Part VII
Cloudiness and Large scale precipitation

1 Cloudiness scheme

Model key LNEBN selects the present cloudiness scheme from routine ACNEBN.
Old scheme ACNEBT + ACNEBCO, controlled by keys LNEBT and LNEBCO has become obsolete.
Statistical scheme (LNEBR, ACNEBR) is reserved for climatologists.
Cloudiness has three origins: large scale over-saturation, subgrid shallow convection over-saturation, and subgrid deep convection over-saturation.

1.1 Large scale and shallow convection cloudiness diagnostic

Condensed water contents is linked to the *potential water vapour excess* $\triangle q_{\text{exc}}$:

$$\triangle q^l_{\text{exc}} \equiv q^l - r^l_c q_{\text{sat}}(T^l, p^l) \tag{1}$$

$$\ell_s = \begin{cases} \ell_{\max}(1 - e^{-\alpha_s \triangle q_{\text{exc}}/\ell_{\max}}) & \text{if } \frac{q}{q_{\text{sat}}} > r_c \\ 0 & \text{otherwise} \end{cases} \tag{2}$$

where

- ℓ_s is the condensed stratiform specific humidity.
- $r^l_c \equiv$ PHUC(JLEV) the critical minimum (mesh averaged) relative humidity producing a cloud. Its profile is computed in APLPAR and is controlling the "triggering" (whether there is a cloud or not). The bigger is HUCOE, the higher are both spatial extension of these clouds and their condensates. HUTIL controls the vertical shape of PHUC, and thus those of large scale clouds. Exponents NPCLO1 and NPCLO2 are normally equal to 1.

$$\mathsf{ZHUC} \equiv r^l_c = 1 - \mathsf{HUCOE}\ {\eta^l}^{\mathsf{NPCLO1}} \left(1 - \eta^l\right)^{\mathsf{NPCLO2}} \left(1 + \sqrt{\mathsf{HUTIL}}\left(\eta^l - 0.5\right)\right) \tag{3}$$

  so that r_c is 1 at the top and the bottom of the atmosphere, and is minimum at an intermediate level.

- α_s is a tunable function (of q_{sat}) controlling the amount of condensates, where a cloud already exists with respect to PHUC:

$$\alpha_s = \frac{\mathsf{QSSUSS}}{\sqrt{1 + (\mathsf{QSSUSV} \cdot q_{\text{sat}})^2}}$$

- $\ell_{\max} \equiv$ QSUSX is the maximum sustainable water contents

The so-called stratiform cloudiness is then

$$n_s = \alpha_{n_s}\sqrt{\frac{\ell_s}{q_{\text{sat}}}} \qquad \epsilon < n_s < 1 - \epsilon$$

where $\alpha_{n_s} \equiv$ QSNEBS is a tunable coefficient.

In order to include sub-inversion cloudiness (shallow convection), equation (1) for the *potential water vapour excess* $\triangle q_{\text{exc}}$ is replaced by the weighted vertical integral:

$$\triangle q^l_{\text{exc}} = \frac{\int_l^{l_b} (q^k - r^l_c q^l_{\text{sat}}) W(r^k) dp^k}{\int_l^{l_b} W(r^k) dp^k} \qquad r^k \equiv \min\left(\frac{s^k - s^l}{D_s}, 1\right), \qquad W(r) \equiv 1 - 3r^2 + 2r^3 \tag{4}$$

So we perform the vertical integral downwards, from current level l down to the level l_b where the difference in dry static energy reaches the tunable threshold $D_s \equiv$ QSSC.

A cubic is used for $W(r)$ instead of the more intuitive e^{-r^2}, with a null derivative for $r = 0$ and $r = 1$ (where $s^k - s^l = D_s \equiv$ QSSC). We get back equation (1) by setting QSSC $= 0.$, so that $l_b \equiv l$.
QSSC has impact only on the shallow convection clouds: it is the vertical extension expressed in dry static energy inside which plumes occur, implying that the saturation function to be used to compute cloud condensates is not that of the current level but of a lower one. QSSC > 0 enables clouds forming at the top of the boundary layer and of the Inter Tropical Convergence Zone.

1.2 Convective suspended water and cloudiness diagnostic

The condensed convective specific humidity is derived from the convective precipitation, which already cares for all subgrid information, including evaporation.

$$\begin{aligned} \ell_c &= \begin{cases} \ell_{\max}(1 - e^{-\alpha_c g \frac{\partial \mathcal{P}_c}{\partial p} \triangle t_{\text{phys}}/\ell_{\max}}) & \text{if } \frac{\partial \mathcal{P}_c}{\partial p} \geq 0 \\ 0 & \text{otherwise} \end{cases} \\ n_c &= \alpha_{n_c} \ell_c \qquad \epsilon < n_c < 1 - \epsilon \end{aligned}$$

where

- ℓ_c is the condensed convective specific humidity.
- $\alpha_c \equiv$ QSSUSC, is a tunable coefficient
- $\mathcal{P}_c$ is the convective precipitation flux
- n_c is the convective cloudiness
- $\alpha_{n_c} \equiv$ QSNEBC is a tunable coefficient

The triggering is controlled by the sign of precipitation flux divergence. QSSUSC controls then the amount of condensates. QSSUSC has linear impact on the cloud condensates until the threshold QSUSX is reached.

1.3 Final combination

Total local condensed water contents ℓ and cloudiness n write then:

$$\begin{aligned} \ell &= \ell_s + \ell_c \\ n &= n_s + (1 - n_s) n_c \end{aligned}$$

Under LNEBNXR, the total cloudiness is computed according to Xu and Randall [1996]:

$$n = \begin{cases} h_r^p \cdot \left(1 - \exp\left[-\frac{\alpha \cdot \ell}{((1-h_r)\cdot q_{\text{sat}})^\gamma}\right]\right) & \text{if } h_r < 1 \\ 1 & \text{otherwise} \end{cases}$$

where

- h_r is the relative humidity
- α, γ, p are tunable coefficients

Under LNEIGE, we compute a partition between solid and liquid phases:

$$\begin{aligned} \ell_w &= \ell(1 - f(T)) \qquad \ell_i = \ell f(T) \\ f(T) &= \begin{cases} 1 - e^{-\frac{1}{2}\left(\frac{T - T_t}{T_0}\right)^2} & \text{if } T < T_t \\ 0 & \text{otherwise} \end{cases} \end{aligned}$$

with

- $\ell_w \equiv$ PQLI is the liquid water contents
- $\ell_i \equiv$ PQICE is the solid water contents
- T_t is the water triple point temperature

- coefficient $T_0 = \mathsf{RDT} = T_t - T_x(e_w - e_i) = 11.82K$ copes for liquid and solid phase coexistence in the neighbourhood of the triple point. T_0 corresponds to the difference between T_t and the temperature of maximum difference between the two saturation curves.

PQLI and PQICE are called "radiative" liquid and solid specific moisture, as they serve as input to the radiation scheme. If LNEIGE is not set, it is supposed that all moisture is liquid: $\mathsf{PQICE} = 0$, $\mathsf{PQLI} = \ell/n_c$.

Partial cloudinesses are then derived as follows.
NTSHM is the transition level between *high* and *medium* clouds, NTSML is the transition level between *medium* and *low* clouds. High, medium, low and convective cloudinesses are

$$\mathsf{PFPLCH} = 1 - \prod_{l=1}^{\mathsf{NTSHM}} (1 - n^l) \qquad \mathsf{PFPLCM} = 1 - \prod_{l=\mathsf{NTSHM}}^{\mathsf{NTSML}} (1 - n^l)$$

$$\mathsf{PFPLCL} = 1 - \prod_{l=\mathsf{NTSML}}^{N} (1 - n^l) \qquad \mathsf{PFPLCC} = 1 - \prod_{l=1}^{N} (1 - n_c^l)$$

while the total cloudiness is

$$\mathsf{PCLCT} = 1 - \prod_{l=1}^{N} (1 - n^l)$$

Under the key LRNUMX (maximum overlap of adjacent radiative clouds), all products at levels $l > 1$ (i.e. excepting the highest model level) are performed on

$$\frac{1 - \max(n^l, n^{l-1})}{1 - n^{l-1}}$$

instead of $(1 - n^l)$.

2 Large scale precipitation scheme

Routine ACPLUIE diagnoses large scale solid and liquid precipitation fluxes.
Saturation is assessed through the difference between wet bulb and local moisture $\delta q \equiv q_w - q \equiv \mathsf{PZDQ}$. We write:

$$\mathcal{P}'_b = \max(0, \mathcal{P}_h - \frac{1}{g} \frac{\delta q}{\triangle t} dp)$$

where $\mathcal{P}_h$ is the large scale precipitation flux at the top of the considered layer, $\mathcal{P}'_b$ is the first guess for the same flux at the bottom of the same layer.

If $\delta q < 0$, there is condensation into the layer, so the first guess will become the final value; but is $\delta q > 0$, the previous equation represents the maximum evaporation into the layer, $\mathcal{P}'_b$ being the minimum value of $\mathcal{P}_b$, in the case the evaporation is performed up to the saturation of the layer.
If $\mathcal{P}_h > 0$, evaporation in unsaturated air is computed with a Kessler [1969]-type formula (revisited with some simplifications). It is derived as follows:

- The falling speed of a drop with diameter D follows the relation

  $$w = a(p, T)\ D^\alpha$$

  with function $a(p, T)$ and parameter α determined empirically.

- drop size distribution follows Marshall-Palmer law:

  $$N(D) = N_0\ e^{-\lambda D}$$

  where N_0 is determined empirically and λ characterizes the intensity of precipitation.

- Laboratory experiments show that the evaporation of a water drop in a sub-saturated atmosphere follows:

$$\frac{dM}{dt} = b(p,T)\ D^{\beta}\ \rho\ (q_{\text{sat}} - q)$$

where M is the mass of the drop and function $b(p,T)$ and parameter β are determined empirically.

Supposing spherical drops, and v_ℓ the specific volume of liquid water, the precipitation flux is

$$\mathcal{P} = \int_0^\infty a(p,T)\ D^{\alpha}\ N_0\ e^{-\lambda D}\ \frac{\pi D^3}{6 v_\ell}\ dD = \Gamma(\alpha+4)\ \frac{a(p,T)\ N_0\ \pi}{6\ v_\ell\ \lambda^{\alpha+4}} \tag{5}$$

with the Γ distribution

$$\Gamma(r) \quad = \quad \int_0^\infty x^{r-1} e^{-x} dx \quad , \qquad r \in I\!R \quad , \qquad\qquad \Gamma(n) = (n-1)! \quad , \quad n \in I\!N$$

and the precipitation divergence linked to evaporation becomes

$$\frac{d\mathcal{P}}{dz} = \int_0^\infty b(p,T) D^{\beta}\ \rho\ (q_{\text{sat}} - q)\ N_0\ e^{-\lambda D} dD = \rho\ (q_{\text{sat}} - q)\ \frac{b(p,T)\ N_0\ \Gamma(\beta+1)}{\lambda^{\beta+1}}$$

In pressure coordinates:

$$\frac{d\mathcal{P}}{dp} = \frac{b(p,T)\ N_0\ \Gamma(\beta+1)}{g\ \lambda^{\beta+1}}\ (q_{\text{sat}} - q) \tag{6}$$

Eliminating λ between (5) and (6):

$$\frac{d}{dp}\mathcal{P}^{1-\frac{\beta+1}{\alpha+4}} = \left(1 - \frac{\beta+1}{\alpha+4}\right) \frac{b(p,T) N_0 \Gamma(\beta+1)}{g} \left(\frac{6\ v_\ell}{\Gamma(\alpha+4) a(p,T) N_0 \pi}\right)^{\frac{\beta+1}{\alpha+4}} (q_{\text{sat}} - q) = A(p,T)(q_{\text{sat}} - q)$$

Considering a standard atmosphere, $T(p)$ profile is determined, and

$$A(p,T) = A(p) = A(p_0)\left(\frac{p_0}{p}\right)^{2.04} \sim A(p_0)\left(\frac{p_0}{p}\right)^{2}$$

Measures yield

$$1 - \frac{\beta+1}{\alpha+4} = 0.452 \sim 0.5$$

Previous approximations are equivalent to keeping measured α, imposing β and tuning the function $b(p,T)$. We have then

$$\frac{d\sqrt{\mathcal{P}}}{dp} = -\frac{C_{\text{evap}}}{p^2}(q_{\text{sat}} - q)$$

where coefficient C_{evap} depends on the cloud type; the bigger C_{evap}, the more outside cloud evaporation occurs. For stratiform clouds, $C_{\text{evap}} \sim 4.8 \cdot 10^6$.

In Arpège-Aladin, precipitation evaporation is estimated with

$$\begin{aligned} \frac{\partial\sqrt{\mathcal{P}}}{\partial 1/p} &= \text{Evap} \cdot (q_w - q) \\ \Longrightarrow \sqrt{\mathcal{P}_b} &= \sqrt{\max(\mathcal{P}_h, \mathcal{P}'_b)} - \text{Evap} \cdot \max\left(0, (q_w - q)\right) \left(\frac{1}{p_h} - \frac{1}{p_b}\right) \end{aligned}$$

The scheme addresses snow by writing

$$\text{Evap} \quad = \quad C_{\text{evap}}(1 - r'_{me}) + C_{\text{evap}} \cdot \mathsf{RV} \cdot r'_{me}$$

where RV is the ratio of evaporation speeds of snow and liquid water.

We consider the (provisional) snow ratio r_{me} (r'_{me}) for evaporation and melting computation, distinct from the snow ratio in the precipitation fluxes r_{mi} (r'_{mi}): for thermodynamic properties, we assume a brutal transition at the triple point, in order to remain consistent with the saturations functions; for "mechanical" properties, as fall-speed depending rates of evaporation and melting/freezing, a temperature dependence is introduced, through function f below.

- in case of evaporation, the snow to water ratio remains constant:

$$r'_{mi} = r^h_{mi} \qquad r'_{me} = r^h_{me}$$

- in case of condensation, we must consider if the temperature is warmer or colder than the triple point T_t:
  - if $T > T_t$, condensation can only occur as liquid and the snow to water ratio decreases:

$$r'_{mi} = r^h_{mi}\frac{\mathcal{P}_h}{\mathcal{P}_b} \qquad r'_{me} = r^h_{me}\frac{\mathcal{P}_h}{\mathcal{P}_b}$$

  - if $T < T_t$, condensation occurs as solid, and the snow to water ratio increases:

$$\begin{aligned} r'_{mi} &= 1 - (1 - r^h_{mi})\frac{\mathcal{P}_h}{\mathcal{P}_b} \qquad r'_{me} = f(T) - (f(T) - r^h_{me})\frac{\mathcal{P}_h}{\mathcal{P}_b} \\ f(t) &= \begin{cases} 1 - e^{-\frac{1}{2}\left(\frac{T-T_t}{T_0}\right)^2} & \text{if } T < T_t \\ 0 & \text{otherwise} \end{cases} \end{aligned}$$

  Function $f(T)$ was used in ACNEBN (see above).

- Melting (if $T > T_t$) or freezing (if $T < T_t$) is obtained by

$$\begin{aligned} \frac{\partial r_{mi}}{\partial 1/p} &= \text{Melting} \cdot \frac{T - T_t}{\sqrt{\mathcal{P}}} = \frac{\partial r_{me}}{\partial 1/p} \\ \text{with Melting} &= C_{\text{melting}}(1 - r'_{me}) + C_{\text{melting}} \cdot \mathsf{RV} \cdot r'_{me} \end{aligned}$$

  discretized in

$$r_{mi} = r'_{mi} + \text{Melting} \cdot \frac{(T_t - T)(\frac{1}{p_h} - \frac{1}{p_b})}{(\sqrt{\mathcal{P}_h} + \sqrt{\mathcal{P}_b})/2} \qquad r_{me} = r'_{me} + \text{Melting} \cdot \frac{(T_t - T)(\frac{1}{p_h} - \frac{1}{p_b})}{(\sqrt{\mathcal{P}_h} + \sqrt{\mathcal{P}_b})/2}$$

r_{mi} gives us the final partition between liquid and solid precipitation:

$$\mathsf{PFPLSL} = \mathcal{P}_b(1 - r_{mi}) \qquad \mathsf{PFPLSN} = \mathcal{P}_b r_{mi}$$

and the corresponding enthalpy fluxes are obtained by multiplying by the corresponding precipitation flux latent heats, respectively

$$\mathsf{PFHPSL} = \mathsf{PFPLSL} \cdot \mathsf{PLHL} \qquad \mathsf{PFHPSN} = \mathsf{PFPLSN} \cdot \mathsf{PLHN}$$

References

E. Kessler . *On distribution and continuity of water substance in atmospheric circulations*, volume 10 n^r 32 of *Met. Mon.* Americal Met. Soc., Boston, 1969. 84pp.

F. Ruiz. Paramétrisation convective avec nébulosité. Rapport de stage, Météo France, CNRM, Juin 1998. DEA d'Océanologie et Météorologie.

Kuan-Man Xu and David A. Randall. A semi-empirical cloudiness parameterization for use in climate models. *J. Atm. Sci.*, 53(21):3103–3119, 1996.