than a century. Both ERB and SRB observations have limitations on their accuracy that make it difficult to obtain a reliable estimate of the energy absorbed in the atmosphere, since the latter is the difference between two large quantities. ERB is measured globally, usually with the same sensor, but the radiation measured from space requires corrections for the spectral sensitivities of the sensors and the angular and diurnal variations in the radiance reaching the satellite. SRB measurements collected from surface sites suffer from an inability to maintain uniform deployment standards and ensure proper calibration amongst the various instruments that are used, in addition to a severe spatial sampling problem. A relatively recent approach to monitoring the surface radiation budget takes care of the sampling problem by using satellite radiance measurements to infer the fluxes at the surface, but it requires the use of radiative transfer models (Schmetz 1989; Pinker et al. 1995).

At present time, knowledge of ERB is far more advanced than that of SRB. The global, annual mean solar flux incident at the TOA is about 1365 W m^{-2}, and its accuracy and year-to-year variability is less than a few tenths of a percent. The fraction reflected to space (albedo) is around 0.30. Its accuracy and year-to-year variability is estimated to be 0.01. While many GCMs show good agreement with observations at the TOA, their surface values tend to be higher than observations (Garratt 1994; Wild et al. 1995; Barker and Li 1995; Ward 1995). For example, the global and annual mean fluxes absorbed at the surface are generally larger and smaller than 170 W m^{-2}, respectively, for the models and observations under study. Global and annual mean atmospheric absorption ranges from 0.16 to 0.29 among these datasets, equivalent to a flux difference of 45 W m^{-2}. A discrepancy of similar magnitude was also found under clear-sky conditions among a large number of GCMs (Randall et al. 1992).

Therefore, two critical issues need to be addressed. First, it is necessary to narrow the large gap in our knowledge of the partitioning of the solar energy between the atmosphere and surface. This requires better and more consistent observations. Second, it is necessary to determine why there is a discrepancy between models and observations, if a discrepancy remains after the observations are better established. Related to the second issue is the role of clouds in atmospheric absorption, which is currently a topic of considerable contention. The debate was ignited by recent studies claiming that solar radiation absorbed by clouds has been substantially underestimated (Cess et al. 1995; Ramanathan et al. 1995; Pilewskie and Valero 1995). However, these results and the methods on which they are based have been challenged (Chou et al. 1995; Li et al. 1995a; Stephens 1995; Arking et al. 1996; Ackerman and Toon 1996; Li and Moreau 1996; Imre et al. 1996; Arking 1996). The amount of the claimed underestimation by clouds is of the order of 25 W m^{-2}, comparable to the average discrepancy between models and observations (Garratt 1994; Wild et al. 1995). Other comparisons between models and observations show a discrepancy of similar magnitude in the clear-sky surface flux (Barker and Li 1995; Arking 1996). The question of whether the discrepancy in atmospheric absorption between models and observations is due primarily to clouds or to clear-sky absorption is important. Almost universally among models, the effect of clouds on atmospheric absorption, when globally averaged, is quite small, as shown in this study. Having a small effect on atmospheric absorption, however, does not imply that clouds absorb little solar radiation, only that for whatever absorption occurs, the bulk of it is in place of clear-sky absorption (Stephens 1996). As a result, clouds can alter the profile of atmospheric heating. While the vertical distribution of atmospheric heating is important, this study is confined to vertically integrated absorption, on which new observational data can be brought to bear. Thus, the second issue concerns the net effect of clouds on atmospheric absorption of solar radiation, whether it is small, as in the models, or large, as found by some investigators (Cess et al. 1995; Ramanathan et al. 1995; Pilewskie and Valero 1995).

Fully resolving the two issues is a tremendous undertaking, especially with regard to the spatial and temporal variation of SED. This study attempts to shed light on the two issues from the perspective of comparison between four observational datasets, one using ground-based and three using satellite-based estimates of SRB, and four GCMs. Comparisons among these datasets and estimates of data uncertainties help us see what are the common features among the models and where they differ, and where models show consistent differences with respect to the observations. The study is limited to global and zonal mean comparisons, and is intended to serve both as an overview and also to present many unreported results. A brief historical perspective on the development of SED is given in the next section. Section 3 describes the four observational and four model datasets. Global and zonal comparisons are presented in sections 4 and 5, respectively, and a summary in section 6.

2. Historical perspective

Prior to the space-borne earth observation era inaugurated in the 1960s, SED estimates were based solely on surface measurements. Simple models of radiative transfer in the atmosphere were used to infer TOA fluxes from the surface measurements. Surface radiation is among the few meteorological variables that have been observed since the last century (Hunt et al. 1986). On the basis of very limited observations at different latitudes, Abbot and Fowle (1908) obtained the first estimate of the global annual mean planetary albedo, 0.37, and near-surface (below 1800 m) absorption, 0.42 (all numbers are normalized to the incoming solar flux at the TOA). Similar estimates of SED were obtained by investigators in the 1920s and 1930s (cf. Table 3.2 of Budyko 1982). Spatial and temporal variations in SED were first addressed by Simpson (1929). More extensive analyses were made in the middle of this century (Liou 1980; Houghton 1954; Budyko 1956; London 1957) based on increased surface observations, more sophisticated radiative transfer theory, and the beginning of laboratory studies. Houghton (1954) estimated the fraction of reflection to space and absorption by the surface and by the atmosphere to be 0.34, 0.47, and 0.19, respectively. London (1957) obtained similar values and estimated the following contribution of various components of the vertical column: TOA reflection 0.35 (of which 0.07, 0.24, and 0.04 are due to air molecules, clouds, and surface, respectively), atmospheric absorption 0.175 (0.16 due to atmospheric constituents and 0.015 due to clouds), and surface absorption 0.475. All of these estimates apply to the Northern Hemisphere, where most surface measurements were made. Sasamori et al. (1972) computed the SED for the Southern Hemisphere (0.35, 0.45, and 0.20). The most extensive and complete compilations of the global surface energy balance (SEB) were carried out by Budyko (1982) and his colleagues. They generated several versions of an SEB atlas depicting the monthly mean global distribution of various SEB components, including SRB. Empirical relationships involving conventionally measured meteorological variables (e.g., cloud amount, sunshine duration, etc.) were used. With improving techniques and a growing set of observations, their estimates of solar flux absorbed at the surface increased (Budyko 1982). Their latest estimates of the SED are TOA albedo of 0.30, surface absorption of 0.46, and atmospheric absorption of 0.24, which are identical to the satellite-based estimates of Li and Leighton (1993). However, the most recent ground-based estimate of surface absorption by Ohmura and Gilgen (1993) is as little as 0.42, coincident with the earliest estimate of Abbot and Fowle (1908).

Since 1960, meteorological satellites have contributed to a radical improvement in our knowledge of ERB (House et al. 1986). In contrast to ground-based observation, space-borne observation has the advantages of global and uniform coverage. From the space-borne radiometers of the first [Television Infrared Observation Satellites (TIROS) type] and second generations (*Nimbus-3*, Environmental Science Services Administration, and National Oceanic and Atmospheric Administration series), a global mean planetary albedo was found to be around 0.30 (Vonder Haar and Suomi 1971; Stephens et al. 1981; Gruber et al. 1983). This number is significantly lower than the presatellite estimates but is in fairly good agreement with the later observations by more advanced sensors (Hartmann et al. 1986; Ramanathan 1987; Barkstrom et al. 1989). The geographical distributions of the TOA albedo for the four seasons were obtained by Raschke et al. (1973). These early estimates of regional radiative fluxes contain large uncertainties due in part to the crude treatment of the dependence of satellite radiance measurements on viewing geometry (Arking and Levine 1967; Raschke et al. 1973). More meticulous monitoring of the spatial and temporal variations in TOA albedo was accomplished by the radiometers of the third generation, including the ERB sensors aboard *Nimbus-7* (Jacobowitz et al. 1984) and the Earth Radiation Budget Experiment (ERBE) sensors aboard three satellites (Barkstrom et al. 1989). One of the major advances is the development of improved angular dependence models (Taylor and Stowe 1984; Suttles et al. 1988). Nevertheless, angular correction is still the primary source of uncertainty in ERB measurements (Arking and Vemury 1984; Stuhlmann and Raschke 1987; Suttles et al. 1992; Wielicki et al. 1995; Li 1996).

Since satellites measure only the radiative fluxes that exit the entire atmosphere–surface system, surface and atmospheric radiation budgets cannot be directly determined. Considerable success has been achieved in the retrieval of solar SRB from ERB measurements, as reviewed by Schmetz (1989) and Pinker et al. (1995). Tens of retrieving algorithms have been proposed, which are of three types: empirical relationships (Fritz et al. 1964; Tarpley 1979; etc.), parameterized schemes (Gauthier et al. 1980; Chou 1989; Cess et al. 1991; Li et al. 1993; etc.) and full radiative transfer models (Möser and Raschke 1983; Pinker and Ewing 1985; Stuhlmann et al. 1990; Bishop and Rossow 1991; etc). The first satellite-based evaluation of SED was made by Hanson et al. (1967) over the United States for the spring of 1962. Multiple years of global data on SED are now available from both operational meteorological satellites (Pinker and Laszlo 1992; Darnell et al. 1992; Rossow and Zhang 1995) and experimental radiation satellites (Li and Leighton 1993; Breon et al. 1994). Global mean surface absorptance estimated from these satellite observations ranges from 0.46 to 0.50. For a planetary albedo of 0.30, global mean atmospheric absorptance therefore varies from 0.20 to 0.24.

In GCMs, SED is generally computed by a simplified radiative transfer model (RTM) with input parameters provided by the GCM. Since GCMs generally do not reproduce cloud properties well, and since clouds are the most important factor in determining the SED, the SED from a GCM is usually not reliable. However, the majority of GCMs may have been tuned to produce "sound" values deemed by modelers for such highly averaged quantities as global and annual mean SED. Regardless, the modeled SED can help us to understand feedback processes and to evaluate and improve the performance of a GCM. To evaluate GCM performance, we need not only reliable observations of the SED, but also the variables that influence the SED. SED is mainly modified by cloud (fractional cover, thickness, height, microphysical parameters), water vapor (amount and vertical distribution), aerosols (amount, vertical profile, size distribution, and optical properties), and surface albedo (including its spectral and angular dependencies). To date, many of these variables can be derived from satellite observations. For example, extensive cloud information is available from the International Satellite Cloud Climatology Program (ISCCP) (Rossow and Schiffer 1991). Vertically integrated precipitable water (Liu et al. 1992) and cloud water amounts have been retrieved from both infrared and microwave sensors (Lin and Rossow 1994; Greenwald et al. 1993; Liu and Curry 1993; Weng and Grody 1994). Aerosol optical thickness over oceans has been inferred from the Advanced Very High Resolution Radiometer (Rao et al. 1989). Global surface albedo data have been developed from the TOA clear-sky measurements (Staylor and Wilber 1990; Li and Garand 1994). Having these values, one is able to interpret the difference between modeled and observed SED in terms of the treatment of various physical processes and radiative transfer algorithms (Barker et al. 1994; Barker and Li 1995; Kiehl et al. 1994; Wild et al. 1995; Ward 1995; among others). The common finding of the comparisons is that modeled global planetary albedo agrees reasonably well with satellite observations, but the partition between the atmosphere and the surface differs markedly. In a word, the existing knowledge of SED is inadequate. A critical examination of the various estimates is thus long overdue.

3. Data

Four sets of observations and the output of four GCMs are compared in this study. One of the observational datasets uses measurements of surface insolation from the worldwide pyranometer network (Ohmura and Gilgen 1991), known as the Global Energy Balance Archive (GEBA), along with TOA measurements from ERBE (Barkstrom et al. 1989). The other three observational datasets are entirely satellite-based, with surface fluxes derived from ISCCP, using the algorithms of Pinker and Laszlo (1992) and Rossow and Zhang (1995), and from ERBE using the algorithm of Li et al. (1993). Hereafter, they are simply referred to as ISCCP/Pinker, ISCCP/Rossow, and ERBE/Li.

The four models include the Canadian Climate Centre's GCM (CCC/GCM2), the Colorado State University GCM (CSU/GCM), the National Center for Atmospheric Research's Community Climate Model (NCAR/CCM2), and the National Aeronatics and Space Administration's Goddard Earth Observation System (NASA/GEOS-1). The results of the first three models were taken from the control runs for the Atmospheric Model Intercomparison Project (AMIP), which provided observed monthly mean SST and sea-ice extent from January 1979 to December 1988 (Gates 1992). NASA/GEOS-1 was run in a data assimilation mode for the period March 1985–February 1990, with input from observed pressure heights (essentially, mean layer temperatures), humidity, winds, and sea level pressure from satellite, balloon-borne, and ground-based measurements (Schubert et al. 1993). These model datasets were selected partly because of their availability and partly because detailed comparisons against surface observations and/or satellite estimation have been conducted. No regard was paid to SED values in making the selections, and thus the datasets are considered typical.

a. Observational datasets

GEBA is a database containing about 150 000 station months of data collected at up to 1600 surface sites (Ohmura and Gilgen 1991). The main source of the radiation data is the World Radiation Data Center at St. Petersburg, Russia, where surface radiation measurements from the world radiation network are gathered. GEBA data are also selected from periodicals, monographs, data reports, and unpublished data. After rigorous quality tests, monthly mean fluxes are computed and archived. These data, together with empirical relationships based on standard meteorological data (cloud amount, sunshine duration, etc.), were employed by Ohmura and Gilgen (1993) in a reevaluation of the global SRB. Due to the poor spatial sampling of the surface albedo measurements, they estimated surface albedo using digitized land-use information, monthly mean snow and ice data, cloud cover data, and a limited number of albedo measurements for some typical surface types. Since the GEBA dataset does not compile separate averages for clear-sky conditions, it does not yield information on the effects of clouds on surface flux and, hence, on atmospheric absorption. We combine the estimate of SRB made by Ohmura and Gilgen (1993) based on actual GEBA measurements and empirical calculations with ERBE TOA data to determine atmospheric absorption. The combined dataset is designated as ERBE/Ohmura for simplicity. ERBE is a dataset of satellite measurement of TOA fluxes and the data used here cover the period January 1985–December 1989.

The ISCCP/Pinker dataset (version 1.1) covers the period March 1985–November 1988 (Whitlock et al. 1995). Atmospheric transmittance was calculated from cloud attributes (primarily amount and thickness), water vapor, ozone, aerosol, surface albedo, and snow/ice cover, using tabulated results of delta-Eddington radiative transfer calculations (Pinker and Laszlo 1992). The cloud optical thickness used in ISCCP/Pinker was not taken directly from the ISCCP output, but rederived from ISCCP radiances. Many other input parameters were taken from the ISCCP dataset, which includes analyses from a suite of operational weather satellites (Rossow and Schiffer 1991). The quality of the SRB data was evaluated by comparison against GEBA surface observations (Li et al. 1995b; Whitlock et al. 1995) and an independent dataset (ERBE/Li). Relative to GEBA, the majority of the regional estimates of net surface flux (downward positive) are accurate to within ± 20 W m^{-2}, with an overall bias of 10 W m^{-2}. Large errors occur over polar and desert areas due to inadequate spectral and angular corrections of satellite radiances and larger errors in precipitable water retrieved from the TIROS Operational Vertical Sounder (TOVS) (Li 1995). The positive bias is mainly due to the use of the Lacis and Hansen (1974) parameterization of shortwave radiative transfer, which underestimates water vapor absorption relative to a line-by-line calculation (Ramaswamy and Freidenreich 1992; Li 1995).

The ISCCP/Rossow dataset was derived using a modified version of the radiative transfer code of the Goddard Institute for Space Sciences GCM (Zhang et al. 1995; Hansen et al. 1983). Although ISCCP/Pinker and ISCCP/Rossow employ the same input dataset, their TOA fluxes are different. The TOA fluxes of ISCCP/Pinker were obtained from ISCCP radiance measurements with angular and spectral corrections, while those of ISCCP/Rossow were computed with their radiative transfer model using a large number of input parameters, including cloud optical thickness provided by ISCCP. At the time of writing, ISCCP/Rossow data were available from April 1985 to January 1989 at a resolution of 280 km in space and 3 h in time for every third month. The retrieved surface downwelling fluxes were compared with observations from both field experiments, such as the First ISCCP Regional Experiment/Surface Radiation Budget and the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment, and operational observation networks, such as those in GEBA. The comparisons show moderate positive biases of 10–20 W m^{-2} (Rossow and Zhang 1995).

The ERBE/Li dataset covers the period from 1985 to 1989 with a spatial resolution of 2.5° in latitude and longitude (Li and Leighton 1993). The inversion algorithm employed is a parameterization developed from extensive radiative transfer modeling. It involves fewer input and output parameters and much less computation than the ISCCP/Pinker and ISCCP/Rossow algorithms. The major input parameter is the TOA irradiance or albedo converted from the broadband ERBE radiances that were calibrated on board the satellites, compared to the narrowband ISCCP radiances that require postflight calibration. The quality of ERBE/Li data was assessed by comparison with the GEBA (Li et al. 1995b) and ISCCP/Pinker (Li 1995) datasets. The overall comparison against GEBA shows no bias and a standard difference of about 25 W m^{-2}, which is attributed mainly to inadequate sampling of surface measurements (Li et al. 1995b). However, appreciable regional errors exist in some tropical land areas when biomass burning is widespread (Z. Li 1997, manuscript submitted to *J. Climate*). Potentially large errors may also occur in the polar regions resulting from extremely low water vapor and unreliable angular correction (Li 1996) and scene identification (Li and Leighton 1991) for the ERBE measurements.

b. Model datasets

The CCC/GCM2 output data were produced for AMIP and are described by McFarlane et al. (1992). The model computes solar radiative fluxes with a two-spectral interval version of Fouquart and Bonnel's (1980) algorithm, where the solar spectrum is split at 0.7 μm. Extensive assessments of the radiative characteristics of the model were conducted by Barker et al. (1994) and Barker and Li (1995). A comparison against the ISCCP cloud climatology, ERBE TOA radiation budget, and ERBE-based surface albedos reveals several deficiencies in the model's radiative transfer scheme (Barker et al. 1994). Of consequence to the zonal mean analysis are an under- (over) estimation of ocean albedo at high (low) latitudes, a too dry atmosphere, too many high clouds in the Tropics, and too few low clouds in the extratropical storm-track regions. Relative to ERBE/Li, CCC/GCM2 systematically under- (over) estimates atmospheric (surface) absorption by 18 W m^{-2}, much of which is attributed to inaccurate computation of water vapor absorption and neglect of aerosols (Barker and Li 1995).

The CSU/GCM output data were also produced in connection with AMIP, using a version of the model described by Fowler et al. (1996) and Fowler and Randall (1996). The model uses a bulk cloud microphysics scheme, encompassing five prognostic variables that relate hydrologic processes to radiative processes via parameterizations. The radiative transfer scheme is described by Harshvardhan et al. (1987), which, for solar radiation, is based on Lacis and Hansen (1974) for computation of clear-sky absorption and scattering, and on the delta-Eddington approximation (Joseph et al. 1976) for radiative transfer in cloudy layers. Comparisons of model output against SSM/I water vapor and cloud water data, ISCCP cloud data, and ERBE radiation data reveal numerous shortcomings despite considerable improvements over the original version. Notably, cloud (especially high cloud) amounts were overestimated, leading to a too-strong shortwave cloud radiative forcing (CRF) at the surface (defined as the net flux at the surface averaged under all-sky conditions minus that averaged under clear-sky conditions).

The NCAR/CCM2 is generally described by Hack et al. (1993), and its radiative transfer scheme is described by Briegleb (1992). There are 18 spectral intervals in the shortwave region (0.2–5.0 μm), and atmospheric absorption due to water vapor, ozone, carbon dioxide, and oxygen are calculated using parameterizations. The delta-Eddington approximation is applied to the optical properties of cloud droplets obtained from the parameterization of Slingo (1989). A comparison of the TOA radiation budget against ERBE shows an overall good agreement, but a considerable discrepancy in the Northern Hemisphere summer, where the shortwave CRF is underestimated (Kiehl et al. 1994). The discrepancy is attributed to the use of too-large cloud droplets over land (Kiehl 1994) and underestimation of cloud amount and cloud optical thickness (Ward 1995). These differences could lead to an overestimation of the surface shortwave flux in the northern summer midlatitudes by as much as 100 W m^{-2} (Ward 1995).

The NASA/GEOS-1 dataset is the output of a reanalysis of observational data for the period March 1985–February 1990, using the NASA/GEOS-1 data assimilation system, which consists of an atmospheric GCM and a three-dimensional multivariate optimal interpolation scheme (Schubert et al. 1993). The observational data come from meteorological measurements made at the surface and from radiosondes, aircraft, ships, and satellites, and include pressure heights (essentially temperature), humidity, winds, and sea level pressure. Cloud variables and radiative terms are prognostic variables computed by the model, and the radiative transfer scheme is the same as that in the CSU/GCM (Harshvardhan et al. 1987). As with other GCMs, NASA/GEOS-1 produces too much cloud cover over the deep convective tropical regions, and too little over the midlatitude storm tracks (Schubert and Rood 1995).

4. Comparison of global annual means

The global annual mean SED for the eight datasets is shown in Table 1, in terms of both absolute values and fractions of the TOA incident solar flux. As one might expect, the agreement at TOA is much better than at the surface. The TOA net flux ranges from 94.8 to 111.5 W m^{-2}, corresponding to a planetary albedo interval of 28.7%–32.6%. The ERBE values are the best available estimates at this point, but the span of the planetary albedos listed in Table 1 is within the range of values obtained from various satellite experiments (cf. Table 2 of Rossow and Zhang 1995). In contrast, atmospheric and surface absorption show considerable variation among the datasets. The flux absorbed in the atmosphere ranges from 56 W m^{-2} (NASA/GEOS-1) to 98 W m^{-2} (ERBE/Ohmura), corresponding to 16.2% and 28.7% atmospheric absorptance. Likewise, the surface-absorbed flux ranges from 142 W m^{-2} (42%) to 191 W m^{-2} (55%), a difference of 49 W m^{-2} (13%), which exceeds the range of historical estimates. Even if the two extreme values are discarded, the maximum difference among the remaining fluxes absorbed in the atmosphere is still as large as 24 W m^{-2}. At this point, therefore, we do not have a reliable estimate of how the net energy absorbed by the climate system is partitioned between the atmosphere and surface. Of interest here is that atmospheric absorption in the GCMs is smaller than the satellite estimates, which, in turn, are smaller than the GEBA surface estimate.

It seems to be a paradox that the global annual mean surface net solar flux based on ERBE/Ohmura derived from GEBA (Ohmura and Gilgen 1993) is substantially less than that from ERBE/Li (Li and Leighton 1993), whereas the overall comparison between ERBE/Li and GEBA data shows a bias error near zero (Li et al. 1995b). There are two potential reasons for this. First, direct measurements of surface insolation are only available at a limited number of stations, usually located in populated regions. Over the vast areas of remote land and oceans, there are almost no observations and thus Ohmura and Gilgen (1993) resorted to empirical relationships to infer insolation from conventional meteorological observations. Since such relationships depend on location and season, substitution of the relationships developed for regions with observations to regions lacking observations could yield unreliable estimates of surface insolation. For the same reason, there are unknown uncertainties in the satellite-based estimates over regions where there are no ground-truth observations. These uncertainties are, however, conceived to be smaller than those aris-

TABLE 1. Global annual mean solar energy disposition under all-sky conditions. Both absolute values (W m^{-2}), and relative values (in the parentheses) are given.

Sources	ERBE/ Ohmura	ERBE/ Li	ISCCP/ Rossow	ISCCP/ Pinker	CSU/ GCM	CCC/ GCM2	NCAR/ CCM2	NASA/ GEOS-1
Solar constant	1365	1365	1366	1357	1365	1365	1370	1380
Reflected to space	101.3 (29.6)	101.3 (29.6)	111.5 (32.6)	99.5 (29.3)	110.3 (32.3)	108.6 (31.8)	94.8 (27.6)	98.4 (28.5)
Absorbed in atmosphere	98.0 (28.7)	83.1 (24.4)	65.0 (19.0)	68.6 (20.2)	60.9 (17.8)	57.4 (16.8)	67.6 (19.7)	56.0 (16.2)
Absorbed at surface	142.0 (41.7)	157.0 (46.1)	165.1 (48.3)	171.1 (50.4)	170.2 (49.9)	175.0 (51.3)	180.6 (52.6)	190.6 (55.2)

ing from the empirical calculations. Second, the values of surface albedo used are different. Conventional measurements of surface albedo have very poor spatial representation. In order to obtain surface net flux from insolation, a surface albedo dataset compiled from multiple ground-based sources was employed by Ohmura and Gilgen (1993). Likewise, as the ERBE/Li data do not contain surface downwelling fluxes, a comparison with GEBA entails surface albedos that were estimated from ERBE (Darnell et al. 1992).

To avoid the uncertainties associated with the use of empirical relationships for computing SRB, Arking (1996) reanalyzed the SED based exclusively on actual measurements from ERBE and GEBA. His global and annual results are identical to those from ERBE/Li. While this bolsters the credibility of the satellite-based product, we cannot claim that the estimates of ERBE/Li are totally correct because of the limited number and skewed distribution of surface observation. In particular, large errors were found for all the satellite-based estimates of SRB over regions with heavy loading of strong absorbing aerosols such as those produced from biomass burning (Z. Li 1997, manuscript submitted to *J. Climate*). Comparisons between satellite-retrieved and surface-observed SRB over different latitude zones reveal a significant overestimation by the three satellite approaches in tropical regions with abundant biomass burning. Since the overestimation occurs in dry season only over a small portion of the tropical land, its impact on global and annual mean SRB is small, less than 2 W m^{-2} (Z. Li 1996, manuscript submitted to *J. Climate*). Even smaller is its impact on the TOA radiation budget (Chylek and Wang 1995).

Table 2 lists the global annual mean values of SED for clear skies. Unfortunately, clear-sky data are not available from GEBA. Just like all-sky values, the consistency among data from different sources for clear-sky condition is better for TOA reflection than for atmospheric and surface absorption. Maximum discrepancy in the atmospheric- (surface) absorbed flux is 30 W m^{-2}, three times that in the TOA reflected flux (10 W m^{-2}). Also similar to the all-sky condition, atmospheric (surface) absorption simulated by GCMs is systematically weaker (stronger) than absorption inferred from satellites. The TOA-reflected fluxes or albedos simulated by GCMs are slightly smaller than those from satellite observations.

Comparison between Tables 1 and 2 reveals that clear-sky reflection is about one-half of all-sky reflection, and clear-sky atmospheric absorption is about the same as all-sky absorption. Global and annual mean values of the shortwave CRF are shown in Table 3. The agreement in shortwave CRF is somewhat better than for all-sky or clear-sky flux, especially for atmospheric absorption, implying that the impact of clouds on atmospheric absorption is similar in the eight datasets. Overall, therefore, the presence of clouds does not significantly alter atmospheric absorption, which is in agreement with some recent analyses of observational data (Li et al. 1995a; Imre et al. 1996; Arking 1996). This was attributed by Ramanathan et al. (1996) to assuming homogeneous and clean clouds (no aerosols) together with an arbitrary cutoff in wavelength for computing Lorertzian line absorption in radiation models. But we believe that the near-zero effect is caused largely by the overlapping of absorption bands due to water vapor and cloud droplets (Davies et al. 1984).

Surface shortwave CRF differs considerably among the models (differences as much as 22 W m^{-2}), but is more consistent among the satellite-based prod-

TABLE 2. Same as Table 1 but for clear-sky conditions.

Sources	ERBE/ Li	ISCCP/ Rossow	ISCCP/ Pinker	CSU/ GCM	CCC/ GCM2	NCAR/ CCM2	NASA/ GEOS-1
Reflected to space	52.7 (15.5)	57.8 (16.9)	53.6 (15.8)	47.1 (13.7)	49.5 (14.5)	48.2 (14.0)	50.2 (14.5)
Absorbed in atmosphere	79.1 (23.2)	66.6 (19.5)	65.2 (19.2)	57.7 (16.9)	50.6 (14.9)	66.7 (19.4)	59.7 (17.3)
Absorbed at surface	209.0 (61.3)	217.2 (63.6)	220.5 (65.0)	236.6 (69.3)	240.8 (70.6)	228.1 (66.5)	235.1 (68.1)

TABLE 3. Difference between Tables 1 and 2.

Sources	ERBE/ Li	ISCCP/ Rossow	ISCCP/ Pinker	CSU/ GCM	CCC/ GCM2	NCAR/ CCM2	NASA/ GEOS-1
Reflected to space	48.0 (14.1)	53.7 (15.7)	45.9 (13.5)	63.2 (18.5)	59.1 (17.3)	46.6 (13.6)	48.2 (14.0)
Absorbed in atmosphere	4.0 (1.2)	−1.6 (−0.5)	3.5 (1.0)	3.2 (1.0)	6.7 (2.0)	0.9 (0.3)	-3.7 (-1.1)
Absorbed at surface	−52.1 (−15.3)	−52.1 (−15.3)	−49.4 (−14.6)	−66.4 (−19.5)	−65.8 (−19.3)	−47.5 (−13.9)	-44.5 (-12.9)
CRF ratio*	1.09	0.97	1.08	1.05	1.11	1.02	0.92

*CRF ratio denotes cloud radiative forcing ratio defined as the ratio of surface CRF to TOA CRF.

ucts (a range of only 3 W m^{-2}). The ratio of surface CRF to TOA CRF is also shown in Table 3, which was used by Cess et al. (1995) and Ramanathan et al. (1995) to infer a cloud absorption anomaly. As they noted, the values of the ratio from radiation models used in both GCM and satellite retrieval algorithms are around 1.0, at variance with the 1.5 that they obtained at a few sites. Although the ratios are close, their corresponding surface/atmosphere absorbed fluxes are quite different. This suggests that the differences do not originate from the computation of radiative transfer in cloudy atmospheres, but in clear atmospheres. This conforms with our observational finding that the ratio is generally around 1.0 except for a small fraction of the cases, primarily in the Tropics, whose ratios are around 1.5 (Li et al. 1995a; Li and Moreau 1996). These large ratios were later proved to be an artifact resulting from the influence of biomass burning aerosol on the retrieval of clear-sky SRB (Z. Li 1996, manuscript submitted to *J. Climate*).

5. Comparison of zonal monthly means

Since the surface-based GEBA dataset is quite nonuniform in its geographic distribution, the vast majority of measurement sites being in midlatitude continents, zonal comparisons are made only between the global datasets from satellite and GCMs. Figure 1 presents clear-sky zonal mean solar fluxes reflected to space by the atmosphere–surface system. The agreement is generally within 10–20 W m^{-2} except for ISCCP/Pinker, which differs significantly from the others south of 40°S in January with deviations as large as 100 W m^{-2}. The large differences more likely stem from the erroneous cloud identification by the old ISCCP scene identification scheme. Although ERBE TOA fluxes are generally considered the most reliable among the various sources, ERBE clear-sky values in the polar region are not necessarily more accurate than others due to the unreliable identification of clear-sky pixels by the ERBE scene identification scheme. For example, the steeper augmentation of the ERBE value near 70°N is likely an artificial effect of the incorrect prescription of ice boundaries (Li and Leighton 1991). In addition, the angular dependence model employed by ERBE (Suttles et al. 1988) for converting radiance into irradiance suffers large errors over snow/ice scenes (Li 1996). The values of the three satellite products are generally closer to each other and slightly larger than those simulated by the three models.

As the clear-sky TOA-reflected flux is modified primarily by surface albedo, comparison of zonal mean surface albedo is presented in Fig. 2. It is seen that modeled surface albedos are lower than most satellite-based estimates outside polar regions, leading to too much reflection to space. The values of ISCCP/Pinker are lower than other satellite estimates and some modeled values, due presumably to an underestimation over bright scenes in arid and snow/ice-covered regions (Whitlock et al. 1995). In addition, the larger dispersion in TOA reflection over the summer polar regions (Fig. 1) is in line with the larger dispersion in surface albedo from various sources.

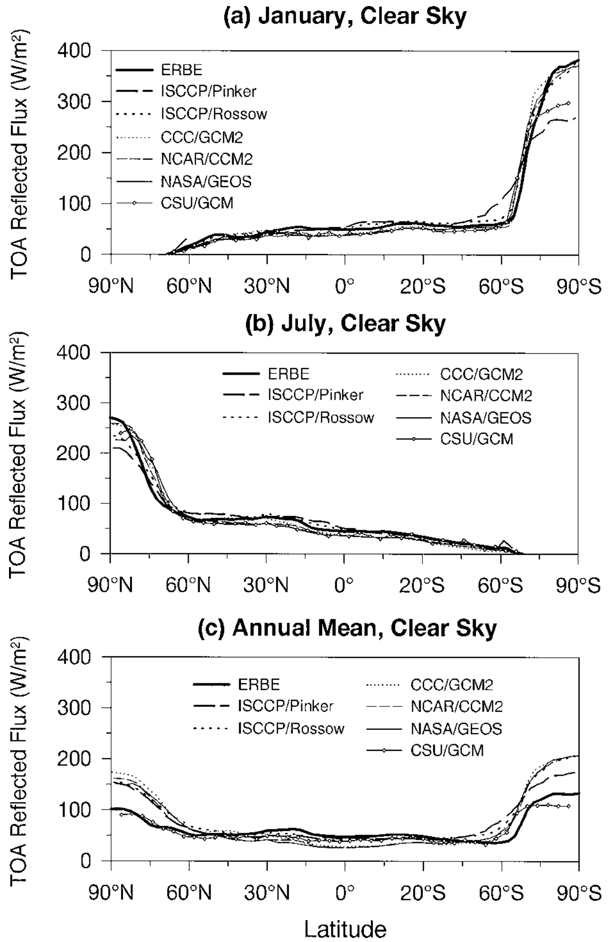

FIG. 1. Comparison of the zonal mean flux reflected to space at the top of the atmosphere under clear-sky conditions for (a) January, (b) July, and (c) annual mean.

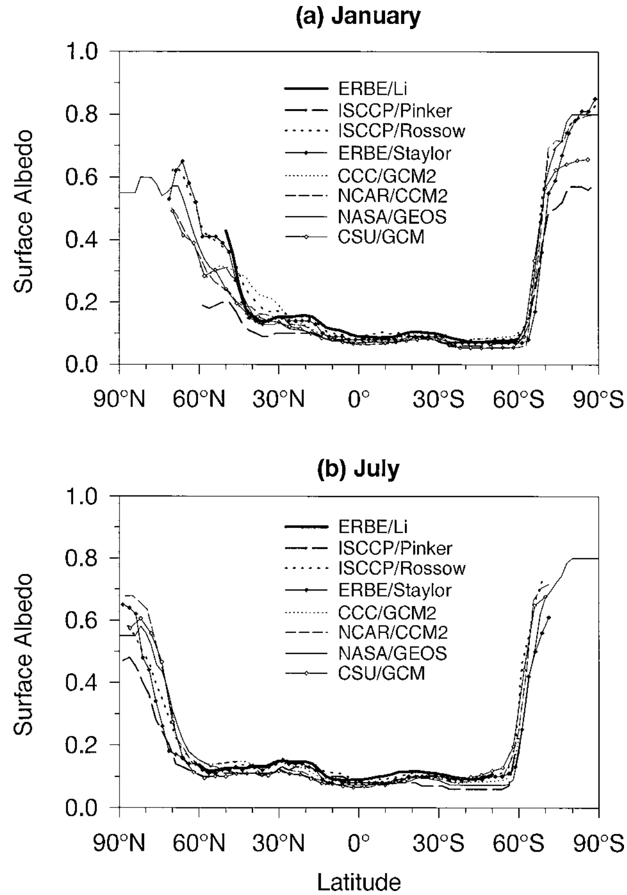

FIG. 2. Comparison of the zonal mean surface albedo for (a) January and (b) July.

Unfortunately, it is difficult to appraise the quality of these data. In polar regions, numerous problems are encountered in the determination of surface albedo by means of both remote sensing and model simulation. The frequent presence of extensive cloud cover, small radiometric contrast between clouds and bright snow/ice-covered surfaces limits the ability of remote sensing surface albedo from space. Model simulated surface albedos suffer even larger uncertainties because of overall poor performance in predicting snow cover and freezing/melting events by GCMs. Therefore, development of a more reliable dataset of surface albedo in the polar region is urgently needed to evaluate and improve the performance of GCMs in the polar regions. Another common problem with respect to the surface albedo occurs in boreal forest regions where many GCMs produce too high surface albedos with the presence of snow cover (Barker et al. 1994; Betts 1996, manuscript submitted to *J. Geophys. Res.*) Besides, some GCMs tend to have too low surface albedos over deserts relative to the satellite estimates (Barker et al. 1994).

The same comparison as in Fig. 1 but under all-sky conditions is shown in Fig. 3. Again, all satellite values do not differ very much. As a matter of fact, the agreement between ERBE and ISCCP/Pinker is even better for all sky than for clear sky, whereas ISCCP/Rossow deviates from ERBE more for all sky than for clear sky. Rossow and Zhang (1995) found that the difference increases linearly with cloud amount at a rate of 1 W m^{-2} per 10% difference in cloud amount. They attributed the dependence partially to the use of different angular dependence models by ERBE and ISCCP (Rossow and Zhang 1995). A more marked feature of Fig. 3 is that the model results disagree significantly with the satellite results. The former are substantially less than the latter in the summer midlatitudes, but moderately more in the Tropics. Since the values of their clear-sky counter-

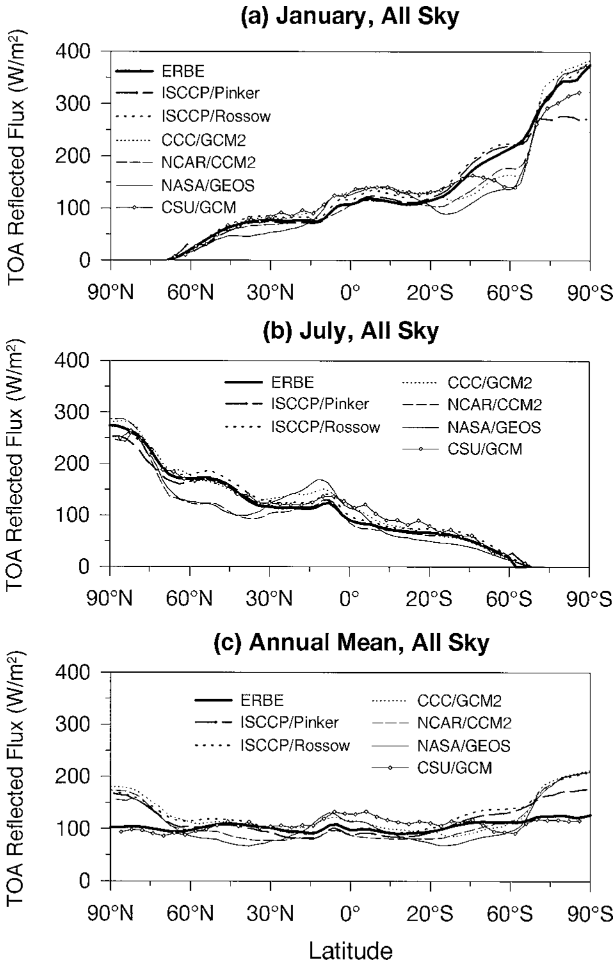

FIG. 3. Same as Fig. 1 but under all-sky conditions.

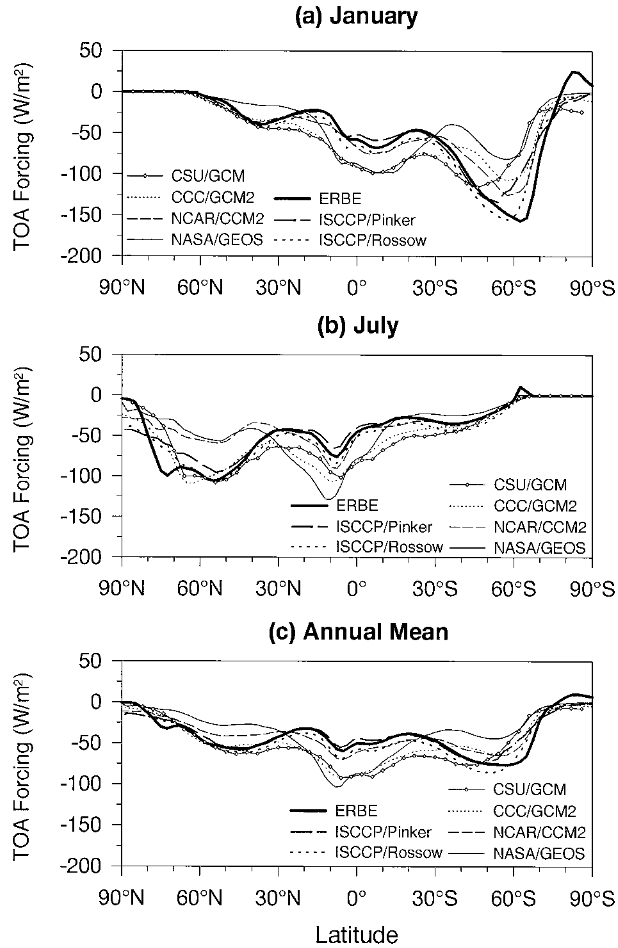

FIG. 4. Comparison of zonal mean cloud radiative forcing at the top of the atmosphere.

parts are similar (cf. Fig. 1), the discrepancies must result from incorrect simulation of clouds.

To gain further insight into the effect of clouds, comparison of the zonal mean TOA CRF is presented in Fig. 4. The magnitudes and discrepancies of the CRF reach maxima over the latitude zones controlled by two major cloud regimes, namely, the ITCZ in the Tropics and storm tracks in the midlatitudes. Good agreement is found among the satellite values, which, however, differ markedly from model values. The majority of the models generate too many clouds over the Tropics, leading to excessive CRF, and too few clouds in the midlatitudes, leading to insufficient CRF. This phenomenon is common to almost all GCMs participating in the AMIP (G. Potter 1996, personal communication), but the causes vary from one model to another. By comparing with the regional distributions of cloud amounts from ISCCP and TOA fluxes from ERBE, Barker et al. (1994) found that the deficiencies of the CCC/GCM2 mainly occur over oceans. The model tends to produce too much high-cloud cover over warm oceans (SST < ~25°C) and too little low-cloud cover over cool oceans (SST < ~25°C) due presumably to the use of an inadequate cloud convection scheme. Over land, agreement in cloud amount and CRF is much better. However, the opposite is true for NCAR/CCM2, which generally agrees better with satellite observations over oceans than over land. This is attributed partially to the assignment of a too-large radius for continental cloud droplets (Kiehl et al. 1994). Based on observational evidence (Han et al. 1994), Kiehl et al. (1994) reduced the cloud-effective radius from 10 μm to 5 μm over continental regions, which eliminates approximately half of the bias in CRF over land. The remaining difference over land was attributed to deficient liquid water content (Kiehl et al. 1994). Ward (1995) found that NCAR/CCM2 underestimates cloud cover in the midlatitude marine

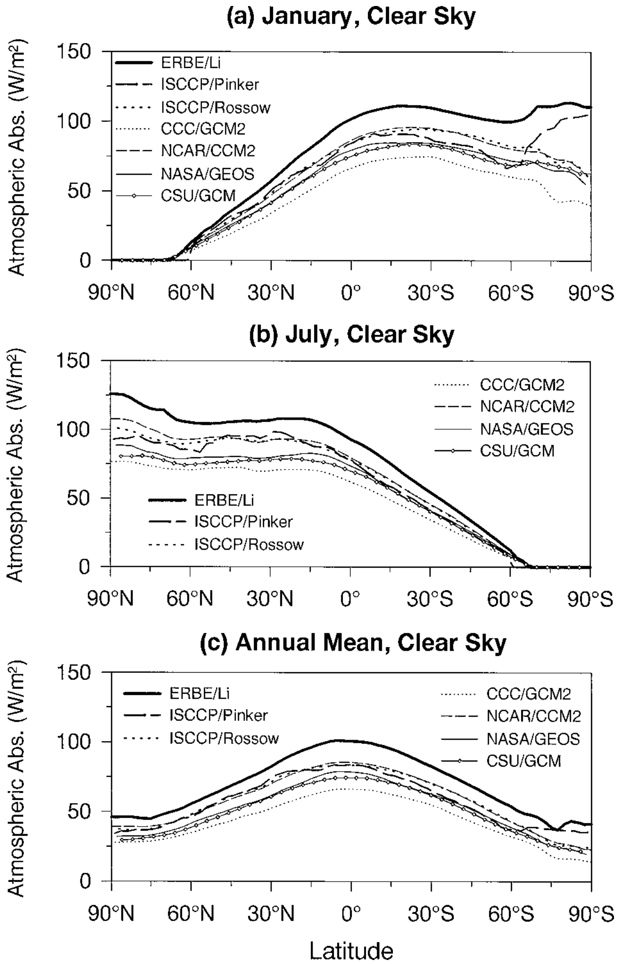

FIG. 5. Comparison of the zonal mean flux absorbed in the atmosphere under clear-sky conditions for (a) January, (b) July, and (c) annual mean.

regions, compared to the ISCCP cloud climatology. Unlike other GCMs, NCAR/CCM2 generates reasonable cloud cover in the Tropics. The stronger CRF of the CSU/GCM was explained, in part, by its inability to simulate partial cloud cover (Fowler and Randall 1996).

Figure 5 compares the fluxes absorbed in the atmosphere under clear-sky condition. The discrepancy is substantial, over 30 W m^{-2} in the summer hemisphere (a relative difference of up to 50%). GCM values are generally lower than satellite-based ones, except for NCAR/CCM2, which agrees well with ISCCP/Rossow. More amazingly, a large amount of the difference arises from a seemingly straightforward calculation for water vapor absorption. Table 4 presents the results of pure water vapor absorption computed by some conventional methods, including the Lacis-Hansen scheme (LH); LOWTRAN 5, 6, and 7 (L5, L6, and L7); the line-by-line method (LBL); and the median value of many schemes adopted in radiative transfer models involved in the intercomparison of radiation codes used in climate models (ICRCCM) (Fouquart et al. 1991). The difference between two widely used codes, L7 and LH, is as large as 30 W m^{-2} for a solar zenith angle of 30°. The benchmark value from LBL is in the middle. More important, the median value of ICRCCM is significantly lower than LBL, implying that the majority of models underestimate water vapor absorption. The actual amount of underestimation is even larger than that indicated by Table 4, since the enhanced absorption due to scattering is not accounted for in Table 4. From Fig. 5 and Table 4, it is inferred that water vapor absorption plays an important role in the discrepancies. The water vapor scheme used in NCAR/CCM2 compares well with LBL (Briegleb 1992), leading to the strongest absorption among the four models. Were aerosols ignored in ISCCP/Pinker, their estimates would be close to those of CSU/GCM and NASA/GEOS, as they all employ the LH scheme. This is seen clearly from the tendency for the atmospheric absorptance of ISCCP/Pinker to gradually approach the model results as aerosol loading decreases from the Tropics to high latitudes. In the polar region, atmospheric absorption escalates

TABLE 4. Solar atmospheric absorption by water vapor only (W m^{-2}) for the midlatitude summer atmosphere computed by different methods (surface albedo = 0.2). The results of Ramaswamy and Freidenreich (1992) and ICRCCM from Fouquart et al. (1991).

SZA	LBL	LH	L5	L6	L7	ICRCCM
30°	178.1	162.3	161.0	185.7	191.3	167.0
75°	71.4	63.6	59.5	73.3	74.4	64.2

SZA: Solar zenith angle; LBL: line-by-line; L: Lowtran; ICRCCM: intercomparison of the radiative codes used in climate models.

again because of the increased TOA daily insolation, multiple reflection between clouds and bright surfaces and extremely low water vapor content. A revised LH scheme was proposed by Ramaswamy and Freidenreich (1992) that renders water vapor absorption almost identical to LBL. If this revised LH scheme were employed, the estimates of ISCCP/Pinker would be very close to those of ERBE/Li (Li 1995). After the water vapor scheme is modified and a reasonable amount of aerosol is introduced, the GCMs under study are also expected to produce atmospheric absorption similar to that of ERBE/Li. To some degree, one may regard the difference between ISCCP/Pinker and CSU/GCM or NASA/GEOS as an approximate measure of the aerosol effect, and the difference between NCAR/CCM2 and CSU/GCM or NASA/GEOS as the errors introduced by using the LH scheme; the sum of the two errors are comparable, overall, to the differences with ERBE/Li. Of course, the differences shown in Fig. 5 are not exclusively due to these two factors. For example, a too dry atmosphere simulated by CCC/GCM2 is an additional factor causing too-weak absorption in the atmosphere by CCC/GCM2 (Barker and Li 1995).

The comparison of all-sky atmospheric absorption bears a strong resemblance to Fig. 5, as seen in Fig. 6. This is expected as the RTMs used in both GCMs and satellite-retrieving algorithms generate little extra atmospheric absorption by clouds. Figure 7 presents the comparison of zonal-mean atmospheric CRF, the difference in atmosphere-absorbed flux between all-sky and clear-sky conditions. It is generally less than 10 W m^{-2}, but diverges significantly among the various datasets. Most of the results show that clouds have a slight warming effect on atmospheric absorption, whereas those of NASA/GEOS and ISCCP/Rossow show otherwise. Whether a cloud has a cooling or warming effect depends on many factors, such as

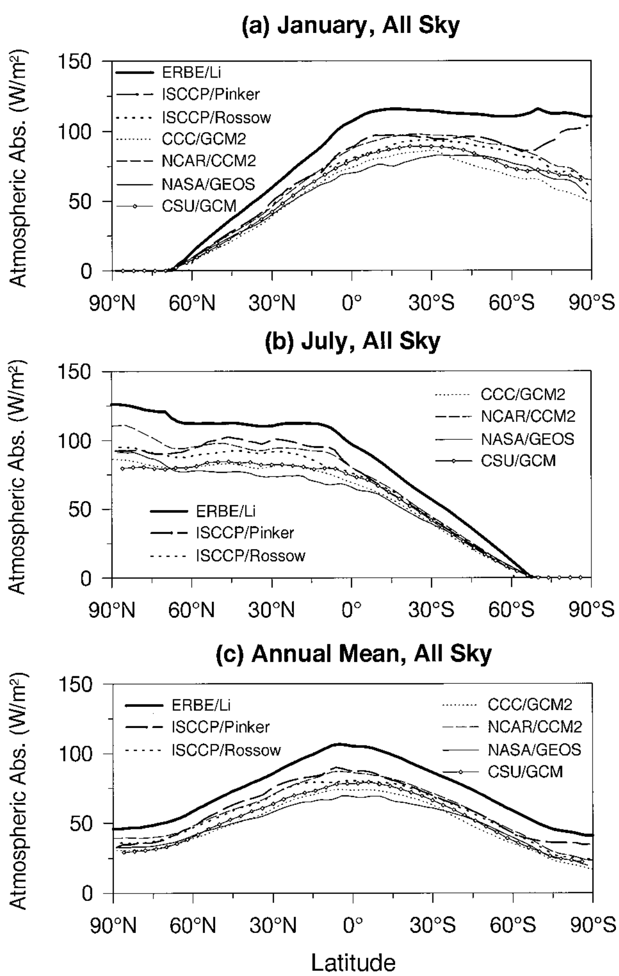

FIG. 6. Same as Fig. 5 but under all-sky conditions.

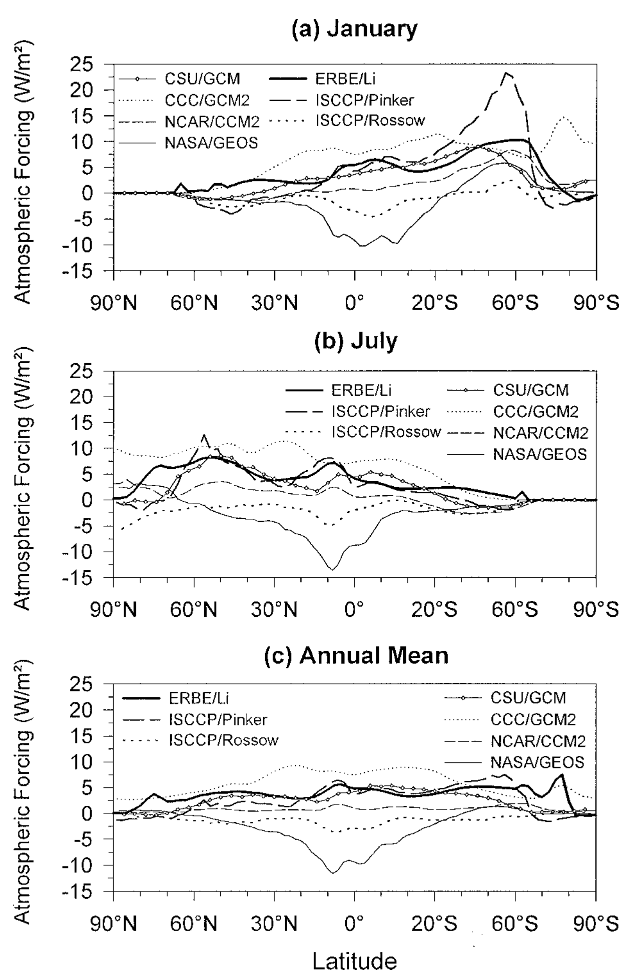

FIG. 7. Comparison of the zonal mean cloud radiative forcing in the atmosphere for (a) January, (b) July, and (c) annual mean.

cloud type and altitude, vertical distributions of absorbers (water vapor, aerosol), surface albedo, solar zenith angle, etc. (Chou et al. 1995; Li and Moreau 1996). Since the properties of the surface and clear atmosphere are relatively stable and should be similar among the datasets under study, discrepancies in Fig. 7 are most likely related to varying cloud conditions, especially cloud height. As cloud height increases, total atmospheric absorption decreases. This is due to backscattering by the cloud, which shields the solar photons from absorption by atmospheric column below the cloud. When the cloud is high enough, cloudy atmospheric absorption becomes even lower than clear-sky atmospheric absorption. Therefore, the extremely negative CRF of NASA/GEOS in the Tropics might indicate that the simulated tropical cloud altitudes are too high. This conjecture is consistent with the finding that the longwave CRF, which is most sensitive to cloud height, is overestimated over the Tropics (Schubert and Rood 1995).

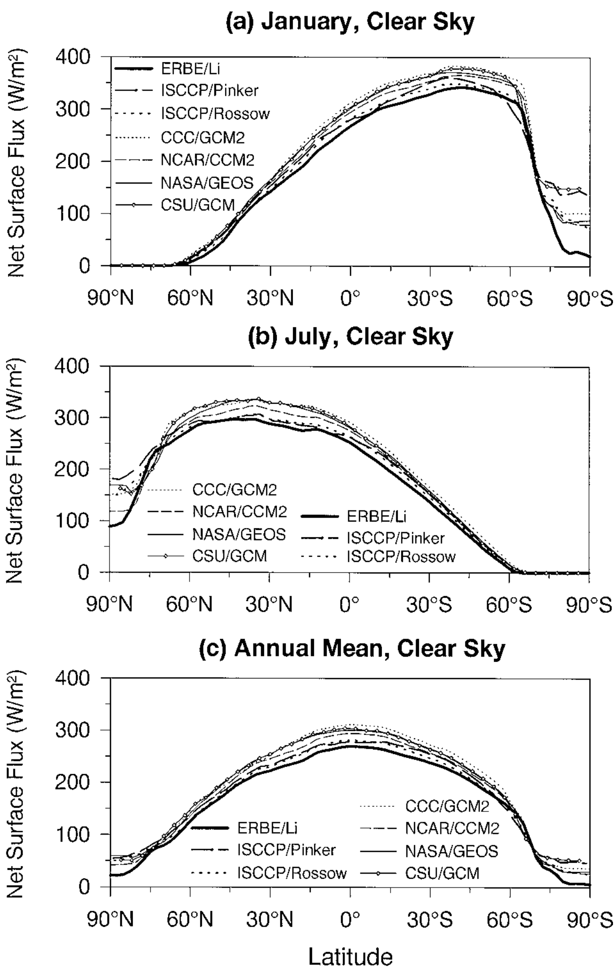

FIG. 8. Comparison of the zonal mean flux absorbed at the surface under clear-sky conditions for (a) January, (b) July, and (c) annual mean.

Since the sum of TOA reflection, atmospheric absorption, and surface absorption is equal to the solar flux incident at the TOA, the comparison of surface-absorbed flux depends entirely on the comparisons of TOA reflection and atmospheric absorption. Under clear-sky conditions (Fig. 8), the difference in surface-absorbed flux is dominated by the difference in atmospheric absorption. The contrast in surface net flux between models and satellites is slightly more striking than in atmospheric absorption. Satellite-based surface net solar fluxes are systematically and significantly higher than model simulations. The largest difference is between ERBE/Li and CCC/GCM2, which amounts to over 40 W m^{-2} in the Tropics. As mentioned earlier, a substantial amount of the discrepancy was accounted for by the deficiencies in the CCC/GCM2 identified by Barker and Li (1995). After several modifications, agreement improves significantly (to within 5 W m^{-2} over land). The close agreement between NASA/GEOS and CSU/GCM is expected as they used the same radiative transfer code (Harshvardhan et al. 1987). Ward (1995) investigated the differences between NCAR/CCM2 and ISCCP/Pinker. Neglect of aerosol in NCAR/CCM2 leads to the larger values of surface insolation, while possible cloud contamination of the ISCCP clear scenes may cause those of ISCCP/Pinker to be too small. In addition, oceanic albedos used in NCAR/CCM2 appear to be too low relative to the satellite-based estimates of ISCCP/Pinker and Li and Garand (1994).

The same comparison but for all-sky conditions is presented in Fig. 9. The differences in all-sky surface net flux are the superimposition of the systematic discrepancies in clear-sky surface flux and the regional (zonal) discrepancies in all-sky TOA flux associated with incorrect simulation of clouds by GCMs. As a result, relative to clear-sky values, the magnitudes of the differences are enlarged in the summer midlatitudes due to the underestimation of storm-track clouds by GCMs and lessened in the Tropics due to the overestimation of tropical clouds by GCMs, which counteracts the overestimation under clear-sky conditions.

6. Summary

Solar radiation is the driving force of the earth's climate system. Of the total radiation intercepted by the earth, part is reflected to space, part absorbed in the atmosphere, and part absorbed at the earth's surface. The state of the knowledge of the solar energy disposition can be traced back to the last century and can be divided into two distinct periods: presatellite and postsatellite, separated around the 1960s. In the first period, studies were focused on the surface radiation budget using a small number of surface measurements and empirical relationships between radiative fluxes and conventional meteorological parameters. With the aid of simple radiative transfer algorithms (often parameterizations), the earth radiation budget at the top of the atmosphere was inferred. At that time, knowledge on ERB was much worse than SRB. For example, the planetary albedo had been thought to be 0.35 and higher. Soon after the commencement of space-borne observations in the 1960s there began a dramatic advancement in our knowledge of ERB. The estimate of planetary albedo became more precise and stabilized at around 0.3. Regional and temporal variations in ERB were monitored systematically by the scanning radiometers aboard *Nimbus-7* and ERBE for over a decade, and surrogate ERB data (based on narrowband measurements) are now available from operational satellites over two decades. Although the advance in knowledge of ERB has fostered a renewed interest in SRB and the application of remote sensing techniques to monitoring it, our current knowledge on SRB has fallen behind ERB, and so has the radiation budget in the atmosphere (ARB).

The current state of knowledge on SED is examined here by comparing eight datasets: one based on surface observations that are extended globally using empirical relationships, three based on estimates from satellite measurements, and four based on estimates from GCM simulation. Comparisons were made for global and annual means, and zonal and monthly means, under both clear- and all-sky conditions. Overall, the agreement at the top of the atmosphere is much better than at the surface and in the atmosphere. Global and annual mean TOA albedos generally agree to within 0.02, whereas atmospheric absorptance differ by more than 0.1. In terms of the global and annual mean flux absorbed at the surface, the maximum difference is nearly 50 W m^{-2}. More important, surface fluxes computed by models are usually larger than ground-based observations and satellite-based estimates. Satellite- and ground-based values agree well in most circumstances, except for regions affected by strongly absorbing aerosols. Since such an effect is limited to a portion of the continental areas, it does not significantly alter zonal and global mean solar radiation budgets. The best estimates of global and annual mean fluxes absorbed at the surface and in the atmosphere and reflected to the space are 157 W m^{-2}, 83 W m^{-2}, and 101 W m^{-2}, respectively (Li and Leighton 1993), assuming a solar constant of 1365 W m^{-2}. It is, however, very difficult to assign uncertainties to these estimates. The discrepancies in SRB between satellite-based estimation and model simulation are of the order of 20–25 W m^{-2}, comparable to those found from direct comparisons between model simulations and surface observations (Arking 1996). As differences of similar magnitude also exist under clear-sky conditions, it is argued that the model deficiency stems mainly from clear-sky calculations. The analyses of zonal comparisons further suggest that the parameterizations of water vapor absorption and neglect of absorbing aerosols may be major factors causing the under- (over) estimation of atmospheric (surface) absorption.

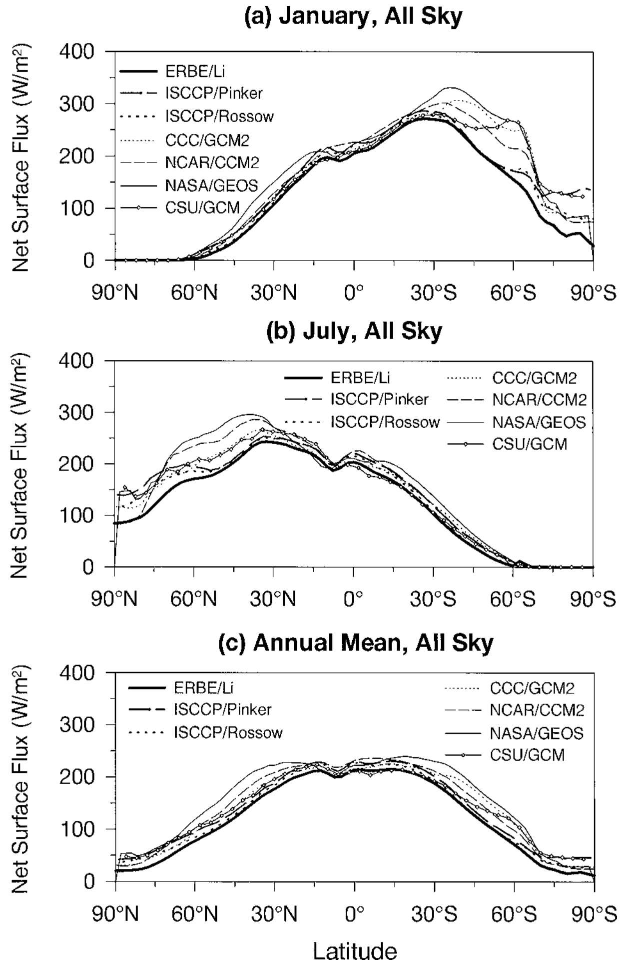

FIG. 9. Same as Fig. 8 but under all-sky conditions.

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