Impact of a statistical cloud scheme on the results of the "Arpège-Climat" model

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Abstract

We describe further tests that have been performed using the diagnostic subgrid scale cloud scheme of Ricard and Royer (1993) in the climate version of the Arpège/IFS model("Arpège-Climat"). This cloud parameterization is based on a statistical condensation scheme which is coupled with a level 2 subgrid scale turbulence parameterization, thus allowing to represent in a consistent way the interactions between cloud fraction, liquid water content and turbulence. This scheme has been implemented in the T42 - L30 version of "Arpège-Climat". Experiments have shown that the resulting cloud cover is sensitive to the value of the temperature threshold used to mark the transition between liquid and solid precipitation. We have adjusted this threshold temperature as a kind of tunable parameter, in order to obtain balanced radiative fluxes at the top of the atmosphere in a simulation for October conditions. We have then performed validation experiments for January and July, and compared the results with those of a control simulation with the standard cloud parameterization of the Arpège model, which is based on an empirically specified critical humidity profile. The new statistical scheme, though containing very few tuning parameters, gives cloud profiles that compare very favourably with those given by the old scheme. It leads to a slight increase in the high cloud cover and decrease in the middle and low cloud. Further experiments have been made using the radiation code of Fouquart and Morcrette instead of the Geleyn and Hollingsworth type radiative scheme used in the Arpège model.

1 Introduction

The parameterization of clouds in Atmospheric General Circulation Models (AGCMs) is of considerable scientific importance since it determines in a large part the sensitivity of the model to various external perturbations. Parameterization schemes used in GCM are still very often of a diagnostic type, and based on empirical relationships for computing the cloud cover as a function of the large scale variables such as humidity, vertical velocity or static stability (Slingo and Slingo, 1991). For instance, in the climate model ("Arpège-Climat") used at Meteo-France (Dreveton et al, 1993; Déqué et al, 1994), cloud cover, cloud optical properties, large-scale condensation and precipitation, and small-scale turbulent diffusion, are computed separately in different subroutines. The stratiform cloud cover is evaluated from the relative humidity, by using a critical humidity profile. This critical humidity profile is used as a tunable parameter that can be adjusted in order to obtain realistic cloud cover values and balanced radiative fluxes at the top of the atmosphere. The total water content of clouds is computed as proportional to the vertical gradient of the saturation specific humidity along a moist adiabat. The shallow convection is parameterized through a modified expression of the Richardson number (Geleyn, 1987). Stratiform precipitation occur when the relative humidity exceeds 100% (Kessler, 1969). The main advantage of these diagnostic methods is that they are easy to implement and can be tuned to produce realistic simulations.

More physically based methods, which have recently received considerable attention, are based on the formulation of fully prognostic cloud and liquid water evolution equations (Sundqvist, 1978; Le Treut and Li, 1989; Heise and Roeckner, 1990; Tiedtke, 1993; Ose, 1993; Fowler et al, 1994). An advantage of these methods is that the physical hypotheses are explicitly formulated, and cloud microphysical processes can be more readily introduced, though still in rather crude and simplified forms, because of lack of adequate data on the large scale for the different cloud types. However one drawback of this approach is its much larger complexity and greater computational cost.

A middle way in-between these two extremes is to use some kind of statistical hypothesis for the subgrid-scale distribution of variables related to cloud properties. Such a kind of scheme has been developed by Sommeria and Deardorff (1977) and by Mellor (1977), and used by Bougeault (1982). The cloud fraction and the liquid water content are dependent on the small scale turbulence. A statistical scheme based on these ideas has been developed for use in a General Circulation Model by Ricard and Royer (1993), generalising a similar approach by Smith (1990). The main advantage of this scheme is that it computes simultaneously in a systematic and theoretically consistent way the cloud fraction, the liquid water content, and the vertical diffusion coefficient, while remaining sufficiently simple and computationally affordable for use in long-term simulations. This scheme was originally introduced and tested in the former 20 level - T42 climate model used at Meteo-France (Planton et al, 1990). Since these first experiments the climate model has been replaced by a new version based on the Arpège/IFS code developed jointly by Météo-France and ECMWF (Courtier et al, 1991; Geleyn et al, 1994). The purpose of the present paper is to report recent experiments that have been performed using the statistical cloud scheme in the Arpège code.

2 The statistical cloud scheme

As the statistical cloud scheme has been described in detail in Ricard and Royer (1993), only a summary of the basic physical principles of the scheme is provided here.

2.1 Description of the scheme

2.1.1 The statistical condensation formalism

A method for computing the fractional cloudiness and liquid water content of clouds based on a statistical representation of subgrid-scale fluctuations has been formulated first by Sasamori (1975), using a Gaussian distribution for the vertical displacements of air parcels, and this method has been later adapted for use in a GCM by Hense and Heise (1984). The basic idea of a statistical cloud scheme is that the large-scale variables represent only the first moments (or ensemble mean) of the statistical distribution which describes the variations of the quantity of interest inside a grid-box volume. The existence of such fluctuations can cause air parcels to locally reach saturation while the large-scale water vapour concentration is still below its saturation value, thus producing a partial (or fractional) cloud cover.

A consistent formalism for expressing the subgrid-scale condensation has been developed by Sommeria and Deardorff (1977) and Mellor (1977), based on the assumption of a given joint distribution for the fluctuations of humidity and temperature. The condensation process is best formulated using the variables (introduced by Betts, 1973) which are conserved in the condensation process: the total water content $q_w = q_v + q_l$ (where q_v and q_l are the specific humidity for water vapour and liquid water), and the liquid water potential temperature $\theta_l = \theta - \frac{\theta}{T}\frac{L}{c_p}q_l$ (where T and θ are the temperature and the potential temperature). Assuming that the fluctuations from the grid-scale means $\overline{q_w}$ and $\overline{\theta_l}$ are sufficiently small for linearizing the saturation pressure relationship, the condensation process can be expressed as a function of a single variable $s = \frac{a}{2}(q'_w - \alpha_1\theta'_l)$, where the primes denote departure from the large-scale mean, a and α_1 are functions of the saturation specific humidity. Assuming a given statistical distribution for this variable s, and normalizing this distribution to unit variance by a change of variable $t = s/\sigma_s$ (where σ_s is the standard deviation of s), the fractional cloud cover R and liquid water content $\overline{q_l}$ are given by the following integrals:

$$R = \int_{-Q_1}^{+\infty} \mathbf{G}(t)\,dt = \mathrm{F}_0(Q_1) \tag{1}$$

$$\frac{\overline{q_l}}{2\sigma_s} = \int_{-Q_1}^{+\infty} (Q_1 + t)\ \mathbf{G}(t)\,dt = \mathrm{F}_1(Q_1) \tag{2}$$

The variable

$$Q_1 = \frac{a\Delta\overline{q}}{2\sigma_s} \tag{3}$$

with $\Delta\overline{q} = \overline{q_w} - q_s\left(\overline{T_l}\right)$ is a normalized distance from the saturation specific humidity $q_s\left(\overline{T_l}\right)$, and its opposite plays the part of a normalized condensation threshold (condensation occurs if $t > -Q_1$).

In order to evaluate these integrals one needs to make some hypothesis about the shape of the distribution $G(t)$ of the subgrid fluctuations. Though the original formulation was based on the

Gaussian model, Bougeault (1982) has shown on the basis of both observations and results from a high order turbulence model, that negatively skewed distributions (e.g. from the Gamma family) might be more appropriate.

While the statistical condensation model has found many applications for cloud prediction in small-scale or mesoscale modelling (Yamada and Mellor, 1979; Redelsperger and Sommeria, 1986; Musson-Genon, 1987), it is only recently that its possible application to large-scale models has been considered. A parametrization of cloud cover based on such a statistical scheme, using a triangular distribution, has been developed by Smith (1991) and applied in the UKMO model. The main problem for the application of the statistical formalism is that one needs to provide information about the spread of the distribution, namely the standart deviation σ_s of the humidity fluctuations. Smith (1991) has chosen to close his system empirically at the first order by relating the partial cloud cover to a critical humidity threshold RH_c, as often used in GCMs, and has shown that with the triangular distribution used in his statistical scheme, this is equivalent to specifying the empirical relationship:

$$\sigma_s = (1 - RH_c)/\sqrt{6}\, q_s(T,p)$$

where $q_s(T,p)$ is the saturation specific humidity.

However, using the definition of s, one can relate σ_s to second-order moments:

$$\sigma_s = \frac{a}{2}\left[\overline{q_w'^2} - 2\alpha_1 \overline{q_w'\,\theta_l'} + \alpha_1^2 \overline{\theta_l'^2}\right]^{\frac{1}{2}}$$

In small-scale models these second-order moments are usually computed by using turbulence models closed at different orders. The interactions between cloud cover and subgrid scale turbulence are thus fully taken into account. Ricard and Royer (1993) have proposed to use a similar approach for large-scale models. They have chosen as a starting point to use a second-order closure turbulence model. The main advantage of this model, which is referred as the Level 2 model in the hierarchy of turbulence models devised by Mellor and Yamada (1974), is that it has been extensively studied in the litterature, that it leads to closed form algebraic relationships for the computation of the turbulent quantities , and that it will be straightforward to extend it into the widely used level 2.5 model. Using this level 2 model, with the inclusion of the influence water vapor and liquid water and turbulent fluxes (as in Yamada and Mellor, 1979), it can be shown that σ_s can be computed by the following relationship:

$$\sigma_s = \frac{al}{2}\sqrt{B_2 \widetilde{S_h}}\left|\frac{\partial \overline{q_w}}{\partial z} - \alpha_1 \frac{\partial \overline{\theta_l}}{\partial z}\right| \qquad (4)$$

In this formula l is a neutral length scale representing the typical vertical size of turbulent eddies; $\widetilde{S_h}$ is a stability function, which represents the influence of vertical stratification, and the last term represents a vertical gradient of the large-scale humidity variables. The neutral length scale is specified as a function of height z according to the formulation of Blackadar (1962) in the lower layers: $l = \frac{f(z)}{(1/(kz)+1/\lambda)}$ multiplied by a slowly decreasing function of height $f(z) = 1/(1+(z/h)^2)$ so as to reduce it gradually in the higher atmospheric levels. For the values we have used ($\lambda = 120m$, $H = 1000m$) the mixing length profile reaches a maximum value of about 80 m at 1 km height and decreases smoothly to less than 10 m above 5 km.

With the assumptions of the level 2 model (i.e. steady state of the turbulent kinetic energy) the stability function $\widetilde{S_h}$ is simply an algebraic decreasing function of the gradient Richardson number Ri. It expresses the fact that more stable (unstable) stratifications will reduce (increase) the size of turbulent eddies compared to the neutral case. The relationship obtained for σ_s seems to make sense from a physical point of view since it can be interpreted as the typical fluctuation produced by the vertical gradient of s over the size of a turbulent eddy.

As was shown by Smith (1990), by using the conservative variable the effect of subgrid scale condensation on buoyancy is automatically taken into account, and we have shown (Ricard and Royer, 1993) that the modified Richardson number can be expressed as:

$$Ri = Rih + F_2(Q_1)Ric \tag{5}$$

where Rih and Ric the "usual" and "complementary" Richardson numbers can be easily computed from the vertical gradients of the large-scale variables according to their definitions:

$$Rih = \beta g \frac{\frac{\partial \overline{\theta_l}}{\partial z} + C_{T_0}\frac{\partial \overline{q_w}}{\partial z}}{\left(\frac{\partial \overline{u}}{\partial z}\right)^2 + \left(\frac{\partial \overline{v}}{\partial z}\right)^2} \qquad Ric = \beta g a D(z) \frac{\frac{\partial \overline{q_w}}{\partial z} - \alpha_1 \frac{\partial \overline{\theta_l}}{\partial z}}{\left(\frac{\partial \overline{u}}{\partial z}\right)^2 + \left(\frac{\partial \overline{v}}{\partial z}\right)^2}$$

Since Q_1 according to its definition is a function of σ_s and σ_s is indirectly a function of Q_1 through the dependence of the stability function on the modified Richardson number, we obtain an implicit equation for Q_1, which can be solved by an iterative method.

$$Q_1 \sqrt{\widetilde{S_h}(Q_1)} = \frac{\Delta \overline{q}}{l\sqrt{B_2}\left|\frac{\partial \overline{q_w}}{\partial z} - \alpha_1 \frac{\partial \overline{\theta_l}}{\partial z}\right|}$$

Once Q_1 has been determined by solving this equation by an iterative method the Richardson number can be computed by Equ. (5), then the stability function and the vertical turbulent exchange coefficients, the standard deviation σ_s (Equ.3), and finally the cloud fraction and liquid water content by (1) and (2).

Compared to the parameterization of Smith (1990), which included the influence of subgrid scale condensation upon turbulence by means of its influence upon stability expressed by the Richardson number (5), but not the influence of turbulence upon cloud since σ_s was only a function of saturation specific humidity, our scheme allows a complete two-way interaction of cloud formation and turbulence, in the framework of the balance between production and dissipation of the turbulent kinetic energy (stationarity) assume by the level-2 closure. One noteworthy feature of our statistical scheme is that it has very few empirical (or tuning) parameters. The coefficients of the turbulence model are fixed at the values recommended by Mellor and Yamada (1982). The only parameters that can be adjusted are the vertical profile of the turbulent mixing length l, the residual level of turbulence in very stable stratifications, and the shape of the statistical distribution. In all our experiments we have used an asymetric exponential distribution.

2.2 Details concerning the influence of stability and subgrid-scale turbulence

2.2.1 case of a very stable atmosphere

If the Richardson number is greater than the critical Richardson number, the scheme behaves like an "all or nothing" scheme :

- when the relative humidity is lower than 100%, there is no cloud fraction and no liquid water.

- when the relative humidity is higher than 100%, the cloud fraction is equal to 1. and the water vapour over the saturated value is converted to liquid water.

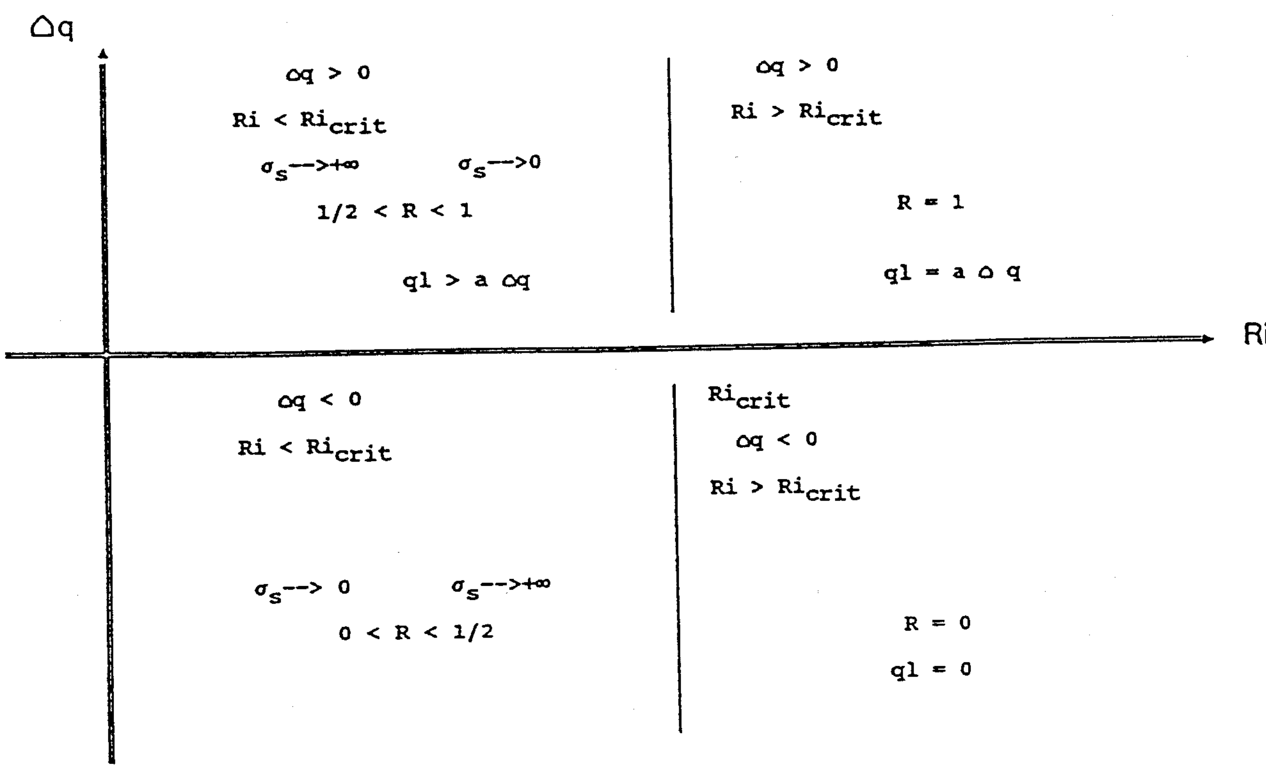

Figure 1: Table showing the different behaviours of the scheme

2.2.2 case of a slightly stable, neutral or slightly unstable atmosphere

The behaviour of the scheme sharply depends on the relative humidity (over or below 100%) and on the subgrid-scale turbulence, measured by σ_s.

- when there is no saturation (Hu $\leq$ 100%).

If there is no subgrid-scale turbulence ($\sigma_s = 0$), there is no cloud fraction and no liquid water. As the subgrid-scale turbulence grows, the cloud fraction and the liquid water content get larger. The cloud fraction is 0.5 when the subgrid- scale turbulence is very large (the value of 0.5 is dependent on the statistical distribution for the s variable ; it is valid if $G(s)$ is Gaussian).

- when there is saturation (Hu $\geq$ 100%).

As in the latter case, if there is no subgrid-scale turbulence ($\sigma_s = 0$), the behaviour of the scheme is the same as in the very stable atmosphere (cloud fraction equals 1. and the liquid water content is the water vapour in excess to the saturation value). As the subgrid-scale turbulence grows, the cloud fraction decreases and the liquid water content increases. The cloud fraction is 0.5 when the subgrid-scale turbulence is very large.

The table 1 page 6 summarizes the different regions in a plane Ri $\times \Delta \overline{q}$, where $\Delta \overline{q}$ is proportional to $(q - q_{sat})$.

2.2.3 Remark about the non-unicity of the solution

The cloud cover, the liquid water content and the turbulent fluxes depend on the large scale environment, i.e. mainly on the vertical shear of the horizontal wind $(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2)$, the liquid water potential temperature vertical gradient $(\frac{\partial \overline{\theta_l}}{\partial z})$, the total water vertical gradient $(\frac{\partial \overline{q_w}}{\partial z})$, $\Delta \overline{q}$ and the master mixing length ℓ.

Ricard and Royer (1993, Figure 3 page 1099) noticed that the large scale environment is not producing a unique solution in every cases : in some regions of the plane $Ri_h \times Ri_c$ (regions D and G on the figure 3 p. 1099), the scheme could produce 2 (physically) acceptable solutions. In the more frequent case (negative complementary Richardson number Ri_c), one solution has virtually no cloud and a very weak subgrid-scale turbulence and the other solution gives a larger amount of clouds associated with a stronger subgrid-scale turbulence. The only way to discriminate between those solutions is the continuity (in time) of the Richardson number.

It is important to note that those regions have to be crossed when the atmosphere goes from very stable conditions to more unstable conditions. This hysteresis effect, due to the presence of liquid water, could therefore have a non- negligible effect and should be analysed with care.

3 Successive implementations of the cloud scheme

3.1 Implementation in the Emeraude model

The statistical cloud scheme had been previously implemented in the T42-20 level version of the Emeraude model (Ricard and Royer, 1993). We have abandoned the Kessler-type large scale precipitation scheme, because the condensation of the whole water vapour over the saturation value is not consistent with the use of a statistical cloud scheme (the cloud cover could not be higher than 50%).

We have chosen to parameterize the large-scale condensation and precipitation with the scheme used by Smith (1990). The precipitation rate is dependent on the liquid water content calculated by the cloud scheme: only a fraction of the liquid water precipitates (supersaturations are authorized). The condensation process depends on the phase of water. The conversion of cloud liquid water is based on the parameterization of Sundqvist (1978):

$$\dot{q}_l = -R\left[c_T\left\{1-\exp\left[-\left(\frac{q_l/R}{c_w}\right)^2\right]\right\}+C_A.PLH\right]$$

where :

$c_T = 10^{-4}s^{-1}$

$c_w = 8.10^{-4}kg.kg^{-1}$

$c_A = 1m^2.kg^{-1}$

PLH is the mass flux of precipitation falling into the layer from above.

The rate of change of cloud ice is obtained with the assumption of a constant fall speed ($v_F = 1$ $m\ s^{-1}$) of ice cloud precipitation :

$$\dot{q}_l = A - Bq_l$$

with :

$A = \frac{PLH}{\rho\Delta z}$

$B = \frac{v_F}{\Delta z}$

$\rho =$ density of the layer

$\Delta z =$ thickness of the layer

The results showed a different distribution of clouds in the vertical. Clouds were simulated at a higher altitude with the statistical cloud scheme. With regard to radiative effects, the temperatures in the middle and higher troposphere were warmer, which was closer to the climatology. The total cloudiness field was found to be more realistic: it diminished the wrong contrast between oceans and continents ; it reduced the surface of regions with a very low amount of clouds (under 10%), giving a better agreement with the climatology.

3.2 Implementation in the first version of the "Arpège-Climat" model

The implementation of the statistical scheme in the "Version 0" (based on Arpege cycle 8) of the "Arpege-Climat" model has been done by Castejon and Gérard (1992). The results were similar to those with the Emeraude model (higher average level of clouds and its radiative effects, less widespread regions with cloud cover $\leq 10\%$). Some problems became very apparent (they did not occur so clearly with the Emeraude model) : apparition of very high clouds at unexpected places (above the gulf of Bengal), reduced rates of precipitation (convective and stratiform ones).

3.3 Implementation in the latest version of the "Arpège-Climat" model

The "Version 1" of the "Arpège-Climat"model (based on ARPEGE cycle 11) is characterized by minor differences with the "Version 0" model. One of those differences is a modification of the radiative properties of clouds ; another one is a re-calibration of the critical humidity profile in order to achieve balanced radiative fluxes at the top of the atmosphere (in annual mean). The critical humidity profile used in the reference simulation is:

$$Hc = 1. - Hu_{coe}\sigma^{n_1}(1-\sigma)^{n_2}\left(1+\sqrt{Hu_{til}}(\sigma-0.5)\right)$$

where Hc is the critical humidity, $\sigma = P/P_s$, $Hu_{coe} = 1.5$, $n_1 = 1$, $n_2 = 1$, $Hu_{til} = 4.5$
(see figure 2)

3.3.1 Description of the experiments

The "Arpège-Climat" climate model is global and spectral. It is similar to the French operational Arpège model (Geleyn et al, 1994), but there are some differences, like the parameterization of ozone (O_3) and the use of additional levels in the stratosphere (Déqué et al, 1994). The model has been run at truncation T42, i.e. the associated Gaussian grid has 64 points in latitude and 128 points in latitude.

In this paper, we compare the results obtained with the statistical cloud scheme to those obtained in a reference simulation with the standard physical parameterizations. To this end, the model has been run for 60 days, keeping only the last 30 ones, so as to simulate a complete month. The "tuning" experiments simulate an intermediate month (October ; starting date : 1/09/1978). The January (resp. July) experiments start on 30/11/1978 (resp. 29/05/1979).

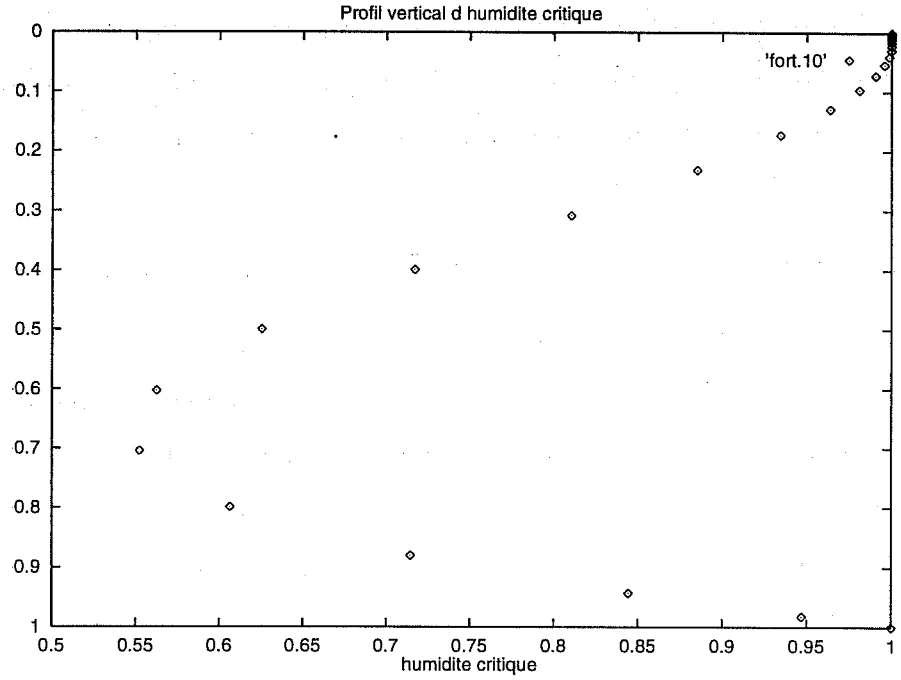

Figure 2: Vertical profile, as a function of pressure, of critical humidity used in the reference simulation

3.3.2 Tuning of the radiative fluxes

Because of these minor changes, the initial simulations with the statistical scheme gave at first unsatisfactory results. The global total cloudiness was too low (51% in October) and the radiative budget at the top of the atmosphere was not well balanced : the incident solar flux was 15 W/m^2 higher than the Outgoing Longwave Radiation (O.L.R.). Such radiative unbalance could lead to a systematic drift of the model climate in the case of long term simulations.

These problems were solved by tuning of the temperature threshold between the liquid and the ice phases. Figure 3 displays a non-linear increase of the cloudiness as the threshold decreases. When the threshold goes beyond - 25°C, the cloudiness increases dramatically : from 56.3% with a -25°C threshold to 63.8% with a -28°C threshold. In the latter case, the simulation becomes rather unrealistic since the O.L.R. becomes larger than the incident solar flux.

All the subsequent experiments with the statistical cloud scheme have been performed with a -25°C threshold, which appears to produce the most realistic results. This value is close to the middle of the range of the probability curve for ice crystal existence of pure crystalline clouds given by Sundqvist (1993), but may appear rather too low according to the observations presented at this workshop by S. Ballard.

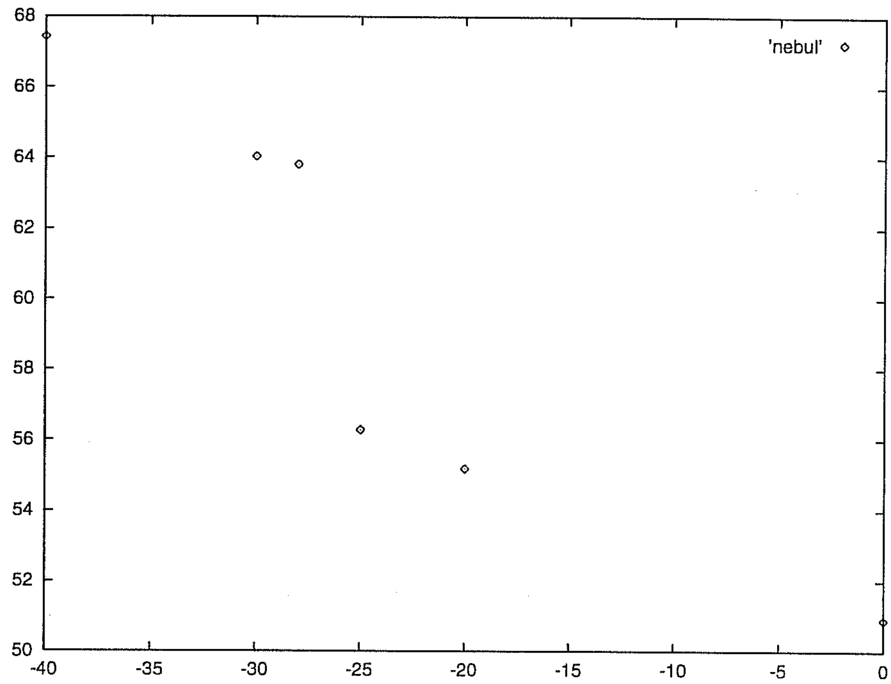

Figure 3: Cloudiness as a function of the temperature threshold between the liquid and the ice phases.

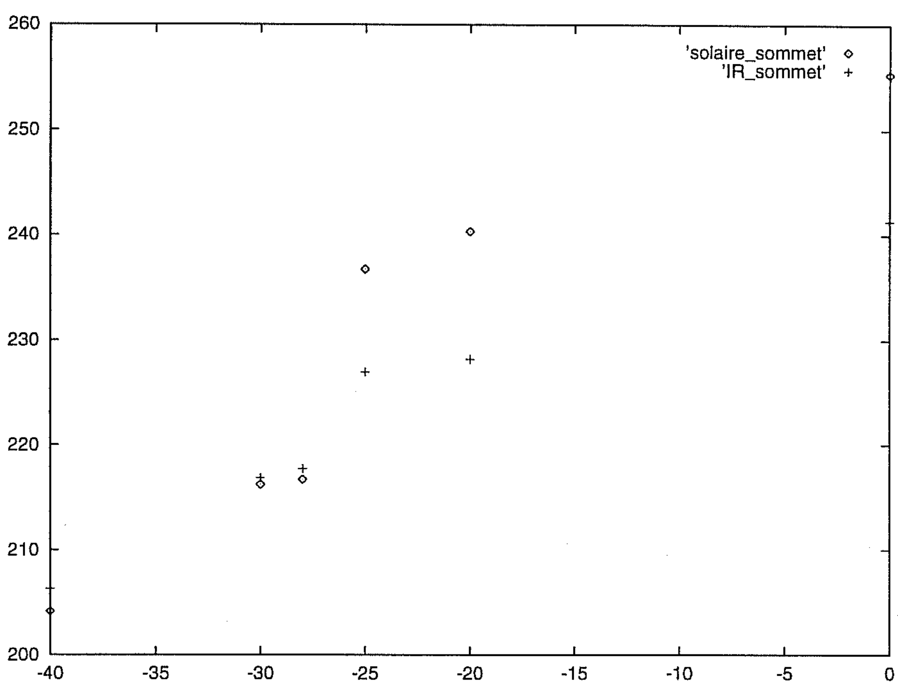

Figure 4: Radiative fluxes at the top of the atmosphere as a function of the temperature threshold between the liquid and the ice water phases. Diamonds : incident solar radiation ; '+' : OLR

4 Results

4.1 Cloudiness

The total cloudiness is slightly higher with the statistical scheme than with the reference (58,7% against 56.4% in January ; 56.4% against 54.4% in July). However, this small increase is not really significant. We have to study in more detail the spatial and vertical structures.

4.1.1 Vertical distribution of clouds

The vertical distribution is different with the statistical scheme, both in January and July ; there is a larger amount of higher clouds, and a smaller amount of middle and lower clouds (except on the 2 lowest levels of the model). The "average" level of clouds is therefore somewhat higher with the new cloud scheme.

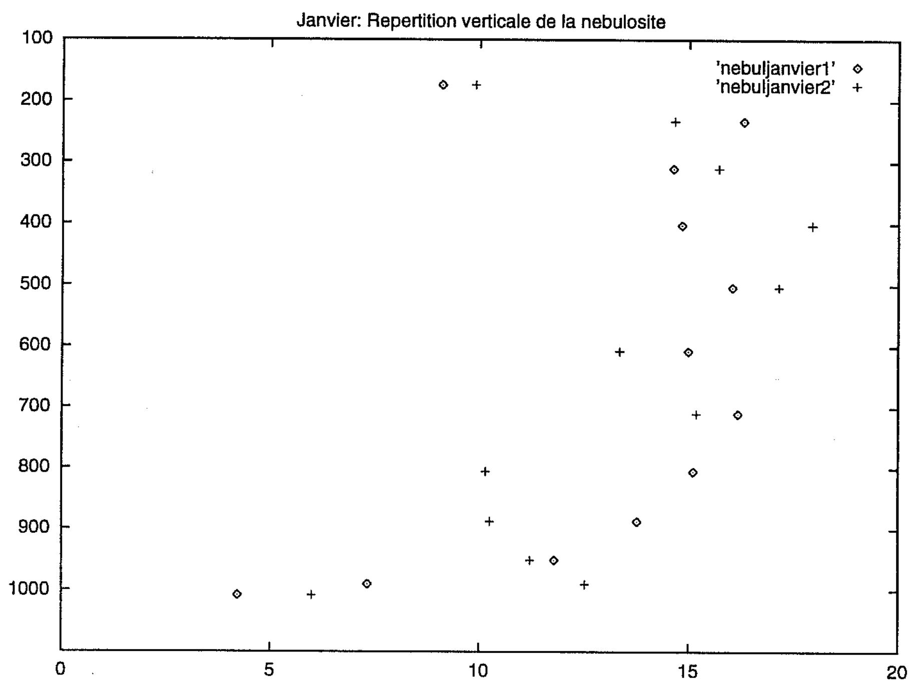

Figure 5: Global mean cloudiness on the model levels (January). Diamonds : Reference ; '+' : cloud scheme

4.1.2 Spatial distribution of clouds

The spatial distribution is not very different with the statistical cloud scheme or with the reference scheme: regions with a large (resp. small) amount of clouds are approximately located at the same places.

The most striking difference can be seen over the sub-tropical anticyclonic regions. Zones with a very low amount of clouds (less than 10%) are much less widespread with the new cloud scheme

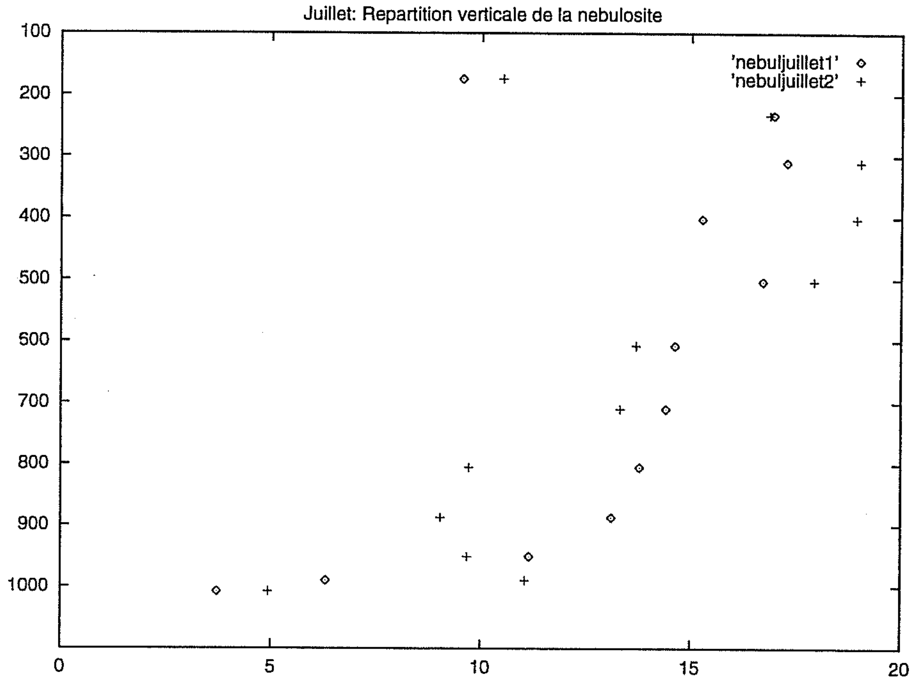

Figure 6: Global mean cloudiness on the model levels (July). Diamonds : Reference ; '+' : cloud scheme

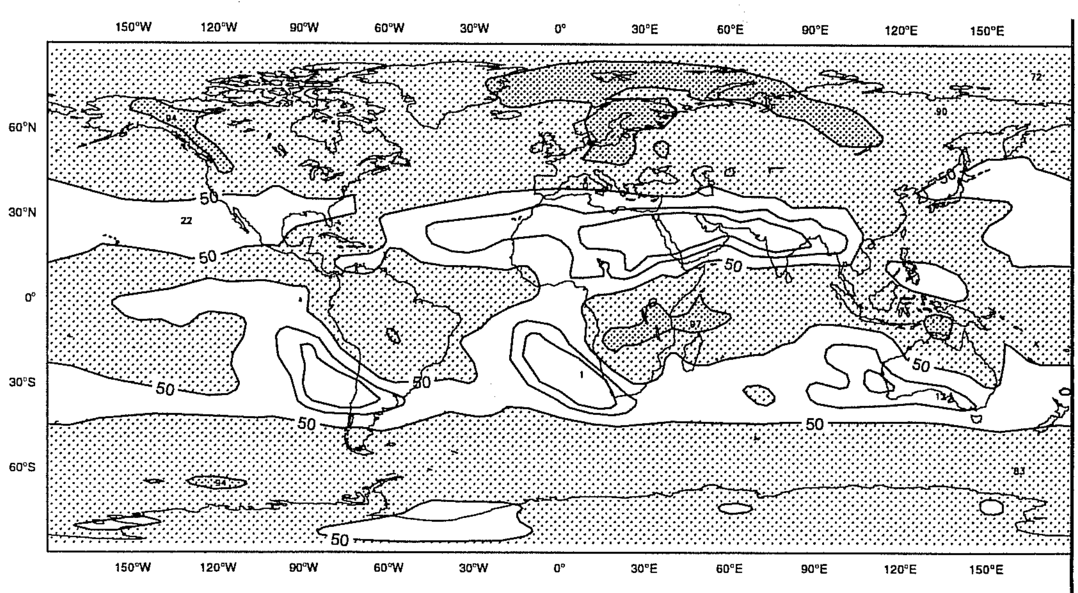

Figure 7: Total cloudiness (isolines 10%, 20%, 50% et 90%) - Reference (January)

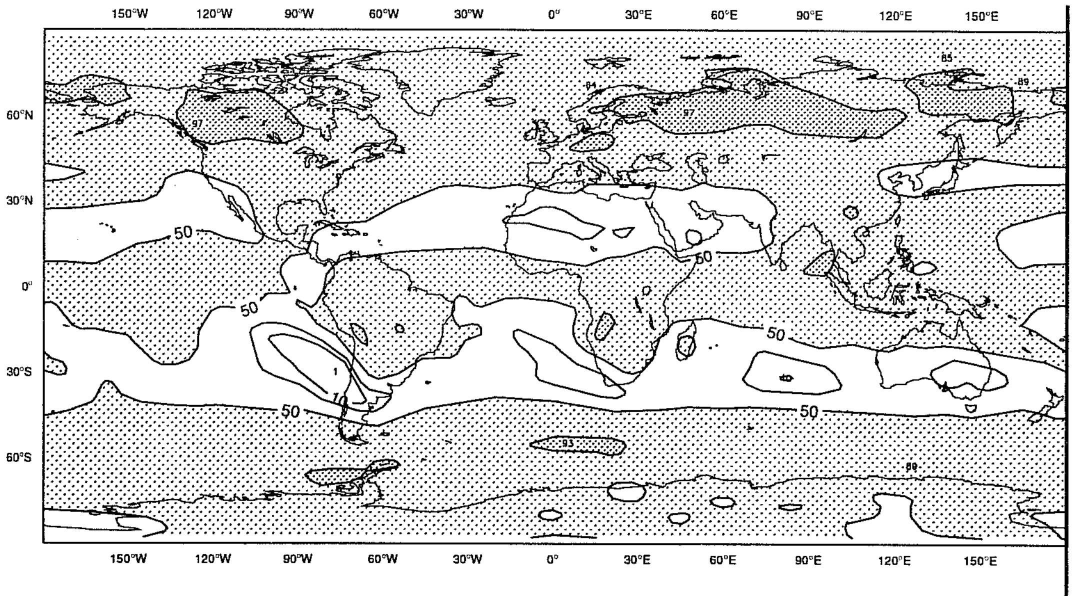

Figure 8: Total cloudiness (isolines 10%, 20%, 50% et 90%) - Statistical cloud scheme (January)

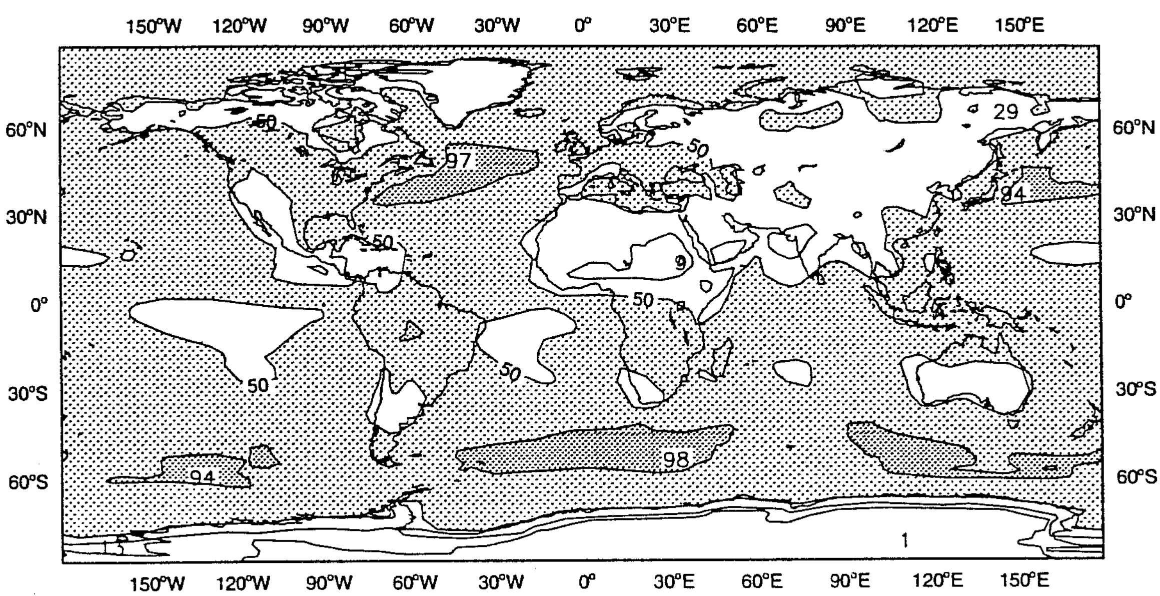

Figure 9: Total cloudiness (isolines 10%, 20%, 50% et 90%) - ISCCP Climatology (DJF)

(especially over the eastern part of the subtropical anticyclones). With the reference scheme such nearly cloud-free areas are seen in January (figure 7) over the Sahara, the Arabic peninsula, Iran (southern part), India (northern part), Chile (off the coast), Namibia and Australia (off the western coast), and in July over the Sahara, the Mediterranean sea (eastern part), the Middle East, the CEI (southern part), South Africa and Madagascar. These zones are drastically reduced with the statistical scheme (figure 8) : only over the Sahara and the Kalahari deserts in July and no such zone in January.

Zones with a large amount of cloudiness are fairly similar. One cannot explain the differences in the vertical distribution of clouds from these total cloudiness charts.

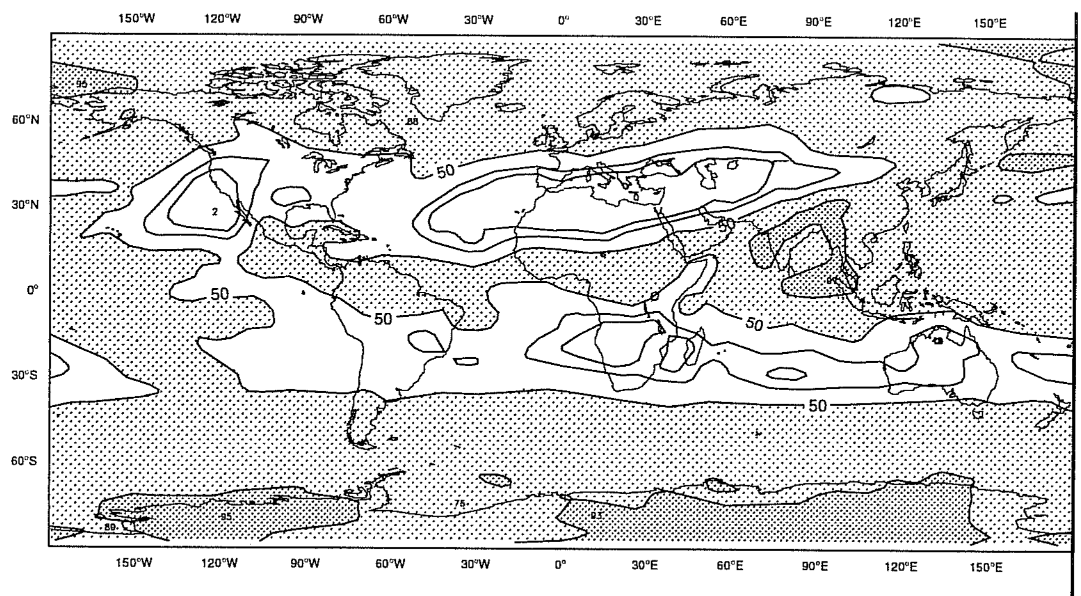

Figure 10: Total cloudiness (isolines 10%, 20%, 50% et 90%) - Reference (July)

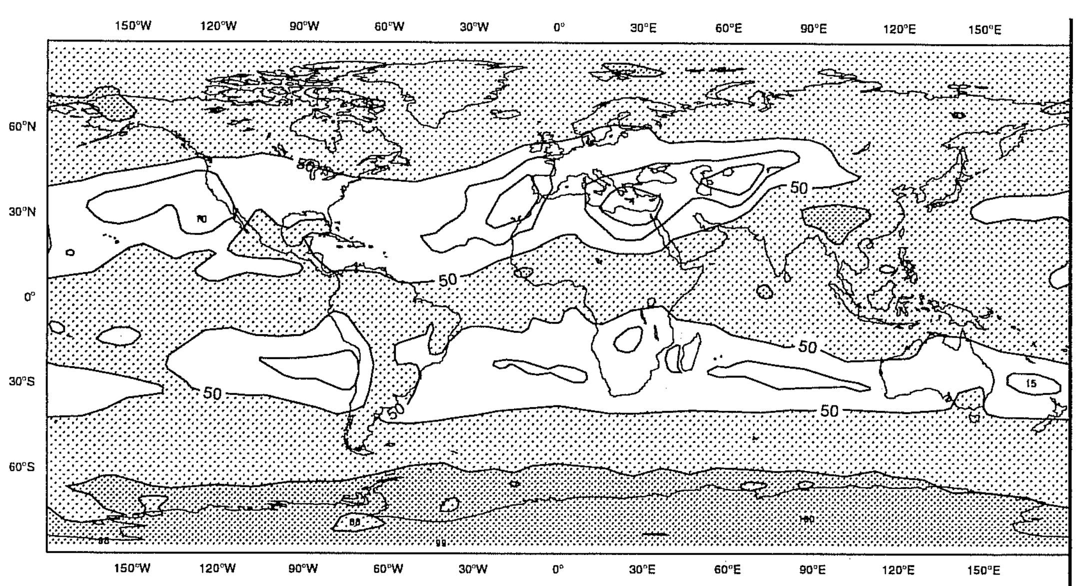

Figure 11: Total cloudiness (isolines 10%, 20%, 50% et 90%) - Statistical cloud scheme (July)

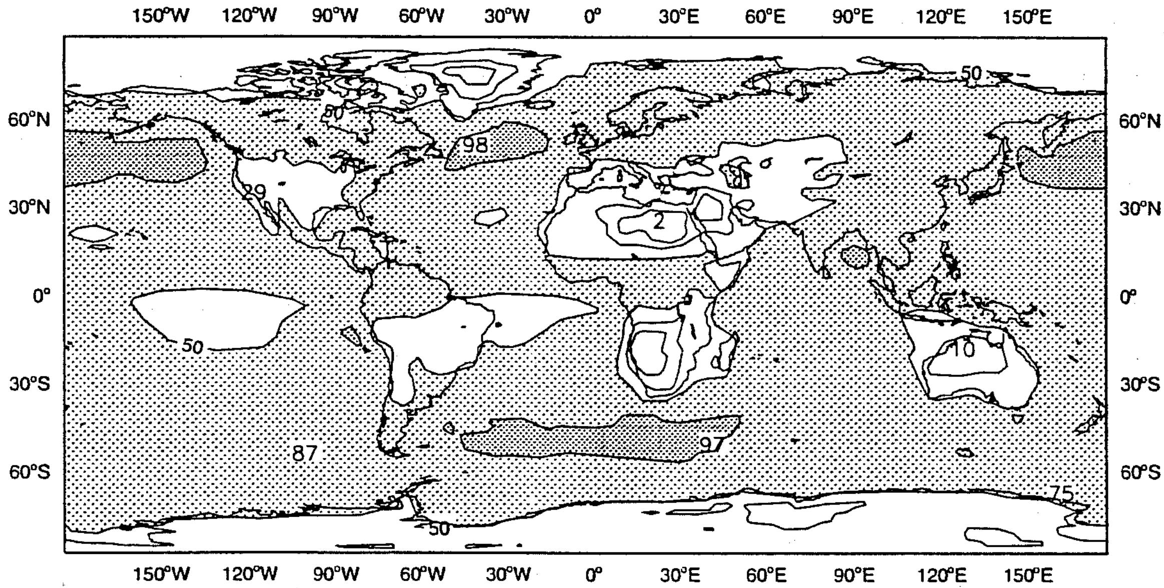

Figure 12: Total cloudiness (isolines 10%, 20%, 50% et 90%) - ISCCP Climatology (JJA)

4.2 Cloud water content

The consistent computation of the cloud fraction and cloud water content, which are linked by the choice of a given statistical distribution is one of the major advantages of the new cloud scheme. In order to see more clearly the differences between the two cloud parameterizations, we have chosen to compute the liquid water content by unit cloud area (total water content divided by fractional cloud cover at each grid-point). The figures 13 and 14 represent the vertical integration of this quantity for the January simulation (Results for July are similar).

The difference between the two cloud parameterizations is rather striking. With the reference scheme, in which the water content of cloud is proportional to a vertical gradient of saturation humidity, the maximum cloud water content is found between the tropics and decreases sharply in the middle and high latitudes. Such a distribution is consistent with the strong influence of temperature on the saturating humidity. With the statistical scheme the maxima of cloud water content are located in the middle latitudes in the regions of moving cyclones. This distribution can be interpreted as due to the broadening of the statistical distribution in regions of increased turbulent exchange. The use of a water content by cloud area prevents a direct comparison of these results with those of satellite observations. Additional diagnostics will be necessary to assess which cloud water distribution is the more realistic. Such a difference in the cloud water content can have an impact on cloud optical properties and radiative fluxes.

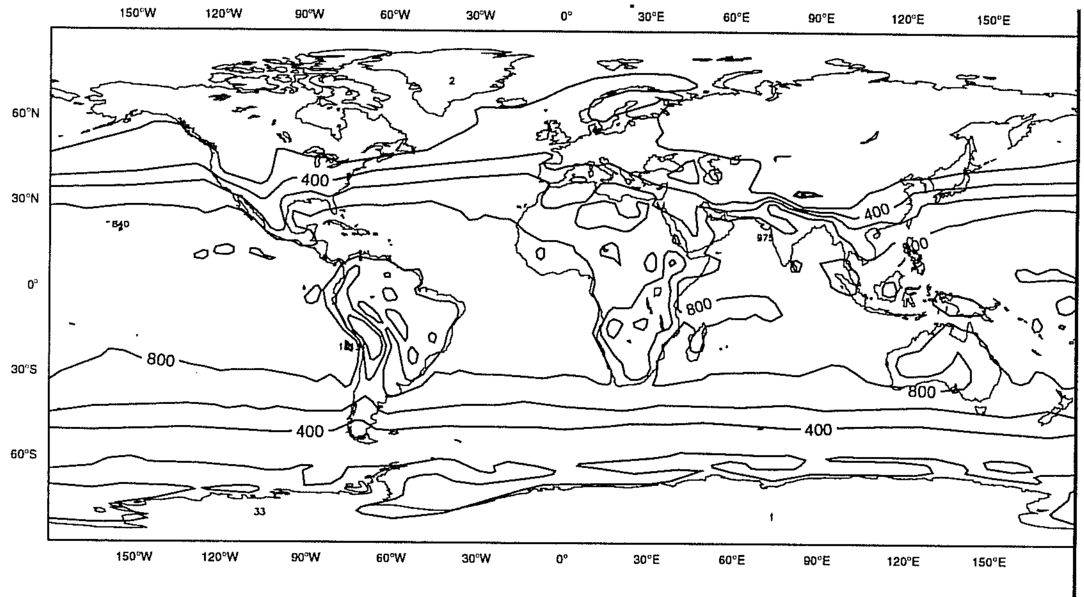

Figure 13: Vertically integrated cloud liquid water content by unit cloud area (isoline spacing 200 E-3 kg/m2) - Reference (January)

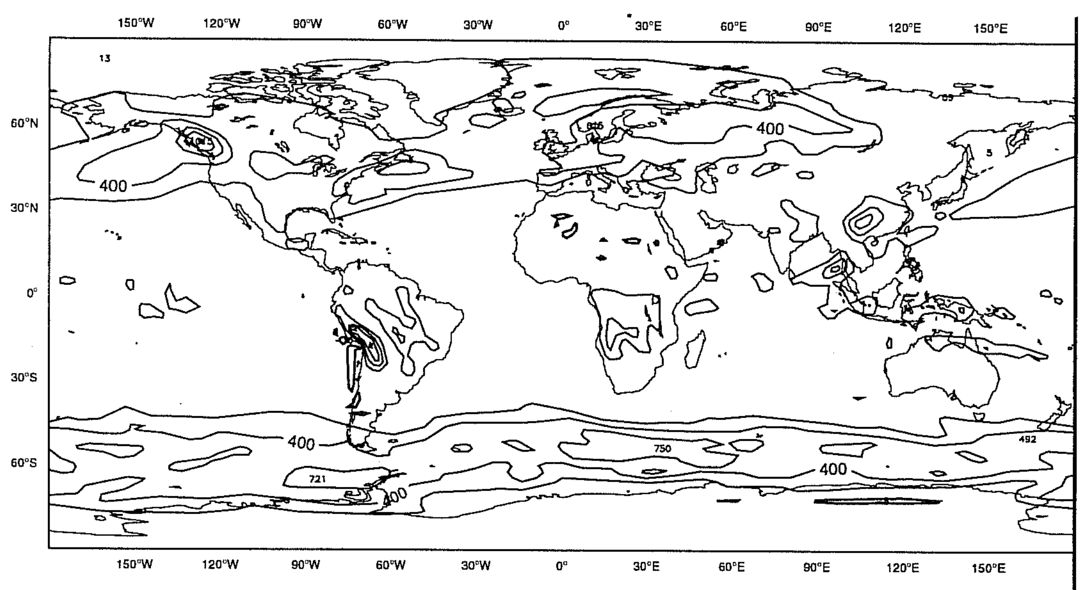

Figure 14: Vertically integrated cloud liquid water content by unit cloud area (isoline spacing 200 E-3 kg/m2) - Statistical cloud scheme (January)

4.3 Radiative fluxes

In the reference, as well as in the experiment with the new cloud scheme, the radiative fluxes at the top of the atmosphere are approximately balanced. There is a small excess of incident solar radiation in January (+8.6 W/m^2 with the reference, +10.55 W/m^2 with the statistical cloud scheme), while there is more O.L.R. in July (-9.02 W/m^2 with the reference, -4.02 W/m^2 with the statistical cloud scheme). Since the "average level of clouds" is lower (and therefore warmer) with the reference, the magnitude of the radiative fluxes is larger with the reference scheme.

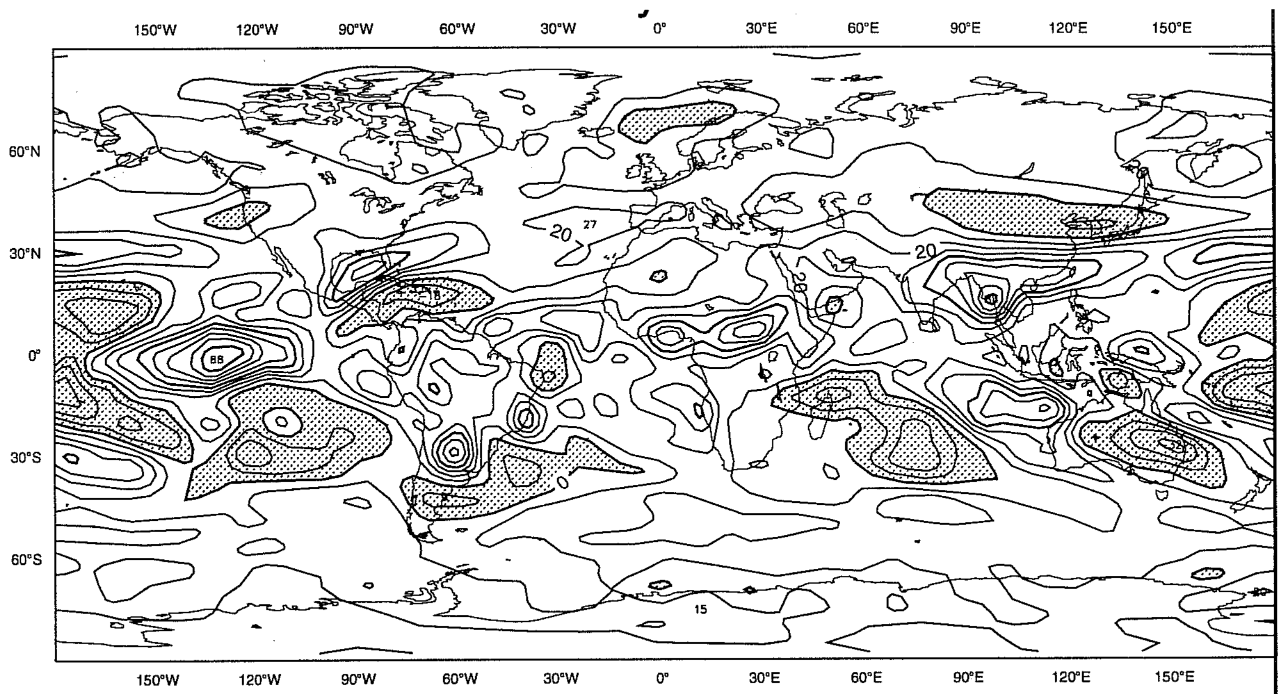

Figure 15: OLR : difference (positive downward) between the statistical scheme and the reference (January)

The O.L.R. charts display (figures 15 and 16) the critical zones where the clouds are significantly higher with the new scheme :

- Equatorial regions. In January, clouds over Africa, Indonesia, Eastern Pacific and South America are higher with the statistical scheme. These zones are deep-convective regions.

In July, the same convective regions produce the same effects.

- over many other regions, the clouds simulated by the new scheme are lower than those simulated with the reference. Over Europe in July, clouds are at a significantly lower altitude with the statistical scheme.

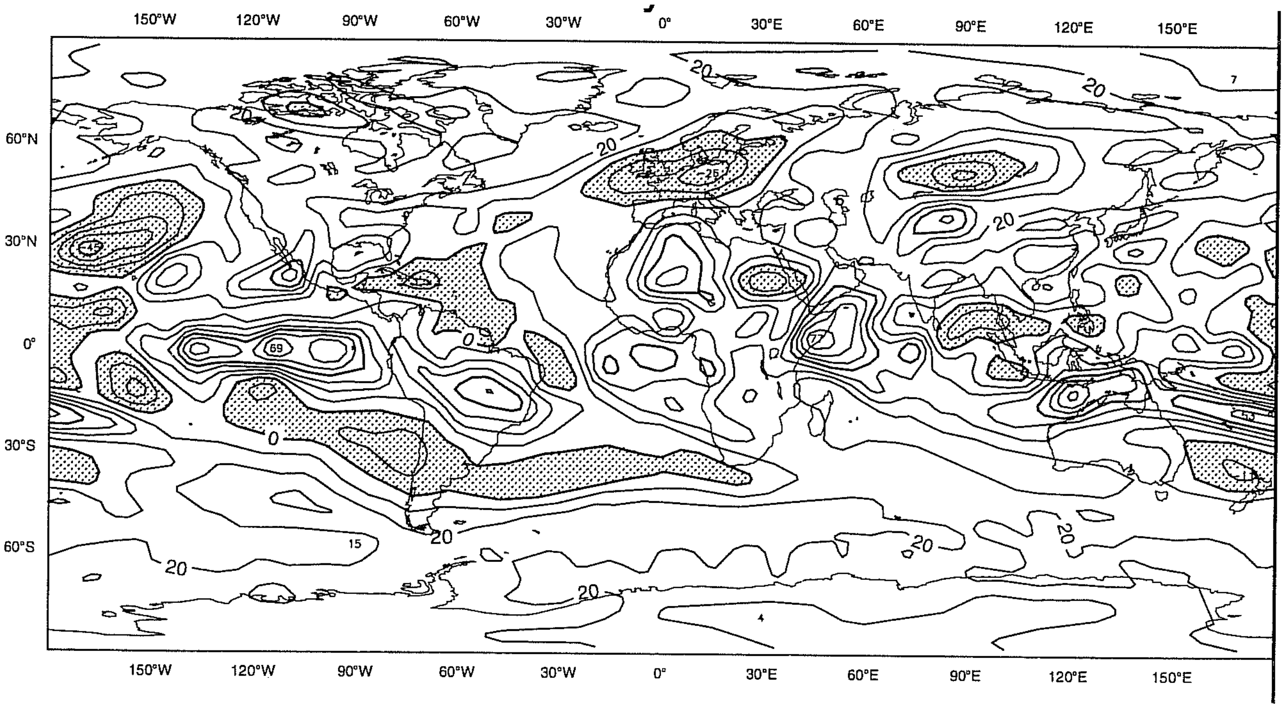

Figure 16: OLR : difference (positive downward) between the statistical scheme and the reference (July)

4.4 Coupling with the FMR radiation scheme

The FMR radiation scheme has been developed by J.J. Morcrette (1991) at the ECMWF. It has been implemented in the "Arpège-Climat" model by Ph. Dandin (Dandin and Morcrette, 1994). Simulations with the FMR and the cloud scheme were performed for an intermediate month (October).

The impact on the precipitation rate is important : the hydrological cycle is more active with the FMR radiation scheme, which is much closer to the reference ; 2.91 mm/day with FMR instead of 2.47 mm/day with Geleyn and Hollingsworth (and 2.93 in the reference). The increase comes from the convective precipitation (2.50 mm/day with FMR, 2.02 mm/day with Geleyn, 2.34 mm/day in the reference) and not from the stratiform precipitation that is almost unchanged. This effect is displayed on figure 17 : the impact on the total precipitation rate is especially visible over South America, Africa and Indonesia.

The impact on the higher cloud cover is important too : the vertical depth of the high clouds is reduced (see figure 18).

These positive impacts may be caused by the lower level of the tropopause obtained with the FMR radiation scheme (see figure 19), and also by the reduction of the cold bias at the tropopause.

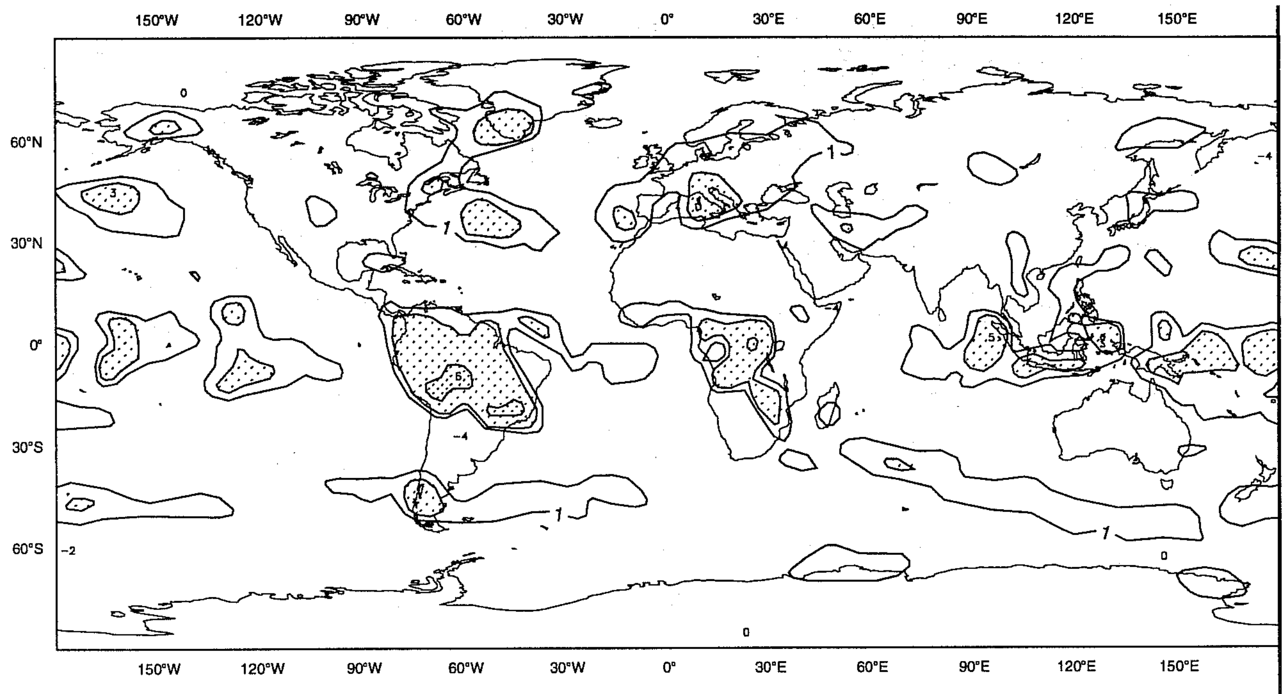

Figure 17: Precipitation : difference between (statistical scheme + FMR) and (statistical scheme + Geleyn radiative scheme

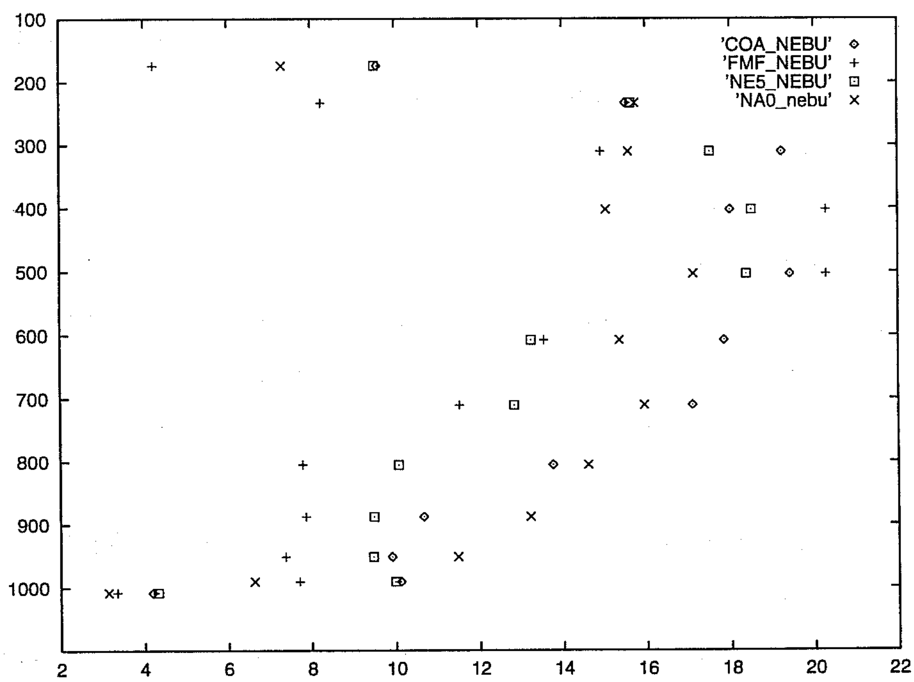

Figure 18: Global mean cloudiness on the model levels
COA : statistical cloud scheme with increased convective entrainment rate in the lower layers (see section 5); FMF : statistical cloud scheme with the FMR radiation scheme ; NE5 : statistical cloud scheme with the Geleyn's radiation scheme; NA0 : reference

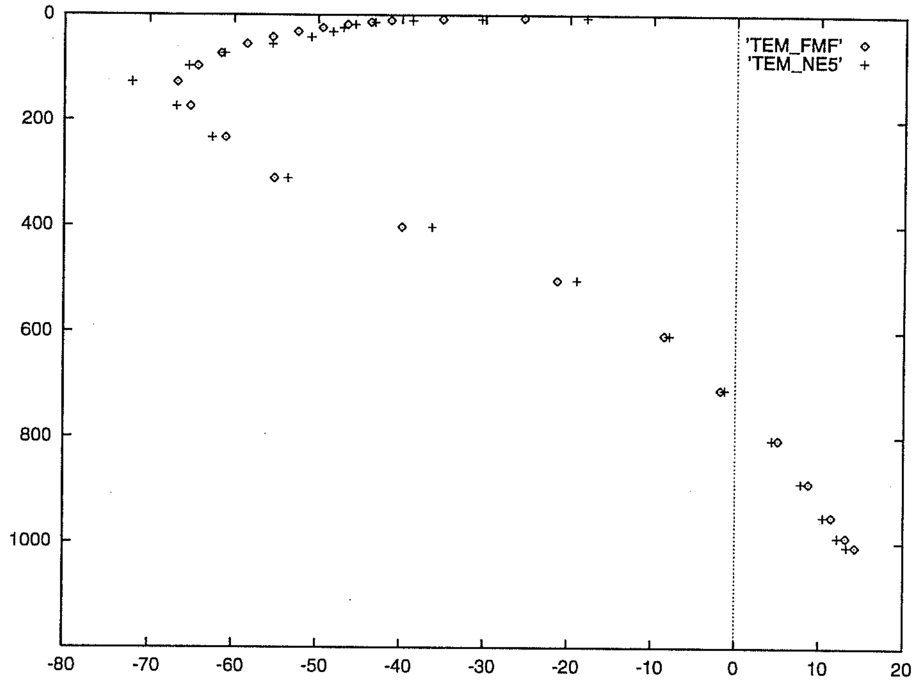

Figure 19: Global mean temperature on the model levels

5 Conclusions and perspectives

The results obtained with the statistical cloud scheme in the Arpège-Climat model have shown that by adjusting a single parameter, the threshold temperature between liquid and solid precipitation, the statistical scheme is able to produce cloud amounts that are in good agreement with the observations. In these experiments this threshold was simply chosen to mark a brutal transition between ice and water. A more gradual transition could be implemented by specifying a simple functional dependence of the ice and water fractions as a function of temperature. The preliminary results obtained with the FMR radiation code seem very encouraging, though there are still some problems that need to be addressed, such as the proper balance of the radiative fluxes at the top of the atmosphere. A recent modification in the convective parameterization, proposed by J.F. Geleyn (Geleyn et al, 1994), by introducing a variation with height of the convective entrainment rate, has also been tested together with the statistical cloud scheme. The resulting modification of convection can also have a strong impact on the vertical distribution of clouds (figure 18) which then become more similar to the reference simulation. The repartition between stratiform/convective precipitation becomes closer to the reference, with a stronger stratiform rate and a weaker convective rate. These experiments indicate clearly that the cloud cover is strongly influenced by radiation and convection, so that the development of a cloud scheme cannot be considered independently from the other physical parameterizations.

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References

Blackadar, A.K. 1962: The vertical distribution of wind and turbulent exchange in a neutral atmosphere. J.Geophys.Res, 67, 3095-3102.

Bougeault, P., 1982: Cloud ensemble relations based on the gamma probability distribution for the higher order models of the Planetary Boundary Layer. J.Atmos.Sci., 39, 2691-2700.

Bougeault, P., 1985: A single parameterization of the large scale effects of deep cumulus convection. Mon. Wea. Rev., 113, 2108-2221.

Castejon, C. et E. Gérard, 1992: Impact de la paramétrisation de la nébulosité dans le modèle Arpège. Note de Travail de l'ENM No 347, Ecole Nationale de la Météorologie, 42 av. Coriolis, 31057 Toulouse, France.

Courtier, Ph., C. Freydier, J.F. Geleyn, F. Rabier and M. Rochas, 1991: The ARPEGE Project at METEO-FRANCE. In: Proc. ECMWF Workshop, Numerical methods in atmospheric modelling, 9-13 Sept 1991, vol 2, p 193-231. ECMWF, Shinfield Park, Reading, U.K.

Dandin, Ph. and J.J. Morcrette, 1994: Effect of a New Radiation Scheme in the Arpège-climat AGCM. To appear In: Research Activities in Atmospheric and Oceanic Modelling, 1994 issue, A. Staniforth (Ed.), CAS/JSC Working group on numerical experimentation, WMO/ICSU/IOC Joint Scientific Committee for the WCRP, Geneva.

Déqué M., C. Dreveton, A. Braun and D. Cariolle, 1994: The ARPEGE/IFS atmosphere model: a contribution to the French community climate modelling. Climate Dyn., 10, 249-266.

Dreveton, C., M. Déqué and J.F. Geleyn, 1993: Interactions of physical parameterizations in the climate version of the ARPEGE/IFS model. Beitr.Phys.Atmos.,66, 283-303.

Fowler, L. D., D. A. Randall and S. A. Rutledge, 1994: Liquid and ice cloud microphysics in the CSU General Circulation Model. Part I: Model description and simulated microphysical processes. (Submitted to J. Climate)

Geleyn, J.F., 1987: Use of a modified Richardson number for parameterizing the effect of shallow convection. J.Met.Soc.Japan, Special NWP Symposium Volume, 141-149.

Geleyn, J.F. and A. Hollingsworth, 1979: An economical method for the computation of the interaction between scattering and line absorption of radiation. Beitr. Phys. Atmosph., 52, 1-16.

Geleyn, J.F., E. Bazile, P. Bougeault et al, 1994: Atmospheric parameterization schemes in Météo-France's ARPEGE N.W.P. model. ECMWF Seminar, Parametrization of Sub-Grid Scale Physical Processes, 5 to 9 September 1994, ECMWF, Shinfield Park, Reading, U.K.

Heise, E. and E. Roeckner, 1990: The performance of physically based cloud schemes in General Circulation Models. Beitr.Phys.Atmosph., 93, 1-14.

Hense, A. and E. Heise, 1984: A sensitivity study of cloud parameterizations in General Circulation Models. Beitr.Phys.Atmos., 57, 240-258.

Kessler, E., 1969: On the distribution and continuity of water vapor substance in atmospheric circulations. Meteorol.Monogr., 10, No 32, Amer.Meteor.Soc.

Le Treut, H. and Z. X. Li, 1988: Using Meteosat data to validate a prognostic cloud water scheme. Atmos.Res., 21, 273-292.

Louis, J.F., 1979: A parametric model of vertical eddy fluxes in the atmosphere. Bound.Layer Meteor., 17, 187-202.

Mellor, G., 1977: The Gaussian cloud model relation. J.Atmos.Sci., 34, 1483-1484.

Mellor, G. and T. Yamada, 1974: A hierarchy of turbulence closure models for planetary boundary layers. J.Atmos.Sci., 31, 1791-1806.

Mellor, G. and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. Review of Geophysics and Space Physics, 20, 851- 875.

Morcrette, J.J., 1991: Radiation and cloud radiative properties in the European Centre for Medium Weather Forecasting system. J. Geophys. Res., 96, 9121-9132.

Musson-Genon, L., 1987: Numerical simulation of a fog event with a one-dimensional boundary layer model. Mon.Weath.Rev., 115, 592-607.

Ose, T., 1993: An examination of the effects of explicit cloud water in the UCLA GCM. J.Meteor.Soc.Japan, 71, 93-109.

Planton, S., M. Déqué and C. Bellevaux, 1991: Validation of an annual cycle simulation with a T42-L20 GCM. Climate Dyn., 5, 189-200.

Redelsperger, J.L. and G. Sommeria, 1986: Three-dimensional simulation of a convective storm: sensitivity studies on subgrid parameterization and spatial resolution. J.Atmos.Sci, 43, 2619-2635.

Ricard, J.L., 1992: Etude d'un schéma statistique de génération de nuages et son introduction dans un modèle de circulation générale. Thèse de doctorat de l'Université Paul Sabatier Toulouse III.

Ricard, J.L. and J.F. Royer, 1993: A statistical cloud scheme for use in an AGCM. Ann.Geophysicae, 11, 1095-1115.

Ricard, J.L., 1994: Impact du schéma "Ricard" de nuages sur les résultats du Modèle Communautaire. In: Atelier de Modélisation de l'Atmosphère, Toulouse, 29 Nov - 1 Dec 1994.

Sasamori, T., 1975: A statistical model for stationary atmospheric cloudiness, liquid water content, and rate of precipitation. Mon.Weath.Rev., 103, 1037-1049.

Slingo, A. and J. M. Slingo, 1991: Response of the NCAR CCM to improvements in the representation of clouds. J.Geophys.Res., 96, 15341-15357.

Smith, R.N.B., 1990: A scheme for predicting layer clouds and their water content in a G.C.M.. Quart.J.R.Met.Soc., 104, 677-690.

Sommeria, G. and J.W. Deardorff, 1977: Subgrid-scale condensation in models of non-precipitating clouds. J.Atmos.Sci., 34, 344-355.

Sundqvist, H., 1978: A parameterization scheme for non-convective condensation including prediction of cloud water content. Quart.J.R.Met.Soc., 104, 677-690.

Sundqvist, H., 1993: Inclusion of the ice phase of hydrometeors in cloud parameterization for mesoscale and largescale models. Beitr.Phys.Atmosph., 66, 137-147.

Taylor, K. E. and S. J. Ghan, 1992: An analysis of cloud liquid water feedback and global climate sensitivity in a General Circulation Model. J. Climate, 5, 907-919.

Tiedtke, M., 1984: The effects of penetrative cumulus convection on the large scale flow in a general circulation model. Beitr.Phys.Atmosph., 57, 216-239.

Tiedtke, M., 1993: Representation of clouds in large-scale models. Mon.Weath.Rev., 121, 3040-3061.