A Quantitative Assessment of the Impact of Increase in CO$_2$ Concentration on Baroclinic Instability

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Carbon dioxide:
- Linear structure
- No permanent electric dipole moment
- Oscillating dipole moment

Transmission spectrum of CO₂

Petty (2006), Fig. 9.12

Petty (2006), Fig. 9.13
Radiative-Dynamical Core Interaction

Input data:
- Temperature
- Pressure
- Active gasses
- Optical properties of clouds & aerosols

Radiative transfer scheme

Output data:
- Radiative fluxes
- Heating (cooling) rate

GCMs structure
Physical packages
Dynamical core

Source: Center for Multiscale Modeling of Atmospheric Processes
CO\textsubscript{2} Radiative Impacts on Baroclinic Instability

The same initial conditions + Different CO\textsubscript{2} concentrations

Radiative transfer model → Baroclinic instability test case → Dynamical core

Aim: Assessing the impact of increase in CO\textsubscript{2} concentration on baroclinic instability
Initial Conditions for Five Experiments

- Initial CO$_2$ concentrations:
  1. 0 ppm
  2. 250 ppm
  3. 500 ppm
  4. 750 ppm
  5. 1000 ppm

- Integration time: 20 days

Start on 1 January 2009-12UTC, Integration time: 20 days
Radiative Transfer Scheme

Setting:

- No aerosol & clear sky
- No PBL, surface albedo set to 0.3
Performance of Baroclinic Instability

Jablonowski & Williamson baroclinic instability test case (2006)

1. Defining the initial condition

Innovation:

The Newtonian relaxation $\frac{\partial T}{\partial t} = -k(T - T_a)$

Radiative transfer scheme $\frac{\partial T}{\partial t} = -\frac{1}{\rho C_p} \frac{\Delta F(z)}{\Delta z}$

2. Triggering baroclinic instability by adding a perturbation to an analytically defined zonal mean jet in a steady state
Dynamical Core

DCASL Based on potential vorticity

CASL
(Mohebalhojeh and Dritschel, 2004)

DCASL
(Mohebalhojeh and Dritschel, 2007 & 2009)

DCASL
(Mohebalhojeh et al, 2016)

Vertical coordinate: $\xi = F(\sigma, \theta)$
The contribution of $CO_2$ to net radiative heating is rather small in the troposphere, because the emitted radiation is reabsorbed at nearby levels with the same temperature. There is only a net radiative heating at the tropopause, where the temperature profile has a minimum value. It is also interesting to note that the CO2 radiative impacts on baroclinic instability are saturated at 750 ppm.
Eddy Activity & CO₂ Radiative Effects

For all experiments baroclinic instability forms at t=9 and eddy’s decay (barotropic phase) begins at t=20 by decreasing EKE and the falling growth rate. The values of EKE and sigma decrease with increasing of CO₂ concentrations. In addition, the different slope of diagrams show how eddies are formed and propagated for different amounts of CO₂. So baroclinic instability can be divided into five periods t:0-9, t:9-12, t:12-16, t:16-18 and t:18-20.

\[ EKE = \frac{1}{2}(u'^2 + v'^2)^{1/2} \]

\[ \sigma = \frac{\log(K'_t/K'_0)}{t} \]
Eddy kinetic energy is derived from the eddy available potential energy, which is dependent on the mean available potential energy. Since the meridional temperature gradient is proportional to the mean available potential energy, it can affect eddy kinetic energy indirectly. The panels indicate the CO$_2$ radiative impacts on the meridional potential temperature gradient from 0 to 90N latitude and from 120W to 180W longitude at the 850 level. As expected, the meridional potential temperature gradient is reduced by increasing CO$_2$ concentrations through cooling in the lower latitudes at this level.
Cooling Rate Difference at $t=0$

The baroclinic instability has not yet started at $t=0$, so the differences among the panels are due to increase in the atmospheric optical depth, through which the lower (upper) levels of the troposphere become colder (warmer) by increase in $\text{CO}_2$ concentration.
Since the interaction between radiation and the dynamical core is a two way relationship, large eddies relax the atmosphere to the initial state by meridional and vertical heat fluxes at t=9 (as instability forms). The polar heat flux changes the temperature field trough which affects radiative cooling rate. The reduction of radiative impacts is seen at t=9 compared to t=0 is seen in all the five experiments.
In the following days, eddy activities influence the temperature field which changes the cooling rate through thermodynamic energy equation.
Eddy-Mean Flow Interaction

**Eliassen–Palm flux** *(Edmon et al, 1980)*

\[
F = \{F_y, F_p\} \quad ; \quad F_y = -u'v', \quad F_p = f \frac{v'\theta'}{\partial \theta / \partial p}
\]

- **Meridional momentum flux**
  - \(F_y\)
  - \(KE \rightarrow KZ\)

- **Meridional temperature flux**
  - \(F_p\)
  - \(AZ \rightarrow AE\)

**Eliassen–Palm flux divergence**

\[
\nabla \cdot F = v'q' = \frac{\partial}{\partial y} (-v'u') + \frac{\partial}{\partial p} (f \frac{v'\theta'}{\partial \theta / \partial p})
\]

- **Source**
  - \(\nabla \cdot F > 0\)
  - Acceleration

- **Sink**
  - \(\nabla \cdot F < 0\)
  - Deceleration
Eddy-Mean Flow Interaction at t=9

Zonal mean zonal wind at the 200 hPa level; Jet: red solid line, EP flux: black arrow, EP flux divergence: shaded

The divergence (convergence) zone is proportional to the energy source (sink) region, which is attenuated by increase in CO₂ concentration. As baroclinic instability forms (t=9), eddy activity is saturated at the lower troposphere. Furthermore, the CO₂ radiative impacts weaken the upward EP flux (proportional to the poleward heat flux, centered at latitude 50N) and subsequently the eddy’s growth at lower levels.
The wave packets propagate in meridional direction by forming the divergence and convergence zones sequentially. This propagation is directly related to the slope of EKE diagrams at all the five experiments. In the first experiment, the meridional propagation of waves occur with a higher energy level. In the other words, the consecutive areas of source and sink provide the growth of eddies more rapidly.
At this stage, the vertical extension of divergence (convergence) zones to the middle levels of the atmosphere affects the acceleration of wind (deceleration) and the upper-level jet. As this process continues, the jet becomes stronger (weaker) according to divergence (convergence) zones. For increasing CO₂ concentration and attenuation of the divergence zones, the meridional and vertical propagation of eddies are decreased at the lower levels, thereby weakening the jet.
Eddy-Mean Flow Interaction at $t=18$

Zonal mean zonal wind at the 200 hPa level; Jet : red solid line, EP flux : black arrow, EP flux divergence: shaded

As the EP flux vectors become equatorward (proportional to the momentum fluxes), acceleration increases at the lower levels. So the eddy activity is saturated in the upper troposphere and the decay phase starts. This stage is proportional to the downward slop of the growth rate diagram.
Eddy-Mean Flow Interaction at $t=20$

Zonal mean zonal wind at the 200 hPa level; Jet: red solid line, EP flux: black arrow, EP flux divergence: shaded

Proportional to the primary energy in each experiment (for different CO$_2$ concentrations), the atmospheric structure changes from baroclinic to barotropic by the conversion of eddy kinetic energy to mean kinetic energy at $t=20$. 
Quantitative assessment
Quantitative assessment

Maximum zonal-time average of upward (FP) and equatorward (FY) components of the EP flux, divergence (DIV_F) and convergence (CONV_F) of the EP flux and the jet (UZO)
Results

• Increase in the concentration of CO$_2$ decreases the meridional temperature gradient and thus reduces the eddy kinetic energy at lower atmospheric levels.

• Increase in the CO$_2$ concentration causes the growth rate, meridional and vertical eddy propagation and upper-level jet to weaken.

• The CO$_2$ radiative impacts on baroclinic instability are saturated at 750 ppm.
References