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SOLAR RADIATION MODEL RAD1

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SUMMARY

In the solar radiation model, RAD1, the radiative transfer equation for a cloudless, plane-parallel atmosphere is solved using the successive-scattering iterative procedure of Davé and Gazdag in which the scattering phase function is expressed in a Fourier series. The model includes Rayleigh and Mie scattering along with absorption by aerosol, ozone, water vapor, carbon dioxide and oxygen. The solar spectrum is divided into discrete spectral intervals (83 in the present model), and the vertical column is divided into up to 500 layers depending upon the optical thickness of the atmosphere. The direct solar flux and the upward and downward diffuse fluxes for each spectral interval at each level are computed taking into account all orders of scattering.

INTRODUCTION

The Lawrence Livermore Laboratory solar radiation model, RAD1, is modeled after the Braslau-Davé solar radiation model.^[1] We assume a cloudless, plane-parallel atmosphere in which there is absorption by oxygen, ozone, water vapor, carbon dioxide and aerosol along with molecular and Mie multiple scattering. At the bottom of the atmosphere, there is a Lambert surface, which scatters radiation isotropically. The solar spectrum between 0.285 μm and 2.5 μm is divided into 83 unequal intervals, and the radiation in each interval is treated as monochromatic in the calculation. Polarization is not taken into account since it makes no significant difference in the flux values.^[2] Heating rates are computed by taking the divergence of the net downward flux.

THE RADIATIVE TRANSFER EQUATION

We first define the various parameters which appear in the radiative transfer equation. The direction of propagation is denoted by the parameters $\mu = |\cos\theta|$ and ϕ where θ is the angle which the direction makes with the local zenith, and ϕ is the azimuth angle referred to an arbitrary meridian plane. The downward direction is represented by $-\mu$.

The vertical (normal) optical depth τ of the atmosphere at a height of h km above the ground is the sum of several vertical optical depths which are due to absorption and scattering by molecules and aerosols. Thus

$$\tau = \tau^{(s,r)} + \tau^{(s,m)} + \tau^{(a,m)} + \tau^{(a)} \quad (1)$$

The optical depth is defined as 0 at the top of the atmosphere and τ_b at the bottom of the atmosphere. The Rayleigh-scattering optical depth $\tau^{(s,r)}$ is taken to be proportional to the pressure at that height. Values of $\tau_b^{(s,r)}$ as a function of wavelength are given by Howard et al. [3] The quantities $\tau^{(s,m)}$ and $\tau^{(a,m)}$ are respectively the scattering and absorption optical depths due to aerosol. Values for these quantities can be computed using Mie theory. [4] The quantity $\tau^{(a)}$ is the optical thickness due to absorption by various gases.

The intensity of the radiation at the level τ in the direction μ, ϕ is denoted by $I(\tau; \mu, \phi)$. The basic radiative transfer equation for monochromatic radiation in a plane-parallel, nonhomogeneous atmosphere has the form:

$$\mu \frac{dI(\tau; \mu, \phi)}{d\tau} = I(\tau; \mu, \phi) - \omega(\tau) J(\tau; \mu, \phi) \quad (2)$$

where the source function $J(\tau; \mu, \phi)$ is given by

$$J(\tau; \mu, \phi) = \frac{1}{4} e^{-\tau/\mu_0} P(\tau; \mu, \phi; -\mu_0, \phi_0) F + \frac{1}{4\pi} \int_{-1}^{+1} \int_0^{2\pi} P(\tau; \mu, \phi; \mu', \phi') I(\tau; \mu', \phi') d\mu' d\phi' \quad (3)$$

It is assumed above that πF is the incident solar flux at $\tau = 0$ in the direction $-\mu_0, \phi_0$. The boundary conditions are

$$I(0; -\mu, \phi) = 0 \quad (4a)$$

and

$$I(\tau_b, +\mu, \phi) = 0 \quad (4b)$$

That is, there is no diffuse radiation incident at the top or the bottom of the atmosphere.

The albedo of single scattering $\omega(\tau)$ and the normalized scattering phase function $P(\tau; \mu, \phi; \mu', \phi')$ of a unit volume containing air and aerosol are defined as

$$\omega(\tau) = [\Delta\tau^{(s,m)} + \Delta\tau^{(s,r)}] / \Delta\tau \quad (5)$$

and

$$P(\tau; \mu, \phi; \mu', \phi') = T(\tau) M(\mu, \phi; \mu', \phi') + [1 - T(\tau)] R(\mu, \phi; \mu', \phi') \quad (6)$$

where the turbidity factor $T(\tau)$ is defined as

$$T(\tau) = \Delta\tau^{(s,m)} / [\Delta\tau^{(s,m)} + \Delta\tau^{(s,r)}] \quad (7)$$

The quantities $M(\mu, \phi; \mu', \phi')$ and $R(\mu, \phi; \mu', \phi')$ are the normalized Mie and Rayleigh scattering phase functions, respectively. The parameters μ and ϕ indicate the direction of the scattered radiation, and the parameters μ' and ϕ' indicate the direction of the incident radiation.

METHOD OF SOLUTION

By expanding the scattering phase function, intensity and source function in Fourier series forms, [5,6] the double integral in Eq. (3) reduces to a single integral since the integration over azimuth angle can be carried out analytically. The intensity takes the following form:

$$I(\tau; \mu, \phi) = \sum_{n=1}^N I^{(n)}(\tau; \mu) \cos[(n-1)(\phi_0 - \phi)] \quad (8)$$

By definition, the downward and upward diffuse fluxes at a level τ are given by

$$F_d(\tau) = 2\pi \int_0^1 I^{(1)}(\tau; -\mu) \mu d\mu \quad (9)$$

and

$$F_u(\tau) = 2\pi \int_0^1 I^{(1)}(\tau; \mu) \mu d\mu \quad (10)$$

Consequently, we need only evaluate the first term of the series ($n=1$).

The first term on the right-hand side of Eq. (3) represents the contribution to the source function due to primary scattering of the direct solar radiation by the unit volume located at the level τ . A successive-scattering iteration procedure is used to solve the radiative transfer equation using this first term as the initial value. After dividing the atmosphere into layers (up to 500), values are obtained for $I^{(1)}(\tau, \pm\mu)$ at increments of 2° in θ (91 values per level) for each level. The iteration procedure continues until values of $I^{(1)}(\tau_b, -\mu)$ on successive iterations

agree within given limits for all values of μ . The present convergence criterion is

$$I_{m+1}^{(1)}(\tau_b; -\mu) / I_m^{(1)}(\tau_b, -\mu) < 1.002$$

where m is the number of iterations.

Once values are obtained for $I^{(1)}(\tau, \pm\mu)$, the effect of ground reflection is incorporated using a procedure introduced by Chandrasekhar.^[7] This results in increasing $I^{(1)}(\tau; \pm\mu)$ by an amount $I^{*(1)}(\tau; \pm\mu)$, assuming isotropic scattering at the earth's surface with a given surface albedo.

COMPUTATIONAL DETAILS

Values of pressure, temperature, aerosol number density, and densities of ozone and water vapor are specified for several selected heights in the model (usually 1 km intervals up to 25 km, then 2.5 km or 5 km intervals up to 50 km). Each layer is then subdivided into sublayers, each with the same optical properties, such that the total optical depth of each sublayer $\Delta\tau$ is less than 0.02. There may be as many as 500 sublayers, depending upon the total optical depth of the atmosphere. The amount of carbon dioxide in a layer is assumed to be proportional to the pressure thickness of the layer.

Following McClatchey et al.,^[8] the concentrations are scaled using the scaling factors:

$$(p/1013)^{0.9} (273/T)^{0.45} \quad \text{for water vapor}$$

and

$$(p/1013)^{1.75} (273/T)^{1.375} \quad \text{for oxygen and carbon dioxide}$$

where p and T are the pressure (mb) and temperature (°K).

Three types of aerosol size distribution functions may be used in the model: (1) discontinuous power law^[9]

$$\begin{aligned} n(r) &= C & r_{\min} &\leq r \leq r_m \\ n(r) &= C(r_m/r)^{(v+1)} & r_m &\leq r \leq r_{\max} \end{aligned} \quad (11)$$

(2) modified gamma function^[4]

$$n(r) = ar^\alpha \exp(-br^\gamma) \quad r_{\min} \leq r \leq r_{\max} \quad (12)$$

or (3) lognormal function^[10]

$$\begin{aligned} n(r) &= [\sigma r(2\pi)^{1/2}]^{-1} \exp[-(\ln r - \ln r_m)^2 / 2\sigma^2] \\ r_{\min} &\leq r \leq r_{\max} \end{aligned} \quad (13)$$

The size distribution is assumed constant with height. The aerosol particles are assumed spherical with a complex index of refraction $m = n - ki$, which may vary with wavelength.

The vertical optical depth due to gaseous absorption $\tau^{(a)}$ is the sum of the optical depths for absorption by ozone, water vapor, oxygen and carbon dioxide. The optical depth for a given spectral region due to absorption by ozone is given by $\alpha_z c_{zi}$ where c_{zi} is the ozone amount (cm-atm)

above the i -th level and α_z is the absorption coefficient of ozone for the spectral region considered.

The transmission factor $T_{wi}(\mu_o)$ due to c_{wi} gm cm⁻² of scaled water vapor amount above the i -th level is obtained by reading the value of transmission factor corresponding to $\log_{10} \alpha_w + \log_{10} (c_{wi}/\mu_o)$ from Fig. 1. The values of $T_{xi}(\mu_o)$ and $T_{ci}(\mu_o)$ due to c_{xi} and c_{ci} km scaled depths (or scaled slant path lengths) of oxygen and carbon dioxide respectively can also be obtained following a similar procedure. These values are then inserted into the following expression to obtain the total gaseous absorption attenuation of the direct solar radiation reaching the i -th level:

$$\exp(-\tau_i^{(a)}/\mu_o) = T_{wi}(\mu_o) T_{xi}(\mu_o) T_{ci}(\mu_o) \exp(-\alpha_z c_{zi}/\mu_o) \quad (14)$$

Since the Bouguer-Langley law is not applicable for the spectral widths used, it is not possible to define accurately the gaseous absorption optical thickness $\Delta\tau_i^{(a)}$ of the i -th layer due to water vapor (or oxygen or carbon dioxide). However, as the total scattering optical thickness of the terrestrial atmosphere in the region of such absorption is generally small (~ 0.1), the direct solar radiation and the primary-scattered radiation are the prime contributors to flux terms at all levels of the atmosphere. With this in mind, $\Delta\tau_{y,i}^{(a)}$ corresponding to the y -th gas is defined as follows:

$$\Delta\tau_{y,i}^{(a)} = -\mu_o \log_e [T_{y,i+1}(\mu_o)/T_{y,i}(\mu_o)] \quad (15)$$

This definition of $\Delta\tau_i^{(a)}$ has been shown by Braslau and Davé^[1] to be valid for regions of fairly high transmission ($\sim 90\%$). Otherwise, there

is some small modification of the flux divergence picture of the lower troposphere only.

The spectrally integrated flux components are estimated to be accurate within about $\pm 0.5\%$.^[1] The solar heating rates are computed by taking the divergence of the net downward flux. A typical calculation requires about 20 minutes on a CDC-7600 computer. Although the iterative procedure used in this model is very time consuming and is limited to a cloud-free atmosphere, it does allow for a high degree of accuracy in the calculation of solar fluxes (direct and diffuse) for cases with highly peaked phase functions.

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Fig. 1. Dependence of the transmission factor on the scaled amount of gas in the path. The quantity C represents the scaled amount of gas (or the scaled path length). The quantity $\log_{10} \alpha$ represents the absorption parameters of a gas for the spectral region under investigation. (After Braslau and Davé [1]).

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