A PROGNOSTIC CLOUD WATER PARAMETERIZATION FOR GLOBAL CLIMATE MODELS

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ABSTRACT

We describe an efficient new prognostic cloud water parameterization designed for use in global climate models. The scheme allows for life cycle effects in stratiform clouds and permits cloud optical properties to be determined interactively. The parameterization contains representations of all important microphysical processes, including autoconversion, accretion, Bergeron-Findeisen diffusional growth, and cloud/rain water evaporation. Small-scale dynamical processes, including detrainment of convective condensate, cloud top entrainment instability, and stability-dependent cloud physical thickness variations, are also taken into account. Cloud optical thickness is calculated from the predicted liquid/ice water path and a variable droplet effective radius estimated by assuming constant droplet number concentration. Microphysical and radiative properties are assumed to be different for liquid and ice clouds, and for liquid clouds over land and ocean.

The parameterization is validated in several simulations using the Goddard Institute for Space Studies (GISS) general circulation model (GCM). Comparisons are made with a variety of data sets, including ERBE radiative fluxes and cloud forcing, ISCCP and surface-observed cloud properties, SSM/I liquid water path, and SAGE II thin cirrus cover. Validation is judged both on the basis of the model's depiction of the mean state and of diurnal, seasonal and interannual variability and the temperature dependence of cloud properties. Relative to the diagnostic cloud scheme used in the previous GISS GCM, the prognostic parameterization strengthens the model's hydrologic cycle and general circulation, both directly and indirectly (via increased cumulus heating). Sea surface temperature (SST) perturbation experiments produce low climate sensitivity and slightly negative cloud feedback for globally uniform SST changes, but high sensitivity and positive cloud feedback when tropical Pacific SST gradients weaken with warming. Changes in the extent and optical thickness of tropical cumulus anvils appear to be the primary factor determining the sensitivity. This suggests that correct simulations of upward transport of convective condensate and of Walker circulation changes are of the highest priority for a realistic estimate of cloud feedback in actual greenhouse gas increase scenarios.

1.) Introduction

Almost all climate GCMs agree that total cloud cover should decrease while cloud height increases in response to a greenhouse gas-induced warming (Del Genio, 1993). This agreement is surprising, since there is no fundamental basis for predicting the sign of cloud cover changes. Despite this consensus, GCM estimates of cloud feedback range from strongly positive to weakly negative. Much of the disagreement can be traced to the models' differing representations of cloud radiative properties and the resulting optical thickness feedback. There are essentially three different approaches to the parameterization of cloud optical properties in GCMs:

(1) *Implicit*: Many GCMs prescribe cloud optical thickness (or shortwave albedo and long-wave emissivity) as a fixed function of altitude. Since cloud height increases with warming, this induces an implicit cloud optics feedback. The feedback is typically positive because optical thickness decreases with height in most models and the albedo effect dominates the greenhouse effect of clouds globally. But negative feedback examples also exist, e.g., in models that prescribe extensive, thick cumulus anvils (cf. Cess et al., 1990).

(2) *Diagnostic*: Several GCMs incorporate temperature-dependent optical thickness, based on the instantaneous condensation needed to eliminate supersaturation or the adiabatic liquid water content of a lifted cloud (Betts and Harshvardhan, 1987). Although this has been viewed as a step forward in parameterization, it is actually the worst possible approach. Such schemes parameterize the source, but none of the sinks, of cloud water and imply constant cloud physical thickness and particle size. Consequently, albedo systematically increases with warming, and as a result these models tend to be biased toward negative cloud optics feedback.

(3) *Prognostic*: The most recent trend in GCMs is to carry cloud water content as a prognostic variable, thus permitting storage of cloud water and life cycle effects as well as interactive optical properties (Sundqvist, 1978; Roeckner et al., 1987; Smith, 1990; LeTreut and Li, 1991; Tiedtke, 1993; Fowler et al., 1995). This approach is in principle the most physically realistic. But it requires the parameterization of complex microphysical, dynamic and radiative processes, thus introducing a number of degrees of freedom absent from the simpler approaches. Not surprisingly, GCMs with prognostic schemes can produce either positive, negative or nearly

neutral cloud feedback.

Because of the wide variety of processes that must be accounted for and the requirement for simple representations of these processes to permit long model integrations, prognostic cloud parameterizations cannot yet be considered superior to the simpler approaches in their ability to predict cloud feedback and climate sensitivity to perturbations. However, the implicit and diagnostic approaches are dead-end philosophies, with very limited potential to either improve in response to advances in understanding or to shed light on the physics of cloud feedback. Models with prognostic schemes can instead be regarded as laboratories that enable us to assess which cloud processes are most important to changing climate. Ideally, then, such models can not only exploit new information but can also guide the strategy for future observations and theoretical studies.

Problems that limit the performance of prognostic cloud parameterizations fall into five broad categories: (1) Poorly understood cloud processes (e.g., cloud top entrainment instability). (2) Inadequate observations of cloud properties for validation (e.g., ice water content). (3) Cloud physics that is computationally not feasible to simulate in GCMs (e.g., evolution of drop size distributions). (4) Physics that is understood on the cloud scale but not on the GCM grid scale (e.g., relationship of cloud cover to relative humidity). (5) Deficiencies in other parts of the GCM (e.g., surface and boundary layer fluxes and large-scale dynamical transports of moisture and heat). Consequently, it is impossible to construct a parameterization that does not contain arbitrary tuning parameters. This opens a Pandora's box of potential abuses that can give a parameterization the appearance of performing well while actually obscuring the issue of whether it contains the appropriate physics for its intended applications. For this reason we propose several "rules of the road" that should be satisfied by any parameterization used in climate models:

(1) Tuning parameters must not be arbitrary functions of latitude, altitude, etc. Ideally they should be functions of a model-predicted quantity based on known physics or at least empirical evidence, but in the absence of either, tuning parameters must be globally uniform constants. This reduces the possibility of misleading apparent validation of the scheme via a well-chosen comparison with a particular data set. In the long run, climate model development and observing

program design are better served by highlighting model-data discrepancies than by artificial agreement based on arbitrary assumptions.

(2) Validation of the scheme must be performed against multiple data sets, not only those selected to give favorable comparisons. For example, accurate simulation of top-of-the-atmosphere radiative fluxes is a necessary but not sufficient condition for validation. Since the fluxes are the composite effect of spatially and temporally varying cloud cover, cloud optical properties, water vapor, etc., and each of these will vary in a climate change, validation of each individual quantity must be conducted to ensure that accurate radiative fluxes are not simply the result of compensating errors in several parameters.

(3) For climate models, which are used to predict change, much of the validation must be done against variability (both sense and magnitude) of the current climate (diurnal cycle, seasonal cycle, ENSO) rather than simply against the mean state. This is the model analog to the issue of accuracy vs. precision for scientific instruments: The accuracy of the absolute value of an observed (simulated) parameter is often worse, but less important, than our ability to detect (predict) changes in that parameter to high precision. As a simple example, the cloud feedback exhibited by different GCMs in SST perturbation experiments bears no relationship to the models' cloud forcing for the current climate (Cess et al., 1990). The cloud feedback in turn is smaller than the accuracy of current radiative flux data sets, yet such data are still valuable constraints for GCMs because of the variability they document. Of course variability may be a function of the mean state itself, but if variability is directly validated against observations, errors in sensitivity caused by errors in the mean state will be evident. This is not to say that the mean state is unimportant; errors in the mean state will contribute, e.g., to climate drift in a coupled atmosphere-ocean model. Thus, both the mean state and variability must be examined.

With these considerations in mind, in this paper we describe a new prognostic cloud water budget parameterization that has been implemented in the GISS GCM. The details of the scheme are presented in Section 2. Section 3 describes the model's mean state, spatial variability, and the effect of the cloud scheme on the general circulation. Section 4 documents its simulated temporal variability on diurnal, seasonal and interannual time scales. Section 5 presents the scheme's simulation of the temperature dependence of cloud properties and its sensitivity to prescribed SST perturbations. In Section 6 we discuss the implications of our work for future observations and for understanding actual climate sensitivity to realistic climate forcings.

2.) Model description

The baseline GISS GCM, Model II, is described in Hansen et al. (1983). The new cloud parameterization was implemented in an updated version of the GCM, run at 4°x 5° horizontal resolution with 9 vertical levels. Aside from the prognostic cloud scheme, the updated GCM differs from Model II in that it contains improved parameterizations of moist convection (Del Genio and Yao, 1993), the planetary boundary layer (Hartke and Rind, 1995), and ground hydrology (Rosenzweig and Abramopoulos, 1995), and uses the quadratic upstream scheme for advection of heat and moisture (similar to that described in Prather, 1986). The GCM was run on an IBM RISC6000 580 workstation, requiring about 8.3 CPU minutes per simulated day; 10-15% of the CPU time is used for parameterized moist processes (convection + stratiform clouds). The scheme is thus efficient and suitable for long climate simulations. Validation and sensitivity assessment in this paper make use of three runs conducted with the model: (1) A 6-year simulation with climatological SSTs, the results being averaged over the final 5 years; (2) A 10-year AMIP run with actual SSTs for the period 1979-1988; (3) Several 1-year perpetual July runs with prescribed globally uniform or non-uniform SST perturbations, the results being averaged over the last 7 months. A number of shorter sensitivity tests have also been conducted to isolate impacts of certain aspects of the model physics. Details of the prognostic cloud water budget parameterization are described below.

a.) *Cloud water formation and evolution*

To parameterize stratiform cloud generation, we follow the approach of Sundqvist et al. (1989). We divide the gridbox into a cloudy part (with fractional cloudiness b and relative humidity $U_s = 1$) and a clear part (with relative humidity U_o). The gridbox mean relative humidity is then

$$U = bU_s + (1-b)U_0 \tag{1}$$

As in Sundqvist (1978), net latent heating of the gridbox due to stratiform cloud phase changes (Q) equals the condensation heating in the cloudy part (Q_c) minus the evaporation of cloud water (E_c) and rain water (E_r) in the clear part, i.e.,

$$Q = bQ_c - (1-b)(E_c + E_r). \tag{2}$$

The continuity equation for the dimensionless cloud water content m can then be written

$$\begin{aligned}\frac{\partial m}{\partial t} &= A(m) + \frac{b^{Q_c}}{L} - (1-b)\frac{E_C}{L} - P + S_s \\ &= A(m) + \frac{Q}{L} + (1-b)\frac{E_r}{L} - P + S_s,\end{aligned} \tag{3}$$

where t is time, A(m) is the large-scale advection of cloud water, P is the conversion rate of cloud water to precipitation, L is the latent heat of condensation/deposition, and $S_s = S_d + S_e$ is the sub-grid-scale dynamical source/sink of cloud water due to convective condensate detrainment (S_d) and cloud top entrainment instability (S_e). The water vapor continuity equation thus includes a sink term -Q/L, and the thermodynamic energy equation a source term proportional to Q/c_p, determined by (2) rather than by the instantaneous phase change required to eliminate supersaturation that is characteristic of diagnostic schemes.

Sundqvist shows that stratiform latent heating can be expressed in terms of the gridbox mean relative humidity tendency as

$$Q = \frac{M - L_{q_s}\dfrac{\partial U}{\partial t}}{1 + \dfrac{U\,\varepsilon\, L^2 q_s}{Rc_p T^2}} \tag{4}$$

where M is the convergence of available latent heat into the gridbox (including the effects of temperature and pressure changes), q_s is the saturation specific humidity, ε is the ratio of the molecular weights of water vapor and dry air, R is the gas constant for dry air, c_p is the specific heat of dry air, and T is temperature. If we assume that the total source of water vapor from dynamic convergence and evaporation M + (1-b)(E_c+E_r) is divided into a part bM that condenses into the already cloudy fraction of a gridbox, and another part (1-b)(M+E_c+E_r) that increases the cloud cover and the relative humidity of the clear fraction, then it can be shown that (Sundqvist et al., 1989)

$$\frac{\partial U}{\partial t} = \frac{2(1-b)^2\ (U_S - U_{00})(M + E_c + E_r)}{L\left[2q_s(1-b)\ (U_s - U_{00})\ + m/b\right]} \tag{5}$$

(U_{oo} is defined below.) (5) is used to calculate the heating term (4), which is then used to predict the tendency of cloud water (3) at each physics timestep (1 hour).

b.) *Cloud cover and morphology*

We specify a threshold relative humidity U_{oo} below which stratiform cloud formation does not occur, and we assume that the relative humidity of the clear fraction increases as the cloud fraction increases (Sundqvist et al., 1989) according to

$$U_0 = U_{00} + b(U_s - U_{00}). \tag{6}$$

There is uncertainty in the calculation of saturation humidity at cold temperatures due to the complexity of the ice phase initiation process. As temperature decreases, the relative contributions of heterogeneous and homogeneous freezing (which require liquid water saturation) and deposition directly from vapor to ice (which requires only ice saturation) systematically vary. To account for this we define U_{oo} with respect to the saturated vapor pressure over liquid water (e_{sw}) for temperatures above -35°C, and with respect to the mixed phase pseudo-adiabatic process proposed by Sassen and Dodd (1989) for lower temperatures:

$$e_s / e_{sw} = 5.36x10^{-3}T(K) - 0.276. \tag{7}$$

We take U_{oo} = 0.6 for all clouds.

From (1), the stratiform cloud fraction is then diagnosed as

$$b = \frac{U - U_0}{U_s - U_o} \tag{8}$$

Although it is plausible that clear-sky relative humidity should be positively correlated with cloud cover on climatic time scales, there is no direct observational support for the use of (6) on an instantaneous basis or for the concept of a threshold relative humidity. Recent analyses of the dependence of cloudiness on relative humidity on GCM grid scales in the upper troposphere, using GOES 6.7 μm data (Soden and Bretherton, 1993), and in the lower troposphere, using radiosonde data in tandem with a mesoscale model (Walcek, 1994), suggest that a threshold relative humidity does not exist and that cloudiness is an almost linear function of large-scale relative humidity, with significant scatter. The GCM, despite its 60% threshold, produces some cloudiness at drier humidities because (1) the saturation reference at cold temperatures (equation 7) is less than that for water saturation, and (2) convective clouds in the GCM depend on cumulus mass flux rather than relative humidity. Nonetheless, the GCM underpredicts/overpredicts cloud cover at low/high relative humidity. On the other hand, the parameterization (6)-(8) performs satisfactorily compared to cloud ensemble model statistics (Xu and Kreuger, 1991).

Other approaches to parameterization such as cloud cover based on subgrid-scale deviations of temperature and moisture are equally plausible and theoretically preferable, but the specifics are similarly unconstrained by data. The arbitrary nature of assumed subgrid-scale variations in global climate models can in fact have significant unintended impacts on cloud feedback (Miller and Del Genio, 1994). Thus, at the current time there is no clear choice for the best way to predict cloud cover variations in GCMs, other than to ensure that they are statistically positively correlated with relative humidity variations.

Although (8) is typically interpreted as the cloud cover (i.e., the horizontal area fraction covered by cloud as viewed from above), it is actually the fraction of the gridbox volume occupied by cloud. In GCMs vertical resolution is too coarse to resolve many clouds, particularly layered stratus and cirrus in stable environments. As a result, if clouds are assumed to fill the gridbox vertically, cloud cover is underestimated while optical thickness (τ) is overestimated. For all but the optically thinnest clouds, the net radiative effect is an underestimate of solar reflection, since reflectance increases less than linearly with optical thickness.

We therefore distribute the cloud fraction b evenly in all three dimensions in stable situations. This allows for the possibility of cloud physical thickness less than the GCM layer thickness for the purpose of estimating optical thickness. The cloud cover (b') and cloud optical thickness (τ') used for radiation c alculations are thus given by

$$\begin{aligned} b' &= b^{2/3} \\ \Upsilon' &= b^{1/3}\Upsilon. \end{aligned} \tag{9}$$

In gridboxes in which moist convection has occurred, the environment is assumed to be disturbed and the clouds more vertically than horizontally developed; in such cases the original b and τ are used for radiative purposes instead. In the lowest model layer, an analogous choice is made in the presence/absence of cloud top entrainment instability (see Section 2d). This approach is at least qualitatively consistent with the observed tendency for layered stratus incidence to increase with stability (Klein and Hartmann, 1993). (9) is clearly not a complete solution to the problem of inadequate vertical resolution, since we perform only one radiation calculation per layer, while (9) implies a subgrid radiative flux divergence. However, it has a positive impact on the simulated global cloud cover and radiation balance, quantities that are biased low and high, respectively, in most GCMs.

Radiation computations in the GISS GCM are performed once per gridbox for either clear or cloudy conditions. For this purpose the box is determined to be either clear or 100% cloud-covered by comparing the fractional cloud cover determined by the cloud parameterization to a random number between 0 and 1 (Hansen et al., 1983). Fractional cloudiness in time is thus used as a proxy for subgrid-scale spatial fractional cloudiness. A single random number is chosen for the entire grid column; this is tantamount to a maximum overlap assumption. A sensitivity test using a different random number for each layer, which produces statistics more like those for random overlap instead, increases the global cloud cover by several percent and reduces solar absorption by about 2 W m^{-2} globally but affects the zonal mean by no more than 7 W m^{-2} at any latitude.

c.) *Cloud microphysics*

Autoconversion of cloud water to precipitation should be an increasing function of the density of condensate inside the cloud. The cloud water density within the cloud is given by $\mu = m\rho/b$. Precipitation formation is then parameterized as

$$P = C_0 m \left\{ 1 - \exp\left[-\left(\frac{\mu}{\mu_r} \right)^4 \right] \right\} + C_1 m P_r \tag{10}$$

where μ_r is a critical cloud water content for the onset of rapid conversion, C_o is the limiting auto-conversion rate for large μ, C_1 is an efficiency factor for accretion of cloud water by precipitation, and P_r is the precipitation flux entering the layer from above. (10) is similar to expressions suggested by Sundqvist et al. (1989) and Smith (1990) but with a larger exponent in the autoconversion term. This provides a sharper transition from weakly to strongly precipitating clouds but has a relatively minor effect on the simulation.

Our parameterization differs from that of previous models in three important ways:

(1) We use (10) for both liquid and ice phase clouds, differentiating between the two only via different values of μ_r, because stratus and cirrus are simply different manifestations of the same microphysics operating under different parameter settings. The UKMO GCM, for example, invokes a different representation for ice clouds that produces immediate precipitation (Smith, 1990). In climate change simulations with this GCM, the resulting short lifetime of ice clouds has important effects on the predicted cloud feedback (Mitchell et al., 1989), yet observational support for systematically different lifetimes for ice and liquid clouds does not exist.

(2) Precipitation formation is easier in maritime clouds than continental clouds, all other things being equal, because of the larger cloud condensation nucleus (CCN) concentration and resulting smaller droplet sizes over land (Twomey, 1977). We therefore adopt different values of μ_r for liquid phase clouds over land and ocean.

(3) The limiting autoconversion rate C_o is related to the coalescence and sedimentation rates

of droplets in static conditions. But clouds often form in regions of strong rising motion, which inhibits sedimentation. We therefore parameterize C_o as a decreasing function of the large-scale vertical velocity w (the gridbox mean minus any environmental subsidence due to moist convection) in regions of uplift according to

$$C_0 = \begin{cases} C_{00} 10^{-w/w_0} & (w >= 0) \\ C_{00} & (w < 0) \end{cases} \tag{11}$$

In the current version of the model, the microphysical constants are set to the following values: μ_r = 0.5 (liquid, ocean), 1.0 (liquid, land), 0.1 (ice) g m^{-3}, C_{oo} = 10^{-4} s^{-1}, w_o = 1 cm s^{-1}, C_1 = 1 m^2 kg^{-1}. The critical cloud water contents are chosen to be comparable to observed upper limits (Stephens et al., 1978; Hobbs and Rangno, 1985; Heymsfield and Donner, 1990), consistent with the assumption that cloud water removal by precipitation approximately balances production by condensation in the mature stage of the cloud life cycle. The limiting autoconversion rate is specified based on microphysics calculations which indicate that stratiform clouds typically require several hours to reach the precipitating stage (Mason, 1971). C_{oo} is the same for liquid and ice; the actual autoconversion rate differs for the two phases only to the extent that μ exceeds μ_r more easily for ice than for liquid. The accretion constant is chosen arbitrarily to make accretion competitive with autoconversion only for massive precipitating cloud systems.

We use a single prediction equation for all condensate regardless of phase. We assume that all clouds in a gridbox form as liquid when the temperature $T > T_o$, where T_o = -4°C over ocean and -10°C over land, based on observations compiled by Hobbs and Rangno (1985). For T < -40°C, all clouds form as ice. In between, the probability P_i of ice formation in a given gridbox layer is given by

$$P_i = 1 - \exp\left[-\left(\frac{T_0 - T}{15} \right)^2 \right] . \tag{12}$$

The choice of phase is then made by comparing P_i to a random number. (12) implies equal probability of liquid and ice formation at temperature T_o-12.5°C, i.e., -16.5°C (ocean) and -22.5°C (land). Falling snow melts in the layer in which the 0°C isotherm is crossed.

After the initial decision to form liquid or ice in a given layer, mixed phase processes can change the phase if ice falls into a lower layer containing supercooled liquid water. We parameterize Bergeron-Findeisen diffusional growth of the ice phase at the expense of the liquid phase via the "seeder-feeder" process by allowing a layer with supercooled water to glaciate if sufficient ice falls into it from above. We compute the probability of glaciation as

$$P_g = \{1 - \exp[-(M_i / M_l)^2]\}\{1 - \exp[-(C_0 C_B \Delta t / 2)^2]\}, \tag{13}$$

where M_i and M_l are the mass of ice entering the layer and the mass of supercooled liquid in the layer, respectively, and

$$C_B = 1 + \exp\left[-\left(\frac{T+15}{10}\right)^2\right], \tag{14}$$

with T in °C. P_g is compared to a random number to determine whether glaciation actually occurs in a given layer and timestep. Upon glaciating, the value of C_o used in the autoconversion estimate (11) for that layer also increases by the factor C_B.

The first term in (13) is designed to limit the occurrence of the Bergeron-Findeisen process when only trace amounts of ice are falling into a supercooled region. Thus, given a multilayer cloud with ice at the top and supercooled liquid below, the cloud can gradually glaciate from the top down as ice phase mass and sedimentation increase. Such clouds can then go through a life cycle in which the ice phase is increasingly preferred as the cloud ages. The second term in (13) allows for maximum probability of Bergeron-Findeisen growth near T = -15°C, where the difference between the saturation vapor pressures with respect to liquid and ice is large. The frequency of occurrence of the Bergeron-Findeisen process in the GCM is displayed in Fig. 1; it is most important at midlevels in the tropics and summer midlatitudes, and in the lower troposphere in the winter midlatitudes. Diffusional growth due to the presence of mixed phase clouds in a single layer can occur if condensate is detrained from a cumulus updraft into an anvil cloud of different phase.

The combined result of (12) and (13) is that the fractional occurrence of ice vs. liquid varies with temperature as shown in Fig. 2. Supercooled liquid persists down to temperatures approaching -40°C over land, consistent with *in situ* observations compiled by Feigelson (1978). Over ocean, the liquid phase disappears more rapidly with decreasing T and is almost nonexistent below -30°C, consistent with SMMR retrievals (Curry et al., 1990). The behavior in Fig. 2 differs from that assumed in the UKMO GCM, in which the transition from liquid to ice occurs completely between 0°C and -15°C (Smith, 1990). Cloud feedback in that GCM (Mitchell et al., 1989; Senior and Mitchell, 1993) may be negatively biased as a result (Li and LeTreut, 1992). It is worth noting that in the GISS GCM, negative feedbacks due to differing ice vs. liquid cloud life-times may be minimized in any case because (1) we assume the same limiting autoconversion rate for ice and liquid in the absence of observations to the contrary, and (2) the Bergeron-Findeisen process parameterization shortens the lifetime of supercooled liquid clouds underlying ice clouds.

Evaporation of cloud water is neglected in many GCMs, but is important to the extent that clear air is turbulently entrained into the cloud. Unfortunately, this is a complex dynamical problem which defies easy parameterization. To incorporate at least the basic microphysics, we define the droplet evaporation rate as (Twomey, 1977; Schlesinger and Oh, 1993)

$$\frac{-1}{t_{ed}} = -\frac{1}{r}\ \frac{dr}{dt}\ = \frac{1-U_0}{(K_1+K_2)\ r^2}\ , \tag{15}$$

where

$$K_1 = \frac{L^2 \rho_w}{kR_v\ T^2}\ ,\ K_2 = \frac{R_v\ T\ \rho_w}{De_s\ (T)}\ . \tag{16}$$

In (15) and (16), r is the droplet radius, p_w is the density of water, k is the thermal conductivity of air, R_v is the gas constant for water vapor, and D is the diffusivity of water vapor in air. The droplet radius is diagnosed from the cloud water content (Section 2e).

The cloud water evaporation rate on the GCM grid scale (t_e-1) is much less than this, because only a small fraction of the cloud mixes with clear air at any time, and droplets in the fraction that does are exposed to a relative humidity $U_o < U < 1$ because of the mixing. We thus set $t_e^{-1} = \alpha\ t_{ed}^{-1}$, with $\alpha << 1$ being a free parameter that incorporates not only the dynamical uncertainties but also the complexity associated with the presence of a spectrum of droplet sizes, each with a different evaporation rate. There is little guidance as to the appropriate magnitude of α. A water budget study of upper troposphere cumulus anvils suggests that cloud water evaporation is a small contributor (Gamache and Houze, 1983), but entrainment is thought to significantly dilute the properties of low-level marine stratus (Hanson, 1991). We take $\alpha = 10^{-3}$, which makes cloud evaporation an important sink for liquid clouds but generally unimportant for ice clouds because of the temperature dependence of e_s (Fig. 3). Plausibly, α might be made a function of stability instead. The cloud water evaporation rate in energy units is then estimated as

$$E_c = \alpha L \frac{m/b}{t_{ed}} = L \frac{m/b}{t_e}. \tag{17}$$

Rain (snow) evaporation (sublimation) does not affect cloud water content directly but does so indirectly by changing the gridbox relative humidity. We parameterize it following Sundqvist (1978) as

$$E_r = \frac{g}{\Delta p}(U_s - U_0) L p_r, \tag{18}$$

where g is the acceleration of gravity and Δp the layer pressure thickness. Precipitation that does not evaporate falls to the ground in one physics time step (1 hour), i.e., there is no precipitation budget in the model. This is a good approximation for rain falling from low levels but not for ice crystals precipitating from high altitude. It is not yet obvious whether the complexity of a prognostic snow budget is justified in climate models, however.

d.) Subgrid-scale cloud dynamical processes

We ignore advection of cloud water by the large-scale dynamics in this version of the GCM, i.e., A(m) = 0 in (3). The justification for this is twofold: (1) Cloud water contents are typically 1-2 orders of magnitude less than the water vapor content of a gridbox, so cloud water has little effect on the overall water transport. (2) The residence time of cloud water in the atmosphere (approximately $C_o^{-1} \approx 10^4$ s) is much less than that of water vapor (10^5-10^6 s). Thus, over the lifetime of a typical cloud, a wind of 50 m s^{-1} would be required to transport a substantial fraction of the cloud water even one gridbox horizontally in a model with 4°x 5° resolution. Vertical transport of cloud water is assumed to roughly offset sedimentation, which is approximately true for droplets of radius 10 μm and typical large-scale vertical velocities of several cm s^{-1}. The effect of variable vertical velocities is crudely accounted for via the parameterized dependence of C_o on w in (11).

Several subgrid-scale dynamical processes associated with vigorous vertical motion can have noticeable effects on cloud water content and optical properties, however. In mesoscale cirrus anvils associated with deep convective clusters, for example, convective condensate is transported vertically and is partly detrained into the anvil. An analysis of the water budget of a GATE cluster suggests that a significant fraction of the anvil water is detrained from the cumulus updraft rather than produced locally by stable ascent and condensation within the anvil itself (Gamache and Houze, 1983). This too is a complex dynamical problem, requiring information on cumulus updraft speeds and convective droplet size distributions. The prediction of such quantities is currently outside the computational scope of climate GCMs used for long integrations. We therefore simply assume that the water condensed at any level above the 550 mb level in deep convective updrafts in the GCM cumulus parameterization (m_c) is added to any existing stratiform cloud water at those levels, i.e., $S_d = m_c/\Delta t$ in (3), where $\Delta t = 3600$ s is the physics timestep. In other words, upper troposphere convective condensate is effectively "detrained" into a stratiform anvil and evolves according to the cloud water budget equation (3) rather than precipitating immediately, as does other convective condensate.

This coupling between the GCM's cumulus and stratiform cloud parameterizations has a dramatic effect on mean cloud water contents in the tropical upper troposphere (Fig. 4) and also produces realistic tropical cloud forcing variability (see Section 4c). With the cloud water budget and anvil detrainment, convective clusters in the GCM can have finite life cycles, with the anvil persisting after the convection has ceased. Examination of histograms of the lifetimes of tropical convective systems simulated by the GCM indicates that more of them have high cloud persisting for 2-7 hours after the initiation of deep convection than is the case when the Model II diagnostic cloud parameterization is used.

Although dilution by entrainment is crudely accounted for by our cloud water evaporation parameterization (17), in certain situations entrainment may catastrophically dissipate a cloud deck as a result of cloud top entrainment instability (CTEI). Unfortunately, the proper instability criterion for CTEI is a matter of considerable controversy. Randall (1980) and Deardorff (1980) derived an instability criterion based on the ratio of the equivalent potential temperature jump, or equivalently the moist static energy jump Δh, across the cloud top interface to the jump in total water content $\Delta(q+m/b)$. Defining $\gamma = (L/c_p)(dq_s/dT)_p$, $\delta = 1/\varepsilon - 1 = 0.608$, $\kappa = c_pT/L$, and $ß = [1+(1+\delta)\gamma\kappa]/(1+\gamma)$, the criterion for CTEI can be written

$$k = \Delta h / L\Delta(q + m/b) > k_{min}, \qquad (19)$$

where $h = c_pT+gz+Lq$ and

$$k_{min} = \kappa / \beta \approx 0.23. \qquad (20)$$

But observations of the transition from marine stratus/stratocumulus to scattered trade cumulus suggest that nearly overcast conditions persist even when (19)-(20) is satisfied (Kuo and Schubert, 1988). Kuo and Schubert suggest that the instability criterion is correct, but that slow growth rates in the marginal instability regime allow the cloud deck to survive for several hours. Betts and Boers (1990) suggest a transition at $k \approx 0.53$ instead on the basis of the available observations. MacVean and Mason (1990) and Siems et al. (1990) argue, however, that the criterion (19)-(20) is incorrect, and derive more restrictive instability criteria. The MacVean-Mason approach, for example, yields

$$k > k_{max} = \frac{(1+\gamma)[1+(1-\delta)\kappa]}{2+[1+(1+\delta)\kappa]\gamma} \approx 0.70. \qquad (21)$$

Recent numerical simulations by MacVean (1993) suggest a continuum of possibilities, with liquid water e-folding times of order 10^4 s when $k \approx k_{min}$ and 10^3 s when $k \approx k_{max}$.

Experiments with an early version of the cloud water budget parameterization produced the result that almost all low cloud was dissipated in the tropics and subtropics when (19)-(20) was used as an instability criterion. Thus, based on the available evidence, we have implemented the following parameterization for CTEI. When $k > k_{min}$, we mix air between the cloud top layer and the layer above in sufficient quantity to dissipate a fraction f of the cloud water in one physics timestep Δt, with

$$f = 1 - e^{-\sigma(k)\Delta t} \tag{22}$$

and

$$\sigma(k) = 2x10^{-4}\left(\frac{k - k_{min}}{k_{max} - k_{min}}\right)^5 s^{-}1. \tag{23}$$

Thus, $S_e = -fm/\Delta t$ is the CTEI cloud water sink in (3). The parameterization (22)-(23) allows for increasing cloud dissipation as k increases, but at a rate somewhat slower than in the simulations of MacVean (1993), since it acts in addition to the cloud water evaporation represented by (17). We allow CTEI to take place at any altitude, but it occurs almost exclusively in the first model level within the planetary boundary layer (PBL).

Figure 5 shows the geographical distribution of CTEI in the GCM. CTEI occurs mostly over the subtropical and tropical oceans (Fig. 5, top), with increasing frequency of occurrence with increasing distance from the west coasts of the continents. This is precisely the pattern expected for the stratocumulus-trade cumulus transition. Because of the restrictive instability criterion we use, however, the fraction of the cloud water mixed on average per physics timestep is only 10-30% of the total (Fig. 5, bottom). As a result, CTEI has only a moderate influence in the GCM in the current climate. Despite its secondary role, we include a CTEI parameterization in the GCM because its importance might increase in a warming climate and thus affect climate sensitivity.

Because of the unique cloud dynamics of the PBL, we parameterize the cloud morphology as follows. When cloud exists in the first model layer, we set

$$b' = \begin{cases} 1 & (k < k_{min}) \\ b + (1-b)e^{-\sigma(k)\Delta t} & (k > k_{min}) \end{cases} \tag{24}$$

for radiation purposes, while the optical thickness seen by radiation is

$$\tau' = \tau \frac{b}{b'} . \tag{25}$$

(24)-(25) imply that the cloud fully occupies the gridbox horizontally in stable conditions, with the fractional cloudiness occurring only in the vertical. In very unstable conditions, $b' \rightarrow b$ and $\tau' \rightarrow \tau$, i.e., the cloud is vertically developed and fractional cloudiness occurs only in the horizontal. If CTEI does not occur but moist convection originates in layer 1, we assume that $b'=b$ and $\tau'=\tau$ for any simultaneous stratiform clouds as in other GCM layers.

e.) *Cloud radiative properties*

Given a prediction of the instantaneous cloud water content, we can allow the visible optical thickness to vary in a self-consistent manner. For the wavelengths and particle sizes of interest, the extinction efficiency is almost independent of size parameter, so the optical thickness takes the simple form

$$\tau \approx \frac{3\mu\Delta Z}{2\rho_w r_e} , \tag{26}$$

where Δz is the GCM layer physical thickness and r_e the effective radius of the droplet size distribution (Hansen and Travis, 1974). Note that the cloud morphology prescriptions (9) and (25) are equivalent to assuming a cloud layer thickness of $b^{1/3}\Delta z$ and $(b/b')\Delta z$, respectively, rather than Δz in (26). Once the visible τ is estimated, infrared emissivity is then determined according to the spectral dependence predicted by Mie theory (Hansen et al., 1983), guaranteeing self-consistent shortwave and longwave radiative properties.

We diagnose particle size from the predicted cloud water content. Ignoring for simplicity the difference between the effective radius and volume-weighted mean droplet radius (r), the cloud droplet concentration N is given by

$$\mu = N\rho_w \frac{4}{3}\Pi r^3 \tag{27}$$

for spheres. Observations suggest that for liquid phase clouds, constant N is a good approximation for low to moderate τ (Slingo et al., 1982; Han et al., 1994). This implies that r increases as $\mu^{1/3}$. We fit this behavior to data of Stephens et al. (1978), taking

$$r = r_0(\mu/\mu_0)^{1/3} \tag{28}$$

with r_o = 10 μm at μ_o = 0.25 g m^{-3} (corresponding to N ≈ 60 cm^{-3}) over ocean. Over land, where there are many more CCN, we set r_o = 7 μm (N ≈ 170 cm^{-3}) instead. For ice clouds, we fit (28) to the data of Platt (1989), although the fit is less satisfactory. Fewer particles act effectively as ice nuclei, so ice crystals tend to be larger than liquid droplets. We use r_o = 25 μm at μ_o = 4.2x10^{-3} g m^{-3} (N ≈ 0.06 cm^{-3}) for all ice clouds. In this case r is the radius of an equivalent sphere, i.e., the Mie scattering phase function is used. The data of Nakajima et al. (1991) and Han et al. (1994) suggest that r for liquid clouds does not increase indefinitely with μ, perhaps due to the onset of precipitation. We therefore set r = r(μ_r) when $\mu > \mu_r$ for liquid clouds. (28) is also used in calculating the cloud droplet evaporation rate (15), i.e., we ignore the difference between effective and mean radius.

Frequency histograms of effective radius resulting from this parameterization are shown in Figure 6. The mean low cloud liquid droplet radius is about 8 μm over ocean and 6 μm over land, but the distribution is broad. The cutoff at 14 μm represents the efficient precipitation threshold; this threshold is more commonly reached for marine clouds, which precipitate more easily, than for continental clouds. Low level ice clouds, for which data are sparse, typically have quite large particle sizes (40-80 μm). High level ice clouds exhibit a bimodal distribution, with a peak near 60 μm due to thick cumulus anvil clouds (mostly occurring near the anvil base) and a 5-15 μm population of thin cirrus. The diagnosed effective radii for low liquid and high thin ice clouds are somewhat smaller than observed (Han et al., 1994; Wielicki et al., 1990) for several reasons: (1) We do not distinguish between volume-weighted mean radius and effective radius; (27)-(28) are more appropriate to the former. For a standard gamma distribution of droplet size with an effective variance of 0.2, a typical value for stratus, $r_e \approx 1.3r$ (Han et al., 1994). (2) The parameterization depends on the predicted liquid water content μ; we show in the next section that the current GCM underpredicts liquid water path (the vertical integral of μ) relative to microwave-retrieved values in several regions of thick cloudiness.

3.) Mean state

Energy balance and hydrologic cycle parameters simulated by the model are listed in Table 1. Global top-of-the-atmosphere (TOA) radiation budget and cloud forcing components are within 5 W m^{-2} of observations except for longwave cloud forcing, which is 5-10 W m^{-2} weaker than that inferred by Nimbus-7 and about 15 W m^{-2} less than that retrieved by ERBE. About half of this discrepancy can be explained by the GCM's underestimate of high cloud cover (see Section 5). The remainder may be an observational bias, because the satellite data are derived from comparisons of cloudy and clear regions and thus the cloud forcing includes the effect of higher humidities within clouds; the GCM performs offline clear sky calculations even in cloudy gridboxes and thus isolates the true cloud effect. TOA and surface shortwave cloud forcing are almost identical in the GCM.

Total cloud cover in the GCM is a few percent less than that estimated by ISCCP, with most of the underestimate occurring over ocean. The GCM correctly simulates the large land-ocean difference in total cloud cover but not the significant January-July difference over land, for reasons that will be discussed later. Low cloud cover is larger than estimated by surface observers by about 5%, and high cloud cover lower than estimated by ISCCP by the same amount. The GCM is generally too dry and too cold in midlatitudes, especially in the middle and upper troposphere; it is somewhat too wet in the tropical upper troposphere. Liquid water path is somewhat smaller than observed, although the uncertainty in the data is large. Liquid water path is greater than or comparable to ice water path in the subtropics, but the ice phase dominates elsewhere. Precipitation by stratiform clouds accounts for about 1/3 of the total globally, and about 15% near the equator; the latter is less than estimated from budget studies of tropical cloud clusters, but a great improvement over Model II, which has no mechanism for condensate detrainment into anvils and thus has virtually no stratiform precipitation in the tropics.

Figures 7-13 display the geographic distributions of several GCM-simulated quantities and differences between the GCM and observations, including TOA absorbed shortwave and outgoing longwave radiation (ASR, OLR; differences only), TOA shortwave and longwave cloud forcing (C_s, C_l), total, high, and low cloud cover (TC, HC, LC), and cloud liquid water path (LWP). ERBE data (Barkstrom, 1984) are used to validate radiation quantities, ISCCP C2 and C1 data (Rossow and Schiffer, 1991) for total and high cloud cover, respectively, the surface cloud observation data set of Warren et al. (1986, 1988) for low cloud cover, and the SSM/I retrieval of Lin and Rossow (1994) for liquid water path.

It is important to note specific problems with individual data sets. ERBE cloud forcing is less accurate than its global TOA radiation fluxes because the former requires separation of cloudy and clear scenes. This is a severe problem in the polar regions, where cloud detection over snow and ice is sufficiently difficult to produce the incorrect sign of cloud forcing (Cess, personal communication); we restrict comparisons to latitudes equatorward of 60°. ISCCP also has a detection problem over snow and ice, but of unknown magnitude; we return to this question later in this section. Surface cloud observations are of poorer quality over lightly-traveled ocean regions, such as the southern midlatitudes, than over land. SSM/I liquid water path is available only over ocean because of the variable microwave surface emissivity of land. Retrievals by different groups differ completely, even as to the sign of the latitudinal gradient. These differences may be caused by different algorithm assumptions about whether clear points are included in the average, what cloud temperature is assumed, and how column water vapor is retrieved (Lin and Rossow, 1994). In heavily precipitating regions, the retrieved cloud water path probably includes a partial contribution from precipitation-sized droplets. For the GCM, only the cloud water path is included; for convective clouds, whose cloud water content is not predicted, we convert the prescribed optical thickness for liquid parts of the cloud to a proxy liquid water path using a relation suggested by Lin and Rossow. Nonetheless, model-data discrepancies in this quantity in the ITCZ should be viewed with caution.

The TOA radiation balance (Fig. 7) represents the integrated effect of all elements of the simulated climate, while Figures 8-13 permit us to understand these model-data differences in terms of individual cloud types and/or hydrologic/radiative quantities. Errors in simulated temperature and humidity can contribute to the differences in Figure 7 as well. Differences between cloud forcing errors (Figs. 8-9) and TOA radiation errors provide a qualitative measure of clear sky contributions to the total TOA radiation error; these differences are significant only in the longwave. Additional validation of the GCM's upper troposphere water vapor distribution against SAGE II data can be found in Del Genio et al. (1994). We organize the discussion below according to different climate regimes in which different cloud types dominate the radiation signature.

(*a*) *Tropical convection regions*: The GCM overestimates the magnitude of C_s in the ITCZ, especially over ocean. LC is too high over the tropical oceans, while HC is slightly overestimated, but not sufficiently to explain the total C_s error. LWP is also underestimated, but the data-model comparison is very limited here, as discussed previously. This suggests that either model anvil clouds contain too much ice or that excessive low cloudiness contributes too much to C_s. C_l is generally underestimated, but not always in the regions of the maxima in C_l and C_s, while both positive and negative OLR errors occur. This is probably the manifestation of some errors in the exact location of convective centers combined with the model's excessive upper troposphere humidity.

(*b*) *Subtropical/tropical ocean subsidence regions*: These areas, off the west coasts of North America, South America, and Africa, are dominated by low-level marine stratus, which have a noticeable shortwave effect and little longwave signature. The GCM underestimates TC, mostly due to insufficient LC, and thus underestimates C_s in these regions, more so in July than January. This occurs despite the relative absence of CTEI in these regions (Fig. 5). LWP errors are within the observational uncertainty, which illustrates that the microwave is not very sensitive to thin clouds.

(*c*) *Subtropical/midlatitude continents*: The major deficiency in the GCM's cloud simulation occurs over Eurasia. In January, excessive TC is simulated, most of it due to excessive LC, in northern/eastern Eurasia. This is a problem common to many GCMs (Mokhov, personal communication). It may either be an indictment of the ability of GCM boundary layer parameterizations to vent moisture into the free troposphere under stable conditions, or an example of the need to develop cloud cover parameterizations based on stability-dependent subgrid-scale temperature and moisture variances. In July, to the west and at somewhat lower latitude, TC, LC, and C_s are all greatly underestimated. This may have several causes, including the underestimate of shallow cumulus over land by the GCM's convective scheme (see Section 4a) and the underestimate of potential evapotranspiration by the GCM's land surface parameterization. There is also slightly too much OLR and too little C_l and HC in these regions, also suggestive of a deficient local surface moisture source. It is this region which accounts for the GCM's incorrect seasonal cycle of global mean continental cloud cover (see also Section 4b).

(*d*) *Midlatitude storm tracks*: The GCM consistently underpredicts C_s, TC, and LWP off the east coasts of North America and Asia and throughout the Southern Hemisphere midlatitude oceans. HC errors are large relative to LC errors, and C_l is too low as well, which suggests model deficiencies in baroclinic storm-generated nimbostratus. This is also the region in which the model's middle and upper troposphere are substantially drier than observations. Errors in OLR are smaller than those in C_l, despite the low humidity, presumably because of the compensating effect of underestimated upper troposphere temperature at these latitudes. Reduction of 5-10 W m^{-2} in ASR is realized in a sensitivity experiment in which the model's cloud overlap is changed from maximum to effectively random, but the remaining error is insensitive to large changes in the parameterization's microphysical constants. A large reduction in threshold relative humidity for cloud formation does produce improvement, but at the expense of a seriously degraded tropical cloud simulation. There are several reasons to suspect that this problem lies outside the cloud parameterization. The underestimate of water vapor is completely insensitive to any change in the cloud parameterization; the primary water vapor source in the GCM's budget at these latitudes is transport by large-scale eddies (Del Genio et al., 1994). However, the GCM's upper troposphere eddy kinetic energy is about 30% lower than observed, despite the fact that the parameterization increases both eddy kinetic energy and baroclinic conversion relative to the previous diagnostic scheme (see Table 2). Furthermore, a sensitivity experiment in which the new cloud water budget and moist convection schemes are combined with the previous Model II dynamics, PBL, and land surface parameterizations produces about twice as much midlatitude high cloudiness.

(*e*) *Polar regions*: These are presumably regions of primarily boundary layer and mid-troposphere cloudiness, but few data exist. The only unambiguous validation statement that can be made is that the GCM overestimates ASR in the summer polar region, but snow/ice coverage and albedo errors contribute to this to an unknown extent. Much of the error is likely to be due to clouds, however, since the ASR difference is an extension of that in midlatitudes. According to ISCCP, though, the GCM greatly overpredicts TC at the summer pole; either the data are unreliable here, or the simulated optical thicknesses are much too small; we will return to this point later. At the winter pole, the model underpredicts TC according to ISCCP, but again, uncertainties are large.

Further insight into these differences can be gained by examining the cloud radiative properties directly. Figures 14-16 compare simulated and ISCCP-observed (Rossow and Schiffer, 1991) two-dimensional histograms of cloud top pressure and optical thickness for January in selected latitude zones. To compare the GCM to ISCCP, we must take into account biases produced by the ISCCP cloud retrieval algorithm. ISCCP errs in its determination of cirrus properties because (1) it neglects the non-sphericity of ice crystals (as does the GCM) and underestimates their particle size, (2) it does not detect extremely thin cirrus and cannot unambiguously determine the cloud top pressure of the thinnest clouds it does detect, and (3) it does not sufficiently correct upward the cloud top pressures of slightly optically thicker clouds. These differences have been quantified by Liao et al. (1995b) by comparisons between nearly coincident SAGE II and ISCCP pixels. Using these results as a guideline, we "detect" the highest cirrus layer in the GCM as ISCCP would, by ignoring clouds with $\tau < 0.1$, placing clouds with $0.1 < \tau < 0.3$ at the tropopause (as ISCCP does when it cannot determine a cloud top pressure), and placing clouds with $0.3 < \tau < 0.5$ one model layer lower than that at which they actually occur. Furthermore, the GCM is "viewed" top down as the satellite would see it, with only the top pressure of the highest cloud in the column included. The GCM figures are thus an approximation of the ISCCP "detection" of the GCM cloud field rather than the actual GCM cloud distribution. ISCCP also probably underestimates the optical thickness of low clouds in cases of subpixel (~≤ 5 km) fractional cloudiness, but this is difficult to quantify and has not been taken into account in the figures.

Over the tropical oceans (Fig. 14 a, b), the GCM correctly simulates the bimodal optical thickness distribution of high clouds, presumably due to deep convection and thick anvils and associated cirrus. This suggests that the GCM's overestimate of C_s (Fig. 8) at these latitudes is due more to its overestimate of low cloud cover than to an overestimate of ice water content (but see the discussion in Section 6). ISCCP observes cloud top pressures systematically increasing with decreasing τ, while the GCM's high clouds peak near 250 mb, independent of τ. The GCM and ISCCP agree that the dominant cloud type at these latitudes is low-level stratus, with tops near 900 mb. The GCM's optical thicknesses are systematically higher than ISCCP's, but whether this is a real discrepancy or an ISCCP bias is not known; if it is real, then it contributes to the excessive tropical C_s.

Over subtropical oceans (Figs. 14 c,d), the same low-level stratus cloud type is even more dominant in both model and data. This supports our earlier conclusion that the GCM's underestimate of C_s in the eastern ocean marine stratus decks is mostly due to an underestimate of low cloud cover (Fig. 12). There is also a tendency for optically thicker low clouds at slightly lower top pressures; in the data these are probably trade cumulus (750 mb tops), but in the GCM, which underpredicts shallow convection (see Section 4a), these are probably thicker stratus (850 mb tops) instead. Both model and data indicate a broad secondary distribution of high level clouds, but the top pressures are 100-150 mb lower in the GCM than in ISCCP.

Midlatitude winter continental cloudiness consists primarily of midlevel optically thick clouds, typical of nimbostratus, with secondary peaks due to thin cirrus, midlevel moderate τ cloudiness (perhaps altocumulus or altostratus), and a hint of moderate τ stratus at low levels (Fig. 15 a, b). The GCM identifies each of these cloud types, but simulates too much low stratus and too little nimbostratus. This is consistent with our earlier conclusion of excessive low cloudiness in winter over Eurasia. Over midlatitude oceans (Fig. 15 c, d), the GCM simulation is fairly good relative to ISCCP, but with the same problem as in the tropics of similar top pressures for optically thick and thin high clouds. The GCM distribution is too heavily weighted toward low clouds, which suggests that the underprediction of C_l, C_s, TC, HC, and LWP in the storm tracks is due to nimbostratus occurring too infrequently, rather than their optical properties being incorrect. The weakness of the GCM's synoptic storms is consistent with this conclusion.

The GCM predicts a bimodal distribution of Arctic cloud types in daytime in both summer and winter (Fig. 16 a, b). In both seasons the model's primary cloud type is low stratus, but more so in summer. In both seasons these clouds have a broad optical thickness distribution, but in winter there is more optically thin cloudiness. The secondary peak in both seasons is at midlevels, slightly higher in altitude in summer than winter and somewhat optically thicker as well. The winter distribution of nighttime cloudiness (not shown) at both low and midlevels peaks at even lower values of τ. ISCCP does not obtain τ information in winter because of the absence of sunlight, but its Arctic summer distribution completely disagrees with the GCM, with primarily optically thick midlevel cloud and thin cirrus and almost no boundary layer cloudiness (Fig. 16 c). Curry and Ebert (1992) have estimated an Arctic seasonal climatology, consistent with available TOA and surface flux data, that suggests a bimodal (low and midlevel) distribution of clouds and τ varying from about 2 in winter to about 8 in summer. The GCM is in reasonable agreement with this estimate, casting doubt on the ISCCP inference. GCM clouds over Antarctica in summer (not shown) are more like those in Arctic winter, but with even optically thinner low cloudiness. This too disagrees with ISCCP, whose Arctic and Antarctic clouds (not shown) are similar.

Liao et al. (1995a) have examined the latitudinal distribution of the thin cirrus not seen by ISCCP but detected by SAGE II. They find that thin cirrus cloud amounts are typically of order 10-20% in the tropics and summer midlatitudes, and 5-10% elsewhere. If thin cirrus are defined in the GCM as all clouds with top pressures < 550 mb and column $\tau < 0.1$ down to this level, then the GCM produces significantly less thin cirrus: 3-6% in the tropics, 0.5-2% in the subtropics and summer midlatitudes, and 0.1-4% in the winter midlatitudes. If we adopt the less stringent cutoff $\tau < 0.3$, then these amounts approximately double, leaving them within a factor of 2 of the SAGE II result in the tropics but still considerably less at higher latitudes. Certainly the GCM's upper level dry bias must play a role in the midlatitude deficiency of thin cirrus, but this cannot be said for lower latitudes, which are somewhat too moist. One possibility is that the GCM's coarse vertical resolution near the tropopause (2-3 km) prevents the formation of thin, stable, moist layers in which cirrus are commonly observed to form (Starr and Wylie, 1990).

Table 2 compares various indices of the GCM's general circulation with those from a one-year sensitivity experiment in which the prognostic scheme was replaced with the previous Model II diagnostic cloud parameterization. All other model physics is identical in the two runs. The version with the prognostic scheme has a more vigorous hydrologic cycle, with increased precipitation, evaporation, and cumulus mass flux. This produces a stronger general circulation, with intensified Hadley and Ferrel cell streamfunctions and increased eddy kinetic energy; each of these improves the model's agreement with observations. Large-scale energy transports, both vertical and horizontal, increase as well, primarily due to increases in latent heat transports.

To understand these results, we examine the January zonal mean changes in cloud cover, radiative heating, moist convective heating, and stratiform condensation heating that result from the implementation of the prognostic cloud parameterization (Fig. 17). The prognostic scheme dramatically increases low and middle level cloud cover in the tropics and subtropics, while generally decreasing high level tropical cloud cover and all types of cloudiness at higher latitudes (Fig. 17a). The changes in cloud cover are due primarily to increases/decreases in water vapor concentration (not shown) at high/low temperatures rather than changes in temperature itself. Temperature differences (not shown) are small with two exceptions: the prognostic scheme cools the tropical tropopause by 4°C and warms the winter polar region by 3-6°C, both improvements. There may be several reasons for the change in the moisture field. The temperature-dependent cloud water evaporation we employ (Fig. 3) is one candidate. Given the longer residence time of cloud water in a prognostic scheme, there is more opportunity for condensate to evaporate before it has the chance to precipitate. In the diagnostic Model II scheme, cloud water is instantly converted to precipitation, and thus has only one chance to evaporate. In addition, CTEI provides a net increase in atmospheric humidity by removing moisture to the top of the PBL; the loss of near-surface moisture by entrainment mixing is balanced by increased surface evaporation.

The change in moisture field has predictable effects on the diabatic heating distribution (Fig. 17b, c, d). With a steeper humidity profile at almost all levels and more low cloud, there is generally increased longwave flux divergence in the tropics and hence stronger radiative cooling at low latitudes. Shortwave heating differences are less dramatic, but reduced solar heating (consistent with the presence of optically thick anvils) helps produce the colder tropical tropopause. This decrease in the latitudinal gradient in radiative heating is more than compensated, however, by the increased tropical moist convective heating driven by the wetter lower troposphere and a slightly steeper lapse rate. The change in cumulus heating is partly offset by stratiform cloud water evaporation in the middle and lower troposphere, but it is augmented in the upper troposphere by condensation heating associated with anvil cloud formation. The net result is that the latitudinal gradient of tropical diabatic heating increases, driving a stronger Hadley cell.

The midlatitude picture is more complicated. The enhanced latent heat transport is due in part to the steeper humidity gradient, but the transient eddy kinetic energy (EKE) increases as well. This occurs despite the fact that temperature increases at high latitudes, thus decreasing baroclinicity. The higher polar temperatures themselves are due to slightly enhanced poleward dry static energy fluxes at high latitudes. In the face of smaller meridional temperature gradients, the stronger midlatitude eddies exist only because of increases in condensation heating there and associated enhanced longwave heating (Fig. 17b, d). Since mean cloud cover decreases in midlatitudes, the stronger eddies are likely to be due at least in part to more favorable correlations between cloud processes and regions of rising motion. As evidence, eddy generation of available potential energy (APE) at 45°N increases by 6% with the new scheme, with a 50% increase in the condensation contribution more than offsetting a 40% increase in the magnitude of the (negative) radiation contribution. This may be caused by both the variable optical thickness permitted in a prognostic scheme and suppressed autoconversion in regions of rising motion (11). As a result, conversion of APE to EKE increases by 40%.

4.) Temporal variability

a.) Diurnal cycle

Diurnal variations in cloudiness are often ignored in assessments of GCM performance. Indeed, some climate GCMs inexplicably still do not even include the diurnal cycle of insolation. Yet it is possible to induce a cloud feedback without changing cloud cover merely by shifting the cloud cover maximum from day to night, or vice-versa. Furthermore, diurnal cycles differ for different cloud types and thus provide a stringent test for GCM physics. We focus on diurnal variations in total and high cloud cover, which dominate the shortwave and longwave diurnal signal, respectively. In the GCM, total and high cloud cover are indicative of the diurnal behavior of low and middle level cloud cover, respectively, as well. The same is true of satellite data to a great extent for total vs. low cloud, but less so for high vs. middle cloud, although cloud shielding effects in the data may be important.

Figure 18 (upper) shows the diurnal cycle of zonal mean high cloud cover over land as simulated by the GCM (left) and observed by ISCCP (right) for July. The GCM results are averaged over 3-hour intervals of local time to match the ISCCP temporal resolution. Almost identical results are obtained by sampling the GCM at 3-hour intervals, and neither technique qualitatively changes either the amplitude or phase seen at full resolution. The GCM successfully simulates the afternoon high cloud maximum over summer midlatitude land, and the slightly later maximum over tropical land, although the observed maximum at most latitudes is 2-3 hours later than the simulated peak. The GCM also correctly simulates the increase in diurnal amplitude from midlatitudes to tropics, although the GCM midlatitude amplitude is too weak. In January (not shown), the GCM has no clear midlatitude diurnal cycle while ISCCP has a weak afternoon peak.

The diurnal cycle of high cloud over ocean is shown in Figure 18 (lower). ISCCP indicates a strong semidiurnal component over midlatitudes of both hemispheres, with maxima near both dawn and dusk. In the tropics the signal is more diurnal with the dusk maximum dominating. The GCM produces a maximum several hours before dawn at most latitudes, with a secondary maximum several hours before dusk present mostly in the subtropics and higher latitudes. Both the model and ISCCP produce very weak amplitudes (2%) for the diurnal cycle of oceanic high cloud. The GCM's tropical peak is reminiscent of that observed in most of the tropical Pacific, but several hours earlier (Fu et al., 1990). The absence of a GCM evening equatorial maximum, as is observed in the east Atlantic, may be indicative of the model's generation of propagating African easterly waves.

Figure 19 shows the corresponding diurnal cycles of total cloud cover. Except near the equator over land, where the observed diurnal cycle is controlled by the evening maximum in high cloudiness, both the model and ISCCP results are indicative of the diurnal cycle of low cloud. The GCM's diurnal cycle of continental low cloud peaks in morning, however, while that observed by ISCCP peaks generally in early afternoon. Over ocean, the GCM agrees with ISCCP's placement of the diurnal maximum of total cloud slightly before dawn. The GCM also correctly simulates the rather large diurnal amplitude over land and the small amplitude over ocean. Except over tropical land, neither the GCM nor ISCCP indicate dramatic latitudinal variation of diurnal cycle phase; the observed peak is perhaps a bit later in the morning in midlatitudes than in the tropics.

The GCM's misplacement of the continental maximum in total cloud is its most glaring shortcoming. Surface observations of low cloud indicate the same afternoon maximum over land that ISCCP observes in total (and low) cloud (Warren et al., 1986, 1988). The surface observations contain morphological distinctions between cumulus and stratus+stratocumulus+fog. The former peaks in early to mid-afternoon over land, while the latter peaks in the early morning. Since the diurnal cycle of cumulus is about twice as large as that of the stratiform low cloud, the former determines diurnal cycle phase.

In the GCM, the situation is reversed. Shallow convection occurs almost as frequently (20-30%) in the model as in the observations, except in western North America. But the cloud amount when present is only about 5%, as opposed to 25-35% in the data. Thus, only a few percent of the GCM's 25-30% low cloud cover over land is convective, and hence its diurnal cycle is determined by the morning peak in low stratus. Apparently the GCM is too unstable over land in early afternoon, generating deep convection when shallow fair-weather cumulus should dominate; this behavior may be sensitive to errors in ground hydrology or boundary layer fluxes. This is a land problem only; over oceans, the GCM has fairly realistic cumulus mass flux distributions and the correct diurnal phase. In addition, different cloud cover parameterizations may be required for shallow and deep convection. In the GCM, both are specified as equal to the fraction of layer mass that convects, but the mass flux/area ratio may differ from shallow to deep cumulus given their systematically different updraft vertical velocities.

b.) *Seasonal cycle*

The seasonal variation of cloud properties is affected by at least four different processes: the migration of the ITCZ and the rising branch of the Hadley cell across the equator, the reduction in equator-pole temperature contrast in summer and associated decline and poleward shift in baroclinic wave activity, the increase in convective instability in summer, and the seasonal melting of snow and sea ice. The first of these has no simple relationship to long-term climate change and sensitivity, hence the climatic irrelevance of hemispheric mean seasonal changes. But the last three are indicative of changes predicted to occur in a warming climate, so seasonality can be a useful validation tool if the effects of individual processes on cloudiness are considered.

Figure 20 shows the zonal mean seasonal cycle of high cloud cover over land (upper) and ocean (lower) simulated by the GCM (left) and observed by ISCCP (right). Over both land and ocean, the dominant feature is the movement of the ITCZ, which lags insolation by 1 month over land and 2 months over ocean in both data and model. The GCM seasonal amplitude is about twice as strong over land as over ocean, somewhat more than the observed land-ocean difference. The maximum seasonal excursion is correctly simulated to be 10°-20° latitude in summer over land, but over ocean the GCM's ITCZ drifts poleward to 20°-25° while the observed ITCZ remains within 10°-15° of the equator.

Of more interest climatically is midlatitude high cloud cover, which is plausibly an index of both baroclinic instability and deep convection (Del Genio et al., 1994). Both the GCM and ISCCP produce a continental peak in late summer at high latitudes which shifts toward spring in midlatitudes, with amplitudes less than half that of the ITCZ migration. ISCCP's midlatitude peak is actually in late winter, while the GCM's is several months later. This is consistent with the

GCM's winter dry bias. Over ocean there is a weaker seasonal amplitude of high cloud in both model and observations, with the GCM's peak occurring in mid-summer, while ISCCP shows a summer peak at high latitudes but a semiannual structure with late summer and early winter peaks in midlatitudes.

Figure 21 shows the corresponding seasonality in total cloud cover. In the tropics, the seasonal cycle of total cloud is dominated by high cloudiness, and is thus similar to that in Figure 20. In middle and high latitudes, low cloud contributes to the seasonal cycle but more so in the model than in the data. Over ocean both model and data indicate a broad maximum of total cloudiness in late fall and winter in the northern hemisphere and late winter into spring in the southern hemisphere, but weaker in the model, consistent with its deficient baroclinic wave activity. There is also a weak secondary northern midlatitude peak in summer. Over land, the GCM disagrees with ISCCP in several ways: (1) Due to the model's excessive Siberian winter cloud cover (cf. Fig. 10), its seasonal cycle in northern midlatitudes is completely out of phase with the observations at 50°-60°N, and several months out of phase at 30°-50°N. (2) In southern midlatitudes, the model's total cloud peaks in winter while ISCCP has a semiannual behavior with an additional late spring peak. (3) In the GCM, polar cloudiness peaks in summer in both hemispheres while ISCCP indicates a winter peak; here ISCCP disagrees with surface climatologies and may be in error (Mokhov and Schlesinger, 1994), although "cloudless" ice crystal precipitation unaccounted for in some data sets complicates the interpretation (Curry and Ebert, 1992).

Seasonal variations in cloud forcing will be documented elsewhere as part of the FANGIO GCM intercomparison activities. Here we only briefly note the major features of the model-ERBE seasonal comparison. For comparison purposes, ERBE zonal mean, area weighted seasonal changes were defined as January minus July and Southern Hemisphere minus Northern Hemisphere; the result is effectively a composite summer minus winter change at each latitude. Relative to this standard, the GCM's RMS cloud forcing differences are 13.7 (C_s) and 12.4 (C_l) W m^{-2}, and the correlation coefficients between the GCM's and ERBE's seasonal variations are 0.88 (C_s) and 0.94 (C_l). The sources of the differences are readily determined by examining Figures 7-13 and 20-21. In the tropics over ocean, where the ITCZ shifts too far poleward in summer, the

GCM produces 15-20 W m^{-2} positive/negative C_s errors at 10°/20° latitude, and comparable errors of opposite sign in C_l. In midlatitudes, underestimates of storm track cloudiness and cloud liquid water content cause the GCM to underestimate the seasonal variation of cloud forcing by a similar amount. At higher latitudes, the GCM's excessive low cloudiness over Eurasia in winter and deficit in summer produce a 20 W m^{-2} error in C_s but almost no error in C_l.

c.) *Interannual variations*

The most well-documented aspect of cloud variations on time scales longer than one year is that due to El Niño - Southern Oscillation (ENSO) perturbations. The ENSO signal in cloud and radiation parameters is strongest near the source, in the tropical Pacific, although weak cloud perturbations may occur elsewhere due to teleconnections.

Figure 22 shows Hovmöller diagrams of OLR and precipitation anomalies at the equator across the Pacific for 1979-1988, with the GCM forced by observed AMIP SSTs. Compared to observations (Kousky and Leetmaa, 1989), the simulation of OLR anomalies for the 1987 ENSO is quite good. Both model and data suggest peak anomalies of about 50 W m^{-2} in early 1987 just east of the dateline; the GCM's peak is about 10° east of the observed peak and persists further into the year. Prior to the El Niño, both model and data show negative anomalies in the central Pacific of 10-30 W m^{-2} and slightly smaller positive anomalies in the west; the peak negative anomaly occurs just west of the dateline in winter 1984, precisely as observed. The model representation of the 1982-83 ENSO is not quite as good but acceptable: the maximum OLR anomaly is about 40 W m^{-2} vs. 60 W m^{-2} observed and is spread over the central and east Pacific rather than focused near 150°W. In general the model interannual variation is somewhat noisier than the observed anomalies throughout the ten year period. The corresponding precipitation anomaly record in Figure 22, though having no reliable observational counterpart, matches most of the features of the OLR record, suggesting that ENSO OLR anomalies are caused by the optically thick anvil clouds accompanying deep precipitating convective systems. Anomalies in low-level wind fields (not shown) are also realistic, suggesting that the GCM produces the correct dynamic response.

Ramanathan and Collins (1991) have examined interannual differences in ERBE TOA shortwave and longwave cloud forcing in the tropical Pacific. The corresponding GCM simulation of the correlation between C_s and C_l differences between ENSO and non-ENSO months is shown in Figure 23. Diagrams such as these simply reflect dynamical shifts in locations of extensive deep convection and thus contain no information about the presence or absence of "thermostat"-type feedbacks on SST change (Fu et al., 1992). Nonetheless, the data provide a useful test of the model's ability to simulate cloud radiative properties; in particular, since longwave and shortwave perturbations are highly correlated, Figure 23 is an indicator of the GCM's success in simulating cumulus anvils. The GCM does an excellent job in reproducing the ERBE results, with C_s differences of up to about ± 70 W m^{-2} and C_l differences of up to ± 50 W m^{-2}; the slope of the best fit is -1.14, almost identical to the -1.20 seen by ERBE. This result may be somewhat fortuitous, given the GCM's simplistic prescription for detrainment of convective condensate and its use of Mie scattering for ice clouds (see Section 6). Surface shortwave cloud forcing in the GCM (not shown) is almost identical to TOA shortwave forcing, with small ($\pm$ 5 W m^{-2}) systematic differences of opposite sign for gridboxes dominated by high and low clouds. Surface longwave cloud forcing (not shown) is small (10-15 W m^{-2}) throughout the tropical Pacific and weakly negatively correlated with shortwave forcing; the small magnitude is realistic given the large specific humidity of the tropical PBL.

5.) Temperature dependence and sensitivity

Increasingly, variability of observed cloud and radiation parameters is being analyzed as a function of temperature variations in the current climate under the (sometimes implicit) assumption that such correlations are directly interpretable in terms of a particular climate feedback. The problem with such assertions is that the observed variability is as likely to be produced by variations in the dynamics as by intrinsic temperature dependence; in general separation of the dynamic from the thermodynamic components has not yet been done because of the absence of accurate global circulation data. Nonetheless, observed temperature dependence, whatever its cause, presents another test for GCM cloud parameterizations. Comparison of such behavior for the current climate with actual simulations of the GCM response to temperature perturbations is a first step in unraveling the dynamic and thermodynamic contributions to cloud variability.

A major unsolved problem in climate is the temperature dependence of cloud optical thickness. *In situ* data (Feigelson, 1978) and adiabatic liquid water behavior (Betts and Harshvardhan, 1987) suggest that cloud liquid water content, and by inference optical thickness, should increase monotonically with temperature. Tselioudis et al. (1992) found, however, that low clouds in the ISCCP data set exhibit this behavior only at cold temperatures, and more so over land than ocean. Elsewhere, τ decreases with temperature instead. Figure 24 shows the simulated temperature variation of low cloud optical thickness for individual GCM layers over ocean. At cold temperatures, τ increases with T; the rate of increase is greater for $T < -20°C$ than for $T > -20°C$, roughly coincident with the center of the ice-liquid transition region (Fig. 2). For $T > 10°C$, τ decreases with T instead, except at the very warmest temperatures. Sensitivity tests indicate that parameterized vertically subgrid-scale cloud physical thickness variations (equations 9, 25) are responsible for this behavior. Figure 24 applies to individual layer optical thicknesses; when total column τ is computed for low clouds with no higher clouds overhead, analogous to the view from satellite, the change from positive to negative $d(\ln \tau)/dT$ occurs near 0°C, and both the change in this quantity with latitude and the difference in land vs. ocean behavior agree fairly well with ISCCP inferences (Del Genio et al., 1995).

Ice water content for high clouds in the GCM (Fig. 25) exhibits a simpler behavior, increasing monotonically with T until leveling off for $T > -25°C$, where the liquid phase starts to become significant. The rate of increase with temperature is fairly consistent with *in situ* ice water content measurements (Heymsfield and Donner, 1990), although the data vary considerably from one region and synoptic situation to another. The GCM probably underestimates the largest ice water contents, consistent with its upper troposphere dry bias in midlatitudes. The Heymsfield and Donner data correspond to cirrus clouds rather than cumulus anvils. For comparison, we separate high clouds in Figure 25 into convective and non-convective situations. GCM "anvils" tend to have systematically higher ice water contents than other high clouds, since they have an ice water source from cumulus detrainment, but their temperature dependence is not markedly different.

It is not clear *a priori* whether such correlations are indicative of feedbacks that would occur in a climate change. As one hypothetical example of a climate change, the FANGIO intercomparison project analyzes GCM response to imposed globally uniform ±2°C changes in SST under perpetual July conditions with fixed sea ice and soil moisture. The climate sensitivity in such experiments is defined as $\lambda = (\Delta F/\Delta T_S - \Delta Q/\Delta T_S)^{-1}$, where ΔT_S is the global mean surface temperature difference between the +2°C and -2°C realizations and ΔF and ΔQ are the corresponding changes in OLR and ASR, respectively, forced by the imposed climate change; by calculating the sensitivity separately for clear skies (λ_c), cloud feedback can be estimated as λ/λ_c (cf. Cess et al., 1990). For each simulation, the GCM was run for one year, with the first 5 months devoted to spinup to the new equilibrium and the final 7 months being averaged to estimate climate changes.

Table 3 compares the sensitivity and feedback contributions for the prognostic cloud water budget parameterization with those for the previous Model II diagnostic cloud scheme. The new version of the GCM calculates clear sky quantities at each gridbox while the old version used clear gridboxes only, but the resulting effects on sensitivity and feedback are unimportant given the gross differences in the two parameterizations. Optical thickness in Model II is prescribed to decrease with height. As a result, F increases with T_s by similar amounts in clear and cloudy regions, because an increase in cloud height (which reduces OLR) in the warmer climate is accompanied by both a decrease in total cloud cover (mostly due to low and middle clouds) and an implicit decrease in column optical thickness (which increases OLR). Q increases dramatically with T_s because of the cloud cover and optical thickness decreases. The latter effect dominates, producing a large climate sensitivity and strongly positive cloud feedback ($\lambda/\lambda_c >> 1$).

The prognostic scheme behaves quite differently. Cloud cover now slightly increases with warming, and while cloud height still increases, a local increase in optical thickness at many model levels more than compensates the effect of the upward shift. Thus, in the new model there is more of an enhanced greenhouse effect (smaller $\Delta F/\Delta T_s$) in the warmer climate, but a decrease in solar absorption (negative $\Delta Q/\Delta T_s$). The latter dominates, producing a very low climate sensitivity and negative cloud feedback ($\lambda/\lambda_c < 1$).

To understand the lower sensitivity of the new parameterization, consider the zonal mean changes in cloud cover and cloud water content (Fig. 26). High cloud cover increases, especially in the tropics, in the warmer climate, but unlike the first generation of GCMs with diagnostic cloud schemes, cloud cover does not uniformly decrease with temperature outside the polar regions below the tropopause. Instead, a complex pattern of cloud cover changes results, with at least two probable causes: (1) Cloud cover increases at low latitudes near 700-800 mb, the level of the trade inversion; this may represent increased venting of boundary layer moisture by shallow cumulus, an effect likely to be captured best by mass flux cumulus parameterizations. (2) Above this level, cloudiness decreases, with the pattern of decrease lying primarily above the 0°C isotherm. This is suggestive of the increase in autoconversion produced by the Bergeron-Findeisen process (see Fig. 1), which decreases the lifetime of mixed phase clouds; as T_s increases, the level at which this process operates preferentially shifts upward and high-altitude ice production increases, causing a local cloud cover decrease. This is a direct result of the use of a prognostic cloud water parameterization. Note, however, that prognostic schemes that neglect the Bergeron process produce a midlevel cloud cover increase with warming instead (Senior and Mitchell, 1993), and thus a low sensitivity for perhaps the wrong reason.

Cloud water content changes with warming generally mirror the pattern of cloud cover changes, but the magnitude of the change is greatest in the tropical cumulus anvil region and not coincident with the location of largest cloud cover change. This suggests that climate changes in anvil microphysical properties are primarily responsible for the negative cloud feedback and low sensitivity of this version of the GCM. By comparison, low level cloud water changes are small and roughly coincident with changes in cloud cover, suggesting that the in-cloud water content response to the climate change is modest.

The extreme sensitivity of the GCM to high cloud feedbacks raises two questions. First, since the GCM is deficient in midlatitude cirrus (Fig. 11), does it underestimate sensitivity by underestimating climate changes in the greenhouse effect of these clouds? To test this proposition, we performed an experiment in which high cloud cover was artificially enhanced outside the tropics by assuming the threshold relative humidity for cloud formation to decrease from 0.6 at the equator to 0.25 at the pole for ice clouds. Climate sensitivity results for this altered version of the GCM are also listed in Table 3. The parameterization change roughly doubles midlatitude high cloud cover and increases C_l by 5 W m^{-2} globally in the current climate. But as Table 3 indicates, the change has virtually no effect on climate sensitivity. There are two reasons for this: (1) Decreases in C_s in the current climate are of similar magnitude and thus offset the longwave changes; (2) Although the mean C_l is different in the two experiments, the climate change ΔC_l is almost identical. This is evidence of our earlier assertion that validation of the mean state by itself contains no information about a GCM's response to perturbations.

The results of this experiment reinforce the notion from Figure 26 that it is the tropical anvil clouds that matter most to climate sensitivity. This raises a second question: Does Figure 26 argue for the concept of a tropical cirrus "thermostat" operating to limit the magnitude of long-term climate change? Unfortunately, the prescribed SST climate change is a poor proxy for greenhouse gas-induced climate change, which may involve changes in SST patterns as well. To explore the effect of such differences, we performed a final experiment in which SST changes were applied uniformly only outside the tropical Pacific. Within the tropical Pacific, SST changes were prescribed so as to average ±2°C in the longitudinal mean, but with the temperature change greatest in the coldest regions; details are given in Ye et al. (1995). The net effect is to drastically reduce the zonal SST gradient in the warmer climate (to about 2°C across the Pacific) while increasing the gradient in the cooler climate. There are several reasons to anticipate such behavior at least qualitatively in an actual climate change: (1) The east Pacific ocean mixed layer is shallower than its western counterpart and thus responds more quickly to perturbations; (2) The thermostat concept, if valid, would require the convective west Pacific to respond less to a perturbation than the mostly non-convective east Pacific.

The effect of the reduced SST gradient is of course to weaken the Walker circulation in the warmer climate. The moisture convergence source of anvil cloud water in the west Pacific is therefore reduced relative to the uniform SST change case, and changes in the east are not sufficient to compensate. Table 3 shows that this version of the GCM, despite identical physics to the standard prognostic version, produces a distinct positive cloud feedback and fairly large climate sensitivity instead, the biggest change being the virtual elimination of negative shortwave impacts in association with warming. Of course, this is still a prescribed SST change experiment; in a real climate change, the SST pattern, Walker cell, and anvil clouds will mutually interact to produce an equilibrium change that may differ from either of the two extremes we have examined here. But the results of these tests demonstrate two important points: (1) Climate sensitivity to greenhouse gas increases can only be determined in the context of actual climate change scenarios with coupled atmosphere-ocean models; (2) The thermostat concept is too simplistic to apply to global climate change, since tropical anvil properties depend on the general circulation and thus the SST gradient rather than merely responding to local changes in SST.

6.) Discussion

Although considerable room for improvement exists in the prognostic cloud water parameterization, its performance in the GISS GCM is encouraging in several respects, especially its ability to simulate the broad features of cloud property differences in different climate regimes, the overall sense and magnitude of variability on several different time scales, and the temperature dependence of cloud properties. Furthermore, the prognostic approach permits a diagnosis of cloud feedbacks in terms of physical processes, thus pointing the way toward strategies for both future model improvements and needed observations. A few of the more important examples are discussed below.

It is apparent from the experiments reported here that realistic simulation of tropical cumulus anvil radiative properties and their variation as climate changes is crucial for a plausible climate sensitivity estimate. The GCM reproduces observed shortwave and longwave cloud forcing variations over ENSO, suggesting at first glance that our parameterization is satisfactory. But we use equivalent spheres to calculate the scattering properties of ice clouds, and realistic ice phase functions tend to produce higher reflectances for the same ice water content (Minnis et al., 1993). Thus, our simple prescription of anvil condensate detrainment from cumulus updrafts is likely to be an overestimate. Most published microphysical and radiative observations of high clouds pertain to thin cirrus in non-convective environments; there is a clear need for more *in situ* data from thick, active anvils in convective environments, including ice water content, precipitation, crystal size, and TOA reflectance. There is also a need for remote sensing techniques capable of determining the global distribution of ice water content. Potential areas for parameterization improvements include more realistic ice phase functions, relating the fraction of cumulus condensate detrained to some measure of instability, and taking into account subgrid vertical velocities on the mesoscale in determining anvil properties (cf. Heymsfield and Donner, 1990). Observations of these quantities are difficult; statistics from cumulus ensemble models may be the only means of obtaining the needed information. Indeed, at this point it is not known whether anvils can be viewed simply as high water content extrapolations of thinner cirrus, or whether completely different parameterizations are required for anvils and other high clouds.

The GCM's climate sensitivity seems to depend less on changes in low clouds, but in part that may be a result of the parameterization's ability to produce both increases and decreases of optical thickness with temperature in different climate regimes. The regional nature of this behavior suggests that at least regional climate response may be sensitive to low cloud optical thickness changes in an actual climate change, while errors in mean low cloudiness may be important for climate drift in a coupled ocean-atmosphere model. The GCM's problems in simulating low stratus are not simple: too much in tropical convectively disturbed areas, but too little in the stable marine stratocumulus regimes west of the continents. The GISS GCM's current vertical resolution (50 mb thickness of the lowest layer) is clearly inadequate to resolve potentially important PBL cloud processes, such as detachment of the boundary layer from the surface by drizzle evaporation and solar absorption. An experimental 18-layer version of the GCM may help in this regard. However, what is really needed is for a consensus to emerge as to the relative importance of these processes and CTEI, and in the latter case, a better understanding of the instability process itself. Our model suggests that optical thickness variations of low clouds are determined primarily by physical thickness variations (Del Genio et al., 1995), but observational confirmation is lacking. Space-borne cloud radars are probably not accurate enough to resolve the required thickness variations, but *in situ* data from field experiments such as ARM and FIRE may shed light on this important result. Over land, errors in low cloudiness are primarily due to underestimates of shallow cumulus mass flux, as well as the cloud cover per unit mass flux. The radiative properties of cumulus are usually an afterthought in GCMs, because deep cumulus occupy such a small area of the globe. The same cannot be said of shallow cumulus, though; more attention needs to be paid to predicting the areal coverage of these clouds.

The cloud types mentioned thus far have already received considerable attention from theoreticians, climate modelers and observationalists. Midlatitude storm clouds have been greatly ignored by comparison, because the study of midlatitude storms is driven by weather prediction rather than climatic considerations. The GISS GCM has known deficiencies in its ability to simulate the dynamics of midlatitude baroclinic waves, so it is possible that our parameterization is adequate for nimbostratus but performs poorly due to incorrect dynamical forcing. But given the paucity of observations of the microphysical and radiative properties of these clouds, evaluation of the parameterization itself is virtually impossible. An important question is whether ice crystals in stratus and nimbostratus in cold seasons need to be parameterized differently from their high cirrus and cumulus anvil counterparts.

Polar cloudiness is currently so poorly observed that any climatological information would be a significant improvement. First order issues include the need to resolve the disagreement between satellite and surface climatologies over the sense of the seasonal cycle, and confirmation of the dominant cloud types in each season. The role of these clouds in the polar surface energy budget is obviously an important consideration for any GCM that attempts to simulate climate change.

Other needs are more general, e.g., a physical basis for predicting cloud cover as a function of subgrid-scale variations in climate parameters rather than as a simple function of relative humidity. This problem is common to all parameterizations, whether diagnostic or prognostic. The prognostic approach to parameterization is not yet preferable to the use of prescribed cloud properties if the goal is simply simulation of the mean state of the current climate, because the prognostic approach creates additional degrees of freedom and feedbacks between different parts of the system not present when fixed properties are used. This is analogous to the statement that coupled atmosphere-ocean GCMs do not yet simulate the current climate better than atmospheric GCMs bounded by prescribed SSTs. But if the goal is to predict change on any time scale, then only the prognostic approach is satisfactory, since it alone attempts to simulate the *physics* of change. The performance of our parameterization relative to diagnostic and prescribed cloud property schemes is favorable enough to conclude that prognostic cloud water should be a feature of all future climate GCMs.

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Table 1. Selected climate parameters simulated by the GCM with the prognostic cloud water parameterization. All quantities are global unless otherwise indicated.

	Annual	January	July
TOA energy balance (W m^{-2})			
Net radiation	3.6	11.6	-6.1
Absorbed shortwave	238.2	244.0	231.3
Net longwave	-234.7	-232.4	-237.4
Shortwave cloud forcing	-53.7	-57.8	-51.3
Longwave cloud forcing	16.9	16.3	17.4
Surface energy balance (W m^{-2})			
Net energy into surface	3.0	10.0	-7.2
Absorbed shortwave	172.3	176.8	166.0
Net longwave	-55.6	-54.3	-55.1
Latent heat flux	-88.8	-89.3	-91.2
Sensible heat flux	-23.7	-21.8	-25.9
Shortwave cloud forcing	-54.2	-58.6	-51.5
Longwave cloud forcing	20.8	21.9	19.7
Cloud cover (%)			
Total (global/land/ocean)	57/45/61	58/50/61	56/43/62
High	15	15	16
Middle	16	16	15
Low	47	49	46
Cloud water path (0.1 kg m^{-2})			
Liquid	0.9	0.9	0.9
Ice	1.5	1.4	1.6
Precipitation, global/land/ocean (mm d^{-1})			
Convective	2.0/1.8/2.3	2.0/1.7/2.3	2.1/2.0/2.3
Stratiform	1.0/0.9/1.1	1.0/1.0/1.1	1.0/0.8/1.1
Precipitable water (mm)	23.4	22.1	25.2
Specific humidity 2°/50°N (g kg^{-1})			
959 mb	14.8/4.6	14.6/2.9	14.6/7.3
634 mb	4.6/1.4	4.4/0.73	4.5/2.6
321 mb	0.49/0.10	0.45/0.04	0.46/0.24
Surface air temperature (°C)	14.0	12.2	15.6
Temperature 2°/50°N (°C)			
959 mb	22/4	22/-4	22/13
634 mb	5/-13	4/-21	4/-2
321 mb	-29/-45	-30/-52	-29/-36
102 mb	-75/-62	-76/-62	-74/-61

Table 2. Selected diagnostics of the general circulation in January simulations with the new prognostic cloud parameterization and the GISS Model II diagnostic cloud parameterization. All quantities are global means unless otherwise indicated.

	Prognostic	Diagnostic
Cumulus mass flux (10^9 kg s^{-1})	1361	1161
Peak N.H. streamfunction (10^9 kg s^{-1})		
Hadley cell/Ferrel cell	177/26	168/17
Diabatic heating (10^{14} W)		
Radiation	-566	-512
Moist convection	384	330
Stratiform phase changes	81	71
Surface sensible heating	111	122
N.H. poleward transport by eddies (10^{14} W)		
Dry static energy	17.9	18.1
Latent heat	11.6	10.0
N.H. upward transport by eddies (10^{14} W)		
Dry static energy	13.1	12.9
Latent heat	15.5	13.1
N.H. tropospheric energy (10^5 J m^{-2})		
Available potential (APE)	89.9	91.0
Eddy kinetic (EKE), transient/stationary	6.9/2.6	5.3/2.4
Tropospheric energy conversions, 45°N (W m^{-2})		
Generation eddy APE	3.2	3.0
by radiation	-2.8	-2.0
by condensation	2.5	1.7
by surface fluxes	3.4	3.3
APE $\rightarrow$ EKE	5.5	4.0

Table 3. TOA radiation balance and cloud forcing differences (W m^{-2}), climate sensitivity (°C-m^2 W^{-1}), and cloud feedback produced by the diagnostic cloud parameterization run in the previous Model II GISS GCM and by 3 different versions of the prognostic cloud water parameterization run in the new GISS GCM under perpetual July conditions with fixed sea ice and soil moisture and subjected to +2 and -2°C SST perturbations.

	Model II	Prognostic cloud water	Enhanced high cloud	Reduced SST gradient
ΔQ				
clear	0.00	0.51	0.57	0.54
total	4.88	-1.28	-1.09	0.41
ΔC_s	4.88	-1.79	-1.66	-0.13
ΔF				
clear	7.72	7.73	8.12	5.58
total	8.16	6.89	7.24	3.90
ΔC_l	-0.44	0.84	0.88	1.68
λ	1.23	0.49	0.49	1.07
λ/λ_c	2.37	0.88	0.91	1.45

FIGURE CAPTIONS

Fig. 1. Zonal mean frequency of occurrence (%) of Bergeron-Findeisen diffusional growth of ice crystals in July as simulated by the GCM. Dashed lines indicate the 0°C and -30°C isotherms.

Fig. 2. Probability of condensate occurring as the ice phase (%) vs. layer temperature (°C), binned at 2°C intervals, for gridpoints over land (solid line) and ocean (dashed line) as simulated by the GCM. Asterisks denote observations compiled by Feigelson (1978) over land.

Fig. 3. Zonal mean difference in January cloud water content (10^{-6} kg/kg) between the GCM control run and a sensitivity experiment with no evaporation of cloud water.

Fig. 4. As in Fig. 3 but for the difference between the control run and a sensitivity experiment with no detrainment of convective condensate from cumulus updrafts into stratiform anvil clouds.

Fig. 5. Frequency of occurrence (%) of cloud top entrainment instability between the first two model layers in the GCM in January (upper) and the mean fraction of liquid water mixed between the layers during CTEI occurrences (lower). Shading indicates occurrence frequencies > 40% in the upper panel and mixing fractions > 20% in the lower panel.

Fig. 6. Frequency histograms of cloud particle effective radius (μm) occurrence diagnosed in the GCM in January for (a, b) liquid phase low-level clouds over ocean and land, respectively, (c) ice phase low-level clouds, and (d) ice phase high-level clouds.

Fig. 7. Differences between GCM-simulated and ERBE-observed TOA absorbed shortwave (left) and outgoing longwave (right) radiation flux for January (top) and July (bottom). The GCM results are 5-year averages with climatological SST; the ERBE data cover January 1986-1989 and July 1985-1988. OLR is defined as positive in this figure.

Fig. 8. GCM-simulated shortwave cloud forcing (left) and GCM-ERBE differences (right) for January (top) and July (bottom). The GCM results are January 1985-1988 and July 1985-1988 averages from an AMIP simulation; gray areas indicate missing/excluded data.

Fig. 9. As in Fig. 8 but for longwave cloud forcing.

Fig. 10. As in Fig. 8 but for total cloud cover (%) and GCM-ISCCP differences. The GCM results are 5-year averages with climatological SST; the ISCCP data are averaged over July 1983-1989 and January 1984-1989 from the C2 data set.

Fig. 11. As in Fig. 10 but for high cloud cover; the ISCCP data are averages over 1985-1990 using visible-IR detection thresholds from the C1 data set.

Fig. 12. As in Fig. 10 but for seasonal average (Dec.-Jan.-Feb. and Jun.-Jul.-Aug.) low cloud cover and differences relative to surface cloud observations. The data are averages over 1971-1981 for land points, and over 1952-1981 for ocean points; gray areas indicate missing/insufficient data.

Fig. 13. As in Fig. 10 but for February and August cloud liquid water path and differences relative to SSM/I. The data are for 1987; gray areas indicate missing/insufficient data.

Fig. 14. Two-dimensional frequency histograms of cloud top pressure (mb) and visible optical thickness over tropical and subtropical oceans in the GCM and ISCCP C1 data. The GCM cloud properties have been binned into the same 5 optical thickness ranges reported by ISCCP. The cloud top pressure categories are different in the model and data: ISCCP reports 7 cloud top pressure categories, while the GCM figures denote the tops of the 9 model levels. The GCM histograms are 5-day averages and have been subjected to the satellite "detection" procedure

described in the text; the ISCCP data are averages for the full month of January 1984. (a) GCM, 0°-15°N ocean; (b) ISCCP, 0°-15°N ocean; (c) GCM, 15°N-30°N ocean; (d) ISCCP, 15°N-30°N ocean.

Fig. 15. As in Fig. 14 but for midlatitude land and ocean regimes: (a) GCM, 30°N-60°N land; (b) ISCCP, 30°N-60°N land; (c) GCM, 30°S-60°S ocean; (d) ISCCP, 30°S-60°S ocean.

Fig. 16. As in Fig. 14, but for polar clouds: (a) GCM, 60°N-90°N ocean, July; (b) GCM, 60°N-90°N ocean, January; (c) ISCCP, 60°N-90°N ocean, July. There are virtually no ISCCP optical thickness data for northern polar regions in January.

Fig. 17. Zonal mean January differences between the control run with the new prognostic cloud water parameterization and a sensitivity experiment using the GISS Model II diagnostic cloud parameterization. (a) Total cloud cover (%); (b) Total radiative heating rate (10^{13} W); (c) Moist convective heating rate (10^{13} W); (d) Stratiform condensation heating rate (10^{13} W).

Fig. 18. Diurnal cycle of July high cloud cover (deviation from the zonal, monthly, and daily average) as a function of latitude and local hour for a single month of the GCM and ISCCP C2 data. The ISCCP estimate uses the IR detection threshold only. (a) GCM, land; (b) ISCCP, land; (c) GCM, ocean; (d) ISCCP, ocean. Units: (%).

Fig. 19. As in Fig. 18 but for total cloud cover.

Fig. 20. As in Fig. 18 but for the seasonal cycle of high cloud cover (deviation from the zonal, annual mean) as a function of latitude and month. The ISCCP estimate uses daytime data only and a combined visible-IR detection threshold.

Fig. 21. As in Fig. 20 but for total cloud cover. The ISCCP estimate uses the visible-IR threshold

for daytime data and the IR-only threshold for nighttime data.

Fig. 22. Hovmöller diagrams of equatorial (5°S-5°N mean) OLR (left) and precipitation (right) anomalies relative to the 1979-1988 mean for the tropical Pacific basin as simulated by the GCM forced with the AMIP SSTs. OLR is defined here to be a negative quantity (cf. Table 1); positive anomalies thus denote less radiation emitted to space.

Fig. 23. April 1987 minus April 1985 and February 1988 minus February 1987 differences in C_s vs. C_l for tropical Pacific (10°S-10°N, 124°E-90°W) gridboxes simulated by the GCM forced with the AMIP SSTs.

Fig. 24. Logarithm of cloud optical thickness for individual GCM layers for low-level clouds over ocean vs. layer temperature (°C), binned into 1°C intervals, for January.

Fig. 25. In-cloud ice water content (g m^{-3}) for individual GCM layers for high-level clouds vs. layer temperature (°C), binned into 1°C intervals, for a 24-hour period in January. Asterisks denote cumulus anvils and plus signs denote other cirrus not associated with deep convection.

Fig. 26. Difference in zonal mean total cloud cover (%) and cloud water content (10^{-6} kg/kg) between GCM perpetual July runs with prescribed globally uniform +2°C and -2°C perturbations in SST.