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A determination of the greenhouse parameter for dry and unpolluted air

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RESUMEN

La extinción relativa de radiación por moléculas de aire limpio y seco, tanto en la región espectral solar como en la infrarroja, ha sido estimada a través del cálculo teórico de un parámetro conocido en general como Parámetro de Invernadero (PI). En una primera aproximación, este parámetro fue calculado considerando el aire terrestre como una mezcla simple de oxígeno y nitrógeno solamente. El método usado aquí se basó en la aplicación, bajo un procedimiento inverso, de un modelo homogéneo, plano-paralelo e independiente del tiempo en el cual se empleó la aproximación de Eddington, tanto en la región solar del espectro como en la región infrarroja para resolver la ecuación de transferencia radiativa, y que tiene el PI como parámetro de entrada. El mejor valor del PI fue estimado ajustando el perfil de temperatura para cuatro tipos de superficies uniformes (nieve, desierto, vegetación y océano) con albedos promedios conocidos en ambas regiones espectrales, adoptando valores para la temperatura superficial del aire, escogidos de acuerdo a un criterio promedio radiativo de un ambiente local o micro-climatológico supuesto. Con este resultado, también fue posible hacer una estimación tanto de la opacidad infrarroja de la capa de aire implicada en este trabajo, como de su coeficiente de extinción en esta región espectral. Los resultados predichos son comparados con resultados obtenidos indirectamente de datos aportados por otros autores. Aunque su validación está sujeta exclusivamente al modelo radiativo aplicado, se concluye que el valor del PI obtenido es más apropiado que el inicialmente encontrado en forma indirecta en la literatura.

ABSTRACT

The relative extinction of solar and infrared radiation by dry and clean air molecules, has been estimated through a theoretical determination of the ratio referred ordinarily as the Greenhouse Parameter (GP). In a first approach, it was calculated assuming that terrestrial air only consists of a simple mixture of oxygen and nitrogen. The method used here is based on the application, in an inverse procedure, of an homogeneous, plane-parallel, and time-independent grey model, which employs the Eddington approximation as a solution to the radiative transfer equation, both in the solar and the infrared spectral regions and, which has the GP value as an input free parameter. The best value of the GP was estimated calibrating the local temperature profile for four types of uniform surface (snow, desert, vegetation and ocean), with average albedos known in these spectral regions, adopting air surface temperature values which were chosen for an assumed micro or local climatological environment according to an average radiative criterion. With this result, it was possible for an estimation of the infrared opacity for the air layer implicated in this model and also the mean extinction coefficient in this spectral range to be calculated. The results predicted are compared with results obtained indirectly from the data provided by other authors. Although its validation is constrained solely to the radiative model applied it seems that the value of the GP obtained is more accurate than the one initially available.

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1. Introduction

The Earth's atmosphere is a mixture of several gases with nitrogen and oxygen predominant and in steady proportions at altitudes below 90 km. Due to gravity most of the total mass of this atmosphere ($\sim 5 \times 10^{18}$ kg) is concentrated close to the surface; 99% of the total mass lies below 30 km (Verniani, 1966; McCartney, 1976). Pertaining to unpolluted and dry conditions, at sea level, atmospheric air is composed of its major constituents, that is to say, nitrogen (78.084%), oxygen (20.946%), argon (0.934%) and variable carbon dioxide ($\sim$ 0.034%), and by its minor constituents, that is to say, neon, helium, methane, nitrous oxide, etc. (Brimblecombe, 1986; Bunce, 1991). Therefore, terrestrial air may be considered roughly as a mixture of 4/5 nitrogen and 1/5 oxygen, confined in a very thin spherical shell with respect to the Earth's radii.

Although an air molecule is not a physical entity this concept or abstraction, however, can be useful in exploring fundamental properties of air. It has been used for general purposes as it will be done in this paper. Table 1 gives some basic physical properties of "air molecules".

TABLE 1. Some basic physical properties (at STP conditions) for "air molecules" (McCartney, 1976).

Molecular weight	28.964
Mass	4.807×10^{-23} g molec^{-1}
Density	1.29 kg m^{-3}
Root-mean-square speed	485 m s^{-1}
Diameter	$\sim 3.7 \times 10^{-10}$ m molec^{-1}
Average spacing between molecular center	$\sim 3.3 \times 10^{-9}$ m
Mean free path	$\sim 6 \times 10^{-8}$ m
Mean collision rate	$\sim 8 \times 10^{9}$ s^{-1}

According to these properties, one can obtain additional information about air molecules in order to understand basic interactions and processes in the radiative transfer of this medium. It travels, for instance, its mean free path in about 1.2×10^{-10} s; by comparing its molecular diameter to a reference wavelength of 5.5×10^{-7} m (0.55 μm), the phase of this wave passes uniformly over the molecule and this, in turn, is an important fact to Rayleigh scattering; during 1.8×10^{-15} s, the period of this wavelength (its frequency is 5.5×10^{14} s^{-1}), the air molecule moves only 0.0023 (8.7×10^{-11} cm) of its own diameter, which is insignificant; in scattering energy from this wavelength, approximately 6.6×10^{4} oscillations take place in its electronic structure during the travel time; etc. (McCartney, 1976). Thus, one can continue searching to acquire more basic or general data for air molecules.

In this sense the purpose of this paper is to attempt to determine theoretically for dry and unpolluted air, in a first approach and in a simple but reasonable way, the ratio of the total vertical opacity in the solar range to the same in the infrared range. In other words, it will estimate the ratio of the solar and infrared extinction's (absorption plus scattering) produced by air molecules. Thus, given the total vertical optical depth in the solar range and known ratio, an evaluation of both the total vertical depth and the mean extinction coefficient in the infrared could be made for these kind of molecules if the height or geometrical thickness of the air layer implicated in this research, is determined.

Generally the relative effectiveness with which any medium extinguishes radiant energy in a certain spectral region in regard to another spectral region, can be measured in terms of a dimensionless number or ratio.

Although the greenhouse concept, strictly speaking, concerns or applies only to the absorption-emission processes of trace gases (water vapor, carbon dioxide, methane, etc) in the atmosphere, this dimensionless number, as will be shown below, is commonly referred to by many authors as the Greenhouse Parameter (GP). This is so since, in this case, it indicates the amount of radiative energy extinguished in the solar in comparison with the amount extinguished in the infrared by the medium, regardless of whether or not this is to be considered actually as a greenhouse gas.

As GP is a pure radiative parameter, independent of time, position and/or geometry of the medium, a conveniently homogeneous, plane-parallel (local) and time-independent one-dimensional radiative model, which has the GP value as an input free parameter, has only been applied. This model along with the method used in this paper will be described briefly in the next three sections.

2. The method

In most publications on radiative transfer calculations, it is assumed that the medium is characterized by a set of radiative and geometric parameters which take a set of given values, depending on the specific medium. Under particular boundary conditions and characteristics of the incident radiation, a quantitative assessment of the diffused radiative fields and also of the temperature distribution field can be made. However, the inverse procedure, in which the attenuated and diffused radiative fields and the temperature profile are known in the medium, is the most suitable and convenient for obtaining the parameter values that characterize it. For example, Durham and Chamberlain (1989) used a one-dimensional convective-radiative model in this way to estimate CO_2 levels in early terrestrial atmospheres. Nevertheless, and unfortunately, up to date, no special consistent and reliable model has been developed due to the requirement of multiple parameters necessary to characterize a medium in a realistic way (Bellman *et al.*, 1965).

In any case, for this problem, the integro-differential radiative transport equation (RTE) has to be solved using alternative methods of solution (Krook, 1955). One of these methods was proposed by Eddington (1926) who used the Legendre polynomial series truncated at the third term. Its applications, range of validity, and limitations have been widely studied by many authors (Henyey and Greenstein, 1941; Krook, 1955; Irvine, 1968; Huang, 1968; Kawata and Irvine, 1970; Shettle and Weinman, 1970; Shettle, 1972; Zdunkowski and Junk, 1974; Irvine, 1975; Wiscombe and Joseph, 1977; Mihalas and Mihalas, 1983; Ibañez, 1979, 1981, 1986; Ibañez and Plachco, 1989). Specifically, this solution is acceptable as far as the scattered radiation phase function does not present neither strong maxima nor strong minima and, the medium is highly dispersive along with an optical depth other than unity (Irvine, 1975; Wiscombe and Joseph, 1977). In view of this approximation, one takes the zero and first order moments in order to obtain the mean intensity and net flux variables for the diffused radiation into the medium.

On the other hand, in literature, the RTE in the Earth's atmosphere is also usually solved in different ways following the considered spectral range, because different physical processes are involved. In the solar range (0.25 μm - 4 μm), scattering (by molecules) and absorption (by trace gases) must be considered in combination for a clear and cloudless atmosphere (with an external radiative source). In the thermal infrared range (4 μm - 100 μm), scattering can be neglected and only absorption-emission (by trace gases) are to be considered (with internal sources). The solar range can be further divided by considering the so-called "visual" part ($\sim 0.25\ \mu$m - $\sim 0.7\ \mu$m) and the "near-infrared" ($\sim 0.7\ \mu$m - $\sim 4 \mu$m). All of these considerations allow us to deal with, in a rough approximation, the spectral dependence of radiative parameters, i.e., the surface albedo, the asymmetry factor, the optical depth and the single scattering albedo.

In connection with this, the thermal response of the medium can reveal a heating and/or a cooling (greenhouse and/or anti-greenhouse effects). In a radiative regime, this thermal response can be analyzed assuming a plausible hypothetical radiative equilibrium as an additional approximation made in this work. The Earth's troposphere is not in fact in such an equilibrium. To account for the total planetary heat balance in this part of the atmosphere it should be stressed that the surface is heated by solar radiation, much of which is transferred to the atmosphere by conduction as sensible heat or through evaporation as latent heat. The surface heating is then distributed through the troposphere by convective (vertical) mixing, and thermal equilibrium is maintained by long-wave cooling to the surface and to space. Obviously, in the context of a radiative regime the above aspects have to be disregarded so that the temperature profile is obtained by solving only the basic equilibrium equation for a grey condition, in the solar and in the infrared regions, assuming a local thermodynamic equilibrium as well.

In this paper, the inverse procedure mentioned above has been applied to achieve theoretically a preliminary estimation of GP magnitude for dry and clean air, using a homogeneous, plane-parallel, and time-independent grey radiative model, which has been processed with the Eddington approximation as a solution for the RTE, both in the solar and in the infrared spectral regions. Furthermore, a calculation of the opacity of this air could be made in the infrared provided its opacity is in the solar range. Also with this latter value the mean extinction coefficient for air in this region was evaluated.

It must be emphasized, however, that the aim of this work is not to reproduce the real temperature-height profile of the Earth's atmosphere. This has been done by other workers. By solving the RTE taking into account the nongrey atmosphere in the solar and infrared ranges, the radiative and radiative-convective models published in the literature (J. Atmos. Sci.) from the mid-60s by Manabe and colleagues, have reproduced to a very good approximation the temperature profile of the Earth's atmosphere.

Not withstanding the above, coherent and physical explanations of the local atmospheric radiative temperature profiles obtained here have been given to support the results, calculated by fixing air-surface temperatures, which values were based on an average radiative criterion for an assumed micro or local climatological environment.

Finally, as mentioned above, in calculating the radiative properties described here, the air has simply been considered as a mixture of nitrogen and oxygen only. In other works, in which a good determination of its average properties like viscosity, mean free path, etc., have been made, the air has been also treated in the simplest approach as a single component gas or as ternary mixture of nitrogen, oxygen and argon (Jennings, 1988). Moreover, it is pointed out that purely grey (or quasi-grey) radiative models are often applied to study general radiative aspects of the real atmosphere of the Earth (Plass, 1956; Shettle and Weinman, 1970; Shettle, 1972; Chamberlain, 1980; Sawyer, 1984; Bohren, 1987) and other terrestrial planets (Sagan and Pollack, 1967; Ingersoll, 1969; Rasool and De Bergh, 1970; Golitsyn and Ginsburg, 1985). The model used in this paper and other models used elsewhere, but more simplified or over-simplified, have been very convenient and practical in inferring the importance of radiative component changes in the greenhouse (global warming) or anti-greenhouse (e.g. Nuclear Winter or its natural analogues) effects, caused by changes in the atmospheric composition and/or by changes in the incident solar flux, etc. (Möller, 1963; Chylek and Coakley, 1974; Hart, 1978;

Chamberlain, 1980; Crutzen *et al.*, 1984; Golitsyn and Ginsburg, 1985; Broyles, 1985; Penner and Haselman, 1985; Peñaloza, 1988a, 1988b; Bridgman, 1989; Ramanathan *et al.*, 1989; Turco *et al.*, 1990).

Essentially, the method used here follows the dictum of William of Ockman. As in the works referred to above as well in other works related to atmospheric modeling (Bryson and Dittberner, 1976), the author has attempted to give a realistic answer using as few variables as possible.

3. The radiative transfer model

Because dry air is non-conservative medium (nitrogen and oxigen molecules are responsibles for about 99% of the atmospheric molecular scattering), it follows the model of Shettle and Weinman (1970) and applied also by Shettle (1972). This model has been utilized for the treatment of the solar part of the incoming radiation from the sun. In their homogeneous, multiple-layered and plane-parallel model, these authors consider the asymmetry factor (g) and single scattering albedo (ϖ_s) independent the optical depth. Additionally, they assume a Legendre polynomial series for the phase function commonly applied in planetary atmospheres (Chandrasekhar, 1960; Busbridge, 1960; Harris, 1961; Sobolev, 1963; Goody, 1964; Irvine, 1963/64, 1975). Here, the phase function is also taken independent of the optical depth.

The radiative transfer equation at frequency ν is

$$\mu \frac{dI_\nu(\tau_\nu,\mu,\phi)}{d\tau_\nu} = -I_\nu(\tau_\nu,\mu,\phi)+$$

$$\frac{\varpi_\nu}{4\pi}\int_o^{2\pi}\int_{-1}^{1} I_\nu(\tau_\nu,\mu,\phi;\mu',\phi')p_\nu(\mu,\phi;\mu',\phi')d\mu' d\phi' + (1-\varpi_\nu)B_\nu(T), \tag{1}$$

where I_ν is the specific intensity at optical depth τ_ν and in the direction $\mu(=\cos\theta)$ and ϕ, ϖ_ν is the single scattering albedo, p_ν is the phase function and, $B_\nu(T)$ the Planck function. Because the specific intensity may be split in attenuated and diffuse fields, i.e., $I_\nu(z,\hat{n}) = I_\nu^a(z,\hat{n}) + I_\nu^d(z,\hat{n})$, the equation (1) in the solar range becomes

$$\mu \frac{dI_s^d(\tau_s,\mu,\phi)}{d\tau_s} = -I_s^d(\tau_s,\mu,\phi) + \frac{\varpi_s}{4\pi}\int_o^{2\pi}\int_{-1}^{1} I_s^d(\tau_s,\mu',\phi')p_s(\mu,\phi;\mu',\phi')d\mu' d\phi'$$

$$+\frac{1}{4}\varpi_s F_{os} p_s(\mu,\phi;\mu_o,\phi_o)e^{-\tau_s/\mu_o}. \tag{2}$$

Introducing the Eddington approximation in the above equation for the solar diffused radiation, i.e., $I_s^d(\tau_s,\mu) = I_{os}^d(\tau_s) + \mu I_{1s}^d(\tau_s)$ and, the phase function $p_s = (1+\hat{\omega}'\cos\gamma)$, where γ, the scattering angle, is $\cos\gamma = \mu\mu' + [(1-\mu^2)^{1/2}(1-\mu'^2)^{1/2}\cos(\phi-\phi')]$, the following solutions are obtained:

$$I_{os}^d(\tau_s) = C_1 e^{-k\tau_s} + C_2 e^{k\tau_s} - \xi e^{-\tau_s/\mu_o}, \tag{3}$$

$$I_{1s}^{d}(\tau_s) = q(C_1 e^{-k\tau_s} - C_2 e^{k\tau_s}) - \beta e^{-\tau_s/\mu_o}, \tag{4}$$

where k, q, ξ and β are parameters given by Shettle and Weinman (1970), with $i = 1$. The constants C_1 and C_2 are determinated from the boundary conditions. Therefore,

$$F_s^d(\tau_s) = 2\pi \int_o^{\pm 1} [I_{os}^d(\tau_s) + \mu I_{1s}^d(\tau_s)]\mu d\mu = \pi[I_{os}^d(\tau_s) \pm \frac{2}{3} I_{1s}^d(\tau_s)], \tag{5}$$

$$F_s^d \downarrow (0) = 0 = 2\pi \int_o^{1} \mu[I_{os}^d(\tau_s) + \mu I_{1s}^d(\tau_s)] d\mu = \pi[I_{os}^d(0) + \frac{2}{3} I_{1s}^d(0)], \tag{6}$$

$$F_s^d \uparrow (\tau_s^*) = \pi[I_{os}^d(\tau_s^*) - \frac{2}{3} I_{1s}^d(\tau_s^*)] = A_s^* \pi [I_{os}^d(\tau_s^*) + \frac{2}{3} I_{1s}^d(\tau_s^*) + \mu_o F_{os} e^{-\tau_s^*/\mu_o}], \tag{7}$$

where, $F_s^d \downarrow (0)$ is the downward solar diffused flux at the top of the layer, $F_s^d \uparrow (\tau_s^*)$ is the upward directed solar diffused flux, τ_s^* is the solar total vertical optical depth of the layer, F_{os} is the initial plane-parallel solar radiation with a cosine angle of incidence μ_o and, A_s^* is the mean solar surface albedo which is assumed constant in the present paper (Lambert surface). In general, the surface planetary albedo varies with the solar incidence angle (Squyres and Verveka, 1982).

The equation (1) in the infrared region is

$$\mu \frac{dI_p(\tau_p, \mu, \phi)}{d\tau_p} = -I_p(\tau_p, \mu, \phi) + B_p(\tau_p), \tag{8}$$

where $B_p(\tau_p)$ is the Planck function in the infrared. The Planck function is (Ibáñez, 1979),

$$B_p(\tau_s) = J_p(\tau_s) + \eta(1 - \varpi_s) J_s(\tau_s), \tag{9}$$

where $J_p(\tau_s)$ is the infrared mean intensity and, $J_s(\tau_s)$ is the solar mean intensity. In Appendix A a derivation of this equation is given. Introducing the greenhouse parameter defined as (King, 1963; Samuelson, 1967; Wildt, 1966; Stibbs, 1971; Morgan, 1973; Harshvardhan and Cess, 1976; Ibáñez, 1979, 1981):

$$\left(\frac{d\tau_p}{d\tau_s}\right) = \left(\frac{\kappa_p dz}{\kappa_s dz}\right) = \left(\frac{\kappa_p}{\kappa_s}\right) = \left(\frac{\tau_p}{\tau_s}\right) = \frac{1}{\eta}, \tag{10}$$

where κ_p and κ_s are the extinction coefficients (absorption plus scattering) in the infrared and in the solar ranges, respectively, the equation (8) becomes

$$\mu \frac{dI_p(\tau_s, \mu, \phi)}{d\tau_s} = -\frac{1}{\eta}[I_p(\tau_s, \mu, \phi) - J_p] + (1 - \varpi_s)(J_s^a + J_s^d), \tag{11}$$

where $J_s^a(\tau_s)$ is the attenuated solar mean intensity and $J_s^d(\tau_s)$ is the solar diffuse mean intensity. As for the solar range case, the infrared specific intensity I_p may be split in attenuated and diffused fields, i.e., $I_p(\tau_s, \mu, \phi) = I_p^a(\tau_s, \mu, \phi) + I_p^d(\tau_s, \mu, \phi)$. However, $I_p^a(\tau_s, \mu, \phi)$ can be neglected since the solar radiation is the dominating energy in its spectrum. Therefore, $I_p(\tau_s, \mu, \phi) \simeq$

$I_p^d(\tau_s, \mu, \phi)$ and $J_p^a \simeq 0$ so that $J_p(\tau_s) = J_p^d(\tau_s)$. With these approximations, the equation (11) becomes:

$$\mu \frac{dI_p^d(\tau_s, \mu, \phi)}{d\tau_s} = -\frac{1}{\eta}[I_p^d(\tau_s, \mu, \phi) - J_p^d] + (1 - \varpi_s)(J_s^a + J_s^d). \tag{12}$$

By applying the Eddington approximation the solution of eq. (12) is

$$I_{op}^d(\tau_s) = -3(1 - \varpi_s)\eta^{-1}\{(1/k^2)(C_1 e^{-k\tau_s} + C_2 e^{k\tau_s}) + \mu_o^2[(F_{os}/4) - \xi]e^{-\tau_s/\mu_o}\}$$

$$-\eta^{-1}C_3\tau_s + C_4, \tag{13}$$

$$I_{1p}^d(\tau_s) = 3(1 - \varpi_s)\{(1/k)(C_2 e^{k\tau_s} - C_1 e^{-k\tau_s})\} + \mu_o[\xi - (F_{os}/4)]e^{-\tau_s/\mu_o} + C_3. \tag{14}$$

The constants C_3 and C_4 are obtained using the same boundary conditions applied to the diffused solar field, i.e.,

$$F_p^d \downarrow (0) = 0 \Rightarrow I_{op}^d(0) = -(\frac{2}{3})I_{1p}^d(0), \tag{15}$$

$$F_p^d \uparrow (\tau_s^*) = A_p^* \pi[I_{op}^d(\tau_s^*) + \frac{2}{3}I_{1p}^d(\tau_s^*)]. \tag{16}$$

In equation (16), A_p^* is the mean infrared surface albedo and the term $\mu_o F_{op} e^{-\tau_s^*/\eta\mu_o}$ that represents the infrared attenuated field, has been neglected. The zero-order moment taken over the solar and the infrared regions, yield the relations $J_s^d(\tau_s) = I_{os}^d(\tau_s)$ and $J_p^d(\tau_s) = I_{op}^d(\tau_s)$.

4. The radiative model for the atmospheric temperature regime

The effective temperature T_o associated with the external solar flux of the incident radiation field F_{os} is given by (Fabry, 1917)

$$\sigma_o T_o^4 = \pi F_{os} \tag{17}$$

From the condition of radiative equilibrium between the incoming external solar flux and the outgoing infrared flux of the atmosphere, it follows that $T_o = T_e$, where T_e is the effective temperature of the Earth's atmosphere.

At any optical vertical depth τ in the atmosphere under local thermodynamical equilibrium (LTE), the absolute temperature $T(\tau)$ is

$$\pi B_p(\tau_s) = \sigma_o T^4(\tau_s), \tag{18}$$

where $B_p(\tau_s)$ is the Planck function in the prevailing atmospheric infrared emission and σ_o is the constant of Stefan-Boltzmann.

Normalizing to T_e, one can find the atmospheric absolute temperature at any vertical depth.

Thus

$$\tilde{T}(\tau) = \frac{T(\tau)}{T_e} = [B_p(\tau)/F_{os}]^{1/4}, \tag{19}$$

or

$$\tilde{T}(\tau) = \{\frac{1}{F_{os}}[J_p(\tau) + \eta(1 - \varpi_s)J_s(\tau)]\}^{1/4} \tag{20}$$

where J_p and J_s are the infrared and solar total mean intensities. Taking into account the diffuse and attenuated component the eq (20) yields

$$\tilde{T}(\tau_s) = \{\frac{J_p^d(\tau_s)}{F_{os}} + \eta(1 - \varpi_s)[\frac{J_s^a(\tau_s)}{F_{os}} + \frac{J_s^d(\tau_s)}{F_{os}}]\}^{1/4}, \tag{21}$$

where $J_s^a(\tau_s) = (F_{os}/4)e^{-\tau_s/\mu_o}$. Eq. (21) can also be written in the form

$$T(\tau_s) = T_e\{I_{op}^d(taus_s) + \eta(1 - \varpi_s)[I_{os}^d(\tau_s) + (1/4)e^{-\tau_s/\mu_o}]\}^{1/4}, \tag{22}$$

where $I_{op}^d(\tau_s)$ is given by equation (13) and $I_{os}^d(\tau_s)$ by equation (3).

Equation (22) establishes that the radiative atmospheric temperature depends on both the infrared and solar diffused radiations fields, on the attenuated solar radiation field, on the radiative or optical properties of the atmospheric medium, and finally, on the effective temperature of the terrestrial atmosphere.

The intrinsic parameter T_e is the only indicator with external nature that may be used to define a general criterion for the existence of a greenhouse or anti-greenhouse effect. Therefore, one can define the Greenhouse Factor, $G(\tau)$, which becomes $\tilde{T}$ if $T_o = T_e$.

If the atmospheric absolute temperature, is maintained above (or below) T_e, then $G(\tau_s) > 1$[or $G(\tau_s) < 1$], there will be a heating or a greenhouse (or cooling or anti-greenhouse) effect. Note that this is not the only criterion for such effect; see, for example, the works by Öpik (1970), Goody and Walker (1972), Chamberlain (1980), Golitsyn and Ginsburg (1985) and Turco *et al.* (1990). The above authors use the effective temperature of the planet surface as the reference temperature.

5. Results and discussion

At first, the model was applied as usual in radiative transfer calculations using the only value for air's GP found in the literature which was in an indirect form. This value is $\eta = 5$ and was calculated using the data provided by Otterman (1983). In choosing an appropriate set of values for the other input parameters for this medium such as the single scattering albedo, asymmetry factor and the total vertical optical depth of the layer in the solar range, the data provided in Table 2 of the work by Shettle and Weinman (1970) were used to obtain also in an indirect form (averaging) its values, that is, $\varpi_s = 0.708$; $g = 0.414$ and $\tau_s^* = 0.28$ (Table 2). The incidence of the external radiation was initially perpendicular to the layer so that $\mu_o = 1$. This value provides the maximum influence in the radiation-matter interaction. For the mean value of the surfaces albedos, in the solar and in the infrared fields, respectively, the following values were found: for snow, $A_s^* = 0.8$ and $A_p^* = 0.5$ (Otterman, 1983); for vegetation, $A_s^* = 0.1$ and $A_p^* = 0.5$ (Otterman, 1983); for ocean, $A_s^* = 0.06$ and $A_p^* = 0.01$ (Goody, 1964; Payne, 1972; Budyko, 1974); and for desert, $A_s^* = 0.33$ and $A_p^* = 0.45$ (Henderson-Sellers and Wilson, 1983).

TABLE 2. Values of the opacity, asymmetry factor and single scattering albedo for an air clear layer and for three different wavelengths in the solar range (Shettle & Weinman, 1970). The solar average values for these three parameters are also given.

	0.76 μm	0.4 μm (blue)	0.7 μm (red)	Average
τ^*	0.492	0.291	0.06	0.280 (τ_s^*)
g	0.553	0.180	0.511	0.414 (g_s)
ϖ	0.124	1.000	1.000	0.708 (ϖ_s)

The results for this first case are shown in Figure 1. The surface air temperatures are evidently unrealistic for each particular surface considered. All four local temperature profiles fall below the atmospheric effective temperature of the Earth (255 K), indicating a cooling or anti-greenhouse effect ($G < 1$) that in turn is not realistic.

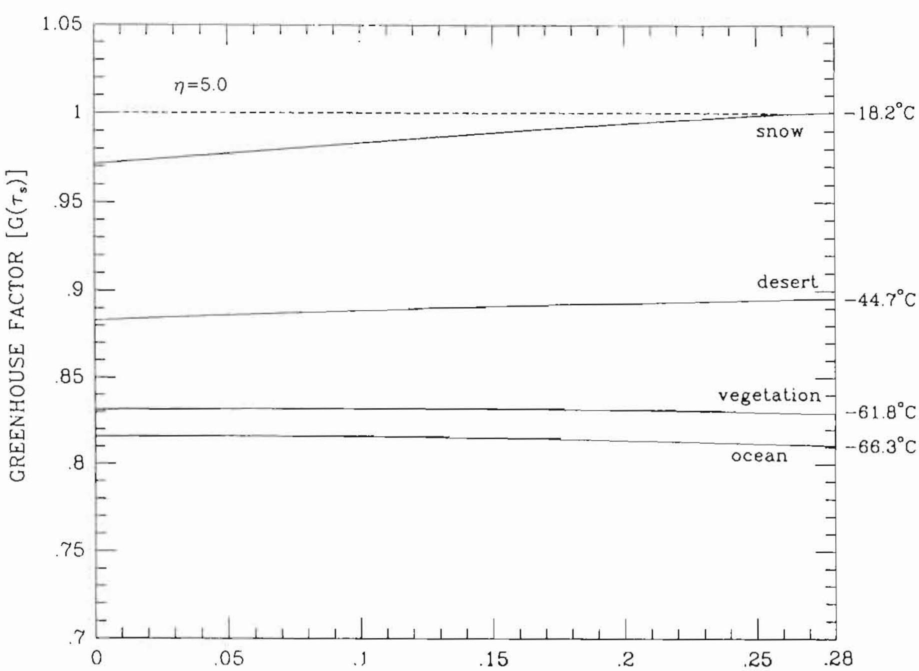

Fig. 1. Temperature profile for a clean and dry ideal air (oxygen and nitrogen) layer with $\varpi_s = 0.708$ y $g = 0.414$, $\mu_o = 1.0$, $\eta = 5.0$ and $\tau_s^* = 0.28$.

Next, the inverse procedure was applied varying the value of the GP looking for a better air surface temperature, depending on the surface taken into account. Figures 2 and 3 show the results for $\eta = 9$ and for $\eta = 10$, for $\mu_o = 1$. In these graphs, the local temperature profiles show a heating or greenhouse effect for desert and snow, but a cooling (anti-greenhouse effect) for vegetation and ocean surfaces. The air surface temperature attained in these cases are somehow unrealistic, mainly for surfaces other than snow (which yields 20.1 C and 27.8 C, respectively). However, in the four profiles for each of these figures, the local atmospheric temperature increases smoothly with the optical depth (or decrease with height), as is realistically expected in the troposphere. Taking this into account, a quasi-constant lapse rate, as in an ideal density-constant atmosphere, was found for all curves (McCartney, 1976).

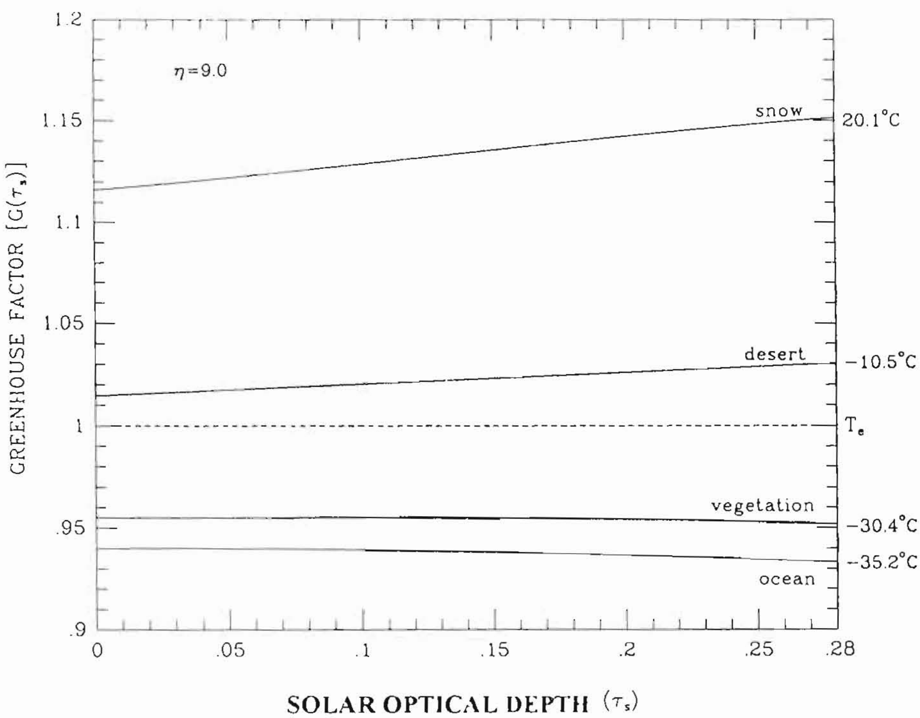

Fig. 2. As Figure 1 for $\eta = 9.0$.

Subsequently the value of the GP was varied up to the air surface temperature matching value of approximately 20 C, for a plane surface covered by vegetation. This is the expected air temperature, in radiative equilibrium, over a surface of this type and that is, in turn, a few degrees higher than the real global time-space averaged air surface temperature for Earth which value is 15 C. The value of the cosine of the incidence angle of the solar radiation was varied to $\mu_o = \sqrt{3}/3$, the amount equivalent to an incident solar radiation isotropic field integrated with a quadrature point, because it appropriately takes into account all angles of incidence to include all daylight irradiation. In this case, a value of $\eta = 26$ was estimated.

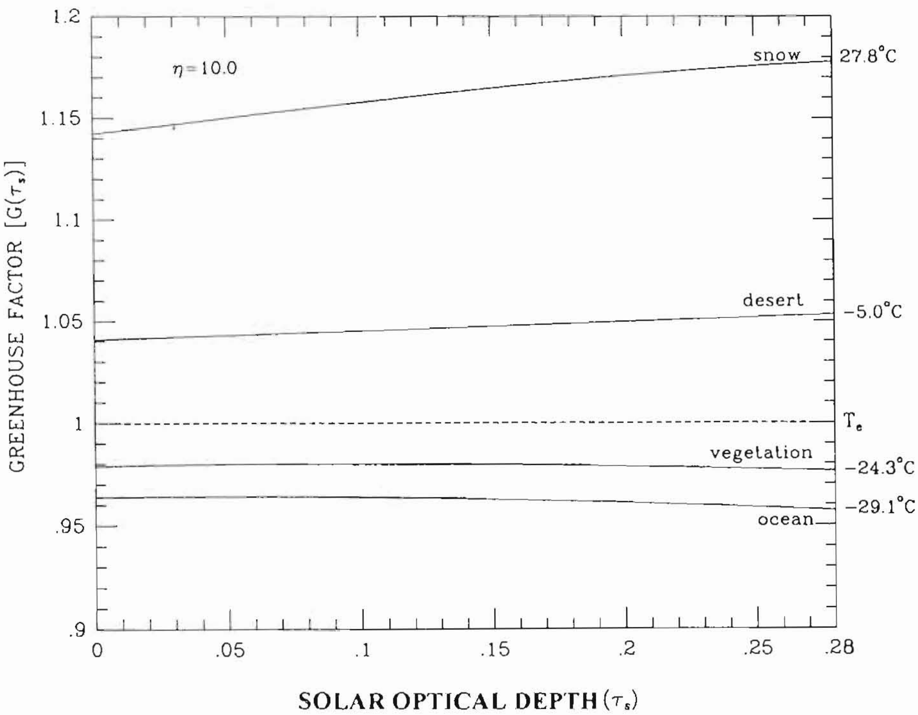

Fig. 3. As Figure 1 for $\eta = 10.0$.

Figure 4 shows the local temperature profiles for this value of GP. The air layer becomes an inverse one because the atmospheric temperature, for all four curves, increases with height, that is to say, decreases with optical depth. This again unquestionably justifies the use of a radiative model since convection is absent under this circumstance. In addition, η is a typically radiative parameter that has therefore, to be calculated using a model of this type. The idea of choosing η as a free parameter is based on the fact that it can vary in a wider range of values (for example between 10^{-3} and 10^{3}) while the other radiative parameters like ϖ_s, and g, can only vary from zero to one, and from -1 to 1, respectively. Therefore, as the temperature profile depends explicitly on η (eq. 22), this variable is very sensitive to the greenhouse parameter.

The air surface temperatures obtained from the above situation are: 17.6 C for ocean, 20.1 C for vegetation and 35.4 C for desert. These values are typical ones for these kinds of surfaces. For snow, the result is 66.0 C that is theoretically acceptable if it supposed that the layer and this surface are at sea level in the tropical zone. At this point, it is interesting to note that according to Vaughan (1979) an air surface temperature up to 42 C has been observed by climbers on Mount Everest at heights of between 6500 and 7000 m, on a white plane glacier surface, along with a good daytime weather, and a transparent (dry) atmosphere. At night and with bad weather, the air surface temperature dropped to -30 C on the same site (Vaughan, 1979). Also, in another glacier, at a height of approximately 5000 m near Bolivar Peak (Venezuelan's Andes), such heating has also been observed during daylight (L. Durán. Personal communication, 1991).

The local temperature curves shown in Figure 4 are above the reference line for the atmospheric effective temperature; therefore, in this case, a greenhouse effect appears in the layer, which in turn, is in agreement with the same effect observed in the Earth's troposphere.

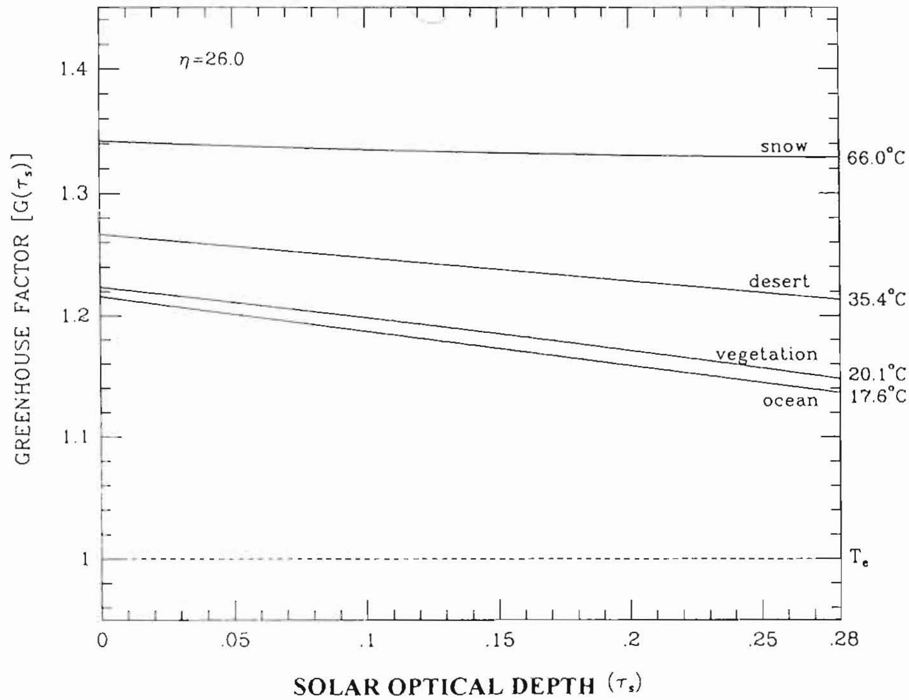

Fig. 4. As Figure 1 for $\mu_o = \sqrt{3}/3$ and $\eta = 26.0$.

Another result obtained in this work was an estimation of the total vertical depth for unpolluted, dry and free-particle air in the infrared region. By using eq. (10), one can calculate a value for τ^*. With $\tau_s^* = 0.28$ and $\eta = 26.0$ the result obtained for τ_p^* is 0.011. This result shows a good agreement with the mean value of the opacity taken over the wavelength range of $0.75 - 0.95\mu$m (near infrared) given by Kondratyev (1969) and, where infrared opacity for this medium is larger than the other one (thermal infrared) for this part of the spectrum (the major atmospheric constituents nitrogen and oxygen, because of their symmetry, possess no electric dipole transitions and hence no strong bands due to them occur in the infrared). Note the larger difference for the infrared opacity between this kind of air (0.011) and the real air of the Earth's atmosphere (~5), which includes water vapor and other greenhouse gases (Goody and Walker, 1972). Research by Fouquart (1988), suggests that since reflectivity is a nonlinear function of sun angle and optical thickness, and has strong variations within the usual spectral intervals, the parameterization of the spectrally-averaged optical thickness must be depend on the sun angle. From his off-line calculation for tropical to subartic atmospheres, and for solar zenith angles between 0 and 80 degrees, he obtained equations to calculate the optical thickness for this air in the intervals of 0.68 and 4μm (near infrared) and of 0.25 and 0.68μm (solar), respectively. By applying these equations, shown in Appendix B, to the value $\mu_o = \sqrt{3}/3$, the approximate values of 0.010 and of 0.20 for the optical thickness in the infrared and the solar ranges, respectively, were found. This indicates that the values of the GP would be 20, close to the value obtained in this work. However, in an earlier paper, Fouquart (1986) estimated a value of 0.007 for the opacity in the near infrared, but the same for the opacity in the solar range. Taking into account these opacities, the value of the GP would be now 28, a value even closer to our GP value of 26.

Similarly, a previous study by Penndorf (1957) indicates that, for a standard atmosphere in an isothermal condition, the optical thickness averaged over the near infrared range from his Table VI, is 0.011. From this table and, on an average basis, a value of 0.22 was found for the optical thickness in the solar range. With these data, a value of 20 for the GP has been calculated being again close to our value.

The value of 5 given indirectly by Otterman (1983) for the GP has been used as a first indicative value of the ratio between solar and infrared optical depth for air. Yet this author does not pretend to give an accurate value for this parameter, because it is not necessary for his work. Hence there is no comparison between this value and that obtained in this paper.

This method of calibration using only a radiative model and radiative "observed" surface air temperature, seems to be not physically coherent because one has to use a radiative-convective model that has shown the importance of convection to reproduce the observed temperature profile in the real lower atmosphere. In this respect, it has to be emphasized that this work has not set out to do this. However, the local temperature curves associated with the best value of the GP estimated are physically coherent from a radiative point of view.

Finally, a comparison among other graphs, corresponding to Figures 1, 2 and 3, respectively, is not valid since an investigation for the best value of the GP has been the main aim of this study.

6. Summary and conclusions

The relative extinction of solar and infrared radiation in the Earth's atmosphere by air molecules, in a particle-free, dry and unpolluted state, has been estimated. This was calculated by taking into account, in a first approach, that air only consists of the main or major components (oxygen and nitrogen). This was done by means of numerical experiments using a homogeneous, plane-parallel, and time-independent grey radiative model, which was processed with the Eddington approximation as a solution to the radiative transfer equation, both in the solar and the infrared spectral regions. To calibrate this model, four types of surface (snow, desert, vegetation and ocean) were used with average albedos known in these spectral regions, adopting air surface temperatures according to values, which were based on an average radiative criterion, for a supposedly micro or local climatological environment.

It was found that the ratio of solar extinction and infrared extinction, referred to in general as Greenhouse Parameter (GP), has a value of 26, approximately five times the value of 5 estimated indirectly from the data provided by Otterman (1983). This value correlated very well with those of approximately 28 and 20, estimated from the data given by Fouquart (1986, 1988) and Penndorf (1957). Although its validation is constrained to the radiative model applied, therefore, it seem that the value of 26 is more accurate than the one initially available and, it can be used in more refined models describing physical properties of dry and clean air.

With this value of the GP, it was possible to obtain the infrared opacity for the air layer involved in this model. The result was 0.011 corresponding to the near-infrared interval. This means that the extinction coefficient is greater in the solar than in the infrared, leading to the conclusion that this medium is optically thinner in the infrared spectral region than in the solar one. The result referred to above for infrared opacity, in the near-infrared, is in a good agreement with those given by Kondratyev (1969), Fouquart (1986, 1988) and, Penndorf (1957), respectively. Both grey or mean extinction coefficients can be calculated via eq 10 if the geometrical depth or height of the layer is given. In accordance with McCartney (1976), a constant-density or homogeneous atmosphere model for standard conditions at sea level, would have a height of 8.00 km. Bearing this value in mind for the geometrical depth or thickness of the air layer, and applying eq. 10, the mean extinction coefficient of air molecules, in the near infrared region, is 1.375×10^{-3} km^{-1}. Note that the mean Rayleigh total scattering coefficient for pure air (at 288.15 K and 1013 mb), taken over the same spectral range, is 0.860×10^{-3} km^{-1} (McCartney, 1976. See his Appendix I). Hence, the extinction coefficient by absorption is 0.515×10^{-3} km^{-1} in the same range (see eq. A2). With this same value for the geometrical depth and same equation, the mean extinction coefficient of air molecules, in the solar region, is 3.500×10^{-2} km^{-1}, that is to say, 26 times that of the near infrared. Note that the mean Rayleigh total scattering coefficient for pure air (at 288.15 K and 1013 mb), taken over the solar range, is 3.350×10^{-2} km^{-1} (McCartney, 1976. See his Appendix I). Thus the extinction coefficient by absorption is 0.150×10^{-2} km^{-1} in this range (see eq. A2). All of these results confirm or corroborate the correctness of the value obtained here for the GP of the pure air.

Both the computed and measured air surface temperatures of 66 C and 42 C for snow and ice. respectively, can be explained due to the high albedo in the solar range (0.8) and the middle albedo in the infrared region (0.5) for a white surface of this type, together with the opacities of the air molecules in the solar as well as in the infrared regions. Ice is very weakly absorptive in the visible (minimum absorption at $\lambda = 0.46$ μm) but has strong absorption bands in the near infrared (Warren, 1982). It seems paradoxical that a larger surface albedo gives a smaller air surface temperature; more solar radiation reflected at the surface gives smaller surface temperature and, therefore, less thermal or near infrared energy is available to the adjacent air by absorption (that is the well-known positive feedback between temperature and surface albedo). However, note that whatever thermal and/or near infrared energy is available, air molecules extinguish in the solar range 26 times more than in the infrared. In radiative equilibrium, an ice-snow surface emits more energy in the infrared than in the solar range contributing to the radiative heating of the air. This, as a negative feedback between air surface temperature and surface albedo, could in turn strongly heat up the air layer in contact with a white surface like a glacier. Further investigations may show the conditions in which these opposite feedbacks would prevail or dominate each other.

Finally, it is concluded that when the GP value is difficult to evaluate for a different medium, for example CO_2 (Martin, 1987), this method would be suitable.

The values reported here, as basic physical properties of "air molecules", must be included among those found in Table 1.

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APPENDIX A

The basic equation of radiative equilibrium is,

$$\int_0^\infty \kappa_\nu J_\nu d\nu = \int_0^\infty \varepsilon_\nu d\nu, \tag{A1}$$

where J_ν is the total mean intensity (attenuated plus diffused), κ_ν is the total extinction coefficient and ε_ν is the total emission coefficient, all at the same frequency ν.

Equation (A1) implies simply that every element of mass in the medium, in radiative equilibrium, extinguishes exactly as much radiation as it emits, in all frequencies.

Taking into account that the total extinction coefficient can be expressed by,

$$\kappa_\nu = \alpha_\nu + \sigma_\nu, \tag{A2}$$

where, α_ν and σ_ν are the true absorption and scattering coefficients, respectively, at the frequency ν, and, that the total emission coefficient can be expressed by,

$$\varepsilon_\nu = \varepsilon_\nu^t + \varepsilon_\nu^s, \tag{A3}$$

where, ε_ν^t is the true emission coefficient and, ε_ν^s is the emission coefficient by scattering, the equation (A1) can be expressed by,

$$\int_0^\infty (\alpha_\nu + \sigma_\nu) J_\nu d\nu = \int_0^\infty (\varepsilon_\nu^t + \varepsilon_\nu^s) d\nu. \tag{A4}$$

Evidently, the energy extinguished by scattering in all frequencies by an element of mass in the medium appears as energy emitted by the same element scattered in all frequencies; so, from equation (A4),

$$\int_0^\infty \alpha_\nu J_\nu d\nu - \int_0^\infty \varepsilon_\nu^t d\nu = \int_0^\infty \varepsilon_\nu^s d\nu - \int_0^\infty \sigma_\nu J_\nu d\nu = 0. \tag{A5}$$

Therefore, from this equation,

$$\int_0^\infty \alpha_\nu J_\nu d\nu = \int_0^\infty \varepsilon_\nu^t d\nu. \tag{A6}$$

Equation (A6) implies simply that every element of mass in the medium, in radiative equilibrium, absorbs exactly as much radiation as it thermally emits, in all frequencies.

Under this condition together with the condition of local thermodynamics equilibrium and, applying the Kirchhoff-Planck laws, the equation (A6) can be expressed by,

$$\int_0^\infty \alpha_\nu J_\nu d\nu = \int_0^\infty \alpha_\nu B_\nu d\nu, \tag{A7}$$

where B_ν, is the Planck function at the frequency ν.

Integration of the equation (A7) in the solar and infrared spectral ranges, respectively, gives,

$$\alpha_s J_s + \alpha_p J_p = \alpha_s B_s + \alpha_p B_p, \tag{A8}$$

where J_s and J_p are the total mean intensities in the solar and the infrared spectral ranges, respectively. Thus, in radiative equilibrium, this equation means that the absorption in the solar range plus the absorption in the infrared range is equal to the emission in the solar range plus the thermal emission in the infrared range. In this case, by neglecting the thermal emission in the solar range ($\alpha_s B_s \cong 0$) the equation (A8) by this approximation becomes,

$$\alpha_s J_s + \alpha_p J_p = \alpha_p B_p, \tag{A9}$$

and from this equation,

$$B_p = (\alpha_s/\alpha_p) J_s + J_p. \tag{A10}$$

Using the definition of single scattering albedo $[\varpi_s = \sigma_s/\kappa_s]$ and the equation (A2), both in the solar range, the absorption coefficient in the same range, can be written as,

$$\alpha_s = (1 - \varpi_s)\kappa_s. \tag{A11}$$

Substitution of this last expression in equation (A10) yields,

$$B_p = (1 - \varpi_s)(\kappa_s/\alpha_p) J_s + J_p. \tag{A12}$$

The scattering in the infrared can be neglected so that $\sigma_p \cong 0$. Thus the greenhouse parameter, by definition (eq 10) and, applying the equation (A2) in both the solar and the infrared ranges, is reduced to,

$$\eta = \kappa_s/\kappa_p = (\alpha_s + \sigma_s)/(\alpha_p + \sigma_p) \cong \kappa_s/\alpha_p. \tag{A13}$$

Substituting from equation (A13) into equation (A12),

$$B_p = J_p + \eta(1 - \varpi_s) J_s. \tag{A14}$$

APPENDIX B

The parameterization of the spectrally-averaged optical thickness, obtained by Fouquart (1988), depending on the sun angle, for tropical to subartic atmospheres and for solar zenith angles between 0 and 80 degrees, are:

$$\tau_s(\mu_o) = 0.042897 + 0.8907\mu_o - 2.8856\mu_o^2 + 5.2274\mu_o^3 - 4.6917\mu_o^4 + 1.6165\mu_o^5, \tag{B1}$$

for $0.25 \leq \lambda \leq 0.68$ μm, and

$$\tau_p(M) = 0.00697 + 0.01733M - 0.08509M^2 + 0.24826M^3 - 0.30203M^4 + 0.12966M^5, \tag{B2}$$

for $0.68 \leq \lambda \leq 4$ μm, where $M = 1 - \mu_o$.

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