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ABSTRACT
This study presents surface solar radiation flux and cloud radiative forcing results obtained by using a combination of satellite and surface observations interpreted by means of a simple plane-parallel radiative transfer model called 2001. This model, a revised version of a model initially introduced by Gautier et al., relates calibrated radiance observations from space to incoming surface solar flux. After a description of the model, an evaluation is presented by comparison with a more complex model that the authors have developed, the Santa Barbara DISORT Atmospheric Radiative Transfer model (SBDART) based on the discrete ordinate model of Stamnes et al. This evaluation demonstrates this model’s accuracy for instantaneous surface flux when used to retrieve daily (and monthly) surface solar flux. Limitations related to its lack of treatment of the bidirectional reflectance properties of clouds are also discussed and quantified by comparison with SBDART for instantaneous surface solar flux retrievals. The influence of satellite sensor calibration uncertainty is also examined in terms of surface solar flux.

The model has been applied to hourly GOES data collected over the Atmospheric Radiation Measurement (ARM) program’s central cloud and radiation testbed site in Oklahoma during a 14-month period to estimate hourly, daily, and monthly surface solar radiation flux. Comparisons of the model’s results with surface measurements made from pyranometers located at the ARM site indicate good overall agreement. The best results are obtained for daily integrated clear skies with an rms error less than 10 W m$^{-2}$ (or about 3% of the mean value) and a 2.8 W m$^{-2}$ bias. These results indicate that the clear sky model is quite accurate and also that the threshold-based technique to detect cloudy conditions works well for the resolution of the satellite data used in this study. For partly cloudy conditions the comparisons show an rms error of about 20 W m$^{-2}$ (or less than 7% of the mean) and a $-2.5$ W m$^{-2}$ bias. The performance of the model degrades with cloud cover conditions with an rms error of 22 W m$^{-2}$ (or 13% of the mean) and a bias of 13.9 W m$^{-2}$ for overcast conditions. The results improve considerably for monthly average values with an rms error of about 11 W m$^{-2}$ (or 4% of the mean) and a bias of 2.6 W m$^{-2}$ for all conditions.

The model has also been used to evaluate the cloud radiative forcing at the surface and results indicate large values of forcing for the spring and summer reaching daily values over 200 W m$^{-2}$ in May.

1. Introduction
Radiation from the sun drives the energy, water, and biochemical cycles of the earth surface–atmosphere system. Over land, the incident solar radiation flux determines, in large part, the surface temperature and the rate of evapotranspiration, with important consequences on atmosphere–surface interactions and the global hydrologic cycle. The surface solar radiation flux is affected primarily by clouds, but also by aerosols, atmospheric absorbers, scatterers, and, to a lesser extent, by surface conditions.

Various computational methods, based on visible and near-infrared observations from meteorological satellites, have been proposed to estimate surface solar radiation flux (e.g., Tarpley 1979; Gautier et al. 1980; Möser and Rashke 1984; Pinker and Ewing 1985; DeDieu et al. 1987; Darnell et al. 1988). With any of these methods it is now possible to map the surface solar flux accurately (to better than 10% on a daily timescale) over large regions of the globe, but also over the entire globe since global satellite datasets have recently become available from the International Satellite and Cloud Cli-
matology Project. Global, long-term climatologies of surface solar flux generated from satellite observations are now being produced (Whitlock et al. 1995; Charlock et al. 1994; Bishop and Rosow 1991) that can be used to investigate the sensitivity of the climate system to clouds and surface processes.

In the present study, we apply the satellite method of Gautier et al. (1980) to GOES-7 Visible and Infrared Spin Scan Radiometer (VISSR) data acquired during the first year and a half of the Atmospheric Radiation Measurement (ARM) experiment over the first experimental site located in the U.S. Southern Great Plains Cloud and Atmospheric Radiation Testbed (SGP CART). The objectives of this paper are to 1) quantitatively evaluate the method and 2) provide the temporal and spatial average and moments of the variability of surface solar radiation flux during the experiment over an area typically covered by the grid cell of a general circulation model (GCM). This information is expected to help general circulation modelers involved in ARM projects develop improved parameterizations of clouds and their interaction with radiation in their models through validation of their results with our satellite derived parameters.

In this paper, section 2 describes the main features of the radiative transfer model 2001. Section 3 evaluates the theoretical limitations of the method for a variety of atmospheric, surface, and geometrical (illumination and viewing) conditions by comparing its results with those from a more complex radiative transfer model. Section 4 discusses the impact of the satellite sensor calibration on the results of the method. Section 5 presents the methodology applied to compute the surface flux. Section 6 compares the satellite estimates made with our model to the in situ pyranometer measurements and discusses the results. Section 7 presents the first one and a half years of surface solar radiation flux over the ARM site. Section 8 presents analyses of the results concerning the shortwave radiation forcing at the surface. The last section summarizes and discusses the implications of the results presented in the previous sections.

2. Method

The 2001 model, used to compute the surface solar radiation flux, was originally developed by Gautier et al. (1980) and is based on simple physical modeling of the most important radiative processes occurring within the atmosphere, namely scattering and absorption by molecules, clouds, and aerosols. Since the variability of surface solar flux results primarily from changes in solar zenith angle and cloudiness, the method focuses on determining the effect of clouds on surface solar flux since the solar zenith angle can be computed accurately from simple formulas. The method accomplishes this by computing cloud albedo, the governing cloud parameter, from GOES VISSR measurements in the visible spectrum. The repeat coverage of the GOES VISSR data (one observation every hour in this study) allows an adequate sampling of the diurnal cloud variability.

Images composed of 695 × 415 pixels centered on 36.28°N and 96.92°W and covering a region of over 450 000 km² are acquired every hour from GOES-7 over the ARM site. The area studied here is commensurate with the size of a typical GCM grid size. The first step of the computational procedure is to calibrate the radiances measured by the VISSR sensor. This is discussed below. The next step is to estimate the surface albedo. For this, we first derive from a time series of satellite images (typically 15 days) the minimum brightness value for each pixel at each observation time during the day. This minimum value defines a threshold (taken a few counts higher) that is used to classify each GOES VISSR pixel as clear or cloudy. This procedure does not allow us to determine whether the pixel is partially covered by clouds or not. But since we utilize full-resolution data, the error introduced by not modeling the resulting effect of partial cloud cover on the surface solar flux only has an effect for clouds that are smaller than about 1 km, that is, those clouds that have a minor impact on the surface solar flux.

Once the pixel’s nature (clear or cloudy) has been determined, we apply clear and cloudy sky radiative transfer models accordingly.

In clear sky conditions, surface solar flux is expressed as

\[ I_s = S_0 \left( \frac{r}{r_o} \right)^2 \cos \theta \exp(-C_1 \cos \theta)/(1 - C_2 A_s) \]

\[ \times \exp \left[ -a_o \left( \frac{u_o}{\cos \theta} \right)^b_o \right] \exp \left[ -a_s \left( \frac{u_s}{\cos \theta} \right)^b_s \right], \]

(1)

where \( S_0 \) is the solar constant; \( r/r_o \) is the ratio of actual to mean Earth–Sun distance; \( \theta \) is solar zenith angle; \( u_o \) and \( u_s \) are ozone and water vapor amounts, respectively; \( A_s \) is surface albedo; and \( a_o, b_o, a_s, b_s, C_1 \) and \( C_2 \) are coefficients \( (C_1 \) and \( C_2 \) depend on the type and concentration of aerosols and account for Rayleigh scattering). The first two terms, \( r/r_o \) and \( \cos \theta \), are computed from the ephemeris programs of the IPW software (Dozier and Frew 1990). The term \( 1 - C_2 A_s \) accounts for photons that have sustained multiple surface reflections. Equation (1), proposed by Frouin et al. (1989), differs from that of the original model in formulation but not in essence. Ozone and water vapor amounts are specified from climatology, and \( A_s \) is obtained by solving the following equation:

\[ A_{sat}(B_{min}) = a + (1 - a)(1 - a_i)(1 - a_i)A_s, \]

(2)

where \( A_{sat} \) is the albedo measured at the satellite (the surface is assumed to reflect solar radiation isotropically), \( B_{min} \) is the minimum brightness, and \( a_i \) characterizes ozone absorption. The direct and diffuse reflection coefficients \( a \) and \( a_i \) are respectively determined.
from the tables of Coulson (1959). Equation (2) simply states that $A_{\text{sat}}$ is the sum of an atmospheric component (photons reflected back to space without surface reflection) and the signal reflected by the surface and diffusely transmitted to space.

In cloudy sky conditions, the clear sky formulation is modified to account for reflection and absorption by clouds, which are assumed to occur in one layer. Cloudy sky surface solar flux is therefore given by

$$I_c = I_s (1 - A_c - a_c)/(1 - A_c A_s),$$

where $A_c$ is cloud albedo and $a_c$ is cloud absorption. The denominator represents the effect of multiple reflections between the cloud and the surface. This effect is generally small except over snow/ice conditions, which were rarely encountered in this study.

Cloud albedo is obtained by solving the following quadratic equation:

$$A_{\text{sat}} = a + (1 - a)(1 - a_c)(1 - a_c)A_c + (1 - A_c - a_c)^2(1 - a_c)(1 - a_c)A_c,$$

where $A_{\text{sat}}$ is the top-of-atmosphere albedo, assuming that clouds reflect solar radiation isotropically. This equation, in fact, gives $A_c$ in the GOES VISSR solar channel (0.5–0.85 μm). We assume that $A_c$ takes the same value in the spectral interval of total surface solar flux. Depending on liquid water path, the ratio of narrowband to broadband albedo increases or decreases with sun zenith angle but, in general, the difference is small. The cloud absorption parameter $a_c$ is adjustable and parameterized as a function of the cloud albedo, a proxy for the cloud optical depth in this case. The parameterization used can be empirically determined (for instance, in the original version of the model this parameter was expressed as a linear function of cloud albedo with a slope of 0.07) or adjusted to fit more sophisticated radiative transfer models, as described below.

### 3. Model numerical evaluation

An important issue with this type of simple model is its accuracy. Whereby complex radiative transfer models that handle monochromatic radiation interaction with a surface and a layered atmosphere containing scattering and absorbing components are available, these models cannot be efficiently used as the basis of satellite algorithms. Simplifications and/or computational artifacts must be introduced to speed up their execution time and make it commensurate with the amount of data available from space.

One of the main advantages of the model described in the previous section is that it is simple and therefore efficient in its processing time, but also requires a minimum of ancillary information about the state of the atmosphere. It only requires calibrated radiances and illumination and viewing geometry. Legitimate questions can then be asked, such as how do the results from such a model compare with those of more sophisticated models and under which conditions does this model fail to provide acceptable values?

It is to address this type of questions that we have undertaken the validation reported here. To evaluate the simple 2001 model, we compare it with our version of the well-known and widely used discrete ordinate model (DISORT) of Stamnes et al. (1988). Our version of DISORT, called the Santa Barbara DISORT Atmospheric Radiative Transfer model (SBDART) [briefly described in appendix A and more fully described in P. Ricchiazzi et al. 1997, manuscript submitted to Earth Interactions], allows us to specify micro- and macroscale atmospheric properties for multiple investigative runs.

In order to evaluate the accuracy of the radiative transfer computations performed by the 2001 model independently of the input data, we have used SBDART to compute pairs of upwelling top-of-the-atmosphere narrowband radiance in the wavelength region covered by VISSR and surface solar (broadband) radiation flux for a variety of surface, atmospheric, and cloud conditions. These upwelling radiances were input to the 2001 model. Then, the derived broadband surface solar radiation was compared with that computed with SBDART. The results from these comparisons are presented in Fig. 1 for a vegetated surface and ozone amount of 0.3 (cm, NTP). The results are quite similar for other surfaces. To facilitate the interpretation, the conditions characterized by similar optical depths but different atmospheric conditions and illumination and viewing angle geometry have been plotted using different symbols. We can first notice the excellent agreement between the two models for clear conditions ($\tau = 0$), except for a few cases. These cases correspond to a very dry atmosphere containing large amounts of aerosols. The differences result mostly from the different ways of representing the aerosol effects in both ap-
proaches. In 2001, the aerosol effects have been fitted to those produced by the 5-S model of Tanre et al. (1990). These are based on a horizontal visibility parameter that is related to a vertical profile of aerosol particle density. The aerosol model included in SBDART follows the parameterization included in LOWTRAN 7 (Kneizys et al. 1988). This aerosol model contains a slightly different makeup of aerosol constituents and is dependent on the water vapor content of the atmosphere. Hence, the differences we see in the model comparisons are due to the lack of sensitivity of the 5-S aerosol parameterization to extremely dry aerosol conditions, which are rare occurrences in nature, typified for example by dust storms.

In cloudy conditions, for similar optical depths, several sets of solutions exist for SBDART but only one solution exists for the 2001 model. This results from the fact that for one set of atmosphere/cloud conditions, different radiances correspond to different illumination and viewing geometries. Since we assume in the 2001 model that the radiation is isotropic, this model provides one answer for each radiance. These results therefore provide a quantification of the effects of cloud anisotropic reflectance on the estimation of the surface solar flux, when isotropy is assumed. Differences as large as 200 W m\(^{-2}\) (corresponding to some instances to 100% error in the determination of the surface solar flux) can be found for optical depths varying from 2 to 100 for the different geometries tested.

To further clarify these anisotropic effects for plane-parallel cloud conditions and evaluate how they vary with different parameters, we have plotted the anisotropic factor of shortwave flux at the top of the atmosphere for different optical depths and solar and viewing geometries. These are presented in the form of polar plots in Fig. 2 for sun zenith angles of 15°, 30°, and 60°, respectively, and optical depths of 2, 10, 30, and 100. The spatial distribution of anisotropic factors varies dramatically for the investigated conditions showing limb brightening for small optical depths and a rather isotropic field centered around the point of mirror reflection for larger optical depths. Limb brightening dominates for all conditions at large viewing angles (e.g., 60°). From these results a number of conclusions can be reached regarding how these anisotropic effects of cloud reflectance could be corrected for in the 2001 model or any equivalent two-stream model to represent the anisotropy of plane parallel clouds. For large viewing angles (>60°), these results suggest that a limb brightening correction for all optical depths for homogeneous clouds would improve the representation of the 2001 results by our model. From Fig. 2 it is also apparent that an additional anisotropy correction could be included to minimize the anisotropy effects of cloud reflectance. In 2001 this correction cannot be based directly on optical depth (since the optical depth is not known), but could be based on a related parameter such as cloud albedo. For average surface solar radiation flux (daily to monthly), these anisotropy effects are compensated in part by averaging over different geometries, assuming that the averages simulate a daily or monthly average. This is shown in Fig. 3, which presents a comparison of 2001 results averaged for all geometries, for each test condition and the SBDART results. A large reduction in the difference between the two sets of results has occurred with part of the remaining difference due to the difference between the treatment of aerosols (as discussed above for clear sky conditions). One of the difficulties with including such anisotropy corrections is that real clouds are not plane parallel, and therefore even though we might be able to tune 2001 to reproduce SBDART-like results, there is no assurance that these model results will be closer to reality than those presently produced by 2001. As we discuss later, using the standard ERBE Angular Distribution Models (ADMs) does not improve the surface solar irradiance results obtained from 2001.

As mentioned in the previous section, cloud absorption in 2001 is expressed as a function of cloud albedo. To ensure a similarity with SBDART's representation of cloud absorption, in 2001 we have adjusted the relationship between cloud absorption and cloud albedo to be that obtained from the SBDART model at a constant sun zenith angle of zero. This empirical fit is presented in Fig 4. The nonlinearity of the relationship is obvious, especially for large cloud albedo values for which cloud absorption is slightly reduced. Admittedly, performing comparisons between 2001 and SBDART results using an absorption empirically fitted to SBDART does not represent a completely independent validation. Such a validation has yet to be conducted.

Our 2001 model implicitly assumed that the cloud layer is fixed at 1 km and contains droplets of constant effective radii equal to 8 µm. To evaluate the impact of this assumption on the surface solar irradiance computations, we have performed SBDART computations for different cloud height and droplet properties. The differences between the standard 2001 model and the results obtained with modified cloud conditions are presented in Table 1. The values for all of the statistics are very similar for each case with the exception of mean bias error of the two differing cloud heights. This slightly higher bias reflects a high bias at high cloud albedo for all values of \( R_c \). The differences are so small, however, that these parameter variations were not incorporated into our model to maintain its simple nature. Overall, the results presented above show that the 2001 model performs very well under most conditions, when the issue of anisotropy is, at least partly, removed by time-averaging processes. Obviously, a model such as SBDART could theoretically be far superior to 2001, but requires such a large number of input parameters (and therefore assumptions) with regards to cloud properties (microphysical and macrophysical) as well as atmospheric and surface properties that this rapidly outstrips its advantages. Furthermore, a radiance-based
model such as SBDART is extremely processing time consuming and therefore unrealistic for processing large amounts of data, such as those involved in computations of global surface solar flux, or even long-term high temporal resolution datasets, such as the ARM dataset discussed above.

4. Calibration issues

As mentioned above, the input data to 2001 is calibrated visible brightness (or radiance). Obviously, similar radiative transfer-based models have the same calibration requirement. Operational satellite sensors are calibrated before launch but have no usable in-flight calibration system. Therefore, vicarious in-flight calibration must be performed to remedy this deficiency. Although not entirely satisfactory, a number of approaches have been proposed and used in the recent past. While our goal is not to provide an extensive review of the presently available calibration methods, it is important to note that despite recent efforts the calibration of the U.S. operational sensors (VISSR on the GOES satellites and AVHRR on the NOAA satellites) is still uncertain to probably at least 6%–7% (Whitlock
et al. 1990; Whitlock et al. 1993). This value represents an estimate obtained by comparing results from all the available calibration methods and computing their standard deviation. It therefore does not represent an absolute calibration assessment.

A second issue of importance is the drift most visible sensors experience with time. Both VISSR and AVHRR sensors have been found to deteriorate with time, which leads to increasing gain with time. While the source of this deterioration is not entirely clear (deposition on the sensor, optics deterioration), it can be rather large (e.g., 10% per year for a VISSR sensor). These sensors must therefore be routinely monitored to adjust for any such deterioration. Whitlock et al. (1990) have suggested that the gain of the VISSR instrument on GOES-7 has increased in the manner described by Fig. 5. As can be seen, the changes since 1991, after Mount Pinatubo’s eruption, have not been monitored due to the sensitivity of the monitoring technique to any long-term change in atmospheric conditions, particularly high aerosol concentrations. Also, the apparent periodic changes are not completely understood and thus difficult to predict.

The third issue with regards to calibration is that of its impact on the computation of the surface solar flux. To evaluate this impact, Gautier and Frouin (1988) have computed the sensitivity of the surface solar flux to radiometric calibration of the visible instrument. This evaluation was based on an earlier version of 2001, very similar to that presented above. The approach was to linearize the model equation about a reference state and evaluate the changes in surface solar flux with respect to instrument gain ($g$) changes (or sensitivity). For that, the sensitivity of the surface solar flux to gain ($\frac{\partial I_s}{\partial g}$) was evaluated from the model’s equations (1), (2), (3), and (4). When applied to these equations, the operator $\frac{\partial}{\partial g}$ leads to an equation for $\frac{\partial I_s}{\partial g}$ that is a linear combination of $\frac{\partial A_s}{\partial g}$, $\frac{\partial N}{\partial g}$, and $\frac{\partial A_c}{\partial g}$. Further mathematical manipulations lead to the following conclusions.

1) The sensitivity of surface albedo to calibration gain is a linear function of the surface albedo; the brighter the surface, the higher the sensitivity. 2) The sensitivity of the cloud albedo to calibration gain is a nonlinear
function of the cloud albedo, the nonlinearity increasing with surface reflection. 3) The sensitivity of the surface solar flux to a 10% uncertainty in calibration gain varies with surface and cloud conditions with a range of 75 W m$^{-2}$ for totally cloudy skies ($N = 1$) to $-50$ W m$^{-2}$ for solar zenith angle of 0 degree with selected conditions ($A_s = 0.4, A_c = 2, N = 0.5$). Thus, instantaneously, the sensitivity to the calibration can be as large as $-75$ W m$^{-2}$, whereas for monthly average values the error can be reduced to $-15$ W m$^{-2}$ because of error compensations. The bias on the solar flux computation is inversely related to the calibration bias; that is, a negative bias in calibration induces a positive bias on the surface solar flux. This result is very important and not unexpected. As a consequence of the sensor’s deterioration with time, we can expect a negative bias in the sensor’s gain and therefore a positive bias in the computed surface solar flux, independent of the quality of the radiative transfer model used to perform the computations. If not properly interpreted in terms of the sensor’s calibration, such a positive bias could be misinterpreted as a potential cloud effect such as clouds absorbing more solar radiation than is represented in any present radiative transfer model. Therefore, great care has to be taken to correctly calibrate operational sensors that are used in conjunction with radiative transfer models for estimating the surface solar flux.

5. Surface solar radiation flux computations

We focus on the first 14 months of the ARM program and thus on a dataset that extends from 1 March 1993 to 30 April 1994. The experimental site is located at the ARM Southern Great Plains CART site in central Oklahoma.

a. Satellite data

To compute surface solar radiation flux over the first CART site (using the satellite calibration method described above), GOES-7 VISSR visible and near-infrared (solar channel) data were acquired at full resolution (0.9 km at nadir) every hour during daytime. This resolution translates into a 1.25-km resolution at the CART site. The data, 8-bit coded, navigated, but uncalibrated numerical counts, were made available by the ARM Experimental Data Center. In a preprocessing stage, we calibrate and check the data for gross navigational errors and data quality. The navigational errors are only detectable in clear sky conditions by visual inspection of large surface features. The SGP CART site comparison computations presented are made over a 3 pixel $\times$ 3 pixel area to use as high a resolution as possible but still compensate for locational inaccuracies and angular integration (2$\pi$ solid angle) of the surface measurements with which the satellite estimations are compared.

b. VISSR visible sensor calibration

As mentioned above, the calibration of the satellite data used is central to the accuracy of the surface solar flux estimated from these datasets. Considering that the last vicarious calibration made was before Mt. Pinatubo’s eruption (June 1991) and that the VISSR calibration deteriorates and also varies seasonally, it is rather difficult to confidently extrapolate the gain value from June 1991 to 1993 and 1994. It is even more difficult to confidently extrapolate the amplitude and frequency of the apparent oscillation with such a small sample set. We therefore derived a drift only from the data presented in Fig. 5 to perform our extrapolation. The results from this extrapolation provides a gain varying with time in the following manner:

$$g(t) = g_0 + g_1 t,$$

(5) where $g_0$ is the instrument gain at $t = 0$ and $g = 0.009$, $g_1$ is the gain’s rate of change with time (0.00084), and $t$ is the time since launch (years). As for the offset, we assume that it remains equal to zero, as it was in 1991.

Furthermore, although the VISSR sensor is composed of eight detectors that have been found to deteriorate at different rates, we used the gain provided in Eq. (5) for all the detectors. This is probably adequate since NOAA, the operator of the GOES satellite, regularly (and artificially) modifies the gain of some of the sensors to provide an aesthetic nonstripped appearance to the images that it makes available to the public.

c. Surface measurements

The surface measurements used in this study are those that are routinely recorded by the BSRN (Baseline Surface Radiation Network) pyranometer at the SGP CART site with the correction factor introduced in November 1993. The measurements taken at each minute are time averaged for the 20-min period encompassing each satellite acquisition time. This is roughly equivalent to the average time for clouds to traverse the 3 pixels used in the computations of the spatially averaged surface solar flux from the satellite data when a typical wind speed is used.

6. Analysis of results

Figure 6 presents comparisons of some daily time series of surface solar radiation flux measured by the SGP CART pyranometer described above and hourly estimations made with 2001 applied to GOES instantaneous visible radiances every hour. These time series clearly show the high-frequency variability of the pyranometer’s measurements, which results from the presence of small clouds overhead. The series on the left panels present the datasets corresponding to the four best comparisons between the surface measurements and the satellite estimations. The criterion used to select
the best or worse cases is based on the comparison of daily integrated surface solar flux; therefore, some cases are labeled worst when data are missing during a period when clouds vary greatly. The best cases generally are in rather clear or partly cloudy conditions. In cloudy conditions, the satellite estimations capture the changes in solar flux (due to variations in cloudiness) measured by the pyranometers. The worst cases correspond to low solar flux values (thick clouds), to missing satellite measurements for a long period during the day, or to cases where there is an oscillatory variation in the solar flux (cloudiness) and the satellite estimations oscillate out of phase with the pyranometer oscillation; that is, when the pyranometer measures a low solar flux value, the satellite estimation indicates a high solar flux value.

To further quantify the differences between the 2001 computations and the surface observations, the hourly results have been statistically analyzed. A scatterplot of these comparisons is presented in Fig. 7. Also included are the statistics of the comparison of these two datasets. These statistics are presented in two forms. 1) At the bottom right are the mean differences (or the mean of
the differences) between the two datasets and the standard deviation of that difference (expressed in W m\(^{-2}\)) and in percentage of the mean value and 2) at the top left are the equations for the best linear fit, the mean bias, the squared correlation, and the rms error. The datasets compared are very highly correlated (\(R^2 = 0.869\)), have a very small negative bias (\(-11 \text{ W m}^{-2}\)), and have an rms error of 93 W m\(^{-2}\) (or 18 % of the mean value of the entire dataset). This dataset includes clear and cloudy conditions. A few outliers are observed, however, with a small tendency for the satellite estimations to be lower than the pyranometer measurements for low solar flux values, suggesting that the absorption parameterization used in 2001 may be a little too large.

For clear conditions, the 2001 model performs extremely well as shown in Fig. 8. The rms error is 54 W m\(^{-2}\) (or 8.4 % of the mean value of the dataset) and the bias is small (15 W m\(^{-2}\)) and positive. This figure also illustrates the ability of the threshold method to determine clear conditions. Only a few percent of the cases show that the model clearly mistakes a cloudy scene for a clear one (outliers showing larger values from the model than what is measured at the surface by the pyranometer). In no instance does the model estimate that it is significantly cloudy when, in fact, it is clear. Thus, if anything, the model will have a small tendency to provide slightly highly biased results due to the rare misinterpretation of cloudy scenes for clear scenes. In no way does the surface data analyzed suggest a significant overestimation by the model.

To investigate whether 2001 results degrade with increasing sun angle, we have partitioned our datasets into different sun angle conditions. The results are presented on Fig. 9 values of the sun zenith angle varying from 13° to 70°. The plot shows that there is no real tendency for a deterioration at large sun zenith angle. This is not surprising since most radiative transfer model performances deteriorate beyond 80° sun angle. That the largest differences occur at small solar zenith angles results from the fact that the surface solar flux is larger at these small angles.

a. Cloudy conditions

To compare the performances of 2001 in different cloudy conditions, the data have been partitioned into different cloud cover conditions. Comparisons are still based on the 3 × 3 pixel array mentioned above, but to ensure an accurate cloud condition classification, four regimes have been selected based on the number of cloudy pixels in a 9 × 9 pixel array surrounding the ARM-SGP CART site. The classifications used in our computations were clear (no cloudy pixel), mostly clear (less than half of the pixels are cloudy), mostly cloudy...
(more than half of the pixels are cloudy), and cloudy (all pixels are cloudy). Results from this partitioning are presented in Fig. 10 with the statistics for each case. We can see that the statistics deteriorate slightly with the cloudiness and particularly that there is a tendency for the satellite computations to underestimate the values measured at the surface in overcast conditions. This can result from a number of causes. The first one is a misrepresentation of cloud absorption. This is an issue that is still not resolved and the topic of ongoing discussion within the scientific community (see Ramathan et al. 1995; Cess et al. 1995; Li et al. 1995; Stephens 1996; Arking 1996; Arking et al. 1996). Our preliminary work on this issue of cloud absorption indicates that absorption in a 3D cloudy atmosphere is complex and varies with sun angle as well as cloud geometry. The simple parameterization used in this model, which is patterned after a plane-parallel discrete ordinate model (SBDART), is intended to account for some of the processes involved. Since a full analysis of this complex issue still remains to be done, we believed that our simple approach is the best possible choice at the moment. Another possibility is calibration. Indeed, if the gain of the sensor is improperly characterized and assumed to be higher than it should be, the brightness corresponding to cloudy pixels would be too high and so would the cloud albedo. The surface solar flux would, in turn, be underestimated. A departure from linearity of the gain as a function of the count squared could also induce this type of underestimation for large brightness values. One way to address the origin of the underestimation is to investigate daily comparisons partitioned into clear and cloudy conditions. This is done in the following section.

b. Comparisons of daily surface solar flux

The hourly results presented above include some uncertainty due to the cloud anisotropy effects not accounted for in 2001, as was discussed earlier. In order to remove these effects, the hourly data have been integrated over the day using the trapezoidal method with computed sunrise and sunset times. We can therefore expect an improvement in the statistics of the comparisons. The overall results obtained from this integration are presented in Fig. 11 for clear, partly cloudy, overcast, and all conditions.

The cloudy conditions have been partitioned, this time, into two regimes: overcast and partly cloudy since it is rather difficult to come up with stricter criteria that can be applied for several hours of the day. Using this partitioning scheme, we classify daily conditions as either entirely clear, overcast, or if neither, as partly cloudy.

It is interesting to note that the clear conditions are computed with an rms error of less than 10 W m$^{-2}$ (or about 7% of the mean surface measurements). The overcast conditions are computed with an rms error of about 22 W m$^{-2}$ (or about 15% of the mean surface measurements) while the partly cloudy conditions have an rms error of about 20 W m$^{-2}$ (or about 7% of the mean surface measurements). The rather small slope (0.83) determined for the linear fit between the satellite-based estimations and the surface measurements in overcast conditions, together with an inspection of the plotted data suggest that the approach has a tendency to underestimate the surface solar radiation flux in the most cloudy conditions.

c. Sensitivity of results to 2001 assumptions

1) Calibration uncertainty effects

Many potential sources for the differences between the 2001 satellite-based computations and the surface observations have been investigated. The first and obvious one is calibration error, which is readily understood when one considers that the latest available calibration value was obtained in spring 1991, or about 3 years before the data analyzed here. Since it is difficult to directly quantify calibration errors, we have performed our computations with an alternative available calibration source. We have used the GOES-7 VISSR calibration coefficients provided by the International Satellite Cloud Climatology Project (ISCCP) (Rossow et al. 1992a; Rossow et al. 1992b). We found that the computations performed with Whitlock’s calibration provided surface solar irradiance values that compared slightly better with the surface observations than those obtained with ISCCP calibration. The standard deviations were only slightly different with 12.43% and 12.85%, respectively, for the Whitlock and ISCCP calibrations, but the mean biases were significantly different with 10.81 W m$^{-2}$ and $-23.68$ W m$^{-2}$, respectively. For daily integrations Whitlock’s calibration results were similarly better with the standard deviations of 6.95% and 10.26%, respectively, and mean biases of 0.65 and $-9.49$ W m$^{-2}$, respectively.

2) Reflectance anisotropy effects

Another potential source of error that has been discussed above is the cloud reflectance anisotropy effect. While this can be quantified theoretically for plane-parallel clouds, its effect for inhomogeneous three-dimensional clouds is unknown. Here again, to evaluate the performances of our model under different sets of assumptions that are commonly made, we have performed runs of 2001 using the ERBE (Suttles et al. 1988) anisotropic cloud reflectance model and compared the results with those obtained with the standard 2001 assumptions. For this, we have developed an anisotropic cloud reflectance model based on ERBE climatological results and applied this to the cloud albedo calculations of the 2001 model. Here again the results obtained with the 2001 model provide better comparisons with the
Fig. 10. Comparison of measured and calculated surface solar radiation for differing cloud cover types (pixel classifier).
Fig. 11. Daily integrated shortwave comparisons with measurements at the ARM-SGP for completely clear, partly cloudy, and overcast days.
surface observations than those obtained using an ERBE anisotropy correction. These results are summarized in Table 2. They do not suggest that an anisotropy correction is not warranted but only that applying the available one does not lead to improved interpretation of satellite radiance in terms of surface shortwave fluxes using our 2001 model.

3) Cloud absorption effects

Finally, we have evaluated the effects of the cloud absorption assumption in 2001 by comparing the results to those obtained with the previous parameterized absorption \( a_c = 0.07 A^c \) used in our previous models (Diak and Gautier 1983; Frouin et al. 1989). The results obtained again indicate that the present absorption value based on SBDART computations provides slightly better results, which are presented in Table 3.

d. Monthly surface solar flux and its variations

For climate applications, time averaging longer than daily averaging is often performed to characterize the behavior of climate variables. To provide an idea of how the solar flux varies seasonally and annually in the region of the CART site over the scale of a climate model (250–500 km), we have computed monthly averaged values over the entire study area. The accuracy of such monthly estimates can be assessed in a manner similar to that of the hourly and daily values, that is, by analyzing the scatterplot of the pyranometer and the satellite monthly mean surface solar flux values. This scatterplot is presented in Fig. 12 with similar statistical information as for the hourly and daily values. The year and month is also indicated next to each plotted point. This figure shows that monthly mean values are obtained with an rms error of 11 W m\(^{-2}\) (or 6% of the monthly mean value) and a small bias of 2.6 W m\(^{-2}\). These results show that the rms error is reduced dramatically by temporal averaging, suggesting that a large part of the error in the shorter time averages is random (noiselike) and likely due to cloud effects. There still is a bias component to the rms error (not removed by the averaging process), however, which can come from the model or the data and is on the order of about 10 W m\(^{-2}\). This is consistent with the results discussed earlier concerning the calibration effects on the cloudy retrievals, which were of the order of 22 W m\(^{-2}\) because many days that make up the monthly mean are clear or partly cloudy and the monthly rms error can be in large part attributed to the cloudy conditions. The two main outliers that influence the rms error in the monthly mean values are for springtime.

Monthly mean values for the entire area centered on the SGP CART site have been computed for the solar radiation flux and the cloud area cover (N). A plot of the time series of these parameters is presented in Fig. 13. The top panel, which represents the variations of the solar radiation flux, shows the strong seasonal variations of the flux induced by both solar zenith angle changes and cloud effects (cloud cover and type). The phase of the cycle suggests that the sun effects are dominant. The cloud cover variations are presented in the lower panel and also show some seasonality. The minima and maxima are, however, out of phase with those of the solar radiation flux. In particular, the minima of cloud cover are in summer and early fall, while late winter and spring seasons experience the largest cloud cover.

The main result from this analysis is that clouds have
7. Surface solar flux cloud forcing

Another way of quantifying cloud radiative effects is to investigate the difference in surface solar flux between cloudy and clear conditions. This approach was introduced by Charlock and Ramanathan (1985). Since clear and cloudy conditions do not exist at the same time, one has to either rely on models to derive the clear conditions or to compare different time periods, ensuring as much as possible that the sun illumination and the atmospheric conditions (e.g., aerosols and water vapor) between the two datasets compared are similar within the accuracy of the measurements used. To compute the daily cloud forcing, we need the daily clear sky flux estimated instantaneous (or at least hourly) clear flux. Since there are very few days for which all the sunlight hours are clear, we use our model to compute the clear sky flux for each hour. Cess et al. (1995) has proposed a method that consists of taking the envelope of the points on a plot of the downwelling solar radiation flux as a function of the solar zenith angle. If a strict envelope is taken, the clear sky flux can be overestimated because of a few points that correspond to conditions when the surface solar irradiance is the sum of the total flux plus that reflected on the sides of small clouds. Some interpretation therefore has to be made.

For this reason and since we have shown that our model was accurate for clear sky instantaneous surface solar flux (5.6%, see Fig. 8), we investigate the surface solar flux cloud forcing using a combination of surface measurements and model data. Our computations of the surface solar flux cloud forcing are made according to

$$ CF = f_{\text{obs, cloudy}} - f_{\text{obs, clear}}, \quad (6) $$

where $f_{\text{obs, cloudy}}$ is the surface solar flux in clear conditions computed with 2001 and $f_{\text{obs, cloudy}}$ is the surface solar flux in cloudy conditions measured with the pyranometers. We will assume a 5% uncertainty in the modeled clear sky conditions, and this is attributable to an imperfect knowledge of the atmospheric conditions, particularly aerosols. Since typical pyranometer measurements are accurate to about 5%, the uncertainty on the cloud forcing at the surface is expected to be better than 10% for hourly values and 5% for daily values.

We have computed the surface solar flux cloud forcing for both hourly and daily values. The hourly results are presented in Fig. 14 in the form of a histogram for each month analyzed. Since our dataset spans over 14 months, two months have more data than others. As expected, the largest frequencies are for the smallest values of hourly surface solar flux cloud forcing. The months during which the values are the largest are April, May, and June, for which the surface solar flux cloud forcing reaches values up to 680 W m$^{-2}$. May is the month during which the largest cloud forcing values are encountered and large values occur most often. For the entire 14-month study period, the mean and standard deviation for the instantaneous cloud forcings are 147.5 W m$^{-2}$ and 152.8 W m$^{-2}$, respectively.

Similar results are found for the daily surface solar flux cloud forcing (Fig. 15). Here again, May is the month during which the largest number of days have the largest values. The daily surface solar flux cloud forcing reaches values beyond 200 W m$^{-2}$. So, if we were interested in performing an experiment when clouds have the largest effect on the surface solar flux, springtime, and May in particular, would be the best suited, if our limited climatology is representative of the region. During the entire study period, the mean and standard deviation for the daily cloud forcings are 53.1 W m$^{-2}$ and 50.5 W m$^{-2}$, respectively.

8. Discussion and concluding remarks

The results presented above are very encouraging with regards to producing high spatial and temporal resolution climatology of the surface solar flux from satellite observations with our model, 2001. These, once again, demonstrate that it is possible to compute time-averaged surface solar flux from satellite data with an accuracy close to that of surface measurements (made with operational pyranometers) but with a much higher spatial resolution and coverage. This is particularly important for regions such as oceans or remote regions.
Fig. 14. Downwelling hourly shortwave cloud forcing histograms by month.
Fig. 15. Downwelling daily shortwave cloud forcing histograms by month.
where surface measurements are difficult to make on a continuous basis. While results have not been presented here for different types of surfaces, our experience indicates that the model performs as well over ocean and other land type surfaces, as over vegetation (as presented here). Snow surface conditions have special conditions with multiple reflections between the surface and clouds, which need to be handled differently than the simplistic treatment included in 2001.

While rather accurate, the 2001 model also has the advantage of being particularly simple to use in comparison with models such as the delta-Eddington or DISORT radiative transfer models, which require a much more detailed characterization of the physical environment being modeled. The 2001 model can be used with any visible channel satellite radiance data (i.e., AVHRR or any geostationary weather satellites) for which a calibration method is available. Because of the way the cloud discrimination is performed, the 2001 model accuracy is expected to deteriorate with the spatial resolution of the input data.

Calibration of the data is an important issue that is not yet solved adequately for the datasets that are made available to the scientific community. Vicarious calibrations with a surface calibration target are possible and useful when clear conditions are not affected by the presence of an unknown amount of aerosols. Other approaches to in-flight calibration must be explored such as intercalibration with other airborne instruments. An acceptable solution needs to be developed soon before long-timescale satellite-based climatologies are developed with inappropriate calibration coefficients.

The effects of clouds on the surface solar flux, which have been computed from a combination of surface measurements and radiative transfer model computations, have been found to be large over the SGP CART site, particularly during the spring season. Whereas most cloud forcing computations have been performed using net flux measurements of satellite observations, here we have computed the cloud effects on the surface downwelling flux solely. This choice was made to avoid errors introduced by uncertainties on the surface albedo. Consequently, our results represent a different way of computing the cloud radiation forcing with data in which the surface albedo is close to 20%, we can therefore expect a reduction by 80% of the net cloud forcing at the surface. This would bring our mean, net instantaneous and daily values to about 30 W m$^{-2}$ and 10 W m$^{-2}$, respectively.

Finally, our monthly mean computations have shown 2%–11% spatial variability (expressed by the standard deviation normalized by the mean) in the surface solar flux over the scale analyzed (roughly that of a GCM grid box), suggesting that over the monthly timescale cloud effects on the surface solar flux can be considered small and random.

The 2001 model is now available to the scientific community. It is currently being used to compute the surface solar fluxes for the BOREAS experiment (Gu and Smith 1995).

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APPENDIX

Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) Model

The discrete ordinate method (DISORT, Stamnes et al. 1988) provides a numerically stable algorithm to solve the equations of plane-parallel radiative transfer in a vertically inhomogeneous atmosphere. The intensity of both scattered and thermally emitted radiation can be computed at different heights and directions. Parameters that define the radiative properties of the atmosphere, such as the single scattering albedo and asymmetry factors of clouds and aerosols, are computed from well-established theories of radiation scattering. The computation of gaseous absorption is based on models in the LOWTRAN 7 (Kneizys et al. 1988) atmospheric transmission code. Our version of DISORT allows general specification of the model atmosphere and enables us to treat a very wide variety of atmospheric radiation problems.

The storage and computing time required by DISORT is considerable, and increasing the spectral resolutions, number of layers, and streams slows the execution time significantly. Therefore, we implement the parameters in a modular structure. The gaseous absorption coefficients, aerosol parameters, and solar spectrum in the wavelength region 0.25–33.0 μm are provided at a spectral resolution of 20 cm$^{-1}$. For more accurate calculation of narrowband flux and/or intensity in the gas absorption bands, where higher spectral resolution is required, the K distribution fitting of HITRAN-92 dataset is used.

The distribution of temperature, pressure, and composition may be selected from a set of standard atmospheric profiles or specified from detailed radiosonde profiles. The single scattering albedo, asymmetry factor, and the extinction efficiency are calculated from a Mie scattering code. The current version of the code assumes the surface is a Lambertian reflector. A BRDF that ap-
plies to different surface types will be implemented for increased accuracy with intensity calculations.

a. Structure of the model

As a compromise between calculation accuracy, computing time, and storage requirements, 40 plane-parallel layers and 16 radiation streams are currently used in the computations. The number of streams can be increased when relative narrow band intensity is calculated. The model consists of modules to

1) read and validate user input;
2) specify solar spectral data and atmospheric profiles;
3) specify sensor filter function including predefined filter types such as AVHRR 1 and 2, Meteosat, GOES, Landsat TM, and SPOT (the code also allows input of a user specified filter function);
4) calculate optical thickness of gas and Rayleigh scattering;
5) calculate optical thickness, single scattering albedo, and asymmetry factor of aerosols and clouds;
6) calculate surface albedo or BRDF; and
7) solve the equation of radiative transfer with DISORT.

b. Optical parameters of the atmosphere

In solving the radiative transfer equations, the parameters required are the optical thickness, single scattering albedo, and asymmetry factor due to gaseous absorption, Rayleigh scattering, aerosols, and clouds. They depend on the atmospheric profiles; total amount and distribution of water vapor, ozone, oxygen, carbon dioxide, and other gases; type and concentration of aerosols; and type and liquid water content of clouds.

1) Atmospheric profiles

We have adopted six standard atmospheric profiles from the 5 S atmospheric radiation code (Tanre et al. 1990), which model the following conditions: tropical, midlatitude summer, midlatitude winter, subarctic summer, subarctic winter, and US62. These models are standard atmospheric profiles recognized by the earth-radiation community.

An additional source of information is the TIGR dataset (Chedin and Scott 1985), which contains 1761 radiosonde atmospheric profiles taken between 80°S and 84°N latitude. These data can be used to provide a range of typical atmospheric profiles that may be realized at any given latitude.

2) Gas absorption and Rayleigh scattering

Gaseous absorption is computed using the models provided by Kneizys et. al. (1988). Absorption by H₂O (vapor), O₃, CO₂, N₂O, CO, O₂, N₂, CH₄, and trace gases are computed at a spectral resolution of 20 cm⁻¹.

3) Aerosols

We use the aerosol dataset from 5 S and LOWTRAN 7. In the former case, the single scattering albedo and asymmetry factor of four basic aerosol particles (dust-like, oceanic, water soluble, and soot) are specified, and the radiative properties of the user-specified aerosol type (continental, maritime, and urban) are calculated by combining the basic aerosol particle types with different weights. In the LOWTRAN 7 version, a database that specifies radiative properties of rural, urban, oceanic, tropospheric, background, aged and fresh volcanic and meteor types of aerosols is used directly. In either case the aerosol optical thickness (which is a function of wavelength) is set by specifying the meteorological visibility at 550 nm.

4) Clouds

Solving the equations of radiative transfer in a cloudy atmosphere requires the knowledge of the cloud single scattering albedo, asymmetry factor, and extinction efficiency. We have computed these parameters with a Mie scattering code for cloud droplets having a lognormal size distribution for a range of equivalent radii between 2 and 64 μm. For flux calculations we use the Henyey–Greenstein approximation of the scattering phase function. The Legendre expansion of the actual Mie scattering phase function can be used to obtain radiation intensity at high angular resolution.

c. Surface conditions

The ground surface cover is one of the most important factors determining the radiation environment of the earth–atmosphere system. In our radiation models we use four basic surface types (clear water, vegetation, snow, and sand) to parameterize the hemispherical spectral reflectivity of the surface. The spectral reflectivity of a large variety of ground objects is well approximated by combinations of these basic types. For example, the fractions of vegetation, water, and sand can be adjusted to generate a new spectral reflectivity representing new or old vegetative growth or deciduous versus evergreen forest. Combining a small fraction of the spectral reflectivity of water with that of sand yields an overall spectral dependence close to wet soil. We expect that six to eight prototypical spectral reflectivities can be combined to get the spectral reflectivity of most ground objects to an accuracy sufficient for most applications.

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