Biogeochemical Model Description

The marine carbon component of the coupled climate-carbon cycle model is based on the Ocean Carbon Model Intercomparison Project (OCMIP; Doney et al., 2006; Najjar and Orr, 1999, http://www.ipsl.jussieu.fr/OCMIP/). The discrete tracer transport equations for a tracer $\varphi$ are given according to Smith and Gent (2004) by

$$\frac{\partial}{\partial t} (1 + \xi) \varphi + L_T(\varphi) = D_H(\varphi) + D_V(\varphi) + F_W(\varphi) + F_B(\varphi)$$

(1)

where the first term on the left side of (1) is the local change of the tracer concentration with the factor $1 + \xi$ associated with changes in surface volume due to undulations, and the second term is the advection of tracers with advection operator $L_T$. On the right side of (1) are the horizontal $D_H(\varphi)$ and vertical $D_V(\varphi)$ diffusion terms, followed by the change of tracer concentration by freshwater flux $F_W(\varphi)$ and by biogeochemical processes $F_B(\varphi)$.

Transfer of gases between ocean and atmosphere reservoirs is calculated as shown in Doney et al. (2006) by

$$\Delta F = F_{ao} - F_{oa} = K_u[(p\text{CO}_2)_a - \beta_T[\text{CO}_2]](1 - f_{ice})$$

(2)

where $\Delta F = F_{ao} - F_{oa}$ is the CO$_2$ gas flux across the air-sea interface, $K_u$ is the wind-dependent air-sea gas exchange coefficient (Wanninkhof, 1992), $(p\text{CO}_2)_a$ is the atmospheric partial pressure of CO$_2$ (which is prescribed for the Late Permian as 3550 ppmv), $\beta_T$ is the temperature-dependent solubility of CO$_2$, and $[\text{CO}_2]$ is the dissolved CO$_2$ concentration of the uppermost layer of the ocean ($([p\text{CO}_2])_o = \beta_T[\text{CO}_2]$). If sea ice is present in high latitudes, then the CO$_2$ gas flux across the sea-air interface is reduced by fractional sea-ice coverage. The ocean carbon cycle model includes seven prognostic variables, i.e. phosphate (PO$_4$), total dissolved inorganic iron (Fe), dissolved organic phosphorus and iron (DOP and DOFe), dissolved inorganic carbon (DIC), total alkalinity (TALK), and dissolved oxygen (O$_2$), that are transported by the ocean model. The parameterization of biological uptake of nutrients $J_{\text{Prod}}$ by Doney et al. (2006) is similar to that used in the Hamburg Model of the Ocean Carbon Cycle (HAMOCC; Maier-Reimer, 1993):
\[
J_{\text{Prod}} = F_T F_N F_1 B \max \left( 1, \frac{z_{ml}}{z_c} \right) / \tau
\] (3)

Nutrient uptake is limited by temperature \( F_T = (T + 2)/(T + 10) \), phosphate (PO₄) and iron (Fe) availability \( F_N = \min \left( \frac{PO_4}{PO_4-K_{PO_4}}, \frac{Fe}{Fe-K_{Fe}} \right) \), and light \( F_I = \frac{I}{I+K_I} \), where \( I \) is the short wave radiation and \( K_1 \) is a light-limitation term (20 W m⁻²). \( B = \min \left( PO_4, \frac{Fe}{rFe:B} \right) \) is a proxy for biomass with \( rFe:B \) representing the ratio of Fe to PO₄ uptake of 5.85 x 10⁻⁴; the optimal timescale for uptake \( \tau \) is set to 15 days. Productivity only occurs above the compensation depth \( z_c \) of 75 m (\( z_{ml} < z_c \)). The carbon cycle model assumes a constant nitrogen-to-phosphate Redfield ratio for particulate organic matter (Redfield et al., 1963). This assumption is based on the fact that the nitrate concentration is generally well correlated to the phosphate concentration in the present-day ocean (see Fig. 8.17 in Millero, 2006). An additional benefit is the significant reduction of computational costs by simulating only PO₄ instead of four tracers (PO₄, NO₃, NO₂, NH₄) (eq. (1)). Changes in the Redfield ratio at the P-Tr boundary cannot be evaluated with high confidence because the relative importance of various phytoplankton groups as contributors to sedimentary organic matter is not well known (Martin and Knauer, 1973; Tappan, 1980; Quigg et al., 2011), and pervasive post-depositional alteration of sedimentary organic matter precludes direct determination of ancient biotic Corg:P ratios.

The uptake of PO₄ is given by the turnover of biomass, modulated by surface solar irradiance, temperature, and macronutrients and micronutrients. The model uses a Martin power-law curve to describe the vertical particulate organic phosphorus flux (\( F_{\text{POP}} \)) profile over the full water column:

\[
F_{\text{POP}} = F_{\text{POP}}(75)(z/75)^{-a}
\] (4)

where \( z \) is the water depth in m and the exponent \( a \) is typically set to 0.9 (Martin et al. 1987), but is changed in this study in one experiment to 0.5 in order to parameterize a deeper particle penetration depth (Buesseler et al., 2007). The flux is altered by scavenged Fe attached to the sinking matter throughout the water column (see Doney et al., 2006, for details).

**Model comparison**

The differences between the Permian simulations with GENIE-1, an earth system model of intermediate complexity, and with CCSM3, a fully-coupled climate model based on general circulation models of the climate subsystems, are of complex nature and related to differences in
model formulations and parameterizations and in resolution. GENIE-1 is described in Meyer et al. (2008) and in Ridgwell et al. (2007).

In GENIE-1, a 2-D energy moisture balance model and a prescribed wind stress field from CCSM3 (Kiehl and Shields, 2005) are coupled to a 3-D non-eddy resolving frictional geostrophic ocean circulation model, whereas in CCSM3, an atmospheric general circulation model is coupled to a comprehensive land surface model, a general circulation model of the ocean, and a dynamic sea ice model (Collins et al., 2006 a, b; Yeager et al., 2006). CCSM3 hence features a more detailed description of the exchange of buoyancy fluxes between the atmosphere and the ocean, which influence the formation of water masses. In addition, the parameterizations of convection, mixing, and diffusion in the ocean and atmosphere are more complex in CCSM3.

GENIE-1 uses 10 vertical layers and is far more diffusive than CCSM3 with 24 active layers for the Permian simulation. Validation against marine observations and sediment records is more problematic in GENIE-1 because large volumes of the ocean are homogenized, whereas biogeochemical processes in the ocean basins are heterogeneous (see also Archer et al., 1999).
Figure DR1: Dissolved oxygen (50-yr mean; in μmol L⁻¹) along an isopycnal surface (ρ = 1.027) for the PTB reference experiment (A), the 2xPO₄ experiment (B), the 10xPO₄ experiment (C), and the 10xPO₄ and enhanced pump experiment (D). The layer depth is about 600 m in the eastern equatorial Panthalassa, corresponding to the center of the oxygen minimum zone in the PTB reference experiment.

References:


