*Solar Energy* Vol. 51, No. 6, pp. 457–465, 1993
Printed in the U.S.A.

0038-092X/93 $6.00 + .00


# SOLAR IRRADIANCE ESTIMATION FROM GEOSTATIONARY SATELLITE DATA: II. PHYSICAL MODELS

M. NOIA,* C. F. RATTO, and R. FESTA
Department of Physics, University of Genoa, Italy

**Abstract**—The use of satellite data to estimate solar irradiance at ground level represents a valid alternative to ground measurements of solar radiation. This paper continues the analysis and evaluation, started in a previous paper, of the best known methods for calculating solar irradiance at the earth's surface using geostationary satellite data. In the previous paper, we examined and compared the so-called statistical models. Now we will consider the physical models and point out the differences between them. Finally, a summary will be made of the assessments and comparisons carried out on the methods described.

## 1. INTRODUCTION

As already stated in the companion paper[1], two different approaches have been developed for estimating solar irradiance incident at ground level using satellite images. The first approach, known as the statistical method, has already been discussed. In this paper we will consider the second approach, known as the physical method, based on radiative transfer models which explicitely describe the scattering and absorption processes operating in the earth-atmosphere system.

Methods of this kind were developed during the eighties by Gautier, Diak, and Masse[2], Möser and Raschke[3], Dedieu, Deschamps, and Kerr[4], and Marullo, Dalu, and Viola[5].

A survey on the physical basis of both the statistical and the physical methods has already been given[1].

## 2. GENERAL DESCRIPTION OF THE PHYSICAL METHODS

The physical methods are based exclusively on physical considerations that allow the radiant energy exchanges taking place within the earth-atmosphere system to be explicitly represented. The main quantities considered are: scattering and absorption coefficients of the clear atmosphere components; cloud albedo and absorption coefficients; surface albedo.

The main advantage offered by the physical methods is, in comparison with the statistical methods, their general nature, since they do not depend on a particular region and may be applied anywhere. In addition, they do not need solar radiation data measured at the earth's surface.

Unfortunately, the physical methods need complementary meteorological data to estimate the various quantities relating to the interactions of solar radiation with the atmosphere. Another drawback of the physical methods is that the digital count values, provided by the satellite, need to be converted into the corresponding flux density of the upward solar radiation emerging from the atmosphere. As a consequence, accurate and updated calibration of the instrument is required.

Moreover, the physical methods share a common difficulty with the statistical methods, caused by the difference between the space and time scales of satellite images and the scales of ground-based measurements. A brief description of the procedures followed to overcome these difficulties appears in the previous paper[1].

## 3. PHYSICAL METHODS

### 3.1 *GDM method*

One of the best known physical methods is that developed by Gautier, Diak, and Masse[2], here referred to as the GDM method.

The GDM method is based on three assumptions:

- water vapor is the principal atmospheric component responsible for the depletion of solar radiation through absorption;
- aerosol effects, although important, are not considered since they are complex, variable and not well known; and
- scattering from the ground is assumed to be isotropic.

The main feature of this method is the formulation of two different models for clear sky and cloudy sky conditions. In order to know which model must be applied, a procedure is followed, during the formulation, to determine whether an image pixel is clear or cloudy.

This is performed by applying a brightness threshold value, obtained through a minimum technique, at every pixel and at every hour using images relative to a series of days. In fact, the effect of a cloud within the field of view of a visible radiometer is that of increasing the measured brightness. When the pixel brightness is lower than the threshold value, the pixel is declared clear and the calculations are made using the clear air model. Otherwise the cloud model is used.

* Present address: AGIP S.p.A.-Drilling and Completion R&D Department, Milano, Italy.

This work is partly based on Mr. Noia's thesis. It is also intended as a contribution to Task 9 (Solar Radiation and Pyranometric Studies) of Solar Heating and Cooling Programme of the International Energy Agency.

Note that Raphael[6] and Raphael and Hay[7] stated that this procedure does not take into consideration variations in albedo with changing zenith angle and changing season. Thus, in their application, this part of the GDM approach was replaced by the T (see[1]) minimum brightness determination.

In the clear sky model, the incident solar radiation at ground level (see Fig. 1) is expressed as a function of the solar zenith angle $\theta$, in the following manner:

$$IG_s = IE_{\downarrow}T(\theta)[1 + \alpha_1(\theta)A] \qquad (1)$$

where $IE_{\downarrow}$ is the flux density of the downward solar radiation incident at the top of the atmosphere; $T(\theta)$ is the atmospheric transmissivity; $\alpha_1(\theta)$ is the reflection coefficient for diffuse radiation (dimensionless); and $A$ is the surface albedo.

The transmissivity is expressed as:

$$T(\theta) = (1 - \alpha)[1 - a(u_1)]$$

where $\alpha$ is the reflection coefficient for beam radiation (dimensionless); and $a(u_1)$ is the absorption coefficient for slant water vapour path $u_1$ for the solar zenith angle (dimensionless).

The reflection coefficients $\alpha$ and $\alpha_1$ as functions of the solar zenith angle and of the satellite zenith angle, respectively, have been taken from Coulson[8].

The surface albedo $A$ is evaluated (see Fig. 1) using the relationship:

$$IE_{\uparrow} = IE_{\downarrow}\alpha + IE_{\downarrow}(1 - \alpha)[1 - a(u_1)] \times [1 - a(u_2)](1 - \alpha_1)A \qquad (2)$$

where $IE_{\uparrow}$ is the solar radiation received by the satellite; and $a(u_2)$ is the absorption coefficient for slant water vapour path $u_2$ for the satellite zenith angle (dimensionless).

Note that $IE_{\uparrow}$ can be derived from the visible satellite measurements by means of a calibration procedure.

The absorption coefficients $a(u_1)$ and $a(u_2)$ appearing in eqns (1) and (2) were calculated through the analytical expressions derived by Paltridge[9] as a function of atmospheric precipitable water path $u$, solar zenith angle $\theta$, and satellite zenith angle $\phi$. The atmospheric precipitable water $u$ is parameterized via an empirical function of the surface dew-point temperature $t_d$, developed by Smith[10].

This clear sky model was subsequently revised by Gautier and Frouin[11] to include ozone absorption, aerosols effects, —parameterized according to Tanrè *et al.*[12]—and multiple reflections.

The new formula for the solar radiation reaching the ground is:

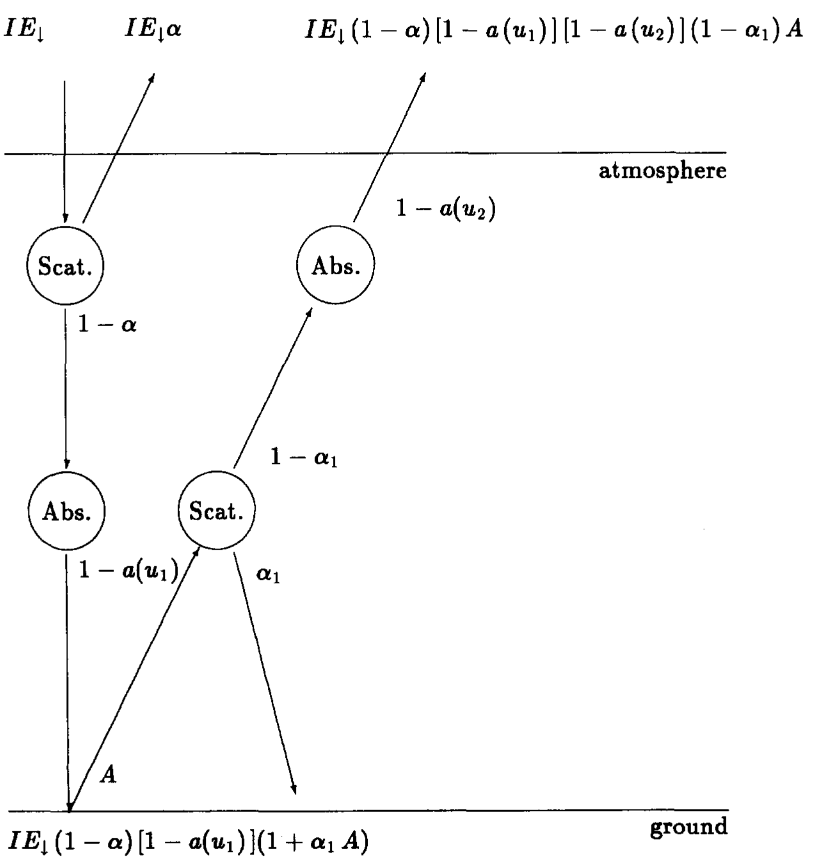


Fig. 1. Scheme of the clear air model developed by Gautier, Diak, and Mass[2].

$$IG_s = IE_{\downarrow}\frac{C_1\exp(-C_2)\cos\theta)(1-O_{z1}) \times (1-O_{z3}(1-a(u_1)]}{1-C_3A} \quad (3)$$

where $O_{z1}$ and $O_{z3}$ are ozone absorption coefficients; and $C_1$, $C_2$, and $C_3$ are empirical constants depending on surface visibility.

In the cloudy sky case, Gautier *et al.*[2]chose a simple treatment valid for stratiform low and middle clouds: these clouds are those which attenuate incident solar radiation the most.

The authors separately considered the absorption above and below the clouds for both the downwelling ($u_1$) and the upwelling paths ($u_2$) (see Figs. 1 and 2).

The incident shortwave at the earth surface is written as:

$$IG = IE_{\downarrow}(1-\alpha)[1-a(u_1)_t](1-A_c)(1-abs) \times [1-a(u_1)_b]. \quad (4)$$

The new variables introduced in this formula are $a(u_1)_t$, the absorption coefficient of short wave radiation above cloud (dimensionless); $A_c$, cloud albedo; $abs$, cloud absorption coefficient; and $a(u_1)_b$, absorption coefficient of short wave radiation below cloud (dimensionless).

The cloud albedo $A_c$ is evaluated through the formula (see Fig. 2):

$$\begin{aligned} IE_{\uparrow} = {} & IE_{\downarrow}\alpha + IE_{\downarrow}(1-\alpha)[1-a(u_1)_t] \\ & \times (1-\alpha_1)A_c[1-a(u_2)_t] + IE_{\downarrow}(1-\alpha) \\ & \times [1-a(u_1)_t](1-A_c)^2[1-a(u_1)_b] \\ & \times A(1-\alpha_1)[1-a(u_2)_t](1-abs)^2 \\ & [1-a(u_2)_b] \quad (5) \end{aligned}$$

where $a(u_2)_t$ is the absorption coefficient of short wave radiation above cloud (dimensionless); and $a(u_2)_b$ is the absorption coefficient of short wave radiation below cloud (dimensionless).

For low and middle clouds, the authors estimated an average of 30% of atmospheric water vapor above the cloud level [i.e., $a(u_1)_t \simeq 0.3a(u_1)$ and so on], assuming that most of water vapor is under cloud base.

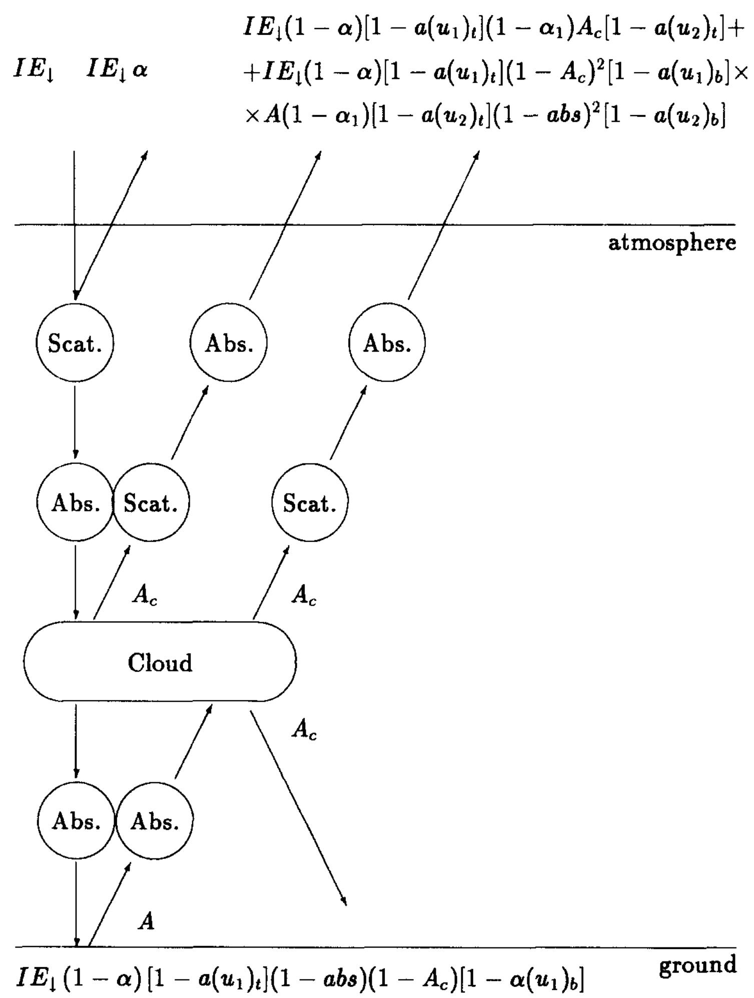


Fig. 2. Scheme of the cloudy atmosphere model developed by Gautier, Diak, and Mass[2].

The cloud absorption coefficient was estimated on the basis of the visible brightness measured by the satellite that indicates the presence (high values) or the absence (low values) of clouds. The authors chose a simple linear relationship between cloud absorption and visible brightness, ranging from zero for no cloud to a maximum of 0.2 for very deep clouds.

Considering the aerosol, ozone, and multiple reflection effects (see section 2), Gautier and Frouin[11] expressed the shortwave radiation incident at the surface in cloudy conditions with an equation similar to eqn (3):

$$IG = IE_{\downarrow} \frac{C_1 \exp(-C_2/\cos\theta)(1 - A_c - abs) \times (1 - O_{z1})(1 - O_{z3})[1 - a(u_1)]}{(1 - C_3 A)(1 - C_4 A_c)}. \quad (6)$$

### 3.2 *MDV method*

The GDM method was reconsidered in 1987 by Marullo, Dalu, and Viola[5]. The MDV method, similarly to the GDM method, presents two different models: a standard atmosphere model and a real atmosphere model.

The first is similar to the clear sky model of the GDM method, the only difference being that it also describes the effects due to a standard aerosol loading. The real atmosphere model differs from the GDM cloudy sky model in its description of the aerosol and cloud effects, assuming the presence of a reflecting nonabsorbing layer which simulates the effects of a nonstandard aerosol loading or the presence of clouds.

In the MDV model, the solar radiation reaching the ground under standard atmospheric conditions $IG_s$ is described with the same expression [eqn (1)] used by Gautier *et al.* However, in order to study the functional relationships $T(\theta)$ and $\alpha_1(\theta)$, Marullo *et al.* used a radiative transfer model developed by Schmetz[13], obtaining the two three-parameter empirical relationships:

$$T(\theta) = a_1\theta^{\gamma_1} + b_1 \quad (7)$$

$$\alpha_1(\theta) = a_2\theta^{\gamma_2} + b_2. \quad (8)$$

Note that this result was obtained assuming a temperature profile, a water vapor content and a three-layer aerosol content that describe Italian meteorological situations under clear sky conditions.

Instead of using eqn (2), Marullo *et al.*[5] evaluate the surface albedo $A$ in agreement with Gautier *et al.*[2], assuming that, in a standard atmosphere, a linear relationship exists between the albedo $A$ and the planetary albedo, defined as $PA = \frac{IE_{\uparrow}}{IE_{\downarrow}}$:

$$PA_s = a + bA. \quad (9)$$

In order to estimate the planetary albedo for a standard atmosphere, $PA_s$, Marullo *et al.* used a series of clear sky data covering the Italian peninsula, assuming the same value for the whole considered area and varying it with the solar zenith angle according to the three-parameter empirical rule:

$$PA_s = \alpha(90 - \theta)^{\gamma} + \beta. \quad (10)$$

If the planetary albedo $PA$ deduced from satellite images is within a 95% confidence limit, when compared with the standard atmosphere value $PA_s$ from eqn (10), the difference is assumed to be related to small variations in the surface albedo: the authors, in fact, follow the procedure relating to the standard atmosphere, described here.

If $PA$ is beyond that confidence limit, the deviation is considered as significant. In this case, Marullo, Dalu, and Viola assumed that any significant deviation of the flux measured at the satellite from the flux evaluated for standard conditions must depend mainly on a variation in the atmospheric particle loading.

The presence of these particles in the MDV method is represented by a nonabsorbing thin reflecting layer inside the atmosphere, assumed to be above the absorption and scattering mechanisms acting in the standard atmosphere (see Fig. 3). In fact, in the MDV model, the "reflection" of this layer may even be negative.

The solar irradiance at ground level is then given by:

$$IG = (IE_{\downarrow} - IE_{\uparrow,r})T(\theta)(1 + \alpha_1 A) \quad (11)$$

where $IE_{\uparrow,r}$ is the radiation back reflected by the aerosol layer. This formula may be compared with eqn (1).

The quantity $IE_{\uparrow,r}$ can be evaluated observing that the reflecting layer causes a variation of the radiance measured by the satellite, given by (see Fig. 3):

$$IE_{\uparrow} = IE_{\uparrow,r} + (IE_{\downarrow} - IE_{\uparrow,r}) \times [\alpha + A(1 + \alpha_1 A)T(\theta)T(\phi)(1 - A_c)] \quad (12)$$

where $T(\phi)$ is the total transmissivity of the atmosphere at the satellite angle of view, and $A_c$ is the albedo of the reflecting layer. Note that Marullo *et al.*[5] assume that $A_c = \frac{IE_{\uparrow,r}}{IE_{\downarrow}}$ in the METEOSAT spectral range.

### 3.3 *MR model*

The method developed by Möser and Raschke[14] was part of Project F of the European Community Solar Energy Research and Development Programme.

This method, applied to METEOSAT images, is based on a radiative transfer model more complex than the simple ones used in the other physical methods: for instance, it needs the inclusion of more parameters in order to describe the state of the atmosphere. Furthermore, the MR method also requires the use of the thermal infrared data, used to estimate the cloud top height, and an input variable of the radiative transfer model. In addition, the MR method does not require calibration of the satellite data.

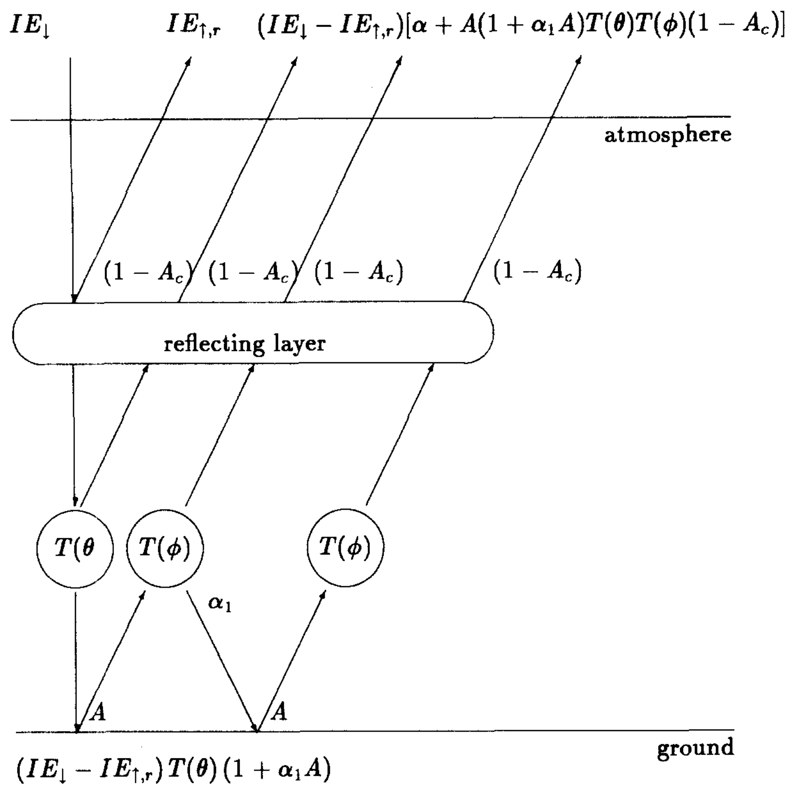


Fig. 3. Scheme of the real atmosphere model developed by Marullo, Dalu, and Viola[5].

Möser and Raschke used the radiative transfer model, developed by Kerschgens *et al.*[15], which is based on a two-stream approximation. The main variables used in this model are the solar zenith angle $\theta$, the cloud top height $h_c$, —estimated from thermal infrared satellite data—the optical depth of clouds $\tau_c$, and the ground albedo $A$. Additional information required by the model includes boundary layer structure (containing the aerosol effects), climatological profiles of temperature, pressure, humidity, and ozone and cloud droplet size distribution.

Calculations performed with this model have shown that solar irradiance at ground level depends to a much lesser extent on absorption by aerosols, water vapor, and ozone than on cloud coverage. Möser and Raschke expressed solar irradiance in the following manner:

$$IG = IG_0 \times IG_n + (1 - IG_n)IG_c \qquad (13)$$

where $IG_0$ = solar irradiance at ground level under clear sky conditions, when the irradiance reaches its maximum value; $IG_c$ = solar irradiance at ground in overcast conditions, when the irradiance reaches its minimum value; and $IG_n$ is the normalized irradiance at ground level.
Note that the quantity $IG_n$, also called fractional cloud cover, can assume values between 0 in overcast conditions and 1 in clear sky conditions. Möser and Raschke observed that $IG_n$ is mainly a function of the solar zenith angle and the optical cloud depth:

$$IG_n = IG_n(\theta, \tau_c).$$

It is possible to write the flux density of solar radiation emerging from the atmosphere in a manner analogous to eqn (13), as follows:

$$IE_{\uparrow} = IE_{\uparrow,c} \times IE_{\uparrow,n} + (1 - IE_{\uparrow,n})IE_{\uparrow,0} \qquad (14)$$

where $IE_{\uparrow,c}$ is the flux density of solar radiation emerging from the atmosphere in overcast conditions, when it reaches its maximum value; $IE_{\uparrow,0}$ is the flux density of solar radiation emerging from the atmosphere in clear sky conditions, when it reaches its minimum value; and $IE_{\uparrow,n}$ is the normalized flux density of solar radiation emerging from the atmosphere.
Note that $IE_{\uparrow,n}$ is equal to 0 in clear sky conditions and 1 in overcast conditions.

It is evident that when the cloud optical depth $\tau_c$ increases, $IE_{\uparrow,n}$ also increases while $IG_n$ decreases; the opposite occurs when $\tau_c$ decreases. For these reasons, Möser and Raschke assumed a direct dependence of $IG_n$ on $IE_{\uparrow,n}$:

$$IG_n(\theta, \tau_c) = IG_n(\theta, IE_{\uparrow,n}).$$

Möser and Raschke assumed that the solar irradiance at ground level in overcast conditions $IG_c$ is equal to zero. With this assumption, eqn (13) may be written as:

$$IG = IG_0 \times IG_n(\theta, IE_{\uparrow,n}). \qquad (15)$$

The value of $IE_{\uparrow,n}$ can be estimated from the visible satellite data. The value of $IG_0$ is wholly determined by the radiative transfer model, which also determines $IG_n$ as a function of $IE_{\uparrow,n}$.

Möser and Raschke correlated $IE_{\uparrow,n}$ to the visible data provided by the satellite by using the following equation:

$$IE_{\uparrow,n} = \frac{B - B_0}{B_c - B_0} \quad (16)$$

where $B$ is the current digital counts provided by the satellite; $B_0$ and $B_c$ are the minimum (corresponding to clear sky conditions) and the maximum (corresponding to overcast conditions) digital counts respectively; and $B_0$ and $B_c$ were determined from sequences of images covering 15–30 days.

### 3.4 *DDK method*

The method developed by Dedieu, Deschamps, and Kerr[4] describes the solar irradiance at ground level by means of a unique expression valid for both clear and cloudy conditions.

The final result was obtained by putting together a model for clear sky conditions and a model in which only the effect of clouds on solar radiation is considered. In fact, the authors assumed that the attenuation of solar radiation due to the atmospheric components, except clouds, is the same both in clear and cloudy conditions.

Dedieu *et al.*, assuming that molecular transmission under cloudy conditions is about the same as under clear sky conditions, combine both the cloud effects and atmospheric transmission in the unique equation:

$$IG = IE_{\downarrow}T(\theta)\frac{1 - PA}{1 - A}. \quad (17)$$

where $T(\theta)$ is the sky transmission factor, accounting for gaseous absorption, Rayleigh and Mie scattering; $PA = \frac{IE_{\uparrow}}{IE_{\downarrow}}$ is the planetary albedo; and $A$ is the surface albedo.

The authors of this model assumed that $T(\theta) \simeq T_0(\theta)$, where $T_0(\theta)$ is the clear sky transmission factor. This assumption was justified as follows:

- most of the ozone absorption is above the clouds and
- saturation of water vapor bands makes water vapor absorption rather insensitive to increase due to the presence of clouds (Davies *et al.*[16]).

Dedieu *et al.*[4] computed $T_0(\theta)$ using the formulae of Lacis and Hansen[17] for atmospheric components, together with the Tanrè *et al.*[12] radiative transfer model for aerosol effects, using standard midlatitude mean values for the climatological parameters.

The albedoes $PA$ and $A$, determined from satellite data, are defined by DDK as albedoes observable in an atmospheric window with no gaseous absorption, while the actual planetary albedo would be smaller because of molecular absorption. The parameter $PA$ is computed, after a calibration procedure, as the ratio between the solar radiation received by the satellite and the extraterrestrial radiation. $A$ is determined by a minimum technique similar to that used by Gautier *et al.*[2].

Cloud scattering is described by the third factor on the right hand side of eqn (17) assuming the isotropy of the radiance reflected by the cloud layer and the surface. The DDK authors also assumed multiple reflections between the cloud base and the ground (see Fig. 4).

Equation (17) clearly shows that the effects of cloud variability on $IG_c$ can be simply derived from continuous monitoring of the planetary albedo $PA$ as viewed by the satellites. Nevertheless, the surface albedo $A$ must be known. In fact, $A$ was determined by the authors as the minimum of the planetary albedo observed by the satellite in a time series of images, long enough to have a good probability of cloud-free conditions.

Note that eqn (17) must be treated with caution over highly reflective surfaces, such as snow covered areas, where $A$ may be close to 1. This is a consequence of the fact that clouds are difficult to distinguish from snow.

Finally, note that a significant aerosol effect is implicitly included in eqn (17). In fact, the main aerosol effect is that of increasing solar radiation back-scatter towards the satellite and, as a consequence, the planetary albedo $PA$. Therefore, this model treats a strong concentration of aerosol as a cloud.

## 4. COMPARISON BETWEEN MODELS

All the models described were tested both by their authors and by other scientists using experimental data.

### 4.1 *GDM method*

In the original application of their model, Gautier *et al.*[2] found that clear day estimations were within 5% of the mean measured radiation, while the estimations for cloudy and completely overcast days were within 14% and 15%, respectively, of the mean measured radiation. For all days, the combined estimations were within 9% of the mean measured radiation.

In their application, Raphael[6] and Raphael and Hay[7] found that this model on average overestimated the measured radiation under clear skies and underestimated the measured radiation under partly cloudy conditions; the model also overestimated the measured radiation under overcast conditions, except in summer. The discrepancy in partly cloudy conditions is attributed (as in the T model: see[1]) to the problem of defining the correct cloud threshold; the overestimation in overcast conditions is again attributed to the inadequate handling of cloud absorption.

On the other hand, the GDM model applied by Raphael and Hay[7] to a series of nine days demonstrated its superior performance over the models HH and T (see[1]) in simulating hourly radiation under partly cloudy and overcast conditions.

After their revision of the T model (see reference[1]), the bias for clear sky conditions was on av-

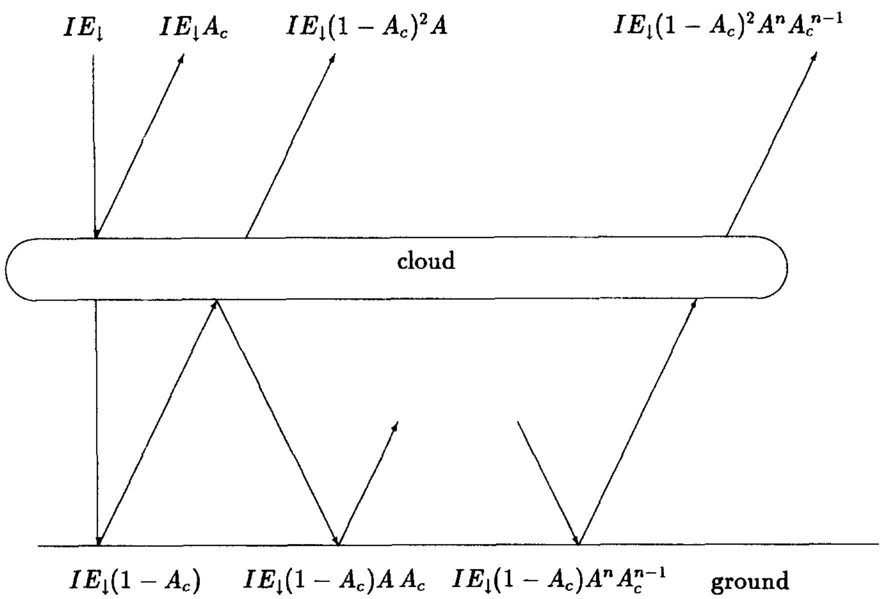


Fig. 4. Scheme of the cloudy atmosphere model developed by Dedieu, Deschamps, and Kerr[4].

erage reduced to zero, with a concomitant decrease in the rms error to a value near ± 5%. This result obtained with the T model was, in this experiment, better than those obtained with HH (see[1]) and GDM models.

### 4.2 *MDV method*

The authors of this method claim that in most cases the difference between the insolation measured by pyranometers and the corresponding results from the model was less than 10% for a wide variety of sky conditions.

### 4.3 *MR model*

The results obtained by Möser and Raschke were affected by a standard error of about 72 W $m^{-2}$ under all sky conditions, decreasing to 60 W $m^{-2}$ in overcast conditions while reaching 82 W $m^{-2}$ in broken cloudiness.

### 4.4 *DDK method*

The authors estimated solar radiation to within an error of ± 19.5% on an hourly basis. This error decreased when averaging the results over a period of one month. In this case the standard error was within ± 6.7%, corresponding to about ± 109 W $m^{-2}$. The absolute errors were minimum for clear sky conditions (± 50 W $m^{-2}$) and maximum for an overcast sky (± 150 W $m^{-2}$), with an intermediate value (± 100 W $m^{-2}$) for partly cloudy skies.

## 5. CONCLUSIONS

We have concluded the analysis and evaluation—started in a previous paper[1]—of the main methods used to estimate global solar irradiance at ground level from satellite data.

In this paper we have considered four methods based on the physical description and parameterization of atmosphere conditions.

Two of these methods (GDM and MDV) use distinct models for clear sky and cloudy sky conditions ("standard" and "real" atmosphere in the MDV model). In the clear sky case, the two approaches are conceptually very similar, both principally taking into account the effects of precipitable water on the absorption coefficients (standard aerosol loading is also considered by MDV). In the second case, the GDM method separately considers the absorption characteristics above and below the clouds, whereas the MDV model introduces a nonabsorbing thin reflecting layer above the standard atmosphere. Both models give estimations with errors to within 5% in the clear sky case and with greater errors, about 15%, in the case of cloudy skies.

A third model (MR) assumes cloud top height, optical depth of the cloud, and ground albedo as main atmospheric parameters. Its performance ranges from a standard error in the ground global irradiance of 60 W $m^{-2}$ with overcast conditions to 82 W $m^{-2}$ with broken cloudiness.

The fourth model (DDK) uses a unique expression for both clear and cloudy sky conditions, putting together a clear sky model with absorption with a cloudy sky model with no absorption and no molecular scattering. The estimation errors obtained on an hourly basis are ± 19.5%, which become ± 6.7% when averaging the results over a period of one month.

By comparing statistical and physical models, some general conclusions can be drawn. The statistical approaches have some advantages due to their simplicity. Unfortunately, their relative operational efficiency is penalized in terms of their lack of generality, since there is no prior guarantee that the regression equation coefficients determined for one location will still be valid for another.

On the other hand, physical models have the advantage of conceptually describing the status of the atmosphere, and thus do not depend on a particular region and may be applied anywhere. Moreover, they do not need solar radiation data measured at the earth's surface. Nevertheless, the physical methods do need accurate and updated instrument calibration in order to allow the conversion of the digital count values, provided by the satellite, into the corresponding flux density of the upward solar radiation emerging from the atmosphere. Furthermore, validity of the models is often dependent on the satellite itself and the field of view for which they have been developed. Finally, they require supplementary physical information about atmosphere status, which is not always available.

Both the physical and the statistical methods share the difficulty that the—space and time—scales of satellite images differ from those of ground-based measurements.

Finally, we have to point out that most methods—both statistical and physical—have been tested in particular cases and/or localities: a very general comparison cannot be made at present. Even the existing comparisons[6,7,18,19] only consider a few models and/or a few areas at a time. In fact, it is likely that the "best" method to use depends very much on both the status of the atmosphere and the studied area.

*Acknowledgments*—Comments and suggestions by Dr. A. Zelenka, Swiss Meteorological Institute, Zürich, were greatly appreciated.

## NOMENCLATURE

- $A$ surface albedo of earth
- $A_c$ cloud albedo (GDM and MDV)
- $abs$ cloud absorption (GDM)
- $a(u_1)$ and $a(u_2)$ absorption coefficients for short wave radiation (dimensionless) (GDM)
- $a(u_1)_t$ and $a(u_2)_t$ absorption of short wave radiation above cloud (dimensionless) (GDM)
- $a(u_1)_b$ and $a(u_2)_b$ absorption of short wave radiation below cloud (dimensionless) (GDM)
- $B$ brightness, i.e., digital counts measured by the satellite, proportional to the upward solar radiation emerging from the atmosphere
- $B_c$ overcast sky threshold brightness (MR)
- $B_0$ clear sky threshold brightness (MR)
- $C_1, C_2, C_3, C_4$ empirical coefficients related to the surface visibility (GDM)
- $h_c$ cloud top height (MR)
- $IE_\uparrow$ flux density of the upward solar radiation emerging from atmosphere and received by the satellite, W m$^{-2}$
- $IE_{\uparrow,c}$ flux density of solar radiation emerging from the atmosphere in overcast conditions (MR)
- $IE_{\uparrow,0}$ flux density of solar radiation emerging from the atmosphere in clear sky conditions (MR)
- $IE_{\uparrow,n}$ normalized flux density of solar radiation emerging from the atmosphere (dimensionless, $0 \le IE_{\uparrow,n} \le 1$) (MR)
- $IE_{\uparrow,r}$ the radiation back reflected by the aerosol layer (MDV)
- $IE_\downarrow$ flux density of the downward solar radiation incident on the atmosphere, W m$^{-2}$
- $IG$ flux density of the downward solar radiation incident on the earth's surface, W m$^{-2}$
- $IG_c$ flux density of the downward solar radiation incident on the earth's surface under overcast atmospheric conditions (C, MR, and DDK), W m$^{-2}$
- $IG_n$ normalized flux density of the downward solar radiation incident on the earth's surface (dimensionless, $0 \le IG_n \le 1$) (MR)
- $IG_s$ flux density of the downward solar radiation incident on the earth's surface under standard atmospheric conditions (GDM and MDV), W m$^{-2}$
- $IG_0$ flux density of the downward solar radiation incident on the earth's surface under clear atmospheric conditions (MR), W m$^{-2}$
- $O_{z1}, O_{z3}$ ozone absorption coefficients (GDM)
- $PA$ planetary albedo, i.e., $\frac{IE_\uparrow}{IE_\uparrow}$
- $PA_s$ planetary albedo under standard atmospheric conditions (MDV)
- $T(\phi)$ atmospheric transmittance for ascending solar radiation, i.e., $\frac{IE_\uparrow}{A \times IG}$ (MDV)
- $t_d$ surface dew point temperature (GDM)
- $T(\theta)$ atmospheric transmittance for descending solar radiation, i.e., $\frac{IG}{IE_\downarrow}$
- $T_0(\theta)$ clear sky transmission factor (DDK)
- $u$ atmospheric precipitable water (GDM)
- $u_1$ slant water vapor paths for solar zenith angle (GDM)
- $u_2$ slant water vapor paths for satellite zenith angle (GDM)
- $\alpha$ reflection coefficient for beam radiation (dimensionless) (GDM)
- $\alpha_1$ reflection coefficient for diffuse radiation (dimensionless) (GDM and MDV)
- $\theta$ solar zenith angle
- $\phi$ satellite zenith angle
- $\tau_c$ optical depth of clouds (MR)

## REFERENCES

1. M. Noia, C. F. Ratto, and R. Festa, Solar irradiance estimation from geostationary satellite data: I. Statistical models. *Solar Energy* **51**, 449–456 (1993).
2. C. Gautier, G. Diak, and S. Masse, A simple physical model to estimate incident solar radiation at the surface from GOES satellite data, *J. Appl. Meteorol.* **36** (1980).
3. W. Möser and E. Raschke, Mapping of global radiation and of cloudiness from meteosat image data, *Meteorol. Rdsch.* **36** (April 1983).
4. G. Dedieu, P. Y. Deschamps, and Y. H. Kerr, Satellite estimation of solar irradiance at the surface of the Earth and of surface albedo using a physical model applied to meteosat data, *J. Clim. Appl. Meteorol.* **26** (1987).
5. S. Marullo, G. Dalu, and A. Viola, Incident short-wave radiation at the surface from meteosat data, *Il Nuovo Cimento*, **10C**, 77–90 (1987).
6. C. Raphael, Models for estimating solar irradiance at the Earth's surface from satellite data: An initial assessment, Technical Report, Atmospheric Environment Service, Downsview Ontario (1983). (Unpublished manuscript.)
7. C. Raphael and J. E. Hay, An assessment of models which use satellite data to estimate solar irradiance at the Earth's surface, *J. Clim. Appl. Meteorol.* **23**, 832–844 (1984).
8. K. L. Coulson, Characteristics of the radiation emerging from the top of a Rayleigh atmosphere, 1 and 2, *Planet. Space Sci.* **1**, 256–284 (1959).
9. G. W. Paltridge, Direct measurements of water vapour

absorption of solar radiation in the three atmosphere, *J. Appl. Meteorol.* **30,** 156–160 (1973).

10. W. L. Smith, Note of the relationship between total precipitable water and surface dew point, *J. Appl. Meteorol.* **5,** 726–727 (1966).
11. C. Gautier and R. Frouin, Downward longwave irradiance at the ocean surface using satellite data. In: *Proc. of the 3rd International Collection on Spectral Signature of Objects in Remote Sensing,* Les Arcs, France (1985).
12. D. Tanrè, M. Herman, P. Y. Deschamps, and A. De Leffe, Atmospheric modeling for space measurements of ground reflectances, including bidirectional properties, *Appl. Optics* **18,** 3587–3594 (1979).
13. J. Schmetz, On the parameterization of the radiative properties of broken clouds, *Tellus A.* **36,** 417 (1984).
14. W. Möser and E. Raschke, Incident solar radiation over Europe estimated from meteosat data, *J. Appl. Meteorol.* **23,** 166–170 (1984).
15. M. Kerschegens, U. Pilz, and E. Raschke, A modified two stream approximation for computations of the solar radiation budget in a cloudy atmosphere, *Tellus A* **30** (1978).
16. R. Davies, W. L. Ridgway, and K. Kyung-eak, Spectral absorption of solar radiation in cloudy atmosphere: a 20 $cm^{-1}$ model, *J. Appl. Meteorol.* **41,** 2126–2137 (1984).
17. A. Lacis and J. E. Hansen, A parametrization for the absorption of solar radiation in the Earth's atmosphere, *J. Atmospheric Science* **31,** 118–133 (1974).
18. G. Czeplak, M. Noia, and C. F. Ratto, An assessment of a statistical method to estimate solar irradiance at the Earth's surface from geostationary satellite data, *Renewable Energy* **1,** 737–743 (1992).
19. A. Zelenka *et al.,* Techniques for supplementing solar radiation network data, Technical Report, International Energy Agency, #IEA-SHCP-9D-1, Swiss Meteorological Institute, Krahbuhlstr. 58, CH-8044 Zurich, Switzerland (1992).