On Solar Energy Disposition: A Perspective from Surface Observation, Satellite Estimation and GCM Simulation

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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 data sets in three categories: surface observation, satellite estimation and GCM simulation. The discrepancies among these contemporary estimates of SED are so large that a 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. Many current GCMs can 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, we still have rather poor knowledge on the surface radiation budget (SRB) and the atmospheric radiation budget (ARB) 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. With reference to surface observations, all satellite-based SRB products are more accurate in the mid-latitudes than in the tropics where aerosols from biomass burning may deteriorate the accuracy. GCM-simulated SRB and ARB are not only subject to large regional uncertainties associated with clouds, but also to the systematic errors of the order of 25 Wm^{-2} due primarily to the neglect of aerosol and inaccurate computation of water vapor absorption. Analyses of various data sets suggest that a SED based on ERBE satellite data appears to be relatively more reliable which suggests a 30% reflection to space, 25% absorption in the atmosphere and 45% 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, the vertical distribution of moisture and aerosols, and surface optical properties. Feedbacks involving these variables and the SED are 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 hydrological 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 for driving the atmospheric circulation, especially in the tropics, and is comparable in magnitude to the solar radiation directly absorbed by the atmosphere. A recent sensitivity study with a GCM shows that modifying the partitioning of solar energy between the atmosphere and surface could substantially alter the modeled fields of cloud cover, temperature, precipitation, humidity, and the 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 and Ramanathan 1986; Ramanathan 1987) and at the surface (Suttles and Ohring 1986; Wielicki et al. 1995). Together they determine how much of the solar energy 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, unevenly distributed over the globe, 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 measurements are global, 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 inability to maintain uniform deployment standards and insure 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 (Pinker et al. 1995)

At present time, our knowledge of ERB is far more advanced than that of SRB. The global, annual mean solar flux incident at the TOA is 1365 Wm^{-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. Estimates of the fraction of the incident TOA flux absorbed at the surface, on the other hand, range from 0.42 to 0.50, as shown in this study. While models tend to show fairly good agreement with observations of TOA albedo, their proportions of surface flux tend to be higher, ranging from 0.50 to 0.55 (Garratt 1994; Li and Barker 1994; Wild et al. 1995). The residual between TOA and surface flux, which is the fraction absorbed in the atmosphere, therefore ranges from 0.20 to 0.28 for observations, and 0.15 to 0.20 for models. In terms of the flux absorbed in the atmosphere, the discrepancy between the models and observations is up to 45 Wm^{-2}.

There are two critical issues that need to be addressed. First, it is necessary to narrow the uncertainty in that part of the SED that deals with 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 showing that solar radiation absorbed by clouds have 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 by other studies (Chou et al. 1995; Li et al. 1995; Stephens 1996; Imre et al. 1996; Arking et al. 1996; Arking 1996). The amount of the claimed underestimation by clouds is of the order of 25 Wm^{-2}, comparable to the average discrepancy between models and observations. Other comparisons between models and observations show a discrepancy of similar magnitude in the clear-sky surface flux (Barker and Li 1995). 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 amongst models, the effect of clouds on atmospheric absorption, when globally averaged, is quite small, as shown in this study. Thus, the second issue requires that one resolve the question of whether or not clouds play a significant role in atmospheric absorption.

This study attempts to address the two issues from the perspective of comparison between four observational data sets---one using ground-based and three using satellite-based estimates of SRB---and four general circulation models (GCMs). Comparisons amongst these data sets and estimates of data uncertainties help us see what are the common features amongst 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 data sets. 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 were 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 1800m) 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 (c.f. 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 (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 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 (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) was carried out by Budyko (1982) and his colleagues. They generated several versions of 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 employed. With improving techniques and a growing set of observations, the estimates of solar flux absorbed at the surface increased (Budyko 1982). Their latest estimate of the SED coincides with an satellite-based estimate: a TOA albedo of 0.30, surface absorption of 0.46, and atmospheric absorption of 0.24 (Li and Leighton 1993). However, a recent ground-based estimate of surface absorption by Ohmura and Gilgen (1993) is as little as 0.42.

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 (TIROS-type) and second generations (Nimbus 3, ESSA and NOAA series), a global mean planetary albedo was found to be around 0.30 (Vonder Haar and Suomi, 1971, Stephens et al. 1981, Gruber 1985). This number is significantly lower than the pre-satellite estimates and is also in fairly good agreement with the later observations by more advanced sensors (Hartmann et al. 1986, Ramanathan 1987). 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 Earth Radiation Budget (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 (ADMs) (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 (Schmetz 1989, Pinker et al. 1995). Multiple years of SRB data have been derived 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). 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 therfore ranges from 0.20 to 0.24.

In general circulation models (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. The modeled SED can, however, help us 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 distribution, and distribution of size 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 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). Aerosol optical thickness over oceans has been inferred from the Advanced Very High Resolution Radiometer (AVHRR) (Rao et al. 1989). Global surface albedo data have been developed from the TOA clear-sky measurements (Staylor et al. 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 remarkably. Recent studies (Cess et al 1995, Ramanathan et al 1995) and Cess (private communication 1995) suggest that the global, annual mean atmospheric and surface absorption are 0.26 (90 Wm^{-2}) and 0.43 (148 Wm^{-2}), respectively.

In a word, the existing knowledge on SED is so diverse that no numbers of wisdom on SED are available. A critical examination of the various estimates is thus long over due.

3. Data

Four sets of observations and the output of four GCMs are compared in this study. One of the observational data sets utilizes measurements of surface insolation from the world-wide pyranometer network (Ohmura and Gilgen 1991), known as the Global Energy Balance Archive (GEBA), along with TOA measurements from ERBE (Barkstrom et al 1989). This data set will be referred to as ERBE/GEBA. The other three observational data sets 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. (1993a). Hereafter, they are referred to as ISCCP/Pinker, ISCCP/Rossow and ERBE/Li.

The four models include the Canadian Climate Center's GCM (CCC/GCM2), the Colorado State University GCM (CSU/GCM), the National Center for Atmospheric Research's Community Climate Model (NCAR/CCM2) (NCAR/CCM2), and the NASA'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 through 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 data sets were selected partly because of the data availability, and partly because detailed comparisons against surface observations and/or satellite estimation have been conducted.

3.1 Observational Data Sets

ERBE/GEBA consists of two independent data sets. GEBA is a data base 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 (WRDC) 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 re-evaluation 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 data set 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. ERBE is a data set of satellite measurement of TOA fluxes, available form the NASA Langley Distributed Active Archive Center (DAAC), and the data used here covers the period January 1985 to December 1989.

The ISCCP/Pinker data set (version 1.1) covers the period March 1985 through November 1988, and is available from the WCRP/SRB project operated at the NASA/Langley Research Center (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 a delta-Eddington radiative transfer model (Pinker and Laszlo 1992). The cloud optical thickness used in ISCCP/Pinker was not taken directly from the ISCCP output, but derived from ISCCP radiances. Many other input parameters were taken from the ISCCP data set, which includes analyses from a suite of operational weather satellites (Rossow et al. 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 data set (ERBE/Li, 1995). Relative to GEBA, the majority of the regional estimates of net surface flux (downward positive) are accurate to within ±20 Wm^{-2}, with an overall bias of 10 Wm^{-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 (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 (Li 1995).

The ISCCP/Rossow data set was derived using a modified version of the radiative transfer code of the GISS GCM (Zhang et al. 1995, Hansen et al. 1983). Although ISCCP/Pinker and ISCCP/Rossow employ the same input data set, 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 hours in time for every third month. The retrieved surface downwelling fluxes were validated against observations from both field experiments, such as FIRE/SRB, TOGA-COARE, and operational observation networks, such as those in GEBA. The comparisons show moderate positive biases of 10-15 Wm^{-2}. About 10 Wm^{-2} is attributed to the crude treatment of aerosols, and the remainder to the narrower band-pass of the pyranometer measurements (0.3-2.5 ÿm) (Rossow and Zhang 1995).

The ERBE/Li data set covers the period from 1985 through 1989 with a spatial resolution of 2.5 degrees in latitude and longitude (Li and Leighton 1993). The inversion algorithm employed is a simple parameterization resulting from extensive radiative transfer simulations for a variety of conditions. It involves fewer input and output parameters and less computation than the ISCCP/Pinker and ISCCP/Rossow algorithms, and it is based on ERBE radiances, which are broadband measurements with on-board calibration, as opposed to ISCCP radiances, which are narrow-band and require post-flight calibration. The quality of ERBE/Li data was evaluated by comparison with GEBA (Li et al. 1995b) and ISCCP/Pinker (Li 1995) data. The overall comparison against GEBA shows no bias and a standard deviation of about 25 Wm^{-2}, which is attributed mainly to inadequate sampling of surface measurements (Li et al. 1995b). However, moderate regional errors exist due to the use of a fixed aerosol amount, and there is a potentially large bias in the polar regions resulting from extremely low water vapor and unreliable correction of the ERBE radiances for scene identification and viewing geometry (Li 1996).

3.2 Model Data Sets

The CCC/GCM2 output data was produced for AMIP and is described by McFarlane et al. (1992). The model computes solar radiative fluxes with the two-spectral interval version of Fouquart and Bonnel's (1980) algorithm, where the solar spectrum is split at 0.7 ÿm. 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, too dry an atmosphere, too much high clouds in the tropics, and too little low clouds in the extra-tropical storm-track regions. Relative to ERBE/Li, CCC/GCM2 systematically under(over)-estimates atmospheric (surface) absorption by more than 30 Wm^{-2}, most likely due 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. (1995) and Fowler and Randall (1995).

The model uses a bulk cloud microphysics scheme, encompassing five prognostic variables that relate hydrologic processes with 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. Comparison of model output against SSM/I water vapor and cloud water data, ISCCP cloud data, and ERBE radiation data reveal considerable shortcomings. 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 by Briegleb (1992). There are 18 spectral intervals in the shortwave region (0.2 to 5.0 ÿm), 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 good agreement in general, but a considerable discrepancy in the northern summer hemisphere, 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 mid-latitudes by as much as 100 Wm^{-2} (Ward 1995).

The NASA/GEOS-1 data set is the output of a re-analysis of observational data for the period March 1985 through 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). In common with other GCMs, NASA/GEOS-1 produces too much cloud cover over the deep convective tropical regions, and too little over the mid-latitude storm tracks (Schubert 1995).

4. Comparison of Global, Annual Means

The global, annual mean SED for the eight data sets are shown in Table 1, as absolute values and as 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 Wm^{-2}, corresponding to a planetary albedo range of 28.7 to 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 amongst the data sets. The flux absorbed in the atmosphere ranges from 56 Wm^{-2} (NASA/GEOS) to 98 Wm^{-2} (GEBA), corresponding to 16.2% and 28.7% atmospheric absorptance. Likewise, the surface absorbed flux ranges from 142 Wm^{-2} (42%) to 191 Wm^{-2} (55%), a difference of 49 Wm^{-2} (13%)! Remarkably, the disparities among these seemingly state-of-the-art numbers exceed those amongst 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 Wm^{-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, 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 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 at the GEBA sites 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 change with 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 the regions where there are no ground-truth observations. These uncertainties are, however, conceived to be smaller than those arising 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 data set compiled from multiple ground-based sources was employed (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).

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 amongst 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 Wm^{-2}, three times that in the TOA reflected flux (10 Wm^{-2}). Also similar to 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 8 data sets. This is not surprising, inasmuch as there is considerable overlap in the absorption bands of water vapor and water droplets throughout the solar spectrum (Davies et al. 1984). As a result, the presence of clouds does not significantly alter atmospheric absorption (Chou et al. 1995, Li et al. 1995a).

Surface shortwave CRF differs considerably amongst the models (differences as much as 22 Wm^{-2}), but is more consistent amongst the satellite-based products (a range of only 3 Wm^{-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. In fact, Li et al. (1995a) found that the ratio is around 1, albeit strong variations with location and season, using global surface and satellite observations.

5. Comparison of Zonal, Monthly Means

5.1. Satellite-surface comparison

As mentioned earlier, all satellite products were validated against GEBA data (Li et al. 1995b, Rossow and Zhang 1995, Whitlock et al. 1995), but no zonal comparisons were presented. Figure 1 presents a comparison between ERBE/Li and GEBA for various latitude zones having GEBA stations. The comparison should not be regarded as the comparison between the *zonal means* of observed and inferred surface insolation because sampling is insufficient to represent the real zonal mean values, especially in the tropics and polar regions (see Fig. 1 of Li et al. 1995b). GEBA stations are only located over a small portion of the land area. Nevertheless, the strong dependence of bias on latitude shown in Fig. 1 is revealing: almost zero in the mid-latitudes (30^o - 65^o) and up to 30 Wm^{-2} or more in the tropics (positive) and the polar region (negative). Statistical uncertainties of the bias errors can be estimated by the ratio of the standard deviation over the square root of sample number, both of which change dramatically with latitude. In the mid-latitudes, the number of samples is much larger and the standard deviation is much smaller than in other regions. Therefore, the small bias errors in the mid-latitudes are more reliable than the underestimation in the polar region and the overestimation in the tropics. Interestingly, similar latitudinal trends are also found in the comparisons of ISCCP/Pinker and ISCCP/Rossow against GEBA, although these satellite values are generally higher than observations in almost all latitude zones (Fig. 2).

The most striking and common feature of the three comparisons is the significant overestimation in the tropics. A plausible explanation is the biomass burning that is popular over extended areas of the tropical landmass, especially in Africa and South America where the majority of the tropical GEBA stations are located. These regions are known to have widespread biomass burning to remove dry vegetation, in a systematic conversion of forests to agricultural and pastoral lands (Crutzen and Andreae 1990, Cahoon et al. 1994). For example, during the dry season of 1987, there were 350,000 fires in the Amazon basin which resulted in abundant smoke and haze clouds extending over thousands of kilometers (Setzer and Pereira 1991). These fires produce large amounts of graphitic (black) carbon, a strong absorber of solar radiation in the solar spectrum (Chylek et al. 1984, Ramaswamy and Kiehl 1985). The single scattering albedo (ÿ) for biomass burning is generally between 0.8 and 1.0 (Lenoble 1991) depending on the concentration of graphitic carbon. The concentration is modified mainly by combustion efficiency, which is related to the type of biomass burned, weather conditions, and the moisture content of the burning biomass (Kaufman et al. 1994). The value of ÿ for the biomass burning in tropical forests was estimated to be 0.90 ± 0.01 from the Biomass burning Airborne and Space-borne Experiment in Amazonas (BASE-A) (Kaufman et al. 1992). Aerosol optical thickness (ÿ) from biomass burning is also highly variable depending on the age of a fire, distance from a fire and prevailing wind direction, *etc*. At the burning site, ÿ can be as large as 2.1 at 500 nm shortly after a burn starts (Kaufman et al. 1992). While lack of actual fire information does not allow to unambiguously attribute the overestimation to biomass burning, the conjecture is supported partly by the finding that overestimation occurs only during dry season with little cloud present (Li, manuscript in preparation). Treatments of aerosols in the development of all SRB products are very crude, due to the lack of aerosol data over large time and space scales. No satellite product was developed to account for biomass burning.

As pointed out earlier, the differences shown in Figs. 1 and 2 do not represent the bias errors in the satellite-based estimates of zonal mean solar fluxes. In fact, if the explanation of biomass burning holds, the *real* zonal mean bias errors in the satellite-based estimates are much smaller than those suggested in Figs. 1 and 2, since biomass burning exerts influence primarily over land during the dry season only. Tropical landmass encompasses about one fourth the area of the tropics (30^{o}S to 30^{o}N) which occupies half the area of the Earth. Assuming that the effect of biomass burning extends across the entire tropical continents for half year, the upper limits of the bias errors in the estimates of ERBE/Li associated with biomass burning are estimated to be 4 Wm^{-2} over the tropics and 2 Wm^{-2} over the globe.

5.2 Satellite-Model Comparison

Despite uncertainties in the satellite-based products, they are still the best global data sets for model assessment. Figure 3 presents clear-sky zonal-mean solar fluxes reflected to space by the atmosphere-surface system from satellites and GCMs. The agreement is generally within 10 to 20 Wm^{-2} except for ISCCP/Pinker, which differs significantly from the others south of 40^{o}S in January, with deviations as large as 100 Wm^{-2}. The large differences may stem from inadequate narrow-to-broadband conversion and erroneous cloud identification by the old ISCCP scene identification scheme. Although ERBE TOA fluxes are generally considered the most reliable amongst 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^{o}N is likely an artificial effect of the prescription of ice bound (Li and Leighton 1991). Besides, the angular dependence model employed by ERBE (Suttles et al. 1989) for converting radiance into irradiance suffers large errors in the polar region (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. 4. 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 the underestimation over bright scenes in arid and snow/ice covered regions (Whitlock et al. 1995) which may be related to narrow-to-broadband conversion (Li 1995). In addition, the larger dispersion in TOA reflection over the summer polar regions (Fig. 3) is in line with the larger dispersion in surface albedo from various sources. 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 limit 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 many GCMs. Therefore, development of a more reliable data set of surface albedo in the polar region is pressing to improve the performance of GCMs in this region.

The same comparison as in Fig. 3 but under all-sky condition is shown in Fig. 5. 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 differs 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 Wm^{-2} per 10% difference in cloud amount. They attributed the dependency partially to the use of different angular dependence models (ADMs) by ERBE and ISCCP (Rossow and Zhang 1995). A more marked feature of Fig. 5 is that the model results disagree significantly with the satellite results. The former are substantially less than the latter in the summer mid-latitudes, but moderately more in the tropics. Since the values of their clear-sky counterparts are similar (c.f. Fig. 3), 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. 6. 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 mid-latitudes. Good agreements are found among the satellite values which differ significantly, however, from model results. The majority of the models generate too much clouds over the tropics leading to excessive CRF, and too little clouds in the mid-latitudes leading to insufficient CRF. This phenomena is common to almost all GCMs participated in the AMIP (Potter, private communication), but the causes vary from one model to another. By comparing to 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 simulate too much high clouds over warm oceans (SST > ~25^{o}C) and too little low clouds over cool oceans (SST < ~25^{o}C). Over land, agreements in cloud amount and CRF are much better. However the opposite is true for NCAR/CCM2 which generally agrees better with satellite observations over oceans than over land. This was attributed partially to the assignment of too large radius for continental cloud droplets (Kiehl et al. 1994). Based on some observational evidences (Han et al. 1994), Kiehl et al. (1994) reduced the cloud effective radius from 10 ÿm to 5 ÿm 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). Besides, Ward (1995) found that NCAR/CCM2 underestimates cloud cover in the mid-latitude marine regions, with respect 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 produce partial cloud cover (Fowler and Randall 1995).

Figure 7 compares the fluxes absorbed in the atmosphere under clear-sky condition. The discrepancy is stunning, over 30 Wm^{-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 substantial 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, 7 (L5, L6, L7), the Line-by-Line method (LBL), and the median value of many schemes adopted in radiative transfer models involved in the Inter-Comparison of Radiation Code used in Climate Models (ICRCCM) (Fouquart et al. 1991). The difference between two widely used codes, L7 and LH, is as large as 30 Wm^{-2} for a solar zenith angle of 30^o. The benchmark value from LBL is in the middle. More importantly, the median value of ICRCCM is significantly lower than LBL, implying that the majority of models underestimate water vapor absorption. Actual amount of underestimation is even larger than that indicated by Table 4, since the enhanced absorption due to scattering events is not accounted for in Table 4. From Fig. 7 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 CSU/GCM and NASA/GEOS, as they all employ the LH scheme. This is seen clearly from the tendency that the atmospheric absorptance of ISCCP/Pinker gradually approaches toward the model values as aerosol loading decreases from the tropics to high latitudes. In the polar region, atmospheric absorption escalates again because of the 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 leads to 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. 7 are not exclusively due to these two factors. For example, 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. 7, as is seen from Fig. 8. This is expected as the RTMs used in both GCMs and satellite retrieving algorithms render atmospheric absorption dependent weekly on clouds. Figure 9 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 Wm^{-2}, but diverges significantly among the various data sets. The majority of the results suggest that clouds have a slight warming effect on atmospheric absorption, whereas those of NASA/GEOS and ISCCP/Rossow show otherwise. Whether a cloud has cooling or warming effect depends on many factors, such as cloud type and altitude, vertical distributions of absorbers (water vapor, aerosol), surface albedo, solar zenith angle, etc. (Li et al. 1995a, Chou et al. 1995). Since the properties of the surface and clear atmosphere are relatively stable and should be similar among the data sets under study, discrepancies in Fig. 9 are most likely related to varying cloud conditions, especially cloud height. As cloud height increases, total atmospheric absorption decreases. This is because the dominant scattering effect of cloud droplets shields absorption of solar photons in the atmospheric column below the cloud. When 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 could 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, 1995).

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 condition (Fig. 10), the difference in surface 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 Wm^{-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 ameliorates significantly (within 5 Wm^{-2} over land). The close agreement between NASA/GEOS and CSU/GCM is envisaged 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 ISCCP/Pinker too small. Besides, 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 condition is presented in Fig. 11. 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 mid-latitudes due to the underestimation of storm-track clouds by GCMs and lessened in the tropics due to the overestimation of tropical clouds by GCMs that counteracts the overestimation under clear-sky conditions.

5. Summary

Solar radiation is the driving force of the Earth's climate system. It is disposed of in three manners: reflection to space, absorption in the atmosphere and absorption at the Earth's surface. The history and current state on the knowledge of the solar energy disposition (SED) are reviewed and examined in this paper.

Studies on SED dated back to the last century with two distinct periods: pre-satellite era and satellite era (since the 1960's). In the first period, studies were focused on the surface radiation budget (SRB) 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 (ERB) 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 observation programs in the 1960s ensued a revolution in our knowledge on ERB. The ever soaring and dancing value of planetary albedo was lowered and stabilized at around 0.3. Later, regional and temporal variations in ERB have been monitored meticulously by the scanning radiometers aboard NIMBUS-7 and ERBE for over a decade. Surrogate ERB data were available from operational satellites for two decades. Although the advance in the knowledge of ERB fosters the studies of SRB, thanks to the advent of remote sensing techniques, our current knowledge on SRB still falls far behind ERB, and so does the radiation budget in the atmosphere (ARB).

The current state of knowledge on SED is examined here by comparing eight data sets: one from surface observations, three from satellite observation and estimation, and four from GCM simulation. Comparisons were made for global and annual means, 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 over 0.1. In terms of the global and annual mean flux absorbed at the surface, the maximum difference is nearly 50 Wm^{-2}. More importantly, surface fluxes computed by models are usually larger than those inferred from satellite measurements. Satellite-based estimates agree well with surface-based observations in the mid-latitudes where the majority of surface radiation stations are located. However, they are persistently higher than ground observations in the tropics. The disagreement was tentatively explained by the influence of biomass burning in the tropics. Since such an effect is limited to a small portion of continental area, it does not alter significantly zonal and global mean solar radiation budgets. The best estimates of global and annual mean fluxes absorbed at the surface and in the atmosphere are in the neighbourhood of 155 Wm^{-2} and 85 Wm^{-2} respectively, assuming a global annual mean flux of 101 Wm^{-2} reflected to space and a solar constant of 1365 Wm^{-2}. It is, however, very difficult to assign uncertainties to these estimates. The estimated fluxes absorbed in the atmosphere and at the surface are at systematic variance with those generated by models. The discrepancies are in the order of 20 Wm^{-2} to 25 Wm^{-2}, comparable to those found from the direct comparisons between model simulations and surface observations. As the differences of similar magnitude also exist under clear-sky conditions, it is argued that the model deficients stem mainly from clear-sky calculations. The analyses of zonal comparisons further suggest that the use of dated schemes for water vapor absorption and neglect of absorbing aerosols are the two major factors causing the under(over)-estimation of atmospheric (surface) absorption.

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Table 1. Global annual-mean solar energy disposition under all-sky conditions. Both absolute values (Wm^{-2}) and relative values (in the parenthesis) are given.

Sources	GEBA /ERBE	ERBE/ Li	ISCCP/ Rossow	ISCCP/ Pinker	CSU/ GCM	CCC/ GCM2	NCAR/ CCM2	NASA/ GEOS1
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 Atmos.	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)

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

Sources	ERBE/ Li	ISCCP/ Rossow	ISCCP/ Pinker	CSU/ GCM	CCC/ GCM2	NCAR/ CCM2	NASA/ GEOS1
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 Atmos.	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 Table 1 and 2.

Sources	ERBE Li	ISCCP Rossow	ISCCP Pinker	CSU/ GCM	CCC GCM2	NCAR CCM2	NASA GEOS1
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 Atmos.	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 (CRF) ratio defined as the ratio of surface CRF to TOA CRF.

Table 4. Solar atmospheric absorption by water vapor only (Wm^{-2}) for the mid-latitude 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^o	178.1	162.3	161.0	185.7	191.3	167.0
75^o	71.4	63.6	59.5	73.3	74.4	64.2

SZA: Solar Zenith Angle, LBL: Line-By-Line, L: Lowtran, ICRCCM: International Comparison of the Radiative Codes used in Climate Models.

Figure Captions

Fig.1 Zonal comparison of surface insolation from the surface net fluxes of ERBE/Li and the surface albedos estimated with the Staylor algorithm against ground-based insolation measurements archived in GEBA (a) number of data samples in each 2.5^o latitude zone, (b) mean difference (satellite minus surface), (c) standard difference.

Fig.2 Zonal comparison of monthly mean surface insolation between the satellite-based estimates of ISCCP/Rossow (a) and ISCCP/Pinker (b) against ground-based observations.

Fig. 3 Comparison of the zonal mean flux reflected to space at the top of the atmosphere under clear-sky conditions for January (a), July (b) and annual mean (c).

Fig. 4 Comparison of the zonal mean surface albedo for January (a) and July (b).

Fig. 5 Same as Fig.4 but under all-sky conditions.

Fig. 6 Comparison of zonal mean cloud radiative forcing at the top of the atmosphere.

Fig. 7 Comparison of the zonal mean flux absorbed in the atmosphere under clear-sky conditions for January (a), July (b) and annual mean (c).

Fig. 8 Same as Fig.7 but under all-sky conditions.

Fig. 9 Comparison of the zonal mean cloud radiative forcing in the atmosphere for January (a), July (b) and annual mean (c).

Fig.10 Comparison of the zonal mean flux absorbed at the surface under clear-sky conditions for January (a), July (b) and annual mean (c).

Fig.11 Same as Fig.10 but under all-sky conditions.