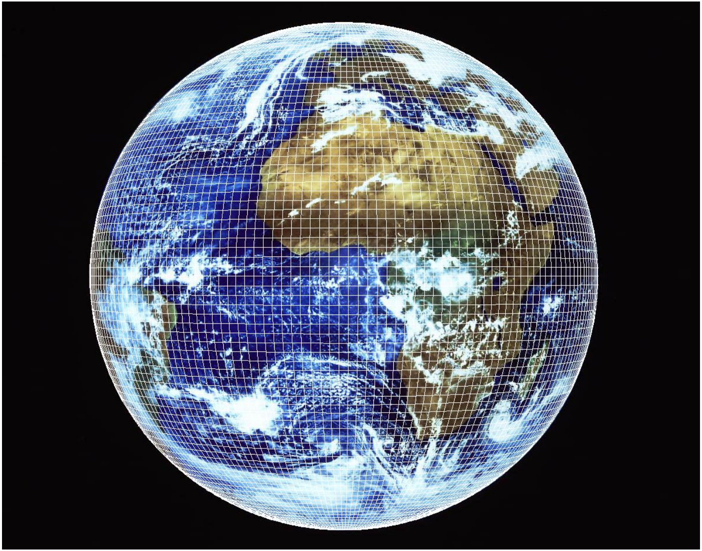
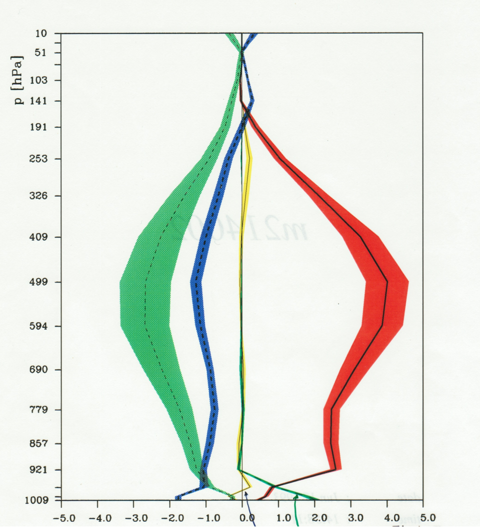
Parameterization of Convection



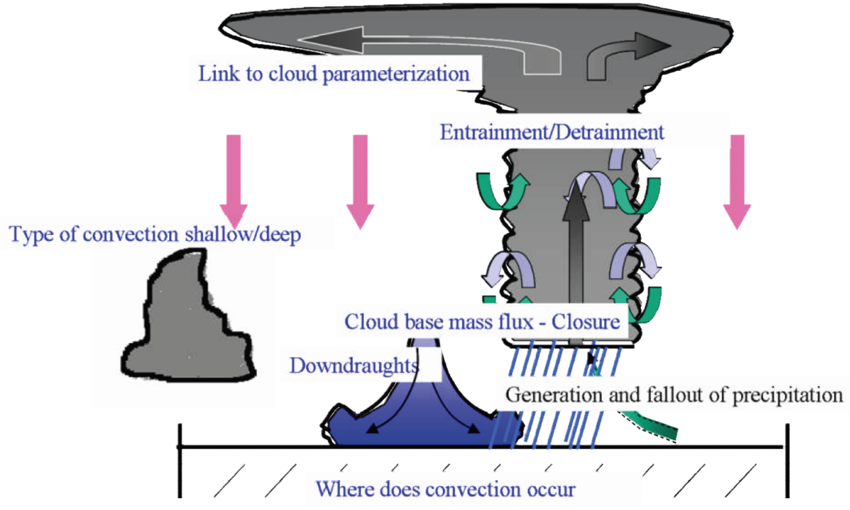
Components of the heat budget equation



after Lohmann and Roeckner, JGR, 1995

- Convection (sum of convective transport and condensation)
- Phase changes in stratiform clouds
- Radiation
- Boundary layer eddy flux
- Large-scale advection

Schematic of a GCM convection scheme



Bechtold, ECMWF lecture

Types of convection

- Shallow convection: non-precipitating boundary layer clouds, e.g. trade wind cumuli under a subsidence inversion, in which downdrafts and precipitation can be neglected
- Deep or penetrative convection: Most convection schemes are developed for these clouds. They precipitate and downdrafts are important
- Mid-level convection: Convection originating at levels above the boundary layer, e.g. rainbands at warm fronts and cumulus clouds that form in the warm sector of extratropical cyclones

Allow only one type of convection in a given grid cell at one time step

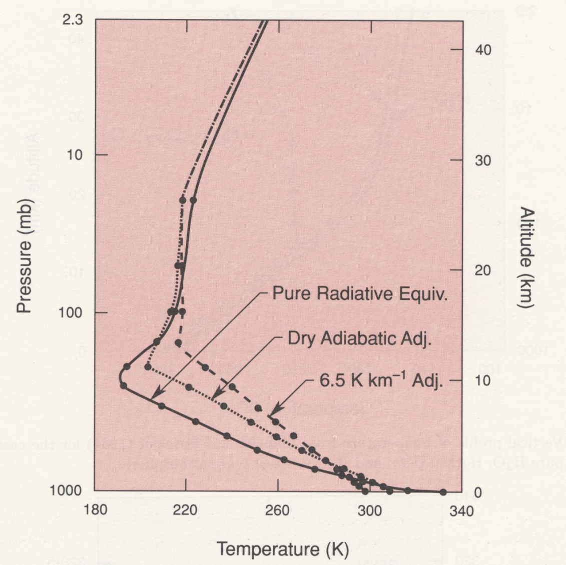
Triggering of convection

- The thermal structure resulting from pure radiative equilibrium

$$c_p \frac{\partial T}{\partial t} = Q_{rad} \tag{1}$$

  results in a lapse rate which is even steeper than the dry adiabatic lapse rate Γ_d (see next slide)
- Thus, radiation destabilizes the atmosphere, causing convection to occur
- In a dry atmosphere the equilibrium thermal profile would follow Γ_d
- In most places in the Earth's atmosphere, cloud formation complicates the equilibrium thermal profile

Different lapse rates (Fig. 10.5 Trenberth, 1992)



- Assuming radiative convective equilibrium
- Vertical temperature profile agrees well with the observed thermal structure of the atmosphere with a tropopause at 15 km

Closure assumptions

- "Closure" means to link convective processes to grid-scale processes
- Start with the prognostic equations of dry static energy, $s \equiv c_p T + \Phi$ (Φ = geopotential height; c_p =heat capacity at constant pressure) and the specific humidity q:

$$\frac{\partial \overline{s}}{\partial t} + \vec{v} \cdot \nabla s + \overline{\omega} \frac{\partial \overline{s}}{\partial p} = -\frac{\partial \left(\overline{\omega' s'} \right)}{\partial p} + L(\overline{C} - \overline{E}) + Q_{rad} \qquad (2)$$

$$\frac{\partial \overline{q}}{\partial t} + \vec{v} \cdot \nabla q + \overline{\omega} \frac{\partial \overline{q}}{\partial p} = -\frac{\partial \left(\overline{\omega' q'} \right)}{\partial p} - (\overline{C} - \overline{E}) \qquad (3)$$

- where C = rate of condensation, E = rate of evaporation and small-scale horizontal transport terms are neglected
- Determination of Q_{rad} in the next parameterization lecture.
- Terms on the left hand side (LHS): resolved by the global model.

Closure and types of parameterizations

Closure:

- Subgrid-scale processes on the RHS must be parameterized
- First terms on the RHS: Subgrid-scale vertical transport of heat and moisture and latent heat release/consumption are the basis for the cumulus parameterization.

Types of parameterizations

- Adjustment schemes (e.g. Manabe et al., 1965)
- Mass flux schemes (e.g. Arakawa and Schubert, 1974; Emanuel, 1991; Tiedtke, 1989)

Main questions:

- How is convection initiated? Using CAPE or moisture convergence?
- If a cumulus mass flux is used, how is the flux determined?

Moist adiabatic adjustment scheme

- Simplest and oldest approach (Manabe et al., 1965).
- Based on the observation that convection tends to adjust the vertical temperature profile to the moist adiabat
- Conserves the column integrals of s, q
- The adjustment procedure cools the surface and warms the upper troposphere, i.e. mixes warmer air near the surface up through the atmosphere.
- In dry atmosphere, adjust to the dry adiabat Γ_d.
- In moist atmosphere, adjust to the moist adiabat Γ_m.

Procedure of the moist adiabatic adjustment

- Perform moist adiabatic adjustment in any layer with $\Gamma > \Gamma_m$ and $q > q_s$ (saturation specific humidity).
- The excess moisture in the layer is rained out, hence q is adjusted as well.
- Since the combined energy $c_p \Delta T + L\Delta q$ must be conserved, the following condition has to be satisfied:

$$\int (c_p \Delta T + L\Delta q) dp = 0. \tag{4}$$

- Iterative process:
  1. adjust layers in a column
  2. check integral constraint above
  3. re-adjust column until the above constraint is met.

Advantages/limitations of adjustment schemes

Advantages:

- Very simple scheme (easy to implement, not expensive)
- Based on observations

Disadvantages:

- Instantaneous adjustment, no life cycle of convective clouds
- Adjustment is local and does not depend on large-scale processes or surface processes

Overcome some disadvantages by using a "soft" adjustment that includes a time scale and a reference profile other than Γ_m

Mass flux approach

Most popular way of approximating convection in today's global models.
Assumptions:

- A grid box is populated by a range of cumulus clouds.
- Each cloud transports heat s and moisture q.
- Collective behaviour of this spectrum of clouds can be represented by a bulk cumulus cloud.
- The cloud mass flux M_c is defined as the amount of air transported in the vertical direction of the cloud.
- Trigger convection using CAPE and overcoming CIN

Parcel model for deep convection

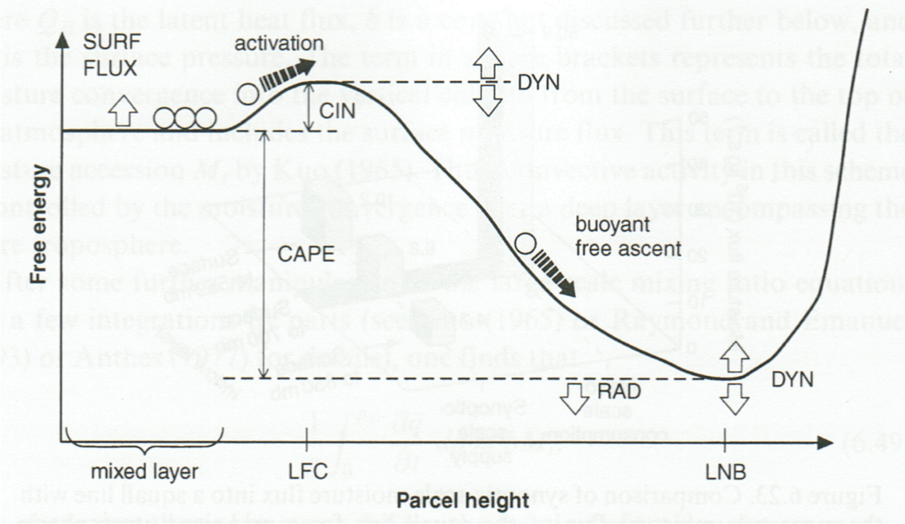


Fig. 6.24, Stensrud, Parameterization schemes, Cambridge Univ. Press, 2007

Mass flux

- M_c is given by the vertical velocity ω in the p-system times the cloud fractional area A_c:

$$M_c = -A_c\omega_c \tag{5}$$

- The mass flux in the surrounding environment M_e is given as:

$$M_e = -(1 - A_c)\omega_e \tag{6}$$

- The total mass flux M for the large-scale region is given as:

$$\overline{M} = M_c + M_e \tag{7}$$

- Obtain the subgrid-scale flux of heat and moisture from:

$$-\overline{\omega' s'} = M_c(s_c - \overline{s}) \tag{8}$$

$$-\overline{\omega' q'} = M_c(q_c - \overline{q}) \tag{9}$$

Mass flux schemes

- Assume that s and q of the environment can be approximated by the large-scale (predicted) values of $\overline{s}$ and $\overline{q}$
- Determine M_c, s_c and q_c from balance equations for mass, heat, moisture, and cloud liquid water.
- Mass balance equation and moisture budget:

$$\frac{\partial M_c}{\partial p} = \epsilon - \delta \; ; \; \frac{\partial (M_c q_c)}{\partial p} = \epsilon \overline{q} - \delta q_c - C \tag{10}$$

ϵ = entrainment rate of environmental air into the cloud

δ = detrainment rate of cloud air into the environment

C = condensation rate.

- Example for cloud base mass flux (Tiedtke, 1989):

$$M_c(p_b) = -\frac{1}{g} \int_{p_b}^{p_s} \nabla_h \cdot \vec{v} q dp + E_s - C \tag{11}$$

E_s = evaporation of moisture at the surface

Balance equations for mass flux schemes

- 4 balance equations: mass, heat, moisture and liquid water
- These equations depend on: entrainment ϵ, detrainment δ, condensation, precipitation formation
- These processes must be parameterized.
- ϵ and δ are often used for "tuning"
- E.g., $\epsilon =$ const., and $\delta \neq 0$ only at cloud top.
- Closure for the mass flux scheme: specification of the cloud base mass flux $M_c(p_b)$
- Assuming a quasi-equilibrium:

$$\frac{\partial \mathrm{CAPE}}{\partial t} \approx 0 \tag{12}$$

i.e. generation of CAPE by large-scale processes is balanced by the consumption of CAPE by convection

Advantages/limitations of mass flux schemes

Advantages:

- More physically based than adjustment schemes
- They can be extended to include downdrafts or a spectrum of clouds (Arakawa and Schubert, 1974)
- Once M_c is determined it can also be used to transport chemical tracers or momentum

Limitations:

- Lack of knowledge concerning the fundamental cloud properties
- Processes like mesoscale circulations, life cycle of convection, interaction of shallow and deep convection are still missing in most schemes
- Needs to be adjusted for convection in extratropical frontal systems which is driven by wind shear instead of by buoyancy

Trigger in ECHAM5: Moisture convergence

Shallow or deep convection
Necessary condition: moisture convergence

$$\int_{sfc}^{top} \left[\left(\frac{\partial q}{\partial t} \right)_{adv} + \cancel{\left(\frac{\partial q}{\partial t} \right)_{vdif}} \right] \frac{dp}{g} > 0$$

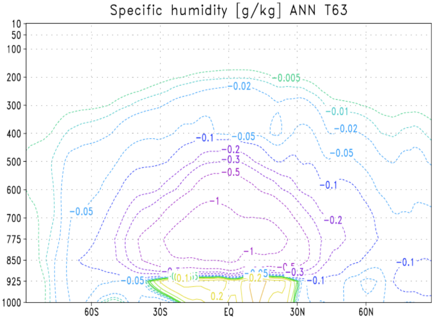
Moisture budget below cloud base

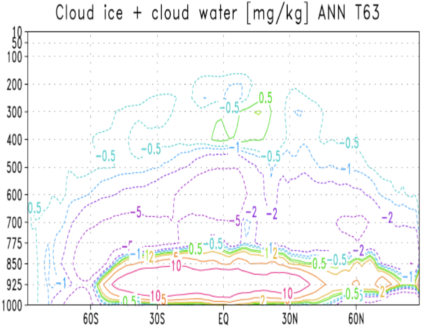
$$\int_{sfc}^{cbase} \left[\left(\frac{\partial q}{\partial t} \right)_{adv} + \cancel{\left(\frac{\partial q}{\partial t} \right)_{vdif}} + \left(\frac{\partial q}{\partial t} \right)_{conv} \right] \frac{dp}{g} = 0$$

Omit vertical diffusion tendencies for closure for convection, but account for vertical diffusion only after the convection scheme

Roeckner, pers. comm., 2010

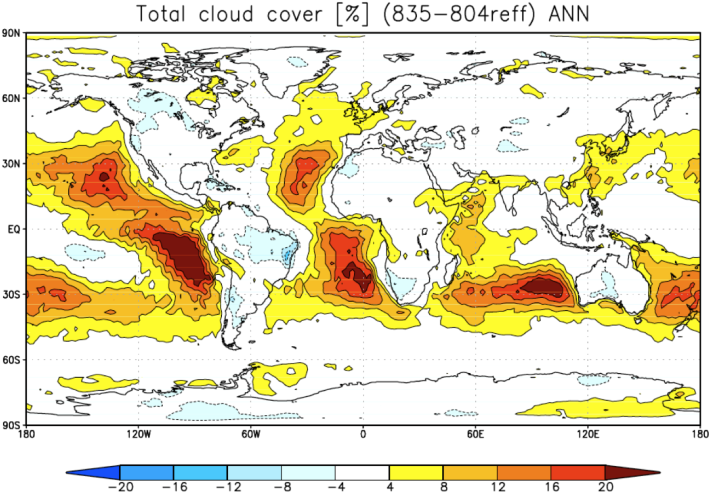
Impact on specific humidity and cloud water (no vdiff - standard)





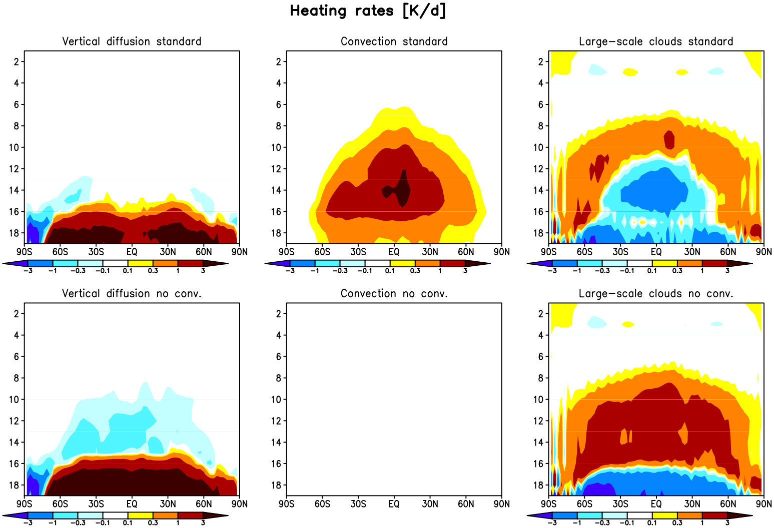
Roeckner, pers. comm., 2010

Impact on cloud cover (no vdiff - standard)

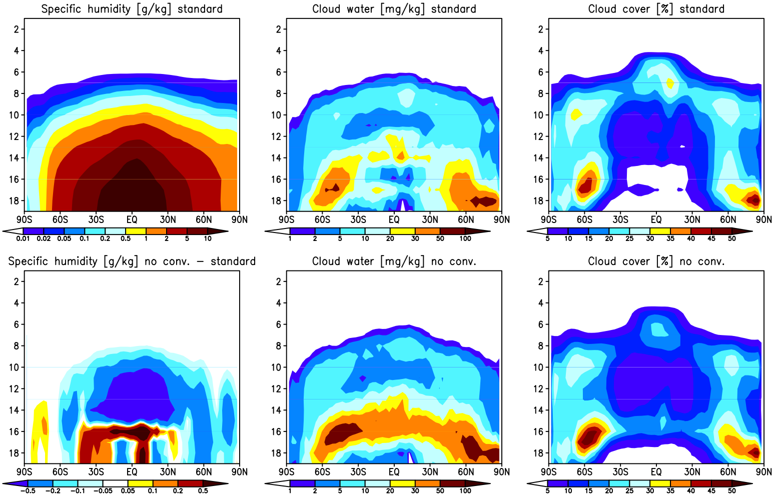


Roeckner, pers. comm., 2010

What happens without a conv. param.?



What happens without a conv. param.?



Superparameterisation/Multi-scale modeling

Idea: In each GCM column insert a 2D cloud-resolving model (CRM) with 2-5 km horizontal resolution and cyclic boundary conditions (Grabowski, 2003; Randall et al., 2003) $\Rightarrow$ cloud processes are resolved at higher resolutions

Advantages:

- More realistic representation of convection processes
- (Maybe) more realistic climate runs

Disadvantages:

- High computational costs
- Parameterisation issue remains (gap at other scale, only "convection permitting" not "convection resolving")
- Different stages of complexity