Iron fertilization in the ocean and consequences for the global carbon cycle

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Abstract

It has been suggested that fertilizing the ocean with iron can stop the continuing increase of atmospheric carbon dioxide by enhancing the biological uptake of carbon. This would decrease the surface ocean partial pressure of carbon dioxide, thus forcing the absorption of carbon dioxide from the atmosphere. Using a five-box model of the ocean circulation, we study the response of the ocean due to iron fertilization and its consequences for the carbon dioxide atmospheric partial pressure. We simulate the fertilization in the North Atlantic, the low latitudes and the Southern Ocean, assuming a strong productivity in each area of experimentation. We find that the Southern Ocean is the most effective area for reducing atmospheric carbon dioxide, with a drop of 28 ppm in carbon dioxide atmospheric partial pressure after 150 years of continuous fertilization.
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1. Introduction

In the aftermath of the industrial boom of the late nineteenth century, human activities such as fossil fuel burning, cement production and tropical deforestation have led to an increase in the emission of carbon dioxide into the atmosphere. An increase from 280 ± 10 ppm before the industrial revolution to 373 ppm in 2002 has been observed[1], meaning a spectacular jump of 93 ± 10 ppm in less than 200 years. This jump is similar to that of the transition from the glacial to the interglacial period, with the difference that the glacial-interglacial changes occurred over hundred of thousands of years [1].

The growth of carbon dioxide in the atmosphere leads to an increase of the earth’s temperature, due to its nature as a greenhouse gas. It acts by trapping the reflected radiation from the earth and thus making the earth warmer. A direct consequence is the observed rising of the sea level at an average rate of 1 to 2 mm/year over the past 100 years resulting from the melting of icebergs at high latitudes. This rise is significantly larger than the rate averaged over the last several thousand years and causes several climatic problems like floods and hurricanes which bring disaster to the population. The increase of the earth’s temperature and its consequences is called global warming and constitutes a challenge for the scientific community of this century. It is necessary to find a solution to this problem.

In the late 1980s, the oceanographer John Martin [2] showed the role of iron in boosting the photosynthesis process at the ocean surface. The micro-organisms responsible are phytoplankton. They fix the dissolved carbon to form organic matter, thus forcing more carbon dioxide from the atmosphere to be trapped in the ocean. Martin suggested the possibility of sequestering enormous quantities of carbon dioxide from the atmosphere by seeding vast areas of open ocean with iron and thus decreasing the quantity of carbon dioxide in the atmosphere.

This revolutionary idea has gripped the attention of the scientific community as a possible solution against global warming. Several experiments have been carried out, during which a bloom of phytoplankton was observed, thereby confirming the effect of iron in boosting photosynthesis. However, due to the short time scale of those experiments (a few weeks), it was difficult to conclude anything about the drop of atmospheric carbon dioxide and its effect in the ocean after several years of fertilization. Considering the time scale of the ocean circulation (the overturning circulation of the ocean has a time scale of order 1000 years [3]), the best way to test this idea is perhaps by modelling.

Using a five-box model of the ocean circulation, we will simulate the fertilization experiment in three areas of the sea surface: the North Atlantic, the low latitudes and the Southern Ocean. We will try to find the response of the ocean and of the atmospheric carbon dioxide after a few hundred years of simulation. We compare the results in each fertilization zone to determine the most effective zone for sequestering CO₂.
2. Iron fertilization in the ocean

2.1 The ocean as the major $CO_2$ container

About 70% of the world’s surface is covered by oceans. Carbon dioxide is more present in the ocean than in the atmosphere. Indeed, if we look at the pre-industrial ocean-atmosphere carbon, 98.5% of the total $CO_2$ is found in the ocean and only 1.5% in the atmosphere. By contrast, we find only 0.6% of world’s oxygen in the ocean [4]. The reasons for the large amount of $CO_2$ in the oceans are its high solubility, (which is thirty times that of oxygen) [1], the chemical reactions it undergoes with water to give carbonate and bicarbonate ions, and biochemical reactions with micro-organisms in the sea. $CO_2$ is a greenhouse gas which acts by trapping the reflected radiation from the Earth. The more $CO_2$ trapped in the ocean, the colder the climate.

Observations show that the concentration of DIC (dissolved inorganic carbon) increases from the surface to deep ocean. The DIC is the sum of the concentrations of $CO_2$, bicarbonate ($HCO_3^-$) and carbonate ($CO_3^{2-}$) ions:

$$DIC = [CO_2] + [HCO_3^-] + [CO_3^{2-}]$$  \hspace{1cm} (2.1)

Carbon dioxide in the ocean undergoes the following reactions:

$$CO_2 + H_2O \iff H_2CO_3 \iff H^+ + HCO_3^- \iff 2H^+ + CO_3^{2-},$$  \hspace{1cm} (2.2)

which gives 88% of $HCO_3^-$, 10.9% of $CO_3^{2-}$ and only 0.5% of $CO_2$ for the total DIC [1].

We have around 1970 umol$^1$/kg of DIC at the surface and 2280 umol/kg in deep ocean [1], which is approximately 12% lower at surface than at depth [4] [5]. This surface to deep difference is due to the biological and the solubility pumps [5], which we shall now discuss.

2.1.1 The solubility pump

The net transport of DIC from the surface to the deep ocean due to temperature and salinity is called the solubility pump. This mechanism is driven by two factors:

- The thermohaline circulation due to the formation of deep water at high latitudes where the water is cooler and denser with high salinity. The thermohaline circulation is an overturning circulation in which the warm water flows polewards and is subsequently converted into cool water that sinks and flows towards the equator in the interior (see figure 2.1). In the North Atlantic, we have the North Atlantic Deep Water (NADW) which penetrates southward at depths between 1500m and 2500m [1]. In the Southern Ocean we have the Antarctic Bottom Water (AABW) which flows northwards into the Indian, Pacific and Atlantic basins upwelling at 3500 m [6]. Radiocarbon measurements show that the thermohaline circulation turns over all the deep water in the ocean every 600 years [6] or 1000 years [3].

\hspace{1cm}$^1$One umol=$10^{-6}$ mol
Section 2.1. The ocean as the major $CO_2$ container

The solubility of carbon dioxide is a strong inverse function of sea water temperature. This means that carbon dioxide is more soluble at high latitudes which are cooler than at low latitudes. Consequently, at high latitudes the formation of deep water together with the higher solubility of carbon dioxide results in the transportation of carbon dioxide from the surface to the deep ocean.

2.1.2 The biological pump

The biological pump is a suite of biological processes that transport carbon from the surface zone to the ocean’s interior. It includes the carbonate and the soft tissue pump:

- The carbonate pump is related to the formation of calcium carbonate $CaCO_3$ from calcium ($Ca^{2+}$) and carbonate ions (see equation 3.4) at the sunlit surface. Calcium carbonate forms the shells of micro-organisms living at the sea surface such as coccolithophores, foraminifera or pteropods.

$$Ca^{2+} + CO_3^{2-} \leftrightarrow CaCO_3 \quad (2.3)$$

Calcium carbonate dissolves at a rate dependent upon local carbonate chemistry. As this process is slower than the synthesis process, and because the particulate material is sinking,
Section 2.1. The ocean as the major $CO_2$ container

The carbonate pump transports material from the surface of the ocean to its depths.

- The soft tissue pump transports the dissolved organic carbon formed by photosynthesis by physical processes such as downwelling or sinking. Remineralization converts organic matter back into dissolved inorganic matter deep in the ocean. This leads to a net transport of DIC from the surface to the depth (see figure 2.2).

![Figure 2.2: The soft tissue pump](image)

Figure 2.2: The soft tissue pump: part of the organic matter produce by photosynthesis is remineralized at the surface, where $CO_2$ returns back to the atmosphere, and another part sinks into the ocean interior.

Photosynthesis is a process by which plants, phytoplankton and some bacteria use the energy from sunlight to produce glucose, with the release of oxygen. It is arguably the most important biochemical pathway known; nearly all life depends on it. The conversion of unusable sunlight energy into usable chemical energy is associated with the action of the green pigment called
Section 2.2. The role of iron

chlorophyll. The equation of photosynthesis is:

$$6CO_2 + 6H_2O \iff C_6H_{12}O_6 + 6O_2,$$

with $C_6H_{12}O_6$ representing the organic matter.

The organisms responsible for photosynthesis at the ocean surface are phytoplankton. In addition to $CO_2$ and water, they need nutrients such as phosphorus, nitrogen and iron to produce organic matter. The reaction is presented as follows [7]:

$$\sqrt{Fe} \ 106CO_2 + 16HNO_3 + H_3PO_4 + 78H_2O \iff C_{106}H_{75}O_{42}N_{16}P + 150O_2,$$

with nitrogen in the form of $HNO_3$ and phosphorus in the form of $H_3PO_4$. The organic matter of composition $C_{106}H_{75}O_{42}N_{16}P$ is built with a stoichiometric ratio of C, N, P and $O_2$ as: $C:N:P:O_2 = 106:16:1:150[7]$. The stoichiometric ratio between a suite of elements is the ratio of these elements in the organic matter. Recent research has expanded this ratio as: $C:N:P:O_2:Fe = 106:16:1:150:0.001$. This means that each unit of iron can fix 106,000 units of carbon, 16,000 units of nitrate and 1,000 units of phosphate.

The quantity of iron needed is very small compared to the quantity of phosphate or nitrate; but without iron, phytoplankton will not survive.

### 2.2 The role of iron

An observation of the annual phosphate as well as the annual nitrate (see figures 2.3 and 2.4) levels at the ocean surface shows that these nutrients are not uniformly distributed. We have areas where concentrations of nitrate and phosphate are close to zero and others with high concentration. The areas with the highest concentrations are the Southern Ocean and the Subarctic Pacific with 1.5 to 2.0 umol/kg, followed by the Arctic zone with 1.0 umol/kg. Normally, we expect those regions to be biologically very active and to have the highest population of plankton. Curiously, some of them have a very weak population of phytoplankton. This can be seen by observing figures 2.3, 2.4 and 2.5 which shows that the Southern Ocean, the Subarctic Pacific and the Equatorial Pacific, all regions of high nutrients concentrations, are very poor in chlorophyll (pigmentation due to photosynthesis carried by phytoplankton). Those regions are described as High Nutrients Low Chlorophyll (HNLC). The reason for this is the limited quantity of iron in these regions. This is due to the large distance between them and the large deserts (Kalahari, Sahara and Arabian deserts) which consequently cannot supply enough iron as they do for the rest of the ocean [10]. The North Atlantic is at the same latitude as the Subarctic Pacific but is supplied with quantities of iron dust from the Sahara desert in Africa. Consequently the North Atlantic is not an HNLC region. Iron comes into the ocean in the form of dust transported by wind and plays a vital role as a micronutrient for photosynthesis and the growth of phytoplankton.
Section 2.2. The role of iron

Figure 2.3: Observed annual mean phosphate concentration at the ocean surface in umol/kg [8]

Figure 2.4: Observed annual mean nitrate concentration at the ocean surface in umol/kg. This image clearly shows the high levels of nitrate in the subarctic Pacific, the equatorial Pacific and the Southern Ocean [8].
Section 2.3. The concept of iron fertilization

The previous observation led to the idea that one could improve the growth of phytoplankton by fertilizing the surface ocean with iron, and as a consequence, extract more CO$_2$ from the atmosphere.

2.3.1 A bit of history

The oceanographer John Martin was the first to propose the idea of iron fertilization in order to manage global warming when he declared in the late 1990s: “Give me half a tanker of iron and I’ll give you another ice age” [11] His studies indicated that it was indeed a scarcity of iron micronutrients that was limiting phytoplankton growth in some area of the ocean like the Southern Ocean. He also gave an explanation for the glacial-interglacial climate change as being the consequence of a bloom of phytoplankton due to higher concentration of iron dust in the ocean. Martin hypothesized that restoring high levels of plankton photosynthesis could slow or even reverse global warming by sequestering enormous volumes of CO$_2$ in the Ocean [11].

A few months after the eruption of Mount Pinatubo in 1991 in the Philippines, the environmental scientist Andrew Watson analysed global data from the eruption and calculated that it deposited approximately $4 \times 10^{10}$g of iron dust into the Southern Ocean [12]. The minerals were washed...
into the oceans, where the iron fertilized the plankton which enabled them to fix and metabolize more \( CO_2 \). This fertilization event generated an observed global increase of \( O_2 \) and decline of \( CO_2 \) in the atmosphere [12], perhaps showing the evidence of Martin’s hypothesis.

### 2.3.2 What is iron fertilization?

As we said before, iron fertilization consists of adding iron to the upper ocean to enhance the growth of phytoplankton and, as a consequence, boost photosynthesis to fix more \( CO_2 \) and slow down global warming. This \( CO_2 \) contributes to the formation of organic matter which is expected to sink to the deep ocean. In fact, just approximately 30% of this carbon-rich biomass sinks below 200 meters into the colder water strata below the thermocline [4]. Part of this fixed carbon continues falling into the abyss and the rest is dissolved and remineralized. However at this depth, the carbon is suspended in the deep ocean and effectively isolated from the atmosphere for centuries. Most of this sinking happens at high latitudes because that is where the deep water is formed.

NASA and NOAA estimate a decline of the phytoplankton population in the last 25 years of at least 6%. Simply returning this population to its previous levels of health and activity could therefore annually sequester 2 to 3 billion more tons of \( CO_2 \) than are being removed today.

Iron fertilization targets are areas that have persistently high levels of major nutrients, such as nitrate and phosphate, but are also weakly photosynthetic (HNLC) regions. An important example is the Southern Ocean, which has the largest repository of unused macronutrients in surface water, and plays a key role in the formation of intermediate and deep water [13]. These areas might represent a non-negligible ratio of about 16% of the total surface ocean [14].

### 2.4 Some iron fertilization experiments

Some experiments have been performed to test the efficiency of iron fertilization. These experiments consisted of adding quantities of iron nutrients in the form of iron sulfate to some of the HNLC areas. For example,

- the SEEDS field experiments in the Western and Eastern subarctic North Pacific [15],
- the SERIES experiment in the Northeast Subarctic Pacific [16],
- the SOIREE experiment in the Polar Southern Ocean [13],
- the experiments in the Equatorial Pacific Ocean [17].

The majority of these experiments have resulted in the study of only the evolution phase of the bloom except in the Northeast Subarctic Pacific where an increase in POC (particulate organic carbon) flux at 50 m depth has been observed [16]. All of these showed arguably the importance of iron in boosting phytoplankton growth and the consequent fixation of more nutrient and \( CO_2 \).
in the biomass at the ocean surface. However, little is known about the net effect on the air-sea carbon dioxide flux and atmospheric $pCO_2$.

The scientific community is divided about the efficiency of iron fertilization. The entrepreneur Michaels [11] claims that he can cancel out the world contribution to atmospheric $CO_2$ increases from burning fossil fuel by fertilizing 16 million square miles of the Southern Ocean with 8.1 million tons of iron [11]. Supporters of this idea also think that the growth of phytoplankton population will have a positive impact on the food chain by providing more food for other species.

On the other hand, some researchers think that this is not the best idea. As the population of phytoplankton grow due to iron nutrient, other species like zooplankton grows also by eating the phytoplankton. Therefore, the biomass remains at the surface from where all the fixed $CO_2$ returns back to the atmosphere following decomposition by bacteria. At present it is difficult to predict the consequences in the ocean ecology due to iron fertilization.

The previous experiments were performed in a relatively small area compared to the total surface of the ocean, typically ten by ten kilometers and usually lasting only few weeks. This makes prediction of the ocean response based on these experiments difficult.

In order to learn more, we will next model the carbon cycle in the context of simple box models with circulation. Our models are similar to the three-box model of Sarmiento and Toggweiler [3], or that of Siegenthaler and Wenk [5].
3. Modeling the ocean circulation

Phenomena such as the thermohaline and thermocline circulations, the Antarctic Bottom Water as well as the North Atlantic Deep Water and the surface ocean movements driven by winds such as the Gulf Stream, make the understanding and prediction of ocean features difficult. One attempt to study this is perhaps by modeling, providing that the model takes into consideration the most important characteristics of the ocean, depending on the research problem.

Restricting ourselves to iron fertilization experiments, a good model should take into consideration the difference in concentration of DIC, $PO_3$ and $NO_4$ from the surface to the deep ocean. This can be done by separating the ocean into boxes, each box representing a particular part of the ocean. The simplest such model is a two-box model.

3.1 Two-box model of the ocean carbon cycle

In a two-box model, the ocean is divided into two parts: the surface and the deep ocean. This division is controlled by the difference in concentration of elements such as DIC, $PO_3$ and $NO_4$, which play a key role in the global carbon cycle. Observations show that the concentration of those elements increases from the surface to the deep ocean and is driven by two factors known as the biological and solubility pumps (see section 2.1).

The model takes into account (figure 3.1) the net biological uptake from the surface to the deep ocean and the mixing terms between the reservoirs.

Taking $V_s$ and $V_d$ to be respectively the volume of the surface and deep ocean in $m^3$, $DIC_s$, $O_{2s}$ and $PO_{4s}$ respectively the concentration of dissolved inorganic carbon, oxygen and phosphate at the surface box in umol/$m^3$ and finally $DIC_d$, $O_{2s}$ and $PO_{4s}$ respectively the dissolved inorganic carbon, oxygen and phosphate at the deep box, we have:

1. In the surface box

The equations of time rate of change of DIC, $PO_4$, $O_2$ are:

\[
\begin{align*}
V_s \frac{dDIC_s}{dt} &= v(DIC_d - DIC_s) - F_{DIC} + \phi_{CO_2}, \\
V_s \frac{dO_{2s}}{dt} &= v(O_{2d} - O_{2s}) - F_{O_2} + \phi_{O_2}, \\
V_s \frac{dPO_{4s}}{dt} &= v(PO_{4d} - PO_{4s}) - F_{PO_4},
\end{align*}
\]

with $F_{DIC}$ and $F_{PO_4}$ the biological sinks due to photosynthesis, $\phi_{CO_2}$ and $\phi_{O_2}$ respectively the net flux of carbon dioxide and oxygen into the ocean due to gas exchange with the atmosphere:

\[
\phi_{CO_2} = K\beta_{CO_2}(pCO_{2atm} - pCO_{2ocean}),
\]
Section 3.1. Two-box model of the ocean carbon cycle

Figure 3.1: Two-box model of the Ocean. \( F \) represents the net biological uptake from the surface to the deep box. \( K \) and \( v \) in \( m^3/s \) respectively represent the mixing term or vertical exchange rate between the surface box and the atmosphere and between the surface box and the deep box.

\[
\phi_{O_2} = K \beta_{O_2} (pO_{2atm} - pO_{2ocean}).
\] (3.3)

\( pCO_{2atm}, pO_{2atm}, pCO_{2ocean} \) and \( pO_{2ocean} \) represent the partial pressure of carbon dioxide and oxygen in the atmosphere and in the ocean. \( \beta_{CO_2} \) and \( \beta_{O_2} \) are respectively the solubility of carbon dioxide and oxygen in \( \text{mol/m}^3 \cdot \text{atm} \).

2. In the deep box

The equations of time rate of change of DIC, \( PO_4 \), \( O_2 \) are:

\[
\begin{align*}
V_d \frac{dDIC_d}{dt} &= v(DIC_s - DIC_d) + F_{DIC}, \\
V_d \frac{dPO_4_d}{dt} &= v(PO_4_s - PO_4_d) + F_{PO_4}, \\
V_d \frac{dO_2_d}{dt} &= v(O_2_s - PO_2_d) + F_{O_2}.
\end{align*}
\] (3.4)
At steady state, the system of equations 3.4 becomes the system of equations 3.5:

\[
\begin{align*}
\upsilon(DIC_d - DIC_s) &= F_{DIC}, \\
\upsilon(PO_{4d} - PO_{4s}) &= F_{PO_4}, \\
\upsilon(O_{2d} - PO_{2s}) &= F_{O_2}.
\end{align*}
\]  

(3.5)

DIC, $PO_4$ and $O_2$ are in stoichiometric ratio of C:P:O$_2 = 117:1:-170$ [3] such that:

\[
O_{2d} - O_{2s} = rP : O_2 \times (PO_{4d} - PO_{4s}),
\]

(3.6)

with $rP:O_2$ the stoichiometric ratio between phosphate and oxygen. From equation 3.6 we can find the concentration of oxygen in the deep ocean.

Taking $O_{2s} = 234$ mmol/m$^3$, $PO_{4d} = 2.1$ mmol/m$^3$ and using $P:O_2 = -170$ [1], we obtain:

\[
O_{2d} = 234 - 170 \times (2.1 - 0)
\]

\[
= -123 \text{mmol/m}^3.
\]

(3.7)

This result (equation 3.7) is clearly wrong because it is impossible to have a negative concentration of oxygen in the ocean. The conclusion is that our model is incorrect. Why?

We assumed that the concentration of nutrients is uniform at the ocean surface. This is incorrect, because at high latitudes we find much more unutilized nutrients, known as preformed nutrients [1] than at low latitudes. From figures 2.2 and 2.3 we observe that the concentration of phosphate and nitrate at high latitudes ranges from 1.5 to 2.0 umol/l, whereas in low-latitude oceans it is close to zero. This difference is related to chemical and physical processes in which both regions are involved. Additionally, the North Atlantic Deep Water (NADW) and the Antarctic Bottom Water (AADW) currents driven by the thermohaline circulation (section 2.2), result in the surface to deep differences in oxygen, nutrients and DIC.

We then need to modify our model to consider the thermohaline circulation and the difference in nutrients between the low and high-latitude surface oceans. We divide the surface ocean into two regions (the low-latitudes and the high-latitudes), thus creating a three-box model.

### 3.2 Three-box model of the ocean carbon cycle

This model was described by Sarmiento and Toggweiler in 1984 [3] to show the importance of the thermohaline circulation and the high-latitude surface ocean in the decrease of carbon dioxide atmospheric partial pressure ($pCO_2$) during the last ice age. From this model, Sarmiento and Toggweiler suggested that the decrease of $pCO_2$ during the last ice age was the response to an increase in the net high-latitude productivity coupled with a possible decrease of the thermohaline overturning.

The model (figure 3.2) is made up of:
Section 3.2. Three-box model of the ocean carbon cycle

1. a box representing the low-latitude surface ocean of thickness 100 m [3],

2. a box representing the high-latitude surface ocean, denser and cooler of thickness 250 m. It represents 16% of the whole surface ocean,

3. a box representing the deep ocean to which all materials from the surface sink, and get remineralized after a certain period of time.

![Diagram of the three-box model of the ocean carbon cycle]

Figure 3.2: Three-box model of the ocean: $P_l$ and $P_h$ represent the sinking particle and dissolved inorganic matter flux from the low latitudes and the high latitudes in mol/s. $K_l$ and $K_h$ are respectively the gas exchange coefficient between the low-latitude surface and the atmosphere and between the high-latitude surface and the atmosphere. $T$ is the thermohaline circulation and $f$ is the vertical exchange rate between the high-latitude and the deep ocean in m$^3$/s.

We consider that most of the interactions between the surface and the deep ocean in terms of solubility happen at the high latitudes.

Taking $V_l$, $V_h$ and $V_d$ to be respectively the volume of the low-latitude surface, high-latitude surface and deep ocean in m$^3$, $\phi_{CO_{2l}}$, $\phi_{O_{2l}}$, $\phi_{CO_{2h}}$, $\phi_{O_{2h}}$ are the net flux of carbon dioxide and oxygen into the ocean, due to gas exchange with the atmosphere at low and high latitudes, the equations of time rate of change of DIC, $PO_4$, $O_2$ are:

1. **In the first box:**
V_l \frac{dDIC}{dt} = -P_lDIC + T(DIC_d - DIC_l) + \phi_{CO2}, \tag{3.9}
\frac{dO_2l}{dt} = -P_lO_2 + T(O_{2d} - O_{2l}) + \phi_{O_2}, \\
V_l \frac{dPO_{4l}}{dt} = -P_lPO_{4} + T(PO_{4d} - PO_{4l}).

2. In the second box:
\[
\begin{align*}
V_h \frac{dDIC_h}{dt} &= f(DIC_d - DIC_h) + T(DIC_l - DIC_h) \\
&\quad - P_hDIC + \phi_{CO2h}, \\
V_h \frac{dO_{2h}}{dt} &= f(O_{2d} - O_{2h}) + T(O_{2d} - O_{2l}) \\
&\quad - P_hO_2 + \phi_{O_2h}, \\
V_h \frac{dPO_{4h}}{dt} &= f(PO_{4d} - PO_{4h}) + T(PO_{4d} - PO_{4l}) - P_hPO_{4}. \tag{3.10}
\end{align*}
\]

3. In the third box
\[
\begin{align*}
V_d \frac{dDIC_d}{dt} &= (T + f)(DIC_h - DIC_d) + P_hDIC + P_lDIC, \\
V_d \frac{dO_{2d}}{dt} &= (T + f)(O_{2h} - O_{2d}) + P_hO_2 + P_lO_2, \tag{3.11} \\
V_d \frac{dPO_{4d}}{dt} &= (T + f)(PO_{4h} - PO_{4d}) + P_hPO_{4} + P_lPO_{4}.
\end{align*}
\]

The net fluxes of carbon dioxide and oxygen in the ocean at the low latitudes are given by:
\[
\phi_{CO2l} = K_l \beta_{CO2}(pCO_{2atm} - pCO_{2oceanl}), \tag{3.12}
\]
\[
\phi_{O_2l} = K_l \beta_{O_2}(pO_{atm} - pO_{2oceanl}). \tag{3.13}
\]

The net flux at the high latitudes is:
\[
\phi_{CO2h} = K_h \beta_{CO2}(pCO_{2atm} - pCO_{2oceanh}), \tag{3.14}
\]
\[
\phi_{O_2h} = K_h \beta_{O_2}(pO_{atm} - pO_{2oceanh}). \tag{3.15}
\]

At steady state, the system of equation 3.11 becomes:
\[
\begin{align*}
(T + f)(DIC_d - DIC_h) &= P_hDIC + P_lDIC, \\
(T + f)(O_{2d} - O_{2h}) &= P_hO_2 + P_lO_2, \tag{3.16} \\
(T + f)(PO_{4d} - PO_{4h}) &= P_hPO_{4} + P_lPO_{4}.
\end{align*}
\]
Now let us evaluate the concentration of oxygen in the deep ocean. From equation 3.16 we can write:

\[ O_{2d} = \frac{P_{hO_2} + P_{lO_2}}{T + f} + O_{2h}. \]  

(3.17)

Using the stoichiometric ratio between oxygen and phosphate, \(O_2:P\) and taking \(O_{2h} = 316 \text{ mmol/m}^3\), \(PO_{4h} = 1.3 \text{ mmol/m}^3\) [1]

\[ O_{2d} = O_{2h} + rO_2 : P(PO_{4d} - PO_{4h}) \\
= 316 - 170 \times (2.1 - 1.3) \\
= 180 \text{ mmol/m}^3. \]  

(3.18)

This is close to the observed value and shows the viability of the model [1]. From equation 3.9 for the time rate of change of phosphate in the first box at steady state, and equation 3.11 for phosphate, we have:

\[ PO_{4h} = \frac{TP_{O4d} + fPO_{4d} - P_{hPO_4}}{T + f}. \]  

(3.19)

If we strengthen the biological pump by increasing \(P_{hPO_4}\), \(PO_{4h}\) decreases, meaning that there is more uptake of DIC from the second to the third box. Also a variation of \(P\) or \(f\) implies a variation of \(PO_{4h}\).

This model shows the importance of the high-latitude ocean in the uptake of carbon dioxide from the surface to the ocean interior, and also the role of the thermohaline circulation. As we said before, Sarmiento and Toggweiler [3] showed using this model that the drop in \(pCO_2\) during the last ice age was the consequence of a strong biological pump. This must have resulted as shown by equation 3.19 in a decrease of \(PO_{4h}\). Also it happens that, for the DIC to be stored for a long time in the deep ocean, it is necessary to have a slow circulation so that the DIC in the ocean interior will not return quickly to the surface where it will escape back to the atmosphere.

Now we know the reason for the cool weather during the last ice age. The question is to understand the phenomenon responsible for it. One idea is that the strength of the biological pump during the last ice age was due to the presence of more iron dust in the ocean [1][11]. Unfortunately, this model cannot provide us with more insight about iron fertilization experiments. We need to move to a more detailed one. Specifically, we need to separate the high latitude box into two, the North Atlantic and the Southern Ocean, because of their different responses to iron fertilization. The result is a five-box model that we shall discuss below.

### 3.3 Five-box model of the ocean carbon cycle

The boxes include (figure 3.4):

1. The s box for the Southern Ocean, where the water from the interior upwells carrying dissolved materials to the surface and creates the Antarctic Bottom Water (AABW) current. Part of this water from the interior advects in the middle ocean.
2. The I box for the low-latitude surface ocean. Here the water is warmer and the photosynthesis process is more effective because of the presence of more light. Meanwhile, the population of plankton in this area is not so high because of the lack of nutrients. It is also the area where the exchange rate between the atmosphere and the surface ocean is more important. Gases such as carbon dioxide and oxygen are released back to the atmosphere easily.

3. The N box for the North Atlantic. Here the water from the surface sinks to the deep ocean, because of the difference in density (due to the temperature and salinity), carrying dissolved materials from the surface to the deep.

4. The D box for the deep ocean. Here the dissolved organic matter can remain for a thousand years [3] before being carried back to the surface by the overturning ocean circulation known as the thermohaline circulation.

5. Finally the M box, for the middle ocean. This part of the ocean is called the thermocline (see figure 3.3). It is a transition layer between the mixed layer (or surface layer) at the surface and the deep water layer, based on temperature. In the thermocline, the temperature decreases rapidly from the mixed layer temperature to the much colder deep water temperature. The mixed layer and the deep water layer are relatively uniform in temperature while the thermocline constitutes a transition between the two.

The circulation pattern of our model is as follows. Water from the deep upwells to the Southern Ocean. It is then advected to the middle ocean, from where it goes to the North Atlantic and sinks back to the deep ocean. Considering mass conservation, the equations for phosphate ($PO_4^-$) and DIC are as follows:

1. In the s box:

   \[
   \begin{align*}
   V_s \frac{dPO_4s}{dt} &= \text{Mix}_s(PO_4d - PO_4s) + \text{Mix}_{ms}(PO_4m - PO_4s) \\
   &\quad + M_s(PO_4d - PO_4s), \\
   V_s \frac{dDIC_s}{dt} &= \text{Mix}_s(DIC_d - DIC_s) + \text{Mix}_{ms}(DIC_m - DIC_s) \\
   &\quad + M_s(DIC_d - DIC_s) + \Phi_{CO_2s}. \\
   \end{align*}
   \]  

2. In the I box:

   \[
   \begin{align*}
   V_l \frac{dPO_4l}{dt} &= \text{Mix}_{lm}(PO_4m - PO_4l) + M_l(PO_4m - PO_4l), \\
   V_l \frac{dDIC_l}{dt} &= \text{Mix}_{lm}(DIC_m - DIC_l) + M_l(DIC_m - DIC_l) \\
   &\quad + \Phi_{CO_2l}. \\
   \end{align*}
   \]
Figure 3.3: A simple temperature-depth ocean water profile. We can see that temperature decreases with increasing depth. The thermocline is the layer of water where the temperature changes rapidly with depth. This temperature-depth profile is what we might expect to find in low to middle latitudes.

3. **In the m box:**

\[
\begin{align*}
V_m \frac{dPO_{4m}}{dt} &= \text{Mix}_{ms}(PO_{4s} - PO_{4m}) + \text{Mix}_{lm}(PO_{4l} - PO_{4m}) + \text{Mix}_{md}(PO_{4d} - PO_{4m}) + (M_n + M_l - M_s)PO_{4d} + M_sPO_{4s} - M_lPO_{4m} - M_nPO_{4m}, \\
V_m \frac{dDIC_m}{dt} &= \text{Mix}_{ms}(DIC_s - DIC_m) + \text{Mix}_{lm}(DIC_l - DIC_m) + \text{Mix}_{md}(DIC_d - DIC_m) + (M_n + M_l - M_s)DIC_d + M_sDIC_s - M_lDIC_m - M_nDIC_m.
\end{align*}
\]
Section 3.3. Five-box model of the ocean carbon cycle

Figure 3.4: Five box model of the ocean. n represents the North Atlantic, s the Southern ocean, m the middle ocean, l the low latitude box and d the deep ocean box. $M_n$, $M_l$, $M_l + M_n$, $M_n + M_l - M_s$, $M_s$ are the advection terms. $\text{Mix}_{ns}$, $\text{Mix}_s$, $\text{Mix}_{lm}$, $\text{Mix}_{md}$, $\text{Mix}_n$ represent the diffusion or mixing terms.

4. **In the d box:**

\[
\begin{align*}
V_d \frac{dPO_{4d}}{dt} &= \text{Mix}_s(PO_{4s} - PO_{4d}) + \text{Mix}_{md}(PO_{4m} - PO_{4d}) \\
&\quad + \text{Mix}_n(PO_{4n} - PO_{4d}) - (M_n + M_l - M_s)PO_{4d} \\
&\quad + (M_l + M_n)PO_{4m} - M_sPO_{4d}, \\
V_d \frac{dDIC_d}{dt} &= \text{Mix}_s(DIC_s - DIC_d) + \text{Mix}_{md}(DIC_m - DIC_d) \\
&\quad + \text{Mix}_n(DIC_n - DIC_d) - (M_n + M_l - M_s)DIC_d \\
&\quad + (M_l + M_n)DIC_m - M_sDIC_d.
\end{align*}
\]  

(3.23)

5. **In the n box:**

\[
\begin{align*}
V_n \frac{dPO_{4n}}{dt} &= \text{Mix}_n(PO_{4d} - PO_{4n}) - (M_l + M_n)PO_{4n} \\
&\quad + M_nPO_{4m} + M_lPO_{4d}, \\
V_n \frac{dDIC_n}{dt} &= \text{Mix}_n(DIC_d - DIC_n) - (M_l + M_n)DIC_n \\
&\quad + M_nDIC_m + M_lDIC_l + \Phi_{CO_{2n}}.
\end{align*}
\]  

(3.24)

The question now is to find where to perform iron fertilization in order to decrease the maximum of $CO_2$ from the atmosphere. It is at the Southern Ocean, the North Atlantic or the low latitude surface ocean?
From the circulation pattern of our model, it appears that the warm water from the low latitudes as well as the middle ocean water are advected into the North Atlantic. In this area (the North Atlantic), the water which carries nutrients and DIC from the low latitude and middle ocean sinks into the deep ocean. The result is the transportation of nutrients and DIC in the deep ocean. From this fact, the probability for the Particulate Organic Carbon (POC) from iron fertilization in this region to sink in the deep ocean is high and could result in a net decrease of $pCO_2$ in the atmosphere.

Another important region for iron fertilization could be the Southern Ocean. Here the upwelling of deep water with nutrients and DIC makes this area rich in nutrients. The nutrients can be fixed into organic matter after photosynthesis. The organic matter will then sink or move, driven by the thermocline in the middle ocean or perhaps in the North Atlantic, and return back into the deep ocean. Since the overturning circulation of the ocean is relatively long (1000 years [3]), this could result in a transportation of DIC in the ocean interior.

For the low-latitude zone to be efficient, it is necessary that the advection term $M_l$ be very high. Otherwise, all the POC produced by iron fertilization would be converted back into dissolved organic matter in the surface or sometimes the middle ocean and all the fixed carbon dioxide will return back to the atmosphere. This phenomenon is amplified by the fact that the water here is warm and therefore less soluble for carbon dioxide.

Thus, the best area to experiment with iron fertilization seems to be the North Atlantic and the Southern Ocean.

In the next section, using a code obtained with the use of the systems of equations 3.20, 3.21, 3.22, 3.23, 3.24, we will simulate the iron fertilization experiment in the North Atlantic, the low latitudes and the Southern Ocean. We could therefore verify our hypothesis and give an answer to the question of where to do the experiment.
4. Simulation of iron fertilization with a five-box model

We are going to test the efficiency of iron fertilization in the ocean with the five-box model. We consider the ocean and the atmosphere together as a system in which the total concentration of carbon remains constant. This means that if the concentration of carbon dioxide decreases in the atmosphere, it increases in the ocean and vice versa.

We will perform the simulation by forcing the concentration of nutrients at the ocean surface to decrease to zero. This is an exaggeration of what happens when enough iron is added at the surface of the ocean[13]. Our experiments thus provide an upper limit to the effect of adding enough iron to the ocean surface. The experiment will be performed in the Southern Ocean (experiment 1), the low-latitude surface ocean (experiment 2) and the North Atlantic (experiment 3). The analysis of the numerical output obtained will enable us to find an answer to our question: where should we perform iron fertilization to get atmospheric $pCO_2$ to decrease most?

4.1 Running the simulation

First of all, we set the values of the mixing and advection coefficients to be:

$M_n = 10 \text{ Sv}$, $M_s = 10 \text{ Sv}$, $\text{Mix}_s = 0 \text{ Sv}$, $\text{Mix}_{lm} = 100 \text{ Sv}$, $\text{Mix}_n = 5 \text{ Sv}$, $\text{Mix}_{ms} = 40 \text{ Sv}$, $\text{Mix}_{md} = 0 \text{ Sv}$, $M_l = 0 \text{ Sv}$. This shows that the mixing terms between the Southern and the deep ocean, the middle and the deep ocean as well as the low latitude and the North Atlantic are neglected. The nutrients (phosphate) and DIC are initialized to their mean concentration values, which are: $PO_4 = 2.17 \text{ umol/kg}$ and DIC = 2280 umol/kg. The partial pressure of carbon dioxide in the atmosphere is initialised to its pre-industrial value of 278 ppm [1]. We run the code in three steps:

1. We initialised the concentration of phosphate and DIC to be the same and equal to their mean values everywhere in the ocean. The goal of this step is to redistribute their values in each box according to their different properties. After running the code for 1000 time steps for DIC and phosphate ($PO_4$) in the deep and middle box, each time step representing one year and 160 time steps in the South, the North and the low latitudes box, each time step representing 1 month, we plot the curve of DIC and $PO_4$ with respect to time (figure 4.1). At this stage, we fixed the partial pressure of carbon dioxide in the atmosphere to be constant with time. The concentrations of $PO_4$ and DIC obtained at steady state for each box are close to the real values measured in the ocean. These values increase from the surface to the depth and constitute a good approximation to the actual ocean behaviour. The steady state is obtained faster for the surface ocean (North Atlantic, Southern Ocean and low latitude ocean) than for the middle and the deep ocean. This is due to the fact that the ocean interior is bigger and takes a longer time to adjust. Another observation is

\[^{1}\text{1 Sv} = 10^6 \text{m}^3/\text{s}\]
that it takes a longer time for DIC to equilibrate compared to $PO_4$. The reason for this is the high solubility of DIC which makes it less free to move.

We are now going to test the stability of our model with time. This will constitute the second step of the simulation.

Figure 4.1: Step 1. (a) and (b) represent the variation of DIC and phosphate ($PO_4$) in the s (South), the l (low) and the n (North) box. The steady state is reached after 60 months for DIC and after 20 months for $PO_4$. The concentrations of DIC and $PO_4$ are in mol/kg. At steady state, the concentration of DIC and $PO_4$ are respectively 2150 umol/kg and 1.5 umol/kg in the s box, 2130 umol/kg and 0.55 umol/kg in the n box and 2100 umol/kg and 0 in the l box. (c) and (d) represent the variations of $PO_4$ and DIC in the m (middle) and d (deep) boxes. The concentration of DIC and $PO_4$ at steady state are respectively 2170 umol/kg and 0.9 umol/kg for the m box and 2300 umol/kg and 2.4 umol/kg in the d box.
2. This step is called the control simulation. We are going to test the stability of our model by observing the variation of the concentration of DIC and $PO_4$ with time. Also, at this stage, we allow the partial pressure of carbon dioxide to vary freely as total ocean-air inorganic carbon is constant.

Figure 4.2 shows the result obtained. We can see that the concentrations of various elements remain constant and are equal to the values we had at steady state at the previous step, showing the stability of our model. Also the concentration of carbon dioxide in the atmosphere changes very little— we have a variation of 4 ppm in 2500 years. This corroborates the observed constant concentrations of DIC in the n, s, and l box. Now that we have the concentrations of DIC and phosphate in the different areas of the ocean, we...
can start our fertilization experiment. This will constitute the final step.

3. The values of DIC and phosphate obtained at steady state during the previous step are used here as initial values. We force nutrients (phosphate) to zero, by imposing a strong productivity of organic matter due to photosynthesis. This productivity is set by the following equation:

\[
\text{production} = \frac{PO_{4\text{surface}} - P^*}{\Delta t},
\]

where \(PO_{4\text{surface}}\) is the concentration of phosphate at the surface in umol/kg and \(P^*\) is the observed \(PO_4\) concentration. For the fertilization experiment, we set \(P^* = 0\) in the region of depletion, while setting \(P^*\) to its observed value in the rest of the ocean. Figures 4.3 and 4.4 show the curves of DIC and phosphate (\(PO_4\)) we obtained after simulating the fertilization in the North Atlantic, the low-latitudes surface ocean and the Southern Ocean. We can see that the effect of the productivity is to decrease gradually the nutrient concentration in each fertilization area. The concentration of DIC and \(PO_4\) in the deep ocean remains constant. This is due to the fact that the deep ocean represents more than 80% of the total ocean volume. As a consequence, it acts like a reservoir. The supply of nutrients from the surface due to fertilization is less significant compared to the total quantity of nutrients in the deep ocean.

4.2 Discussion of the results

4.2.1 Fertilisation in the low latitude

Figure 4.5 shows a decrease in \(pCO_2\) from 278 ppm to 274 ppm after 2500 years of simulation, when the curve achieves steady state. This result shows that fertilization in the low latitudes surface ocean is not efficient. The reason for this is the poor concentration of nutrients. From chapter 2 we learnt that concentrations of phosphate and nitrate in this area are close to zero, due to the high mixing with the middle ocean, resulting in the transportation of nutrients into the ocean interior (the mixing term between the low and the middle ocean is 100 Sv and represents the highest exchange rate). As a consequence, the population of phytoplankton is very low despite the presence of enough light and iron brought by the dust from large deserts such as the Sahara. Our result suggests that an addition of iron will simply be a waste.

4.2.2 Fertilization in the North Atlantic

Figure 4.6 (a) shows the variation of atmospheric \(CO_2\) after depletion in the n box. We observe a decrease of 35 ppm in the partial pressure of \(CO_2\) from 280 to 245 ppm at steady state. It is difficult to have a better result because of the advection from the middle ocean (\(M_n = 10\) Sv) which brings back nutrients at the surface and thus decreases the overall productivity of the area. Meanwhile, this result is much better than what we had for the low latitudes where nothing significant happens.
Figure 4.3: Variation of $PO_4$ and DIC due to fertilization. We can observe the depletion of nutrients in the n box. In the l box everything remains constant.

We have 267 ppm of atmospheric partial pressure after 150 years of fertilization, which is 6 ppm closer to the 260 ppm observed after the last ice age [5]. This result is pretty good; but it is the best? We are now going to test the fertilization in the Southern Ocean.
4.2.3 Fertilization in the Southern Ocean

The variation of the atmospheric $CO_2$ in the s box is shown in figure 4.6 (b). The graph shows a variation of the partial pressure, from 278 ppm to 246 ppm at steady state. We can decompose this in three parts: firstly, we have a strong drop from 278 ppm to 255 ppm, with a slope close to the vertical, during the first hundred years. Then, the curve becomes smoother with a drop of the partial pressure from 255 ppm to 247 ppm after 1500 years. For the last part of the curve the slope is close to the horizontal, meaning that it is approaching steady state. These successive changes can be explained by the DIC behaviour in the Southern Ocean (see figure 4.4). Since the advection from the deep to the middle ocean is zero \( M_n + M_l - M_s = 10 - 10 = 0 \), all the sinking water from the North Atlantic is advected in the Southern ocean, bringing more DIC. Part of this moves to the middle ocean and another part equilibrates after a period of time with the DIC taken up by the biological pump. Due to the difference in time frame, when the Southern Ocean is fertilized, the DIC is removed quickly during the first year (see figure 4.4) and equilibrates later with that advected from the ocean interior, corresponding to the change of slope for the atmospheric $pCO_2$ (figure 4.6 (b)).

We observe a decrease of the partial pressure from 278 ppm to 250 ppm over 150 years, which gives a drop of 28 ppm on figure 4.6 (b). This is more than the double what we had for the Northern Atlantic simulation, and shows the strength of the fertilization in the Southern Ocean.

Figure 4.7 shows atmospheric $pCO_2$ decreasing in all three fertilization scenarios for 150 years. The efficiency of the Southern Ocean in sequestering atmospheric $pCO_2$ is more than twice that of the North Atlantic.

It appears that the uptake of $CO_2$ in the Southern Ocean is very efficient for the first 500 years of fertilization, as compared to the North Atlantic and the low latitudes. In particular, after 150 years of experiment, we have a drop of 28 ppm of atmospheric $pCO_2$ for the Southern Ocean, only 11 ppm for the North Atlantic and almost nothing for the low latitudes. Meanwhile, after 2500 years of fertilization, when both curves have reached steady state, we observe the same drop of atmospheric $pCO_2$ for the North Atlantic and the Southern Ocean, but this time frame is not interesting because it is very long. The reasons that the Southern Ocean and the North Atlantic are more efficient in the absorption of atmospheric $CO_2$ is that they are areas of deep water formations and contain more unutilized nutrients compared to the low-latitude surface ocean. The more the nutrients at the ocean surface, the stronger the biological production and the taken up of $CO_2$ from the atmosphere to the ocean.

Consequently, fertilization in the Southern Ocean presents the fastest solution in the uptake of $CO_2$ in the atmosphere and is therefore the most effective.

Throughout this study, we have consider the parameters of the ocean (mixings and advections terms) to be constants. What happens if those parameters change? It is clear today that, the change of climate during the glacial-intergalacial period was not only a consequence of a strong productivity of the ocean surface due to iron dust, but also a change in the mixings and advections terms between the high latitudes surface ocean and the deep ocean. We are now going to test the behaviour of atmospheric carbon dioxide after a change in the ocean parameter. This is called the sensitivity studies.
4.3 Sensitivity studies of the ocean

We perform Southern Ocean fertilization for a range of $\text{Mix}_s$ (0, 20 Sv, 40 Sv, 60 Sv, 80 Sv) and $\text{Mix}_n$ (0, 15 Sv, 30 Sv, 45 Sv) values. Figures 4.8 (a) and (b) show the output of the simulation. It appears that increasing mixing between the Southern and deep ocean boxes results in a decrease of the atmospheric partial pressure of $CO_2$, meaning that more carbon dioxide is taken up from the atmosphere by the ocean. Atmospheric $pCO_2$ decreases by 33 ppm for $\text{Mix}_s = 0$, by 40 ppm for $\text{Mix}_s = 20$ Sv and by 55 ppm for $\text{Mix}_s = 80$ Sv.

After 150 years, changing mixing between the North Atlantic and the deep ocean ($\text{Mix}_n$) has little impact on atmospheric $pCO_2$. Meanwhile, a change may occur after a longer time frame. This difference in the sequestration of carbon dioxide after changing $\text{Mix}_s$ and $\text{Mix}_n$ can be explained by the size of the Southern Ocean which represents 0.22% of the total surface ocean, and which is very big compared to the 0.01% for the North Atlantic and by the larger quantities of nutrients at the Southern Ocean surface compared to the North Atlantic. Therefore, the Southern Ocean contains and transports more nutrients and DIC than the North Atlantic. A change of in Southern Ocean parameters results in a more significant atmospheric $CO_2$ drawdown.

The values of the parameters that we chosen for the fertilization simulation, $\text{Mix}_s = 0$ and $\text{Mix}_n = 5$ Sv were guessed and are not probably the most suitable ones. However, our results suggest the critical importance of the Southern Ocean in the absorption of carbon dioxide from the atmosphere.
Figure 4.4: Variation of $PO_4$ and DIC due to fertilization. We observe the depletion of nutrients in the s box. In the m and the d box, the concentration of phosphate ($P_{mid}$, $P_{deep}$) and DIC ($DIC_{mid}$, $DIC_{deep}$) is almost constant after 200 years. For the fertilization in the s box, we observe an increase of the concentration of DIC from 10 months onward. This is due to the upwelling of DIC from the deep ($M_s = 10$ Sv) which balances with the uptake of DIC due to fertilization.
Figure 4.5: We can see that the variation of $CO_2$ is the same in the control simulation (a) and in the fertilization of the low latitudes surface ocean (b).

Figure 4.6: (a) We have a decrease of the partial pressure of 11 ppm after 150 years and 35 ppm after 2500 years, which is very much greater than the 4 ppm drop during the control simulation (figure 4.5 (a)). (b) We observe a decrease of the partial pressure of 28 ppm after 150 years, and 35 ppm after 2500 years.
Figure 4.7: S represent the variation of the partial pressure of CO\textsubscript{2} in the atmosphere due to fertilization in the Southern Ocean, L for the low-latitude ocean and N for North Atlantic.

Figure 4.8: Mix\textsubscript{s,0}, Mix\textsubscript{s,20}, Mix\textsubscript{s,40}, Mix\textsubscript{s,60} and Mix\textsubscript{s,80} represent respectively Mix\textsubscript{s} = 0, 20, 40, 60 and 80 Sv. Mix\textsubscript{n,0}, Mix\textsubscript{n,15}, Mix\textsubscript{n,30} and Mix\textsubscript{n,45} represent respectively Mix\textsubscript{n} = 0, 15, 30 and 45 Sv. (a) The atmospheric carbon dioxide decrease more with increasing values of Mix\textsubscript{s}. (b) The decrease of atmospheric carbon dioxide remains the same while changing Mix\textsubscript{n} from 0 to 45 Sv after 150 years of fertilization in the Southern Ocean.
5. Conclusion

From our study it appears that the low-latitude surface ocean is not an area to be considered for the fertilization experiments. This area, despite the presence of sunlight and enough iron brought naturally by large deserts, has the lowest abundance of nutrients and chlorophyll (see chapter 2.2). An addition of iron will simply be a waste. The Northern Atlantic is more interesting in that it is possible to sequester up to 11 ppm of atmospheric $CO_2$ in 150 years. The Southern Ocean is the most suitable area with a decrease of $CO_2$ partial pressure in the atmosphere of 28 ppm after 150 years of continuous fertilization. Meanwhile, for a longer time frame, both the North Atlantic and the Southern Ocean give the same results (we have a drop of 35 ppm of atmospheric $pCO_2$ after 2500 years). The reasons that the Southern Ocean and the North Atlantic are more efficient is that they are areas of deep water formation and contain more unutilized nutrients, which could be used during fertilization to fix more $pCO_2$ than at the low-latitude surface. The Southern Ocean is more efficient than the North Atlantic because it is bigger and has the highest surface nutrients. For each experiment, we forced a strong biological productivity, dropping the concentration of nutrients to zero in the area of fertilization. It is very difficult to achieve that in reality but such simulations give an understanding of the ocean behaviour. The results we obtained are the outcome of the assumptions we made for the values of the mixing and the advection parameters. A change of these values may result in a different response of the ocean.

Despite the uncertainty about the mixing parameters, the Southern Ocean has a key role in the sequestration of carbon dioxide, because of its high surface nutrient concentration and its important role in the formation of deep water.

It is now clear that if we consider seeding the ocean with iron as a quick solution to stop global warming, the Southern Ocean is the zone to consider, though we cannot predict, for a given period, the quantity of $CO_2$ which will be removed from the atmosphere because of the simplicity of our model. Another question then arises: for how long should we do the experiment? In other words, if we stop the fertilization would the partial pressure of $CO_2$ in the atmosphere return to its previous value? What would happen to all that iron and what would be the ecological consequences? These are the questions we cannot give an answer to now, but shall consider for further studies.
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Bibliography


