Changes to the Moist Processes in CAM2

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Abstract.

1. include latent heat of fusion
2. cloud fraction calculation simplified
3. ice effective radius is function of $T$ only
4. separate cloud liquid and cloud ice variables
   - same ice fractions for microphysics and optics.
5. cloud liquid and ice transported and sedimentsed
   - ice velocity determined from effective radius
6. rain and snow are produced by liquid and ice, respectively
   (a) all condensation processes use same ice fraction
   (b) snow melts if $T > T_{\text{melt}}$ in any layer
   (c) rain never freezes
   (d) evaporation equation same for all processes
      - different coefficients for convective and stratiform precipitation
      - rain and snow evaporate proportionally
   (e) correction to ZM scheme so that precipitation is always equal to the integral of the net production

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1. Cloud fraction

The cloud fraction parameterization has been simplified significantly. Cloud fraction $f_c$ depends on a relative humidity threshold in the
same way as in CAM2
\[ f_c = \left( \frac{RH - RH_{min}}{1 - RH_{min}} \right)^2. \]  

However, \( RH_{min} \) does not depend on stability. At present, \( RH_{min} = 0.9 \), unless \( cM_{c2}^k > 10 \) mb/day for high cloud or \( cM_{c2}^k > 100 \) mb/day for low cloud, in which case \( RH_{min}^k = 0.8 \). Here, \( M_{c2} \) is the mass flux from the Hack convection scheme and \( c = 864 \) g converts \( \text{kg/m}^2/\text{s} \) to \( \text{mb/day} \). Note that only the low cloud term is actually important since the Hack convection scheme is not very active at upper levels. This term is mostly active in the midlatitude storm tracks.

The marine stratus parameterization based on Klein and Hartmann (1993) is also included. If the ocean fraction is greater than 0.5 and \( \min(d\theta/dp) \leq -0.125 \) K/mb below 750 mb, the stratus cloud fraction is
\[ f_s = 0.057 \times (\theta_{700} - \theta_s) - 0.5573 \quad 0 \leq f_s \leq 1, \] (2)
where \( \theta_s \) and \( \theta_{700} \) are the potential temperatures at the surface and at the midpoint nearest 700 mb. This cloud is assigned to the layer \( k_{min} \) below the interface at which \( \min(d\theta/dp) \) occurs after bounding by the relative humidity at \( k_{min} \) and the layer below.
\[ f_s = \max(f_s, \min(RH^{k_{min}}, RH^{k_{min}+1})) \] (3)
The final cloud fraction is then
\[ f^{k_{min}}_c = \max(f^{k_{min}}_c, f_s) \] (4)

2. Effective radii

The effective radius for ice clouds is now a function only of temperature, following Kristjánsson et al. (2000), as shown in Figure 1.

The effective radius for cloud water drops has been changed to 14 microns over ocean (land fraction < 0.5) and 11 microns for warm air over land. The land radius increases linearly from 11 to 14 microns over the temperature range \([-10, -30] \) C. Note that this is the same dependance as in CAM2 and does not necessarily correspond the range over which the cloud ice fraction increases from 0 to 1. (BAB)

We should really determine the land liquid radius and merge it with the ocean radius according the fractional areas.

3. Stratiform cloud ice

Cloud liquid and ice are assumed to coexist with a temperature dependent fraction
\[ f_i = \frac{T - T_{max}}{T_{min} - T_{max}} \quad T_{min} \leq T \leq T_{max} \] (5)
with \( f_i(T < T_{min}) = 1 \) and \( f_i(T > T_{max}) = 0 \). The bounds are adjustable constants with current settings \( T_{min} = -40 \) C and \( T_{max} = -10 \) C. These settings are different from both the cloud (-20, 0) and radiation parameterization (-30, -10) settings in CAM2.

Liquid and ice mass mixing ratios (\( \ell \) and \( I \)) are independently advected, diffused, and transported by convection. The detrained liquid from the ZM convection is all added to the cloud liquid, since the ZM scheme does not have an ice phase. After the convection and sedimentation (see below), the liquid and ice are recalculated from the total cloud condensate
\[ \ell_n' = (\ell_n + I_n)(1 - f_i) \] (6)
\[ I_n' = (\ell_n + I_n)f_i. \] (7)
The heating due to the change in cloud ice is
\[ Q^k = L_f \frac{I_n - I_n'}{\delta t}. \] (8)

The stratiform cloud condensate tendency is computed and partitioned according to \( f_i \).
Ice particles:
effective radius and terminal velocity

Figure 1. Top, ice effective radius versus temperature. Bottom, ice velocity versus radius (left) and temperature (right); the Stokes terminal velocity is shown in black and the actual velocity in red.
The excess heating due to cloud ice production instead of cloud liquid production is included with the evaporation and freezing of precipitation below.

The cloud microphysics are unchanged from CAM2, except that the two thresholds for converting in-cloud ice to precipitate (snow) have been doubled.

- $icrit_c = 10 \times 10^{-6}$
- $icrit_w = 8 \times 10^{-4}$

The applied threshold is $f_iicrit_c + (1 - f_i)icrit_w$.

4. Evaporation and freezing of precipitation

The evaporation of precipitation is computed for each source of precipitation using the same expressions. The flux of precipitation $F^{k+}$ on each interface is

$$F^{k+} = F^{k-} + \frac{\delta k p}{g} (P^k - E^k)$$

(9)

where $P^k$ and $E^k$ are precipitation production and evaporation, respectively. $P^k$ is determined by the convection or stratiform microphysics routines and

$$E^k = k_e (1 - c^k) \left( 1 - \min(1, \frac{q^k}{q_s^k}) \right) (F^{k-})^{1/2}$$

(10)

where $k_e$ is an adjustable constant and $c^k$ is the fractional cloud area. The $(1 - c^k)$ factor represents an overlap assumption that precipitation falling into the existing cloud in a layer does not evaporate. For convective precipitation, $k_e = 3 \times 10^{-6}$ and for stratiform precipitation, $k_e = 1 \times 10^{-5}$. Two bounds are applied to $E^k$:

1. $E^k \leq \frac{\delta k - q}{\delta t}$, to prevent supersaturation;
2. $E^k \leq \frac{P^k}{\delta t}$, to prevent $F^{k+} < 0$, note that precipitation is not permitted to evaporate in the layer in which it forms;

4.1. Snow production and evaporation

Exactly the same procedure is applied to snow,

$$F^{k+}_s = F^{k-}_s + \frac{\delta k p}{g} (P^k_s - E^k_s - M^k)$$

(11)

where $P^k_s = f_i P^k$ is the snow production, $f_i(T)$ is the ice fraction, $M^k$ is the melting rate and

$$E^k_s = E^k F^{k+}_s / F^{k-}_s$$

(12)

so snow evaporates in proportion to the fraction of snow in the precipitation flux on the upper interface.

Falling precipitation is not permitted to freeze. Snow is produced only by the assumed ice fraction $f_i$ in the production term. Snow does not melt unless it falls into a layer with $T^k > 0$ C, in which case $M^k = F^{k+}_s \frac{q}{\delta p}$ so that all the snow melts.

The net heating rate due to freezing, melting and evaporation of precipitation is

$$Q^k = -L_v E^k + L_f (P^k_i - E^k_s - M^k)$$

(13)

This is the method by which the heating due to $L_f$ is included for all condensation processes. For convective precipitation, $P^k_i \equiv P^k$, while for stratiform precipitation, $P_i = f_i C^k$ where $C^k$ is the net condensation rate in the cloud. Both the cloud ice fraction and the snow production fractions are determined by $f_i$, with $P^k_s$ coming from the cloud ice.

For stratiform precipitation, the above equations are iterated once to allow the first estimate of the heating to change $T$ and consequently $q_s$ (but not $f_i$) for the 2nd iteration.
5. Cloud sedimentation

Cloud liquid and ice particles are allowed to sediment using independent settling velocities, similar to the form described by Lawrence et al. (1998)

The liquid and ice settling fluxes are computed at interfaces, from velocities and concentrations at midpoints, using a SPITFIRE solver, Rasch and Lawrence (1998). The resulting flux at each interface is constrained to be smaller than the mass of liquid or ice in the layer above. This constraint does not allow for particles falling into the layer from above.

Sedimenting particles evaporate if they fall into the cloud free portion of a layer. No bound is applied to prevent supersaturation of the layer. This will be accounted for in the subsequent cloud condensate tendency calculation. Maximum overlap is assumed for stratiform clouds, so particles only evaporate if the cloud fraction is larger in the layer above. The overlapped fraction is

\[
f_o = \min\left(\frac{f^k_c}{f^k_c - 1}, 1\right) \tag{14}\]

5.1. Cloud particle velocities

The ice velocity \(v_i\) is a function only of the effective radius \(R_e\), which itself is a function only of \(T\). For \(R_e < 40 \times 10^{-6}\) m, the Stokes terminal velocity equation for a falling sphere is used

\[
v_i = \frac{2 \rho_w g R_e^2}{9 \eta} \tag{15}\]

where \(\eta = 1.7 \times 10^{-5}\) kg m/s is the viscosity of air and the density of air has been neglected compared to the density of water.

For \(R_e > 40 \times 10^{-6}\) m, the Stokes formula is no longer valid and we use a linear dependence of \(v_i\) on \(r = 10^{-6} \times R_e\)

\[
v_i(r) = v_i(40) + (r - 40) \frac{v_{400} - v_i(40)}{400 - 40} \tag{16}\]

where \(v_{400} = 3.5\) m/s is the assumed velocity of a 400 micron sphere, from Byers (1965), Fig 6.6.

The liquid particle velocity depends only on whether the cloud is over land or ocean, as is true of the liquid effective radius. The net liquid velocity \(v_l\) is

\[
v_l = v_l^{\text{land}} f^{\text{land}} + v_l^{\text{ocean}} f^{\text{ocean}} \tag{17}\]

where \(f^{\text{land}}\) and \(f^{\text{ocean}}\) are the land and ocean fractional areas of the cell, respectively. The ocean fraction may contain sea ice. The velocities are \(v_l^{\text{land}} = 2.8\) and \(v_l^{\text{ocean}} = 1.5\) cm/s.

References


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