The Roles of Dry Convection, Cloud-Radiation Feedback Processes and the Influence of Recent Improvements in the Parameterization of Convection in the GLA GCM

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ABSTRACT

The sensitivity of the simulated July circulation to modifications in the parameterization of dry and moist convection, evaporation from falling raindrops, and cloud-radiation interaction is examined with the GLA (Goddard Laboratory for Atmospheres) GCM (general circulation model). Inferences are based on several 47-day summer integrations using the same prescribed boundary forcings. The Arakawa–Schubert cumulus parameterization, together with a more realistic dry convective mixing calculation that allowed moisture, heat and momentum to mix uniformly, yielded a far better intertropical convergence zone (ITCZ) over North Africa than did the previous convection scheme. It also produced a rain-free Sahara desert, which is a well-known feature in observations, but is poorly simulated by a number of GCMs. The physical mechanism for the improvement was identified to be the upward mixing of planetary boundary layer (PBL) moisture by vigorous dry convective mixing, which prevented the buildup of moist convective instability in the PBL.

A modified rain-evaporation parameterization which takes into account the raindrop size distribution, the atmospheric relative humidity, and a typical spatial rainfall intensity distribution for convective rain was developed and implemented. It evaporated about 50% of the convective rain while greatly reducing the evaporation of large-scale rain. As compared to the old scheme which produced no evaporation of convective rain and the maximum possible evaporation of large-scale rain, the new scheme led to some major improvements in the monthly mean vertical profiles of relative humidity and temperature, convective and large-scale cloudiness, rainfall distributions, and mean relative humidity in the PBL.

When the convective cloud-radiation interaction was included by assuming that an entire sigma layer of a grid-box at the detrainment level(s) of Arakawa–Schubert clouds was covered by convective clouds of appropriate optical thickness, some major changes in the surface fluxes, diabatic heating, and orientation of the ITCZ over equatorial America were simulated. The experiment suggests a strong potential for further improvement of the GCM simulations by including more realistic parameterization of cloud-radiation interaction.

1. Introduction

Numerical general circulation models (GCMs) such as those available at the Goddard Laboratory for Atmosphere (GLA) are useful tools for simulating and investigating the behavior of the atmosphere in relation to the slowly varying boundary forcing at the earth’s surface and the time evolution of the atmosphere starting from a specific initial state. However, model deficiencies that make the simulations drift unrealistically from nature, which are present in even the state-of-the-art models of the atmosphere, often cast some doubt on the reliability of results of investigations performed with the models.

For example, one of the outstanding problems of all the GLA GCMs including the current fourth order model has been that the models produced significant rainfall over most of the Sahara desert for all July simulations (Charney et al. 1977; Shukla et al. 1981; Randall 1982; Kalnay et al. 1983; Sud and Smith 1984). It is helpful to point out that the problem of spurious rain in the Sahara desert has been present in other GCMs as well (see, for example, Williamson 1983; Cuninington and Rowntree 1986). Mintz (1984) attributed the anomalous rainfall to an excess of evapotranspiration such as found in the studies by Charney et al. (1977) that, in turn, was caused by excessive prescribed soil moisture that was held constant over the entire period of integration. Recently, a number of studies (Walker and Rowntree 1979; Kurbaatkin et al. 1979; Shukla and Mintz 1982; Sud and Fennessy 1984; Yeh et al. 1984; Carson and Sangster 1985) have investigated the role of soil moisture on Saharan, as well as global rainfall and circulation. Cuninington and Rowntree (1986) also discuss systematic studies of the influence of land surface fluxes and interaction of radiation with water vapor in the atmosphere on Saharan rainfall. All these models show the sensitivity of rainfall to surface fluxes. Accordingly, it was generally felt that
with the use of realistic surface albedo and soil moisture datasets, together with better and more realistic evapotranspiration parameterization, the problem of spurious rain over the Sahara desert should largely disappear. Consequently, the soil moisture dataset produced by Mintz and Serafini (1981) using observed rainfall climatology in a simple hydrologic balance model, a more realistic surface albedo dataset produced by Sud and Fennessy (1982), and the Deardorff (1972) planetary boundary layer (PBL) parameterization for surface fluxes were included. However, despite these improvements of land surface parameterizations, and modified surface albedo and soil moisture boundary conditions as incorporated by Randall (1982), the GLA GCM produced an even worse distribution of rainfall over the Sahara, Saudi Arabia and India. This suggested, among other things, that the prescribed uniform surface roughness over all land, which does some violence to the PBL processes, was questionable; large surface stress may yield too much cross-isobaric convergence from the monsoonal flow over the Sahara desert. Indeed, for the Sahara desert, reduction of land surface roughness led to some significant and desirable changes in the moisture convergence that essentially led to reductions in rainfall over the Sahara desert (Sud and Smith 1985).

However, it must be emphasized that none of the foregoing improvements produced a realistic ITCZ over the Sahara desert or produced a dry Sahara in any of the July simulations.

It can be safely concluded that even though the above improvements were useful and significant, they were not sufficient to eliminate the so-called drift in the model’s rainfall climatology of North Africa. This suggested that the amelioration of the problem of spurious model simulated rainfall over the Sahara desert might be found in other physical parameterizations of the model. The most likely and perhaps the least researched candidates were dry and moist convective processes and cloud-radiation interaction. Therefore, it was logical to examine the possibility of improving these parameterizations. The approach not only led to a satisfactory solution of the problem, but also revealed the critical role of three other basic mechanisms, namely, the dry convective adjustment process, evaporation of falling raindrops, particularly from convective clouds, and cloud-radiation feedback. We shall discuss the GLA model’s sensitivity to these parameterizations in the text of the paper and show how improvements in these have led to numerous improvements in the circulation and rainfall climatology of the July simulations.

2. Model description

The GLA GCM (previously referred to as the GLAS fourth-order GCM) has been described in detail by Kalnay et al. (1983), and has been used in a number of numerical weather prediction and data impact studies (Halem et al. 1982; Baker et al. 1984a,b; Atlas 1987). Nevertheless, this is the first time we are making a concerted attempt to systematically examine changes in the model’s July climatology in response to changes in physical parameterizations. Even though the model parameterizations and structure are discussed by many other scientists, we briefly outline some of the important aspects of the parameterizations in the following subsections.

a. Model structure and hydrodynamics

The version of the GLA GCM used in this investigation has a horizontal resolution of 4° lat by 5° long; it employs nine sigma layers of equal thickness between the top of the model atmosphere (at 10 mb) and the earth’s surface. The model has seven prognostic variables, in particular the zonal and meridional winds, temperature and specific humidity at each of the nine sigma levels in the free atmosphere, as well as the surface pressure, ground temperature and soil moisture at the surface of the earth. The governing equations for all of the prognostic variables are written in flux form and are solved by using fourth-order finite difference equations in space and the Matsuno (Euler backward) scheme for integration in time.

A sixteenth-order Shapiro (1970) filter is applied every 2 h to the sea level pressure and potential temperature on all longitudinal and latitudinal planes. However, the wind fields are filtered only in the longitudinal direction only. The overall influence is that wavelengths longer than four grid lengths, which are resolved accurately, are largely unaffected; however, wavelengths shorter than four grid lengths, which would otherwise be grossly misrepresented by the finite differences, are filtered out while they are still infinitesimal. Thus, filtering avoids the accumulation of energy that would otherwise occur at the shortwave cutoff.

Shallow-water model numerical experiments indicate that Kalnay’s scheme conserves potential enstrophy to an adequate degree of accuracy (Kalnay-Rivas and Hoistma 1979). Moreover, long integrations with the GLA fourth-order model have demonstrated good skill in simulating January and July climatology (Kalnay et al. 1983). In addition, the use of the Euler backward time integration scheme damps most of the gravity waves, and the vertical motion fields are well correlated with the simulated synoptic patterns.

b. Physical parameterizations

The parameterizations of physical processes have evolved from those originally described by Somerville et al. (1974). However, over the years some significant

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1 In all of the experiments that are reported herein, the soil moisture was prescribed.
improvements were made besides those discussed in this paper. The details of all these improvements are given in Kalnay et al. (1983). Specifically they include PBL parameterization by Sud and Abeles (1980), longwave radiation parameterization by Wu (1980) and fourth-order differencing in hydrodynamics by Kalnay and Hoistma (1979). The major physical parameterizations are those for the surface fluxes, dry and moist convective processes and radiative transfer, including cloud-radiation interactions. These are described briefly below.

1) SURFACE FLUXES

The surface fluxes of sensible heat, \( H_s \), evaporation or evapotranspiration, \( E_s \), and surface stress, \( \tau_s \), are calculated from the following bulk aerodynamic equations

\[
H_s = \rho C_D U_s C_U C_H(T_g - T_s),
\]

(1a)

\[
E_s = \beta \rho U_s C_U C_H(q^* (T_g, \rho_i) - q_s),
\]

(1b)

and

\[
\tau_s = \rho C_D U_s^2,
\]

(1c)

where \( \rho \) is the density of air, \( U_s \) the surface wind speed, \( C_U \) and \( C_H \) are bulk aerodynamic coefficients, \( T_g \) and \( T_s \) represent the ground surface and air temperature at the top of surface layer of the planetary boundary layer (PBL), respectively, \( q^* (T_g) \) is the saturation specific humidity at the ground temperature, \( q_s \) the actual specific humidity of air at the top of surface layer, and \( \beta \) a soil moisture availability parameter which is prescribed as a function of soil moisture (Sud and Fennessey 1984). To calculate \( q_s, T_g, \) and \( C_U \) and \( C_H \) we consider the structure of the entire PBL. We assume \( C_H = 1.5 C_U \) and \( C_D = C_U^2 \). However, \( C_D U_s \) is called the drag parameter and is carried as a composite variable in the model. The lowest sigma layer of the model is taken to be the PBL. Typically it has a thickness of about 110 mb. The PBL consists of a shallow surface layer, which has constant fluxes and relatively large gradients of temperature, humidity, and winds, surmounted by a mixed layer which is dominated by boundary layer eddies and has a large eddy transport coefficient, \( K_s \) and small gradients of temperature, humidity and winds.

The flux transport in the mixed layer can be expressed by

\[
H_m = K(q_s - q_a)/Z,
\]

(2a)

\[
E_m = K(q_s - q_a)/Z,
\]

(2b)

where \( H_m \) is the sensible heat flux into the mixed layer which must equal \( H_s \), \( E_m \) the evaporation into the mixed layer which must equal \( E_s \), \( K \) the eddy transport coefficient, \( Z \) the mixed layer height ( \( \sim 1 \) km for 110 mb thickness near the surface), and subscripts \( s \) and \( a \) refer to the top of the surface layer and mixed layer, respectively. By invoking a compatibility condition for fluxes at the interface of the surface and mixed layer from Eqs. (1) and (2), and using buoyancy-dependant formulations of \( C_D \) and \( K \), one can obtain \( T_s \) and \( q_s \), which then enables calculation of the PBL fluxes. The surface wind speed, \( U_s \), which is the vector sum of the zonal and meridonal wind components, is obtained by linear extrapolation of wind from the two sigma levels above. The ground temperature, \( T_z \), which is required to calculate the saturation specific humidity, \( q^*_s = q^*_s(T_z, \rho_i) \), is obtained by solving the ground temperature tendency equation

\[
C_g \frac{dT_z}{dt} = S - R - H_s - \lambda E_s.
\]

(3)

Here \( C_g \) is the heat capacity of a diurnally-interactive slab of soil. It represents the heat capacity of an imaginary slab of moist soil that heats uniformly and yields the same amplitude of diurnal temperature oscillation that an infinitely deep actual mass of earth would, if both were forced by a sinusoidal heat flux input. In Eq. (3), \( S \) is the absorbed shortwave flux, \( R \) the net outgoing longwave flux at the surface, \( E_s \) and \( H_s \) are the net upward latent and sensible heat fluxes at the surface, respectively, \( \lambda \) is the latent heat of evaporation of water. For stability, Eq. (3) is solved using an implicit-backward method of integration over a finite time step which is half an hour in the GCM.

2) DRY CONVECTIVE ADJUSTMENT

The dry convective adjustment is invoked if the potential temperature of a sigma layer exceeds that of the next higher sigma layer. The adjustment mixes the two air masses to yield the same potential temperature for both layers while conserving dry static energy. In this mixing, the moisture and momentum of each layer remain unchanged. This is an arbitrary restriction of the parameterization because, in nature, air mass exchanges during the mixing process must include its accompanying moisture and momentum. The momentum was not allowed to mix for another reason, namely that the subgrid-scale momentum exchange problem, in the presence of clouds, is somewhat complicated (Y. Mintz, personal communication 1985). We shall discuss this later. This type of adjustment has been used in many models (e.g., Arakawa 1972; Manabe 1975; Williamson 1983). In the GCM, this adjustment is carried out before any of the cloud processes. Its influence on the model simulated July circulation and rainfall is one of the issues addressed in this paper.

3) CLOUD PARAMETERIZATION

The model generates both convective and supersaturation clouds using the predicted fields of temperature
and specific humidity. The convective cloud calculation is an adaptation of Arakawa et al. (1969) parameterization for a three-level GCM and requires three cloud types (Arakawa 1972). Convective clouds are restricted to the lowest six sigma layers in the model, which are projected as three layers by strapping sets of two consecutive layers into one. A convective cloud ensemble is formed if the moist static energy at the base of the cloud ensemble exceeds the saturation moist static energy at its top. For deep convection this condition must be satisfied for all the strapped layers that are penetrated by the cloud ensemble. Evaporation of convective raindrops which usually fall in a concentrated region is assumed to be small and is neglected. For further details of the current procedure for strapping and unstrapping, see Helfand (1979) or Kalnay et al. (1983).

Supersaturation (also called large-scale) clouds can appear at any level if the specific humidity at that level exceeds the saturation specific humidity. Clearly, therefore, all supersaturation clouds produce rain. These clouds are generated after the atmosphere has been adjusted by the convective clouds. This sequence accounts for the fact that even though large-scale and convective clouds are generally mutually exclusive, when they occur together large-scale rain follows moist convection. In this particular parameterization, the rainfall from large-scale clouds must first evaporate to saturate a traversed model layer, and only the excess water is allowed to fall as precipitation.

The cloud processes produce precipitation and neutralize moist static instability, as well as eliminate supersaturation conditions, at 30-min time steps in the model. Hence, for a cloud to persist beyond one time step, the other physical and dynamical processes in the model must reinstate the instability and/or supersaturation condition. A more detailed discussion of the cloud parameterization may be found in Arakawa (1969 and 1972) and several earlier papers (Helfand 1979; Shukla and Sud 1981; Sud and Smith 1984; Wolfson et al. 1987).

4) RADIATION AND CLOUD–RADIATION FEEDBACK

The amount of solar radiation absorbed by the atmosphere is computed using a parametric method that is based on the multiple-scattering computations by Lacis and Hansen (1974). In this parameterization the optical thickness and, therefore, the absorption or reflection of shortwave radiation by the atmosphere, depends on the type of clouds, the specific humidity and the zenith angle of the sun. The prescribed values of optical thickness for different types of clouds are summarized in Table 1. The basis for choosing various optical thickness values is given in the paper by Lacis and Hansen (1974). For the cumulus clouds produced by the Arakawa–Schubert (1974) parameterization, an optical thickness of eight was assumed at any layer in which the cumulus cloud was present. At the surface of the earth the absorbed shortwave radiation is reduced by a fraction proportional to the surface albedo. In the stratosphere the absorption of shortwave radiation also depends on the ozone distribution which is seasonally and latitudinally prescribed.

The emission, transmission and absorption of longwave radiation is parameterized according to Wu (1980). The calculation allows for longwave interaction with water vapor, ozone, carbon dioxide, and clouds. Only the water vapor absorption calculation is carried out in detail; transmission functions for ozone and carbon dioxide are precalculated. Clouds predicted by the model behave as blackbodies with respect to longwave radiation, and entire grid volumes are considered to be either fully clear or cloudy. This is a questionable simplification, but it was accepted because the model did not produce fractional cloudiness. In addition, the longwave radiation calculation is performed every 3 h and the fluxes are saved to make temperature adjustments at half-hour time steps.

5) IMPLEMENTATION OF ARAKAWA–SCHUBERT PARAMETERIZATION

The Arakawa–Schubert cloud parameterization has been described and evaluated in a series of four papers. The first paper (Arakawa and Schubert 1974), describes the theory behind the parameterization; the

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Table 1. Radiative properties of clouds.

<table>
<thead>
<tr>
<th>Cloud type</th>
<th>Sigma layers</th>
<th>Optical thickness $(\tau)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Supersaturation</td>
<td>2</td>
<td>1*</td>
</tr>
<tr>
<td>Supersaturation</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>Supersaturation</td>
<td>4</td>
<td>4</td>
</tr>
<tr>
<td>Supersaturation</td>
<td>5 or 6</td>
<td>6</td>
</tr>
<tr>
<td>Supersaturation</td>
<td>7 or 8 or 9</td>
<td>8</td>
</tr>
<tr>
<td>Stopped penetrating convection</td>
<td>4–7 or 5–8 or 6–9</td>
<td>32*</td>
</tr>
<tr>
<td>Midlevel convection</td>
<td>5 or 6</td>
<td>8</td>
</tr>
<tr>
<td>Low-level convection</td>
<td>7 or 8</td>
<td>16*</td>
</tr>
<tr>
<td>Arakawa–Schubert convection</td>
<td>2, 3, 4, 5, 6, 7 or 8</td>
<td></td>
</tr>
</tbody>
</table>

* The optical thickness of large-scale clouds is a function of height. It is less for higher clouds because there is less moisture and clouds are thinner (Lacis and Hansen 1974).

Strapped clouds that occur in four layers have an average optical thickness of eight per layer (Lacis and Hansen 1974).

Low-level clouds have most moisture and have an optical thickness of 16 (Lacis and Hansen 1974)

The authors assumed that all detaining cumulus anvils have an optical thickness of 8 per layer.

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2 A simple fractional cloudiness parameterization was developed after writing this paper. A discussion of the influence of that parameterization will be a subject of a future paper.
second paper further discusses and evaluates the closure assumption (Lord and Arakawa 1980); the third paper evaluates its results against GATE phase III data (Lord 1982); and the fourth paper deals with its application to vertically discrete model GCM (Lord et al. 1982). The parameterization was earlier implemented in the UCLA GCM, which has a prognostic PBL depth (Suarez et al. 1983) and its application to other state-of-the-art GCMs has been extensively studied. We describe briefly the key implications of implementation of Arakawa–Schubert cumulus parameterization in the GLA GCM, which has a fixed PBL depth.

In the nine-sigma level GLA GCM, we allow clouds to detrain in seven layers, sigma layer 8 through sigma layer 2. This gives us seven cloud types, one for each detrainment level. The vertical structure of the thermodynamic variables of temperature, dry static energy, and moist static energy is assumed linear in \( \log p \) except for the boundary layer (sigma level 9 of the model) where specific humidity and potential temperature are assumed to be constant with height. This assumption is consistent with that employed in the UCLA GCM with a variable PBL depth. Calculation of the entrainment parameter \( \lambda \) is done for each cloud type such that the cloud attains a vanishing buoyancy at the particular sigma level. Subsidence associated with convection must be incorporated into the PBL. The UCLA GCM has a variable-depth PBL and its depth changes in response to the cloud base mass flux, \( M_B \). For the GLA GCM with a fixed PBL depth, any mass that is removed from the PBL by cumulus convection is ultimately returned to the PBL from the lowest cloud layer using an upwind scheme to account for subsidence between clouds.

(i) Moist convective adjustment. The Arakawa–Schubert (1974) parameterization handles the deep moist convective instabilities emanating from the cloud base, i.e., the top of the PBL. Any moist convective instability originating at the upper levels is handled by invoking mixing involving heat and moisture (Suarez et al. 1983). Partial mixing occurs when there is conditional instability but only the lower layer is saturated; otherwise, total mixing occurs. The mixing is performed from top to bottom taking two adjacent layers at a time; the process is continued until the entire vertical profile is moist adiabatically stable.

(ii) Dry convective adjustment. The slope of the dry static energy profile represents the vertical stability if the relative humidity is so small that no condensation can occur. A positive (negative) slope shows a stable (unstable) atmosphere at that point. The lower atmosphere gets highly unstable over dry regions in the summer where surface heating warms the PBL air to substantially increase its dry static energy with respect to the air above. The dry convective overturning that occurs mixes the two air masses to yield uniform dry static energy, humidity and momentum. This is the process of dry convective mixing. In the previous version of the GLA GCM only mixing of heat was allowed; however, we have examined the influence of mixing of moisture and momentum as well, because, in nature, mixing air parcels mixes moisture, heat and momentum.

3. Design of the experiments

Several modifications were made to the physical parameterizations of the GLA GCM, but their effects were examined by introducing them one at a time. In each case a 47-day simulation was made, starting from 15 June 1979. The initial conditions of the simulations were from the ECMWF (European Centre for Medium Range Weather Forecasts) analysis of actual observations, and the boundary forcings, i.e., sea surface temperature, soil moisture, surface albedo, and snow and sea ice extents, were all externally prescribed and taken from the standard GLA datasets (Kallay et al. 1983). As will be shown in section 4 (on results), some of the modifications made a very substantial impact on the model-simulated July climate which can be easily interpreted from physical considerations.

We have summarized the studies in Table 2. They were divided into two categories: (a) those performed with the standard model (Kallay et al. 1983) using the three-level cumulus parameterization of Arakawa (1969), identified as experiments S1 through S4; and (b) those performed with the Arakawa–Schubert (1974) cumulus parameterization as taken from the UCLA GCM (Suarez et al. 1983) identified as experiments A1 through A6. The differences between the Arakawa–Schubert (1974) cumulus parameterization and the Arakawa (1969) three-level parameterization are pointed out in appendix A. The control runs with the two cumulus cloud parameterizations are S1 and A1.

Parallel modifications were made to each model version to examine and compare the impact of the two convective parameterization schemes. Experiments S2 and A2 included moisture mixing during the dry convective adjustment process, and an essential requirement that the relative humidity at the top of the PBL exceed 90% for the onset of moist convection (Helfand 1979). The influence of the 90% relative humidity restriction was isolated in another simulation, A3, which was identical to A2 except for the absence of the relative humidity restriction.

Since moist convection will not be sustained in nature unless the latent heat of water vapor is released by condensation, the requirement for a 90% minimum PBL top relative humidity ensures the suppression of convection until a parcel somewhere in the grid area reaches its lifting condensation level (LCL). At 90% grid-averaged relative humidity, it is reasonable to as-
Table 2. July simulation of experiments and the relevant physical parameterizations.

<table>
<thead>
<tr>
<th>Expt. ident. no.*</th>
<th>Dry convection</th>
<th>Cloud base relative humidity restriction for moist convection</th>
<th>Rain-evaporation parameterization</th>
<th>Cloud-radiation interaction with Large-scale</th>
<th>Convective</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Temperature</td>
<td>90%</td>
<td>None</td>
<td>Yes</td>
<td>None</td>
</tr>
<tr>
<td>A2</td>
<td>Temperature and moisture</td>
<td>90%</td>
<td>None</td>
<td>Yes</td>
<td>None</td>
</tr>
<tr>
<td>A3</td>
<td>Temperature and moisture</td>
<td>None</td>
<td>None</td>
<td>Yes</td>
<td>None</td>
</tr>
<tr>
<td>A4</td>
<td>Temperature, moisture and momentum</td>
<td>90%</td>
<td>None</td>
<td>Yes</td>
<td>None</td>
</tr>
<tr>
<td>A5</td>
<td>Temperature, moisture and momentum</td>
<td>90%</td>
<td>Yes</td>
<td>Yes</td>
<td>None</td>
</tr>
<tr>
<td>A6</td>
<td>Temperature, moisture and momentum</td>
<td>90%</td>
<td>Yes</td>
<td>Yes</td>
<td>None</td>
</tr>
</tbody>
</table>

(a) With the Arakawa–Schubert cumulus parameterization

(b) With the Arakawa three-layer cloud parameterization

S1 Temperature only | None | None | Yes | Yes
S2 Temperature, moisture and momentum | 90% | None | Yes | Yes
S3 Temperature and moisture | 90% | None | Yes | Yes
S4 Temperature, moisture and momentum | 90% | Yes | Yes | None

* A stands for Arakawa–Schubert (1974) cumulus cloud parameterization; S stands for Arakawa (1972) three-layer cloud parameterization as described in Kalnay et al. (1983).

Assume that either an unsaturated parcel will be lifted about 100 m by turbulent eddies enabling it to reach its LCL, or that the horizontal inhomogeneity will ensure the existence of some seed regions of 100% or more relative humidity.

Since physically any mixing of air should result in the mixing of dry static energy, moisture, and momentum in a consistent way, it may be hypothesized that a dry convective adjustment scheme which fully mixes heat but leaves moisture and momentum unchanged was not as realistic as one that ensures full mixing of water vapor. Momentum mixing can be complex, but we allowed full mixing of momentum as well. Consequently, the dry convective adjustment was further modified to include full mixing of dry static energy, moisture, and momentum due to zonal and meridional winds, and the additional experiments S3 and A4 were performed to examine the influence of this change.

Experiments S4 and A5 included a simple, but realistic, parameterization of evaporation from falling raindrops. A detailed justification for including this may be found in Sud and Molod (1986). A brief description of the parameterization is given in appendix B. Inclusion of the rain evaporation parameterization

![Figure 1. Observed July precipitation (mm day⁻¹) from Jaeger (1976).](image-url)
Fig. 3. (a) Sensible heat flux for $F_1$ in W m$^{-2}$; (b) sensible heat flux for $A_6$ in W m$^{-2}$; (c) evapotranspiration for $S_1$ in mm day$^{-1}$; and (d) evapotranspiration for $A_6$ in mm day$^{-1}$. 

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was an attempt to address some weaknesses in the model results outside the framework of the Arakawa–Schubert convection scheme.

Finally, experiment A6 included an additional modification of the cloud-radiation. At the level of a detraining convective cloud anvil, an entire grid box was assumed to be filled with a cloud of appropriate optical thickness. It was recognized that this assumption might produce an overestimate of cumulus cloud cover, but considering that the effect at the surface of a cloud in any one layer with a typical optical thickness of about 8 is not too different from that of clouds in all the layers (i.e., shortwave transmission function $e^{-\mu}$ for $\mu \equiv 8$ is less than 2% for overhead sun, where $\mu$ is cosine of solar zenith angle and $\tau$ is the optical thickness which, in our model, is 8 for each sigma layer) the assumption was considered reasonable. Following the results of this investigation, we are now expending a major effort to improve the cloud-radiation interaction by introducing a physically realistic parameterization for fractional cloudiness and nonprecipitating large-scale clouds.

4. Results

In this section we analyze and discuss the effects of the various changes in the parameterizations of dry and moist convection on the circulation and rainfall as simulated by the GLA GCM for July.

One of the most well-known estimates of July rainfall climatology that is based on observations is by Jaeger (1976) and is shown in Fig. 1. In comparison the standard GLA GCM simulated July rainfall is shown in Fig. 2a. Spurious rainfall over the Sahara desert, too much rainfall over the continental United States, and a poor prediction of the ITCZ over the ocean are among the model's weaknesses. Although in the past it was believed that some of the discrepancies in rainfall may be due to a poor simulation of the earth's surface fluxes (Mintz 1984; Cunnington and Rowntree 1986), the model's conspicuous weakness of producing significant convective rainfall over the Sahara desert cannot be fully attributed to local evapotranspiration. In this model, evapotranspiration (Figs. 3c and 3d) is much improved as compared to that obtained by Charney et al. (1977). For instance, large spans of the Sahara desert have evapotranspiration between zero and 1 mm day$^{-1}$. Moreover, the sensible heat flux and evapotranspiration for S1 and A6 are quite similar, which indicates that the circulation and rainfall changes are not produced by earth-atmosphere interaction changes.

When the Arakawa–Schubert (1974) cumulus parameterization was included with the 90% restriction on the PBL-top relative humidity, the rainfall over the Sahara region was reduced somewhat, but it was still unrealistic (Fig. 4a). For the exploded view of rainfall distribution over Africa, see Fig. 5. Excessive rain to the south of the Sahara is produced in the Arakawa–Schubert simulation due to excessive heating that was a direct consequence of the neglect of convective-cloud radiation interaction. However, when in addition to the relative humidity restriction, the moisture was allowed to mix during the dry convective mixing episodes
which are quite vigorous over the Sahara desert in July (simulations S2 and A2), some marked improvements in the precipitation pattern began to emerge (see Figs. 2b and 4b). In particular, a far more realistic ITCZ over North Africa was produced in the simulation A2 (Fig. 4b). Even simulation S2, with the Arakawa (1969) three-layer parameterization, yielded improvement over S1 (Fig. 2b) but the improvement was much less than that in A2 because the midlevel convection from the three-layer scheme continued to produce some rainfall in the Sahara desert. The Arakawa–Schubert scheme, which produces no midlevel convection, did not produce any rain over the Sahara desert. This comparison indicates that the inclusion of the Arakawa–Schubert (1974) cumulus parameterization has an important and beneficial influence on Saharan rainfall, partly because the model employs a moist convective adjustment and large-scale rain parameterization which effectively evaporates rain from midlevel convection in all the levels above the PBL. It is reasonable to expect that rain from midlevel convection over the dry Sahara would evaporate before reaching the ground. Experiment A3, in which the 90% relative humidity restriction was removed, led to some marginal deterioration in the North African ITCZ (Fig. 4c). The rainfall time series showed that spurious Saharan rainfall began to appear in late July. At this point, it was clear that the improvement in the dry convective mixing, used with the Arakawa–Schubert cumulus parameterization, was the primary source of the more realistic ITCZ over North Africa. Indeed, when the same modifications were made to the Arakawa et al. (1969) three-layer parameterization, there was a similar reduction in rainfall over Sahara. However, its rainfall distribution around Sahara was not as impressive (not shown). This indicates that a part of the overall improvement is attributable to the superiority of the Arakawa–Schubert parameterization.

In order to isolate the physical mechanism responsible for producing the improvement in the ITCZ over North Africa, we examined a zonally averaged wind vector map for a Saharan window (Fig. 6) as well as the vector wind maps at the lowest two model-sigma levels (Figs. 7a and 7b). The northward excursion of PBL winds at approximately 20°N carrying moisture from the monsoonal circulation, the subsequent mixing of the moisture into sigma level 8 due to dry convective mixing, and its consequent transport toward the equator, explains the entire water vapor cycle that produced the dry Sahara in the simulation.

Experiment A4, which invoked total mixing of momentum, heat and moisture during dry convection produced the best results yet for both the short-range forecasts and July climatology (Fig. 4d). For a justification of this modification, refer to section 5c. However, it must be pointed out that a comparison of the A4-simulation rainfall with Jaeger’s rainfall estimates shows the following weaknesses: (i) too much rainfall over the land, particularly over the continental United States; (ii) still somewhat unrealistic ITCZ simulation over the ocean, i.e., it is too weak and does not have the proper structure; (iii) too much rainfall over the maritime islands of Southeast Asia; (iv) very little large-scale rainfall and a distorted South Pacific convergence

![Figure 6](image)

**Fig. 6.** Time series of zonally averaged wind vectors for simulation A4, for latitudes near the summer ITCZ (1.0 m s⁻¹ = distance of 1° latitude on the map).
Fig. 7. Global wind vectors for simulation A5 in m s\(^{-1}\): (a) sigma level 8, and (b) sigma level 9. Confluence line for meteorological equator are shown by a thick line.
zone; and (v) too dry an upper atmosphere in the regions of strong moist convection (Fig. 8a). Some of these results are in agreement with earlier findings using the Arakawa–Schubert (1974) cumulus parameterization scheme in other models. Suarez et al. (1983) and Randall et al. (1985) show that the UCLA model with a variable PBL depth also produces no rain over the Sahara desert. We believe that the key mechanism in the UCLA model is also that of vertical moisture mixing produced by the variable depth PBL as well as and the explicit moisture mixing by dry convective adjustment.

In simulations S4 and A5 the evaporation from falling raindrops was included. The parameterized evaporation of rain is a function of the rain intensity distribution, drop size distribution, fall velocity of raindrops, and vapor pressure gradients between the hydrometeors and ambient air (Sud and Molod 1986; also appendix B herein). This parameterization was found to greatly improve both the global distribution of large-scale and convective precipitation. Results are shown for both types of simulations (see Figs. 2d and 4e), but we shall discuss only those for A5 as we have already pointed out the important influence of the Arakawa–Schubert (1974) convection scheme on Saharan rainfall. The overall effect of the new rain evaporation parameterization was to evaporate about 50% of the convective rainfall, and, conversely, as expected, evap-

![Fig. 8. Zonally averaged relative humidity (percent): (a) simulation A4, and (b) simulation A5.](image-url)
orotate less large-scale rain than before because the atmosphere traversed by the large-scale raindrops is left somewhat unsaturated in the modified parameterization. The global effect of the rain evaporation parameterization can be seen on the daily global rainfall plots. A comparison between rainfall reaching the ground and rainfall starting at its source at the cloud detrainment level shows the differences in the amount of rain evaporated by the old and new parameterizations (Figure 9). Examination of the zonally averaged relative humidity in simulation A5 (Fig. 8b) as compared to that for simulation A4 (Fig. 8a) indicates three important differences: (i) the zonal structure of the new relative humidity profile is far more realistic; in particular, it simulates high relative humidity in the upper atmosphere between 10°–14°N, the ascending branch of the Hadley cell; (ii) the zonally averaged relative humidity near the ground is also lower and more realistic (Fig. 8b). A high relative humidity near the ground has been an outstanding weakness of the GLA model, and its improvement is potentially useful in realistic simulations of the hydrological cycle. The new rain evaporation scheme has improved the relative humidity profile by allowing moistening of the layers below cloud detrainment through evaporation of falling raindrops. (iii) Monthly July rainfall field exhibits many improvements (Fig. 4e). These include emergence of a South Pacific convergence zone, better simulation of rainfall over equatorial Africa, and a generally reduced rainfall over convective regions. Large-scale rain also appeared in the Indian monsoon region (not shown), which is also realistic, because it is well known that about half of the monsoon rainfall over India is of large-scale origin (see e.g., Ramage 1971, pp. 101–103; and Rao 1976, Chap. XII). Among the weaknesses of this simulation are excessive evapotranspiration and rainfall globally, dislocated ITZC over the Americas, too high a continental evapotranspiration (Fig. 3d) and too much solar radiation reaching the ground which corresponds with too low a planetary albedo (Fig. 10a).

The latter weakness suggested a need for inclusion of convective cloud-radiation interaction. A simple convective cloud-radiation feedback, as described in section 3, was included in experiment A6. Results show a better planetary albedo distribution (Fig. 10b) reduced solar radiation at the ground (not shown), and reduced diabatic heating of the PBL over land. Globally, net surface radiation balance reduced from 170.8 to 147.5 W m⁻². This reduced the moisture convergence into continental land masses, thereby reducing the rainfall over North America and tropical Africa. It also strengthened the ITZC and produced a more realistic South American rainfall pattern. Figures 11a and 11b show that evaporation of falling rain increases deep diabatic heating in the ITZC regions and reduces it in the PBL, near the surface, particularly from 30° to 60°N, a region dominated by landmass.

In experiment A6 the weakness of excessive rainfall over Southeast Asia and over maritime islands persists. Analysis of moisture convergence in the PBL and evaporation flux from the surface for simulation A6 shows that the total rainfall derives its moisture approximately 40% from evaporation and 60% from convergence. Moreover, because there is clear evidence of reduced rainfall in regions surrounding the monsoon flow, some role of excessive moisture convergence is clear, but that points to a possible weakness in the moisture advection scheme. However, if too much moisture is entrained into the clouds, then the increased moisture convergence will follow the excessive heating of the PBL and the weakness lies in the cumulus scheme. It appears that the moisture advection and cumulus convection schemes are intertwined, and therefore it is difficult to isolate the real cause of these persistent weaknesses without doing further diagnostic work.

**Fig. 9.** Time series of globally averaged precipitation in mm day⁻¹ comparing A4, and A5 simulations [without and with Sud and Molod (1986) rain evaporation], respectively, for (a) convective precipitation, and (b) large scale precipitation.
5. Conclusions

The studies reported in this paper were initiated because large amounts of rainfall were simulated over the Sahara desert in all of the July simulations made with the GLA GCM. This was clearly a deficiency of the model. Following several well-documented studies of the importance of land surface fluxes (Charney et al. 1977; Walker and Rowntree 1979; Sud and Fennessey 1982, 1984; Sud and Smith 1985; Shukla and Mintz 1982; Cunnington and Rowntree 1986), it was hypothesized and generally believed that a realistic model of the biosphere was vital to realistic rainfall climatology. However, when some of the biosphere-related improvements were introduced into the model, and sensitivity studies were conducted with various versions of the GLA GCM, they produced little evidence of solving the problem of rainfall in the Sahara (Randall 1982), and we proceeded to examine the dry and moist convection, and cloud radiation processes as alternative possibilities. These avenues not only led to a solution of the problem, and a clearer understanding of the physical and dynamical processes responsible for the presence of absence of rain over the Sahara desert, but also helped us to improve the model simulations in general.

The key findings can be summarized as follows:

(a) Over North Africa, in July, there is convergence of moisture into the thermal low near the ground both in model simulations as well as in nature. However, in nature, the vigorous dry convective adjustment that follows strong PBL heating in the absence of evapotranspiration mixes the moisture into upper layers of diverging air. If that mixing were not allowed to take place, the moisture would keep accumulating until the steady moist static energy buildup made the PBL moist convectively unstable with respect to the overlying air mass, eventually leading to episodic moist convective outbursts and rainfall. The upward mixing of moisture from the lowest model layer provides an outlet path through the diverging air aloft and in that way it has a major role in eliminating the moisture buildup and spurious rainfall over the Sahara desert.

(b) A 90% relative humidity restriction for the onset of moist convection implies that moist air parcels that initiate cumulus convection must be within 100 m of their lifting condensation level (LCL). This physical condition had some very useful effects on the monthly July rainfall. Without this restriction, spurious rainfall begins to appear in the Sahara in the latter part of July, and the time series shows a slow northward drift of

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Fig. 10. Global planetary albedo for simulations (a) A5 and (b) A6.
simulated ITCZ rainfall as compared to Jaeger's (1976) observations (Fig. 1).

(c) Mixing of momentum along with the other atmospheric quantities during dry convective adjustment did not negatively influence the monthly July simulation. Since in nature a mixing process must include some mixing of all the properties of the air masses, we emphasize that momentum mixing must not be excluded in the dry convective adjustment parameterization, as has been the case in many models (see, for example, Williamson 1983; Halem et al. 1977). We further clarify that countergradient eddy momentum fluxes, which may accompany momentum mixing, do not occur in the simple dry convective mixing process. The coarse resolution of a nine-level GCM makes the parameterization a subgrid scale process. In the future, we propose to use eddy mixing for dry convective mixing following Tiedtke (1983) who developed a simple-eddy-exchange-coefficient dependent moist-convective adjustment scheme to account for shallow convection.

(d) Evaporation from falling raindrops is a natural process. It happens for convective as well as large-scale rain. Two major clues that suggested the need for a new rain evaporation parameterization were 1) simulations giving too dry an upper atmosphere with the Arakawa–Schubert scheme, and 2) spurious rainfall over the Sahara with the Arakawa three-level scheme (standard model). We conjectured that although there may be some rain detrained from clouds over the Sahara in nature, it will evaporate into its dry environment and not reach the ground. The parameterization of rain evaporation was kept simple because the actual subgrid-scale distribution of rainfall is difficult to determine. Inclusion of the rain evaporation scheme following Kessler's (1969) relations resulted in some ma-
ior improvements in the global climatology of rainfall and circulation. Indeed, the relative humidity and moist static energy profiles became far more realistic.

(c) It was evident that excessive rainfall over major continents was caused, in part, by excessive evapo-transpiration, which, in turn, was produced by a large radiation imbalance at the surface. This suggested a need for including convective cloud-radiation interaction. When a simple convective cloud radiation interaction was included by assuming that the entire grid box was covered by detrainment convective clouds, we noted some major improvements in the radiation balance at the earth’s surface, PBL fluxes, and resultant changes in circulation and rainfall. Experiment A6 overcompensates some of the weaknesses of experiment A5, particularly in the summer in the Northern Hemisphere. Currently, more work is in progress to include fractional cloud cover for convective as well as non-precipitating clouds in an attempt to realistically correct for the effect of the cloud-radiation feedback mechanism.

In summary, the systematic examination of various model modifications produced model simulations that represent some real improvements in the July climatology. Our 10-day forecast tests with the newly modified model revealed no significant improvement in the forecast skill with a sample of four forecasts. For further detail of these tests refer to Helfand et al. (1988).

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APPENDIX A

Cumulus Parameterization Schemes Analyzed

The differences and similarities between Arakawa (1969) and Arakawa–Schubert (1974) cumulus parameterizations are described by comparing their closure conditions, respectively, following Arakawa and Chen (1986). Some additional assumptions which are necessary for implementation of these schemes in the GLA GCM, with nine sigma-layers of equal mass in the vertical and fixed boundary layer height, are also discussed.

Following Yanai et al. (1973) the total diabatic heating, $Q_1$, and latent heat (water vapor) removal, $Q_2$, enter the temperature and moisture conservation equations in the following way:

\[
C_p \left[ \frac{\partial \tilde{\theta}}{\partial t} + \tilde{v} \cdot \tilde{\nabla} \tilde{\theta} + \tilde{\omega} \frac{\partial \tilde{\theta}}{\partial \tilde{p}} \right] = \left( \frac{p_0}{p} \right)^{K/C_p} \dot{Q}_1 \tag{A1}
\]

and

\[
L \left[ \frac{\partial \tilde{q}}{\partial t} + \tilde{v} \cdot \nabla \tilde{q} + \tilde{\omega} \frac{\partial \tilde{q}}{\partial \tilde{p}} \right] = -\dot{Q}_2. \tag{A2}
\]

Here an overbar denotes an ensemble mean over the grid domain; $\theta$ and $q$ are potential temperature and specific humidity, respectively; and the rest of the symbols have their usual meaning.

The primary goal of a cumulus parameterization is to develop a realistic criterion for the onset of cumulus convection and then to calculate the overall effect of subgrid-scale cumulus ensemble on the environment, i.e., change in temperature and humidity profile in the vertical. Accordingly, we describe both aspects of both schemes in the following sections.

Arakawa (1969) uses Eq. (A3a) for the onset of cumulus convection and therefore Eq. (A3b) becomes type I constraint:

\[
h_B - h_B^* > 0 \tag{A3a}
\]

\[
h_B - h_B^* = 0 \tag{A3b}
\]

where $h_B = (c_p T + gz + Lq)_B$ is the moist static energy at the cloud base and $h_B^* = (c_p T + gz + Lq^*_B)$ is the moist saturation energy at cloud top level.

The onset condition for cumulus convection in Arakawa–Schubert (1974) is also based on (A3a) but it has been refined. It requires

\[
A(\lambda) - A_c(\lambda) = 0 \tag{A4}
\]

where $A$ is the cloud work function for cloud type $\lambda$ which is equal to the cloud entrainment rate—a particular cloud type is solely dependent on $\lambda$—and $A_c(\lambda)$ is the corresponding critical value of $A(\lambda)$. If $A(\lambda) > A_c(\lambda)$, a cumulus cloud of that type is formed and it adjusts the temperature and humidity in conjunction with other clouds to restore constraint (A4). For further details of the cloud work function (which equals the integrated buoyant energy of the cloud) see Arakawa and Chen (1986).

The second closure assumption, i.e., a type II constraint, both in Arakawa (1969) and Arakawa and Schubert (1974) parameterizations comes from the use of a cloud ensemble model; clouds emerge at the cloud base, grow with entrainment of environmental air by the cloud plume, and dissipate following detrainment of different cloud types at different levels. These processes govern the vertical distribution of $Q_1$ and $Q_2$. The overriding constraint for the vertically integrated mean quantities (denoted by tilde) for the air-columns always hold:

\[
\tilde{Q}_1 - \tilde{Q}_B = \tilde{Q}_2. \tag{A5}
\]

Here $\tilde{Q}_B$ represents all diabatic heating processes except condensation heating. Even though constraint (A5)
applies equally to the Arakawa et al. (1969) and Arakawa–Schubert (1974) scheme, there are some intrinsic differences between them.

The Arakawa et al. (1969) three-layer cumulus parameterization was first employed in the UCLA GCM, Arakawa (1972). It was adapted into the nine-level GLA GCM by strapping the bottom six layers into three pairs starting from the bottom and excluding the top three sigma layers. The resultant three layer cumulus parameterization can have only three types of cumulus clouds: one between the bottom and middle stratified layers called shallow clouds, one between the middle and top layer called midlevel cumulus clouds, and one between the bottom and top layer called deep cumulus clouds. Shallow cumulus clouds do not precipitate while mid- and deep cumulus clouds produce precipitation. The equations for midlevel and shallow clouds are

\[
\frac{\partial h_{1,3}}{\partial t} = f_{m,3}(h_{c,1,3} - h_{2,4}) \tag{A6a}
\]

and

\[
\frac{\partial h_{3,5}}{\partial t} = f_{m,3}(h_{2,4} - h_{3,5}) \tag{A6b}
\]

where \( h \) is the moist static energy; suffixes \( l \), \( l+2 \), \( l \) for midlevel and shallow convection between layers 1-3 and 3-5; and \( f_{m,3} \) represents fractional mass flux for midlevel and shallow convection, respectively; \( h_c \) represents moist static energy of the cloud as opposed to \( h \) which represents moist static energy of the environment.

Equation (A6a) gives the time rate of increase of moist static energy at the cloud detrainment level; it equals the cloud mass fraction, \( f_m \) or \( f_s \) (as the case may be) times the difference between the moist static energies of the detraining cloud and its environment. This cloud mass flux displaces the environmental air which subsides and replaces the mass flux lost to the cloud in the layer below. The corresponding equation for the moist static energy increase at the level of cloud base is (A6b). (See Fig. A1 for a schematic representation of the cumulus cloud processes.) These clouds essentially destroy moist convective instability between the levels involved. Therefore, the magnitude of \( f_{m,3} \) is such that it reinstates constraint (A3) at the end of every physics time step (which is \( \frac{1}{2} \) h for the GLA GCM).

A similar method is used for deep cumulus convection which begins at level 5, entrains environmental air at level 3, and then detrains at level 1. In fact two types, type A and type B, of deep clouds can appear. They are defined as follows:

- **type A**: \( h_5 > h_f > h \)
- **type B**: \( h_5 > h_1 > h_f \).

Figure A1 shows a schematic of the mass exchange among the three layers. Mass flux \( f_d \) is determined from the condition \( h_c = h_f \) at the end of convection whereas the entrainment factor \( \eta \) is determined by assuming that the cloud plume must achieve neutral buoyancy with respect to its environment at the middle layer. For further details on the implementation of Arakawa et al. (1969) parameterization in the GLA GCM see Kalnay et al. (1983).

In the Arakawa–Schubert cumulus parameterization, total cumulus mass flux \( M_c(z) \) is expressed by

\[
M_c(z) = \int_{0}^{\lambda_{max}} M_B(\lambda) \eta(z, \lambda) d\lambda \tag{A7}
\]

where \( \eta(z, \lambda) \) is the cloud mass flux at height \( z \) normalized with the cloud base mass flux \( M_B(\lambda) \) for a cloud entrainment parameter \( \lambda \). For each cloud type, exponential growth of mass and therefore \( \eta(z, \lambda) \) is assumed giving

\[
\eta(z, \lambda) = e^{\lambda(z-z_B)} \tag{A8}
\]

where \( z \) and \( z_B \) are the heights of the cloud plume and

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**Fig. A1:** Schematic diagram showing cumulus fluxes for shallow/mid-level and deep cumulus clouds.
cloud base respectively. For a mass detrainment $D(z)$ per unit height for a cloud with $\lambda = \hat{\lambda}$ we get

$$D(z)dz = -M_s(\hat{\lambda})\eta(z, \hat{\lambda})d\hat{\lambda}. \quad (A9)$$

The relationship between $Q_1$ and $Q_2$ then follows:

$$Q_1(z) = \frac{1}{\rho} \left[ -D(z)L\hat{\lambda}(z) + M_e(z)\frac{\partial s}{\partial z} \right] \quad (A10a)$$

and

$$-Q_2(z) = \frac{L}{\rho} \left[ D(z)(\hat{\lambda}(z) + q^* - q) + M_e(z)\frac{\partial q}{\partial z} \right] \quad (A10b)$$

where $\hat{\lambda}(z)$ is the liquid water mixing ratio, $L$ the latent heat of condensation, $M_e(z)$ the cloud mass flux at level $z$ and $\partial s/\partial z$ is the rate of change of dry static energy in the vertical. Solution of (A10a) and (A10ab) yields the $Q_1$ and $Q_2$ distribution in $z$. Arakawa and Chen (1986) show that if detrainment effects of the first term on the right-hand side of Eq. (C10a) and (C10b) are ignored, one gets a simple relation:

$$Q_1/(\partial s/\partial z) = -Q_2 \left/ \left( \frac{L}{\rho} \frac{\partial q}{\partial z} \right) \right. \quad (A11)$$

For details on the implementation of Arakawa–Schubert cumulus parameterization in the GLA GCM, see section 2b5.

In the Arakawa (1969) parameterization, also, the basic equations (A6a) and (A6b) can be expressed as

$$\frac{\partial h}{\partial t} = f_{ms} \frac{\partial h}{\partial z}$$

and when $Q_1$ and $Q_2$ are represented in A12 in the form (A10a) and (A10b) we get

$$Q_1 = \frac{1}{\rho} f_{ms} \left( \frac{\partial s}{\partial z} \right) \quad (A12a)$$

and

$$Q_2 = -\frac{L}{\rho} f_{ms} \left[ (q^* - q) + \frac{\partial q}{\partial z} \right]. \quad (A12b)$$

If we assume $q^* - q \approx 0$, the equations again give

$$Q_1 \left/ \left( \frac{\partial s}{\partial z} \right) \right. = -Q_2 \left/ \left( \frac{L}{\rho} \frac{\partial q}{\partial z} \right) \right. \quad (A13)$$

which is identical to type II constraint in Arakawa–Schubert (1974). In both parameterizations, the cumulus processes destroy the convective instability which the large-scale circulation and boundary layer fluxes must reinstate in order for future cumulus convection to reoccur.

APPENDIX B

Rain-Evaporation Parameterization

The evaporation of falling precipitation (hereafter evaporation) of convective and large-scale origins is a process that reduces the net precipitation reaching the surface of the earth. The amount of evaporation depends on the temperature, pressure, and relative humidity of the air traversed by the falling raindrops. When rainfall begins, the evaporation is large, but as the air becomes saturated the evaporation diminishes. In a persistent large-scale precipitation episode, the evaporation vanishes as the relative humidity of air approaches 100%. Clearly, therefore, the evaporation should be directly related to the water vapor density deficit between the air and the surface of the raindrops, $(q_a - q_d)$. The physical process of mass exchange between air and falling spheres of rain must take into consideration their diameters, fall velocities and a water vapor diffusion coefficient. For a single raindrop, the expression of Kinzer and Gunn (1951), as given in Table 117 of the Smithsonian Tables (1958), gives a relationship for the rate of change of mass of a falling water drop:

$$\frac{\partial M_i}{\partial t} = 2\pi D \left( 1 + \frac{FD}{2S} \right) K(q_a - q_d) \quad (B1)$$

where $M_i$ is the mass of raindrop $i$ in grams, $D$ is its diameter in meters, $F$ a dimensionless ventilation factor, $S$ the equivalent thickness of transition shell over the raindrop, and $K$ the eddy exchange coefficient in $m^2$ sec$^{-1}$.

Within $\pm 20\%$ accuracy, Kinzer and Gunn (1951) experimentally related the first expression on the right to the raindrop diameter as follows:

$$2\pi D \left( 1 + \frac{FD}{2S} \right) = 2.24 \times 10^3 D^{8/5} \quad (B2)$$

and the second expression to the saturation deficit, $m(m < 0)$, by

$$K(q_a - q_d) = 10^{-5}m. \quad (B3)$$

Using (B2) and (B3) we can rewrite (B1) as follows:

$$\frac{\partial M_i}{\partial t} = 2.24 \times 10^{-2} D^{8/5}m. \quad (B4)$$

Assuming that the raindrops are size-distributed in approximate accord with an exponential law proposed by Marshall and Palmer (1948), $N = N_0 \exp(-\lambda D)$, integration over the entire spectrum of raindrop diameters yields

$$\frac{\partial M}{\partial t} = 2.24 \times 10^{-2} N_0 \int_0^{\infty} e^{-\lambda D} D^{8/5} dD. \quad (B5)$$

For the above raindrop distribution, Kessler (1969) obtained the mass of raindrops per unit volume $M$ in gm m$^{-3}$ from

$$M = \int_0^{\infty} \frac{\pi D}{6} D^3 N_0 e^{-\lambda D} dD = \frac{\pi N_0}{6} \frac{\Gamma(4)}{\lambda^4}. \quad (B6a)$$
where \( \rho \) is the density of liquid water and \( \Gamma(4) = 6 \). Taking \( \tau = 10^6 \text{ gm m}^{-3} \), we get

\[
\lambda = 42.1 N_0^{-1/4} M^{-1/4} \tag{B6b}
\]

Substituting (6b) into (5) and integrating over all diameters yields

\[
\frac{\partial M}{\partial t} = 1.93 \times 10^{-6} N_0^{7/20} M^{13/20} \tag{B7}
\]

In the GLA and other GCMs, a typical physical adjustment time-step of \( 1/2 \) h is used. Through this period, the rainfall rate and therefore the mass of water drops per cubic meter of air, \( M \), is assumed to be constant. The evaporation rate, \( \partial M/\partial t \), merely helps to increase the water vapor density of the air and must equal \( \partial q_a/\partial t \) because the saturation vapor density of evaporating raindrops, \( q_a \), may be held constant. Therefore, we write

\[
\frac{\partial M}{\partial t} = - \frac{\partial}{\partial t} (q_a - q_d) = - \frac{\partial q_d}{\partial t} = - \frac{\partial m}{\partial t} \tag{B8}
\]

Substituting (8) into (7) and integrating for the time step \( \Delta t \), we obtain evaporation efficiency \( \eta \) defined as follows:

\[
\eta = 1 - \frac{m_f}{m_i} = 1 - \exp \left[1.93 \times 10^{-6} N_0^{7/20} M^{13/20} \Delta t\right] \tag{B9}
\]

Subscripts \( i \) and \( f \) refer to initial and final values of water vapor density deficit, respectively. Gunn and Kinzer (1949) developed an expression for the terminal velocity of hydrometeors falling through air; comparison with observations showed it to be accurate within \( \pm 10\% \). Using the Marshall–Palmer (1948) drop-size distribution, Kessler (1969) derived a relation between precipitation intensity and water vapor density in air:

\[
R = 138 N_0^{-1/8} M^{9/8} \exp(kz/2) \tag{B10}
\]

where \( R \) is the precipitation intensity in millimeters per hour, \( N_0 \) is a constant of the Marshall and Palmer raindrop size distribution and equal to \( 10^7 \), \( M \) is the mass of water in raindrops in \( \text{gm m}^{-3} \), \( k \) is a constant that affects the terminal velocity and is equal to \( 10^{-4} \text{ m s}^{-1} \), and \( z \) is the height of raindrops above sea-level in meters.

The evaporation efficiency obtained from (9) and (10) for different rainfall rates at different heights for the typical \( 1/2 \) h period are shown in Fig. B1. Convective rain is considered differently from large-scale rain because it is nonuniform in intensity, particularly, for a typical \( 4^\circ \times 5^\circ \) grid box of a GCM. We consider only an ensemble mean. To evaporate rain realistically, information is needed about the rainfall intensity distribution. We have used the rainfall intensity distribution given by Ruprecht and Gray (1976) derived from observations over the western Pacific region as typical of convective rain anywhere (Fig. B2). For this rainfall

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**Fig. B1:** Evaporation efficiency for different rain amounts for large-scale rain.

**Fig. B2:** Distribution of rain intensity with area. Ruprecht and Gray (1976).

**Fig. B3:** Evaporation efficiency for different rain amounts for convective rain.
intensity distribution, the evaporation efficiency for convective rain as a function of rainfall amount and intensity is shown in Fig. B3. The overall effect of including this rain–evaporation parameterization is to allow the rain to evaporate into the intervening air in a realistic way as opposed to the crude assumption of no evaporation of convective rain and maximum possible evaporation of large-scale rain. The ultimate answer depends upon our ability to obtain fractional area of the convective clouds, but such a parameterization does not exist at this stage.

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