DCMIP2016, Part 3: Idealized Tropical Cyclone

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Abstract. The 2016 Dynamical Core Model Intercomparison Project (DCMIP-2016) held at the National Center for Atmospheric Research (NCAR) in June 2016 utilized test cases of intermediate complexity to compare and explore advances in non-hydrostatic global models. In particular, DCMIP-2016 paired 12 different dynamical cores with a simplified physical parameterization package to shed light on the physics-dynamics interactions in a controlled test environment. The paper presents the results of an idealized tropical cyclone test case, which is the second of the three DCMIP-2016 tests. It is based on an analytically-prescribed vortex in a quiescent background environment that fosters a rapid intensification of the initial tropical cyclone seed. The cyclone intensifies from a weak, balanced vortex to a tropical cyclone over a 10-day period. Moisture and physics-dynamics coupling features prominently in this test, as well as the impact of precipitation and boundary layer parameterization formulations. Variable-resolution setups can be used to model the tropical cyclone at fine grid spacings and assess the impacts of the resolution transition regions on the behavior of the storm. The DCMIP-2016 models are compared at a default horizontal grid spacing of approximately 50 km, and effects of higher horizontal resolutions are explored. The

paper demonstrates how idealized test cases are part of a model hierarchy that helps distinguish between causes and effects in atmospheric models.

1 Introduction

The simplified tropical cyclone test case on a regular-size Earth is based on the work of Reed and Jablonowski (2012, 2011a, c, b). In this test an analytic vortex is initialized in a background environment which is tractable to a rapid intensification of tropical cyclones.

Table 1. List of constants used for the Ideaized Tropical Cyclone test

Constant	Value	Description
X	1	small-planet scaling factor (regular-size Earth)
z_t	15000 m	Tropopause height
q_0	0.021 kg/kg	Maximum specific humidity amplitude
q_t	10^{-11} kg/kg	Specific humidity in the upper atmosphere
T_0	302.15 K	Surface temperature of the air
T_s	302.15 K	Sea surface temperature (SST), 29 C$^\circ$
z_{q1}	3000 m	Height related to the linear decrease of q with height
z_{q2}	8000 m	Height related to the quadratic decrease of q with height
Γ	0.007 K m^{-1}	Virtual temperature lapse rate
p_b	1015 hPa	Background surface pressure
φ_c	$\pi/18$	Initial latitude of vortex center (radians)
λ_c	π	Initial longitude of vortex center (radians)
Δp	11.15 hPa	Pressure perturbation at vortex center
r_p	282000 m	Horizontal half-width of pressure perturbation
z_p	7000 m	Height related to the vertical decay rate of p perturbation
ϵ	10^{-25}	Small threshold value

1.1 Initialization

The background state consists of a prescribed specific humidity profile, virtual temperature and pressure profile. The initial profile is defined to be in approximate gradient wind balance. The vertical sounding is chosen to roughly match an observed tropical sounding documented in Jordan (1958). The background specific humidity profile $\overline{q}(z)$ as a function of height z is

$$\begin{aligned}
&\overline{q}(z) = q_0 \exp\left(-\frac{z}{z_{q1}}\right) \exp\left[-\left(\frac{z}{z_{q2}}\right)^2\right] \quad \text{for } 0 \le z \le z_t \\
&\overline{q}(z) = q_t \quad \text{for } z_t \le z
\end{aligned} \tag{1}$$

The background virtual temperature sounding $\overline{T}_v(z)$ is split into two different representations for the lower and upper atmosphere. It is given by

$$\begin{aligned} \overline{T}_v(z) &= T_{v0} - \Gamma z && \text{for } 0 \le z \le z_t, \\ \overline{T}_v(z) &= T_{vt} = T_{v0} - \Gamma z_t && \text{for } z_t < z, \end{aligned} \tag{2}$$

with the virtual temperature at the surface $T_{v0} = T_0(1 + 0.608\,q_0)$ and the virtual temperature at the tropopause level $T_{vt} = T_{v0} - \Gamma z_t$. The background temperature profile can be obtained from (**??**).

The background vertical pressure profile $\overline{p}(z)$ of the moist air is computed using the hydrostatic balance and (2). The profile is given by:

$$\begin{aligned} \overline{p}(z) &= p_b \left(\frac{T_{v0} - \Gamma z}{T_{v0}} \right)^{g/R_d\Gamma} && \text{for } 0 \le z \le z_t, \\ \overline{p}(z) &= p_t \exp\left(\frac{g(z_t - z)}{R_d T_{vt}} \right) && \text{for } z_t < z. \end{aligned} \tag{3}$$

The pressure at the tropopause level z_t is continuous and given by

$$p_t = p_b \left(\frac{T_{vt}}{T_{v0}} \right)^{\frac{g}{R_d\Gamma}}, \tag{4}$$

which, for the given set of parameters, is approximately 130.5 hPa.

1.1.1 Axisymmetric Vortex

The pressure equation $p(r,z)$ for the moist air is comprised of the background pressure profile (3) plus a 2D pressure perturbation $p'(r,z)$,

$$p(r,z) = \overline{p}(z) + p'(r,z), \tag{5}$$

where r symbolizes the radial distance (or radius) to the center of the prescribed vortex. On the sphere r is defined using the great circle distance

$$r = a \arccos\left(\sin\varphi_c \sin\varphi + \cos\varphi_c \cos\varphi \cos(\lambda - \lambda_c)\right). \tag{6}$$

The perturbation pressure is defined as

$$\begin{aligned} p'(r,z) &= -\Delta p \exp\left[-\left(\frac{r}{r_p}\right)^{3/2} - \left(\frac{z}{z_p}\right)^2 \right] \left(\frac{T_{v0} - \Gamma z}{T_{v0}} \right)^{\frac{g}{R_d\Gamma}} && \text{for } 0 \le z \le z_t, \\ p'(r,z) &= 0 && \text{for } z_t < z. \end{aligned} \tag{7}$$

The pressure perturbation depends on the pressure difference Δp between the background surface pressure p_b and the pressure at the center of the initial vortex, the pressure change in the radial direction r_p and the pressure decay with height within the vortex z_p. The moist surface pressure $p_s(r)$ is computed by setting $z = 0$ m in (5), which gives

$$p_s(r) = p_b - \Delta p \exp\left[-\left(\frac{r}{r_p}\right)^{3/2} \right]. \tag{8}$$

The axisymmetric virtual temperature $T_v(r,z)$ is computed using the hydrostatic equation and ideal gas law

$$T_v(r,z) = -\frac{g p(r,z)}{R_d}\left(\frac{\partial p(r,z)}{\partial z}\right)^{-1}. \tag{9}$$

Again it can be written as a sum of the background state and a perturbation,

$$T_v(r,z) = \overline{T}_v(z) + T_v'(r,z), \tag{10}$$

where the virtual temperature perturbation is defined as

$$\begin{aligned} T_v'(r,z) &= (T_{v0} - \Gamma z)\left\{\left[1 + \frac{2R_d(T_{v0}-\Gamma z)z}{g z_p^2\left[1 - \frac{p_b}{\Delta p}\exp\left(\left(\frac{r}{r_p}\right)^{3/2} + \left(\frac{z}{z_p}\right)^2\right)\right]}\right]^{-1} - 1\right\} && \text{for } 0 \le z \le z_t, \\ T_v'(r,z) &= 0 && \text{for } z_t < z. \end{aligned} \tag{11}$$

The axisymmetric specific humidity $q(r,z)$ is set to the background profile everywhere

$$q(r,z) = \overline{q}(z). \tag{12}$$

Consequently, the temperature can be written as

$$T(r,z) = \overline{T}(z) + T'(r,z), \tag{13}$$

with the temperature perturbation

$$\begin{aligned} T'(r,z) &= \frac{T_{v0} - \Gamma z}{1 + 0.608\overline{q}(z)}\left\{\left[1 + \frac{2R_d(T_{v0}-\Gamma z)z}{g z_p^2\left[1 - \frac{p_b}{\Delta p}\exp\left(\left(\frac{r}{r_p}\right)^{3/2} + \left(\frac{z}{z_p}\right)^2\right)\right]}\right]^{-1} - 1\right\} && \text{for } 0 \le z \le z_t, \\ T'(r,z) &= 0 && \text{for } z_t < z. \end{aligned} \tag{14}$$

Due to the small specific humidity value in the upper atmosphere (10^{-11} kg/kg for $z > z_t$) the virtual temperature equals the temperature to a very good approximation in this region. The formulation presented here is equivalent to the one presented in Reed and Jablonowski (2012).

If the density of the moist air needs to be initialized its formulation is based on the ideal gas law

$$\rho(r,z) = \frac{p(r,z)}{R_d T_v(r,z)} \tag{15}$$

which utilizes the moist pressure (5) and virtual temperature (10). The surface elevation z_s and thereby the surface geopotential $\Phi_s = g z_s$ are set to zero.

Finally, the tangential velocity field $v_T(r,z)$ of the axisymmetric vortex is defined by utilizing the gradient-wind balance, which depends on the pressure (5) and the virtual temperature (11). The tangential velocity is given by

$$v_T(r,z) = -\frac{f_c r}{2} + \sqrt{\frac{f_c^2 r^2}{4} + \frac{R_d T_v(r,z)\, r}{p(r,z)}\frac{\partial p(r,z)}{\partial r}}, \tag{16}$$

where $f_c = 2\Omega\sin(\varphi_c)$ is the Coriolis parameter at the constant latitude φ_c. Substituting $T_v(r,z)$ and $p(r,z)$ into (16) gives

$$v_T(r,z) = -\frac{f_c r}{2} + \sqrt{\frac{f_c^2 r^2}{4} - \frac{\frac{3}{2}\left(\frac{r}{r_p}\right)^{3/2}(T_{v0}-\Gamma z)R_d}{1 + \frac{2R_d(T_{v0}-\Gamma z)z}{g z_p^2} - \frac{p_b}{\Delta p}\exp\left(\left(\frac{r}{r_p}\right)^{3/2} + \left(\frac{z}{z_p}\right)^2\right)}} \quad \text{for } 0 \le z \le z_t,$$

$$v_T(r,z) = 0 \quad \text{for } z_t < z. \tag{17}$$

The last step is to split the tangential velocity (17) into its zonal and meridional wind components $u(\lambda,\varphi,z)$ and $v(\lambda,\varphi,z)$. Similar to Nair and Jablonowski (2008) these are computed using the following expressions,

$$d_1 = \sin\varphi_c\cos\varphi - \cos\varphi_c\sin\varphi\cos(\lambda-\lambda_c) \tag{18}$$

$$d_2 = \cos\varphi_c\sin(\lambda-\lambda_c) \tag{19}$$

$$d = \max\left(\epsilon, \sqrt{{d_1}^2 + {d_2}^2}\right), \tag{20}$$

which are utilized in the projections

$$u(\lambda,\varphi,z) = \frac{v_T(\lambda,\varphi,z)\,d_1}{d} \tag{21}$$

$$v(\lambda,\varphi,z) = \frac{v_T(\lambda,\varphi,z)\,d_2}{d}. \tag{22}$$

A small $\epsilon = 10^{-25}$ value avoids divisions by zero. The vertical velocity is set to zero.

Note: We are currently investigating a test case specification that places the idealized tropical cyclone on an f-plane. This configuration would remove issues associated with β drift of the cyclone and allow for a more direct intercomparison of the simulated storm.

1.2 Grid spacings, simulation time, output and diagnostics

- Moist simulations should be performed at 0.5° resolution with 30 vertical levels for 10 days.
- Plots of minimum surface pressure over the duration of the simulation.
- Experiments could address the coupling frequency between the dynamics and physics.
- A variable resolution simulation should be performed that (a) studies the effect of the tropical cyclone transitioning from fine resolution to coarse resolution and (b) high resolution simulations down to 0.125° over the tropical cyclone.

2 Conclusions

TEXT

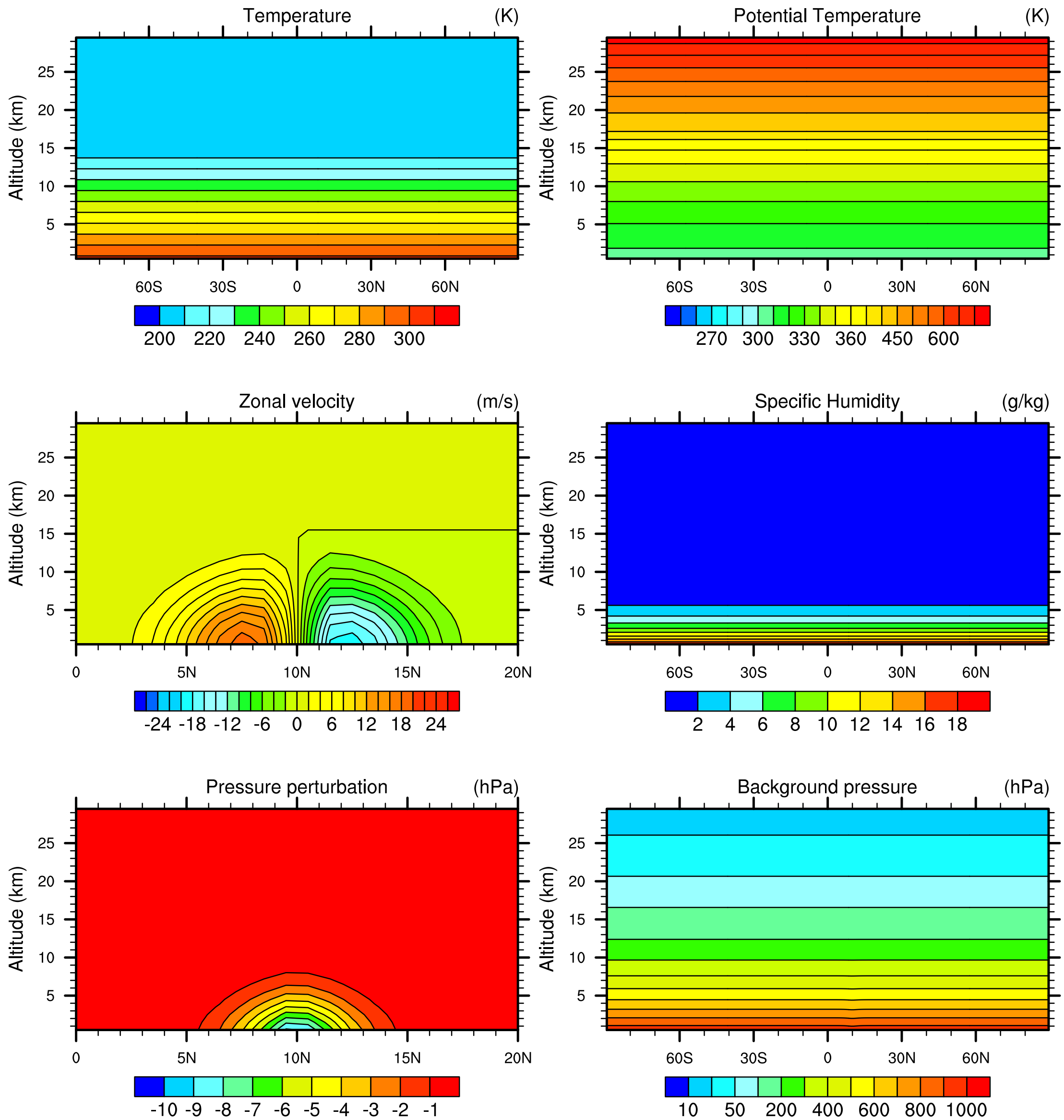

Figure 1. Initial state for the tropical cyclone test.

Appendix A: Alternative Planetary Boundary Layer for Tropical Cyclone Test

An alternative approach for boundary layer mixing has been proposed using a K-profile parameterization formulation for the tropical cyclone test described in Section **??**. The implementation only impacts the manner in which the eddy diffusivity coefficients are calculated. In particular the calculation of K_m in (**??**) is replaced with

$$\begin{aligned} K_m &= \kappa u^* z \left(1 - \tfrac{z}{h}\right)^2 && \text{for } z \le h \\ K_m &= 0 && \text{for } z > h, \end{aligned} \tag{A1}$$

where $\kappa = 0.4$, $u^* = \sqrt{C_d}|\boldsymbol{v}_a|$ and $h = 1$ km. K_E in (**??**) is then defined as

$$\begin{aligned} K_E &= \kappa e^* z \left(1 - \tfrac{z}{h}\right)^2 && \text{for } z \le h \\ K_E &= 0 && \text{for } z > h, \end{aligned} \tag{A2}$$

where $e^* = \sqrt{C_E}|\boldsymbol{v}_a|$. This implementation will be used in supplemental simulations of the tropical cyclone test.

Author contributions. TEXT

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