The diagnostic cloud parameterization scheme

1. Introduction.

This document describes a diagnostic cloud scheme module of the GFDL Flexible Modeling System (FMS) model. The categorization “diagnostic” indicates that cloud water is not a prognostic variable. The vertical distribution of the areal cloud fraction is diagnosed from thermodynamical and (to a lesser extent) dynamical model variables, while the variations in diagnosed cloud fraction and cloud optical depth affect the radiation field, and in turn, the thermodynamical and dynamical fields.

The FMS version of our diagnostic cloud scheme has its roots in an empirical scheme, implemented by Gordon (1992) into the GFDL-EP spectral atmospheric general circulation model (AGCM). However, it resembles, even more closely, a pre-FMS version that was recently employed in a AMIP 2 AGCM simulation and in ocean-atmosphere coupled model integrations.

The Gordon (1992) scheme consists of three components, i.e., (i) the diagnostic cloud fraction; (ii) the treatment of cloud optical depth; and (iii) anchoring of the cloud radiative properties to the cloud optical depth. The first is *diagnostic cloud fraction*. Most of the algorithms for computing the diagnostic cloud fraction in Gordon (1992) are based upon the empirical formulation of Slingo (1987), although some details of the two schemes are different. Global mean optical depths are specified for low, middle and anvil cirrus clouds, while an empirical temperature-dependent formula of Harshvardhan, et al. (1989) is used to calculate the cloud optical depth of non-anvil cirrus clouds. Spectral cloud absorptivities and reflectivities are computed from the cloud optical depths, utilizing algorithms obtained from V. Ramaswamy (1987, personal communication). Longwave emissivity is calculated from the *effective* cloud liquid water and cloud ice water paths (since cloud water and ice concentration are not prognostic variables) and corresponding specified longwave specific absorption coefficients. The *effective* cloud liquid and ice water paths are computed from the cloud optical depths and specified shortwave specific extinction coefficients for ice particles and liquid cloud droplets, assuming that the ratio of cloud liquid to total water path varies linearly with temperature.

The pre-FMS version has evolved from the Gordon (1992) diagnostic cloud scheme by (i) modifying the power law relationship (linear vs. quadratic) between the cloud fraction and the relative humidity and the value of the threshold relative humidity, RH_c; (ii) modifying the parameterization of the marine stratocumulus (MSc) cloud fraction; and (iii) adding an option to calculate single scattering albedo, ω_o, as a piecewise continuous linear function of zonal mean saturation mixing ratio at the model's lowest vertical level, as a crude means of simulating anomalous cloud absorption. In turn, the FMS version of the diagnostic cloud scheme differs from the most recent pre-FMS versions in the following respects: (1) The source code is written as a Fortran 90 module in FMS; (2) Some of the diagnostic cloud algorithms have been modestly revised in order to make them more scalable with vertical resolution; (3) While the FMS algorithms which compute cloud reflectivity and absorptivity are the same as in Gordon (1992), the calculations are now performed in a separate Fortran 90 module coded by Stephen Klein (personal communication, 1999). Also, the computed cloud fractions and cloud-radiative properties are input into the FMS radiative transfer module.

2. FMS Diagnostic Cloud Scheme Algorithms.

The salient features of the FMS diagnostic cloud scheme are now described in greater detail. Two basic sets of calculations are performed for each horizontal grid box: (i) *The areal cloud fraction,* i.e., the fraction of the grid box containing cloud, is calculated at each radiation time step for various cloud types and vertical levels within the vertical column; later, this information is merged to produce a vertical profile of distinct cloud layers and associated areal cloud fractions; (ii) *The cloud optical depth* of each distinct cloud layer is calculated as a function of cloud type and a crude measure of the cloud layer pressure thickness; cloud optical depths of non-anvil cirrus also depend upon the mean temperature of the cloud layer.

The FMS diagnostic cloud scheme is invoked each radiation time step. Generally speaking, either time mean or instantaneous GCM prognostic and diagnostic variables may be input as cloud predictors. We recommend using time mean cloud predictor variables, that have been averaged over the time interval between radiation time steps. Four cloud types found in Slingo (1987), i.e., high, middle and low stratiform clouds, and precipitating convective clouds as well as a fifth cloud type, i.e., shallow convective (or trade wind cumulus) clouds are permitted. Low clouds are further broken down into two subtypes as in Slingo (1987), i.e., (a) those associated with the large-scale ascent of moist air in synoptic weather systems; and (b) marine stratocumulus clouds associated with a temperature inversion. Note that any of the cloud types mentioned above may be switched off.

The distinction between “low”, “middle” and “high” model vertical levels for layer clouds is specified a priori. The low/middle cloud and middle/high cloud boundaries vary with latitude and season, consistent with the climatological tops and bases of high, middle and low clouds that would be specified, if the climatological zonal mean clouds option of FMS were chosen.

The vertical distribution of areal cloud fraction is calculated at full vertical levels for all five cloud types. Their occurrence is restricted to a subset of the full vertical domain. For example, a lower bound on pressure is imposed, to prevent high clouds from occurring very far into the stratosphere. The current default is 50 h Pa. Second, the occurrence of fog, i.e., stratiform clouds very close to the earth’s surface is optional. If fog is excluded, an upper bound on pressure (950 hPa is recommended) is imposed for stratiform and (shallow or deep) convective clouds alike. Conversely, if fog is allowed, stratiform cloud may form all the way down to the model’s lowest full vertical level, while (shallow or deep) convective clouds are excluded only from the lowest vertical level. Markers that indicate the cloud type(s) associated with the cloud fraction(s) *at each vertical level* are stored. As explained later, distinct high, and/or middle and/or low *stratiform* cloud layers, (if they exist), as well as the corresponding cloud fractions, vertical placement and thickness of these layers may be identified from the vertical distribution of stratiform cloud fractions and cloud type markers. Also, cloud type markers associated with each distinct stratiform cloud layer may be generated from this information and stored.

Next, stratiform and (deep and shallow) convective cloud layers are merged. By definition, distinct cloud layers must either be separated by a gap in the vertical or by a change in cloud type from stratiform to convective or vice versa. Thus, distinct cloud layers at adjacent model levels may survive. The current radiative transfer algorithms assume that the cloud fractions associated with all distinct cloud layers are randomly overlapped in the vertical. Initially, it is assumed that each high stratiform cloud layer contains non-anvil cirrus. However, if certain condensation criteria are met, its status is elevated to anvil or even super-anvil cirrus. This categorization affects only the optical depth of the cloud layer, and not the cloud fraction.

Finally, diagnostics are computed such as the low, middle, high and total cloud fractions, and the cloud layer true pressure thicknesses, mass-weighted mean temperatures and vertically integrated water vapor mixing ratios. The cloud layer mass-weighted mean temperature of non-anvil cirrus are used, later on, to calculate their cloud optical depths.

a. Areal cloud fraction algorithms.

Cloud fraction is diagnosed by 15 equations in all. Many of the equations resemble the empirically-based formulae of Slingo (1987) in form, and many of the empirical constants are the same as in Gordon (1992) and Slingo (1987). However, differences will be highlighted.

Relative humidity, RH, is the sole predictor of the cloud fraction of high and middle stratiform clouds and the primary predictor of the cloud fraction for the synoptic subclass of low stratiform clouds over the range over the range $RH_c \leq RH \leq 1$, while the vertical pressure velocity, ω, serves as an auxiliary predictor for the latter subclass of low clouds.The optimal relative humidity-cloud fraction relationship and threshold relative humidity RH_c were found to be rather sensitive to the cumulus parameterization scheme, and thus appear to be model-dependent. In addition, Xu and Krueger (1991) have argued the optimal value of RH_c may be height dependent. A linear relationship and $RH_c = 70\%$ for stratiform cloud fraction worked well in conjunction with moist convective adjustment (MCA) in Gordon (1992). In contrast, a quadratic relationship and $RH_c = 80\%$ was better suited to the relaxed Arakawa-Schubert (RAS) cumulus parameterization, now employed, because RAS generates a moister troposphere than MCA. By allowing RH_c, to vary modestly in the vertical, a further improvement in the *global* *mean* radiative flux balance at the top of the atmosphere, i.e., to within ~5 W m^{-2}, (which is acceptable for our purposes), was achieved. However, the simulated top of the atmosphere (TOA) climatological monthly mean earth radiation budget does contain some significant regional systematic errors. In contrast, an AMIP 2 atmosphere-only integration with the most recent pre-FMS version of the GFDL-EP model, (where EP denotes experimental prediction), simulates the interannual variability of the tropical TOA earth radiation budget quite well, when verified against ERBE and CERES data. In any event, the radiative tuning exercise has led to the current default settings of RH_c, i.e., 0.80 and 0.84, respectively, for $p < 750$ h Pa and $p \geq 750$ h Pa, where p is the pressure.

The high cloud fraction, n_h, is formally given by

$$n_h = \begin{Bmatrix} 0.0 \\ (RH - RH_c)^2/(1.00 - RH_c)^2 \\ 1.0 \end{Bmatrix}, \qquad \text{if} \begin{Bmatrix} RH < RH_c \\ RH_c \leq RH \leq 1.0 \\ RH > 1.0 \end{Bmatrix}, \tag{1}$$

While the FMS diagnostic cloud scheme recognizes anvil and super anvil cirrus as special categories of high clouds, they are treated the same as other high clouds in terms of cloud fraction. What differs is their cloud optical properties. The classification of high clouds as non-anvil cirrus, anvil cirrus or super anvil cirrus is deferred until the vertical distribution of cloud fractions is transformed into distinct stratiform cloud layers.

Similarly, the middle cloud fraction, n_m, is given by

$$n_m = \begin{Bmatrix} 0.0 \\ (RH_e - RH_c)^2/(1.00 - RH_c)^2 \\ 1.0 \end{Bmatrix}, \qquad \text{if} \begin{Bmatrix} RH_e < RH_c \\ RH_c \leq RH_e \leq 1.0 \\ RH_e > 1.0 \end{Bmatrix}, \tag{2}$$

where RH_e = RH is the default environmental relative humidity. Alternatively, the RH may be scaled by a drying factor outside of the convective cloud. Then, the environmental relative humidity is given by

$$RH_e = RH(1.00 - n_{cnv}) \tag{3}$$

where n_{cnv} is the convective cloud fraction. While the diagnostic scheme for n_{cnv} is discussed later, it must be invoked, in practice, before Eq. (3) can be solved.

The low cloud fraction, $n_{l1,}$ for the synoptic subclass of low stratiform clouds, is expressed as a product of two linear, piecewise continuous ramp functions:

$$n_{l1} = A(RH_e)B(\omega) \ . \tag{4}$$

where "A" depends on relative humidity only and "B" on the pressure vertical velocity, ω, only.

$$A(RH_e) = \begin{Bmatrix} 0.0 \\ (RH_e - RH_c)^2/(1.00 - RH_c)^2 \\ 1.0 \end{Bmatrix}, \qquad \text{if} \begin{Bmatrix} RH_e < RH_c \\ RH_c \le RH_e \le 1.0 \\ RH_e > 1.0 \end{Bmatrix}, \tag{5}$$

$$B(\omega) = \begin{Bmatrix} 1 \\ (\omega - \omega_0)/(\omega_1 - \omega_0) \\ 0 \end{Bmatrix}, \qquad \text{if} \begin{Bmatrix} \omega < \omega_1 \\ \omega_1 \le \omega \le \omega_0 \\ \omega > \omega_0 \end{Bmatrix} \qquad , \tag{6}$$

The default settings are $\omega_1 = 0$ and $\omega_0 = 3.6$ hPa h^{-1}. n_{l1} is diminished by weak vertical descent and vanishes for stronger vertical descent. The constraints of ω on reducing the low cloud fraction have been made less stringent in the FMS version than in Gordon (1992), in order to increase the low cloud fraction in the North Pacific and North Atlantic during boreal summer, and in turn, reduce the warm SST summer bias in these regions. On the other hand, at tropical latitudes, the low cloud fraction, and hence negative bias of net absorbed shortwave radiative flux increases slightly over the oceans, outside of the MS_c regions, and in the zonal mean.

The parameterization of marine stratocumulus and stratus clouds, hereafter referred to collectively as MSc clouds, has been modified since Gordon (1992), in an effort to improve its performance. Gordon (1992) employed the linear regression relationship for the most stably stratified lower tropospheric layer, which Slingo (1987) derived for the ECMWF AGCM, utilizing GARP Atlantic Tropical Experiment (GATE)) observed cloud data. Hereafter we shall refer to this as the S scheme. More recently, we have incorporated an approach similar to Philander, et al. (1996), (designated as the PL scheme) whereby a linear regression relationship is derived between *our AGCM's simulated* monthly vertical mean (from $\sigma = .844$ to $\sigma = 1.0$) lower tropospheric dry static stability vs. the observed monthly mean ISCCP low cloud fraction. The vertical mean static stability generates less negative feedback, i.e., destabilization due to cloud top longwave cooling, while the ISCCP data captures MSc clouds in the southeastern tropical Pacific off the west coast of Peru. Generally speaking, the PL scheme simulated subtropical MSc clouds somewhat better than the S scheme in integrations of the most recent pre-FMS version of the GFDL-EP AGCM. However, in simulations by the ocean-atmosphere coupled version of this model, a warm SST drift in the southeastern tropical Pacific develops rather rapidly, rendering even the PL scheme ineffective in that region within a few months.

In practice, various combinations of the PL and S schemes are applied, as discussed later. The principal equation governing n_{12} in the PL scheme may be written as

$$n_{12}(\mathrm{k}^*) = H_{12}(\mathrm{k}^*)\left(\overline{c_l} + \frac{(\overline{c'_l \Delta T'})}{(\overline{\Delta T'^2})} \cdot \Delta T'\right) \tag{7}$$

In Eq. (7), $\Delta T = \Delta T(k_m) = T(k_m) - T_{SFC}$, where $T(k_m)$ is the model-simulated temperature at pre-specified vertical level k_m, whose default value is the layer nearest to 850 h Pa. $\Delta T(k_m)$ is a measure of the local vertical mean dry static stability in the lower troposphere. Also, c_l is the ISCCP monthly mean low cloud fraction. It is derived from the ISCCP D2 monthly mean data, by subtracting off the middle and high monthly mean ISCCP cloud fractions from the ISCCP total cloud fraction. As explained in Gordon, et al. (2000), the ISCCP archive of monthly mean low, middle and high cloud fractions are derived only from satellite infrared radiance data. However, visible radiance data is a much better discriminator of marine stratocumulus clouds, and our procedure incorporates this information. Referring again to Eq. (7), the overbar denotes the long term annual mean climatology of c_l or $\Delta T(k_m)$, while the prime notation signifies climatological monthly mean departures from the respective long term annual means. Note that the cloud layer is placed at model level k = k*, where the lower tropospheric dry tropospheric static stability $-(\partial\theta/\partial p)_{k-1/2}$ is maximum. Thus, in Eq. (7), the Heavaside function factor

$$H_{l2} = \begin{Bmatrix} 1.0 \\ 0.0 \end{Bmatrix}, \qquad \text{if} \begin{Bmatrix} \left(-\frac{\partial\theta}{\partial p}\right)_{k^*\text{-}1/2} > \left(-\frac{\partial\theta}{\partial p}\right)_{crit} \\ -\left(-\frac{\partial\theta}{\partial p}\right)_{k^*\text{-}1/2} \le \left(-\frac{\partial\theta}{\partial p}\right)_{crit} \end{Bmatrix}. \tag{8}$$

serves as an auxiliary constraint on the existence of MSc calculated by Eq. (7). In Eq. (8), θ is the potential temperature, p is the pressure, $\partial\theta/\partial p$ is the lapse rate (K hPa^{-1}) in the most stable layer below 750 h Pa, and $(-\partial\theta/\partial p)_{crit}$ is a critical threshold value of dry static stability. Its default value, 0.0181 K hPa^{-1}, corresponds to a considerably weaker stable stratification than for an inversion. A physical realizability constraint is here, as well as for all other cloud types, e.g., $0 \le n_{12} \le 1$. Also, a few other auxiliary constraints, that affect only whether or not MSc exist, must be satisfied in order for Eqs. (7) and (8) to be applied. (i) Eqs. (7) and (8) are applied only within the latitude band $|\varphi| \le |\varphi_{crit}|$, straddling the equator. The rationale is that the ISCCP algorithms have some trouble distinguishing cloud from ice or snow. The default value of $|\varphi_{crit}|$ is 60^o. (ii) The scheme is applied only over open water, i.e., both land and sea ice grid points are excluded. (iii) It is applied only at grid points where the correlation coefficient, $r(c_l,\Delta T)$ exceeds some threshold value, r_{crit}. Here, $r(c_l,\Delta T)$ represents the correlation between departures of the observed ISCCP low cloud fraction and model-simulated vertically averaged static stability from their respective long term climatological annual mean values. The default value of r_{crit} is 0.40. (iv) An optional constraint, $\omega(k=k_m) < \omega_{crit}$, may be applied as well. Here, the pressure vertical velocity, ω, is an auxiliary predictor and the subscript "crit" signifies a threshold value. For example, the constraint $\omega_{crit} = 0$ acknowledges the tendency for observed subtropical marine stratocumulus to occur where subsidence exists above an inversion layer. However, its use may introduce noise, since the vertical pressure velocity is typically a 2-hour time mean value. The constraint on ω is effectively deactivated by choosing the default value of ω_{crit}, i.e., minus infinity. (v) The final constraint, $\Delta T(k=k_m) \ge \Delta T_{crit}(k_m)$, might be beneficial if the scheme tends to generate MSc clouds in highly convective, relatively unstable regions. Whether or not this is a problem may be model-dependent. Currently, we effectively de-activate this constraint by setting $\Delta T_{crit}(k_m)$ equal to its default value of -100°K.

Alternatively, instead of employing Eqs. (7) and (8) plus the above auxiliary constraints, the MSc cloud fraction found may be computed from Eqs. (9), (10) and (11) below. Conceptually, this latter algorithm resembles the one formulated by Slingo (1987) and employed by Gordon (1992), more closely.

$$n_{12}(k^*) = S\left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} \cdot B(RH_{base}) \tag{9}$$

$$S\left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} = \left\{\begin{matrix} 1.0 \\ \alpha\left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} + \beta \\ 0.0 \end{matrix}\right\}, \text{ if } \left\{\begin{matrix} \left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} > 0.10 \\ 0.05 \le \left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} \le 0.10 \\ \left(-\frac{\partial\theta}{\partial p}\right)_{k^*-\frac{1}{2}} < 0.05 \end{matrix}\right\} \tag{10}$$

$$B(RH_{base}) = \left\{\begin{matrix} 1 \\ (RH_{base} - RH_{min})/(DRH) \\ 0 \end{matrix}\right\}, \text{ if } \left\{\begin{matrix} RH_{base} > RH_{min} \\ RH_{min} \le RH_{base} \le RH_{max} \\ RH_{base} < RH_{min} \end{matrix}\right\} \tag{11}$$

Above, the parameter values $\alpha = 2000.0$ hPa K^{-1} and $\beta = -1.0$ are determined by the conditions that the cloud fraction be 0 and 1 at the two transition points of the piecewise continuous ramp function in Eq. (10), i.e., at $(-\partial\theta/\partial p) = 0.05$ and $(-\partial\theta/\partial p) = 0.10$ K hPa^{-1}. The term *quasi-inversion* layer connotes a fairly highly stratified layer, but not necessarily sufficiently stratified to be characterized as a true inversion layer. The stratification criteria for MSc clouds in Eq. (10) is less stringent than in Gordon(1992), where α = 667.0 hPa K^{-1} and β = -0.667, but considerably more stringent than in Eq. (8). Meanwhile, in Eq. (11), RH_{base} is the relative humidity at the base of the highest contiguous *quasi-inversion* layer in the lower troposphere (i.e., beneath 750 h Pa), The cloud fraction is reduced for base relative humidities below RH_{max} and disabled for base relative humidities less than RH_{min}, and DRH = RH_{max} - RH_{min}. The parameters RH_{max} and RH_{min} assume two sets of values, one for ocean grid points and one for land grid points. The default values over the ocean, i.e., $RH_{max} = RH^{O}_{max} = 0.80$ and $RH_{min} = RH^{O}_{min} = 0.60$, (and hence DRH = DRH^{O} = 0.2) are the same as in Gordon (1992). The same value may be specified over land. The rationale for the ramp function B is to prevent stratocumulus or stratus clouds from forming under extremely dry conditions, e.g., over deserts. However, the scheme may be effectively disabled over land, i.e., restricted to marine stratocumulus and stratus clouds, by specifying the default values $RH_{max} = RH^{L}_{max} = 1.0000$ and $RH_{min} = RH^{L}_{min} = 0.999999$.

As previously mentioned, four options are available for calculating the fraction of MSc / stratus clouds. They are: (i) apply only Eqs. (7) and (8); (ii) apply only Eqs. (9), (10) and (11); (iii) apply both sets of equations, but restrict the application of Eqs. (9), (10) and (11) exclusively to latitudes where Eqs. (7) and (8) are never applied, i.e., generally to relatively high latitudes; or (iv) apply both, but at latitudes where Eqs. (7) and (8) are applied, restrict the application of Eqs. (9), (10) and (11) to vertical columns where Eqs. (7) and (8) yield $n_{12}(k^*) = 0$:

After calculating the vertical distribution of n_{l1} and n_{l2}, utilizing Eqs. (3) through (10), the dominant cloud fraction at each model vertical level in the lower troposphere takes precedence.

$$n_l = max\{n_{l1}, n_{l2}\} \tag{12}$$

An ad hoc diagnostic parameterization for shallow convective cloud fraction n_{shl} has been included. Presumably, the trade wind cumulus regime is more or less orthogonal to the synoptic and marine stratocumulus cloud regimes. The essence of the parameterization is described by Eqs. (13) and (14) below.

$$n_{shl} = C_{shl}A_{max}(RH_e) \tag{13}$$

Eq.(13) is evaluated within the lowest *contiguous,* conditionally unstable layer, $\partial\theta_e/\partial p \geq 0$, (where, θ_e is the equivalent potential temperature and p the pressure), that lies at or above the lifting condensation level, (LCL). ($\partial\theta_e/\partial p \geq 0$ is an approximation to the positive buoyancy condition), Thus, the base of the shallow convective layer is explicitly tied to the LCL, in the FMS scheme in contrast to Gordon (1992), However, the auxiliary lid constraint $p \geq p_{lid}$ (750 hPa is the current default value) usually still has to be invoked. Also, another auxiliary constraint, i.e., $\omega > \omega_1$, corresponding to large scale vertical ascent and (depending upon the value of ω_1) possibly weak vertical ascent, is simultaneously applied. The contiguous shallow convective layer that finally emerges typically straddles one or two vertical levels in the FMS model version with 18 σ levels. The first factor in Eq. (12) is defined by Eq. (14),

$$\left\{ C_{shl} = \begin{array}{l} C_0\text{, at vertical levels that satisfy all constraints} \\ 0\ \text{, at all other vertical levels} \end{array} \right\} \tag{14}$$

In order to compute $A_{max}(RH_e)$, Eq. (5) is first applied to the contiguous layer where $C_{shl} = C_0$, (whose current default value is 0.2), to obtain the vertical distribution of $A(RH_e)$. Then, its maximum, $A_{max}(RH_e)$, is determined and specified throughout this shallow convective layer, i.e., maximum overlap is applied.

Meanwhile, convective cloud fraction in the lower troposphere is given by

$$n_{cnv} = \left\{ \begin{array}{c} 0.0 \\ a + b\ ln\,\mathrm{P} \\ 0.8 \end{array} \right\}, \qquad \text{if} \left\{ \begin{array}{c} P < 0.14 \\ 0.14 \leq P \leq 85.0 \\ 85.0 < P \end{array} \right\}, \tag{15}$$

as in Gordon (1992), where b ~.125 and a ~ -.125 ln [0.14] are empirical constants and P is the model's convective precipitation rate reaching the surface in mm day^{-1}, calculated from the time mean value for the time interval between radiation time steps. Also, Eq. (15) is scaled by 0.25 in the middle and upper troposphere to represent idealized penetrative convective towers. The base and top of the complete convective cloud layer are defined by the contiguous layer in the vertical column where convective precipitation has occurred during the time interval (currently 2 hours in our model) since the previous radiation time step. In practice, we have shut off this parameterization, in an effort to reduce the cumulative low cloud fraction in the tropics.

Eqs. (1) through (15), generate a vertical distribution of cloud fractions of high, middle and low stratiform clouds as well as shallow convective clouds and deep convective clouds. Next, the vertical distributions of low, middle and high stratiform clouds are transformed into a set of distinct stratiform cloud layers by the multi-step process described below. These layers may not necessarily emerge completely in tact, after being merged, later, with the shallow and precipitating convective clouds.

The first step is to identify the three model levels k_h, k_m and k_l, in the upper, middle and lower troposphere, respectively, where the high, middle and low stratiform cloud fractions n_h, n_m and n_l attain their maximum values. Initially, the high, middle and low stratiform cloud layers are assumed to be only one model layer thick. Thus, the cloud fractions are set equal to those maximum values, n_h^{max}, n_m^{max}, and n_l^{max}, i.e.,

$$\left\{\begin{array}{c} n_h(k_h) \\ n_m(k_m) \\ n_l(k_l) \end{array}\right\} = \left\{\begin{array}{c} n_h^{max} \\ n_m^{max} \\ n_l^{max} \end{array}\right\} \tag{16}$$

at levels k_h, k_m and k_l, but equal to zero at all other levels, while the top and base of each cloud layer coincide, and are described by the same indices, i.e., k_h, k_m or k_l, respectively. Moreover, the dominant cloud type markers at model levels k_h, k_m and k_l, are assigned to the respective cloud layers. Of course, fewer cloud layers will be identified if the high and/or middle and/or low stratiform cloud fractions are zero.

The second step is to search for thicker *contiguous* cloud layers straddling model levels k_h,

k_m and k_l, respectively, in which the reductions in cloud fraction from the maximum values do not exceed specified values. Cloud layers that straddle more than one model vertical level may facilitate scalability to finer vertical resolutions. Mathematically, the criteria are:

$$\left\{ \begin{array}{c} n_h(\mathrm{k}) \\ n_m(\mathrm{k}) \\ n_l(\mathrm{k}) \end{array} \right\} \geq \left\{ \begin{array}{ccc} n_h^{max} & - & n_h^{crit} \\ n_m^{max} & - & n_m^{crit} \\ n_l^{max} & - & n_l^{crit} \end{array} \right\} \tag{17}$$

where, n_h^{crit}, n_m^{crit} and n_l^{crit} are the respective high, middle and low cloud fraction threshold values. The search is terminated in each direction, i.e., upwards or downwards as soon as Eq. (17) is violated. Currently, in our 18 level FMS model, $n_h^{crit} = n_m^{crit} = 0.0001$ and $n_l^{crit} = 0.25$. Despite the small value of n_h^{crit}, thick high cloud layers frequently occur in the model tropics under near-saturated conditions. Also, the search is confined to a single class of model levels, i.e., high, middle or low, and is restricted by the appropriate maximum pressure thickness parameter Δp_h^{max}, Δp_m^{max} or Δp_l^{max}. Currently, $\Delta p_h^{max} = \Delta p_m^{max} = 200$ hPa for high and middle clouds, while Δp_l^{max} = only 100 hPa for low clouds. In the 18 level model, the search is effectively restricted to two levels above or below k_h and k_m. The number of low cloud levels affected is more variable, due to the tighter spacing of vertical levels in the planetary boundary layer. When searching downward for thick low stratiform clouds layers, the lowest model level is excluded if fog is permitted; and if fog is excluded, so are all model levels in the fog layer. The mass-weighted vertical mean cloud fraction for each cloud layer may be affected by the merging of stratiform and convective clouds. Therefore, this calculation is deferred until later.

The high stratiform cloud layer classification is refined into non-anvil, i.e., ordinary cirrus, anvil cirrus and super anvil cirrus, viz. the cloud layer type marker. As mentioned earlier, the distinguishing characteristic of these three subclasses is their cloud optical properties, whereas their cloud fraction is unaffected. The scheme checks first for super anvil cirrus within each vertical column. For this subclass of high cloud to exist, an extensive, contiguous portion of the middle and upper tropospheric column must have experienced convective latent heat release at least once during the time interval (currently 2 hours) between radiation (and diagnostic cloud) calculations. This is monitored by a “convective counter”. Two parameters, a convective base parameter cnv_B and a convective top parameter cnv_T whose current values are 0.55 and 0.225, respectively, in sigma coordinate space, are specified. The “convective counter” must register a “hit” for all model levels which lie between cnv_B and the lesser (i.e., higher in z coordinate space) of cnv_T and the actual cloud top of the high cloud layer. If the criterion for super anvil cirrus is not met, two less stringent criteria for anvil cirrus are analyzed. It is sufficient that only one of them be satisifed.The first is that the “convective counter” has been activated at least once since the previous radiation time step, at one or more vertical levels within the high cloud layer. The second is that the large scale condensation rate which has occurred within this high cloud layer since the last radiation time step exceeds a minimum critical threshold, (currently 0.50 mm day^{-1}). If neither of the above two criteria for anvil cirrus are satisfied, then the ordinary, i.e., non-anvil cirrus classification of the high cloud layer and hence the original high cloud layer marker index are retained.

Next, some overlap assumptions are imposed upon cloud fractions at each “low” model level where low stratiform, and/or shallow convective and/or precipitating convective clouds simultaneously exist. Additionally, the cloud type associated with the dominant cloud fraction amongst n_l, n_{shl} and n_{cnv} becomes the cloud type marker there. More specifically, n_l is reset to zero, if $n_l < n_{shl}$, and only the shallow convective cloud type marker is carried forward. Otherwise, n_l and n_{cnv} are randomly overlapped. In this latter case, the low stratiform cloud type marker is carried forward if $n_l \geq n_{cnv}$, while the precipitating convective cloud type marker is carried forward if $n_l < n_{cnv}$. Next, at “low” levels where shallow and precipitating convective cloud types still co-exist, the maximum overlap condition is applied. In other words, if $n_{shl} > n_{cnv}$, the cloud fraction is set to n_{shl} and the shallow convective cloud type marker is carried forward; conversely, if $n_{shl} \leq n_{cnv}$, the cloud fraction is set to n_{cnv} and the precipitating convective cloud type is carried forward. Convective cloud fractions are also randomly overlapped with stratiform middle and high cloud levels. However the middle or high cloud types are re-classified as convective only if $n_{cnv} \geq n_m$ or $n_{cnv} \geq n_{h,}$ respectively.

Sufficient information now exists to identify each distinct cloud layer within the vertical column, as well as the top, base and mass-weighted, vertical mean cloud fraction of that layer, and the total number of distinct cloud layers. A distinct cloud layer may be either stratiform or convective. A distinct stratiform cloud layer may encompass low and middle cloud or middle and high cloud at adjacent model levels. The model level corresponding to the top of a distinct cloud layer must meet the following conditions: either the model level immediately above is cloud free, or the cloud type marker at adjacent model levels changes from stratiform to convective (either shallow or precipitating) or vice versa. Similarly, for a model level to correspond to the base of a distinct cloud layer, the model level immediately below should be cloud free, or the cloud type marker at adjacent model levels should change from convective (either shallow or precipitating) to stratiform or vice versa. Additionally, if the fog option is turned on, the base of any distinct cloud layer which is not classified as fog can extend no lower than the second model level above the earth's surface. Of course, if the fog option is turned off, the cloud base cannot extend into the restricted "fog layer" zone.

Once, the cloud base and cloud top of each distinct cloud layer are determined, the mass-weighted, vertical mean cloud fraction of the layer is calculated and then assigned to all model levels between cloud base and cloud top. Also, a final cloud type marker is assigned to each distinct cloud layer. An adjustment, which could affect the optical properties of the cloud layer, is performed when a distinct stratiform cloud layer straddles adjacent "low" and "middle" or "middle" and "high" model levels. In that case, the cloud type marker designation at cloud base is carried forward. For example, for a stratiform cloud whose base and top are located at a "low" and "middle" model level, respectively, would be classified as a low cloud. The cloud marker would retain information regarding the type of low cloud, e.g., synoptic or MSc. Lastly, the following three diagnostic variables are computed: the mass-weighted cloud layer temperature (which coincides with the cloud top temperature for cloud layers that are one layer thick), the true pressure thickness of the cloud layer and the cumulative water vapor mixing ratio within the cloud layer. The latter variable could potentially affect the cloud optical depth in a future version of the diagnostic cloud scheme.

b. parameterization of cloud optical depth

An approach similar to the one proposed by Harshvardhan et al., (1989) is used to specify / parameterize cloud optical depths. A distinction is made between "cold", "warm" and "cool" high clouds. The FMS version is virtually identical to that employed in Gordon (1992), except that more options for "cold" high clouds are allowed, while "cold" middle or low clouds are excluded. Non-anvil cirrus clouds whose cloud layer mean temperature T_c satisfies the condition $T_c < -10$ C are treated as "cold" clouds. If the anvil/super anvil cirrus option is turned off in the FMS scheme, all high clouds which satisfy $T_c < -10$ C are treated as "cold" non-anvil cirrus clouds, in the calculation of their optical depth. On the other hand, if the cold cloud option is turned off, all high clouds are treated like "warm" anvil cirrus clouds, regardless of T_c,and are assigned the appropriate cloud optical depth. In the most general configuration, wherein both options are turned on, high clouds which satisfy the requirements for anvil or super anvil cirrus are treated as "warm", even if $T_c < 10$ C, while the remainder, i.e., non-anvil cirrus are treated as "warm", "cold" or "cool" (-10 C $< T_c < 0$ C) cirrus, depending upon T_c.

Visible optical depths of "cold" non-anvil cirrus high clouds ($T_c < -10$ C) vary quadratically with the departure of T_c from a very cold reference value T_{c0}:

$$\tau_{sw} = A\ (T_c - T_{c0})^2, \quad T_{c0} < T_c < -10\mathrm{C}, \tag{18}$$

where, $T_{c0} = -82.5$ K. Equation (18) is adapted from Harshvardhan et al., (1989), who based their formulation for cold clouds on the empirical results of Platt and Harshvardhan (1988). But our coefficient $A = 4 \times 10^{-4}$ exceeds the value recommended by Harshvardhan et al., (1989) by approximately a factor of three. Even so, $\tau_{sw} < 0.5$ for many "cold", high clouds. The optical depths of "cool" non-anvil cirrus are obtained by linearly interpolating between the values for "cold" non-anvil clouds with T_c = -10 C and warm cirrus clouds. For this class of clouds to exist, the cold cloud option must be turned on and -10 C < T_c < 0 C.

Distinct fixed values of visible cloud optical depth, τ_{cld}, are specified for various types of low, middle, and high (excluding cold or cool non-anvil cirrus) stratiform clouds, as well as shallow convective and precipitating convective clouds. τ_{cld} is expressed as the product B Δp, where a hypothetical pressure thickness, Δp, is specified in the calculation of τ_{cld}. The current default values of τ_{cld}, B and Δp are listed below in Table 1 as a function of cloud type.

Table 1. Default Values of Specified Fixed Optical Depths for Various Cloud Types.

Cloud Type	τ_{cld}	B	Δp
warm, non-anvil cirrus	2.5	B_{std}	Δp_h
warm, thick non-anvil cirrus	5.0	B_{std}	$2\Delta p_h$
anvil cirrus	5.0	$2B_{std}$	Δp_h
super anvil cirrus	5.0	$2B_{std}$	Δp_h
middle	6.0	B_{std}	Δp_m
low synoptic, with $B(\omega) = 1$	9.0	B_{std}	Δp_l^*
low, synoptic, with $B(\omega) \le 1$	9.0	B_{std}	Δp_l^*
MS_c	9.0	B_{std}	Δp_l^*
shallow convective	9.0	B_{std}	Δp_l^*
precipitating convective with high cloud base	5.0	$2B_{std}$	Δp_h
precipitating convective with middle cloud base	12.0	$2B_{std}$	Δp_m
precipitating convective with low cloud base	18.0	$2B_{std}$	Δp_l

In Table 1, $\Delta p_h = 31.25$ hPa, $\Delta p_m = 75$ hPa, and $\Delta p_l = 112.5$ are prescribed pressure thicknesses of (warm, non-anvil cirrus) high clouds, middle clouds, and low clouds, and following Harshvardhan, et al. (1989), the coefficient for warm, non-anvil cirrus clouds is $B_{std} = 0.08$ hPa^{-1}. Also, $\Delta p_l^* = \Delta p_l$, if $p(k_T - 1/2) \leq \Delta p_l$, $\Delta p_l^* = \max\{1.0 - p(k_T - 1/2), \Delta p_{min}\}$, if $p(k_T - 1/2) > \Delta p_l$, where $p(k_T - 1/2)$ is the cloud top pressure at model half level $k_T - 1/2$ and $\Delta p_{min} = 25$ hPa is the minimum prescribed low cloud pressure thickness in our 18 level model. Thus, an adjustment is made for "thin" low clouds whose tops lie close to the earth's surface. More specifically, their prescribed thickness may not exceed their actual thickness, though a lower bound, i.e., Δp_{min} is imposed. For clouds straddling low/middle or middle/high atmospheric layers, the cloud optical depth assignment is determined by the base of the cloud layer. In a future version of the scheme, the cloud pressure thickness may be computed as a function of the precipitable water (vapor) within the cloud layer to improve scalability for finer vertical resolution.

c. shortwave and longwave cloud radiative properties

Stephens (1978) proposed a parameterization for GCM's which links the shortwave and longwave properties through their mutual dependence on the cloud liquid water path; and the approach was extended to ice clouds by Stephens and Webster (1981). We have adopted a variation of this approach, suggested by V. Ramaswamy (personal communication, 1987), which is quite convenient when cloud water is not a prognostic GCM variable. Namely, the cloud reflectivities (R_{vis} and R_{nir}) and absorptivities (A_{vis} and A_{nir}) for the visible (vis) and near infrared (nir) spectral bands are computed as functions of the shortwave optical depth τ_{SW} and zenith angle, and the longwave emissivity ε_{LW} as a function of an effective infrared absorption optical depth τ_{LW}. However, as previously mentioned, in the FMS, the spectral reflectances and spectral absorptivities are calculated outside of the diagnostic cloud scheme module.

In any case, given τ_{SW}, the liquid water fraction β, and the shortwave specific extinction K_{SWI} and K_{SWL}, for ice particles and liquid cloud droplets, the equation

$$\tau_{SW} = (K_{SWI})(IWP) + (K_{SWL})(LWP). \quad (19)$$

is used to diagnose the effective cloud ice water path (IWP) and cloud liquid water path (LWP). We assume that $\beta = 0$ for $T < 258k$ (pure ice phase), $\beta = 1$ for $T > 268k$ (pure liquid phase), and β varies linearly with temperature for $258k < T < 268k$ (mixed phase). The system is closed by computing the ratio IWP/LWP and eliminating the total cloud water path CWP from

$$LWP = \beta\ (CWP); \quad IWP = (1 - \beta)(CWP). \quad (20)$$

The near-infrared emissivity calculation

$$\varepsilon_{LW} = 1 - \exp\{-b\,\tau_{LW}\}, \quad (21)$$

is still performed within the diagnostic cloud module. Here,

$$\tau_{LW} = (K_{LWI}/b)(IWP) + (K_{LWL}/b)(LWP), \quad (22)$$

b is the diffusivity factor ($b = 1.66$), and K_{LWI}/b and K_{LWL}/b are the longwave specific absorption for ice particles and liquid droplets. We adopted the following values: $K_{SWI} = 74\ m^2\ kg^{-1}$, $K_{SWL} = 130\ m^2\ kg^{-1}$, $K_{LWI} = 100\ m^2\ kg$ and $K_{LWL} = 140\ m^2\ kg$. Note that the factor "b" in the numerator of Eq. (21) and denominator of Eq. (22) cancel.

In FMS, the constant asymmetry parameter is $g = 0.85$, (cf. $g = 0.80$ in Gordon (1992)), while the default single scattering albedo in the near IR is still the constant $\omega_{o,nir} = 0.99420$. Alternatively, if the anomalous cloud absorption option, (Gordon, et al.,2000), is invoked, $\omega_{o,nir}$ becomes a linear, piecewise continuous, monotonically decreasing function of zonal mean saturation water vapor mixing ratio, $<q_s>$, at the lowest model level. $\omega_{o,nir}$ attains its minimum value, 0.970, for $<q_s> \geq 2.0\ \xi\ 10^{-2}$. Cloud absorption is moderately enhanced in the Tropics by this parameterization, which has a modestly favorable impact upon the SST cold bias in the western tropical Pacific of the pre-FMS version of our coupled model. However, its physical basis is controversial.

References

Gordon, C. T., 1992: Comparison of 30-Day Integrations with and without Cloud-Radiation Interaction. *Mon. Wea. Rev.*, **120**, 1244-1275.

Gordon, C. T., A. Rosati and R. Gudgel, 2000: Tropical Sensitivity of a Coupled Model to Specified ISCCP Low Clouds. *J. Climate*, **13**, 2239-2260.

Harshvardhan, D. A. Randall, T. G. Corsetti, and D. A. Dazlich, 1989: Earth Radiation Budget and Cloudiness Simulations with a General Circulation Model. *J. Atmos. Sci.*, **46**, 1922-1942.

Philander, S. G. H., D. Gu, D. Halpern, G. Lambert, N. C. Lau, T. Li, and R. C. Pacanowski, 1996: Why the ITCZ is Mostly North of the Equator. *J. Climate*, **9**, 2958-2972.

Platt, C. M. R., and Harshvardhan, 1988: Temperature Dependence of Cirrus Extinction: Implications for Climate Feedback. *J. Geophys. Res.*, **93 (D9),** 11051-11058.

Slingo, J. M., 1987: The Development and Verification of a Cloud Prediction Scheme for the ECMWF Model. *Quart. J. Roy. Meteor. Soc.*, **113**,899-927.

Stephens, G. L., 1978: Radiation Profiles in Extended Water Clouds. II. Parameterization Schemes. *J. Atmos. Sci.*, **35**, 2123-2132.

Stephens, G. L. and P. J. Webster, 1981: Clouds and Climate: Sensitivity of Simple Systems. J*. Atmos. Sci.*, **38**, 235-247.

Xu, K-M., and S. K. Krueger, 1991: Evaluation of Cloudiness Parameterizations Using a Cumulus Ensemble Model. *Mon. Wea. Rev.*, **119,** 342-367.