A Parametrization of Solar Energy Disposition in the Climate System

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ABSTRACT During the past decade a class of climate models of reduced complexity, called Earth system Models of Intermediate Complexity (EMICs), has been developed. Some of these models employ an energy and moisture balance model (EMBM) as the atmospheric component. However, the solar energy disposition (SED) in the subcomponents of these climate models using an EMBM has never been parametrized in a systematic manner. In this paper, the SED, which is a measure of the amount of solar radiation absorbed in the atmosphere, absorbed at the surface and reflected to space, is first expressed as functions of the surface albedo and the integrated atmospheric reflectivity, transmissivity, absorptivity and cloud amount for a one-layer atmosphere which includes a cloud region and aerosols. Then an atmospheric radiative-convective model is used to parametrize the integrated atmospheric reflectivity and transmissivity in terms of cloud optical depth, aerosol optical depth, precipitable water, and solar zenith angle. Next, the present-day climatology of the SED is calculated using the climatological data (for cloud amount and optical depth, aerosol optical depth, precipitable water and surface albedo) from the International Satellite Cloud Climatology Project (ISCCP), ECMWF 15-year Reanalysis (ERA-15) and the Pathfinder Atmosphere (PATMOS). Since cloud amount data are used from three independent sources, three SEDs are in fact calculated and tested against the SED derived from satellite data. The calculated SEDs are in good agreement with the SED derived from satellite data; thus the parametrized SED presented here is recommended for use in climate models which employ an EMBM or a one-layer atmosphere model.

RÉSUMÉ [traduit par la rédaction] Au cours de la dernière décennie, on a mis au point une classe de modèles climatiques de complexité réduite appelés Modèles de système terrestre de complexité intermédiaire (EMIC). Certains de ces modèles utilisent un modèle d’équilibre de l’énergie et de l’humidité (MEEH) en guise de composante atmosphérique. Cependant, la disposition de l’énergie solaire (DES) dans les sous-composantes de ces modèles climatiques utilisant un MEEH n’a jamais été paramétrée de façon systématique. Dans cet article, la DES, qui décrit les quantités de rayonnement solaire absorbé dans l’atmosphère, absorbé à la surface et réfléchi vers l’espace, est d’abord exprimée en fonction de l’albédo de la surface et de la réflectivité, transmissivité, absorptivité et nébulosité atmosphériques intégrées pour une atmosphère à une seule couche qui inclut une région de nuages et des aérosols. Puis, on utilise un modèle atmosphérique de rayonnement et de convection pour paramétrer la réflectivité et la transmissivité atmosphériques intégrées en fonction de l’épaisseur optique des nuages, de l’épaisseur optique des aérosols, de l’eau précipitable et de l’angle zénithal du soleil. Ensuite, on calcule la climatologie du jour courant de la DES d’après les données climatologiques (sur l’étendue et l’épaisseur optique des nuages, l’épaisseur optique des aérosols, l’eau précipitable et l’albédo de la surface) du Projet international d’établissement d’une climatologie des nuages à l’aide de données satellitaires (ISCCP), d’après la réanalyse de 15 ans (ERA-15) du Centre européen pour les prévisions météorologiques à moyen terme (CEPMMT) et d’après les données du projet Pathfinder Atmosphere (PATMOS). Puisqu’on utilise des données de nébulosité provenant de trois sources indépendantes, on calcule en fait trois DES que l’on compare à la DES dérivée des données satellitaires. Les DES calculées s’accordent bien avec la DES dérivée des données satellitaires. Par conséquent, on recommande d’utiliser la DES paramétrée présentée ici dans les modèles climatiques employant un MEEH ou un modèle atmosphérique à une couche.

1 Introduction
The radiation schemes used in present weather forecast and climate models are as plentiful as the designs in a kaleidoscope. Even in simple climate models which employ an energy balance model (EBM) for the atmospheric component, there is no agreement in the literature on the best way to parametrize radiative processes, including those associated with solar radiation. The main purpose of this paper is to develop a rigorous solar radiation scheme which is suitable
for EBMs. Also, this scheme together with present-day climatological data for clouds, aerosols, precipitable water and surface albedo are used to derive the solar energy disposition (SED) which is tested against the SED derived from satellite data.

EBMs were first developed independently by Budyko (1969) and Sellers (1969). The zonally averaged surface temperature was predicted by considering the incoming solar insolation, outgoing longwave radiation and the energy convergence/divergence induced by the atmosphere–ocean poleward heat transport. This kind of one-dimensional EBM was further developed into a two-dimensional (in the horizontal) EBM which resolves a seasonal cycle and a land–sea configuration (for example, see North et al. (1983)). Since the atmosphere and the underlying surface components such as the ocean, sea ice, land surface and continental ice are not treated separately, the surface temperature is used to represent the thermal state of the atmosphere–surface system. Hence, the incoming solar radiation in this type of EBM is the solar insolation absorbed by the whole atmosphere–surface system, which can be parametrized as a function of the solar insolation at the top of the atmosphere (TOA) and the planetary albedo.

During the past decade, a class of climate models of reduced complexity called Earth system Models of Intermediate Complexity (EMICs) (Claussen et al., 2002) was developed. Some of these models employ an EBM and a moisture balance model (MBM) for the atmospheric component. These models have been used to simulate global warming (Stocker and Schmittner, 1997), the climate of the Last Glacial Maximum (Weaver et al., 1998), the millennial climate variability during a glacial period (Schmittner et al., 2002), and the last glacial inception and rapid ice sheet growth (Wang and Mysak, 2002). In these models, EBMs and MBMs are explicitly coupled to the other climate components, such as the ocean, sea ice, land surface, continental ice, and biosphere (Marchal et al., 1998; Weaver et al., 2001; Meissner and Gerdes, 2002; Wang and Mysak, 2002). Thus, the parametrizations of the solar insolation absorbed by the atmosphere cannot be the same as those in Budyko (1969), Sellers (1969) and North et al. (1983). They have to be modified to take into account the other climate components (Stockert et al., 1992; Schmittner and Stocker, 2001; Fanning and Weaver 1996; Weaver et al., 2001; Wang and Mysak, 2000; Meissner and Gerdes, 2002). However, the parametrizations of the solar insolation absorbed by the atmosphere and the surface in these models are quite arbitrary. Furthermore, the parametrization of the planetary albedo used in all these EBMs was fitted from limited observational data and is thus too simple. The formulae required to describe the physical mechanisms of radiative processes and to analyse further the atmospheric compositions which influence the current climate or influenced past paleoclimates need to be more complex.

At present, since almost all climate forcings have been found and estimated, and satellite observations of the Earth system radiation budget are available (for example, see Li and Leighton (1993)), it is now possible to propose a parametrization of SED in the climate system for those climate models which employ an atmospheric EBM and MBM to simulate the energy and hydrologic cycle of the atmosphere. Furthermore, since SEDs are very different even in several climatological datasets and Atmospheric General Circulation Models (AGCMs) (Li et al., 1997), another purpose of this paper is to derive the SED using present-day climatological observations of cloud amount and optical depth, aerosol optical depth, precipitable water and surface albedo together with a radiative-convective model (RCM). This calculated SED is then compared with the SED derived from the Earth Radiation Budget Experiment (ERBE) (Li and Leighton, 1993; Li et al., 1997) and the Langley Eight-Year Shortwave and Longwave Surface Radiation Budget Dataset (hereafter Langley Dataset) (Gupta et al., 1999).

The structure of this paper is as follows. A review of the SED parametrizations in the above-mentioned EMICs is given in Section 2. The SED parametrization formulae are presented in Section 3. In Section 4, the parametrized SED is calculated using data from some present-day climatological datasets of clouds, aerosols and precipitable water and is tested against the SED derived from satellite data. A summary and discussion are given in Section 5.

2 Reviews
In this section, the parametrizations of the SED in four different EBMs are reviewed. Since the SED in Harvey (1988) is explicitly calculated using a radiation transfer scheme (i.e., is not parametrized), his SED calculation is not reviewed here. Note that we discuss the papers of Stockert et al. (1992) and Schmittner and Stocker (2001) together because the latter paper further develops the EBM used in the former. Also, we discuss Fanning and Weaver (1996) and Weaver et al. (2001) together because the same EBM is used.

a Stockert et al. (1992) and Schmittner and Stocker (2001)
Stockert et al. (1992) use $\kappa Q_{\text{short}}$ to represent the solar insolation absorbed by the atmosphere, where $\kappa = \frac{20}{71}$ and $Q_{\text{short}}$ is the monthly mean solar insolation absorbed by the atmosphere–surface system and is obtained from Table 4c in Stephens et al. (1981). The solar insolation absorbed by the underlying ocean surface is taken to be $(1 - \kappa)Q_{\text{short}}$. Thus, the total solar insolation absorbed by the atmosphere and the underlying surface is $Q_{\text{short}}$. In Schmittner and Stocker (2001), $Q_{\text{short}}$ is taken from the ERBE data (ERBE, 1990) as modified by Trenberth (1997) (see Eq. (3) in Schmittner and Stocker (2001)). The solar insolation absorbed by the underlying surface is not explicitly specified in Schmittner and Stocker (2001). However, they mention that the simple thermodynamic sea-ice component used in Wright and Stocker (1993) is employed. In Wright and Stocker (1993), the solar insolation absorbed by sea ice is calculated as $(1 - \kappa)S$ (see their Eq. (5)), where $S$ is not explained. If $S$ is the same as $Q_{\text{short}}$, the surface albedo of sea ice does not explicitly appear. This is probably the reason why Schmittner and Stocker (1999) parametrized the ice–albedo feedback by prescribing...
the change in the planetary albedo as a function of surface air temperature (Eq. (21) in Schmittner and Stocker (1999)).

**b Fanning and Weaver (1996) and Weaver et al. (2001)**

Fanning and Weaver (1996) calculated the solar insolation absorbed by the atmosphere using \( \frac{S_o}{4} S(1 - \alpha)C_A \), where \( S_o \) is the solar constant, \( S \) is the seasonal distribution of the solar insolation at the TOA (Berger, 1978), \( \alpha \) is the latitudinally dependent planetary albedo (also varying with the time of year (Weaver et al., 2001)) from Graves et al. (1993), and \( C_A \) is an absorption coefficient for the atmosphere (a latitudinal profile ranging from 0.32 to 0.44, as shown in Fig. 3c in Fanning and Weaver (1996)). We note that \( C_A \) has a constant value of 0.3 in Weaver et al. (2001). The solar insolation absorbed by the underlying surface is represented as \( \frac{S_o}{4} S(1 - \alpha)(1 - C_A) \) for the ocean (Fanning and Weaver, 1996). In Weaver et al. (2001) the latter absorption formula is also used for land surface and sea ice. Thus, the total solar insolation absorbed by the atmosphere and the underlying surface is \( \frac{S_o}{4} S(1 - \alpha) \) in both papers. If the planetary albedo \( \alpha \) takes the present-day profiles, the SED cannot change if there are either imposed or internal climate changes. Also, the surface albedo never appears in the absorbed solar insolation term at the surface and hence the ice–albedo feedback is not explicitly taken into account in their model. This latter deficiency is probably the reason why they need to increase the planetary albedo by hand when large-scale continental ice appears and why the surface air temperature change over Greenland, due to the thermohaline circulation change, is too small (Schmittner et al., 2002).


Wang and Mysak (2000) calculated their solar insolation absorbed by the atmosphere as \( Q_{SSW}(1 - \alpha_{a})\), where \( Q_{SSW} \) is the seasonal distribution of the solar insolation at the TOA (Berger, 1978), \( \alpha_{a} \) is the atmospheric albedo (taken as 0.3 + 0.1 sin^2\( \phi \), where \( \phi \) is the latitude), \( a \) is the atmospheric transmissivity of solar radiation (taken as 0.65), and \( B \) is the surface albedo. The solar insolation absorbed by the surface is \( Q_{SSW}(1 - \alpha_{a})a(1 - B) \). The total solar insolation absorbed by the atmosphere and the underlying surface is \( Q_{SSW}(1 - \alpha_{p}) \), where \( \alpha_{p} = \alpha_{a} + (1 - \alpha_{a})a^2B \). In this model, \( \alpha_{p} \) is meant to take into account the reflection processes due to clouds, aerosols, water vapour, CO2, and other gases. It is also assumed that the incident solar radiation is reflected by the atmosphere first; the remainder enters the atmosphere and is absorbed there with an absorption coefficient of 0.35. The portion of incident solar radiation reflected by the surface is partly absorbed in the atmosphere (also with an absorption coefficient of 0.35) without consideration of the atmospheric reflectivity; the rest is reflected to space. Although the ice–albedo feedback is taken into account, the determination of the zonally averaged atmospheric albedo is arbitrary, and the total solar insolation absorbed by the atmosphere is probably too large over regions with high surface albedo.

**d Meissner and Gerdes (2002)**

Meissner and Gerdes (2002) calculated the solar insolation absorbed by the atmosphere as \( (2(\alpha_{a} + \alpha_{c}) + 2a_{s}(1 - (\alpha_{a} + \alpha_{c} + A_{a}))) + A_{c} - 1)Q_{SSW}^{top,down} \), where \( Q_{SSW}^{top,down} \) is the solar insolation incident at the TOA (Berger, 1978), \( \alpha_{a} \) is the albedo of a cloudless atmosphere (0.08), \( \alpha_{c} \) is the albedo of the clouds (a function of longitude and latitude obtained from Haney (1971)), \( A_{a} \) is the albedo of the surface, and \( A_{c} \) is the absorptivity of the atmosphere (0.18). The values of \( \alpha_{a} \) and \( A_{a} \) are taken from London (1957). The solar insolation absorbed by the atmosphere is \( (1 - \alpha_{a})Q_{SSW}^{top,down} \), and the underlying surface is \( (1 - \alpha_{p})Q_{SSW}^{top,down} \), where \( \alpha_{p} = \alpha_{a} + \alpha_{c} + A_{a}(1 - (\alpha_{a} + \alpha_{c} + A_{a})) \).

### 3 An SED parametrization in a one-layer radiation model

We consider a one-layer atmospheric radiation model of the type used by Rasool and Schneider (1971), Sellers (1973) and Jentsch (1991) (see Fig. 1). For this model we assume that:

1. the solar beam entering one region cannot enter another region for any reason;
2. the integrated atmospheric reflectivity, transmissivity and absorptivity for the whole atmospheric column are \( r \) and \( a \) respectively for any clear sky region; \( r_{c} \) and \( a_{c} \), respectively for any cloudy region, for radiation incident from above, and that
   \[
   r + t + a = 1, \tag{1}
   \]
   \[
   r_{c} + t_{c} + a_{c} = 1; \tag{2}
   \]
3. the integrated atmospheric reflectivity, transmissivity and absorptivity for the whole atmospheric column for radiation incident from below are the same as those for radiation incident from above for both clear and cloudy sky conditions; and
4. the surface albedo is given by \( r_{s} \) (Note that iii) is valid when the atmosphere is vertically homogeneous. For vertically inhomogeneous cases, such as in Li et al. (1994), it was found that there exist differences in reflectivity between the vertical inhomogeneous cases and their plane parallel counterparts for overcast stratocumulus clouds. The difference can be as large as 10% for a large solar zenith angle. However, the differences are generally very small (around 1%). The cloud vertical inhomogeneity also causes a 7% increase in cloud absorption.)

On the basis of the above assumptions, and taking into account multiple reflections between the atmospheric layer and the surface (see Fig. 1), it follows that the disposition of unit incident solar insolation into the atmosphere, \( Q_{a} \), is given by
where \( Ac \) is the cloud amount (i.e., cloud fraction).

Similarly, the disposition of unit incident solar insolation at the surface, \( Q_s \), is given by

\[
Q_s = (1 - A_c) \left( r + \frac{r^2 A_e}{1 - r r_s} \right) + A_c \left( a_c + \frac{A_e a_c r_s}{1 - r_c r_s} \right),
\]

(3)

and the disposition of unit incident solar insolation that is reflected to space, \( Q_e \), is given by

\[
Q_e = (1 - A_c) \left( r + \frac{r^2 A_e}{1 - r r_s} \right) + A_c \left( a_c + \frac{A_e a_c r_s}{1 - r_c r_s} \right).
\]

(5)

It is easy to verify that

\[
Q_e + (Q_a + Q_s) = 1,
\]

(6)
i.e., the solar insolation reflected to space, \( Q_e \), plus the sum of the solar insolation absorbed by the atmosphere and at the surface, \( (Q_a + Q_s) \), is equal to the incident unit solar insolation.

We now determine the values of \( r, t, r_c \) and \( t_c \) as functions of cloud optical depth, aerosol optical depth, precipitable water and solar zenith angle, using a one-dimensional RCM which we developed, but which is based on Wang et al. (1981), Shi (1992) and Liang and Wang (1997). Once \( r, t, r_c \) and \( t_c \) are known, \( a \) is determined by

\[
a = 1 - r - t \quad \text{and} \quad a_c = 1 - r_c - t_c.
\]

The broadband reflectivity and transmissivity of the atmosphere depend on the concentrations of trace gases, the different types of aerosols and the different types of clouds. The shortwave radiation is calculated in 14 intervals from 0 to 4.00 microns. The method used to solve the radiative transfer equation with scattering and absorption processes is the delta-Eddington approximation as used in Briegleb (1992). The transmission of gases is calculated by the k-distribution method (exponential sum fitting). HITRAN-92 is used as the absorption database. The vertical distribution of aerosols is fitted from observational data (Hu et al., 2001). Aerosol optical properties are calculated from Mie scattering theory and the refractive indices of aerosols are obtained from Toon et al. (1976) and WCP-55 (1983). Since the effect of ozone change on \( r, t, r_c \) and \( t_c \) is very small, the simple formulae from Lacis and Hansen (1974) are employed to calculate the absorption by ozone.

In the following, the atmospheric reflectivity and transmissivity are decomposed into their climatological values (multi-year global annual means with solar zenith angle at 60°) plus perturbations caused by variations of atmospheric compositions, which are functions of space and time, and solar zenith angle. Thus,

\[
r = r_o + \delta r_o + \delta r_{\text{aerosol}}, Z
\]

(7)

\[
r_c = r_{\text{co}} + \delta r_{\text{aerosol}}, Z + \delta r_{\text{cloud}}, Z
\]

(8)
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\[
\delta r_a = -\frac{0.275 f(Z)}{f(Z) + 3.225}, \tag{11}
\]

where \(f(Z) = \cos(60^\circ)/\cos(Z)\).

For stratospheric aerosol perturbations, we obtain

\[
\delta r_{aerosol, Z} = 5.0659 \times 10^{-2} \tau_a f(Z) \tag{12}
\]
\[
\delta t_{aerosol, Z} = -6.7871 \times 10^{-2} \tau_a f(Z) \tag{13}
\]

and for tropospheric aerosol perturbations, we obtain

\[
\delta r_{aerosol, Z} = 0.1948 \tau_a f(Z) \tag{14}
\]
\[
\delta t_{aerosol, Z} = -0.3579 \tau_a f(Z) \tag{15}
\]

where \(\tau_a\) is the aerosol optical depth.

\[
\delta r_{cloud, Z} = 0.2858 \left( \sqrt{\tau_c Z} - \sqrt{\tau_{co}} \right) - 1.3330 \times 10^{-2} (\tau_c Z - \tau_{co}) + 5.2222 \times 10^{-2} \left( \frac{1}{\tau_c Z} - \frac{1}{\tau_{co}} \right) \tag{16}
\]

\[
\delta t_{cloud, Z} = -0.2652 \left( \sqrt{\tau_c Z} - \sqrt{\tau_{co}} \right) + 1.1881 \times 10^{-2} (\tau_c Z - \tau_{co}) + 1.2222 \times 10^{-2} \left( \frac{1}{\tau_c Z} - \frac{1}{\tau_{co}} \right) \tag{17}
\]

where \(\tau_c Z\) is the product of cloud optical depth \(\tau_c\) and \(f(Z)^{0.78}\) and \(\tau_{co}\) is the climatological value for cloud optical depth (see Table 2).

For precipitable water perturbations, we obtain

\[
\delta t_{H_2O, Z} = -3.8300 \times 10^{-2} \ln \frac{PW_f(Z)}{PW_o} + 2.1585 \times 10^{-2} \left( \sqrt{PW_f(Z)} - \sqrt{PW_o} \right), \tag{18}
\]

where \(PW\) is precipitable water (cm) and \(PW_o\) is the climatological value for precipitable water (cm) (see Table 2). We note that the second coefficient in Eq. (18) is only valid when \(PW\) is given in centimetres.

Thus, once the cloud amount and optical depth, precipitable water, aerosol optical depth and surface albedo are known, the SED consisting of \(Q_a\), \(Q_s\) and \(Q_e\) can be calculated by using the above formulae. In climate models which employ an EBM and an MBM or a one-layer atmospheric model, cloud amount and optical depth can be prescribed as the present-day climatology or parametrized. The aerosol optical depths can be obtained from observational data. Precipitable water is a predicted variable. Therefore, this parametrization of SED can be used in those climate models which employ an EBM and an MBM or a one-layer atmospheric model. In the following section, we will calculate the parametrized SED by using present-day climatological data for the above quantities and test the parametrized SED against the SED derived from satellite data.

4 Data and parametrized SED

a Data

In order to calculate the monthly mean SED for the present-day climatology, the precipitable water is obtained from the European Centre for Medium-Range Weather Forecasts Reanalysis (15 years from 1979 to 1993) (ERA-15). The cloud amount is obtained from three data sources: the Pathfinder Atmosphere Project (PATMOS) from 1981 to 2000, (Stowe et al., 2002); the International Satellite Cloud Climatology Project (ISCCP) from 1983 to 1999 (http://isccp.giss.nasa.gov/products/browsed2.html); and ERA-15; the cloud optical depth can be prescribed as the present-day climatology or parametrized. The aerosol optical depths can be obtained from ISCCP (http://isccp.giss.nasa.gov/products/browsed2.html); and the surface albedo is also obtained from ISCCP (from 1992 to 1996) (http://isccp.giss.nasa.gov/products/browsesurf.html); and the global tropospheric aerosol optical depth is from Pinker and Laszlo (1992). The stratospheric aerosols are not taken into account in this study. The monthly mean \(\cos(Z)\) is calculated by Berger (1978). In the calculation, if \(Z\) is greater than 87.5°, it is set to 87.5°.


b Parametrized SED

We compute \(Q_a\), \(Q_s\) and \(Q_e\) from Eq. (3) – Eq. (5) in which Eq. (7) – Eq. (10) are used, and the perturbations in the
The quantities $Q_e$, $Q_a$ and $Q_s$ are multiplied by the present-day monthly mean solar insolation at the TOA calculated by Berger (1978) to produce the solar energy fluxes in Figs 2, 5 and 6. Although all datasets are two-dimensional in the horizontal and hence a two-dimensional SED has been calculated, only zonally averaged results and globally averaged annual means are presented. Since the investigation of interannual variability is beyond the scope of this paper, only multi-year means are presented.

### SOLAR ENERGY FLUX REFLECTED TO SPACE

The monthly mean solar energy fluxes reflected to space calculated using cloud amounts from the PATMOS, ISCCP and ERA-15 data are shown in Fig. 2 for January, April, July and October. The ERBE fluxes (black curves) are obtained from Li and Leighton (1993), whereas the other fluxes (red, green, blue curves) are calculated by using Eq. (5) inputting the indicated datasets. The thick magenta curve is the reflected flux calculated after the cloud amounts from the three sources were averaged.
of the cloud amounts in the above datasets are also used for calculating the SED. The reason for using three data sources for cloud amount is that the data from these three sources are very different, especially in the two polar regions (see the red, green and blue curves for PATMOS, ISCCP and ERA-15 respectively in Fig. 4).

Generally, the red, green and blue curves for the parametrized solar energy flux reflected to space are fairly close to the black curve for the ERBE satellite data (Li and Leighton, 1993). The differences between those calculated using four cloud amounts (PATMOS, ISCCP, ERA-15 and AVERAGED) and that from ERBE are shown in Fig. 3.

Obviously, different cloud amounts (see Fig. 4) lead to different solar energy fluxes reflected to space, especially in the two polar regions. From the thick magenta curves in Fig. 2 and Fig. 3 for the average cloud amounts, we note that the calculated SED is underestimated (overestimated) in the southern polar region for January (the northern polar region for July). A small bias (underestimation) exists in the tropics in January (Fig. 2a and Fig. 3a), in the sub tropics in the northern hemisphere in April (Fig. 2b and Fig. 3b) and in the sub tropics in the southern hemisphere in July (Fig. 2c and Fig. 3c). A relatively large bias (underestimation) exists south of 20°S in October (Fig. 2d and Fig. 3d). In the polar regions, satellite observations and climatological data are less reliable and the radiation processes are not yet well understood (Curry et al., 1996). The cause of the large discrepancy in the middle and high latitudes of the southern hemisphere in October is not evident.

2 ATMOSPHERIC ABSORPTION

Figure 5 shows the monthly mean solar energy fluxes that are calculated (coloured curves) and derived from satellite data (black curve) absorbed by the atmosphere in January, April, July and October. Again, the largest bias (underestimation) exists in the southern and northern polar regions (see Figs 5a, 5b and 5c). In other regions, the curves for the parametrized atmospheric absorption are quite close to those derived from ERBE.

![Figure 3](image-url)

Fig. 3 Differences between the calculated zonally averaged solar energy fluxes reflected to space at TOA using four cloud amounts (PATMOS, ISCCP, ERA-15, and AVERAGED) and the ERBE fluxes for (a) January, (b) April, (c) July and (d) October. Negative (positive) values represent an underestimation (overestimation) compared to the ERBE values.
(Li and Leighton, 1993), except that there are relatively large differences in April (Fig. 5b). It is interesting to note that the parametrized atmospheric absorptions are essentially the same for different cloud amounts from the three data sources. Even in the two polar regions, the discrepancies resulting from different cloud amounts are small. Thus the atmospheric absorption is relatively independent of cloud amount, which is in agreement with Li and Trishchenko’s (2001) conclusion.

3 SURFACE ABSORPTION

Figure 6 shows the monthly mean solar energy fluxes that are calculated (coloured curves) and derived from satellite data (black curve) absorbed by the surface in January, April, July and October. Here, the solar energy flux absorbed by the surface that is derived from satellite data from the Langley Dataset is also plotted (thick cyan line). The differences (between ERBE and the Langley Dataset) are generally very small except in the two polar regions in January, April and July. Again, the largest bias between the parametrized surface absorptions (red, green, blue and magenta lines) and the absorptions of ERBE (black line) and the Langley Dataset exists in the two polar regions. Generally, the parametrized surface absorptions for the four months shown are slightly larger than those derived from ERBE. Although smaller surface and atmospheric reflections can cause a larger surface absorption, the smaller atmospheric absorptions (see Fig. 5) can also lead to larger surface absorptions.

4 GLOBAL ANNUAL MEAN SED

Table 3 shows the global annual means of parametrized SEDs using three types of cloud amounts and their average and the global annual mean SED derived from ERBE. The calculation using ISCCP cloud amount gives a planetary albedo closer to that of ERBE because the global annual mean ISCCP cloud amount is the highest, while that using the PATMOS cloud amount gives the smallest value for the planetary albedo because the global annual mean PATMOS cloud amount is the smallest. The planetary albedo for the averaged cloud amount is 1.7% less than that of ERBE. A cloud amount change, however, has little effect on the global annual mean atmospheric
absorption. This is also true for different latitude bands, as seen in Fig. 5 and also in Subsection 5 of this section which deals with cloud optical depth changes. Such a small effect occurs because the increase (decrease) in cloud amount could increase (decrease) the atmospheric reflectivity and decrease (increase) the atmospheric transmissivity at the same time (see Table 1). Thus, the atmospheric absorptivity remains almost unchanged.

In fact, the overlapping of the cloud droplet and water vapour absorption bands results in the independence of the atmospheric absorption on clouds. The surface absorption has a simple relation with the solar energy flux reflected to space; i.e., the smaller the energy flux reflected to space due to a smaller cloud amount, the larger the surface absorption.

**5 SENSITIVITY TO CLOUD OPTICAL DEPTH**

Table 4 shows the global annual means of parametrized SEDs for three different cloud optical depths and the global annual mean SED derived from ERBE. (The cloud amount used here and in the following sensitivity studies is the averaged cloud amount.) In the observations, large uncertainty exists for the estimation of cloud optical depth (Pincus et al., 1995). It is quite common that in AGCMs the modelled cloud optical depth is tuned to obtain a better solar energy flux reflected to space (for example, see Yu et al. (1997)). Since ISCCP C2 data give 5.25 as the global annual mean of the cloud optical depth (Zhang et al., 1995), while ISCCP D2 data give 4.06, we decrease or increase the cloud optical depth by 30% to perform the sensitivity study. Here, it is found that the solar energy fluxes reflected to space and those absorbed at the surface are very sensitive to cloud optical depth. The 30% decrease in the cloud optical depth

### Table 3. Global annual means of parametrized SEDs (percent) using cloud amounts from PATMOS, ISCCP, ERA-15 and using an average of cloud amounts from these three data sources. Also given is the global annual mean SED derived from ERBE.

<table>
<thead>
<tr>
<th>Source</th>
<th>Space</th>
<th>Atmosphere</th>
<th>Surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>ERBE</td>
<td>29.7</td>
<td>24.3</td>
<td>46.0</td>
</tr>
<tr>
<td>PATMOS</td>
<td>27.3</td>
<td>22.6</td>
<td>50.1</td>
</tr>
<tr>
<td>ISCCP</td>
<td>29.0</td>
<td>22.3</td>
<td>48.7</td>
</tr>
<tr>
<td>ERA-15</td>
<td>27.9</td>
<td>22.4</td>
<td>49.7</td>
</tr>
<tr>
<td>AVERAGED</td>
<td>28.0</td>
<td>22.5</td>
<td>49.5</td>
</tr>
</tbody>
</table>

![Fig. 5](image) Zonally averaged solar energy flux absorbed by the atmosphere for (a) January, (b) April, (c) July and (d) October.

![Fig. 5](image)
caused the planetary albedo to change from 28.0% to 24.3%; a 30% increase in the cloud optical depth leads to an increase in the planetary albedo from 28.0% to 31.2%. From Eqs (16) and (17), it is clear that an increase (decrease) in the cloud optical depth leads to an increase (decrease) in the reflectivity and a decrease (increase) in the transmissivity. Hence, the resulting change in the atmospheric absorptivity is very small and the large change in the planetary albedo corresponds to a large change in the surface absorption.

6 SENSITIVITY TO AEROSOLS
Table 5 shows the global annual means of parametrized SEDs for three different aerosol optical depths and the global annual mean SED derived from ERBE. The effects of aerosols on the solar radiation budget have not been thoroughly studied. Also, aerosol optical depth observations are still poor, especially over continents. It is thus necessary for us to understand the sensitivity of our parametrized SED to aerosol optical depth. From Table 5, we note that a 50% decrease in the aerosol optical depth leads to a decrease in the planetary albedo from 28.0% to 26.9%; a doubled aerosol optical depth leads to an increase in the planetary albedo from 28.0% to 30.4%. Also, a decrease (increase) in the aerosol optical depth leads to a decrease (increase) in the atmospheric absorption. As a consequence, the surface absorption changes are relatively large.

It is clear that in this study the atmospheric absorption change due to aerosols is larger than that due to clouds. Recent aerosol research results show that the black carbon aerosols from biomass burning can produce a significant positive radiative forcing due to their strong absorptive properties (Kaufman et al., 2002; Li et al., 2001). However, uncertainties in the aerosol radiative forcing still remain large as aerosols have complicated chemical compositions and microphysical properties. Hence, the results in this study can only be considered preliminary.

Table 4. Global annual mean SEDs (percent) derived from ERBE, and parametrized using an averaged cloud amount for three different cloud optical depths. COD is the cloud optical depth from ISCCP.

<table>
<thead>
<tr>
<th>Source</th>
<th>Space</th>
<th>Atmosphere</th>
<th>Surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>ERBE</td>
<td>29.7</td>
<td>24.3</td>
<td>46.0</td>
</tr>
<tr>
<td>0.7 × COD</td>
<td>24.3</td>
<td>22.4</td>
<td>53.3</td>
</tr>
<tr>
<td>1.3 × COD</td>
<td>31.2</td>
<td>22.4</td>
<td>46.4</td>
</tr>
</tbody>
</table>

Fig. 6 Zonally averaged solar energy flux absorbed by the surface for (a) January, (b) April, (c) July and (d) October.
7 SENSITIVITY TO SURFACE ALBEDO
Table 6 shows the global annual means of parametrized SEDs for three different surface albedo fields and the global annual mean SED derived from ERBE. There exist large uncertainties for the surface albedo in both climate models and climatological datasets. In the high latitudes, there are both snow and ice change processes, it is rather difficult to model or observe the surface albedo. For this reason, the surface albedo used for ERA-15 has no annual variation; when producing ERA-40, a surface albedo field with seasonal variations was used, but it is very different from that in ISCCP. Due to the unreliable ERBE scene identification and large uncertainties in the radiative transfer computations in the polar regions, the surface albedo only between 60°S and 60°N is provided by Li and Garand (1994). Therefore, it is worthwhile to carry out this sensitivity study. In Table 6, a 10% decrease in the surface albedo leads to a 2.5% decrease in planetary albedo while a 10% increase leads to a 3.5% increase. (Note: if the decreased surface albedo is less than zero, it is set to zero.) A decrease (increase) in surface albedo leads to a decrease (increase) in atmospheric absorption. This change occurs because a decrease (increase) in the surface albedo reduces (increases) the solar energy flux reflection from the surface to the atmosphere. The change in the surface absorption due to the surface albedo change is larger than that in atmospheric absorption and solar energy flux reflected to space.

5 Summary and Discussion
The radiation schemes for the SED in the EBMs of several EMICs are somewhat arbitrary and vary from model to model. Moreover, the parametrized planetary albedo used in EBMs is fitted from limited observational data and is too simple to take into account the effects of atmospheric compositions. In this study, a significantly improved parametrization of SED (including the planetary albedo) is proposed which is obtained from the application of an RCM. This parametrization of SED is also used to calculate the present-day SED by using some climatological data, and it is also tested against the SED derived from ERBE satellite data (Li and Leighton, 1993; Gupta et al., 1999).

The comparisons of the parametrized SED with that derived from ERBE satellite data show general agreement, although discrepancies do exist. Our parametrized solar energy fluxes reflected to space are much closer to those of ERBE than those shown in Li et al. (1997) (see their Figs 3a and 3b) for zonally averaged distributions. Since this energy flux can be observed with very high accuracy by satellites, our parametrization for this flux is encouraging, assuming its interannual variability is small. The zonally averaged atmospheric and surface energy flux absorptions are also in good agreement with those derived from ERBE (Li and Leighton, 1993). Although the atmospheric and surface absorption in Li and Leighton (1993) are derived by using a parametrization, their results are extensively validated by surface measurements (Li et al., 1995; Wild et al., 1998), and the surface absorption in Li and Leighton (1993) is in close agreement with another estimation derived from satellite data (Gupta et al., 1999). From Table 3, we note that our global annual mean solar energy flux reflected to space is slightly lower than that of ERBE. The atmospheric absorption is also slightly lower than that of ERBE, but very close to that (22.9%) estimated by Gupta et al. (1999). The surface absorption is slightly larger than that of ERBE, which is due to slight underestimation of the TOA reflected flux and atmospheric absorption.

Sensitivity studies show that cloud optical depth changes lead to notable changes in the energy flux reflected to space and that absorbed at the surface, and only to small changes in the atmospheric absorption. Aerosol optical depth changes lead to a relatively large change in the atmospheric absorption, in addition to changes in the energy flux reflected to space and in the surface absorption (see Table 5). Surface albedo changes lead to changes in the TOA reflected energy flux, atmospheric absorption and surface absorption.

The discrepancies between our parametrized SED and that derived from satellite data may be the result of many factors. Discrepancies are relatively large in the polar regions. At present, it is very difficult to find specific reasons for these large discrepancies (Curry et al., 1996). These may occur because polar satellite observations of radiation flux contain errors resulting from extremely low water vapour content, unreliable angular correction and scene identification in polar regions (Li et al., 1997). Surface albedo in the polar regions may produce further uncertainties because of complicated snow and ice processes. Both the observational and modelling studies for aerosols are also poor in the polar regions. In addition, cloud observations have large uncertainties in the polar regions: cloud amounts are quite different for the three datasets (PAT-MOS, ISCCP and ERA-15) (see Fig. 4); cloud optical depth observation may also be unreliable in the polar regions. In addition to the above, we cannot exclude uncertainties resulting from the RCM, since radiation processes in the polar regions have not yet been well resolved (for example, the effects of polar sphericity). It is difficult to find reasons for the underestimation in the southern middle and high latitudes.
in October (Figs 2d and 3d). Obviously, extended satellite observations as well as further observational and modelling studies on the SED are needed.

Despite the discrepancies and uncertainties discussed above, our results nevertheless show that the parametrized SED proposed in this study can be used to derive the present-day climatology of SED by using the climatological data for cloud amount and optical depth, precipitable water, aerosol optical depth and surface albedo. This parametrization, which is a fast and accurate radiation scheme, is applicable to those climate models that employ an atmospheric EBM and MBM and a one-layer atmospheric model for climate and paleoclimate modelling.

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References