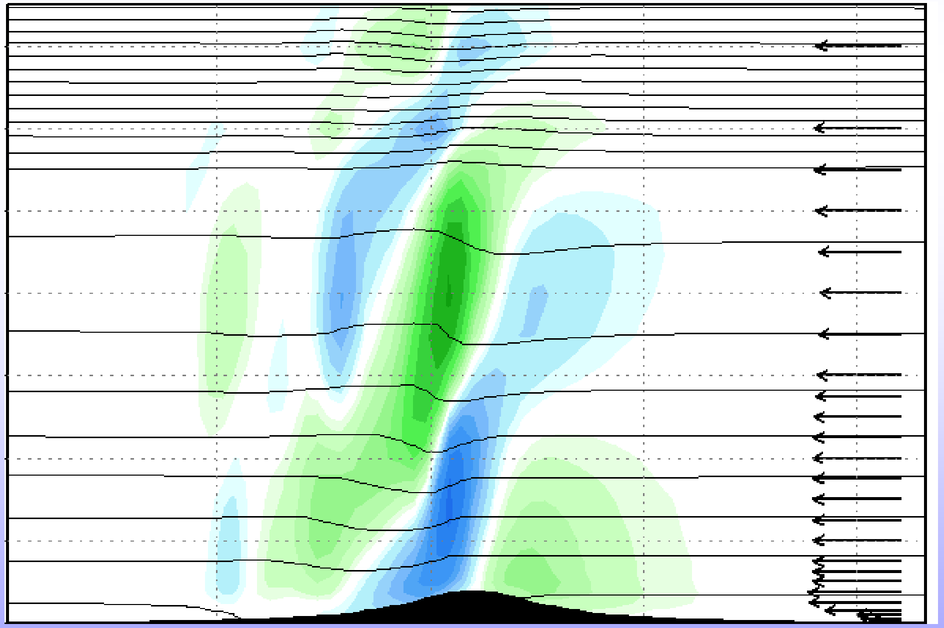
Generation of 3D hydrostatically balanced fields for use in artificial flow simulations

Janek Zimmer

Institute for Meteorology,
University of Leipzig

Ronny Petrik

Max Planck Institute for Meteorology,
Hamburg



Motivation

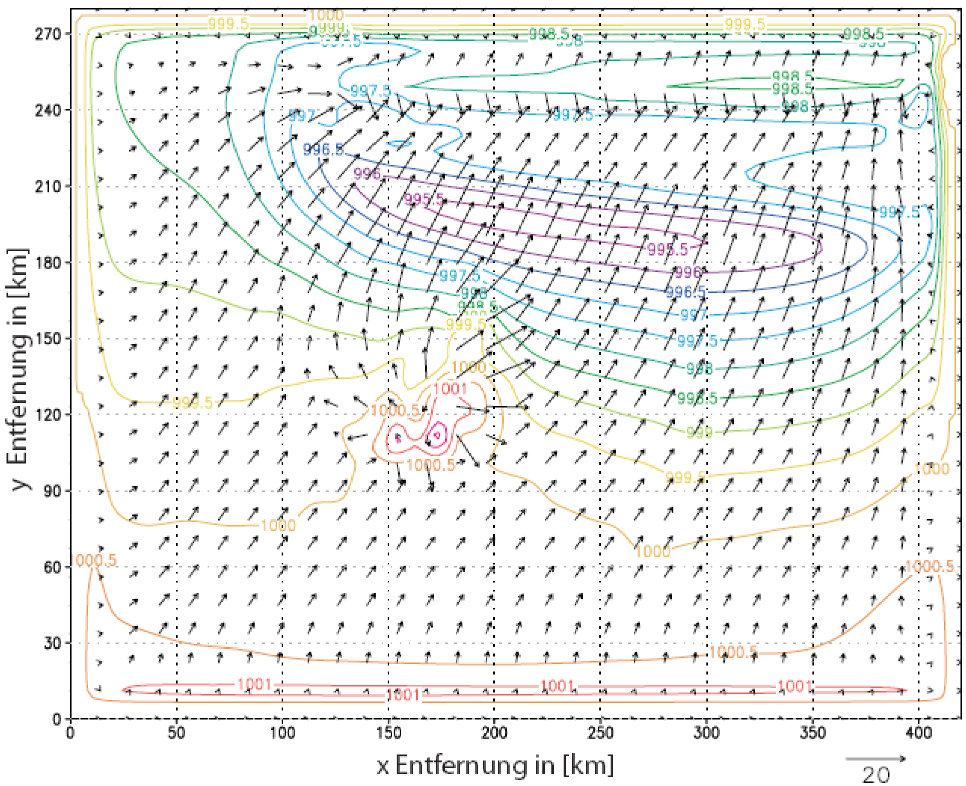
Numerical simulations with ***artificial*** initial and boundary conditions

- to evaluate model behaviour for given atmospheric flow in comparison to
  analytic solutions or
  cases in literature
- to test new numerical schemes while developing the model
- to isolate effects interacting in realistic flows, such as
  - flow modification by orographically structured terrain and synoptic-scale disturbances
  - individual convective cells and cell complexes, e.g. interaction with boundaries (see Bluestein & Weisman (2000) among others)
- to perform sensitivity studies of flow response to
  - atmospheric stratification
  - wind speed and -shear
  - orographic structure
  - namelist parameters (grid spacing etc.)

Motivation: Problem

current LM code:

if Coriolis force terms are activated (*lcori=.TRUE.*),
NO horizontal pressure gradient is applied !



convective environment ($CAPE \approx 2500 Jkg^{-1}$), moderate vertical shear ($\approx 10 ms^{-1}$), temperature disturbance initialised to promote convective initiation (Weisman & Klemp, 1982)

→ flow strongly distorted (pressure gradient established within model interior)
→ no bubble cell sensitivities deducible

Figure:
Surface pressure (coloured contours), wind vectors at lowest model level after 3h

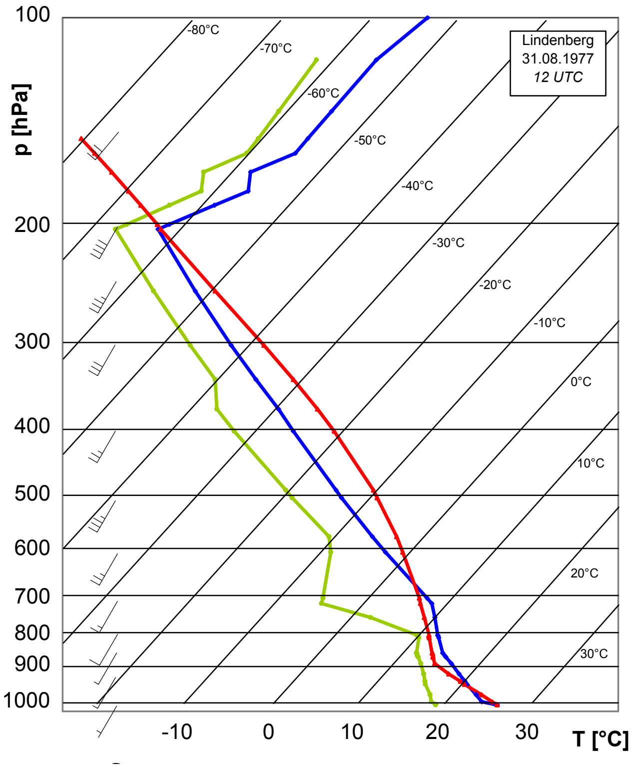
Initialisation procedure: strategy

Generate horizontal pressure field that satisfies

1. geostrophic wind relation
2. wind profile in frictional layer
3. hydrostatic equilibrium in every grid point

out of single vertical profile of
- temperature T
- specific humidity q_v
- wind components u, v
- and surface pressure p_s

either adopted from radiosonde sounding ⟶
or purely artificial



<u>by 3 steps:</u>

- 1D hydrostatic pressure routine
- horizontal pressure gradient according to geostrophic wind relation
- horizontal temperature adjustment

Initialisation procedure

FIRST STEP:

compute vertical pressure distribution (input profile),
according to hydrostatic equation (adopted from LM source code):

decompose pressure into $p = p_0 + p'$ so that:

$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g \qquad \longrightarrow \qquad 0 = -\frac{1}{\rho}\frac{\partial p'}{\partial z} + g\frac{\rho_0}{\rho}\left[\frac{T_v - T_0}{T} - \frac{T_0 p'}{T p_0}\right]$$

→ hydrostatic equation for pressure disturbances,
with the buoyancy term:

$$B = g\frac{\rho_0}{\rho}\left[\frac{T_v - T_0}{T} - \frac{T_0 p'}{T p_0}\right]$$

virtual temperature contribution — pressure contribution

→ final equation for pressure disturbance at level *k-1*:

$$p'^{k-1} = \frac{p'^{k}\left(1 - \frac{1}{2\rho_0^k R T^k}\delta p_0^{k-1}\right) + \frac{1}{2}\left[\frac{T^k(1+0.608q_v^k - q_c^k) - T_0^k}{T^k}\delta p_0^{k-1} + \frac{T^{k-1}(1+0.608q_v^{k-1} - q_c^{k-1}) - T_0^{k-1}}{T^{k-1}}\delta p_0^k\right]}{1 + \frac{1}{2\rho_0^{k-1} R T^{k-1}}\delta p_0^k}$$

Initialisation procedure

SECOND STEP:

apply horizontal pressure gradient
according to geostrophic wind relation:

$$\frac{1}{\rho}\frac{\partial p}{\partial x} = fv \qquad \text{and} \qquad \frac{1}{\rho}\frac{\partial p}{\partial y} = -fu$$

starting above the friction layer (expressed by Ekman height H_E).

In case of frictional force (roughness length z_0>0):
wind speed and direction within Ekman layer derived from

$$f(v - v_g) = K_M \frac{d^2 u}{d^2 z} \qquad f(u - u_g) = K_M \frac{d^2 v}{d^2 z}$$

→ Ekman spiral:

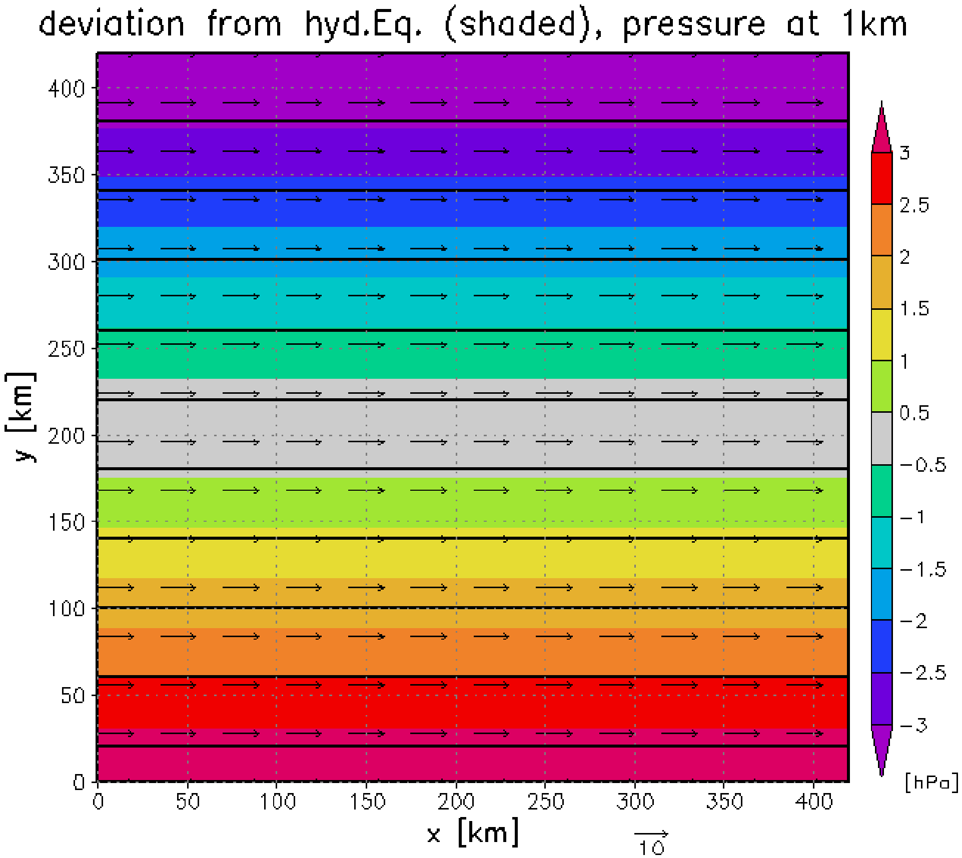
$$u(z) = u_g \left\{ 1 - \sqrt{2} \sin\alpha_0 \exp\left(\sqrt{\frac{f}{2K_M}} z\right) \cos\left(\sqrt{\frac{f}{2K_M}} z + \frac{\pi}{4} - \alpha_0\right)\right\}$$

$$v(z) = u_g \sqrt{2} \sin\alpha_0 \exp\left(\sqrt{\frac{f}{2K_M}} z\right) \sin\left(\sqrt{\frac{f}{2K_M}} z + \frac{\pi}{4} - \alpha_0\right) .$$

Initialisation procedure

***after SECOND STEP**:*

geostrophic balance



Test of hydrostatic balance for profile with strongly sheared (6 *ms^{-1}/km*) westerly flow by

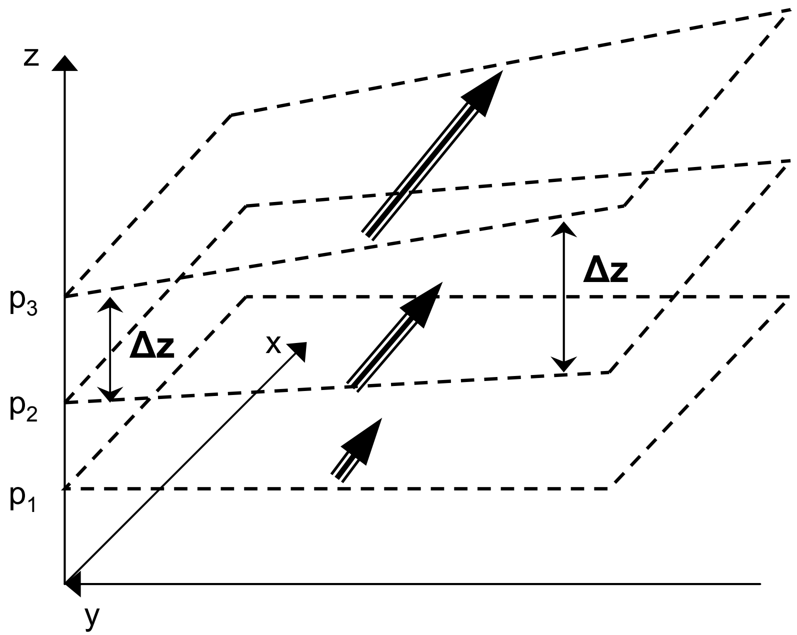
difference between new geostrophic pressure p_{geo} and p_{hydEQ} for every grid point

→ only initial profile (domain center) hydrostatic

Figure:
Pressure (solid contours) and wind vectors at z=1km, deviation from hydrostatic equilibrium (shaded)

Initialisation procedure

hydrostatically unbalanced state due to stretched / shrinked pressure thickness between model levels:



in case of shear:

pressure change with height not constant in the horizontal plane

→ layer-mean **density** has to change as well according to hydrostatic equation!

→ thermal wind relation

→ adjustment of horizontal temperature field required

Initialisation procedure

***THIRD STEP**:*

adjust density distribution (via virtual temperature) assuming that:

if domain average is in hydrostatic balance: $$\frac{\partial \overline{p}}{\partial z} = -g\overline{\rho}$$

then horizontal pressure deviation *p‘* is as well: $$\frac{\partial p'}{\partial z} = -g\rho'$$

(Holton, 1992)

→ derive *ρ‘* (via T_v‘) by inversion of 1D hydrostatic adjustment procedure:

$$p'^{k-1} = \frac{p'^{k}\left(1 - \frac{1}{2\rho_0^k R T^k}\delta p_0^{k-1}\right) + \frac{1}{2}\left[\frac{T^k(1+0.608q_v^k - q_c^k) - T_0^k}{T^k}\delta p_0^{k-1} + \frac{T^{k-1}(1+0.608q_v^{k-1} - q_c^{k-1}) - T_0^{k-1}}{T^{k-1}}\delta p_0^k\right]}{1 + \frac{1}{2\rho_0^{k-1} R T^{k-1}}\delta p_0^k}$$

→ *p‘* known for all levels, temperature T^k and specific humidity q_v^k from lower levels as well

→ solve for T^{k-1} and q_v^{k-1} using iteration

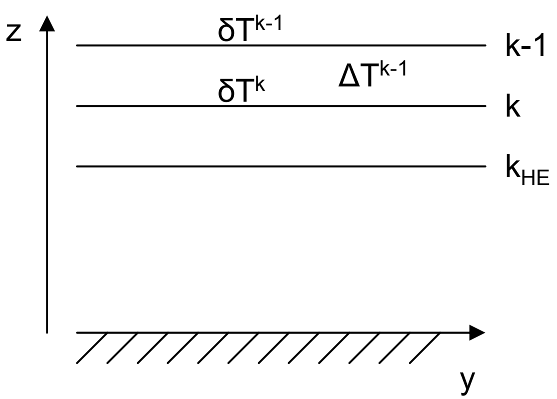
Initialisation procedure

Inversion for temperature yields:

$$T^{k-1} = \frac{-\frac{\delta p_0^k p'^{k-1}}{2\rho_0^{k-1} R} - \frac{1}{2} T_0^{k-1} \delta p_0^k}{p'^{k-1} - \frac{1}{2}(1 + 0.608 q_v^{k-1} - q_c^{k-1}) \delta p_0^k - f^k}$$

$$f^k = p'^k \left(1 - \frac{1}{2\rho_0^k R T^k} \delta p_0^{k-1}\right) + \frac{1}{2} \frac{T^k(1 + 0.608 q_v^k - q_c^k) - T_0^k}{T^k} \delta p_0^{k-1}$$

iterative procedure inevitable, since q_v^{k-1} depends on T^{k-1} !



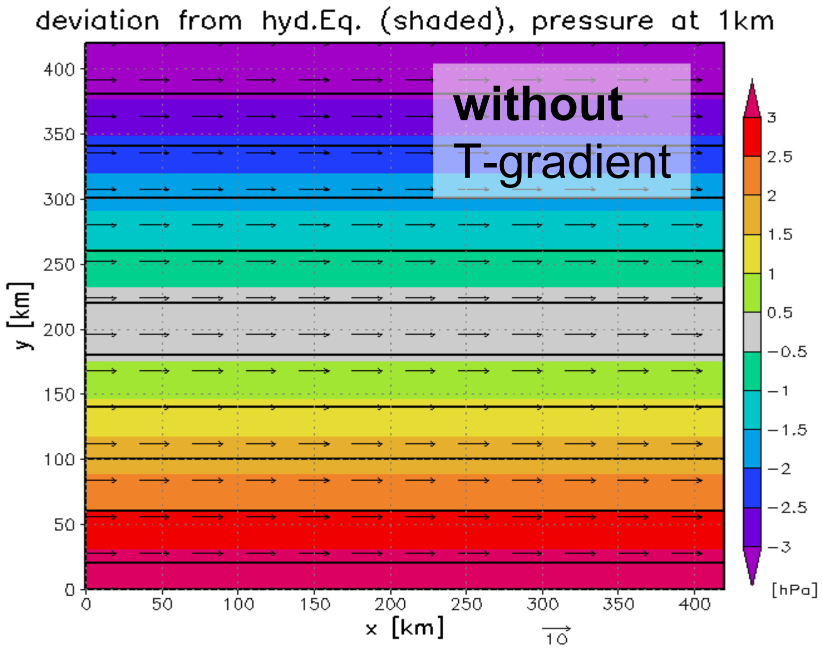
- start at Ekman height H_E (above friction layer), working upwards

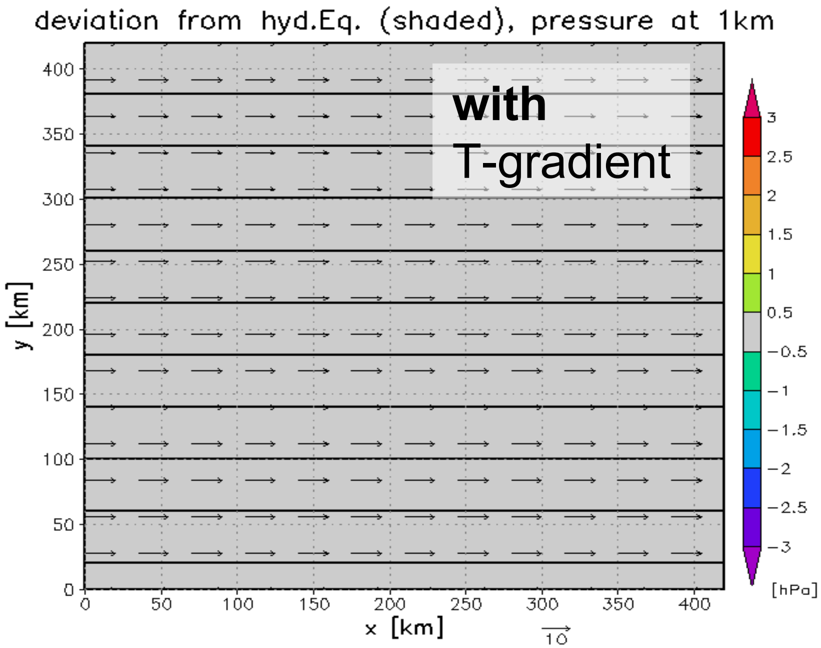
- horizontal temperature disturbance ΔT^{k-1} related to layer between levels k and k-1
→ average value $(\delta T^k + \delta T^{k-1})/2$ equals ΔT^{k-1}

- within Ekman layer: ΔT constant (as well as ∇p)

Initialisation procedure

Result of horizontal density adjustment:
3D hydrostatically balanced flow: strongly sheared test case

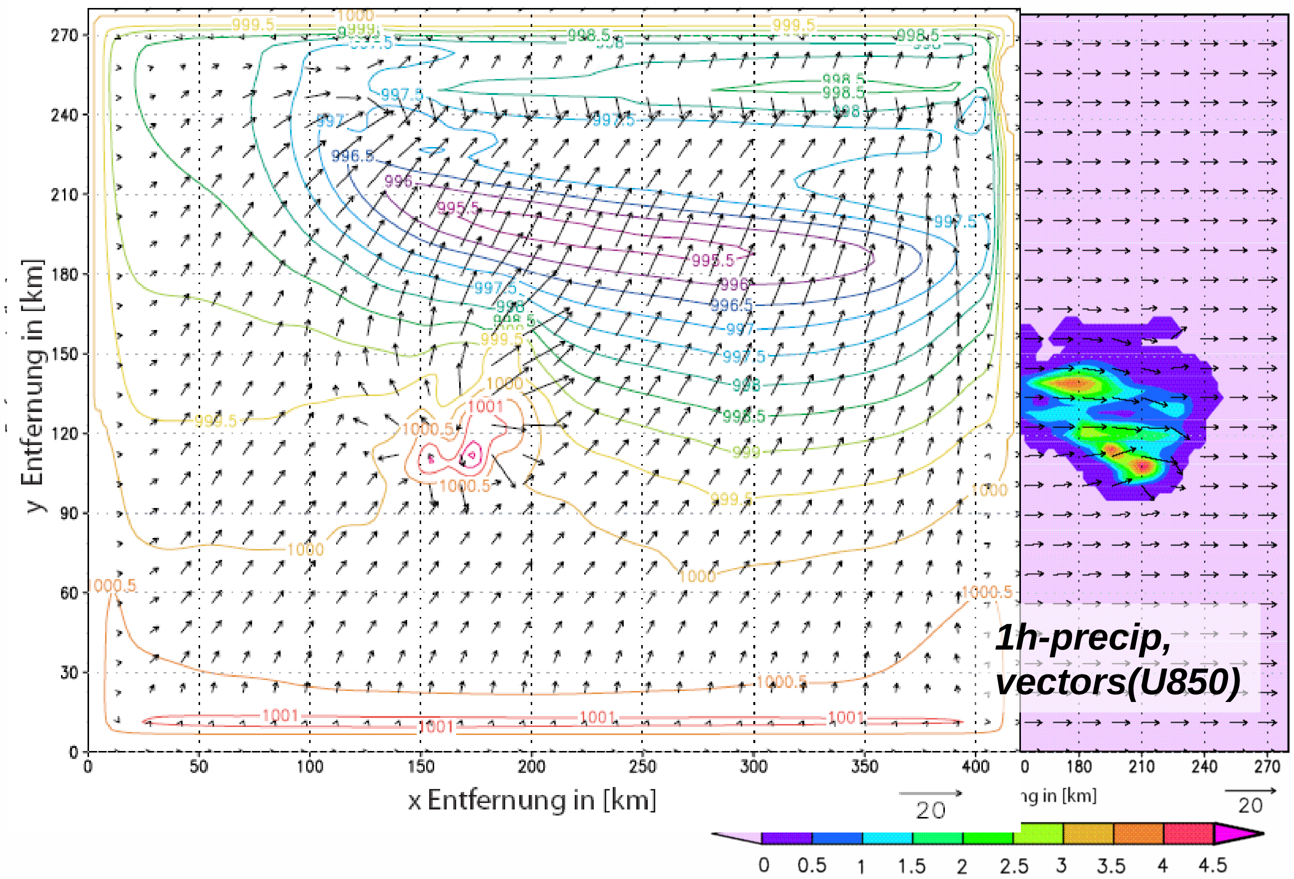




difference between geostrophic pressure p_{geo} and p_{hydEQ} for every grid point vanishes after temperature adjustment!

Initialisation procedure

Result of horizontal density adjustment:
3D hydrostatically balanced flow: convective environment test case



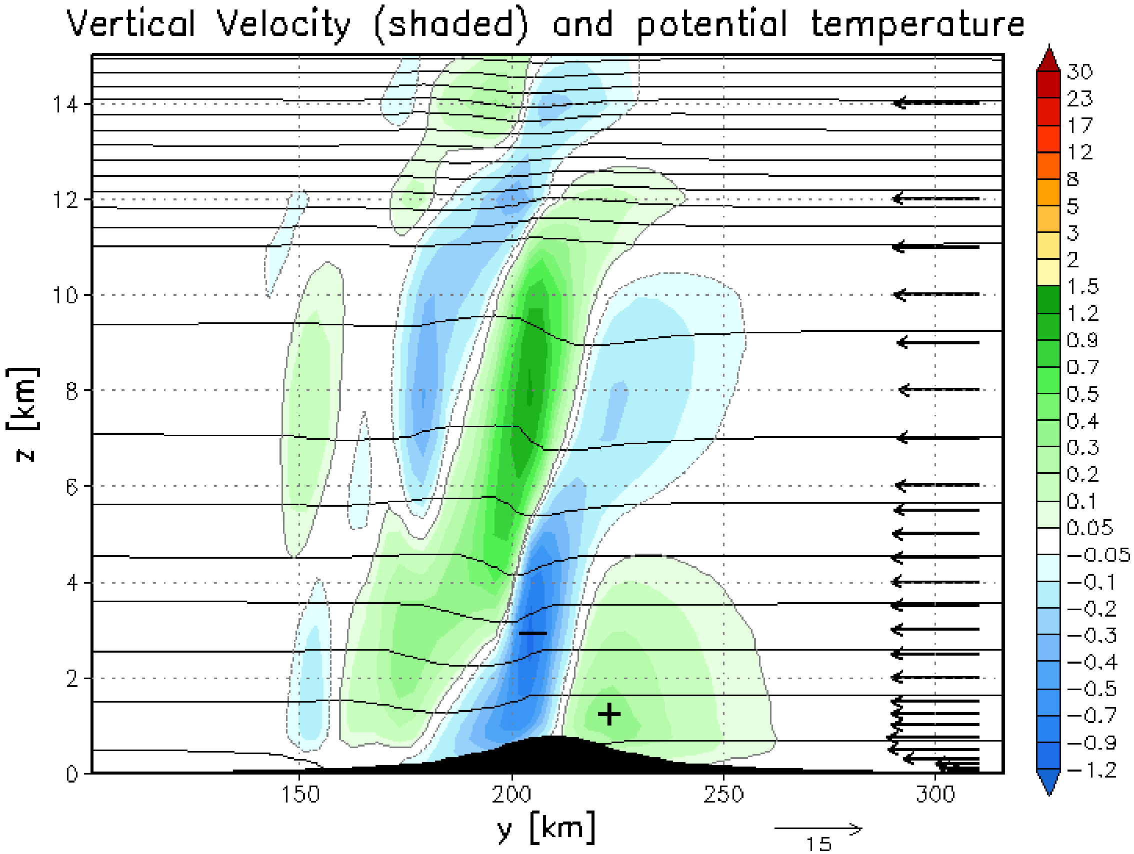
→ no disturbing boundary effects any more

Figure (right): hourly precipitation (shaded), wind vectors at 850hPa level after 3h

Figure (left): Surface pressure (coloured contours), wind vectors at lowest model level after 3h

Coriolis effect - example

Northerly flow impinging on bell-shaped mountain ridge
(slightly stable stratification, nearly saturated, unsheared)

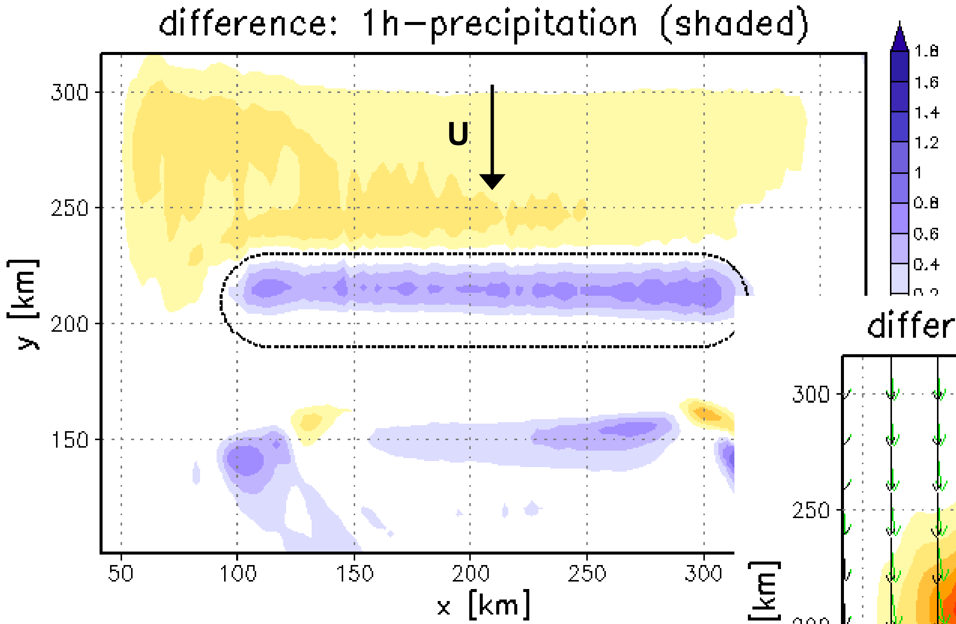


→ wave-like distribution of vertical velocity
→ maximum uplift at windward slope close to theoretical value
$w_{oro}=U\, dh/dx$

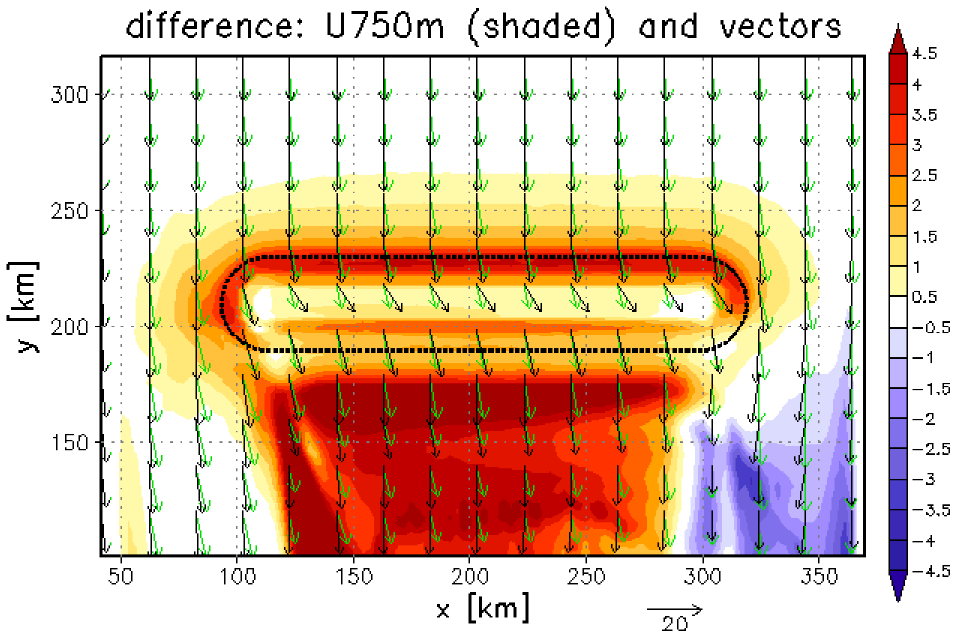
Figure:
Vertical velocity (shaded) and potential temperature (contours, interval 5K)

Coriolis effect - example

difference between runs ***with*** and ***without*** Coriolis terms



→ upstream blocking reduced if Coriolis terms are considered
→ ≈11% higher rain rates at northern slope in this case

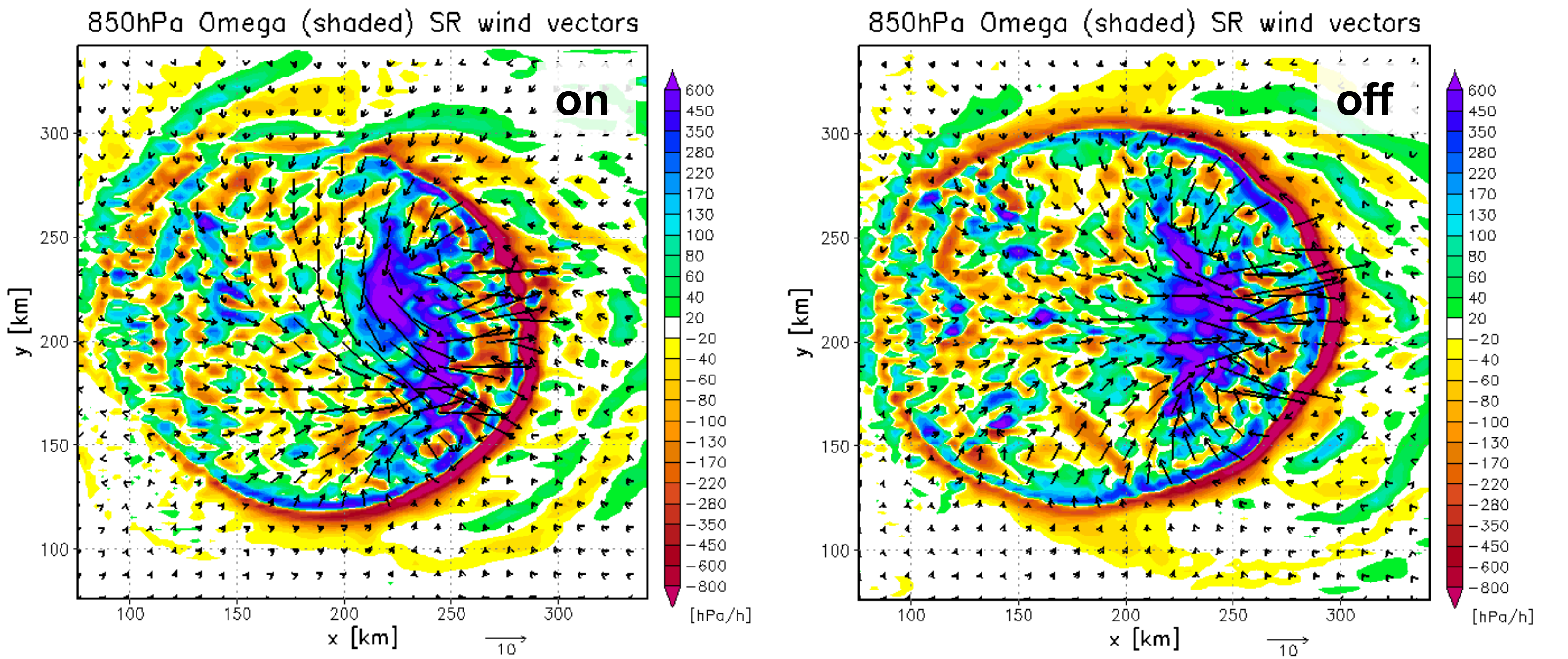


→ speed reduction much smaller
→ leftward deflection by pressure gradient force

(see Pierrehumbert & Wyman, 1985)

Coriolis effect - another example

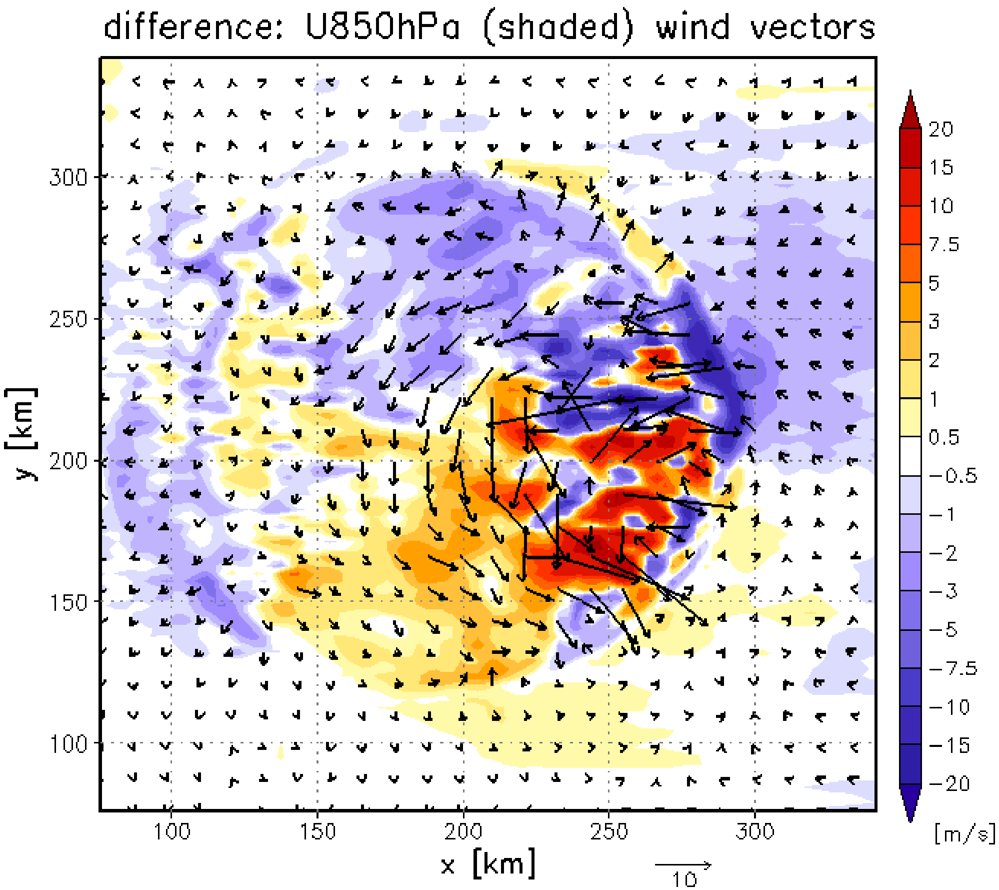
Development of Mesoscale Convective System
(high CAPE environment, moderately sheared westerly flow)



Pressure vertical velocity (shaded) and storm-relative wind vectors at 850hPa after 4h – with and without Coriolis force

Coriolis effect - another example

Development of Mesoscale Convective System
(high CAPE environment, moderately sheared westerly flow)



effect of Coriolis force:

→ symmetry along shear-parallel axis reduced
→ higher wind speeds at southern half of leading edge
→ enhanced convergence
→ signs of mesoscale circulation due to deflection of inflow winds

Figure:
difference of winds at 850hPa (shaded) between Coriolis run and NO-Coriolis-run, and difference vectors

Summary

Current LM's 1D hydrostatic pressure routine has been enhanced for artificial flow simulations ***considering Coriolis force terms*** by

- application of horizontal pressure gradient (geostrophic balance)
- adaptation of wind profile in friction layer (Ekman spiral)
- adjustment of horizontal density to satisfy 3D hydrostatic equilibrium

Test simulations show

- resulting initial and boundary fields remain nearly indistorted
- Coriolis force reduces upstream blocking with flows over orographic barrier
- deflection of wind field by Coriolis force can influence structure and intensity of convective systems

Some problems with new initialisation procedure arise if

- low-level wind shear is very strong (surface pressure affected)
- wind shear abruptly changes sign with height (destabilisation aloft)

THANK YOU FOR LISTENING !

References:

Bluestein, H.B., Weisman, M.L., 2000: The interaction of numerically simulated supercells initiated along lines. *Monthly Weather Review Vol. 128, 3128-3149.*

Holton, J.R., 1992: An introduction to dynamic meteorology. *Academic Press, 511 pp.*

Pierrehumbert, R.T., Wyman, B., 1985: Upstream effects of mesoscale mountains. *Journal of the Atmospheric Sciences Vol. 42 (10), 977-1003.*

Weisman, M.L., Klemp, J.B., 1982: The dependence of numerically simulated convective storms on vertical shear and buoyancy. *Monthly Weather Review Vol. 110, 504-520.*