Universiteit Utrecht

Physics and Astronomy

Impact of Atlantic Meridional Overturning Circulation (AMOC) variability on ocean carbon uptake

Bachelor Thesis

Jelle van Oosterhout

Supervisor:

Prof. Dr. Henk Dijkstra
Institute for Marine and Atmospheric Research Utrecht

June 18, 2020

Abstract

The ocean is an important sink of CO_2, its uptake of this gas is partly controlled by the ocean circulation. In this thesis we develop a carbon cylce model with a variable atlantic meridional overturning circulation. We find with our model that a collapse of the AMOC leads to an increased uptake of CO_2, which in turn leads to a decreased atmospheric CO_2 concentration.

Contents

1 Introduction

The strength of the overturning circulation in the ocean is a important driver for the uptake of atmospheric CO_2 into the ocean surface waters[1].In the same article it is concluded that on a short time scale a (small) decrease in overturning leads to an decrease in outgassing. In an article by Nielsen it is concluded that over a larger time scale a switch in overturning circulation from a high to a low state leads to a net increase in outgassing, which contributes to an increased atmospheric CO_2 concentration[2].

In this thesis we look at the impact of the Atlantic meriodional overturning circulation variability on the uptake of carbon into the ocean. To do that we use two different box models. One simulation which models the carbon-cycle[3] and has a constant overturning and a second model which only simulates the atlantic meridional overturning circulation[4][5].

2 Theory

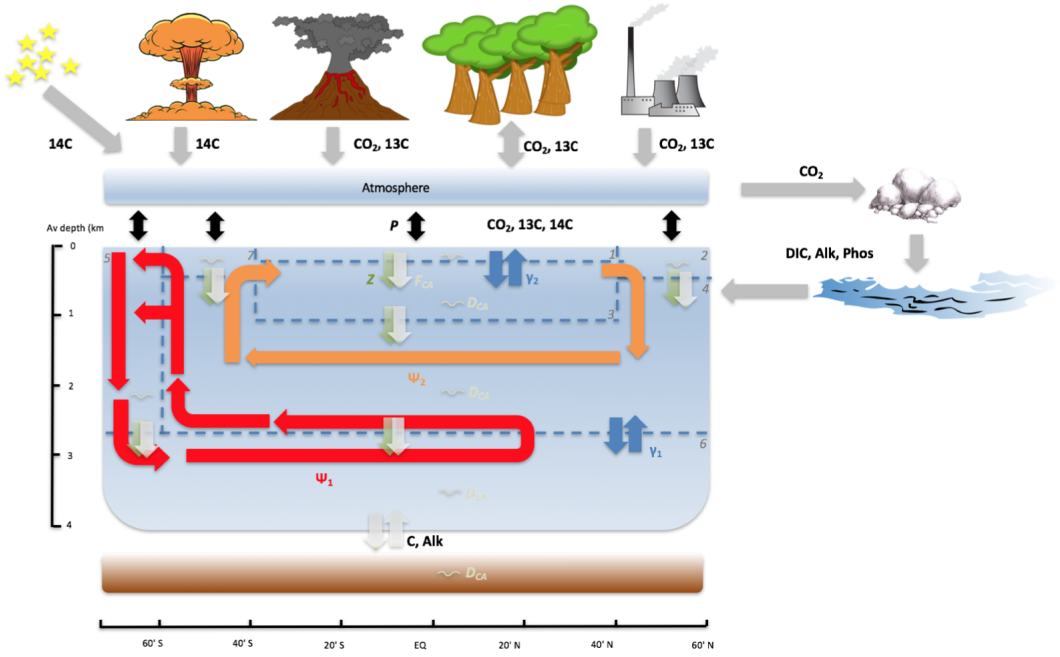


Figure 1: Overview of the processes stimulated in Neil's model, where Ψ_2 (orange arrows) are replaced by Cimatoribus' model, figure from Neil[3]

To model the carbon cycle we use the model developed by Neil [3]. Here the carbon cycle is modeled with the use of seven ocean boxes and one atmosphere box. The equations for the ocean boxes can be split up into the various contributions

name (box number)	latitude	depth
low latitude surface ocean (1)	40°S to 40°N	0 m to 100 m
northern surface ocean (2)	40°N to 60°N	0 m to 250 m
intermediate ocean (3)	40°s to 40°N	100 m to 1000 m
deep ocean (4)	61°S to 60°N	1000 m to 2500 m
southern ocean (5)	80°S to 61°S	0 m to 2500 m
abyssal ocean (6)	80°S to 60°N	2500 m to 6000 m
sub polar southern surface ocean (7)	61°S to 40°S	0 m to 250 m

Table 1: dimensions of Neil's boxes numbers refer to those found in figure1

$$\frac{d(\boldsymbol{C})}{dt} = \left[\frac{d\boldsymbol{C}}{dt}\right]_{phys} + \left[\frac{d\boldsymbol{C}}{dt}\right]_{soft} + \left[\frac{d\boldsymbol{C}}{dt}\right]_{hard} + \left[\frac{d\boldsymbol{C}}{dt}\right]_{gas} + \left[\frac{d\boldsymbol{C}}{dt}\right]_{weath} \tag{1}$$

Where $\boldsymbol{C}$ is a vector composed of the carbon concentration in the seven ocean boxes. In section 2.1 we will look into the first term, physical transport. In section 2.2 we look into the other terms, the carbon cycle composed of soft tissue, hard tissue, ocean-atmosphere gas exchange and weathering.

2.1 ocean circulation

In the sub-polar region (box 7) westerly winds drive both a transport of water mass to the north and an upwelling of deep water by ekman transport. In the north the water cools, sinks and is transported back south.
In Neil this circulation ($\Psi 2$, orange line in figure 1) is taken to be constant. we replace it with a separate model[5][4]. The general overturning circulation (Ψ_2), deep-abyssal mixing (γ_1) and the low-latitude thermohaline mixing (γ_2) are not changed.

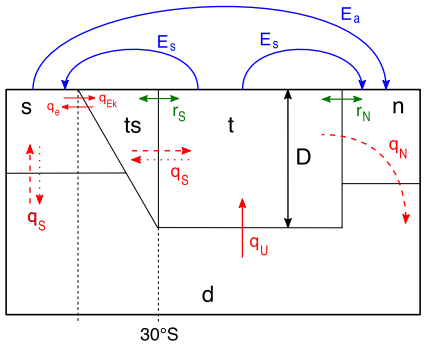


Figure 2: Structure of the box model which is used to calculate $\Psi 2$

In this model the circulation is controlled by winds in the south (as in Neil) and freshwater fluxes at the ocean surface. The model is governed by a set of salt conservation equations, i.e. the total amount of salt in the basin is constant

$$\begin{aligned}
\frac{dS_t}{dt} &= \frac{q_S(\theta(q_S)S_{ts} + \theta(-q_S)S_t) + q_U S_d - \theta(q_N)q_N S_t}{V_t} \\
&+ \frac{r_s(S_{ts} - S_t) + r_N(S_n - S_t) + 2E_s S_0}{V_t} - \frac{S_t A_t}{V_t}\frac{dD}{dt} \\
\frac{dS_{ts}}{dt} &= \frac{q_{Ek}S_s - q_e S_{ts} - q_S(\theta(q_S)S_{ts} + \theta(-q_S)S_t)}{V_{ts}} \\
&+ \frac{r_s(S_t - S_{ts})}{V_{ts}} - \frac{L_{xA}L_y S_{ts}}{2V_{ts}}\frac{dD}{dt} \\
\frac{dS_n}{dt} &= \frac{\theta(q_N)q_N(S_t - S_n) + r_N(S_t - S_n) - (E_s + E_a)S_0}{V_n} \\
\frac{dS_s}{dt} &= \frac{q_S(\theta(q_S)S_d + \theta(-q_S)S_s) + q_e S_{ts} - q_{Ek}S_s}{V_s} - \frac{(E_s - E_a)S_0}{V_s} \\
(A_t + \frac{L_{xA}L_y}{2})\frac{dD}{dt} &= q_U + q_{Ek} - q_e - \theta(q_N)q_N \\
S_d &= \frac{S_0 V_0 - (V_t S_t + V_{ts}S_{ts} + V_n S_n + V_s S_s)}{V_d}
\end{aligned} \tag{2}$$

where S_i, V_i and A_i are the salt concentrations, volumes and areas of the different boxes, E_s and E_a freshwater fluxes, L_{xA} and L_y the length and with of box ts and q_i are the fluxes between the ocean boxes. $\theta(x)$ is the Heaviside function which is defined as zero if x is smaller than zero and one everywhere else.
solving this system in time for some initial values of salt and depth of layer t (D) gives an time evolution for the salt concentration values and the depth of the top layer (D). subsequently the transport fluxes are calculated,

$$\begin{aligned}
q_{Ek} &= \frac{\tau L_{xS}}{\rho_0 |f_S|} \\
q_e &= A_{GM}\frac{L_{xS}}{L_y}D \\
q_U &= \frac{\kappa A_t}{D} \\
q_N &= \eta\frac{\rho_n - \rho_{ts}}{\rho_0}D^2 \\
q_S &= q_{Ek} - q_e
\end{aligned} \tag{3}$$

where density is defined as

$$\rho_i = \rho_0(1 - \alpha(T_i - T_0) + \beta(S_i - S_0)) \tag{4}$$

with α a thermal expansion coefficient and β a salinity contraction coefficient. To implement this in the model from Neil we modify the part governing the physical transport in equation

(1),

$$\left[\frac{dC}{dt}\right]_{phys} = \frac{(\Psi_1 \boldsymbol{T_1} + q_N \boldsymbol{Q_N} + q_U \boldsymbol{Q_U} + q_S \boldsymbol{Q_S} + \gamma_1 \boldsymbol{E_1} + \gamma_2 \boldsymbol{E_2}) \cdot \boldsymbol{C}}{\boldsymbol{V}} \tag{5}$$

or we can use a simpler version where q_N is taken to replace $\Psi 2$

$$\left[\frac{dC}{dt}\right]_{phys} = \frac{(\Psi_1 \boldsymbol{T_1} + q_N \boldsymbol{T_2} + \gamma_1 \boldsymbol{E_1} + \gamma_2 \boldsymbol{E_2}) \cdot \boldsymbol{C}}{\boldsymbol{V}} \tag{6}$$

here $\boldsymbol{C}$, $\boldsymbol{V}$, $\boldsymbol{T_1}$, $\boldsymbol{T_2}$, $\boldsymbol{E_1}$, $\boldsymbol{E_2}$, γ_1 and γ_2 are defined as in Neil[3]
q_S, q_U and q_N will be calculated using salt conservation (2), and $\boldsymbol{Q_N}$, $\boldsymbol{Q_U}$ and $\boldsymbol{Q_S}$ are new sparse matrices defined as:

$$\boldsymbol{Q_N} = \begin{pmatrix} 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & -1 & 1 & 0 & 0 & 0 & 0 \\ 0 & 0 & -1 & 0 & 0 & 0 & 0 \\ 0 & -1 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \end{pmatrix} \quad \boldsymbol{Q_U} = \begin{pmatrix} 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 1 & 0 & 0 & 0 \\ 0 & 0 & 0 & -1 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \end{pmatrix} \tag{7}$$

$$\boldsymbol{Q_S} = \begin{pmatrix} 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 1 \\ 0 & 0 & 0 & -1 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 1 & 0 & 0 & -1 \end{pmatrix} \tag{8}$$

2.2 carbon cycle

Carbon concentrations in the ocean are for a large part governed by two processes, exchange of CO_2 with the atmosphere and the consumption of carbon through biological activity, such as plankton. When this organic matter dies it sinks and partly dissolves back into the ocean water, the rest gets stored in the sediments.

The exchange of CO_2 between the ocean and the atmosphere is assumed to follow Henry's Law

$$[P] = K_H C, \tag{9}$$

where $[P]$ is the partial pressure of the gas in the atmosphere, C the concentration of the gas in the liquid and K_H a constant about how well the gas can dissolve in the liquid (solubility). For modeling purpose it is advantageous to have a flux, i.e. a change in concentration over a certain time period. This flux is estimated by a difference in pCO_2 between the ocean and the atmosphere, a gas transfer velocity (piston velocity) p and a solubility constant K,

$$\frac{dC}{dt} = pK_0(pCO_{2_{at}} - pCO_{2_{oc}}). \tag{10}$$

The second process, carbon consumption in the upper ocean which dissolves back in the deeper and upper layers, is known as the carbonate pump. The carbonate pump is split into two parts: one for soft organisms (second term equation 1) and one for hard shelled organisms(third term equation 1). For the biological production a nutrition based system is chosen, as in Zeebe [6]. Where the upwelling (γ) of phosphate determines the biological consumption of phosphate in the upper ocean with an efficiency term (η) which determines how well the upwelled phospate is utilised,

$$P_{producton} = \eta * \gamma * [P_{intermediate}] \tag{11}$$

and via the redfield-ratios the biological production of the other elements.
Except for the production factor, which is from Zeebe[6], we model the hard and soft tissue carbonate pump as in Neil[3]. Most of the soft organic matter is dissolved back in the upper layers of the ocean, for which an emprical relation is used[7],

$$F = F_{100}(\frac{z}{100})^b \tag{12}$$

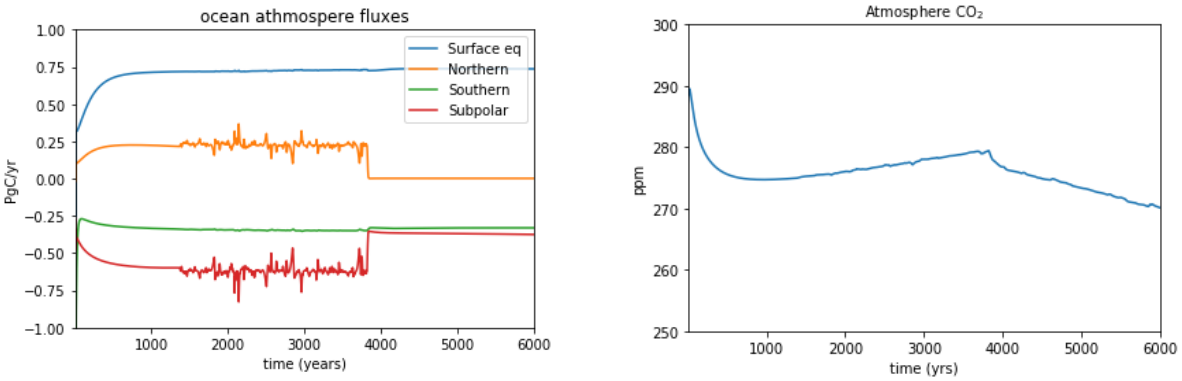
where F is the biological flux of an element at depth z, b a depth scalar and F_{100} the biological production of this element at a depth of 100 meters (as calculated by equation 11). This is implemented in the box model by taking the difference of the influx at the top of a box and the outflow at the bottom,

$$\left[\frac{dC_i}{dt}\right]_{soft} = F_{in} - F_{out} = \frac{Z_t S_t(\frac{d_{ci}}{100})^b}{V_i} - \frac{Z_t S_t(\frac{d_{fi}}{100})^b}{V_i}, \tag{13}$$

Where Z_t and S_t are respectively the production of carbon (via redfield ratio related to equation11) and surface area of the surface box located above box i. d_{ci} and d_{fi} are the height of the ceiling and floor of box i in meters and V_i is the volume of box i in cubic meters. It is assumed in Neil that the amount of hard shelled organic matter produced is related to the amount of soft organic matter via the rain ratio (F_{CA}). This matter is than partly dissolved back into the ocean depending on the concentration of $CaCO_3$ the rest is stored in the sediments and can be dissolved back into the lowest box depending on the concentration of $CaCO_3$. So for the flux between the boxes we get,

$$\left[\frac{dC_i}{dt}\right]_{hard} = \frac{F_{CA} Z_i S_i}{V_i} + (\zeta + \epsilon) CaCO_3 \tag{14}$$

where F_{CA} (=-0.07) is the rain ratio, S and V the production area and volume of the boxes and Z the biological production as calculated in equation 11. Z is only non zero for the top boxes. ζ and ϵ are constant and concentration dependent dissolution terms.



(a) Outgassing and uptake of the upper ocean

(b) Athmospheric CO_2 concentration

Figure 3: A figure with two subfigures

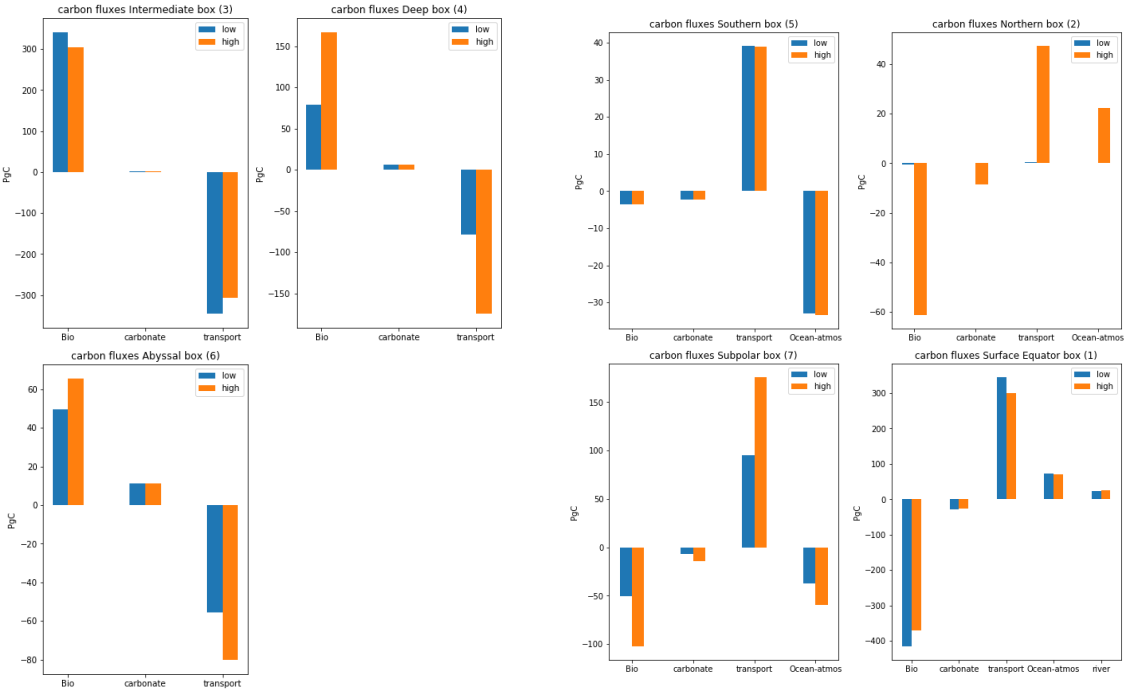
3 Results

In this section we take a look at the impact of an AMOC collapse on the various carbon processes in our combined model. To achieve this collapse we introduce some random noise on the fresh water flux E_a which lead to a switch from the circulation around 4000 years from a high to a low state, and let the model run for a total of 6000 years.

In figure 3a we see that in the first couple of hundred years the fluxes gets to a steady state then around 4000 years (where the switch to low overturning is made) we see an abrupt transition. The uptake in the northern region decreases and the outgassing in the subpolar region also decreases. In the atmospheric CO_2 concentration, figure 3b, we also see a change but here it is not as abrupt because land sources of CO_2 dampen the effect of the ocean.
For comparing the low and high states we have chosen two different hundred year periods, from 1000 until 1100 for the high overturning state and the last hundred years for the low overturning state. A positive flux means that carbon goes into the box, a negative flux means carbon goes out of the box, for the box numbers see figure 1 and table 1.

As can be seen the biggest impact on the carbon fluxes is in the boxes where the transport flux is changed, i.e. box 2, 3, 4 and 7, in these boxes we observe, as can be expected, a change in the carbon fluxes due to transport. In the upper boxes we also find that the reduced upwelling of nutrients leads to a reduced biological production (figure 4b), and thus a reduced export and a reduced influx in the lower boxes (figure 4a)

Due to changed circulation in the ocean the uptake and outgassing of carbon also changes as we could see in figure 4b. Globally this leads to a net increase in uptake of carbon into the ocean, which is composed of a reduced uptake in the northern region (box 2) which is overcompensated by reduced outgassing in the southern sub polar region (box 7).



(a) subsurface boxes

(b) Surface boxes

Figure 4: Uptake and outgassing of carbon from the ocean boxes.

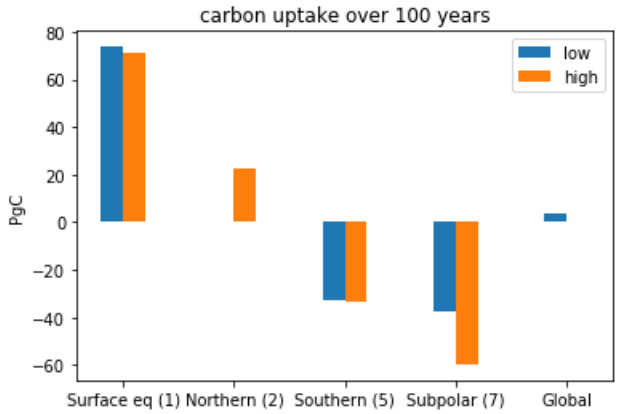


Figure 5: net uptake or outgassing of the ocean for the low and high overturning state

4 Discussion

We have looked how the oceanic carbon cycle reacted to a collapse of the AMOC. The short term reaction is difficult to compare because our switch is abrupt. For the reaction on the longer time scale, table 2, we found that a global increase in uptake is composed of different regional contributions. We can compare our findings with those found by Nielsen (table 3)[2]. Although the exact numbers are difficult to compare, the effects of the change in circulation can be compared.

in the southern regions (box 5 and 7) we find a decrease in the net outgassing which corre-

state	flux	south (5) 80-60°S	subpolar (7) 60-40°S	surface eq (1) 40°S to 40°N	north (2) 40-60°N	global
high	(15 Sv)	-0.33	-0.61	0.73	0.22	0.006
low	(0.1 Sv)	-0.33	-0.37	0.72	0.002	0.03
difference		0	0.24	-0.01	-0,218	0,024

Table 2: average yearly uptake (+) and outgassing (-) of the different boxes in PgC/yr for the high and the low state and the difference between them (low-high). ($Sv = 10^6 m^3$)

Case	flux	90-44°S	44-18°S	18°S to 18°N	18-49°N	49-90°N	global
high	(12.5 Sv)	-0.142	-0.437	-1.053	0.496	0.224	-0.038
low	(4.5 Sv)	-0.089	-0.472	-1.107	0.495	0.180	-0.049
difference		0.053	-0,035	-0,054	-0,001	-0,044	-0.011

Table 3: average yearly ocean-atmosphere fluxes from Nielsen[2] from the high and low state and the difference between them (low-high). ($Sv = 10^6 m^3$) (Note: the signs are changed compared to the paper for comparison)

sponds with Nielsen's region south of 44°S. In the northern region (box 2) we find a decrease in uptake as does Nielsen (49-90°N). In our simulation the equator region (box 1) has a stable uptake, in Nielsen we see that in particular the region from 44°S till 18 °N has an increase in outgassing. The difference in response from the equator region leads to a different global picture, we observe an increase in uptake while Nielsen observes an increase in outgassing. Due to this difference the athmopsheric CO_2 concentration decreases in our model where it increases in Nielsen. One explanation for these differences might be that the low-latitude thermohaline mixing (γ_2 from figure 1) is constant, and as such the response to the switch in circulation is not present in the upper ocean equator region. Developing a model where the thermohaline mixing is also varied might further improve the model.

References

[1] T. DeVries, M. Holzer, and F Primeau. "Recent increase in oceanic carbon uptake driven by weaker upper-ocean overturning." In: *Nature* 542 (2017), pp. 215–218. DOI: https://doi.org/10.1038/nature21068.

[2] Søren B. Nielsen et al. "Two-Timescale Carbon Cycle Response to an AMOC Collapse". In: *Paleoceanography and Paleoclimatology* 34.4 (2019), pp. 511–523. DOI: 10.1029/2018PA003481.

[3] Cameron M. OŃeill et al. "[simple carbon project] model". In: (July 2018). DOI: 10.5281/zenodo.1310161.

[4] Daniele Castellana et al. "Transition Probabilities of Noise-induced Transitions of the Atlantic Ocean Circulation." In: *Scientific Reports* 9.20284 (2019). DOI: https://doi.org/10.1038/s41598-019-56435-6.

[5] Andrea A. Cimatoribus, Sybren S. Drijfhout, and Henk A. Dijkstra. "Meridional overturning circulation: stability and ocean feedbacks in a box model". In: *Climate dynamics* 42 (Jan. 2014), pp. 311–328. DOI: https://doi-org.proxy.library.uu.nl/10.1007/s00382-012-1576-9.

[6] R. E. Zeebe. "LOSCAR: Long-term Ocean-atmosphere-Sediment CArbon cycle Reservoir Model v2.0.4". In: *Geoscientific Model Development* 5.1 (2012), pp. 149–166. DOI: 10.5194/gmd-5-149-2012. URL: https://www.geosci-model-dev.net/5/149/2012/.

[7] John H Martin et al. "VERTEX: carbon cycling in the northeast Pacific." In: *Deep-Sea Research* 34.2 (1987), pp. 267–285.