On the connection between upwelling, opal deposit rates and atmospheric CO_2 concentration

Bachelor's Thesis

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Abstract

The theory that the Southern Ocean dominates the carbon exchange between the sea and the atmosphere has been challenged in the recent years. Strong evidence for this theory has been the opal flux record from the sediment, which is said to be a proxy for enhanced wind-driven upwelling. Opal as upwelling proxy is tested in the fully coupled CESM simulation under the orbital configuration from 113 ka BCE. The results show that the limiting factor for the biogenic silica burial in this region is iron, not silicate as is hypothesized. The concentration of iron in the ocean in the deglaciation period is reduced across the coast of Antarctica. This decrease is found to be caused by enhanced southward meridional water speeds, resulting in a phenomenon of nutrient trapping found in other recently published simulations.

Contents

1 Introduction

The complex interplay of the climate with the atmospheric and oceanic elemental cycles remains unsolved to this day, although much is already known. Evidence from the sediment and ice cores reveals a cycle that dates back approximately half a million years, involving atmospheric partial pressure of CO_2 (pCO_2) in correlation with global temperature and ice volume,[5] most notably Greenland and Antarcitca temperatures (Dansgaard-Oeschger events[4]). What drives this cycle is still not understood, although there are many contesting theories attempting to explain its underlying physical and biogeochemical processes.

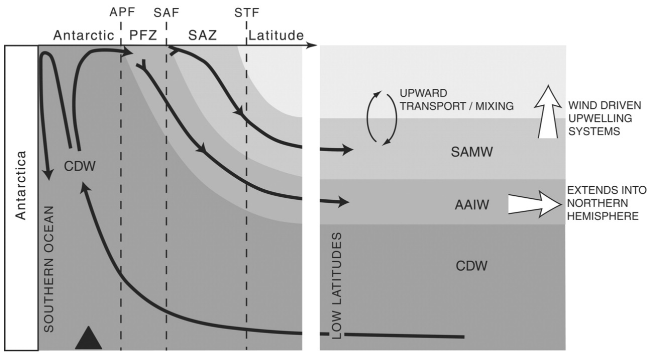
The still, bottom waters of the ocean are where dissolved nutrients such as silicate or phosphate sink down until they are brought to the surface again through upwelling. Those waters also contain dissolved inorganic carbon (DIC) reservoirs, which reach the surface through the same mechanism. The coast of Antarctica is a major upwelling site on Earth, which has lead scientists to believe that the Southern Ocean is responsible for the pCO_2 rise after a glacial period.[1] But is it? In recently published fully coupled Community Earth System Model (CESM) simulations, the Earth's ocean was analyzed under the insolation corresponding to 113 ka BCE.[4] The results showed that there are two sources of atmospheric pCO_2 increase that equally contribute to the overall rising pCO_2 effect. The first, caused by precipitation changes as a result of the Atlantic Meridional Overturning Circulation (AMOC) collapse, is due to increased terrestrial CO_2 outgassing in the tropics. The second, which happens over a longer period of time, is due to increased CO_2 release from the ocean. Surprisingly, the main increase of CO_2 outgassing is observed not in the Southern Ocean, but in the Atlantic. In the Pacific sector, there is a slight increase in outgassing, however the two other sectors in the SO - Indian and Atlantic - show enhanced uptake of the CO_2. This implies that the Southern Ocean does not play as big a role in the rise of the atmospheric pCO_2 as it is believed.

2 Theory

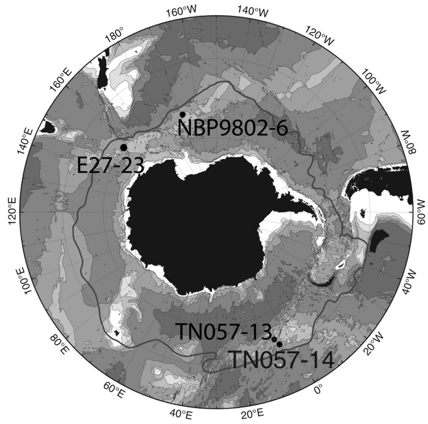
2.1 Biogenic Opal as Upwelling Proxy

A study published in 2009 by *Anderson et al.*[1] suggested a neat way of determining the history of wind-driven upwelling from the sediment core data. The coupling between the Antarctic temperatures and atmospheric CO_2, together with the fact that about 80% of the ocean waters upwell in the Souther Ocean, has lead many scientists to believe that the SO plays a vital role in regulating deep sea CO_2 outgassing into the atmosphere. Until the 2009 study was published, though, there was no direct evidence of the Southern Ocean control. Biogenic opal as a proxy for upwelling was proposed on the basis of diatom activity. There is a linear correlation between the opal burial and the amount of diatoms in the euphotic zone, for which the limiting factor is the concentration of dissolved silicate in the upwelled deep sea water. It was therefore inferred that higher opal flux indicates enhanced wind-driven upwelling. Silicate is not the only material delivered to the surface with this process - DIC from the deep sea accompanies the upwelled nutrients, which provides a chance for CO_2 release into the atmosphere. Hence, a proxy for this process provides a powerful tool that links changes in wind stress to the role of the ocean in the atmospheric carbon variability.

With those claims in mind, the team of scientists analyzed three sediment cores around the shore of Antarctica located south of the Antarctic Polar Front or the APF (fig. 1a). This choice was made based on the relative accumulation rates of opal - the maximum zone of biogenic silica production lays just south of the APF. Moreover, the three cores were taken from the three sectors of the Southern Ocean: Atlantic, Pacific and Indian (fig. 1b). An additional core was also analyzed in the Pacific sector to give a profile of opal fluxes for the period from 90 to 20 ka BCE. The results of the analysis showed increased upwelling during the deglacial period between 17,000 and 10,000 ka BCE. The flux of biogenic opal to the sediment in this time period was increased in all four examined cores, and consistent with other sediment proxy data. The three investigated cores and the resulting fluxes are showed in figure 2.



(a)



(b)

Figure 1: Figures from *Anderson et al.*[1] showing the locations of the sediment cores used for proxy analysis; (a) is a diagram showing the upwelling of deep sea water at the coast of Antarctica; the black triangle indicates the approximate latitude of the sediment cores; (b) shows the location of three analyzed sediment cores, as well as one (TN057-14PC) which was used to create an extensive opal flux record; the black solid line marks the APF.

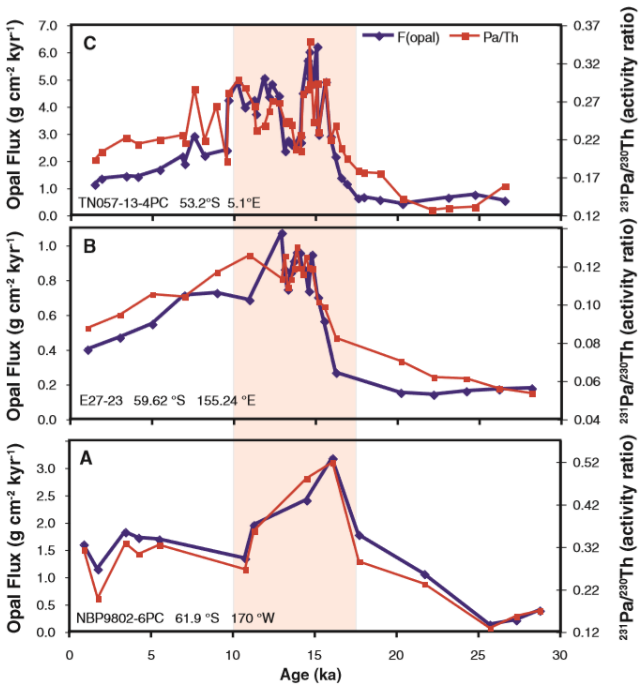


Figure 2: The locations of the cores examined by *Anderson et al.*[1] and the respective plots of the opal flux to sediment.

The results showed clear flux changes in the deglacial period, and the extended profile of opal fluxes demonstrated agreement with other proxy data. However, the recent work of *Nielsen et al.*[4] offers that in the deglacial state of the fully coupled Earth system model there was no increased carbon dioxide outgassing across the SO, except for the Atlantic. This result refutes the theory about the wind-driven upwelling being the source of increased pCO_2 in the deglacials, and puts the biogenic opal as a proxy to question. To uncover which processes regulate the flux of silica into the sediment, I dive into the complex marine cycle of silicate, and determine what else might be carried in the opal flux signal found in the deposits at the bottom of the Southern Ocean.

2.2 Ekman Transport

The ocean is abundant in biological activity which relies on access to light and nutrients. There is one kind of life crucial to the transport of silicon across the ocean waters: diatoms. The phytoplankton that lives in the euphotic zone relies on the silicic acid, nitrite, phosphate and iron, which get upwelled together with the deep sea water.

Upwelling is a process that occurs mostly around the equator and the low latitudes, and is a result of a phenomenon called the Ekman transport, which arises due to wind stress. Ekman transport describes the effect wind stress and the Earth's circulation has on the top 10 to 100 meters of the sea water. It is theoretically derived from the simplified equations of motion:

$$a_x = 0 = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv + \frac{1}{\rho}\frac{\partial \tau^x}{\partial z} \tag{1}$$

$$a_y = 0 = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu + \frac{1}{\rho}\frac{\partial \tau^y}{\partial z} \tag{2}$$

$$a_z = 0 = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g, \tag{3}$$

where $f = 2\sin\Omega$, g is gravitational acceleration, ρ is the mean density of the ocean, x, y, z are zonal, meridional and depth coordinates respectively, and u and v are the zonal and meridional speeds. The force of wind on the sea water is represented as the horizontal stress τ_0 acting on the horizontal surface of the ocean. These equations allow for finding the net Ekman mass transport due to wind stress in the zone where the frictional influence is non-negligible, which stretches from the surface to the depth of about 100m. Assuming latterally constant pressure, we end up with the Ekman balance:

$$0 = fv + \frac{1}{\rho}\frac{\partial \tau^x}{\partial z} \tag{4}$$

$$0 = -fu + \frac{1}{\rho}\frac{\partial \tau^y}{\partial z}. \tag{5}$$

So for the net Ekman mass transport M^{Ek} of a kilogram of water per second per a meter wide section of the water column:

$$M_x^{Ek} = \frac{\tau_0^y}{f} \tag{6}$$

$$M_y^{Ek} = \frac{\tau_0^x}{f}. \tag{7}$$

The water mass displaced from the surface at one point of the ocean must be replaced by water from another source due to mass conservation. Indeed, the Ekman transport induces downwelling (Ekman pumping) or upwelling (Ekman suction), depending on the direction of the wind. In the case of the Southern Ocean, the waters at the coast of Antarctica experience upwelling due to the wind stress of the Westerlies at the lower latitude. The Ekman transport due to the Westerlies pulls the water away from the coast and brings the Circumpolar Deep Water (CDW, see fig. 1a) to the surface. Models of the meridional overturning circulation reveal a global trend of waters at intermediate depths 1000 to 3500m gradually flow towards the south.[6] Hence, the CDW is rich in nutrients, DIC and other various materials that sink into the deep sea all around the global ocean. This is a perfect spot for biological activity, and observations show that indeed, the largest concentrations of the silicic acid and marine diatoms lay at the coast of Antarctica.

2.3 Silicate Cycle

When looking at elemental cycles in the ocean, we construct an equation which relates the time rate of change of a tracer (any element or compound) concentration at a given point and the processes that change this concentration. Such an equation is called transport concentration equation and has a form (for a volume of water at a fixed location):

$$\frac{\partial C}{\partial t} = \left.\frac{\partial C}{\partial t}\right|_{\text{advection}} + \left.\frac{\partial C}{\partial t}\right|_{\text{diffusion}} + SMS(C) \tag{8}$$

The advection term refers to the flow of water through the volume at the location and how that affects the tracer concentration. The diffusion is the flux that occurs to molecular and eddy diffusions. Finally, the term $SMS(C)$ describes the internal sources minus sinks of the tracer.

As simple as the equation looks, the biogeochemical processes in the sea make the relation far from it. Therefore it is useful to divide the ocean in three boxes and investigate the cycle in each of them. Those boxes are the surface (stretching across the depth of the euphotic zone), the water column and the sediment. To fully understand why opal is a good candidate for an upwelling proxy in the SO, in this section I will revise the processes which regulate the amount of silicon in the sea water at these three depths, and bring up theoretical values and available data (based on *Sarmiento et al.*[6]).

In the ocean, silicon exists primarily as silicic acid ($Si(OH)_4$), which is utilized by diatoms to form their cell wall, or frustule. The frustule is composed of biogenic opal (or biogenic silica), $SiO_2 \cdot nH_2O$. The solubility of opal ($[Si(OH)_4]_{SAT}$) is defined as the concentration of silicic acid in sea water that is in thermondynamic equilibrium with opal. This variable is sensitive to temperature, the specific surface area (SSA, defined as the surface area of the opal per unit mass), pressure, pH and the concentration of aluminium in the sediment. The change of opal concentration with time is given by the equation:

$$[SiO_2] = [SiO_2]_{t=0} \exp(-kt) \tag{9}$$

From observations we know that k ranges from $1/4\,\mathrm{d}^{-1}$ to $1/40\,\mathrm{d}^{-1}$. This shows that the upper limit of opal dissolution in the water column, given diatoms sinking at a rate of $100\,\mathrm{m\,d}^{-1}$ over the ocean depth of 5000m, is 70-100%. The silicic acid in the ocean is used almost exclusively by diatoms, and is mostly not passed up the food chain. Moreover, the regeneration of opal to silicic acid happens because of dissolution, instead of biological degradation. The only sink for silicon in the ocean is the burial of the biogenic silica in the sediment. This seems promising for a proxy material, however theory shows that on average only about 3% of the opal production becomes sedimentary rock. Luckily for us, the Southern Ocean proves to be exceptionally good at preserving silica in the sediment.

In the euphotic zone, the diatoms compete with other phytoplankton for resources. What they need is light and iron for the photosynthesis, and nitrate (or phosphate) and silicate to form their frustules. The frustules are theorized as the reason why the diatoms are successful compared to other species of phytoplankton. In general, the macronutrients nitrate and phosphate come in similar concentrations across the ocean, and any growth limitations in observations largely correspond to the limitation of nitrate rather than phosphate. The iron concentration influences diatom's cell size, which in turn changes the Si:N uptake ratios. Cells with high Si:N ratios that grown under low iron conditions are small and grow slower, but still form opal. The global mean production of opal in the euphotic zone is $2.1\,\mathrm{mmol\,m}^{-2}$, but the values vary greatly from $2\,\mathrm{mmol\,m}^{-2}$ in the open ocean to $90\,\mathrm{mmol\,m}^{-2}$ in the Southern Ocean and coastal upwelling regions. Opal export into the water column ranges according temperature: 11% in warm regions to 85% in cold regions, with global average of 47%. The importance of diatoms is well known - it is estimated that they are responsible for between 20% and 90% of organic matter export into the water column.

About 25% of the opal that leaves the surface dissolves in the water column. Here also lays the crucial pathway for resupplying the silicic acid into the upper ocean: the Subantarctic Mode Water (SAMW). This water layer is formed out of the upwelled CDW (see figure 1a) and mixes with the surface, re-supplying nutrients into the euphotic zone. There is an interesting observation to be made about the silicic acid in this region: comparing the global Si:N ratios, we find $\mathrm{N} \gg \mathrm{Si}$, but in the Southern Ocean it is the opposite. The drawdown of the silicic acid from the upwelled CDW is more efficient than the drawdown of nitrate in this region. The coast of Antarctica is a unique spot on Earth in many ways, one of the aspects being the limitation of light and iron. This limitation of iron is theorized to be the reason why diatoms utilize Si to a much greater extent than N in the Southern Ocean, and the large Si:N ratios explain the distribution of the silicic acid on Earth. Indeed, the diatoms are so efficient at utilizing the silicic acid, they completely drain the SAMW South of the Antarctic Polar Front, causing the concentration of Si to be almost nonexistent in the open ocean.

About 25% of opal leaving the euphotic zone arrives in the sediment. In the SO, the efficiency of burial is high (30%), but the flux of the silicic acid out of the sediment is also higher than average (-2.9 to -5.6 $\mathrm{mol\,m^{-2}\,d^{-1}}$). The dissolution rate constant of opal decreases with its age, as the number of reactive sites on the diatom surface declines and they become increasingly blocked by contaminants such as aluminium. The sediment kinetics are largely influenced by opal rain at higher burial efficiencies and by opal chemistry at lower burial efficiencies. Therefore, in the case of the Southern Ocean, the deciding factor in the burial efficiency will be the amount of opal rain falling into the sediment. Since opal solubility also depends on temperature and, as discussed above, in the cold regions, opal export into the water column can reach up to 89% of the silica produced in the euphotic zone, overall the Southern Ocean seems like a perfect candidate for investigating the sedimentary opal signal.

The silicate cycle discussion would not be complete without mentioning iron. This element is a micronutrient, meaning it exists in the sea in concentrations of the order of $\mu\mathrm{mol\,m^{-3}}$. As mentioned above, iron is necessary for the crucial processes of photosynthesis and respiration. Reduced supply of iron may lead to reduced growth rates, and it is crucial for larger phytoplankton such as diatoms, as they rely on the iron supply to compete with more efficient picophytoplankton. Iron is highly reactive, and it gets removed by biological activity and scavenging very fast, so fast that most of the river input gets depleted locally. Therefore, the main sources of iron in the ocean are aeolian dust transport and dissolution from continental margin sediment. The iron concentration increases with depth, meaning that it gets re-supplied into the surface similar to other nutrients. About 75% of the iron supply in the surface comes from upwelling, but the total local iron concentration is dependent on the dust flux into the sea. Iron is limited in the SO due to low aeolian dust transport.

3 Experimental Setup

For the analysis of the silicate cycle, I use the fully coupled Community Earth System Model (CESM) output generated by Søren Nielsen. The model was spun up for a 1000 years, using pre-industrial orbital forcing. After a 1000 years, the simulation branches out in two - NOW and THEN. NOW is a control set, where no perturbations are introduced. THEN runs under orbital forcing of 113,000 years before the Common Era. The model runs for another 1000 years, and the results used for the analysis average over the last 100 of those years.[4] NOW represents the current, pre-industrial Earth, while THEN represents the planet in the warming, deglacial stage. In order to investigate how the perturbation influences the variables within the simulation, the difference THEN-NOW is used.

The following outline of the model setup is based on *Nielsen et al.*,[4] where further description of the parameterization can be found. In the model, the orbital forcing results in the collapse of the Atlantic Meriodional Overturning Circulation (AMOC). Within the first two centuries of introducing the insolation perturbation, the AMOC strength at 26°N decreases from 12.5 Sv to below 5 Sv. Historically, the process of breaking internal waves in the ocean has been represented in models as a constant diffusivity. This process converts turbulent kinetic energy to potential energy and is hypothesized to be influenced by the AMOC collapse. The internal mixing has an influence on the biogeochemistry, which means that constant parameterization of internal wave breaking results in loss of information. Therefore, the output is generated in a model with a recently proposed vertical mixing parameterization, IDEMIX, where the internal wave energy and the dissipation rate are calculated for each grid cell in the sea. The specific model used to generate the output is the coarse resolution version of CESM ver. 1.2. The ocean resolution depends on latitude, as the zonal grid size ranges from 400 km near the Equator to 20 km at the poles. The resolution is also dependent on longitude, varying from 400 km to 40 km. The largest values of meridional grid sizes are at the North Pacific. The resolution also decreases with depth. There are 60 vertical layers, with thickness of 10 m at the surface and of

500m at the bottom. The biogeochemistry is fully coupled with the model, exchanging the carbon between ocean, land and atmopshere. In the ocean, the model includes diatoms, phytoplankton and diazotrophs. The phytoplankton growth is governed by temperature, light and nutrients available in the model: nitrate, phosphate, silicate and iron.

In order to compare the model output with existing observations, I use World Ocean Atlas 2013 (WOA13) observations on the silica concentration in the sea.[8] The WOA13 data is shared online by National Oceanic and Atmospheric Administration (NOAA). The data set contains the statistical mean of all ocean observations available to scientists today. This particular data set averages over a period of a year. I will compare the NOW output with observations of the silica distribution in the ocean. This will shed some light on the agreement of the model results with observations.

4 Results

4.1 Opal and Wind Stress

Figure 3 shows the flux of biogenic opal to the sediment and the zonal wind stress in NOW and THEN-NOW. From these four figures, it is apparent that the flux to the sediment does not increase in the Atlantic sector of the Southern Ocean, and that this change cannot be explained by the wind stress change. Between the longitude of 20°E and 70°E in THEN-NOW, there is a drop in flux, as well as substantial rise in the wind stress. There are also instances of decrease in flux of 0.04 - 0.24 $\mathrm{g\,cm^{-2}\,kyr^{-1}}$ in the Western hemisphere, where the wind stress is exclusively enhanced. This demonstrates that the changes in wind stress alone cannot explain the variance of the opal flux to sediment in THEN-NOW.

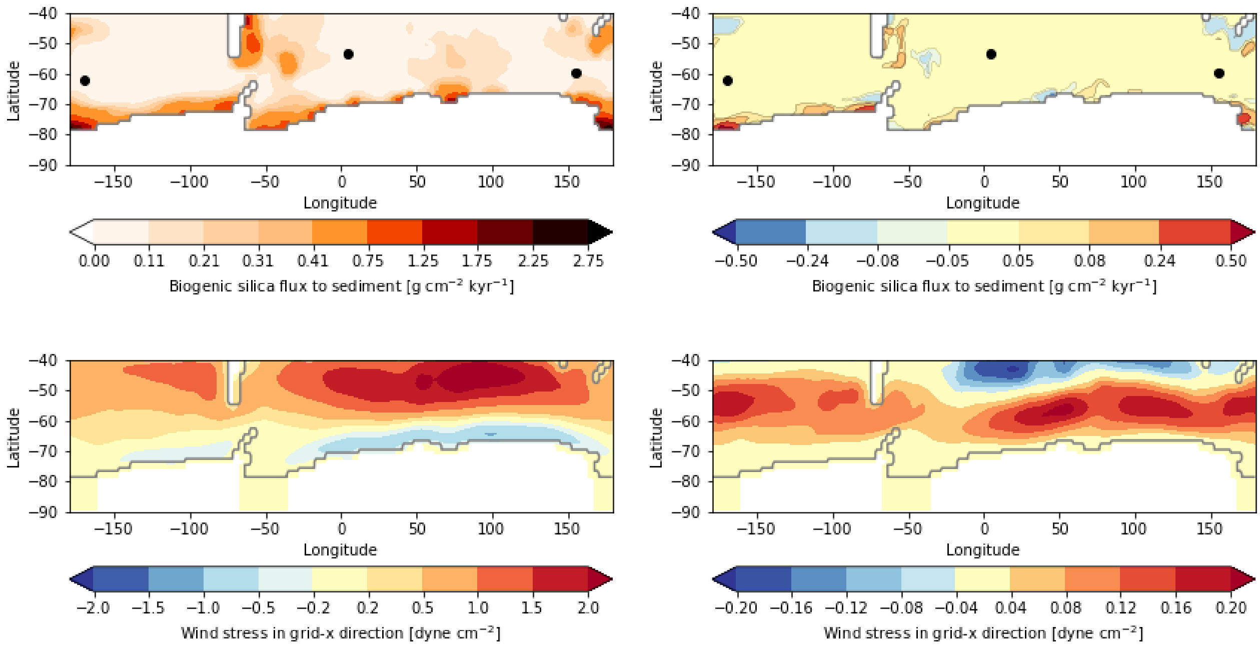


Figure 3: Top: the flux of biogenic silica in NOW (left) and THEN-NOW (right). Bottom: the zonal wind stress in NOW (left) and THEN-NOW (right). The black points represent the three cores investigated by *Anderson et al.*[1] As the changes in the flux are relatively small, the contours are added for easier readability. The opal flux maximum located west of 150°W at the coast of Antarctica has value of 0.72 $\mathrm{g\,cm^{-2}\,kyr^{-1}}$.

4.2 Diatom Growth Limitation

A further insight into what might be the cause of the drop in the flux can be provided by examining the diatom growth limitations. The limitation variable may have a value from 0 to 1, 0 corresponding to no growth, and 1 to full growth of the diatoms. The limitation of 0.5 then means that half the diatoms can grow under these circumstances compared to the growth under limitation equal to 1. The limitation variables in the CESM output vary in depth. The 15 different depths range from 5m to 145 m, in 10m intervals. The depth of main diatom activity is located to determine which layer is most relevant for investigation. This is done by examining the diatom chlorophyll variable showed in figure 4. It is evident that the majority of the diatom activity in the Southern Ocean occurs within the depth of 25m from the surface. Therefore the investigation of the diatom limitations is focused at 15m. The following figures exhibit the variable values at this depth, unless stated otherwise.

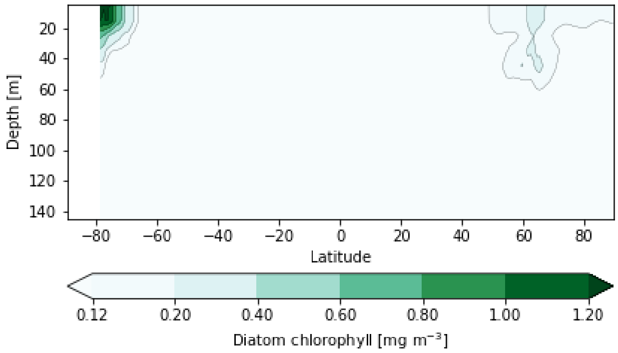


Figure 4: Zonal mean of diatom chlorophyll in the ocean.

The two main nutrients that the diatoms need in order to create their frustules are silicate and nitrate (or phosphate). As discussed in the Silicate Cycle section, the nitrate is more likely to be the limiting nutrient at large, and hence this analysis centers around the nitrate rather than phosphate. Figure 5 shows the silicate limitation in NOW and THEN-NOW, and figure 6 demonstrates corresponding results for the nitrate limitation.

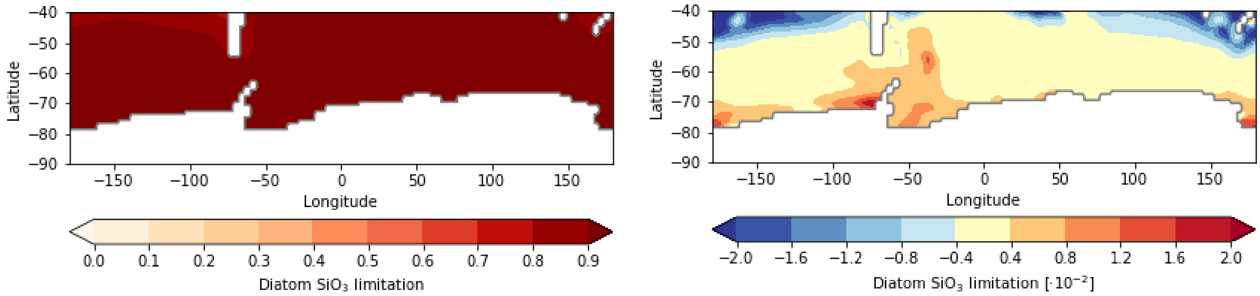


Figure 5: Diatom silicate limitation at the surface in NOW (left) and THEN-NOW (right).

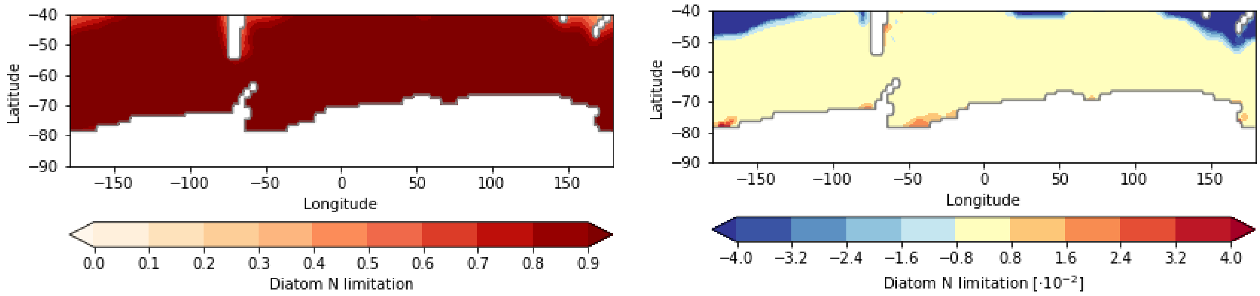


Figure 6: Diatom nitrate limitation at the surface in NOW (left) and THEN-NOW (right).

While the silicate limitation rises across the Southern Ocean by up to 2%, the silicate limitation is

1 nearly everywhere between 50°S and 90°S in NOW, only falling below 0.9 at the coast of South America. The silicate limitation is one that changes the least out of the five limiting variables. If the silicate regulated the opal flux into the sediment, we should observe a drop in the limitation in the vicinity of 50°E. However, there is no drop in the silicate limitation anywhere in the SO, indicating that another limiting factor is in control of the opal flux.

The nitrate exhibits similar high limitation values across the Southern Ocean, with few instances of limitation below 0.9 right by the coast of Antarctica. It shows higher variability than the silicate, going up to 5%. However, there is no decrease in limitation at high southern latitudes. Furthermore, similarly to the silicate limitation, there is no change in the vicinity of the opal flux decrease.

As discussed in the Silicate Cycle section, the two remaining factors that have influence on the growth of diatoms are light and iron. Figures 7 and 8 show the limitations of these variables respectively.

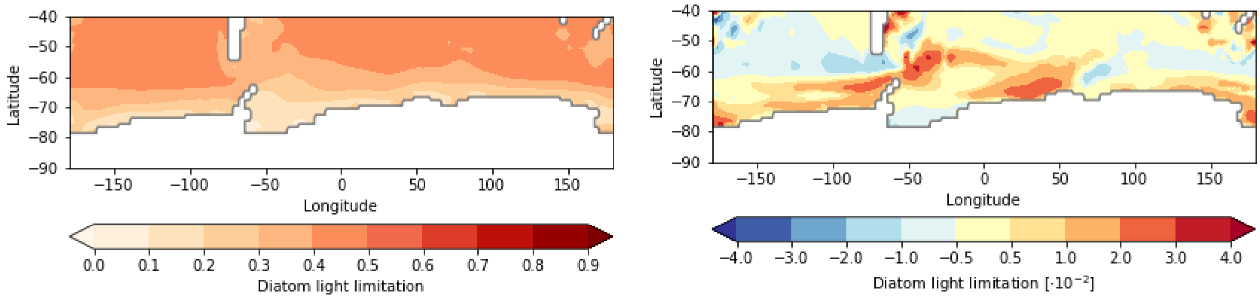


Figure 7: Diatom light limitation at the surface in NOW (left) and THEN-NOW (right).

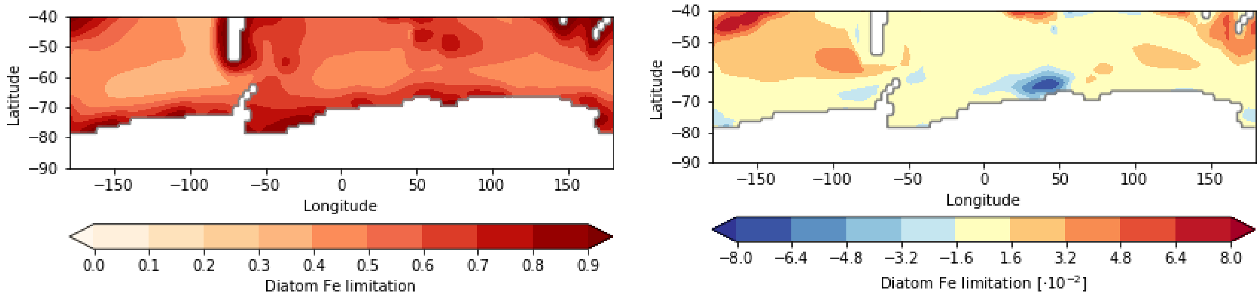


Figure 8: Diatom iron limitation at the surface in NOW (left) and THEN-NOW (right).

The diatom light limitation in NOW shows the phytoplankton cannot grow to full capacity anywhere in the Southern Ocean. The maximum value of the light limitation here is 0.48. The light limitation variability is comparable to the nitrate, going up to 5%. The light limitation is generally enhanced at the coast of Antarctica, and substantially increased between 0° and 50°E. This suggests that light is not the limiting factor for the amount of opal in the sediment.

The limitation of iron is generally high in the SO, with a minimum value of 0.32. It has high values at the coast of Antarctica that climb down steeply further away from the land. In THEN-NOW, the limitation generally falls at the coast, with a minimum of -8%, and rises in the lower latitudes. This variability is the highest out of the factors named so far, and the minimum is located approximately between 40°E and 50°E. Out of the five limiting variables, iron seems to have the potential for explaining the observed drop in the opal flux. A closer look at the iron concentration in the Southern Ocean helps uncover what might be the reason for its decrease in the perturbed Earth system. The iron concentration is showed in figure 9.

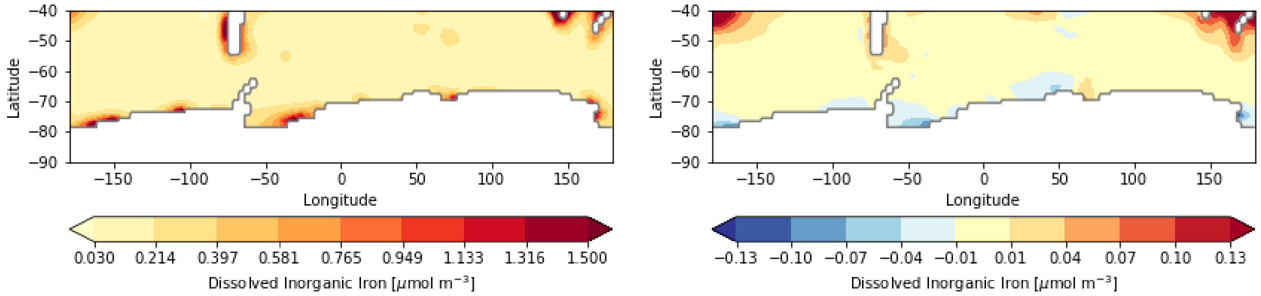


Figure 9: The concentration of iron at the surface in NOW (left) and THEN-NOW (right). In NOW, the maximum concentration is much larger than the rest of the output. The maximum is located between 50°W and 0°, and has a value of 4.6μmolm^{-3}.

4.3 Where Did Iron Go?

Clearly, the concentration of iron falls nearly everywhere at the coast of Antarctica, which is reflected in the diatom iron limitation (fig. 8). As discussed in the Silicate Cycle section, the iron supply in the SO is limited by dust transport. The dust can be brought to the surface of the ocean by the wind, or the dissolution from the sediment. Hence, the three possible sources of change that could have affected the iron concentration are:

1. Rise in ice sheet volume which covers the ocean and makes it impossible for the dust to be brought to the surface by the wind;

2. a) Drop in diffusion or b) change in advection which prevent the iron from spreading out of the sediment into the ocean.

The hypotheses can be tested by investigating the ice fraction, iron dust flux and water speed variables in the CESM output. The zonally averaged iron concentration is presented in figure 10 and is used to determine the depth layer most relevant to the investigation.

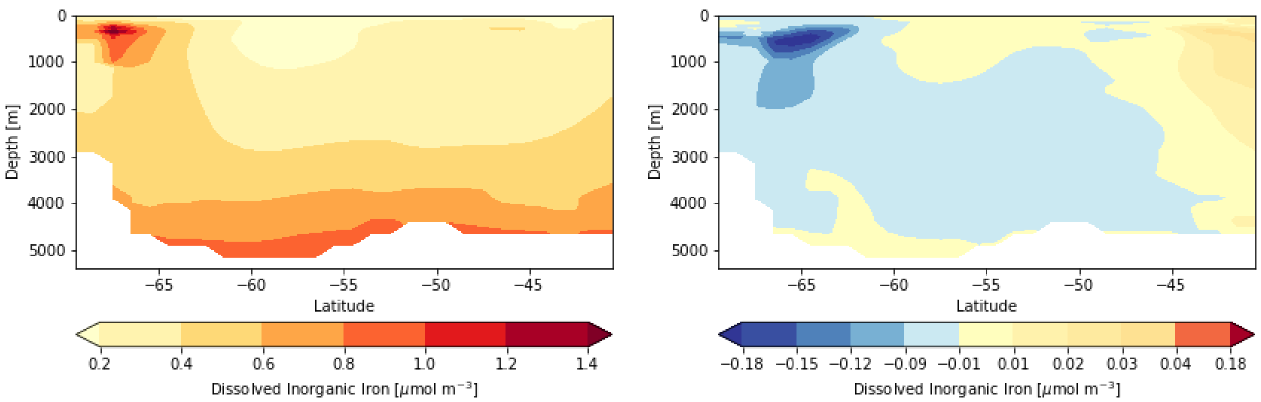


Figure 10: Zonally averaged (20°E to 70°E) iron concentration in NOW (left) and THEN-NOW (right).

Both in NOW and THEN-NOW, the majority of the concentration is located in the vicinity of 500m depth. In the simulation results, the depth of 527.72m is available for investigation. The concentration of iron in the water column is showed in figure 11. Unsurprisingly, figures 9 and 11 resemble each other to a large extent, where the concentration clusters have the same locations, but higher values of concentration as the depth increases. As discussed in the Silicate Cycle section, this is consistent with theoretical behavior of iron, which acts similar to a macronutrient in the sea water, with higher concentrations in the water column.

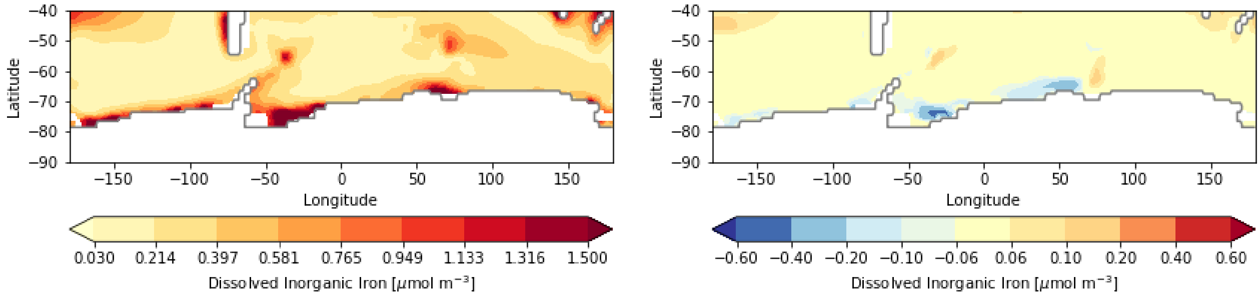


Figure 11: The concentration of iron in NOW (left) and THEN-NOW (right) at 527.72 m. The maximum concentration located between 50°W and 0° is much larger than the rest of the output, equal to $9.0\,\mu\mathrm{mol\,m^{-3}}$.

The ice cover fraction to sea water is showed in figure 12. The fraction of 0 represents no ice, and the fraction of 1 represents full ice cover. The maximum change in ice fraction is between -10% and -20%. There is only a small 1-4% change east of 50°E, which is in the vicinity of the opal flux drop. As it would be expected, the ice fraction comparison exhibits consistency with the changes in light limitation (fig. 7). The general drop of ice fraction suggests that it is not responsible for the fall in iron concentration in THEN-NOW. If ice were to blame, the ice fraction would have to increase, making it harder for the dust from the atmosphere to reach the surface of the Southern Ocean.

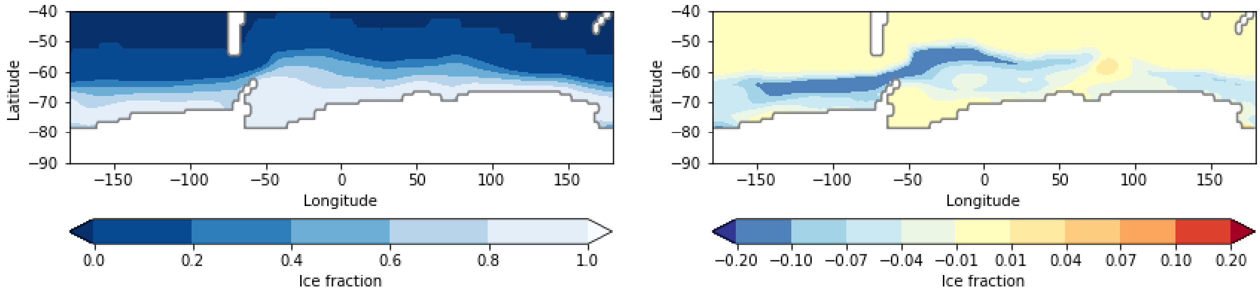


Figure 12: The ice fraction in NOW (left) and THEN-NOW (right).

The second possibility for the source of decreased iron concentration is that some part of the dissolution process was disturbed by the perturbation. The iron dust flux at 527.72 m is showed in NOW and THEN-NOW in figure 13, and the respective dust flux zonal averages in figure 14.

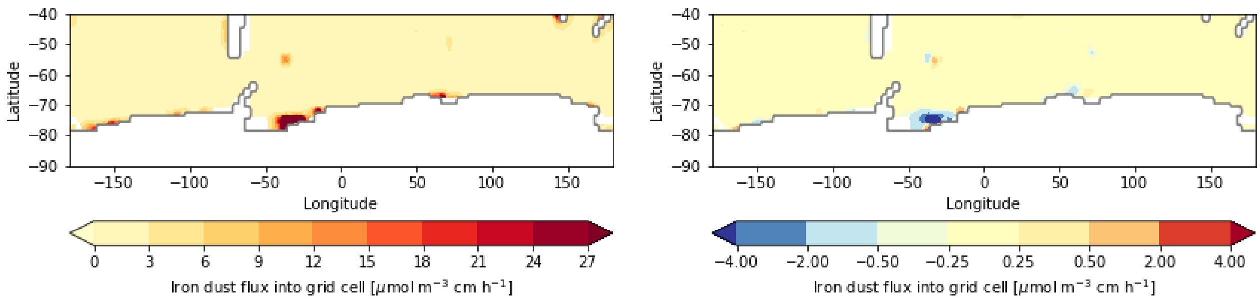


Figure 13: Iron dust flux into the sea water in NOW (left) and THEN-NOW (right). Just as in the case of the iron concentration, the maximum dust flux in NOW is much larger than the rest of the output. Located between 50°W and 0°, the flux has a maximum value of $175.1\,\mu\mathrm{mol\,m^{-3}\,cm\,h^{-1}}$. At the same location in THEN-NOW, the flux has a minimum value of $-17.6\,\mu\mathrm{mol\,m^{-3}\,cm\,h^{-1}}$.

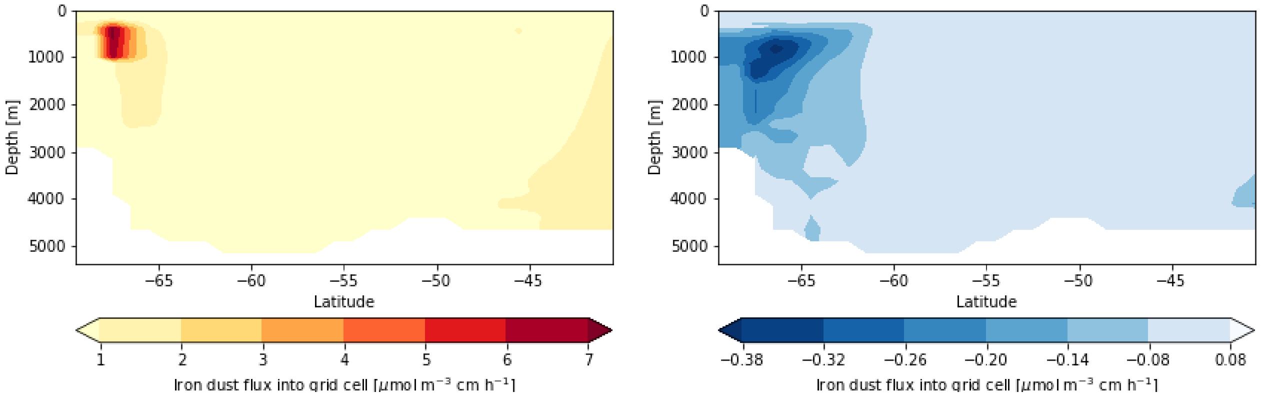


Figure 14: Zonally averaged (20°E to 70°E) iron dust flux into the sea water in NOW (left) and THEN-NOW (right).

As showed in equation (8), the tracer concentration can be affected by advection and diffusion. The diffusion is reflected by the dust flux into cell values right by continental margins. The advection is determined by the speed of the sea water in the interior. It should be noted that the further we move from the continental margin, the more the dust flux signal is affected by advection. Figure 14 shows general drop in the iron dust flux at the coast of Antarctica, but also further in the sea, up until 60°S. This suggests that the reduction of iron concentration in the ocean is not only a result of reduced diffusion, but also of a change in advection caused by a change in water speeds. The meridional speed at 527.72m is showed in figure 15 and zonal average of the speed in figure 16.

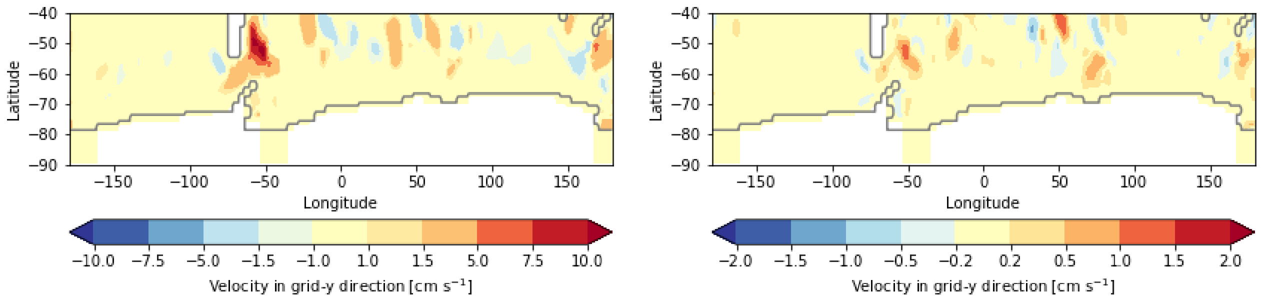


Figure 15: Meridional speed in NOW (left) and THEN-NOW (right).

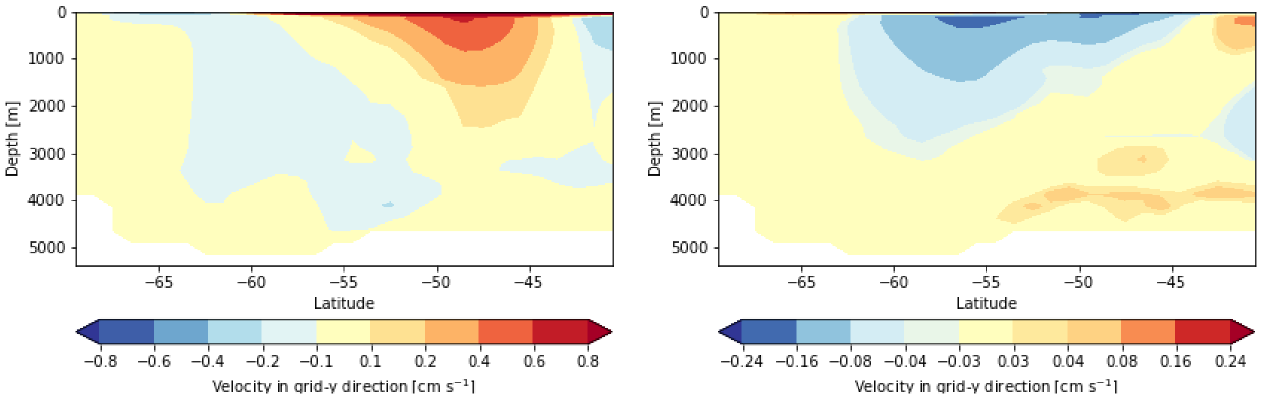


Figure 16: Zonally averaged (20°E to 70°E) meridional speed in NOW (left) and THEN-NOW (right).

The meridional speed shows local strengthening directed both southward and northward across the Southern Ocean. As the zonal mean between 20°E and 70°E in figure 16 demonstrates, the general trend for the meridional speed here is increased southward strength, reaching beyond the latitude of 60°S.

5 Discussion

5.1 Analysis of Results

The results reveal that within the scope of the model, the flux of biogenic silica to the sediment does not change in the open ocean. The changes observed are located at the continental coasts. The changes are largely positive, however there are instances of decreases in opal flux to the sediment. The figure 3 shows that these drops are located at longitudes where the wind stress substantially rises. These results refute the theory that an increase in wind stress will result in an increase of the flux of biogenic silica into the sediment.

The investigation of the limiting factors for diatom growth concludes that the flux of opal to the sediment is limited by iron. As figures 5 and 6 reveal, the diatoms' growth is not limited by silicate and nitrate in the Southern Ocean. The changes of these limitations in THEN-NOW only show increases of limitation values. The iron limitation shows the highes variability in THEN-NOW (of up to 8%), and a general reduction at the coast of Antarctica. The most substantial fall is between 20°E and 70°E, where the largest drop in the opal flux into the sediment is also observed (fig. 3).

So why would iron be a limiting factor for the diatom growth? Figure 9 sheds a light on the concentration of the iron in the Southern Ocean. Indeed, the concentration is decreased nearly everywhere at the coast of Antarctica. Considering the insight from the Silicate Cycle section, there are three hypotheses that could explain the iron concentration drop. Since the iron dust can enter the ocean through aeolian processes, or through continental margin sediment, one of the two processes is responsible for the reduced iron concentration. The aeolian processes are mostly influenced by the sea ice presence in the ocean, preventing the iron dust from entering at the surface. The ice fraction in THEN-NOW from figure 12 is presented in figure 17 with the iron concentration (fig. 9) contours overlaid. As mentioned in the Results section, ice fraction would have to rise in the areas where the iron concentration falls. However, the sea ice is showed to almost exclusively decrease across the coast of Antarctica. This suggests that the aeolian dust transport is not the limiting factor for the iron concentration in the Southern Ocean.

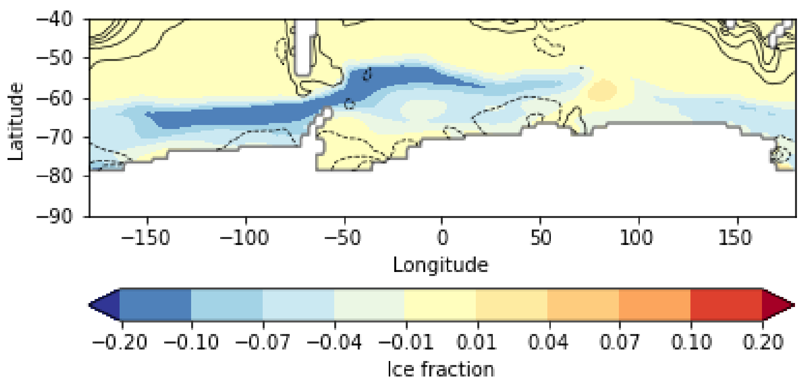


Figure 17: The ice fraction in THEN-NOW. The black contours outline the iron concentration in THEN-NOW at the surface (15m). Solid lines represent increase in concentration, and the dashed lines - decrease in concentration.

The continental margin dust flux is evidently closely correlated with the concentration of iron in the sea, which is not surprising - as discussed in the Silicate Cycle section, the iron concentration reflects local dust delivery pattern, and the aeolian dust delivery to the Southern Ocean is low. Hence, we are left with the continental margin sedimentary sources and this is also reflected in the abnormally high concentration and dust fluxes located between 50°W and 0°. The meridional velocity patterns in figure 15 show a varied response to the perturbation, with some meridional water speeds increasing northward, and some increasing southward. Figure 16, zonally averaged over the approximate location of the iron limitation for the diatoms, reveals a general increase in the northward velocity. Further insight can be gained by overlaying the iron concentration in the water column over the meridional water speeds in THEN-NOW, which can be seen in figure 18. It reveals that the dashed lines of decreased iron concentrations are accompanied by increased northward water speed nearly everywhere in the Southern Ocean. This suggests that the advection of the water at the coast has prevented the iron from spreading outwards into the open ocean, and thereby causing a decrease in its concentration in sea water.

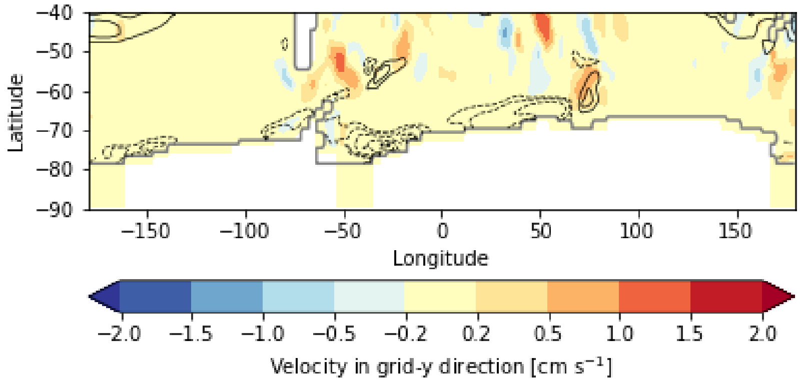


Figure 18: The meridional speed in THEN-NOW. The black contours outline the iron concentration in THEN-NOW in the water column (527.72m). Solid lines represent increase in concentration, and the dashed lines - decrease in concentration.

Clearly, the dust transport inside the water column has affected the iron concentration to a greater extent than the aeolian dust transport. Can it be determined which factor is more to blame for the reduction of the iron concentration and thereby the decrease in iron limitation: diffusion or advection? The zonally averaged dust flux (fig. 14) and meridional speed (fig. 16) can give an insight into this problem. The dust flux right by the continental margin falls from $3.5\,\mu\mathrm{mol\,m^{-3}\,cm\,h^{-1}}$ to $3.12\,\mu\mathrm{mol\,m^{-3}\,cm\,h^{-1}}$, or decreases by 10.9%. The maximum decrease of the meridional speed is 30%, however it is located close to the surface and far from the coast, where the spreading of the iron can be affected. More relevantly, the northward speed of the water at the latitude of 60°S increases from $10\mathrm{ms^{-1}}$ to $14\mathrm{ms^{-1}}$, or 40%. In comparison, the meridional water speed seems to be the key factor which affects the iron concentration. However, the proximity of the dissolution of the sediment indicates that the dust flux from the continental margin is also relevant to the concentration and can cause a substantial change.

The study recently published by *Moore et al.*[3] outlines a phenomenon similar to the one observed here. In the study, the CESM simulations are ran to predict the Earth climate in the future. Accompanied by the strengthening of the westerly winds and reduction in sea ice, the scientists observe nutrient trapping in the Southern Ocean. The trapping of iron is caused by the southward shift in the upwelling zone. Although the concentration of iron generally decreases at the coast of Antarctica, while the study predicts a pattern of an increase of iron in high latitudes and a decrease in low latitudes, the effect of trapping can explain the strongly reduced iron dust flux (fig. 14) and the effect of the meridional speed on the iron concentration.

5.2 Model Limitations

The disagreement of the biogenic silica flux to sediment values predicted by the model and found by *Anderson et al.* unsurprisingly shows that the model does not perfectly reflect reality. The opal flux to the sediment in NOW is predicted to be 0.00-0.11 $\mathrm{g\,cm^{-2}\,kyr^{-1}}$, which is only in agreement with the core TN057-13-4PC. The core E27-23 located 155.24°E is in the vicinity of flux value 0.11-0.21 $\mathrm{g\,cm^{-2}}$ $\mathrm{kyr^{-1}}$, which is quarter to half the value found in observations. Moreover, the model does not predict a change in silicate flux at any core location, which is in direct disagreement with the found observations.

The observations for the silicate concentration in the Southern Ocean, together with the comparison NOW-observations are showed in figure 19. The general trend appears to be the underestimation of the silicate concentration by the model at the coast of Antarctica, and overestimation in the open ocean. This results in a high error of >100% in low latitudes at the surface, where the silicate concentration is low, and relatively low error of 30% at the coast in the water column, which is abundant in silicate.

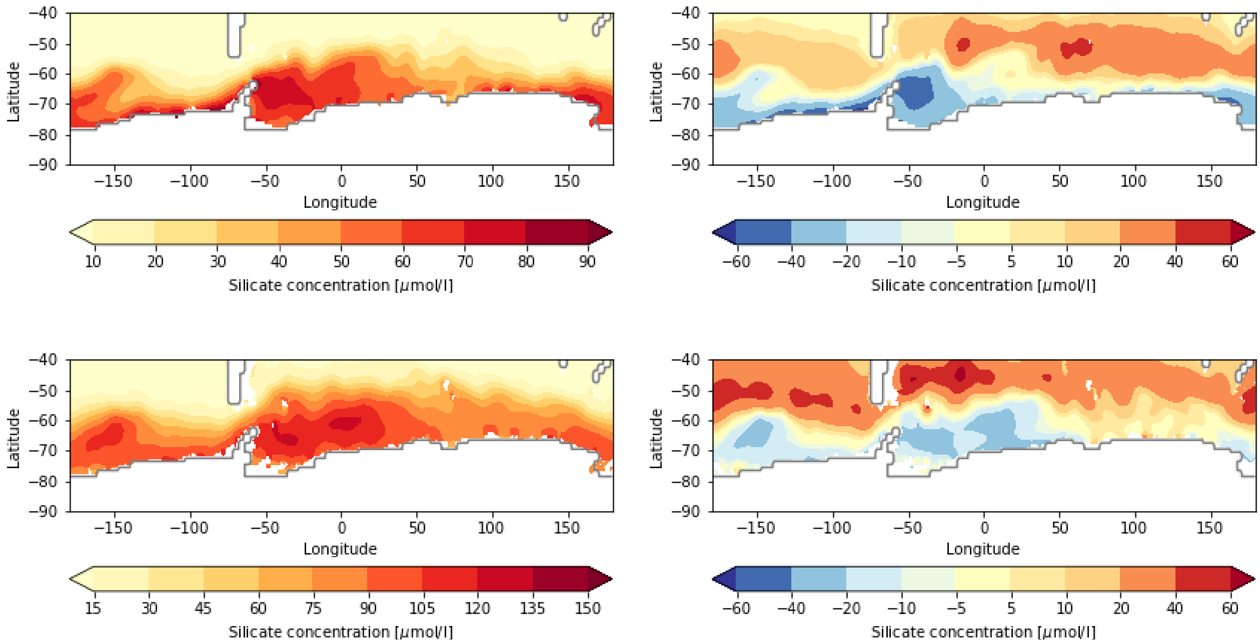


Figure 19: Observations (left) and the difference NOW-observations (right) of the concentration of silicate in sea water at the depth of 15 m (top row) and 500 m (bottom row).

The trend appears to be zonally constant, which allows for a look into the depths of the world's ocean. Figure 20 shows the zonally averaged observations and the difference NOW-observations of the silicate concentration in sea water. It should be noted here that the vertical resolution of the WOA13 data and NOW output differ from each other, and the difference increases with depth. Specific depth values from both were chosen that were either equal or approximately the same. Hence, there is more comparison values for the near-surface depths, and less for the deep ocean (>1000 m), and the deep ocean values also include vertical error. Moreover, the amount of observed data decreases substantially with depth. This is likely the reason why the difference NOW-observations increases significantly at the bottom.

The pattern of underestimation at the coast of Antarctica and overestimation in the open Southern Ocean by the model continues until about 3500 m. The error again varies from high >100% in low silicate concentration areas to low 3% in high concentration regions.

As mentioned in *Nielsen et al.*, the applied perturbation of insolation is abrupt compared to realistic time scales. If the process were to occur slower, the results might be different. Moreover, the model uses a fixed climatology of dust and iron fluxes. Therefore, the influence of the iron dust supply change due to wind stress changes that result from the AMOC collapse cannot be observed within the scope of the model.

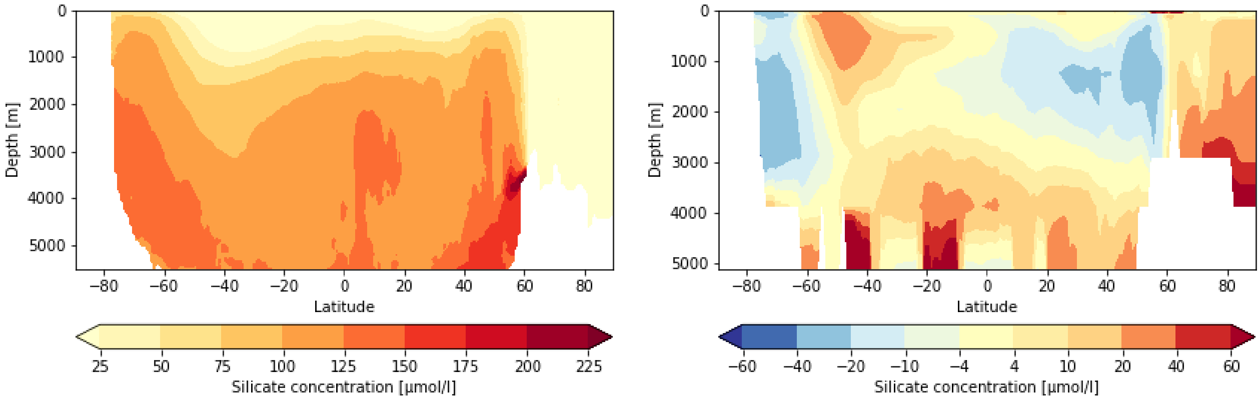


Figure 20: Zonally averaged observations (left) and the difference NOW-observations (right) of the silicate concentration in sea water.

Nevertheless, the realistic output values from the various model variables and the complexity of the model itself suggest that the results can illuminate the areas of ocean research that might need revising.

5.3 Further Research

This work only scrapes the surface of the complexity that is the silicate cycle in the ocean. The scientists have yet to find out many of the aspects of the iron cycle, and how exactly it influences the diatoms, in particular the Si/N and the Si/C ratios.[7] The complexity of the research is limited by the complexity of the model, which only adds to the difficulty of tackling the problem of the key factors influencing the opal flux to the sediment. As it stands now, the possibilities of further work in this discipline are endless.

The first step could be to fully understand the model calculations which determine the values for diatom growth limitations, and approach the problem with this perspective. The insight could put to light the exact complexity limitations of the model. The studies of diatoms show not only variability between the effects of limitations on different diatom species, but also a possibility for co-limitations.[2] This complexity cannot be reflected by the model, and the assessment of the information loss caused by this could more precisely tell how well our model can reflect reality.

The limitation mentioned above based on the time scale of the insolation forcing can be avoided in future research. Soon, there will be models available to Team Ocean which are fully coupled with mixing and biogeochemistry, and which force the perturbation on a more realistic time scale. Repeating the investigation steps taken in this work have potential to show results of a setup resembling the real Earth system even closer.

6 Conclusion

Within the scope of the model, opal is not a good upwelling proxy. The opal in the Southern Ocean sediment is regulated by the concentration and distribution of iron, not silicate. The flux of biogenic silica to the sediment generally increases across the coast of Antarctica in the deglacial state, however there is a substantial decrease between 20°E and 70°E, which is caused by reduced iron concentration. The iron in the Southern Ocean is regulated by the dust flux from the continental margin sediment. The analysis of the processes near the sediment reveal that the change of advection of sea water is the main influence which results in reduced iron concentration at the coast of Antarctica. The increased southward water speeds might be caused by southward upwelling zone shift connected to the planet warming, which was observed in other model studies.

7 Acknowledgements

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