AN INTEGRATED APPROACH TO THE REPRESENTATION OF CLOUD PROCESSES

R.N.B. Smith

Meteorological Office, London Road,Bracknell, Berkshire, RG12 2SZ, England

Summary: A scheme to represent cloud processes in numerical models is presented. Its main feature is the use of thermodynamic and water content variables conserved during cloud water phase changes. The scheme predicts the cloud amount and liquid water content and includes a representation of turbulent mixing in cloudy regions. The proposed approach attempts to treat clouds and their prediction as an integral part of the model's physics.

1. INTRODUCTION

The prediction of non-convective cloud in Atmospheric General Circulation Models (AGCMs) and large-scale Numerical Weather Prediction (NWP) models is usually a simple diagnostic process. This is the case at present in the U.K. Meteorological Office models (Slingo and Wilderspin (1985)). The cloud amount is calculated for input to the radiation scheme from quantities such as relative humidity. However, apart from its role in calculating the radiative fluxes, this cloud has no other direct effect; in particular its thermodynamic consequences are ignored and the turbulent mixing scheme takes no account of the cloud.

It is now well established that cloud, especially in the boundary layer, owes its formation, dissipation and distribution to a combination of physical processes, often in a subtle balance. These processes may include:

(i) large (resolved) scale dynamical motions,

(ii) unresolved mesoscale effects,

(iii) small scale turbulent mixing,

(iv) phase changes of water and associated latent heating,

(v) the formation of precipitation,

(vi) radiative heating or cooling.

We may expect that a model which represents these processes in a consistent manner and as an integral part of its design could predict cloud in a more realistic way. The accuracy of the prediction would obviously depend on the way the physical processes are represented and interact in the model. Like all aspects of a numerical model, there is little that can be claimed for a scheme a priori apart from theoretical consistency. The scheme presented here has yet to be tested in a systematic way and may require further development for satisfactory results.

The remaining sections of this paper provide a theoretical outline of a scheme to include clouds and moist processes in a consistent way in the model's physics.

2. CLOUD WATER CONTENT AND "CONSERVATIVE" VARIABLES

A basic feature of the type of scheme to be described is the use of a variable representing cloud liquid water. Strictly, we should introduce and distinguish the specific cloud liquid water content (q_ℓ) and the cloud frozen water content (q_f). We can then define thermodynamic and water content variables which are conserved during all cloud water phase changes (somewhat imprecisely called "conservative" variables):

the "ice-liquid water temperature"

$$T_\ell \equiv T - \frac{L}{C_p} q_\ell - \left(\frac{L + L_f}{C_p}\right) q_f \qquad (2.1)$$

and the total water content

$$q_t \equiv q + q_\ell + q_f \qquad (2.2)$$

In these expressions T is the temperature and q is the specific humidity. L is the latent heat of condensation and L_f is the latent heat of fusion ($(L+L_f)$ is then the latent heat of deposition.)

T_ℓ and q_t satisfy the prognostic equations:

$$\frac{DT_\ell}{Dt} - \frac{\kappa T\omega}{P} = \frac{Q}{C_p} - \frac{L}{C_p}\frac{d_p q_\ell}{dt} - \frac{(L+L_f)}{C_p}\frac{d_p q_f}{dt} + \frac{L}{C_p}\frac{d_s r_\ell}{dt} + \frac{(L+L_f)}{C_p}\frac{d_s r_f}{dt} \qquad (2.3)$$

$$\frac{Dq_t}{Dt} = \frac{d_p q_\ell}{dt} + \frac{d_p q_f}{dt} - \frac{d_s r_\ell}{dt} - \frac{d_s r_f}{dt} \qquad (2.4)$$

where:

$r_{\ell(f)}$ is the specific liquid (frozen) water content in the form of precipitation;

D/Dt is the time derivative following a fluid element;

Q is the radiative heating rate;

$d_p q_{\ell(f)}$ represents the change of liquid (or frozen) cloud water due to the formation of precipitation (≤ 0);

$d_s r_{\ell(f)}$ represents the change of specific liquid (or frozen) precipitation content due to changes of state of the precipitating water (conversion of precipitation to vapour (≤ 0); or freezing or melting of precipitation).

Note the absence of terms representing cloud water phase changes, as expected. Equations (2.3) and (2.4) are valid for all scales of motion and so no terms have been included to represent the averaged effects of unresolved motion. The effects of turbulent motion on the model's resolved scales will be discussed in the next section. For simplicity the presence of frozen cloud water and frozen precipitation will be ignored in the following sections.

3. TURBULENT MIXING IN CLOUDY REGIONS

If the model variables now represent the resolved scale mean of quantities we have to include the effects of unresolved processes. To represent turbulent transports, the turbulent flux divergence terms

$$-\frac{1}{\rho}\frac{\partial}{\partial z}(\rho\overline{w'T_\ell'}) \qquad \text{and} \qquad -\frac{1}{\rho}\frac{\partial}{\partial z}(\rho\overline{w'q_t'})$$

have to be added to the l.h.s. of (2.3) and (2.4) respectively and similar terms to the momentum equations. We shall assume that radiative and precipitation processes do not have to be directly accounted for in the calculation of turbulent effects. However, by the use of T_ℓ and q_t in place of T and q we are implicitly including direct effects of cloud water phase changes on the turbulence. It is this feature which distinguishes the proposed scheme from the dry turbulence schemes usually employed.

There are several approaches to parametrizing the turbulent fluxes in terms of variables predicted by the model. The mixed layer type of scheme which has been used in the UCLA GCM (Suarez et al. (1983)) has many merits but simple profiles for the model variables and the fluxes have to be imposed in the boundary layer. This scheme has not yet been generalised to deal with partial cloudiness. Higher order turbulence closure schemes (Mellor and Yamada (1982)) show considerable skill in high resolution one-dimensional model simulations. However, they are computationally expensive and their benefits become more marginal when used with the relatively coarse vertical resolutions of current large-scale models.

The U.K. Meteorological Office AGCM and operational NWP models, together with the ECMWF operational model, use relatively simple first order turbulence closure schemes. This type of scheme parametrizes turbulent fluxes in terms of local gradients of the basic model variables and stability dependent turbulent transport coefficients (K_h and K_m for heat/moisture and momentum respectively). It is the modification of this type of scheme to take account of clouds which will now be described. The use of conservative variables for this purpose has been developed by Yamada and Mellor (1979) and Mellor and Yamada (1982).

The basic structure of the scheme remains unchanged from the existing dry formulation. The vertical turbulent fluxes are parametrized as:

$$\overline{w'T_\ell'} = -K_h \left(\frac{\partial T_\ell}{\partial z} + \frac{g}{C_p}\right)$$

$$\overline{w'q_t'} = -K_h \frac{\partial q_t}{\partial z} \qquad (3.1)$$

$$\overline{w'\underline{v}'} = -K_m \frac{\partial \underline{v}}{\partial z}$$

where the turbulent transport coefficients are given by

$$K_{h,m} = \ell^2 f_{h,m}(Ri_m)\left|\frac{\partial \underline{v}}{\partial z}\right| \qquad (3.2)$$

The functions of stability f_h and f_m are empirically specified and are identical to those used in the dry scheme. ℓ is a neutral mixing length given by the Blackadar formula. The definition of the Richardson number is modified to take account of any cloud which may present:

$$Ri_m = \left[\tilde{\beta}_T \left(\frac{\partial T_\ell}{\partial z} + \frac{g}{C_p}\right) + \tilde{\beta}_w \frac{\partial q_t}{\partial z}\right] \Bigg/ \left|\frac{\partial \underline{v}}{\partial z}\right|^2 \qquad (3.3)$$

The modified buoyancy parameters are given by

$$\tilde{\beta}_T = \beta_T - a_\ell \, \alpha_\ell \, C\beta_L$$

$$\tilde{\beta}_w = \beta_w + a_\ell \, C\beta_L \qquad (3.4)$$

where C is the cloud fraction and

$$\beta_T = \frac{g}{T} \quad , \quad \beta_w = \frac{g\delta}{(1+\delta q-q_\ell)} \quad , \quad \beta_L = \frac{L}{C_p} \beta_T - \frac{\beta_w}{(1-\epsilon)} \qquad (3.5)$$

$$\epsilon = 0.622 \quad , \quad \delta = \frac{1}{\epsilon} - 1 = 0.608$$

$$\alpha_\ell = \left.\frac{\partial q_{sat}}{\partial T}\right|_{T=T_\ell} = \frac{\epsilon L q_{sat}(T_\ell,p)}{RT_\ell^2} \qquad \text{(Clausius–Clapeyron)} \qquad (3.6)$$

$$a_\ell = 1/(1 + L\alpha_\ell/C_p) \qquad (3.7)$$

The definition of stability given by (3.3) rests on the hypothesis that the vertical flux of liquid water is a linear combination of fluxes of T_ℓ and q_t:

$$\overline{w'q_\ell'} = a_\ell\, C(\overline{w'q_t'} - \alpha_\ell\, \overline{w'T_\ell'}) \qquad (3.8)$$

In the special cases of no cloud (C = 0) and complete cloud cover (C = 1) the prescription (3.8) reduces to the usual results.

It should be noted that stability now depends on gradients of T_ℓ and q_t and involves latent heating effects. Partial cloudiness is also treated. In the special case of a cloud-free region C = 0, $q_\ell = 0$, $T_\ell = T$, $q_t = q$, $\tilde{\beta}_T = \beta_T$, $\tilde{\beta}_w = \beta_w$ and $Ri_m = Ri$ where Ri is the usual Richardson number. The scheme thus reduces to the currently used dry version in the absence of clouds. In the other extreme of a completely cloudy region (C = 1) it can be shown from (3.3)-(3.7) that $Ri_m < 0$ is equivalent to the inequality:

$$b\frac{\partial h}{\partial z} < C_pT \frac{\partial q_t}{\partial z} \qquad (3.9)$$

where $h = C_pT + Lq + gz$ is the moist static energy and b is a slowly varying function of temperature given by

$$b = [(1+\delta q-q_\ell) + (1+\delta)\alpha_\ell T]/(1+L\alpha_\ell/C_p)$$

When the vertical gradients are taken across the cloud top, inequality (3.9) is the criterion for Cloud Top Entrainment Instability (CTEI) derived by Randall (1980) and Deardorff (1980). With the use of the modified Richardson number the proposed scheme therefore automatically diagnoses CTEI and will enhance the turbulent mixing of boundary layer air with the free atmosphere when the instability occurs. There is evidence to suggest that the mixing of cloudy boundary layer air with the overlying air due to CTEI is an important process to represent in models in order to obtain realistic low cloud amounts (Randall (1985) and Slingo and Wilderspin (1985)).

The turbulent mixing scheme just described should, subject to the constraints of vertical resolution, represent mixing in cloud-free, partially and completely cloudy regions. At the very least it should improve on the use of the existing dry schemes. A lapse rate for temperature less than the dry adiabatic value, as will occur if clouds are present, will be interpreted as static stability by a dry mixing scheme.

4. THE CALCULATION OF CLOUD AMOUNT AND LIQUID WATER CONTENT

The calculation of the cloud amount C and the cloud liquid water content q_ℓ is based on the cloud ensemble concepts developed by Sommeria and Deardorff (1977) and Mellor (1977). A statistical distribution of the thermodynamic and moisture variables is assumed to exist. In the ensemble mean, the amount of cloud and its liquid water content are clearly going to be determined by the mean of the difference of the specific total water content and the saturation specific humidity and also the statistics of the fluctuations of the difference. The mean and fluctuating parts of this quantity are:

$$Q_L = a_\ell(q_t - q_{sat}(T_\ell, p)) \tag{4.1}$$

and $$s = a_\ell(q_t' - \alpha_\ell T_\ell') \tag{4.2}$$

(a_ℓ and α_ℓ are defined by (3.6) and (3.7))

The grid-box mean (strictly the ensemble mean) of s is zero since $\overline{q_t'} = 0$ and $\overline{T_\ell'} = 0$. The standard deviation of s is

$$\sigma_s \equiv a_\ell(\overline{q_t'^2} - 2\alpha_\ell\overline{q_t'T_\ell'} + \alpha_\ell^2\overline{T_\ell'^2})^{1/2} \qquad (4.3)$$

The skewness and higher order moments could also be written in terms of higher order correlations involving q_t' and T_ℓ'. The complete description of the statistics of s is given by a probability distribution function (p.d.f.) G.

The unmeaned liquid water content is

$$q_\ell = \begin{cases} 0 & s \leq -Q_L \\ Q_L + s & s > -Q_L \end{cases} \qquad (4.4)$$

The mean cloud cover is the proportion of ensemble members with liquid water content greater than zero. In terms of G this is given by

$$C = \int_{-Q_L}^{\infty} G(s)ds \qquad (4.5)$$

The cloud liquid water content is similarly given by

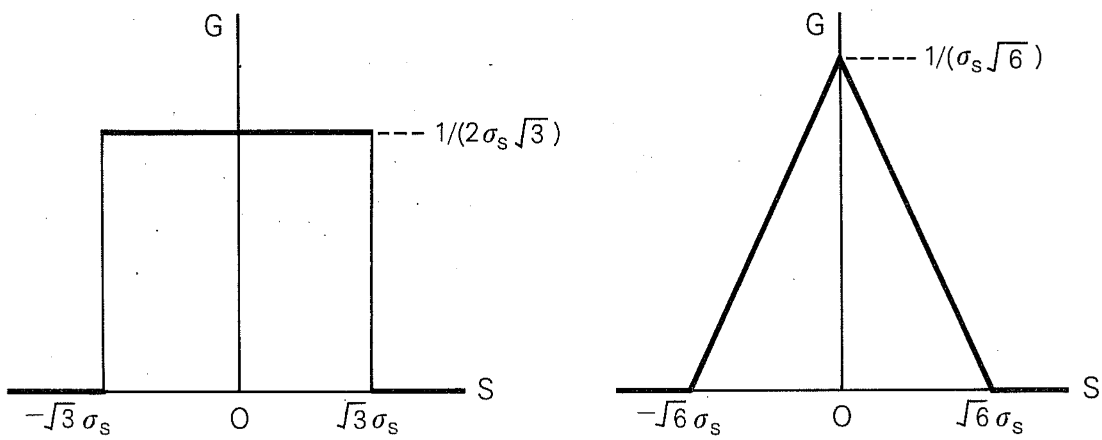
$$q_\ell = \int_{-Q_L}^{\infty}(Q_L + s)G(s)ds \qquad (4.6)$$

Because C and q_ℓ are both determined from G they are not independent: from (4.5) and (4.6) it can be shown that

$$\frac{\partial q_\ell}{\partial Q_L} = C$$

Therefore, if a particular choice of G gives a realistic cloud cover, we may expect the corresponding mean liquid water content to be reasonable. For the case of zero cloud, C = 0 and $q_\ell = 0$ whereas for complete cloud cover C = 1 and $q_\ell = Q_L$. The latter result is the formula which would be used in cloudy regions if partial cloudiness had not been considered.

G needs to be specified to close the system. This is the part of the scheme where a pragmatic and empirical approach is necessary. The p.d.f. is often taken to be a Gaussian with standard deviation σ_S although there has been work experimenting with skewed functions (Bougeault (1982)). However, in the context of an AGCM or operational NWP model we would like C and q_ℓ to be simple functions of Q_L if this can be achieved without sacrificing physical realism. If we use either of the two simple p.d.f.'s shown in Figure 1 we obtain corresponding cloud fractions and liquid water contents which are simple low-order polynomial functions of Q_L and σ_S.



Symmetric 'top-hat' p.d.f. with standard deviation σ_S

Symmetric 'triangular' p.d.f. with standard deviation σ_S

Figure 1. Two simple p.d.f.'s for the cloud ensemble model

The fluctuations whose statistical distribution is described by G are, in the context of a large scale model, not only due to small scale turbulence but also to mesoscale effects. So even if we had a higher order turbulence model to predict $\overline{q_t'^2}$ etc.it may not be appropriate to use the derived σ_S to diagnose the cloud fraction for a grid-box large enough to contain unresolved mesoscale motions. A specification for σ_S, chosen for its simplicity rather than for strong physical reasons, is

$$\sigma_S = \left(\frac{1-RH_C}{A}\right) a_\ell \, q_{sat}(T_\ell,p) \qquad (4.7)$$

RH_C is a critical relative humidity for cloud to occur and A is a constant ($A = \sqrt{3}$ for the "top-hat" p.d.f. and $A = \sqrt{6}$ for the "triangular" p.d.f.). When used with either of the simple p.d.f.'s described above this σ_s gives the cloud fraction as a simple function of relative humidity. Figure 2 illustrates the form of these functions for both p.d.f.'s and shows the quadratic dependence currently used in the diagnostic cloud scheme for comparison. A critical relative humidity of 85% has been assumed in this figure. It remains to be seen whether the cloud amounts produced by this scheme are realistic. If not, alternative p.d.f.'s could be sought. The choice of p.d.f. will be discussed further in section 6.

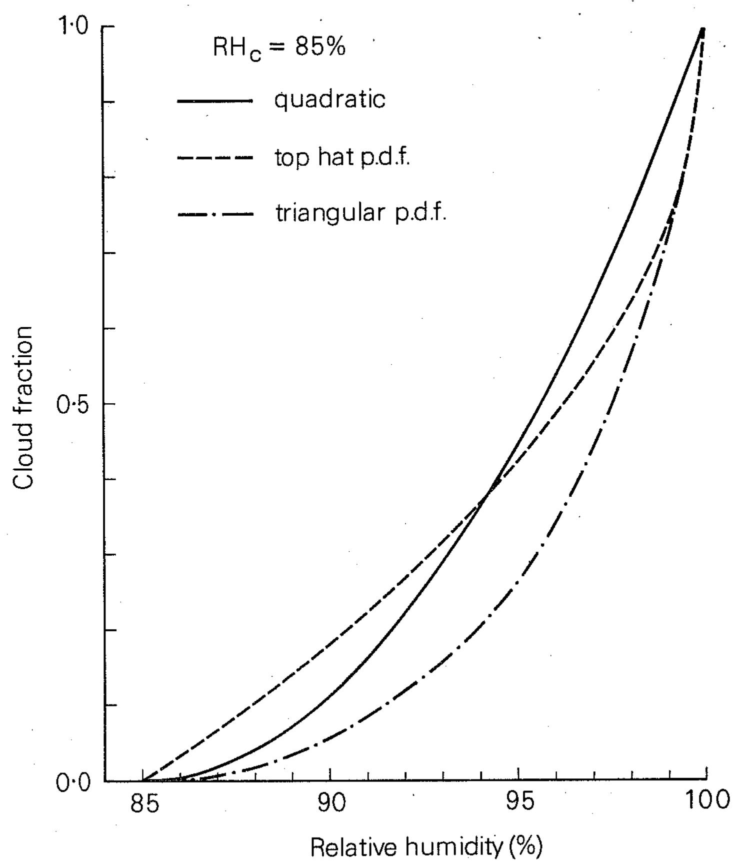


Figure 2. Cloud fraction as a function of relative humidity

5. PRECIPITATION AND RADIATION

Models with an explicit cloud water variable require a parametrization of precipitation somewhat more sophisticated than those without. The scheme must calculate the rate at which cloud water is converted to precipitation and deplete the cloud water content accordingly. One such parametrization has been developed by Sundqvist (1978 and 1981). The local rate of formation of precipitation will be denoted by P_t. With the notation of Section 2:

$$P_t = \frac{-d_p(q_\ell + q_f)}{dt} \geq 0 \qquad [P_t] = \text{kg kg}^{-1}\ \text{s}^{-1}$$

The following formula for the grid-box averaged rate of formation of precipitation P_t is based on Sundqvist's proposal:

$$\bar{P}_t = C(c_t + c_a \bar{P}_r) \left\{ 1-\exp\left[-\left(\frac{\bar{q}_\ell / C}{c_w} \right)^2 \right] \right\} \frac{\bar{q}_\ell}{C} \qquad (5.1)$$

The leading factor C in (5.1) is present because precipitation only forms in cloudy regions. Note that the in-cloud liquid water content $\bar{q}_\ell/C$ is used. The quantities c_t, c_a and c_w are constants, although they could, and probably should, be made to depend on temperature or cloud type. For initial tests the following values will be used: $c_t = 10^{-4}\ \text{s}^{-1}$, $c_a = 1\ \text{m}^2\ \text{kg}^{-1}$, $c_w = 5 \times 10^{-4}\ \text{kg kg}^{-1}$. P_r is the flux of precipitation ($\text{kg m}^{-2}\ \text{s}^{-1}$) reaching the layer from above; the term involving this quantity is a simple representation of the enhancement of the conversion rate due to accretion.

The timescale τ_c defined by

$$\tau_c^{-1} = (c_t + c_a \bar{P}_r) \left\{ 1-\exp\left[-\left(\frac{\bar{q}_\ell / C}{c_w} \right)^2 \right] \right\} \qquad (5.2)$$

is a conversion time for cloud water to precipitation. With this definition we can write

$$\bar{P}_t = \bar{q}_\ell/\tau_c \tag{5.3}$$

Care must be taken in the numerical scheme to ensure that no more than the available cloud water is converted during a timestep.

Denote the rate of evaporation of falling precipitation by P_e so that

$$P_e = \frac{-d_s(r_\ell+r_f)}{dt} \geq 0, \qquad [P_e] = \text{kg kg}^{-1}\ \text{s}^{-1}$$

using the notation of Section 2. A simple parametrization of P_e is used (Sundqvist (1981)):

$$\bar{P}_e = c_e(1-\bar{R})\ (\bar{P}_L)^{1/2} \tag{5.4}$$

$\bar{R}$ is the grid-box average relative humidity and $\bar{P}_L$ is the grid-square average flux of precipitaton entering the layer (kg m^{-2} s^{-1}). c_e is a constant taken to be 2 x 10^{-5} SI units ($kg_{(water)}^{1/2}$, $kg_{(air)}^{-1}$ m $s^{-1/2}$). The amount of evaporated precipitation must be limited to the amount of precipitation falling through the layer.

Finally it should be mentioned that the presence of an explicitly predicted cloud liquid water content enables liquid water optical path lengths to be calculated immediately. Cloud optical properties may then be computed interactively rather than specified. This would remove a constraint which is present in most AGCM radiation schemes at present.

6. DISCUSSION

It may legitimately be asked what advantage there is in a scheme such as described above over the simpler diagnostic cloud schemes. The preceding sections have shown how cloud can be made an integral part of the model. This is so in the sense that:

(i) the cloud has an associated and consistently derived liquid water content,

(ii) the thermodynamic and phase change effects of the cloud are represented,

(iii) the turbulent mixing scheme takes account of the presence of cloud and can recognise cloud top entrainment instability,

(iv) the model's non-convective rainfall is directly formed from the predicted cloud,

(v) cloud optical properties can be calculated from the cloud liquid water content.

However, the scheme should not be seen as being in conflict with the work currently being done to improve diagnostic cloud schemes (e.g. Slingo J.M. (1985)). Indeed this work could provide useful information for the further development of the scheme described in this paper. If the dependence of cloud amounts on identified quantities such as relative humidity, vertical velocity and thermal stability can be confidently established then we might be able to incorporate this knowledge in the p.d.f. G introduced in section 4. The precise form of G will have to remain simple enough to provide easily computable cloud amounts and liquid water contents. However, the low order moments of G such as its standard deviation and skewness could perhaps be parametrized in terms of the quantities identified as important for cloud prediction. This is a rather vague speculation as yet; the basic scheme should be assessed before further work along these lines is done.

REFERENCES

Bougeault, P. 1982: Cloud ensemble relations for use in higher order models of the planetary boundary layer. J. Atmos. Sci., 39, 2691-2700.

Deardorff, J.W. 1980: Cloud top entrainment instability. J. Atmos. Sci., 37, 131-147.

Mellor, G.L. 1977: The Gaussian cloud model relations. J. Atmos. Sci., 34, 356-358, 1483-1484.

Mellor,. G.L. and Yamada, T. 1982: Development of a turbulence closure model for geophysical flows. Rev. Geophys. and Space Phys., 20, 851-875.

Randall, D.A. 1980: Conditional instability of the first kind, upside-down. J. Atmos. Sci., 37, 125-130.

Randall, D.A. 1985: Parametrization of Boundary Layer Cloudiness. Proceedings of the ECMWF Workshop on Cloud Cover, 26-28 November 1984.

Slingo, A. and Wilderspin, R.C. 1985: Simulation of cloudiness with the U.K. Meteorological Office 11-Layer AGCM. Proceedings of the ECMWF Workshop on cloud Cover, 26-28 November 1984.

Slingo, J.M. 1985: Cloud cover experimentation in the ECMWF model. Proceedings of the ECMWF Workshop on Cloud Cover, 26-28 November 1984.

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

Suarez, M.J., Arakawa, A. and Randall, D.A. 1983: The parametrization of the planetary boundary layer in the UCLA GCM: Formulation and results. Mon. Weather Rev., 111, 2224-2243.

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

Sundqvist, H. 1981: Prediction of stratiform clouds: results from a 5-day forecast with a global model. Tellus, 33, 242-253.

Yamada, T. and Mellor, G.L. 1979: A numerical simulation of BOMEX data using a turbulence closure model coupled with cloud ensemble relations. Quart. J. Roy. Met. Soc., 105, 915-944.