# SENSITIVITY OF THE RADIATION FIELD TO CLOUD COVER

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## INTRODUCTION

More than a complete cloud experiment which would have needed a general circulation model (GCM) with fully interactive clouds and radiation, this paper describes a study of the sensitivity to changes in the cloud cover of the radiation code presently used in the GCM of the Laboratoire de Météorologie Dynamique (LMD). We have the ambition of conducting such complete an experiment in the near future, however, as in the LMD's GCM the radiation is still calculated for a zonal cloud climatology, the present sensitivity study must be considered as a test of the consistency of the radiation code and not as a realistic climate study.

## I - DESCRIPTION OF THE RADIATION CODE

The longwave radiation code has been taken from Katayama (1974) ; it is based on global emissivities calculation and includes water vapeur carbon dioxide and ozone absorption. Clouds are considered as black bodies (except the cirrus clouds for which we set a grey emissivity between 0.3 and 1.0), with random overlapping ; the e type absorption is presently not taken into account and the effect of temperature inhomogeneities upon absorption are also ignored. This is certainly an important limitation and a more accurate long wave radiation code is being developed. As shown by Chou and Arking (1980), the sensitivity of the water vapor absorption to temperature variations can be handled very accurately by using only two sets of parameters : one for the band centers and one for the band wings. However the temperature variation is more particularly important for the $CO_2$ absorption for which a similar parameterization has not yet been derived. We are, now, performine a critical study of the different longwave parameterizations with the objective of finding out which degree of sophistication is actually needed for the purpose of climate modeling.

1 - The shortwave radiation code is called SUNRAY. A complete description can be found in Fouquart and Bonnel (1980). Its principal characteristics are the following

. water vapour, carbon dioxide, ozone and cloud drop absorption are taken into account

. scattering is solved by means of the Exponential Kernel Approximation

. partial cloudiness is considered

. the computational scheme does not depend on the parameterization of gaseous absorption

### 1.1 - Distribution of the amount of absorber.

Interactions between scattering and molecular band absorption are dealt with by the method of the distribution of the amount of absorber (see also Geleyn and Hollingsworth (79)). We define it as the probability distribution p(u) du that a photon contributing to the continuum intensity in the conservative case, has encountered a total amount of absorber between u and u + du. Thus, the monochromatic intensity $I_\nu$ corresponding to the absorption coefficient $k_\nu$ is

$$I_\nu = I_c \int_0^\infty p(u) e^{-k_\nu u} \, du \quad : (1)$$

Obviously, the distribution as well as the intensities is a function of optical depth and directions. Over a finite spectral interval the mean intensity $I_{\Delta\nu}$ can be related to the mean transmission function $\psi_{\Delta\nu}(u)$

$$I_{\Delta\nu} = I_c \int_0^\infty p(u) \, \psi_{\Delta\nu}(u) \, du. \qquad (2)$$

The variation of the absorption coefficient with temperature and pressure can be dealt with by the "scaling approximation" or by a generalization of the Curtis Godson approximation (Buriez and Fouquart (1980)

$$d\bar{u}(z) = \left(\frac{p(z)}{p_0}\right)^{\alpha_v} \left(\frac{T(z)}{T_0}\right)^{\beta_v} u(z)\, dz \qquad (3)$$

where u(z) is the density of absorbing molecules at altitude z.

With this method, the transmission functions do not need to be fitted by a sum of exponentials like it is in most of the shortwave parameterizations (Lacis and Hansen (1974) for example). In SUNRAY, we approximate the averaged transmission functions by means of Pade Approximants

$$\psi(u) = \int_{0,25}^{4\,\mu m} F_0(\lambda)\, \psi_\lambda(u) d\lambda \Big/ \int_{0,25}^{4\,\mu m} F_0(\lambda) d\lambda = \sum_{i=0}^{N} a_i u^i \Big/ \sum_{i=0}^{N+1} b_j u^j \qquad (4)$$

The distribution function (du) can be calculated using an inverse Laplace Transform technique, but for the purpose of climate modeling that method would be much too time consuming. In their radiation code, for the ECMWF' GCM, Geleyn and Hollingsworth (1979) make the following approximation

$$I_{\Delta v} \simeq I_c\, \psi_{\Delta v}\, (\langle u \rangle) \qquad (5)$$

where $\langle u \rangle = \int_0^\infty u p(u)\, du = \mu,$ $\qquad (6)$

is the mean amount of absorber.

Because of the properties of the Laplace Transform, $\langle u \rangle$ is also equal to

$$\langle u \rangle = -\frac{d}{dk} \left[ \mathrm{Ln} \left(\frac{I(k)}{I_c}\right) \right]_{k=0} \qquad (7)$$

Theoretically, it can be shown that the mean intensity is given exactly by this approximation in case of weak absorption ; however, for strong absorption the transmission depends on another moment of the distribution function :

$$\mu_{1/2} = \int_0^\infty u^{1/2}\, p(u)\, du \qquad (8)$$

This is directly related to the square root regîme of the clear case.

As a consequence of this dependence of the equivalent amount of absorber on the regime of absorption, the approximation which makes use of the mean amount of absorber overestimates absorption in the regions of strong absorption. We obtained a significant improvement in SUNRAY where the effective amount of absorber is numerically calculated from equation (7) which is now written as

$$u_{ef} = k^{-1} \ln (I_c/I(k))$$

and k is fixed such as, for the whole solar interval

$$\psi(u) \simeq a_1 + a_2\, e^{-ku}$$

The accuracy of this last fit is of course rather bad but it allows a rather good determination of $u_{ef}$ which does not depend strongly on the particular choice of k.

### 1.2 - Multicloud case

We checked our radiation code by comparing its results with those of Lacis and Hansen's code but also using models with a high spectral resolution ($\Delta\lambda = 0{,}01\ \mu m$).

When we used the approximation of a single effective amount of absorber for a model of atmosphere with several cloud layers we found some inaccuracy in the heating rate corresponding to the upper (Cirrus) clouds ; for the lower clouds the results were much better ; this is related to the distribution of water vapour with height and can easily be corrected.

Let us consider the simple two clouds model (the case of high ground reflectivity is quite similar). Obviously, in the upward light above the upper cloud (1) part of the photons (say $I_1$) have been directly reflected by the upper cloud without interaction with the underlying atmosphere, another part ($I_2$) have passed through cloud 1 and undergone one (or several) reflection on cloud 2. The relative importance of $I_2$ compared to $I_1$ depends upon the optical thickness of cloud 1.

Since the water vapour concentration rapidly decreases with height, the photons directly reflected by cloud 1 cannot encounter a large amount of absorber, in opposition to those reflected by cloud 2. In other words if $p_1(u)$ and $p_2(u)$ are the distributions of absorbing amount associated with $I_1$ and $I_2$ respectively, $p_1(u)$ is centered around weak values of u compared to $p_2(u)$. Moreover, because of the multiple reflections between clouds 1 and 2 and because of the repartition of the absorber with height $p_2(u)$ is much smoother than $p_1(u)$.

Now, we can write the overall distribution p(u) as

$$Ip(u) = I_1 \, p_1(u) + I_2 \, p_2(u) \qquad (9)$$

where $I = I_1 + I_2$

The mean reflected intensity (or flux) is thus

$$I_{\Delta\nu} = I_1 \int_0^\infty p_1(u) \, \psi_{\Delta\nu}(u) \, du + I_2 \int_0^\infty p_2(u) \, \psi_{\Delta\nu}(u) \, du \qquad (10)$$

which has to be approximated by

$$I_{\Delta\nu} \simeq I_1 \, \psi_{\Delta\nu}(u_{e_1}) + I_2 \, \psi_{\Delta\nu}(u_{e_2}) \qquad (11)$$

with $u_{e_2} > u_{e_1}$.

Note that a direct calculation of the mean intensity using an overall effective amount of absorber $u_e$ leads to inaccurate results since for weak absorption $u_e \simeq u_{e_2}$ but for strong absorption $u_e \simeq u_{e_1}$.

For a system of n cloud layers, to be quite logical, we would have to separate the contributions of each individual cloud to the overall reflected light, however as already noticed the multiple reflections which occur below the uppermost cloud have a smoothing effect on $p_2(u)$ so that even for a multicloud system the mean intensity can still be calculated with a reasonable accuracy as in (11) where now, $I_2$ stands for the contribution to the total reflected light I of the photons which have passed through cloud 1 and undergone reflections on cloud 2, 3 etc...

3.3 - Molecular scattering

Rayleigh scattering contributes for a significant part to the planetary albedo. In a GCM in which partial cloudiness is considered it cannot be neglected. However the total optical thickness of the atmosphere is weak (we found $\tau_R \simeq 0{,}06$ as the best fit to the computations using high spectral resolution) and it seems reasonable to neglect the interactions between absorption and scattering.

Since the Rayleigh phase function is symmetrical we equally divide the diffuse radiation between the reflected and transmitted parts. The reflection R and transmission T of a layer of optical thickness $\tau$ are thus given by

$$R \simeq (1 - e^{-\tau/\mu_0})/2 \qquad (12)$$

$$T = 1 - R \qquad (13)$$

with $\tau/\mu_0 \ll 1$, the reflection is approximately

$$R \simeq \tau/2\mu_0 \qquad (14)$$

we used this approximation for all $\mu_0$, with a slight modification needed to avoid too large reflectivities for $\mu_0 \ll 1$

$$R \simeq 0.5\ \tau/\ (\mu_0+\tau) \qquad (15)$$

In our computational scheme we compute layer transmissions T and reflections at interface R in such a way that the multiple reflections between two successive layers cannot be exactly taken into account. The combined reflectivity and transmissivity of two successive layers 1 and 2 are then approximated by

$$R_{12} = R_1 + R_2 - R_1R_2 \qquad (16)$$

$$T_{12} = (1-R_1)\ (1-R_2) = T_1T_2 \qquad (17)$$

where $R_1$ and $R_2$ stand for the reflections of layer 1 and 2 respectively.

A comparison with the exact reflectivities of the whole atmosphere, computed by the Spherical Harmonics method is given in table ( I ) for four solar zenith angles and two ground albedos. The approximate reflectivities result from the adding of the eleven layers of the GCM. So that the multiple scatterings are approximately taken into account using (16). This expression overestimates the combined reflectivity for $R_2 \neq 0$ with a maximum discrepancy which occurs for $R_2 \simeq 0,5$ for the layer zenith angle. This corresponds to a 2%error on the global reflectivity for $\theta_0 = 15°$, and about 1 % when averaged over a whole day. Keeping in mind that the lowest values of the daily averaged solar zenith angle are associated with high ground albedo and that high reflectivities at low latitudes are due to the presence of clouds our approximation appears reasonable.

TABLE I - Reflectivity of the whole atmosphere for Rayleigh scattering Exact calculations have been performed with the Spherical Harmonics method. Mean Rayleigh optical thickness : $\tau_R = 0.06$.

| $\theta_0$ | 15° | 60° | 75° | 85° | ground albedo $R_g$ |
|---|---|---|---|---|---|
| Exact<br>Approximate | 0.0301<br>0.0297 | 0.0564<br>0.0552 | 0.106<br>0.102 | 0.250<br>0.229 | 0 |
| Exact<br>Approximate | 0.504<br>0.515 | 0.518<br>0.528 | 0.544<br>0.551 | 0.614<br>0.617 | 0.5 |

The radiative effects of aerosols could be handled in a similar way. The major problem concerns the geographical repartition of aerosols, their optical properties and their concentration. It is not clear whether calculations including the aerosol effect is needed when we have so few informations about their properties and their concentration.

Cloud drop absorption

The absorption coefficient of liquid water is very small in the visible but it increases in the near infrared and, since there are about 30 % of the incoming solar flux in the 1μm to 4μm spectral range, the cloud drop absorption is generally not negligible.

Table (II) shows the results of high spectral resolution calculations using the spherical Harmonics Method, and the complete distribution of amount of absorber. The heating rate profiles are shown for the case without (HR1) and with cloud drop absorption (HR2) for two different cloud optical thicknesses. Obviously the major effect occurs in the cloud layer itself where the heating rate due to the cloud drop absorption rises 3.5°/day for $\tau_1 = 10$ (5°/day for $\tau_1 = 20$); the decrease of the cloud transmissivity is still noticeable principally for large optical thicknesses; in the upper atmosphere, the influence of the cloud absorption on the reflectivity appears very weak. In order to take the cloud drop absorption into account in the solar radiation scheme one has to define some equivalent single scattering albedo $\omega_e$ for the global solar range. Empirically, we found that $\omega_e$ is not very sensitive to the phase function but increases as a function of the optical thickness $\tau$ (this is directly related to the saturation of the liquid water absorption bands); $1 - \omega_e$ can be approximated with an accuracy better than 5% for the range $0 \leq \tau \leq 20$ using the empiral formula

$$1 - \omega_e(\tau) = 1.1 \times 10^{-3} + 4.10^{-3} \exp ( - 0.15\tau )$$

With such an accuracy, the uncertainty upon the heating rate remains of the order of 0.1 to 0.2°/day in the cloud layer and is negligible elsewhere.

TABLE II : Heating rate (°/day). Mid Latitude Winter, $\Theta_0 = 0°$.
A stratus cloud of optical thickness $\tau_1$ fills layer 8
HR1 : high resolution computations ($\Delta\lambda = 0.01\mu$) without cloud drop absorption
HR2 : high resolution computations with cloud drop absorption
SUNRAY : global computations with approximate cloud drop absorption.

| | $\tau = 1$ | | | $\tau = 10$ | | |
|---|---|---|---|---|---|---|
| LAYER | HR1 | HR2 | SUNRAY | HR1 | HR2 | SUNRAY |
| 1 | 6.15 | 6.15 | 5.80 | 6.79 | 6.79 | 6.35 |
| 2 | 1.25 | 1.25 | 1.25 | 1.57 | 1.57 | 1.51 |
| 3 | .65 | .65 | .65 | .76 | .76 | .75 |
| 4 | .69 | .69 | .82 | .73 | .73 | .86 |
| 5 | 1.31 | 1.31 | 1.35 | 1.38 | 1.34 | 1.39 |
| 6 | 1.63 | 1.63 | 1.65 | 1.76 | 1.72 | 1.74 |
| 7 | 1.29 | 1.29 | 1.33 | 1.53 | 1.47 | 1.52 |
| 8 | 1.94 | 2.56 | 2.58 | 2.52 | 6.01 | 6.30 |
| 9 | 1.49 | 1.39 | 1.31 | .77 | .61 | .66 |
| 10 | 1.30 | 1.23 | 1.22 | .70 | .57 | .64 |
| 11 | 1.25 | 1.19 | 1.20 | .69 | .56 | .64 |

2 - Model sensitivity to cloud cover

We also estimated the response of the radiation model to changes in cloudiness. Presently in the GCM, the cloud cover that is used in the radiative computations, is not generated by the model but fixed at climatological mean values taken from the cloud climatology of London (1957). In the present experiment, temperature and humidity profiles for the Northern hemisphere in winter and summer are also taken from London (1957). We distributed the clouds over three different layers corresponding to low, middle and high clouds. We also increased from eleven to fourteen the number of levels of the model for radiative computations in order to fall in with London's data. Moreover, for each cloud layer, the cloud optical thickness that is necessary for evaluating the shortwave fluxes, have been set fixed such as to restitute the mean experimental cloud reflectivities and transmittivities given by London.

2.1 - Fluxes at the top of the atmosphere

Figure (1) shows the variation of the zonal albedo for winter and summer in the Northern hemisphere. The calculated values are in good or even excellent agreement with the experimental values of Ellis and Vonder Haar (1976).
However the calculated albedo is slightly overestimated at low latitudes and there may be two reasons for this discrepancy : (i) the mean experimental cloud reflectivities and transmittivities of London may be overestimated, (ii) the cloud cover associated with convective activity may be overestimated in the cloud climatology. Any way, the discrepancy is not important as the difference in the absorbed solar energy does not exceed 10 $Wm^{-2}$.

Figure (2) compares the latitudinal profile of the infrared outgoing flux measured by Ellis and Vonder Haar, and those calculated from our radiation model for two values of the high cloud emissivity. Those two values (0.3 and 1.0) can be regarded as the lower and higher limits of the range over which the emissivity of cirrus clouds is found to vary according to recent observations (Platt et al., 1980,Griffith et al., 1980). Calculated longwave fluxes exhibit larger discrepancies when compared to the observed ones. The temperature and humidity profiles are obviously much more important for longwave than for shortwave calculations. However, the principal modulator of the longwave fluxes is the cloud cover. The good agreement with observations in the shortwave range suggests that the total cloud cover is probably good or at least consistent with the satellite observations. Consequently, the disagreement in the longwave range should be due to errors in the cloud top altitudes or in the distribution of the total cloudiness over the different kinds of clouds. Obviously, there are too

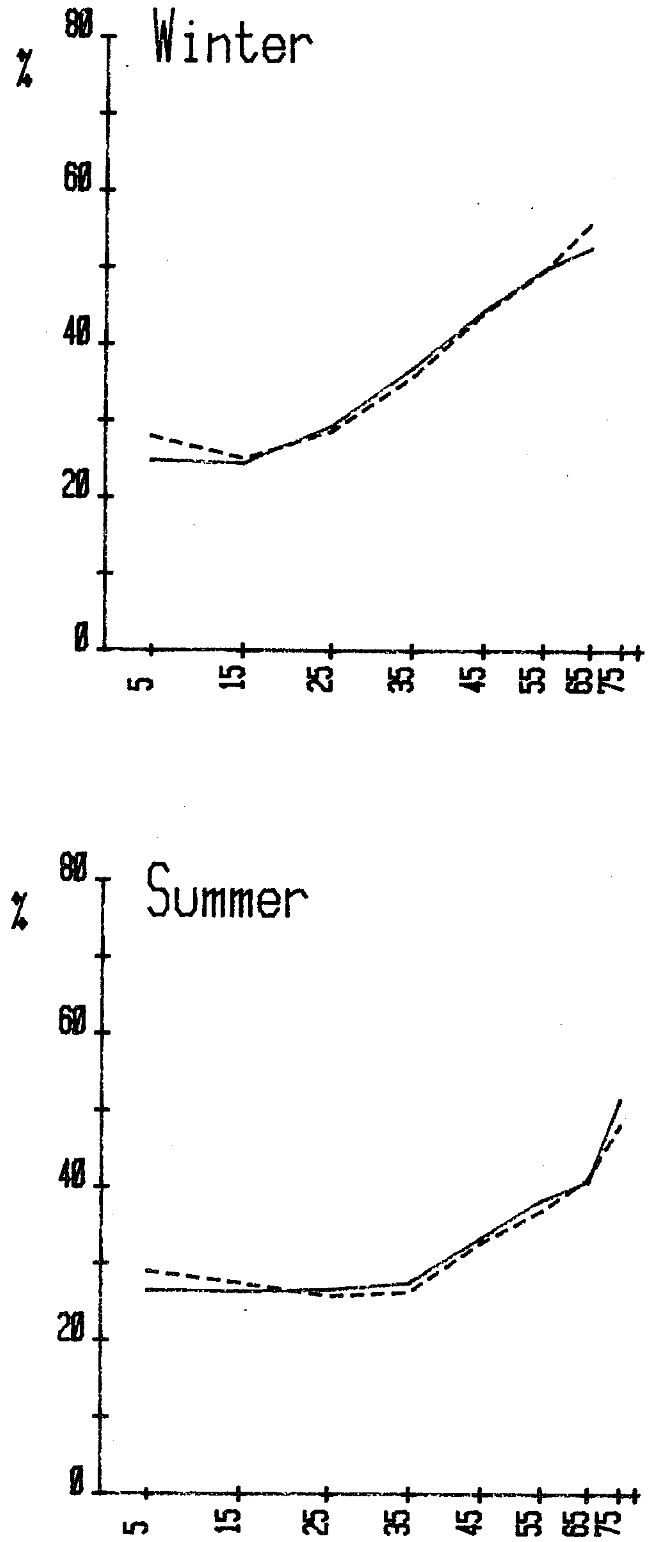


Figure 1 : Planetary albedo as a function of latitude.
( ---- calculated, —— Ellis and Vonder Haar)

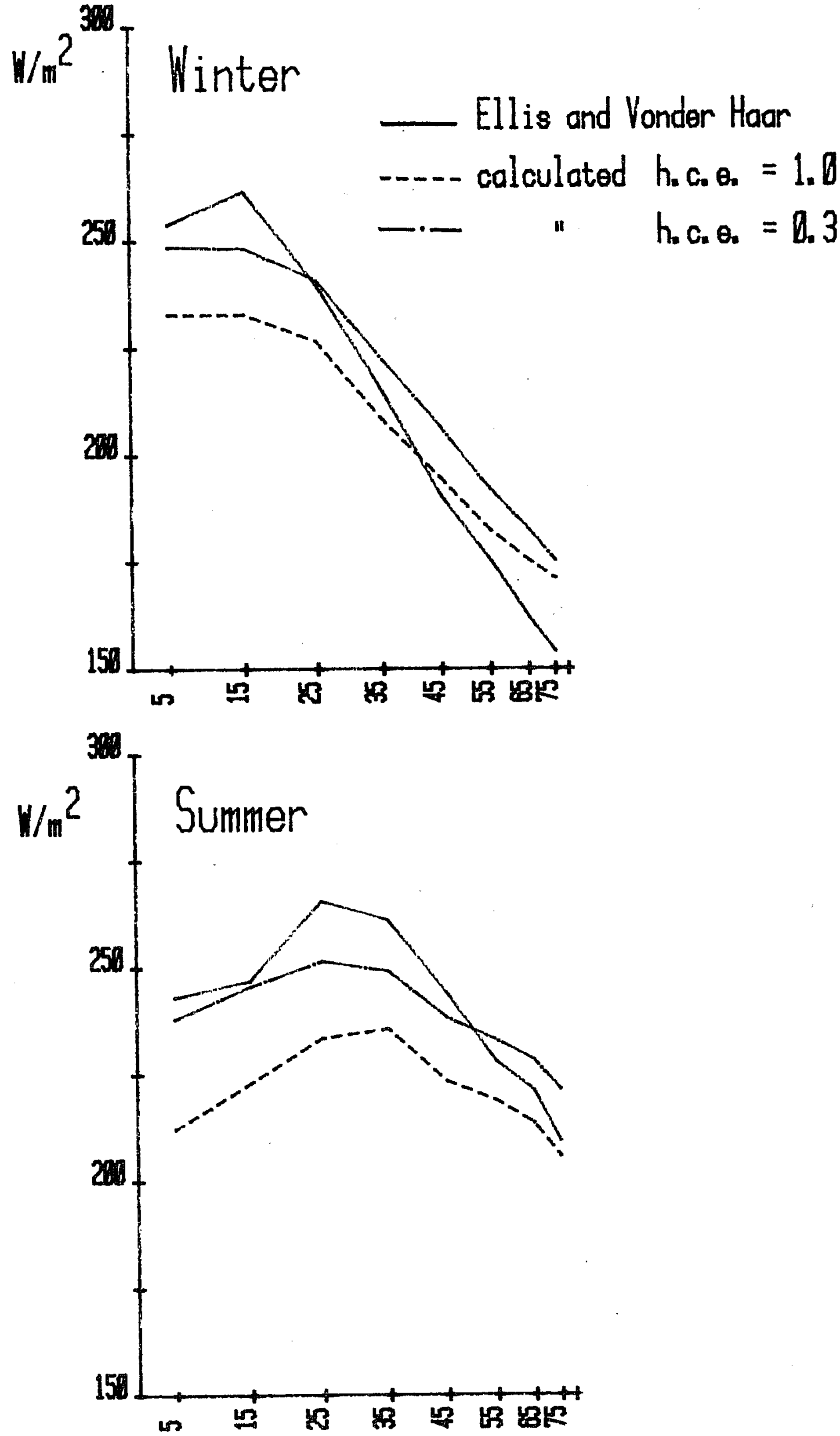


Figure 2 : Outgoing infrared flux as a function of latitude (h.c.e. stands for high cloud emissivity)

many uncertainties in the climatological data, and particularly in the cloud climatology. Presently, it is almost impossible to state whether the disagreement is due to the input data or the computational procedure itself. In other words, the satellite observations cannot be used to check and adjust the radiation calculations without a serious improvement in the cloud climatology.

## 2.2 - Changes in cloud cover

Climate sensitivity to cloud cover has generally been studied using either energy balance climate models (Van den Dool, 1980) or radiative convective models (Schneider, 1972, Coakley, 1977) with a general agreement on the overall effect, namely that an increase in the cloud cover would decrease the surface temperature.

Nevertheless, Cess (1976) correlated directly ground temperature and cloud cover with the outgoing longwave flux obtained from the radiative budget experimental data of Ellis and Vonder Haar (1976). He obtained the some what surprising result that the longwave greenhouse effect and shortwave albedo effect cancelled out. Such different a result is partially due to the method used by Cess for deriving his correlation. His method was biased by the variation of the cloud top temperature with latitude ; another bias lies in the fact that Cess actually observed variations in both cloud cover and cloud top temperature, while the other authors have kept the cloud top temperature fixed when varying the cloud cover in their calculations. More recently, Ohring and Clapp (1980) and Hartmann and Short (1980) have demonstrated that the albedo effect is actually larger than the greenhouse effect. This result has been obtained from satellite data by directly estimating the relative variation in the observed outgoing longwave flux and shortwave albedo $dF/d\alpha$ independently of any determination of the cloudiness.

In this paper, we do not only estimate the global sensitivity of our radiation model to changes in the total cloud cover, but we also discriminate between the influences of low, middle and high clouds, and we study the zonal variation of the climate sensitivity parameter.

At the top of the atmosphere, the radiative budget can be written as

$$B(\varphi) = Q_{abs}(\varphi) - F(\varphi)$$
$$= I_0(\varphi)\,[\,1 - \alpha(\varphi)\,] - F(\varphi)$$

where $I_0$ is the incoming solar flux, $\varphi$ the latitude, $\alpha$ the albedo of surface-atmosphere system, and F the outgoing longwave flux. We define a sensitivity parameter related to cloud species i as

$$\delta_i(\varphi) = \frac{\partial B(\varphi)}{\partial n_i}$$

where n stands for the cloud cover. Practically, $\delta$ is estimated by simply increasing $n_i$ by 10 percent in each latitude zone. The cloud radiative properties, cloud top altitude and temperature are kept constant as we deal with partial derivatives which only consider variations in the cloud cover.

Figure (3) shows the latitudinal variation of $\partial A/\partial n_i$ where A is the shortwave albedo. This variation mainly results from a competitive effect between the increase of the ground albedo with latitude and the sensitivity of the cloud albedo to the solar zenith angle. For random overlapping clouds, we have

$$\frac{d\alpha}{dn_i} = \alpha_i - \alpha_0 \prod_{j\neq i} (i - n_j)$$

where $\alpha_i$ is the albedo of the i-th cloud and $\alpha_0$ the albedo for the clear atmosphere. In summer, except for the very high latitudes, London's values for ground albedo do not change much with latitude, neither does the daily averaged solar zenith angle. In winter, the solar zenith angle effect is much stronger. However, as shown in Figure (3), this effect dominates only for high clouds, since for low and medium clouds it is damped out by the overlap of higher clouds giving the predominant effect to the ground albedo variation.

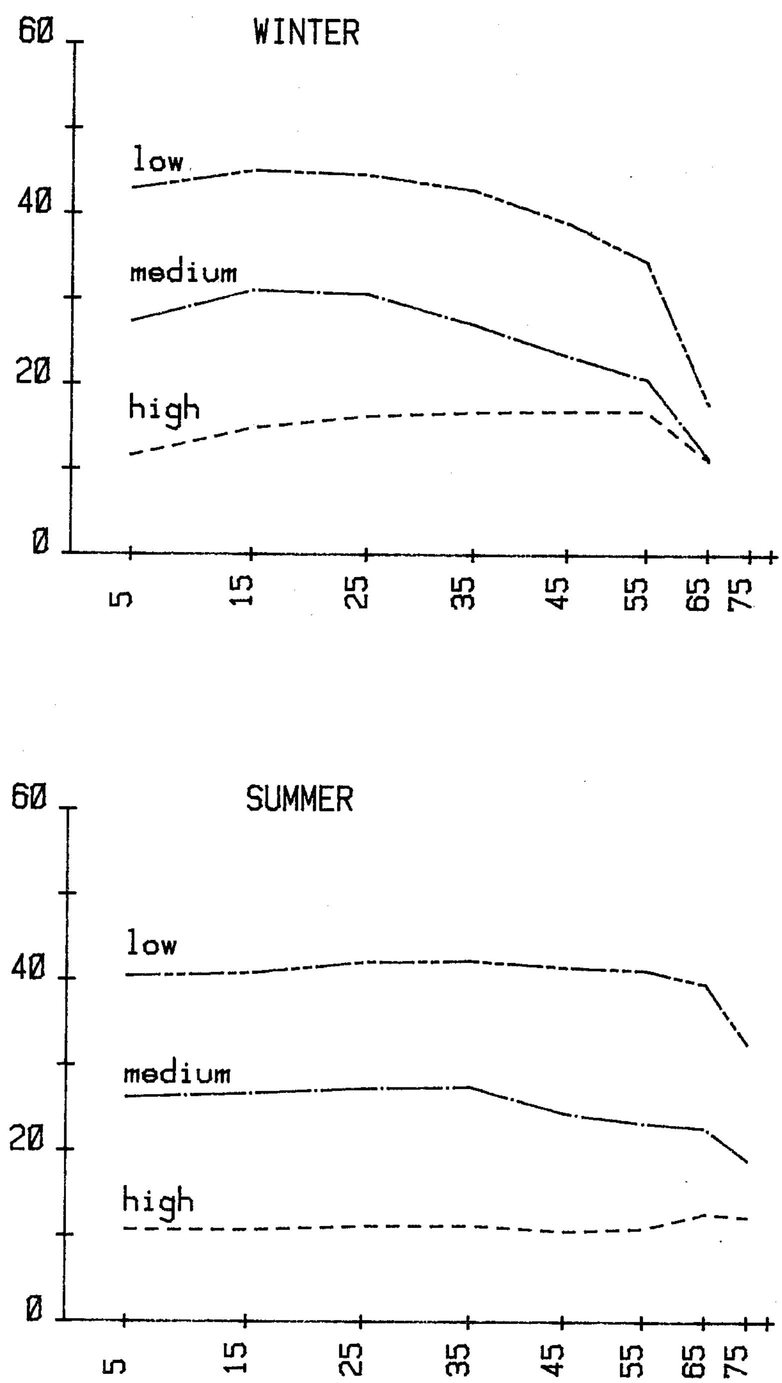


Figure3: $\partial\alpha/\partial n_i$ as a function of latitude

Figure (4) shows $\partial F/\partial n_i$ as a function of latitude. The so called greenhouse effect strongly depends on the temperature difference $T_s - T_c$ between the surface and cloud top temperatures. For high clouds, this difference is maximum in the tropics and rapidly decreases with latitude, giving $\partial F/\partial n_H$ its maximum value in the 10 - 20° latitude belt.

The resulting sensitivity parameter is shown in Figure (5) where we consider the two cases of Figure (2) ($\varepsilon_H = 1$ and $\varepsilon_H = 0.3$). Middle and low clouds tend to cool the surface except at high latitudes in winter ($\varphi > 50°$). Drawing a conclusion for high clouds is not as easy. Thick high clouds tend to heat the surface, but most of authors report observations of cirrus emissivities much less than unity. For the rather common value of 0.3, cirrus clouds show almost no effect on the radiative budget in our computations. However, it is to be noticed that we did not change the shortwave radiative properties of cirrus clouds, mostly because of lack of information on the topics. Following Paltridge (1980), our values for high cloud reflectivities correspond to emissivities closer to 0.3 than to unity. Obviously the relationship between infrared emissivity and shortwave albedo for cirrus clouds need to be determined more precisely since they are the major source of uncertainty.

From Figure (5), it is possible to derive (i) the latitudinal profile of $\bar{\delta}$, the vertically averaged value of $\delta$, and (ii) $\bar{\bar{\delta}}$ the globally averaged value of the sensitivity parameter. Table {III} shows the variation of $\bar{\delta}$ with latitude where $\bar{\delta}$ is averaged over the different cloud layers,

$$\bar{\delta}(\varphi) = \sum_i n_i \ \delta_i(\varphi)$$

From this table, it appears that a change in the cloud cover which respects the London's repartition between low, middle and high clouds tends to decrease the surface temperature everywhere except at high latitudes in winter. The globally averaged sensitivity parameter strongly depends on the cirrus cloud emissivity but is negative in any case. As seen from Table IV, which presents values of the sensitivity parameter obtained from previous studies, our results, albeit derived from a different method, are in good agreement with results by Schneider, Ellis, and Ohring and Clapp.

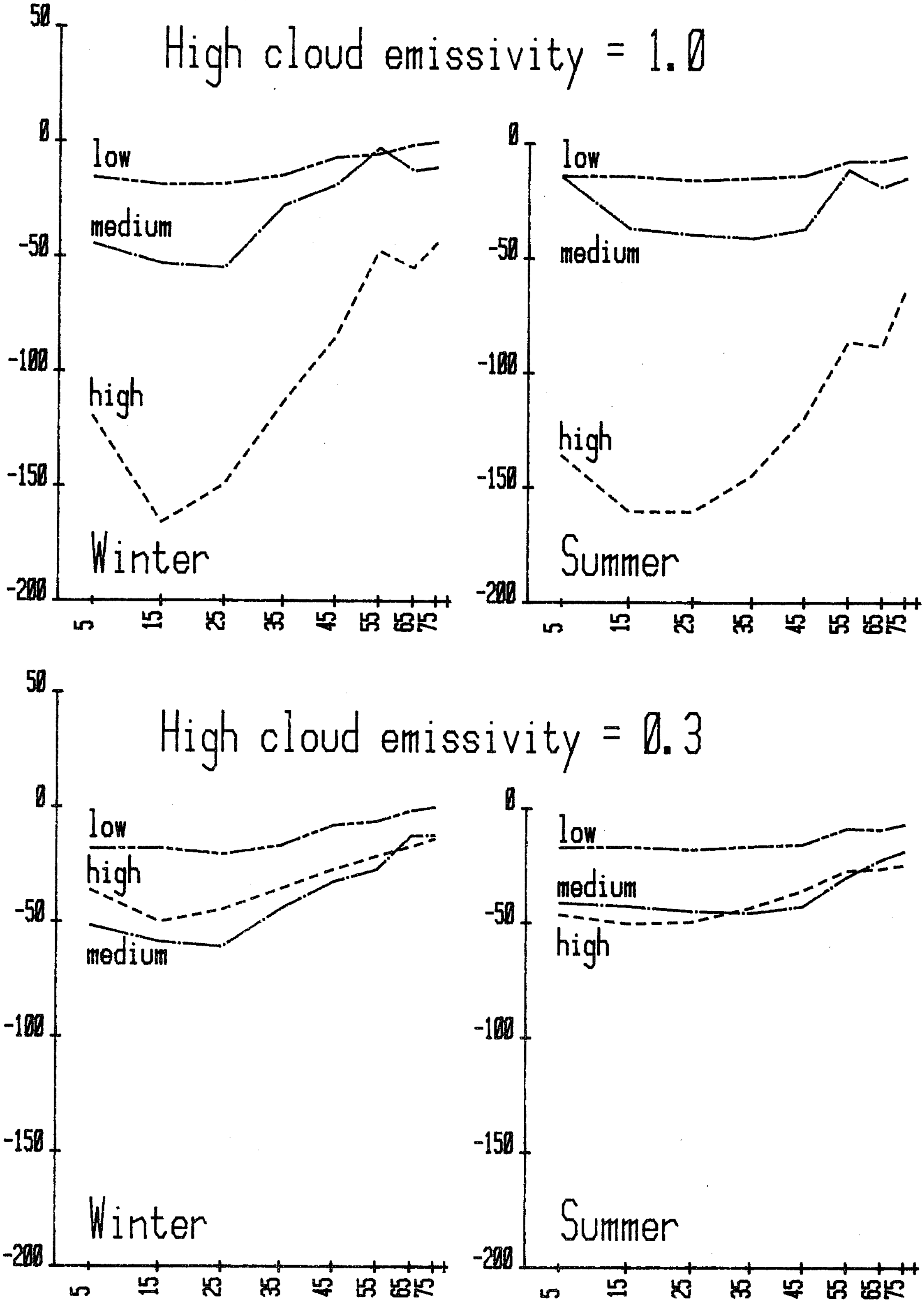


Figure 4 : $\partial F/\partial n_i$ as a function of latitude.

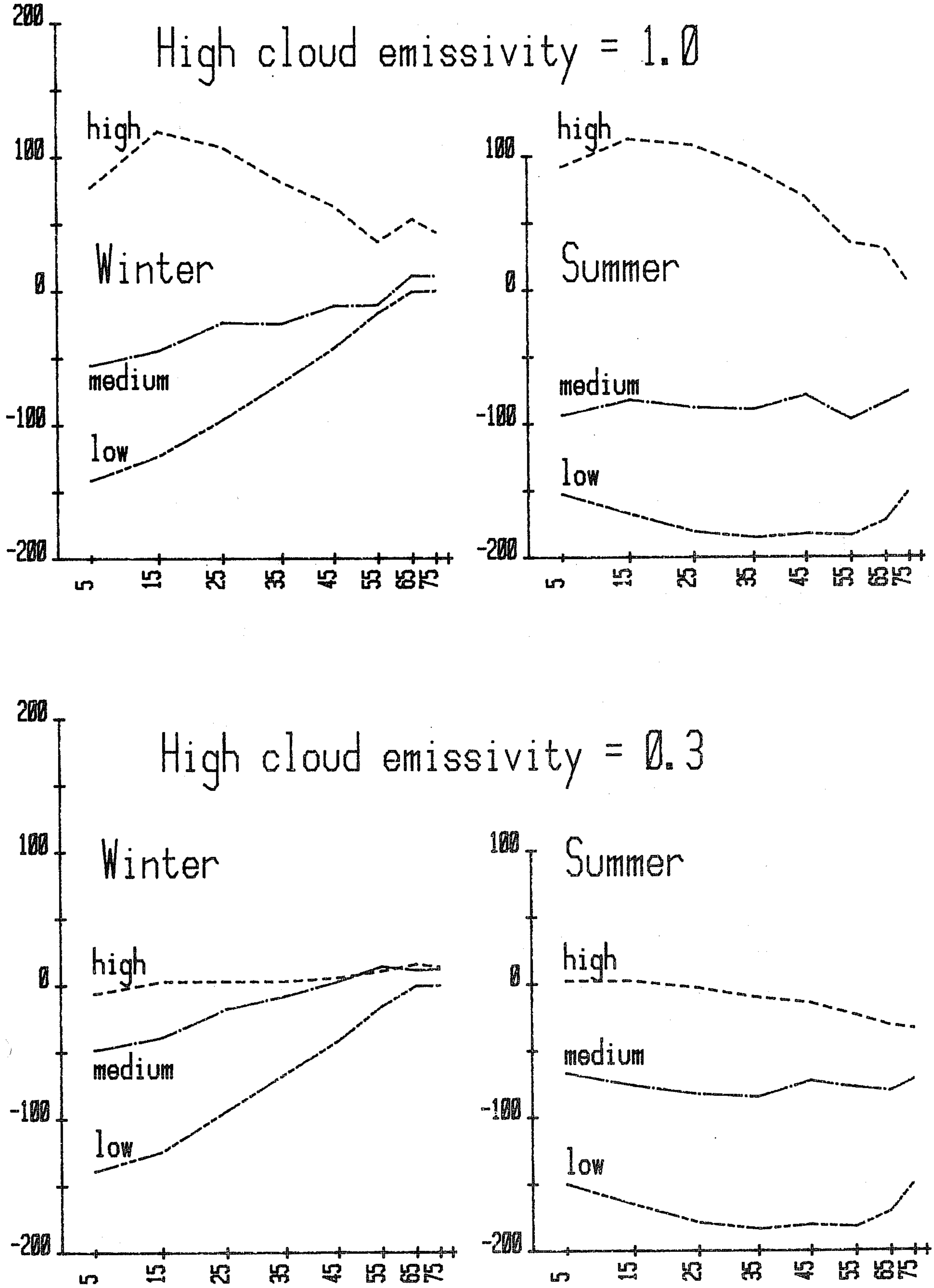


Figure 5 : Sensitivity parameter as a function of latitude.

| Latitude | H.c.e. = 1.0 Winter | H.c.e. = 1.0 Summer | H.c.e. = 0.3 Winter | H.c.e. = 0.3 Summer |
|---|---|---|---|---|
| 0 - 10 | - 35 | - 47 | - 47 | - 57 |
| 10 - 20 | - 19 | - 40 | - 31 | - 56 |
| 20 - 30 | - 14 | - 40 | - 26 | - 53 |
| 30 - 40 | - 15 | - 46 | - 25 | - 57 |
| 40 - 50 | - 9 | - 68 | - 17 | - 80 |
| 50 - 60 | - 6 | - 86 | - 5 | - 95 |
| 60 - 70 | + 10 | - 82 | + 3 | - 97 |
| 70 - 80 | + 6 | - 82 | + 3 | - 92 |
| 0 - 80 | - 14 | - 54 | - 23 | - 66 |
| annual | - 34 | | - 45 | |

Table III : Sensitivity parameter ($W/m^2$)
integrated over altitude in latitude belts,
integrated over altitude and latitude,
annually averaged value
( H.c.e. stands for "high cloud emissivity" )

| | |
|---|---|
| Schneider (1972) | - 58 |
| Cess (1976) | + 2.6 |
| Ellis (1978) | - 35 |
| Ohring, Clapp (1980) | - 65 |

Table IV : Sensitivity parameter ($W/m^2$) according to previous studies.

CONCLUSION

A change in cloud cover must be considered in the context of a global climate variation due to a change in some internal or external parameter. In this case, the cloud cover might change as might the cloud top and base temperatures, and cloud distributions both with latitude and altitude ; and our present results might well be reversed. In this connection, Wetherald and Manabe (1980) have reported on such a climate study that they have performed with a GCM with fully interactive cloudiness and physical processes. They found that, in response to an increase of the solar constant, due to different compensation mechanisms between cloud amount and altitude according to latitude, the changes of cloud cover have a relatively minor effect on the sensitivity of the climate of their model. It is obviously impossible to draw any conclusion of that kind from our present study which only shows that the present steady state of the climate corresponds to the predominance of the albedo effect over the greenhouse effect.

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