Coupling between the Carbon Cycle and Physical Processes on multiple scales in the past and present

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Joint Mathematics Meeting
January 2008

Atmospheric Carbon Dioxide
Measured at Mauna Loa, Hawaii

NH summer

NH winter
Radiative Balance

Incoming = Outgoing

\[ I = \epsilon \sigma T^4 \]

\( \epsilon \) = emissivity

30% is reflected
... the rest is absorbed

Earth intercepts 1.37 kW/m²

... and later re-emitted at a longer wavelength (infrared)

How could the mean temperature of the planet change?

1. Incoming Insolation - small fluctuations
2. Reflected - albedo variations - with climate
3. Emitted radiation - emissivity affected by atmospheric constituents
Atmospheric greenhouse gases (CO2) and the earth’s climate have been amazingly stable!
The last 65 m yrs

From deep sea sedimentary cores.

Lowering of CO2 -- by slower spreading of continental plates or faster uplift and weathering.

Source: Zachos et al, Nature 2005
Paleocene-Eocene Thermal Maximum (PETM) as recorded by benthic foraminifer isotopic records (Zachos et al. 2003).

The decrease in carbon isotopic ratios is indicative of a rapid increase in atmospheric CO2 and CH4 coincident with a 5 degree warming.

Additional greenhouse carbon would be absorbed by the oceans, resulting in reduced pH, depression in the carbonate saturation horizon, and dissolution of sea floor calcium carbonate.
Carbon Dioxide and Temperature Records

- 10 deg variation in temperature
- 100 ppm variation in carbon dioxide

The last 600K yrs
The last millennium

We’re presently experiencing in one of the largest and most abrupt perturbations to climate and ecology

Temperature

Millennial Northern Hemisphere (NH) temperature reconstruction (blue) and instrumental data (red) from AD 1000 to 1999, adapted from Mann et al. (1999).

The temperature commensurate with the present CO2 levels will be evidenced in the coming century.
\[ F = k \cdot s \cdot \Delta pCO_2 \]

- \( F \) = Air-sea flux of CO2
- \( k \) = Gas exchange coeff
- \( s \) = Solubility
- \( \Delta pCO_2 \) = Air-sea difference in partial pressures

**Gas exchange coefficient**

**Wind speed**

**Anthropogenic emissions 7.1gt, Oceanic uptake 2.3 gt, Terrestrial sink 1.8 gt**
Section - Atlantic Ocean

Dissolved CO2 distribution

\[
\text{DIC} = \text{CO}_2(\text{gas}) + \text{H}_2\text{CO}_3 + \text{HCO}_3^- + \text{CO}_3^{2-}
\]

The ocean holds an enormous capacity for CO2. But the equilibration between atmosphere and ocean is slow due to the chemical buffering of CO2 and slow mixing.
Marine Biological Pump

Mean vertical profiles of oceanic properties

Plots of temperature, salinity, dissolved oxygen content and nitrate content as a function of water depth at GEOSECS station 214 in the North Pacific (32°N, 176°W). Potential temperature rather than the temperature measured in situ is given. The salinity profile in this region of the ocean shows a pronounced minimum at a depth just over 600 meters. This intermediate water (as it is called) forms in the northern Pacific and sinks and flows latitudinally beneath the waters of the warm temperate ocean. The level of this minimum is shown in the other diagrams by a dashed line. Dissolved oxygen is utilized by animals and bacteria living in the deep sea to “burn” the organic debris falling from the surface. Thus all deep waters are deficient in oxygen compared to the amount they received from the atmosphere before descent. As oxygen is consumed, nitrate is produced. Thus the shapes of the profiles of
What limits our ability model biological productivity?

1. Vertical flux of nutrients
   Mixed layer dynamics
   Mechanisms/scales for upwelling?

2. Ecosystem modeling
   Need productivity & fluxes, but measure concentrations.

NPZD...

\[ \frac{DN}{Dt} = -P_{growth}(\text{light}, N, P) + R \]
\[ DP/Dt = +P_{growth}(\text{light}, N, P) \]
\[ -Z_{growth}(P, Z) \]
\[ -\text{sinking} - \text{mortality} \]
\[ DZ/Dt = Z_{growth}(P, Z) \]
\[ -\text{detritus formation} - \text{mortality} \]
\[ DD/Dt = \text{detritus formation} - R \]

very large number of parameters
Modeling

\[ P = \text{Hydrostatic pressure} \]
\[ Q = \text{Nonhydrostatic pressure} \]
\[ P = p + \delta q \]

\[
Dt u + Ro^{-1} (p_x + \delta q_x - f v + Ro \delta bw) = F^x \\
Dt v + Ro^{-1} (p_y + \delta q_y + f u) = F^y \\
Dt w + Ro^{-2} \delta^{-2} (\rho^{-1} p_z + g + \delta q_z - \delta bu) = F^z \\
u_x + v_y + Ro w_z = 0
\]

Hydrostatic \( \delta \to 0 \)

\[
p_z + \rho g = 0 \\
w_z = -Ro^{-1} (u_x + v_y)
\]

Free-surface height

...and density \( \Rightarrow p \)

Nonhydrostatic (\( \delta \) not \( \to 0 \))

\[
Dt w + Ro^{-2} \delta^{-1} (q_z - bu) = F^z
\]

Well-posed with open boundaries

\[ b \equiv 2\Omega \cos \phi \]
\[ f \equiv 2\Omega \sin \phi \]

\[ Ro = \frac{U}{\Omega L} \]
\[ \delta = \frac{D}{L} \]

Mahadevan et al., 1996a,b, Mahadevan & Archer 1998
Surface nitrate / biological production is sensitive to vertical velocities, and hence, model resolution.

There are fronts .... and smaller fronts.

Mahadevan and Archer, (2000)
At higher (1 km) model resolution, we find that:
The largest vertical velocities $O(100\text{m/day})$ occur where the Rossby number becomes $O(1)$.
Circulation not in geostrophic / thermal-wind balance -- has a large vertical component.
A closer look at a single feature

Surface Density

Frontogenesis

Rossby Number = \text{Rel vor} / f

Vertical Velocity at 15 m

A simpler model for circulation in the vertical plane

Semi-geostrophic: higher order in Ro

\[ b = \frac{-g \rho}{\rho_0} F_2^2 \frac{\partial^2 \psi}{\partial z^2} + 2 S_2^2 \frac{\partial^2 \psi}{\partial z \partial y} + N^2 \frac{\partial^2 \psi}{\partial y^2} = -2Q^g, \]

where \( N^2 = b_z \), \( S_2^2 = -b_y = fu_{yz} \), \( F_2^2 = f(f - u_{yy}) \)

Potential vorticity

\[ q_{2D} = \frac{1}{f} (F_2^2 N^2 - S_2^4) \]

\[ Q^g = (Q_1^g, Q_2^g) = \left( -\frac{\partial u_y}{\partial x}, -\frac{\partial u_y}{\partial y} \cdot \nabla b \right) \]

< generally positive, but when it changes sign, this is not solvable

\textbf{Loss of balance} -- leads to vertical motion and mixing.
Spatial heterogeneity: sea surface chlorophyll and temperature

Variance spectra
Slope is a measure of patchiness

Surface Chl is always more patchy than temperature. Why?

Mahadevan & Campbell, 2002
What we have achieved with global carbon cycle models

We can model the present, large-scale, seasonally varying, distributions of phytoplankton and air-sea flux of CO2

Do we need to get the biology right at these scales?
Are our carbon cycle models equipped to handle the effects of climate change?

Labrador Sea salinity 1950-2002, Dickson et al., 2002

Issues
dynamical - vertical transport
biological - net community production and export