



Supplement of

Southern Ocean controls of the vertical marine $\delta^{13} C$ gradient – a modelling study

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SI 1A Particulate organic carbon cycling in HAMOCC2s

Particle export production in the surface layer (the euphotic zone), P_{POC} , P_{opal} or P_{CaCO3} [mol L⁻¹ yr⁻¹], is described by Michaelis-Menten kinetics for nutrient uptake as follows:

$$P_{POC} = \frac{v_{max}^{POC} \cdot [PO_4^{3-}]^2 \cdot r_{C:P}}{K_S^{POC} + [PO_4^{3-}]}, \ P_{opal} = \frac{v_{max}^{opal} \cdot [Si(OH)_4]^2}{K_S^{opal} + [Si(OH)_4]} \text{ and } P_{CaCO3} = P_{POC} \cdot R \cdot (1 - \frac{\frac{P_{opal}}{P_{POC}}}{S_{opal}})$$

- 5 where K_S^{POC} and K_S^{opal} are the half saturation constants for POC and opal respectively, $r_{C:P}$ is the Redfield ratio between carbon (C) and phosphorus (P) and V_{max}^{POC} and V_{max}^{opal} [yr⁻¹] are the maximum uptake rate of phosphate and silicic acid respectively. *R* is the maximum rain ratio C(CaCO₃):C(POC) and is set to the constant value 0.3405. P_{CaCO3} increases when $\frac{P_{opal}}{P_{POC}}$ lowers below threshold value for onset of CaCO₃ production S_{opal} . Furthermore, V_{max}^{POC} , V_{max}^{opal} , K_S^{POC} and K_S^{opal} are prescribed depending on sea surface temperature following Heinze et al. (2003).
- 10 The sinking of the exported biogenic particles is described by a mass balance between gains and losses, for POC:

$$\frac{dPOC_{settle}}{dt} = gains - losses. With$$

$$\frac{dPOC_{settle}}{dt} = P_{POC} - \frac{w}{\Delta z_0} \cdot POC_{settle} - r_{POC} \cdot POC_{settle}, \quad \text{for the surface layer,}$$

$$\frac{dPOC_{settle}}{dt} = w \cdot \frac{\partial POC_{settle}}{\partial z} - r_{POC} \cdot POC_{settle}, \quad \text{for all layers below the surface layer}$$

- This is done in the same way for CaCO₃, opal, and clay although clay is considered chemically inert and thus not degraded or dissolved. In the above set of equations, gains equal P_{POC} , P_{opal} or P_{CaCO3} and dust input for clay. Losses in the surface layer come from sinking and degradation/remineralisation, where the sinking term involves the sinking velocity w (3 m d⁻¹ in the control run, adjusted in the POC sinking rate experiments) and the thickness of the euphotic zone Δz_0 (50 m). The degradation term consists of a constant degradation rate r_{POC} (2.69 yr⁻¹) - with equivalents r_{opal} (1.23 yr⁻¹) or r_{caCO3} (6.76 yr⁻¹) for the other biogenic particles. Below the surface layer, gains are determined by the input of sinking particles from above (for POC, $w \cdot \frac{\partial POC_{settle}}{\partial z}$) and continued losses to inorganic carbon and nutrients through degradation/remineralisation (for POC, $r_{POC} \cdot POC_{settle}$). Remineralisation of POC requires oxygen availability above a minimum of 1e⁻⁵ moles/l. In addition to degradation CaCO₃ and opal dissolution is simulated, which depends on carbonate saturation and opal saturation respectively. Throughout the water column, POC losses are mirrored by source terms for the inorganic dissolved species of the water column (for TAlk, DIC, phosphate, and oxygen). The remainder of particles which are not subject to degradation within the water
- 25 column are deposited onto the ocean sediments. Here, the lowermost water column layer and topmost sediment layer directly interact through a sediment balance of deposition and redissolution/remineralisation.

SI 1B Calculation of gas transfer fluxes

Separate fluxes F_{up} and F_{down} are calculated by splitting the gas transfer formulation into two parts. The gas transfer formulation for CO₂ can be described as: $F_A = k_w ([A]_{water} - [A]_{air})$, where for A=CO₂, k_w is the specific gas exchange rate, [A]_{water} is the model surface-ocean free CO₂ concentration, and [A]_{air} is the atmospheric CO₂ concentration. In the model,

5 $F_{up}=k_w*[A]_{water}$ and $F_{down}=k_w*[A]_{air}$ and $F_{net}=F_A=F_{up}-F_{down}$. This splitting up is useful for the calculation of air-sea fractionation of the carbon isotopes, as the equilibrium fractionation factor only needs to be multiplied with F_{up} when calculating the effects of the air-sea gas exchange on δ^{13} C. It also proved to be useful for our discussion, because the total amount of exchange F_{u+d} influences δ^{13} C.

The freely evolving atmospheric concentration of O_2 , ¹³CO₂ and CO₂ is modelled by a one-layer box over each grid point, for which zonal average atmospheric concentrations are calculated at every time-step. Because of the annual model time-step, gas transport is simulated through meridional diffusion only, where the assumption is that the atmosphere is longitudinally well-mixed.



Figure S1 Surface map of the annually constant sea ice thickness in the model in meter.



Figure S2 Pre-industrial $\delta^{13}C$ data based on Eide et al. (2017b) at 200 m depth (top), a Pacific transect at 150° W (bottom left) and an Atlantic transect at 30° W (bottom right)



Figure S3 Atmospheric development of $\delta^{13}C^{atm}$ and pCO_2^{atm} during the (Global) sensitivity experiments. Note that the high POC sinking rate experiment was continued for an additional 10 000 years as compared to the other experiments. Small ongoing drift for 100 000s of years occurs in all experiments due to the very long equilibration timescale between burial and weathering for $\delta^{13}C$ (Roth et al., 2014).



Figure S4 The pCO_2 difference [ppm] between the surface ocean and the atmosphere for the model control run, 'Gas fast' and 'Gas slow' experiments. Negative values indicate a carbon flux into the ocean. The actual flux depends amongst others on the size of the pCO_2 difference and the air-sea gas exchange rate.



Figure S5 Modelled slow gas exchange sensitivity experiment δ^{13} C of DIC [‰] difference with the model control run: global experiments (a) and (c) and SO-only experiments (b) and (d), at 25 m depth (a) and (b) and as a Pacific transect of δ^{13} C difference (c) and (d). The sign of the change equals the sign of the δ^{13} Cdiseq.



Figure S6 Air-sea ${}^{13}C$ [gC m⁻² yr⁻¹] flux for the V_{max} experiment (left) and model control run (right). Negative values indicate carbon uptake by the ocean.



Figure S7 POC export production [gC m⁻² year⁻¹] for the Vmax experiment (left) and model control run (right).



Figure S8 The basins used to make Fig. 4. A=North Atlantic, B=South Atlantic, C=Southern Ocean, D=South Pacific, E=North Pacific, F=Indian Ocean.

SI 3 References

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